

The deposition and alteration history of the northeast Syrtis Major layered sulfates

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January 7, 2019

Abstract

Ancient stratigraphy on Isidis Basin's western margin records the history of water on early Mars. Noachian units are overlain by layered, basaltic-composition sedimentary rocks that are enriched in polyhydrated sulfates and capped by more resistant units. The layered sulfates – uniquely exposed at northeast Syrtis Major – comprise a sedimentary sequence up to 600-m thick that has undergone a multi-stage history of deposition, alteration, and erosion. Siliciclastic sediments enriched in polyhydrated sulfates are bedded at m-scale and were deposited on slopes up to 10°, embaying and thinning against pre-existing Noachian highlands around the Isidis basin rim. The layered sulfates were modified by volume-loss fracturing during diagenesis. Resultant fractures hosted channelized flow and jarosite mineral precipitation to form resistant ridges upon erosion. The structural form of the layered sulfates is consistent with packages of sediment fallen from either atmospheric or aqueous suspension; coupling with substantial diagenetic volume-loss may favor deepwater basin sedimentation. After formation, the layered sulfates were capped by a “smooth capping unit” and then eroded to form paleovalleys. Hesperian Syrtis Major lavas were channelized by this paleotopography, capping it in some places and filling it in others. Later fluvial features and phyllosilicate-bearing lacustrine deposits, sharing a regional base level of ~2300 m, were superimposed on the sulfate-lava stratigraphy. The layered sulfates suggest surface bodies of water and active groundwater upwelling during the Noachian–Hesperian transition. The northeast Syrtis Major stratigraphy records at least four distinct phases of surface and subsurface aqueous activity spanning from late Noachian to early Amazonian time.

Key Points

- The eroded stratigraphy at Northeast Syrtis Major includes m-scale layered, sulfate-enriched sedimentary basin deposits up to 600 m thick.
- Layer dips <10° and topographic-high embayment are consistent with deepwater deposition; volume-loss fractures indicate diagenetic dewatering.
- Strata also including Syrtis Major lavas and fluvio-deltaic phyllosilicates record ≥3 fluvio-lacustrine episodes into the early Amazonian.

Contents

1	Introduction	3
2	Geologic context	4
2.1	Physiography	4
2.2	The Northeast Syrtis Noachian plains	6
2.3	Layered sulfates	6
2.4	The Syrtis Major lavas	9
3	Methods	9
3.1	Conceptual Approach: Formation mechanisms for the layered sulfates	9
3.2	Digital elevation models and dataset registration	12
3.3	Regional mapping	14
3.4	Orientation measurements	16
4	Results	18
4.1	The layered sulfates	18
4.1.1	Basic morphological and mineralogical characteristics	18
4.1.2	Extent	18
4.1.3	Bedding orientations	19
4.2	Boxwork polygons	23
4.2.1	Fracture morphology	23
4.2.2	Fracture orientations	25
4.2.3	Dip changes at small spatial scale	25
4.2.4	Fracture mineralogy	27
4.3	Penetrative alteration	29
4.4	Capping units	29
4.4.1	Morphology	29
4.4.2	Structural characteristics	31
4.4.3	Spectral and thermophysical characteristics	33
4.4.4	Thickness of the smooth capping surface	33
4.4.5	Relationship with the Syrtis Major lavas	33
4.5	Late fluvial history	35
4.5.1	Channels atop the lava flows	35
4.5.2	Fluvial and lacustrine deposits	35
4.5.3	Basin-floor deposits	37
4.5.4	Draping valley fill	40
4.5.5	Elevation alignment of fluvial features	40
5	Discussion	42
5.1	Deposition of the layered sulfates	42
5.2	Post-depositional alteration of the layered sulfates	44
5.2.1	Interpretation of boxwork fractures	44
5.2.2	Implications for sediment size and amount of burial	45

5.2.3	Mineralization of fractures and implications for water volume	46
5.3	Comparison with other Martian layered deposits	47
5.4	Containing basin for the layered sulfates	48
5.5	Modification by fluvial erosion, lavas, and late lake deposits	49
6	Summary, Conclusions, and Future Work	51
7	Acknowledgements	54
	Bibliography	55
Table 1	HiRISE and CTX scenes used in elevation models	13
Figure 1	Context map	5
Figure 2	Elevation datasets	7
Figure 3	Geologic map	8
Figure 4	Depositional settings	11
Figure 5	Representative CRISM spectra	15
Figure 6	Key morphologies	17
Figure 7	Layer thicknesses	18
Figure 8	Structural model of sulfate thickness	20
Figure 9	Cross-sections of structural model	21
Figure 10	Regional orientation measurements	22
Figure 11	Southeast-dipping stratigraphy	24
Figure 12	Boxwork examples	26
Figure 13	Boxwork orientations	27
Figure 14	Attitudes within boxwork	28
Figure 15	Throughgoing alteration	30
Figure 16	Cap unit surface	31
Figure 17	Cap unit thickness	32
Figure 18	Thermal inertia	34
Figure 19	Cap unit relationships	36
Figure 20	Delta and fluvial systems	38
Figure 21	Basin Floor deposits	39
Figure 22	Draping unit	41
Figure 23	Boxwork schematic	45
Figure 24	Emplacement and alteration history	52

1 Introduction

A major focus of Mars science is deciphering the nature and drivers of changing environmental conditions during the planet's early history. Orbital observations of late-Noachian fluvial and alluvial systems (e.g. *Howard et al.*, 2005; *Irwin et al.*, 2005; *Fassett and Head*, 2005; *Andrews-Hanna and Lewis*, 2011; *Schon et al.*, 2012) and phyllosilicate and carbonate alteration of igneous bedrock by surface and subsurface waters (e.g. *Bibring et al.*, 2006; *Ehlmann et al.*, 2008a, 2011; *Mustard et al.*, 2009; *Murchie et al.*, 2009), as well as *in-situ* examination of lacustrine deposits and groundwater diagenesis by the *Opportunity* and *Curiosity* rovers (*Squyres et al.*, 2004; *Grotzinger et al.*, 2013), present powerful evidence of the active role of liquid water on early Mars. From this initial active hydrosphere, the Martian climate underwent a secular drying through the Hesperian period to arrive at the cold, arid environment that prevailed through most of the Amazonian period. However, the timing and character of this global shift is unclear.

The character of environmental change during the Noachian and Hesperian is best constrained by analyses of stratigraphic sections that span portions of this time interval. Stratified sedimentary deposits with hydrated minerals stand out as key environmental records (*Gendrin et al.*, 2005; *Niles and Michalski*, 2009; *Milliken et al.*, 2010; *Ehlmann and Mustard*, 2012; *Grotzinger et al.*, 2012). The stratigraphy exposed at the northeast margin of the Syrtis Major lavas [Figure 1, NE Syrtis] presents an opportunity to examine environmental change from approximately the early Noachian to the Hesperian, and possibly Amazonian, in a sequence of geologic units whose timing is well-constrained. The lower units were emplaced after the Isidis basin-forming impact around 3.9 Ga (*Mangold et al.*, 2007; *Mustard et al.*, 2007, 2009) and are capped by Hesperian Syrtis Major lavas (*Hiesinger and Head*, 2004). The units record a characteristic change in Martian igneous materials, including a transition from low-Ca pyroxene bearing units to high-Ca pyroxene bearing units as well as a high Fo#, olivine-enriched deposit whose formation is related to the Isidis impact (*Mustard et al.*, 2005, 2007, 2009; *Koeppen and Hamilton*, 2008; *Baratoux et al.*, 2013). Importantly, the NE Syrtis stratigraphy contains most of the hydrated mineral diversity recognized on Mars in an organized stratigraphic sequence. Layered sulfates with jarosite ridges are superposed over carbonate-bearing units, which are superposed over clay-bearing units (*Ehlmann and Mustard*, 2012). This stratigraphic sequence records a transition from neutral-alkaline (clay- and carbonate-forming) to acidic (iron sulfate-forming) aqueous environments that corresponds to a global pattern indicating increasing aridity (*Bibring et al.*, 2006). Thus, the mesas of NE Syrtis stratigraphy represent a rare temporally-constrained and in-place record of changing hydrological conditions during the Noachian–Hesperian transition.

The thick layered sedimentary sulfates represent a major change in formation style from the impact- and volcanically-emplaced units dominating the rest of the stratigraphy (*Ehlmann and Mustard*, 2012; *Bramble et al.*, 2017). The lower Noachian clay and carbonate units, exposed regionally to the north over an area spanning hundreds of thousands of square kilometers, have

38 been well-studied by prior workers (e.g. *Mangold et al.*, 2007; *Ehlmann et al.*, 2009; *Mustard et al.*,
39 2009; *Michalski et al.*, 2010; *Viviano-Beck et al.*, 2014; *Ehlmann and Edwards*, 2014). However, only
40 the basics of sulfate mineralogy (polyhydrated sulfates, jarosite), texture (ridged, layered), and
41 stratigraphic position have been previously reported (*Ehlmann and Mustard*, 2012). What is the
42 extent of this sulfate unit? How did it form? What controls the layering, ridges, and specific sul-
43 fate mineralogy? These questions hold particular significance because at the time of this writing,
44 the NE Syrtis landing site is under consideration by the Mars-2020 rover mission and the sulfates
45 and Syrtis Major lavas are the key extended mission target.

46 In this paper, we examine the structural geology of the layered sulfates at NE Syrtis to deter-
47 mine their emplacement mechanism. We comprehensively map the sulfate unit's extent, thick-
48 ness, bedding characteristics, ridge characteristics, and mineralogy. We further examine the con-
49 tact relationships with units above and below, evaluate the capping materials, and determine the
50 temporal relationship with regional fluvial features. We then evaluate these observations of the
51 NE Syrtis Major layered sulfates critically against the range of possible formation mechanisms
52 and propose a multistage formation and modification history that implies a significant role for
53 water on the surface of Mars over a long period of time.

54 **2 Geologic context**

55 **2.1 Physiography**

56 The study area is situated on the western rim of Isidis Basin, about 40 km southwest of Jezero
57 crater and along the northeastern margin of the Syrtis Major volcanic province [Figure 1 and
58 Figure 2]. The layered sulfates are exposed just inside the sharp topographic inflection that marks
59 the 1100-km diameter inner ring of Isidis Basin, as defined by *Mustard et al.* (2007), based on the
60 concentric tectonic expressions of post-Isidis faulting that comprise the Nili Fossae (*Wichman and*
61 *Schultz*, 1989; *Ritzer and Hauck*, 2009).

62 Broadly, both the Syrtis Major lavas south of the study area and the bedrock peneplain of
63 Noachian units extending north of the study area gently slope into Isidis basin. East of the study
64 area, elevations decline into the knobby plains of Isidis basin and the Vastitas Borealis formation
65 (*Ivanov and Head*, 2003; *Ivanov et al.*, 2012).

66 In contrast, the study area itself contains a relatively abrupt topographic step from highland
67 units at -0.5 to 0 km, which define the inner ring of Isidis basin, to flat-bottomed valleys with
68 floors at less than -2.5 km on the Mars Orbiter Laser Altimeter (MOLA) datum (*North Basin* and
69 *Deep Basin* on Figure 2). The valleys are bounded by exposures of highland crust that form moun-
70 tains up to 1.5 km higher than the surrounding terrain. The degree of east-west topographic vari-
71 ation within the study area contrasts with smooth east-west slopes into the Isidis basin to the
72 north and south. The steep basin rim in our study area could be inherited from basin formation
73 or modified by valley erosion. A major valley cutting across the innermost Nili Fossae graben

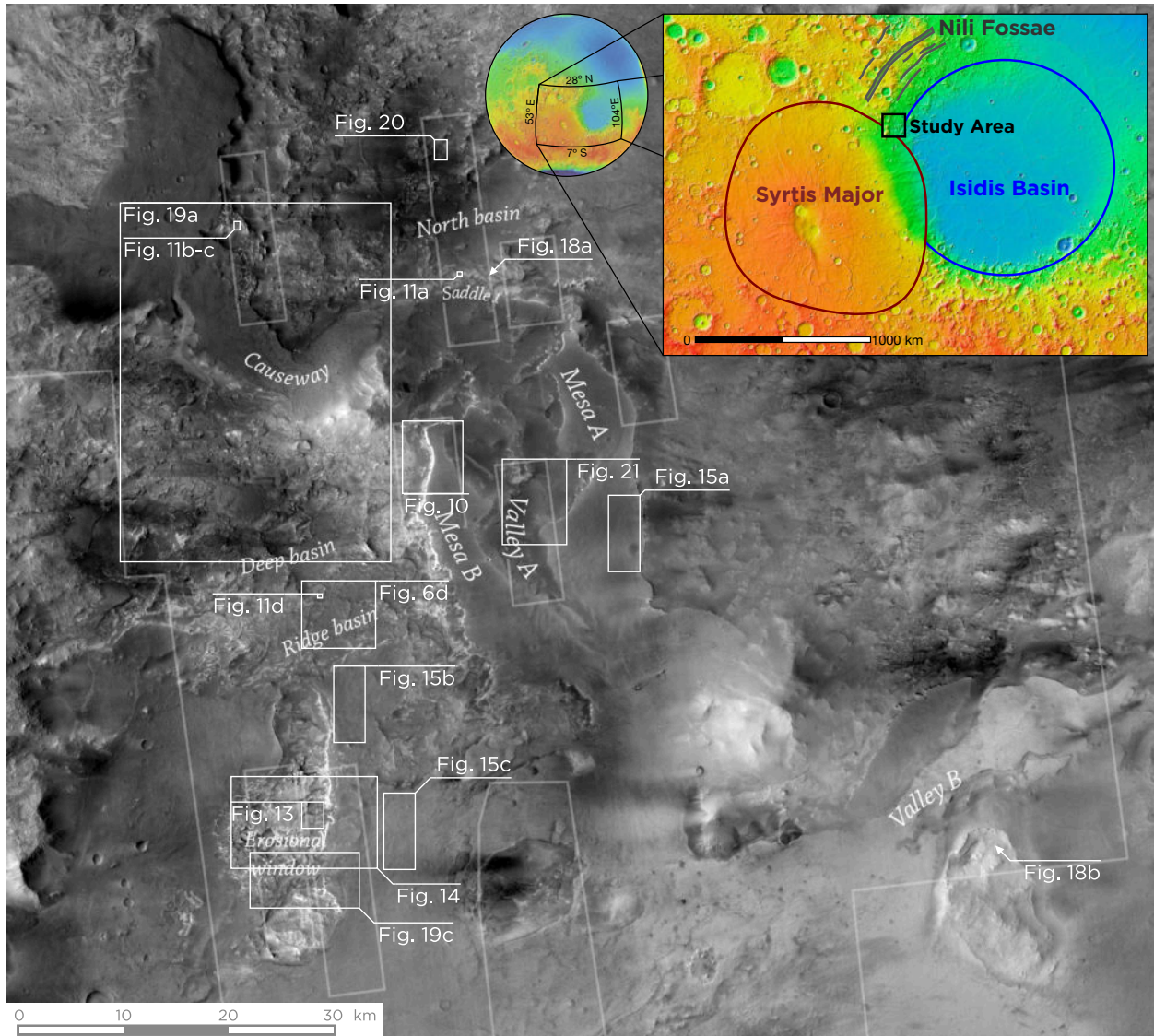


Figure 1: CTX mosaic of the study area showing the location of elevation models and figures referred to in text. The unofficial names used to refer to physiographic features in this study are shown.

74 and Isidis basin rim [Figure 2, nicknamed “I-80” in *Harvey and Griswold* (2010)] channelizes dis-
75 tal Syrtis Major lavas in the northwest portion of the study area. Southeast of the study area,
76 a topographic scarp at ~-3.5 km is cut into the outer edge of the Syrtis Major lavas. This scarp
77 continues southward and represents erosional modification of the basinward edge of the Syrtis
78 Major volcanic province (*Ivanov and Head*, 2003).

79 **2.2 The Northeast Syrtis Noachian plains**

80 We will refer to the bedrock peneplain of Noachian units within the study area as the NE Syrtis
81 plains. The NE Syrtis plains are the lowest exposed stratigraphic units and consist of two litho-
82 logic units. A low-calcium pyroxene- and Fe/Mg phyllosilicate-enriched bedrock (the ‘basement’)
83 comprises the lowermost unit. The phyllosilicate-bearing hydrated basement contains exposures
84 of megabreccia that were formed and/or modified by the Isidis-Basin-forming impact (*Mustard*
85 *et al.*, 2009), which occurred in the Early to Middle Noachian (*Werner*, 2008; *Frey*, 2008). The
86 basement is unconformably overlain by a later-emplaced, olivine-enriched unit partially altered
87 to carbonate (*Hoefen et al.*, 2003; *Hamilton and Christensen*, 2005; *Mangold et al.*, 2007; *Koeppen*
88 *and Hamilton*, 2008; *Mustard et al.*, 2007, 2009; *Ehlmann et al.*, 2008a, 2009). Both of these units
89 are determined to be Noachian in age because they are disrupted by tectonics associated with the
90 opening of the Nili Fossae graben (shortly after Isidis; *Wichman and Schultz* (1989)) and are dis-
91 sected by Noachian-aged valley networks, including the Jezero valley system (*Fassett and Head*,
92 2005, 2008; *Mangold et al.*, 2007; *Ehlmann et al.*, 2008b; *Schon et al.*, 2012; *Goudge et al.*, 2015).
93 These Noachian units extend over more than 100 000 km² in the northwest part of Isidis Basin.

94 *Bramble et al.* (2017) completed a geomorphic mapping study of the beveled NE Syrtis plains
95 directly to the north of the study area (in the area of interest for the candidate Mars 2020 landing
96 site), focusing on relationships between the Nili Fossae basement, olivine-carbonate unit, and
97 overlying capping mesas. Though these two units in the NE Syrtis plains are clearly distinct and
98 well-resolved from orbital mapping, this study treats them collectively as “Noachian Plains” units
99 stratigraphically below the sulfates.

100 **2.3 Layered sulfates**

101 The geologic features targeted by this study are layered basaltic-composition materials enriched
102 in polyhydrated sulfates relative to other materials. These sulfate-bearing layered deposits are re-
103 ferred to as the “layered sulfates.” The unit is preserved and exposed only within this study area, at
104 the NE margin of the Syrtis Major lava flows; it unconformably overlies the Noachian plains units
105 (*Ehlmann and Mustard*, 2012). The sulfates are layered at meters-scale, thick (>300 m in several
106 places), and exposed recessively beneath a cliff-forming capping unit(s), previously interpreted
107 to be the Syrtis Major lavas. These layered domains have mineral signatures of polyhydrated sul-
108 fates. In parts of the region, the sulfates are cut by a boxwork of polygonal raised ridges, which
109 show near-infrared spectral signatures of jarosite, signifying acidic (pH < 4) aqueous conditions

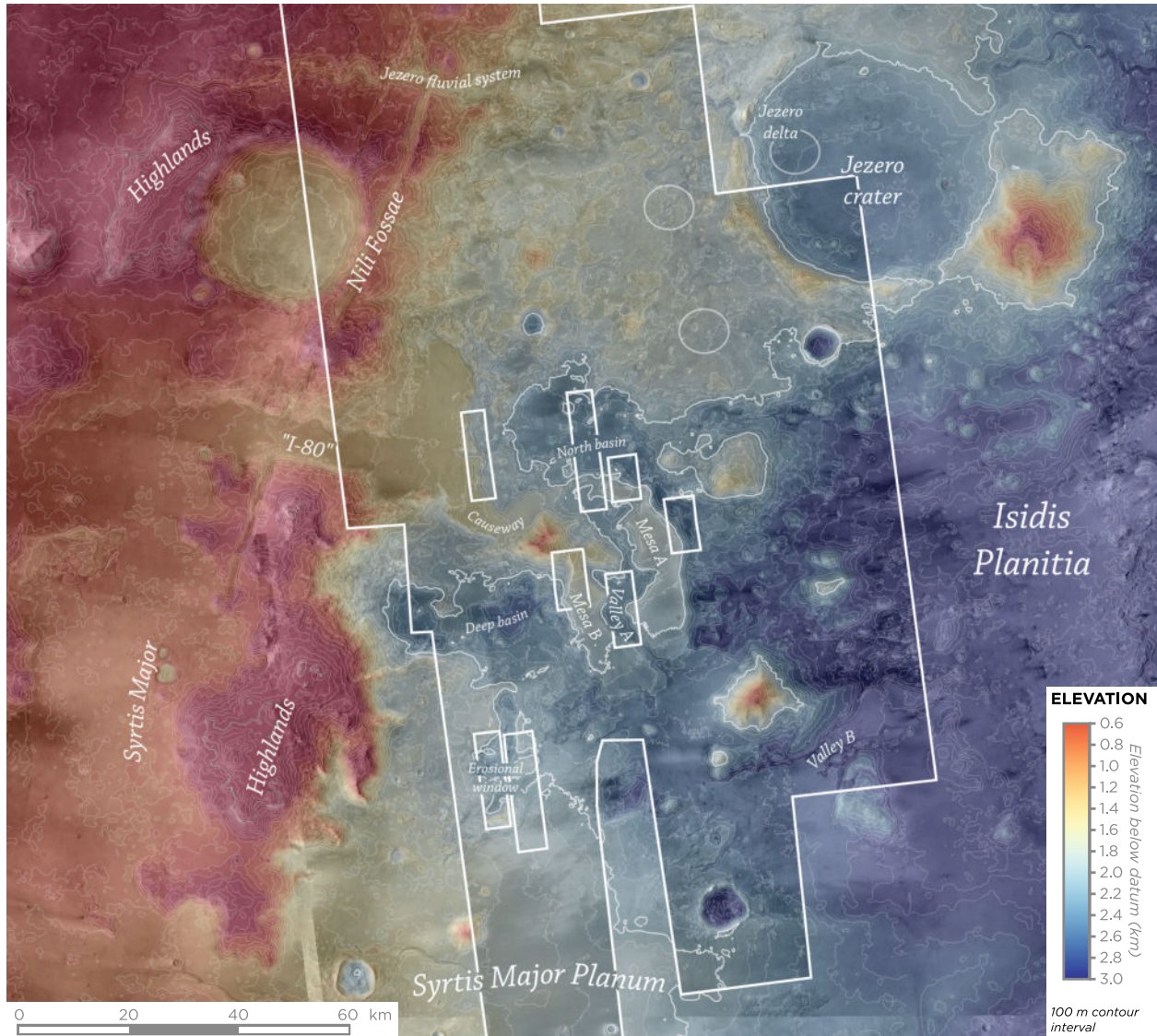


Figure 2: Context map of the NE Syrtis region. The white outlines nf 1-nf8 and the larger regional outline show the nine elevation datasets prepared for the study using HiRISE and CTX stereo images, respectively [Table 1]. A CTX mosaic basemap is color shaded by elevation to show the broad topographic context of NE Syrtis. For elevation shading, our study-created elevation models are supplemented by HRSC elevation models outside the study area. The names of major physiographic features, including unofficial names assigned in this study, are shown. The working landing ellipses for the Jezero and NE Syrtis landing sites (as of late 2017) are also shown.

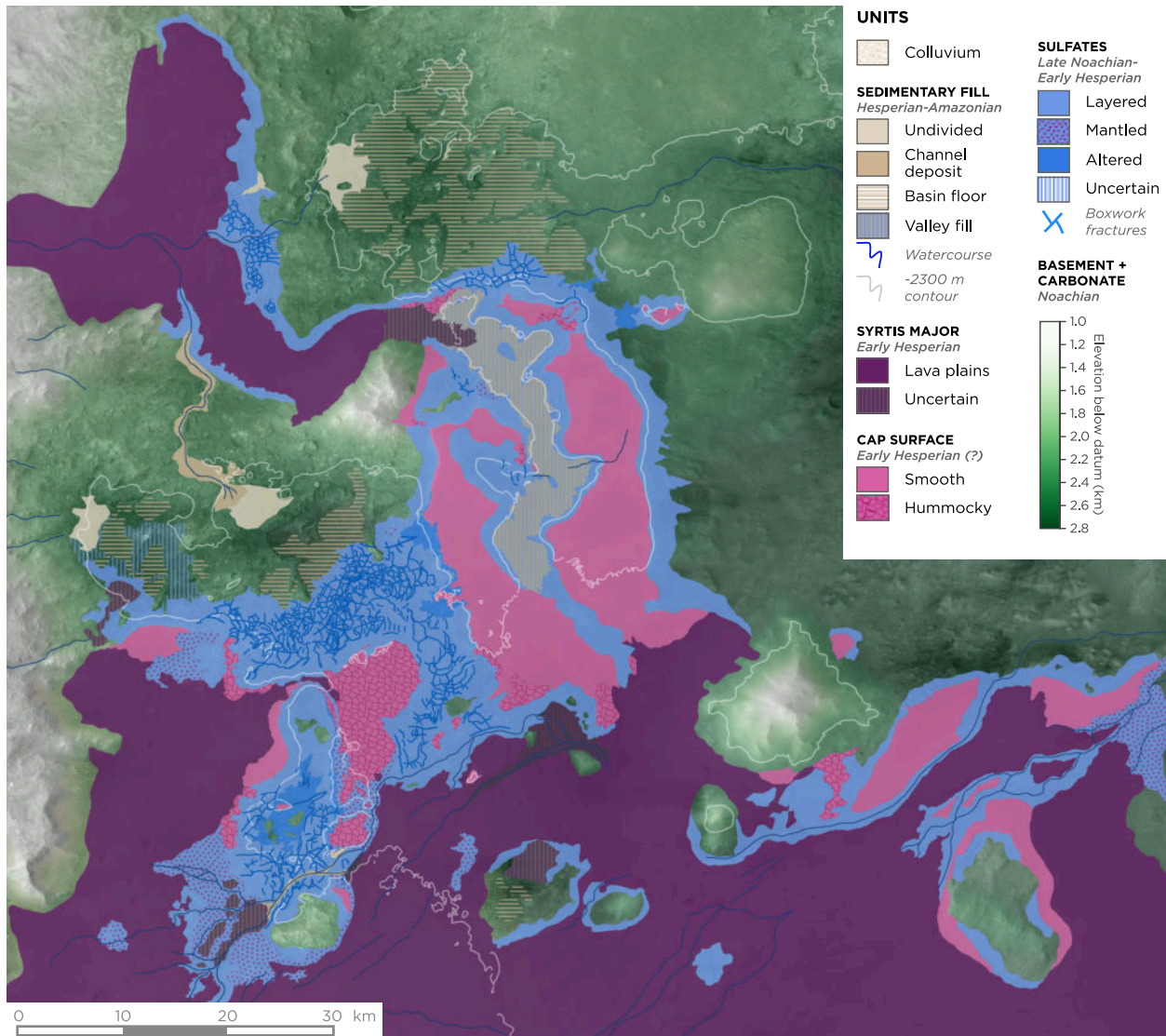


Figure 3: Geologic map of the layered sulfates, Syrtis Major lavas, and capping and sedimentary units discussed in this study. This map compliments that of *Bramble et al. (2017)*, which focuses on basement morphologic units in the Northeast Syrtis plains, just to the north of this map area. Here, the basement and olivine-carbonate unit of the northeast Syrtis regional stratigraphy are undivided, while sulfates, capping materials, and Hesperian/Amazonian sedimentary deposits are detailed.

110 and a distinctive change relative to the clay and carbonate minerals formed earlier in the NE Syr-
111 tis plains (Ehlmann and Mustard, 2012). The sulfates are exposed at the southeast margin of the
112 area mapped by Bramble *et al.* (2017) (their *feature-bearing slope* and *raised boxwork ridges*). The
113 formation of these layered sulfates and their modification by subsequent events are the focus of
114 this work.

115 **2.4 The Syrtis Major lavas**

116 To the south, the Isidis rim is covered by the Syrtis Major volcanic province, a ~1100 km (E-W)
117 region of effusive lava flows averaging ~500 m thick (Hiesinger and Head, 2004) which extends
118 eastward into Isidis basin. Syrtis Major lava flows are sourced in the vicinity of Nili Patera and
119 Meroe Patera near 8°N, 67°E (Fawdon *et al.*, 2015) and descend westward into Isidis Basin, ex-
120 tensively blanketing its rim to the south of the study area (Ivanov *et al.*, 2012). The Syrtis Major
121 edifice has been dated to the early Hesperian by crater counting, with model ages ranging from
122 3.4 Ga (Skok *et al.*, 2010) to 3.6 Ga (Hiesinger and Head, 2004). The Syrtis Major lava flows are
123 enriched in high-Ca pyroxene and distinct from the low-Ca pyroxene basement (Baratoux *et al.*,
124 2007; Skok *et al.*, 2010; Clenet *et al.*, 2013), but in contrast to the Noachian plains to the north,
125 no crystalline hydrous minerals are seen in orbital infrared imagery of the Syrtis Major lavas.

126 The Syrtis Major lavas are the stratigraphically highest unit in portions of our study area, and
127 their relatively unaltered character suggests that they postdate pervasive aqueous alteration in
128 the region. Nevertheless, the margin is eroded by numerous fluvial channels and valleys that
129 point to Hesperian and/or Amazonian surface waters, at least episodically after lava emplace-
130 ment. Mangold *et al.* (2008a) identified outflow channels inscribed on the surface of the Syrtis
131 Major lava flows near the southern margin of the study area, where canyons and channels are cut
132 into the edge of the Syrtis Major lava plains. These form an outflow system originating west of
133 the study area and flowing south and east towards Isidis basin, demonstrating hydrologic systems
134 postdating the emplacement of the Syrtis Major sequence.

135 **3 Methods**

136 **3.1 Conceptual Approach: Formation mechanisms for the layered sulfates**

137 The polyhydrated sulfates within the layered sulfate unit are not indicative of specific aqueous
138 geochemical conditions (Ehlmann and Mustard, 2012), and jarosite within the ridges indicates
139 only precipitation from acidic waters (Papike *et al.*, 2006; Ehlmann and Mustard, 2012). Precipi-
140 tation of sulfate minerals can occur both subaerially during evaporitic deposition (e.g. Hurowitz
141 *et al.*, 2010) and during “alteration” by circulating groundwater, either by bulk reaction or pore-
142 occluding cementation (e.g. Siebach and Grotzinger, 2014). Consequently, both polyhydrated sul-
143 fates and jarosite-bearing ridges may record an alteration signature and do not uniquely define
144 the original depositional environment for these sulfate-bearing sediments.

145 Examination of physical characteristics of the layered sulfates provides a separate set of met-
146 rics for use to understand the formation and evolution of the unit. A wide range of potential
147 mechanisms has been invoked for the deposition of layered rocks on Mars (*Grotzinger and Mil-*
148 *liken, 2012*), each of which possesses distinctive structural characteristics that are potentially
149 observable at orbital scale [Figure 4].

150 Volcanic origin scenarios such as lava flows, ash flows, and ash falls have been proposed for
151 layered deposits elsewhere on Mars (e.g. *Kerber et al., 2012; McCollom et al., 2013*). These em-
152 placement mechanisms do not require abundant surface water; their potential alteration to sul-
153 fates could be enhanced by the circulation of volcanic hydrothermal fluids (e.g. *Kaasalainen and*
154 *Stefánsson, 2011*). Lava and ash deposits are typically thick and internally jointed (*Bondre et al.,*
155 *2004*), and lavas are usually erosionally resistant.

156 Layered sedimentary sulfate deposits can form by the primary precipitation of evaporites,
157 such as in playa environments. Alternatively, layered sediments can arise from a variety of phys-
158 ical sediment-transport mechanisms, implying a wide range of depositional settings. Processes
159 such as fluvial, shallow lacustrine, and aeolian sedimentation are dominated by sediment trac-
160 tion; in arid environments, their deposits are often found together and along with evaporites (*Al-*
161 *Masrahy and Mountrney, 2015*). On Mars, rover (e.g. *McLennan et al., 2005; Grotzinger et al., 2015*)
162 and orbital (*Milliken et al., 2014*) studies indicate the presence of sulfates in lacustrine, evapor-
163 ite playa, and eolian deposits. The fallout of sediment from suspension forms a distinct set of
164 deposits. Loess sedimentation implies the fallout of particles from aerial suspension, similar to
165 distal ash fall (e.g. *Smalley, 1966*); “duststones” formed by this process are found on Mars (*Lewis*
166 *et al., 2008; Bridges and Muhs, 2012*). Alternatively, icy particles deposited from aerial suspen-
167 sion could form sublimation residues (*Michalski and Niles, 2012*). Deep lacustrine sedimentation
168 is dominated by sediment density currents and fallout from aqueous suspension (e.g. *Stow and*
169 *Piper, 1984*). The different geological processes that could form the layered sulfates imply radi-
170 cally different surface environments and water budgets at the time of emplacement.

171 Contact relationships and bedding orientations are key measures of the internal geometry of
172 sedimentary sequences that distinguish depositional processes and their timing. Certain types
173 of sedimentary sequences have characteristic limitations on the distribution of bedding orienta-
174 tions (i.e., strike and dip) imposed by their depositional process. Other important criteria include
175 the assessment of sedimentary onlap and downlap onto pre-existing surfaces versus bedding en-
176 tirely concordant or draping topographic highs. Traction-current sediments typically form near-
177 flat layers and onlap pre-existing topography as they aggrade, while fallout of suspended sediment
178 (e.g. ash falls, duststones, and deep lacustrine sediments) form draping, concordant layers. Sub-
179 aqueous basin-margin sedimentation occupies an intermediate case, where dipping sediments are
180 emplaced by both density currents and fallout from suspension; these sediments are sometimes
181 base-concordant but often thin over, onlap, and in some cases embay basement highs (*Mitchum*

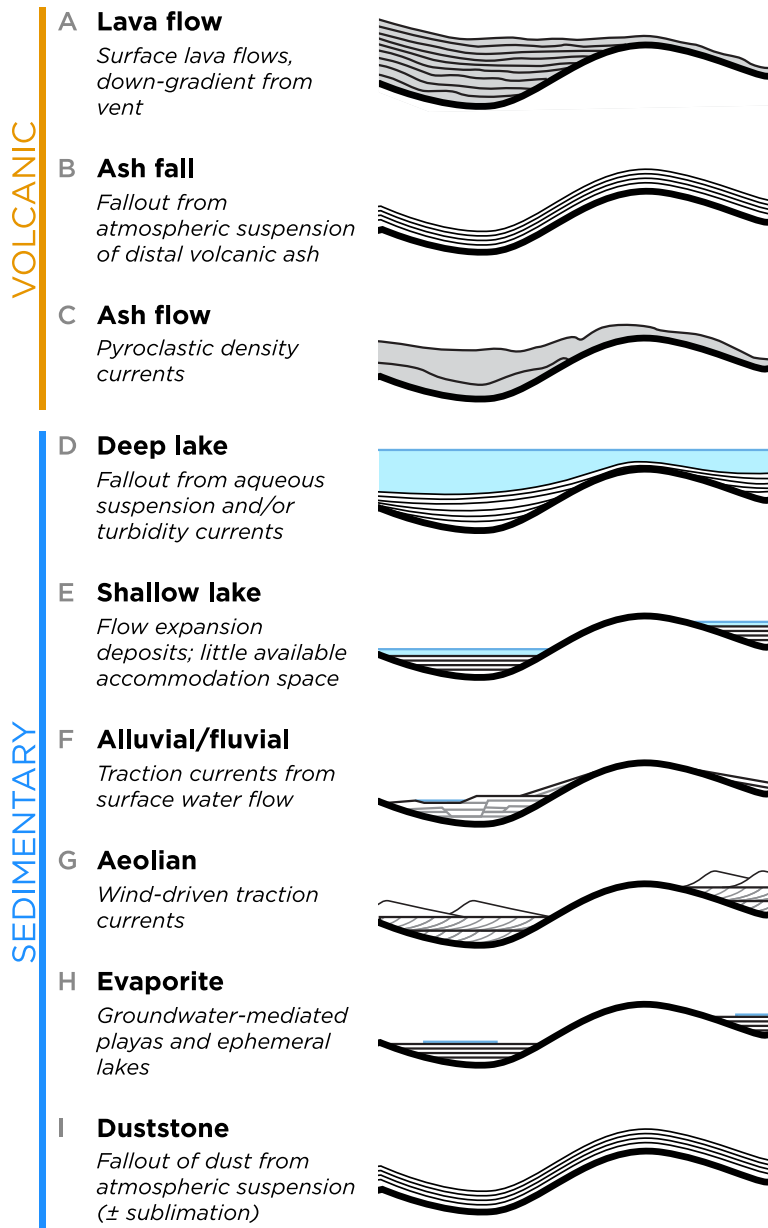


Figure 4: A graphical summary of depositional settings proposed for layered deposits on Mars, with applicability to the layered sulfates. Each potential mechanism varies in the structure and style of bedding predicted, which can be diagnostic of the unit's original form.

182 *et al.*, 1977).

183 In this work, we attempt to determine the formation mechanism and post-depositional his-
184 tory of the layered sulfate unit primarily from its structural characteristics.

185 **3.2 Digital elevation models and dataset registration**

186 Images from the HiRISE instrument on the Mars Reconnaissance Orbiter (*McEwen et al.*, 2007)
187 were acquired covering key parts of the study area. Overlapping pairs of images were acquired
188 with one near-nadir and one oblique image for stereo convergence angles of 15°-30° [Table 1].
189 These stereo pairs were processed using standard pipelines, and digital elevation models (DEMs)
190 were created in the SOCET SET software using the techniques described in *Kirk et al.* (2008). In
191 this pipeline, the stereo images are individually photometrically corrected and horizontally and
192 vertically controlled to the Mars Orbiter Laser Altimeter (MOLA) datum. MOLA shot data are
193 used to correct the elevation of ground control points, and the gridded MOLA DEM is used to
194 anchor the DEM solution. In total, eight HiRISE stereo models were constructed, typically cov-
195 ering 5 × 10-20 km areas of the surface in overlapping HiRISE scenes at a ground sample spacing
196 of 1 m/px. Relative vertical accuracy is ~0.25 m in textured areas of the scenes (based on the “ex-
197 pected precision” metric of *Kirk et al.* (2003) and their estimate of 0.2 px typical image-registration
198 accuracy within SOCET Set). The resulting elevation models were used to create 0.25 m/px or-
199 thorectified images aligned exactly to the DEM.

200 In addition to the HiRISE DEMs, a single CTX DEM covering the entire study area was con-
201 structed using 21 images from 11 overlapping stereo pairs. The dataset was prepared in SOCET
202 SET using methods that closely followed the procedures used for HiRISE. The resulting DEM has
203 a 10 m/px horizontal scale and ~5-20 m vertical fidelity, varying based on the specific stereo pair.

204 The scale gap between HiRISE and MOLA can produce systematic bias when MOLA data is
205 sparsely sampled, especially in the presence of N-S (along-track) sloping topography. Our CTX
206 DEM is not susceptible to such bias due to its much larger coverage area, allowing its use as an
207 external check on the whole-image tilt of the HiRISE DEMs. Elevation models nf1, nf2, and nf4
208 had negligible slope, but nf3 and nf5 at the northern margin of the study area had artificial south-
209 ward slopes of ~0.25 and 0.38°, respectively, corresponding to elevation differences of 30-100 m
210 within *North Basin* relative to CTX. HiRISE elevation models nf6-nf8 were explicitly controlled
211 to the CTX DEM during creation, removing this source of error. These small tilts do not signifi-
212 cantly impact the dip magnitudes measured in this study, but they do limit the precision with
213 which true horizontality can be recovered.

214 The internal quality of stereo DEMs varies based on the stereo convergence angle between
215 scenes. Also, image-matching algorithms perform better on areas with fine-scale surface features.
216 In general, DEM quality is much higher in areas with significant slopes and high local contrast.
217 DEM errors are summarized by the “Figure of Merit” dataset produced by SOCET SET. Errors
218 can additionally be visually inspected using contour lines (following procedures described by *Kirk*

Table 1: HiRISE and CTX scenes used in elevation models

ID	Nadir	Oblique	CA ^a (°)	EP ^b (m)
<i>HiRISE</i>				
NF1	PSP_009217_1975	ESP_027625_1975	18.4	0.15
NF2	ESP_018065_1975	ESP_019133_1975	22.5	0.12
NF3	ESP_026280_1975	ESP_027902_1975	13.0	0.21
NF4	PSP_002809_1965	PSP_006000_1965	24.9	0.11
NF5	ESP_013041_1975	ESP_030025_1975	17.3	0.16
NF6	ESP_021612_1975	ESP_021757_1975	11.8	0.24
NF7	ESP_027269_1970	ESP_042671_1970	21.3	0.13
NF8	ESP_047194_1965	ESP_046983_1965	12.9	0.22
<i>CTX (single multistrip elevation model)</i>				
	G21_026280_1976	D02_027902_1975	13.2	5.1
	B01_010206_1975	B03_010628_1974	15.7	4.3
—	<i>same as above</i>	G02_019133_1977	13.5	5.0
	G09_021612_1972	G09_021757_1972	11.9	5.7
	B18_016720_1978	B18_016786_1978	22.0	3.0
	B19_016931_1975	B19_017076_1975	14.0	4.8
	D14_032504_1996	D14_032649_1996	14.8	4.5
	P15_006778_2002	D17_033849_2002	22.9	2.8
	D14_032715_1995	D15_033137_1996	20.2	3.3
	G11_022680_1976	G12_022746_1976	23.2	2.8
—	P05_002809_1975	P13_006000_1974	24.8	2.6

— Not acquired as a stereo pair

^a Convergence angle

^b Expected vertical precision (assuming resolution of 0.25 m/px for HiRISE and 6 m/px for CTX)

219 *et al.* (2003)). Areas with errors were avoided for our quantitative analyses, but all areas have
220 sufficient data quality for 3D visualization. The CTX DEM suffers from noise for images with low-
221 contrast or poor stereo separation. This manifests as noisy, discontinuous contours and uneven
222 topography in 3D model views. The poorer results of this registration are propagated through
223 our elevation models, yielding large error ellipses on some CTX-derived bedding poles [see e.g.
224 Section 3.4].

225 Each topographic dataset (consisting of a DEM, quality metrics, and aligned orthoimages)
226 is warped to a transverse mercator projection centered on 76.5° to retain angular conformality
227 and approximate true scale over the study area. The datasets were coregistered using significant
228 shared landmarks to build a unified geodetic framework tied to regional CTX imagery. The result
229 is a network of aligned images forming a regionally consistent basemap of the study areas. Other
230 imagery datasets, such as thermal inertia, imaging-spectrometer, and non-stereo HiRISE images,
231 were aligned to this framework.

232 The DEM and imagery basemap was integrated into a 3D computer vision system with a
233 NVIDIA 3D Vision system used for stereo display. HiRISE and CTX stereo pairs were examined
234 in their original viewing geometry using SOCET Set photogrammetry software, and synthetic
235 stereo reconstructions [e.g. Figure 6a] were created using the OSGEarth 3D toolkit to examine
236 the region from arbitrary oblique viewpoints.

237 **3.3 Regional mapping**

238 The morphological character of the layered sulfates and surrounding units was evaluated in detail
239 within the 8 HiRISE stereo pairs used in the project, and their local character was correlated with
240 larger-scale features visible in CTX orthoimages. Regional mapping across all images focused pri-
241 marily on the internal character of the layered sulfates and on the nature of capping units. Map
242 units were identified based on their morphological characteristics, and small-scale features from
243 HiRISE scenes were extrapolated into CTX data. Morphological identification of map units is
244 augmented by Compact Reconnaissance Imaging Spectrometer for Mars (CRISM) hyperspectral
245 data (*Murchie et al.*, 2007). In areas with available CRISM coverage, morphological mapping was
246 cross-correlated with spectral data for verification. Spectra were ratioed to relatively featureless
247 terrains to accentuate the compositional differences in the area of interest. These ratioed spectra
248 were classified broadly into spectral types. The sulfate units have characteristic absorptions for
249 jarosite and polyhydrated sulfates [Figure 5].

250 Contacts between map units, morphological features indicative of channels, and over 850 km
251 of boxwork fractures were mapped. Within the layered sulfates, surface traces of bedding were
252 captured for the structural interpretation of the layered sulfates. Polygonal swatches of surficial
253 units were captured to estimate the orientation of unit surfaces. Mapped contacts were correlated
254 with the topographic dataset to create elevation constraints on the top and bottom of the layered
255 sulfates, which can be interpolated throughout the study area to form a 3D structural model of

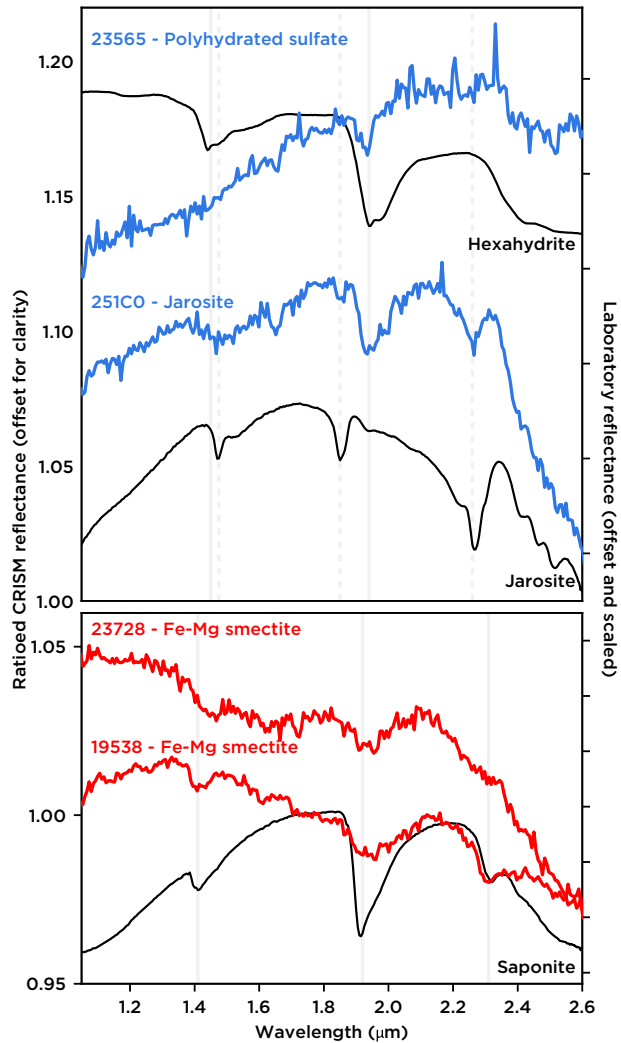


Figure 5: Examples of CRISM spectra used to verify new mapping of the layered sulfates and surrounding units at several locations within the study area. CRISM scenes are identified by 5-digit ID, and mineral identification is shown. Laboratory reflectance spectra shown for comparison are from Brown RELAB, available at http://psf.uwinnipeg.ca/Sample_Database/. The top panel shows sulfate identifications, and characteristic absorptions for polyhydrated sulfates (solid) and jarosite (dotted) are shown as vertical lines. The lower panel shows smectite clay identifications in late sedimentary deposits [Section 4.5.3], with characteristic absorptions for smectite clay shown as vertical lines.

256 the layered sulfates before large-scale erosion of the unit.

257 **3.4 Orientation measurements**

258 To understand stratal relationships, bedding orientations are typically evaluated alongside outcrop-
259 scale observations of angular relationships along a contact. However, where the contact is ob-
260 scured (as is often the case in remote sensing), the local nature of the contact between sedimen-
261 tary packages must be assessed using bedding orientation differences observed in portions of
262 units in isolation. This approach has been employed since the advent of seismic stratigraphy
263 (*Mitchum et al.*, 1977), including on Mars. For instance, angular relationships and approximate
264 unit bounding surfaces have led to the identification of downlap surfaces associated with pro-
265 grading sedimentation in the nearby Jezero Crater deltaic system (*Schon et al.*, 2012; *Goudge et al.*,
266 2017) and other deltas on Mars (*Lewis and Aharonson*, 2006; *DiBiase et al.*, 2013).

267 In order to compare structural data collected from differently oriented hillslopes and DEMs
268 with different levels of inherent accuracy in the NE Syrtis sulfates and capping units, we have de-
269 veloped a new statistical approach and software pipeline for making orientation measurements in
270 the presence of errors. This technique is described in detail in [*Quinn and Ehlmann* (2017); *Quinn*
271 *and Ehlmann* (2018); submitted]. In brief, we use principal-components regression to fit planes
272 to DEM data points, form error distributions for planar estimates, and convert these errors to
273 major- and minor-axis error bounds in spherical coordinates. These errors are hyperbolic error
274 bounds on a nominal plane, or ellipsoidal error bounds around the normal vector to the plane.
275 This new technique allows us to build a regional database of comparable bedding orientation mea-
276 surements of geologic units with statistically rigorous error bars to assess the quality of calculated
277 strikes and dips. Uncertainties are higher for CTX than HiRISE, reflecting high levels of noise in
278 low-relief areas and between poorly-registered image pairs in the CTX DEM.

279 Within individual images, several techniques were used to find the local orientations of geo-
280 logic units at high resolution. In several cases, beds could be traced and evaluated individually.
281 In other cases, individual bedding traces do not form a 3D exposure adequate to capture a unique
282 bedding orientation. To mitigate this, closely-spaced individual measurements were grouped un-
283 der the assumption of parallel bedding within the same stratigraphy. The resulting jointly fitted
284 orientation was tested against the residuals of each measured plane. Groupings with low out-
285 of-plane variation are accepted as likely representative of the true dips of a consistent package
286 of beds. An additional method used for several resistant units primarily exposed at the surface
287 is the measurement of slopes of the current topographic surface. Absent differential erosion,
288 which would impart tilt, surface orientations can be examined alongside outcrop-traced bedding
289 measurements.

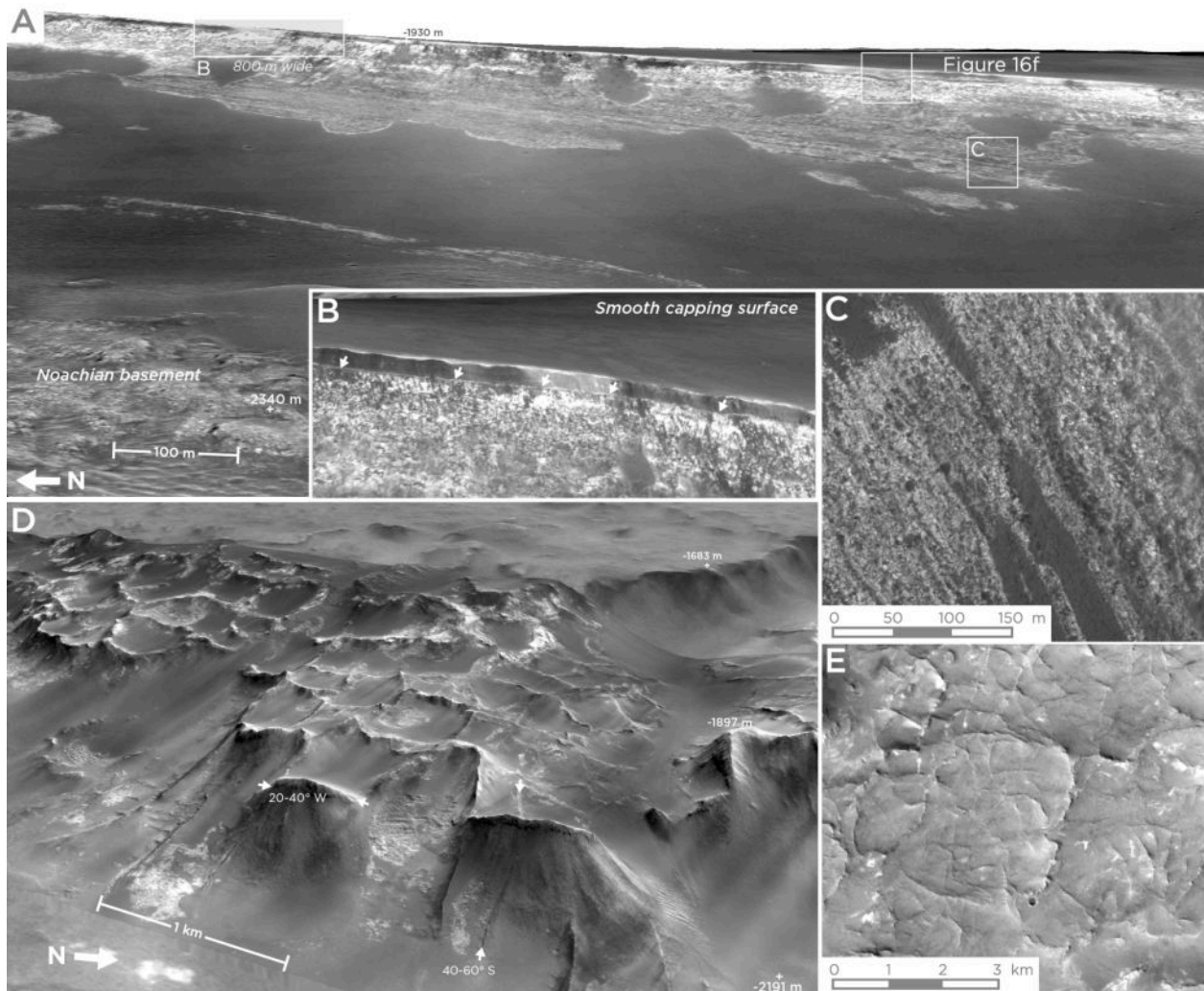


Figure 6: Examples of key morphologies of the layered sulfates. (a) HiRISE stereo (nf6) perspective image of a 2.7 km segment of the west face of Mesa B (no vertical exaggeration), showing a 400 m thick stratigraphy of layered sulfates beneath the “smooth capping surface.” The beds in this stratigraphy dip consistently at $\sim 7^\circ$ to the southwest [Figure 11]. (b) Oblique view ($2\times$ vertical exaggeration) of an 800-meter segment of the bright top contact of the layered sulfates beneath the darker “smooth capping surface.” Arrows show truncations of bedding by the contact. (c) Plan-view image of typical bedding features in the layered sulfates. (d) HiRISE stereo (nf2) perspective image of boxwork polygons, which cover large parts of the study area and have a characteristic scale of ~ 500 m on a side. The approximate dips (within error) of several relatively planar boxwork fracture segments are shown. (e) CTX map view of boxwork ridge network in the *Ridge Basin* area of the unit.

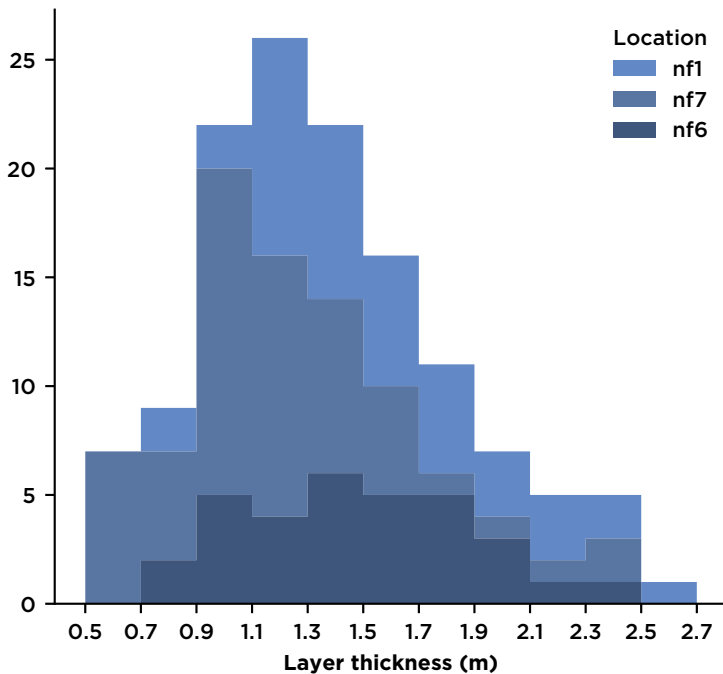


Figure 7: Histogram of layer thicknesses measured for the layered sulfates in three HiRISE scenes with well-exposed, consistently dipping beds, with a mean of 1.38 m and a standard deviation of 0.44 m for 131 layer thicknesses. A single sequence of beds was measured within each image using the method shown in *Lewis et al.* (2008); dip corrections of 4.0° for nf1, 6.9° for nf6, and 3.5° for nf7 were applied. Locations of measured beds in nf6 are shown in Figure 11.

290 4 Results

291 4.1 The layered sulfates

292 **4.1.1 Basic morphological and mineralogical characteristics** The general character of the layered sulfates has been reported in previous investigations (*Ehlmann and Mustard, 2012*) and was
 293 used along with CRISM to map their full extent in the CTX and HiRISE basemaps. The sulfates
 294 are light-toned where exposed, exhibiting a polyhydrated sulfate infrared spectral properties and
 295 continuous, meter-thick bedding. Layers can be followed for several kilometers in a few cases
 296 [Figure 6a] but are often obscured by dark debris. Layers identified in HiRISE are tightly distributed
 297 around a mean thickness of 1.4 m (range: 0.5 m to 2.7 m) [Figure 7]; these layers may
 298 package finer-scale lamination at below HiRISE scale.

299 The layered sulfates have been eroded to their present form by fluvial activity [Section 4.5],
 300 and capping units and post-depositional alteration appear essential for their preservation. The
 301 layered sulfates are recessive in general, and are only preserved where beneath more resistant
 302 caprock or buttressed by polygonal ridges, forming slopes otherwise. Locally, ridged [Figure 6c]
 303 regions of the layered sulfates show strong signatures of jarosite mineralogy (as previously reported
 304 for select locations in *Ehlmann and Mustard (2012)*) and are more resistant to erosion. We
 305 have also found jarosite in a horizon with localized pervasively altered sulfate that appears texturally
 306 distinct [Section 4.3]. Craters are not generally preserved in the sulfate unit, potentially
 307 because the surface of the unit is continuously refreshed by scarp retreat.
 308

309 **4.1.2 Extent** The layered sulfates are exposed over ~100 km east-west along the transition zone
310 between the Syrtis Major lava flows and the Noachian Plains. They extend from ~74-76°E and
311 ~14-16°N. Their northern boundary can be mapped but the southern boundary is not fully char-
312 acterized, as it lies beneath the Syrtis Major lava flows. The southernmost well-characterized
313 exposure is within an erosional window at 74°E, 14°N. The layered sulfates range in elevation
314 from -1600 m to -2800 m at their southeastern eroded margin. The highest exposures of the
315 layered sulfates broadly coincide with the elevation of the regional topographic step that defines
316 the inner ring of Isidis basin.

317 We construct a 3D model of the layered sulfates using simple structural rules constrained by
318 mapping data. First, observed contacts between the layered sulfates and other map units provide
319 direct elevation constraints on the top and bottom boundaries of the layered sulfates [Figure 8a].
320 Contacts with the Syrtis Major lavas and other capping units represent the top surface of the
321 sulfates, and contacts with the NE Syrtis plains units form its base. In select locations, both
322 contacts are exposed in close proximity, allowing a direct measurement of thickness. Unconfor-
323 mities where a capping surface instead directly meets the NE Syrtis plains constrain both the top
324 and bottom surfaces of the sulfates, i.e., indicating places where they are absent. Additional con-
325 straints on the bottom surface of the sulfates are imposed in some covered areas to improve the
326 smoothness of the overall result and mitigate the shallowing bias on the sulfates imparted by
327 contacts on localized Noachian basement highs.

328 Second, the interpolated surfaces are constructed using kriging, a statistical procedure com-
329 monly used for interpolating geological surfaces (e.g. *Oliver and Webster*, 1990; *Caumon et al.*,
330 2009). The kriging model is implemented on a 200 m grid using the Python program PyKrige.
331 Separate kriged surfaces for the top and bottom of the layered sulfate are interpolated beneath
332 present topography and projected into space to model the extent of the layered sulfates in areas
333 they likely occupied in the past but were since eroded. The uncertainty of each surface is calculated
334 based on the RMS distance to the nearest control point.

335 The interpolated structural surfaces provide key limits on the regional form and thickness of
336 the layered sulfates. The sulfates form a thick regional package that has been partially eroded in
337 some areas. Cross-sections extracted along arbitrary axes through the structural model provide a
338 means to assess the performance of the modeling approach and evaluate the deposit-scale char-
339 acter of the layered sulfates [Figure 9]. The layered sulfates have a variable thickness throughout
340 the study area and embay and thin to zero against inherited highs of Noachian highlands topog-
341 raphy. The sulfates have a mean thickness of 197 m, varying from 0 to a maximum of over 600 m
342 over the topographic depression we term Deep Basin [Figure 8b]. Our structural model suggests
343 a prior total volume of layered sulfates of 1245 km³ within the ~5800 km² polygonal area with
344 high-quality mapping constraints, with a volume of 460 km³ (37%) still buried. These volume
345 estimates constitute a minimum constraint on the original volume of the layered sulfates.

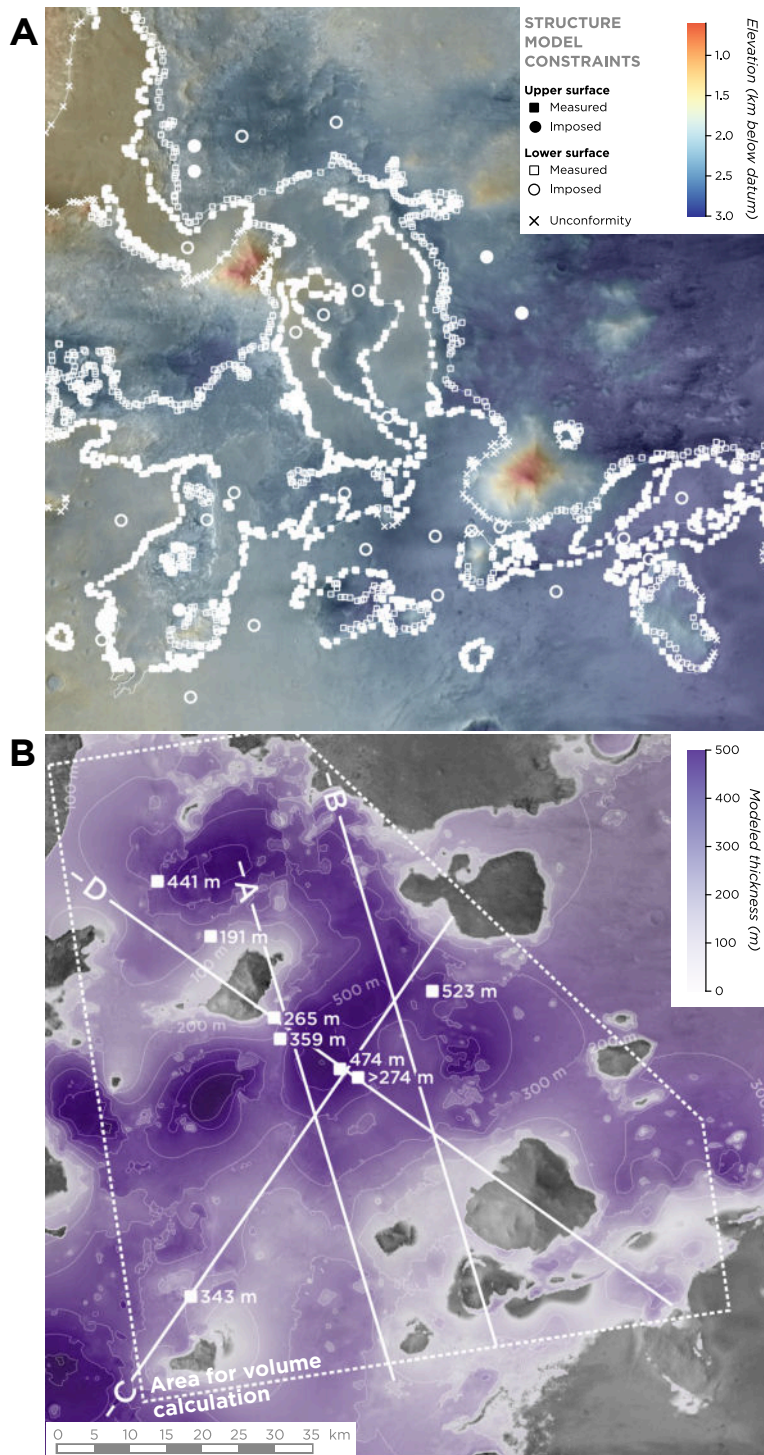


Figure 8: Maps of parameters derived from structural modeling of the top and bottom surfaces of the layered sulfate. (a) Point constraints used for the kriging model underlying the structural model (filled symbols also shown in Figure 9). These points are at the edges of the surface exposure of the layered sulfate and constrain the 3D volume of the unit relative to present topography. Additional constraints (open symbols) are imposed on the top and bottom surface of the layered sulfates to maintain consistent thickness in areas without exposures. (b) Modeled thickness of the layered sulfate prior to erosion of NE Syrtis mesas, showing the tapering of the deposit away from the center of the study area. Locations with known thicknesses (measured from the top to bottom of sulfate exposures) are shown numerically atop the kriged surface. Section lines for Figure 9 are shown.

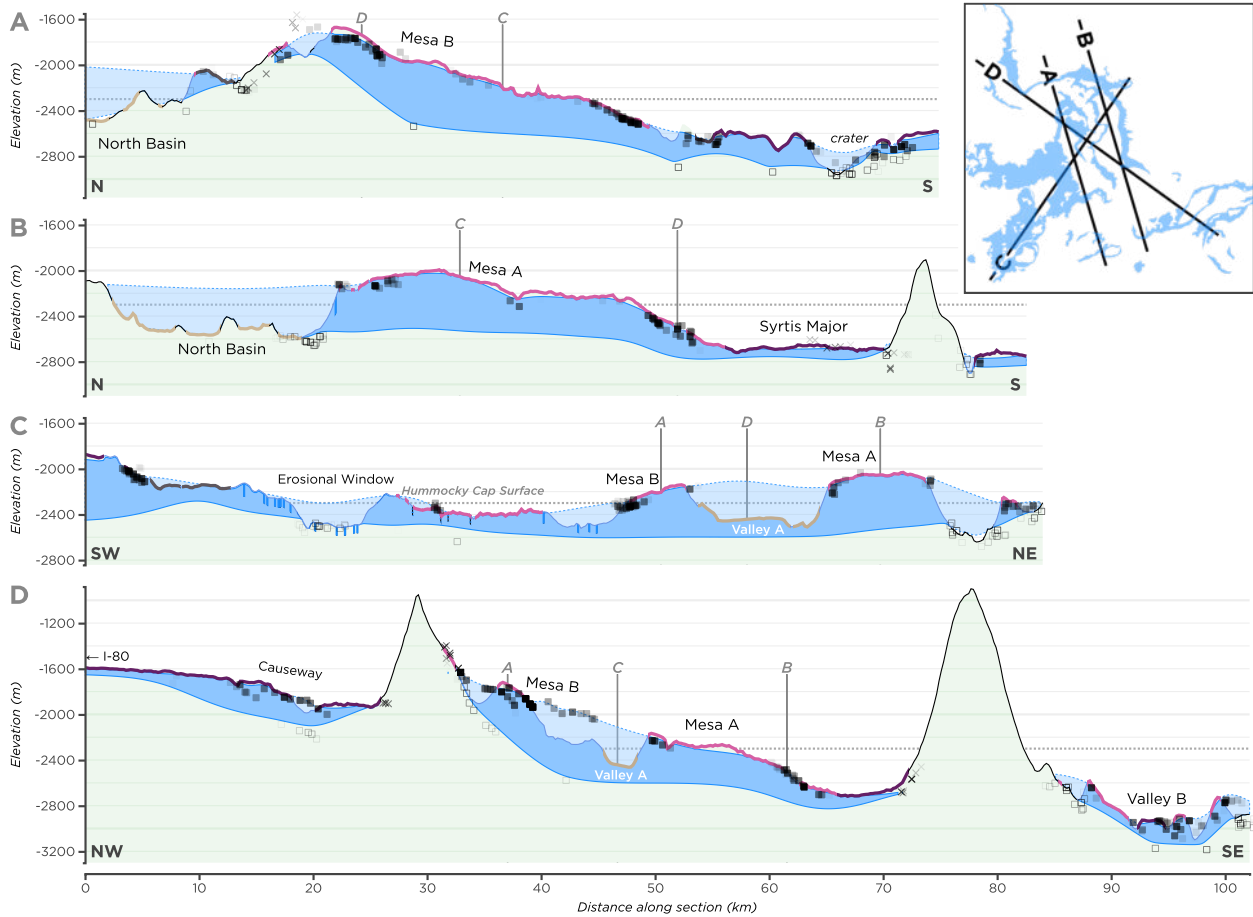


Figure 9: Cross-sections of the study area showing profiles through the layered sulfates created using the structural model [Figure 8]. Model constraints are shown on the section using the symbology of Figure 8, faded with distance from the section line, with constraints up to 2km away shown. All sections are at the same scale, with 10× vertical exaggeration. (a) A north–south transect through *Mesa B* showing the draping of the sulfates atop basement topography. (b) A north–south transect across *Mesa A* and the adjacent low-elevation *North Basin* and *Syrtis Major* lava flow surface. (c) A southwest–northeast transect showing the erosional window at the southwest of the study area, and *Mesas A* and *B*. (d) A northwest–southeast transect including both of the basement buttes and paleovalley relationships shown in Figure 19.

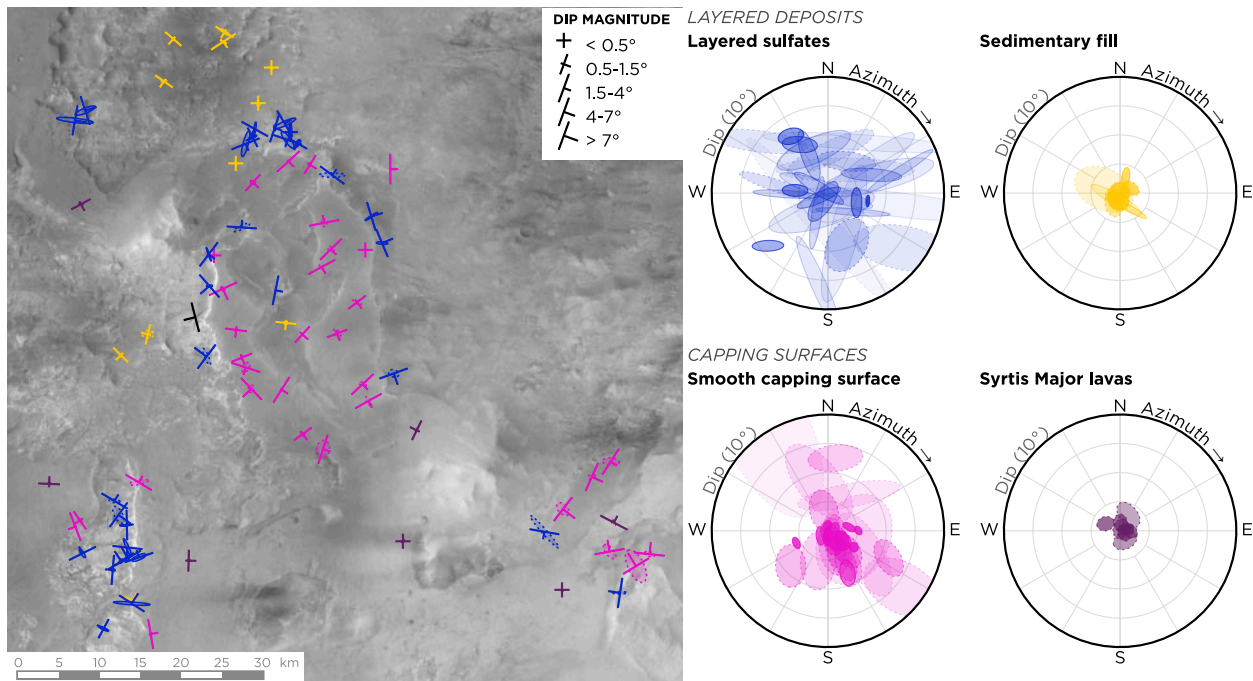


Figure 10: Representative bedding orientations for the study area showing the variation of dips within each class of units. On the map (left panel), nominal bedding orientations are underlain by error ellipses showing the direction of errors. Measurements derived from lower-resolution CTX data have a dotted error ellipse. These ellipses correspond to the data plotted to the right as poles to bedding on upper-hemisphere stereonet. Measured bedding orientations for sulfates and later layered deposits (e.g. late fluvial features) are shown along with surface orientations of the Syrtis Major lava flows and the smooth capping surface.

346 **4.1.3 Bedding orientations** Resolvable bedding traces in the layered sulfates were challenging
 347 to assess as they are only exposed in some areas, where broad hillslopes or the interiors of box-
 348 work domains are stripped of debris cover. In many cases, instead of surface-exposed bedding
 349 planes that are ideal for measurement, the strikes and dips of sulfates were constructed from
 350 the trace of linear features on hillslopes. The quality of local bedding exposures provides varying
 351 levels of confidence in bedding between areas [Section 3.4]. A regional set of representative bed-
 352 ding orientations for sulfates and surface orientations for capping units was determined using
 353 measurements along outcrops that had low errors and agreed with surrounding measurements
 354 [Figure 10].

355 Bedding within the layered sulfates always dips at low angles ($<10^\circ$ and usually $<7^\circ$) in a variety
 356 of directions. There is no systematic trend in dips that might suggest that the orientation of
 357 the sulfate unit was modified by large-scale tilting or regional folding. Instead, the relatively
 358 wide range of dips measured within the layered sulfates suggests that the unit was not deposited
 359 as an originally flat surface. Some areas of the layered sulfates are nearly flat, such as near the
 360 northern boundary of the unit in nf3 and nf5. Despite this, truly flat deposition is well outside
 361 of the confidence bounds for many individual measurements. Both CTX and HiRISE produce
 362 orientations with low-angle nonzero dips with a magnitude greater than measurement errors.

363 The layered sulfates onlap basement topography at both deposit- and bed-scale. The unit thins
364 over and embays basement highs [Figure 9], and bedding often dips away from high topography
365 as well. Bed-scale contact relationships are not directly resolvable from remote sensing but can be
366 inferred from dip relationships: while individual beds may thin somewhat over localized highs,
367 bedding dips shallower than and projecting into adjacent basement topography imply an onlap-
368 ping relationship.

369 Relatively steep dips of 5-10° are persistent over large areas of the layered sulfates. For exam-
370 ple, the ~7 km west-facing hillslope in nf6 exposes a stratigraphy dipping 5-7° to the southwest
371 [Figure 11]. Grouping bedding traces across the entire hillslope yields a maximum orientation
372 error of <1°, consistent with a planar, dipping stratigraphy. Dips slightly steeper than the under-
373 lying basement surface suggest that the sulfate package may downlap at its base. The shallower
374 slope of the capping surface above the sulfates [Section 4.4], which dips to the south instead of
375 southwest, implies a low-angle unconformity between the two. Bedding truncations at the base
376 of the cap are visible in HiRISE imagery [Figure 6b].

377 Although layers within ~5 km packages are often consistently oriented, these exposures differ
378 substantially in orientation across the study area. In nf7, 5 km to the west of nf6, dips are con-
379 sistent ~2-3° to the east. In nf1, still further west, layers dip ~5° northeast. Orientations may
380 change gradually within a single stratigraphic package or at unconformities between relatively
381 planar packages but this cannot be further addressed with current data.

382 Some exposures of the layered sulfates are less internally consistent, with variable bedding
383 orientations at sub-kilometer lateral scale. In the northern part of nf6, dips shallow from ~7°
384 (southwest-directed) to ~4° (south-directed) [Figure 11]. In the northern part of the *Erosional*
385 *Window* (covered only by CTX topography), east-dipping exposures contain bedding traces that
386 are kinked relative to their best-fitting plane, signifying localized variation. Discordant layer ori-
387 entations at hundred-meter to 1-km scale are particularly evident in areas with abundant box-
388 work polygonal fractures, which we discuss further in Section 4.2.3.

389 **4.2 Boxwork polygons**

390 Boxwork polygonal ridges are a striking feature of parts of the layered sulfates. They occur in many
391 areas including the central *Ridge Basin*, along the northern erosional margin of the unit, and in
392 the southern *Erosional Window*. While geographically widespread in occurrence, they are found
393 only patchily throughout the unit. For simplicity, we will discuss these ridge-forming features as
394 fractures while we discuss the data supporting this classification.

395 **4.2.1 Fracture morphology** The boxwork polygons have a characteristic length scale of ~500 m
396 and have strong positive relief with exposures defining ridges standing as much as 30 m above
397 polygon interiors. The features have a significant vertical dimension within the layered sulfates:
398 some single ridges continue through an elevation range of greater than 200 m [Figure 14]. In

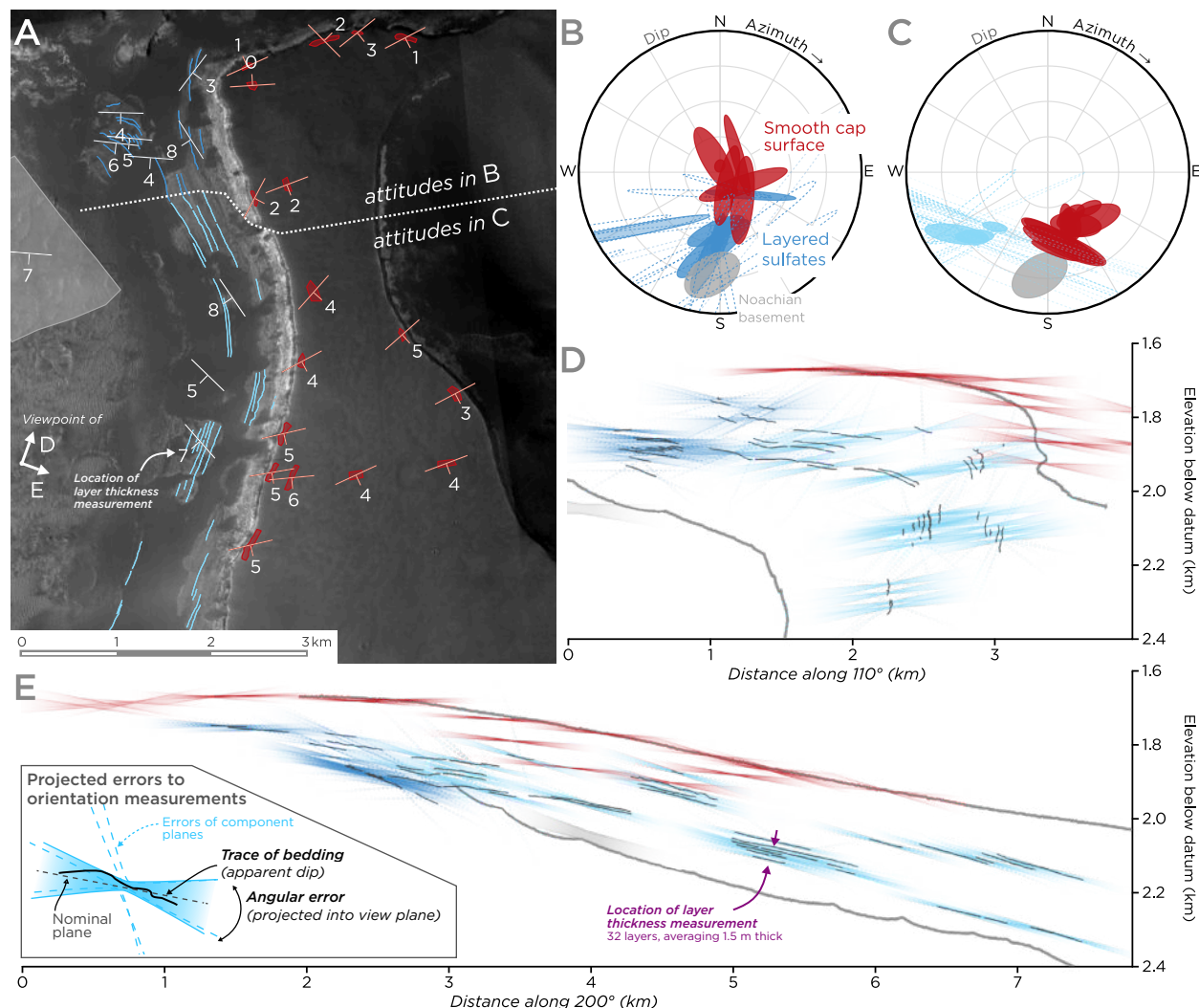


Figure 11: Bedding orientations collected for the top and west face of *Mesa B* in HiRISE orthomosaic nf6, showing the surface slope of the smooth capping surface in red and the layered sulfates in blue. (a) Map view of bedding orientations. (b and c) upper-hemisphere stereonet projections showing poles to bedding with error ellipses for the north and south parts of (a), respectively. Dotted, unfilled ellipses represent components of grouped measurements. (d and e) Projections of the entire stratigraphy along orthogonal axes as shown in (a), with apparent dips of bedding traces and hyperbolic error bounds to planar fits. The difference in orientation of the capping surface and layered sulfates suggests a low-angle unconformity at the top of the layered sulfates. The difference in dips between the northern and southern section corresponds to a “kink” in projected bedding traces in d.

399 nf2 and nf5, the boxwork features penetrate the entire ~400 m thickness of the unit [Figure 6d].
400 The different examples of polygonal fractures vary markedly in thickness. Some ridges are thin
401 and have little topographic expression, while others are thick, prominent, and shed boulders [Fig-
402 ure 12].

403 Morphological features of the polygonal ridges suggest that they may be filled fractures in-
404 stead of single-origin injective features. Many boxwork ridges show light-toned bands parallel to
405 and on either side of the fracture centerline [Figure 12b-d]. This parallel-sided geometry varies
406 from ~5 m [Figure 12b and c] to up to 20 m in width [Figure 12d]. Several examples show addi-
407 tional features such as a light-toned central zone ranging from ~1 to ~30 m across [Figure 12d]
408 or a diffuse bright-toned halo ~2-20 m on each side of the fracture itself [Figure 12c and d]. No
409 cross-cutting relationships are observed at fracture intersections, suggesting that the fractures
410 were filled separately from their propagation.

411 Parallel-sided fills are indicative of channelization of fluid along pre-existing fractures. Rela-
412 tively narrow, zoned fractures surrounded by wide zones of light-toned material texturally con-
413 tinuous with the groundmass of the unit [e.g. Figure 12d] suggest the formation of a zone of
414 alteration around a relatively narrow original fracture. The considerable width range of altered
415 zones along fractures suggests that the intensity of fluid channelization along boxwork fractures
416 varied significantly within the sulfate unit.

417 **4.2.2 Fracture orientations** Over 850 km of boxwork ridges exposed within the layered sulfates
418 were digitized as lines atop regional CTX imagery [Figure 3]. Ridge orientations were measured
419 across the dataset, and intersection angles were calculated for any point where three or more
420 digitized fracture segments met. Angles were calculated at 10 m-radius buffers around each in-
421 tersection, which corrects for digitization noise and changes in orientation within this radius.

422 Boxwork ridges commonly intersect at right angles, both crossing and forming “T-junctions”
423 [Figure 13]. They do not have 120° orientations characteristic of progressively annealed drying
424 polygons and permafrost patterned ground (*El-Maarry et al.*, 2010) or cooling polygons where
425 rock tensile strength exceeds fracture propagation stresses (*Shorlin et al.*, 2000). Individual ridges
426 can run for several km, over which they range from relatively straight to curvilinear. In some areas
427 (particularly the *Ridge Basin*) ridges follow gentle arcs with radii of ~8-10 km. However, these
428 circular trends are commonly disrupted by cross-cutting fractures. The dominant map pattern of
429 the boxwork ridges is of a coarsely gridlike, sometimes weakly concentric, network.

430 The summary of fracture orientations suggests a complete lack of preferred orientation for
431 the fractures [Figure 13b]. This suggests that the features were not formed by injection or in a
432 regionally consistent stress field. The curvilinear nature of individual fractures makes it difficult
433 to assess their dip in many cases, and most appear to be near vertical. However, some apparently
434 planar fractures have steep (40-60°) non-vertical dips [Figure 6d].

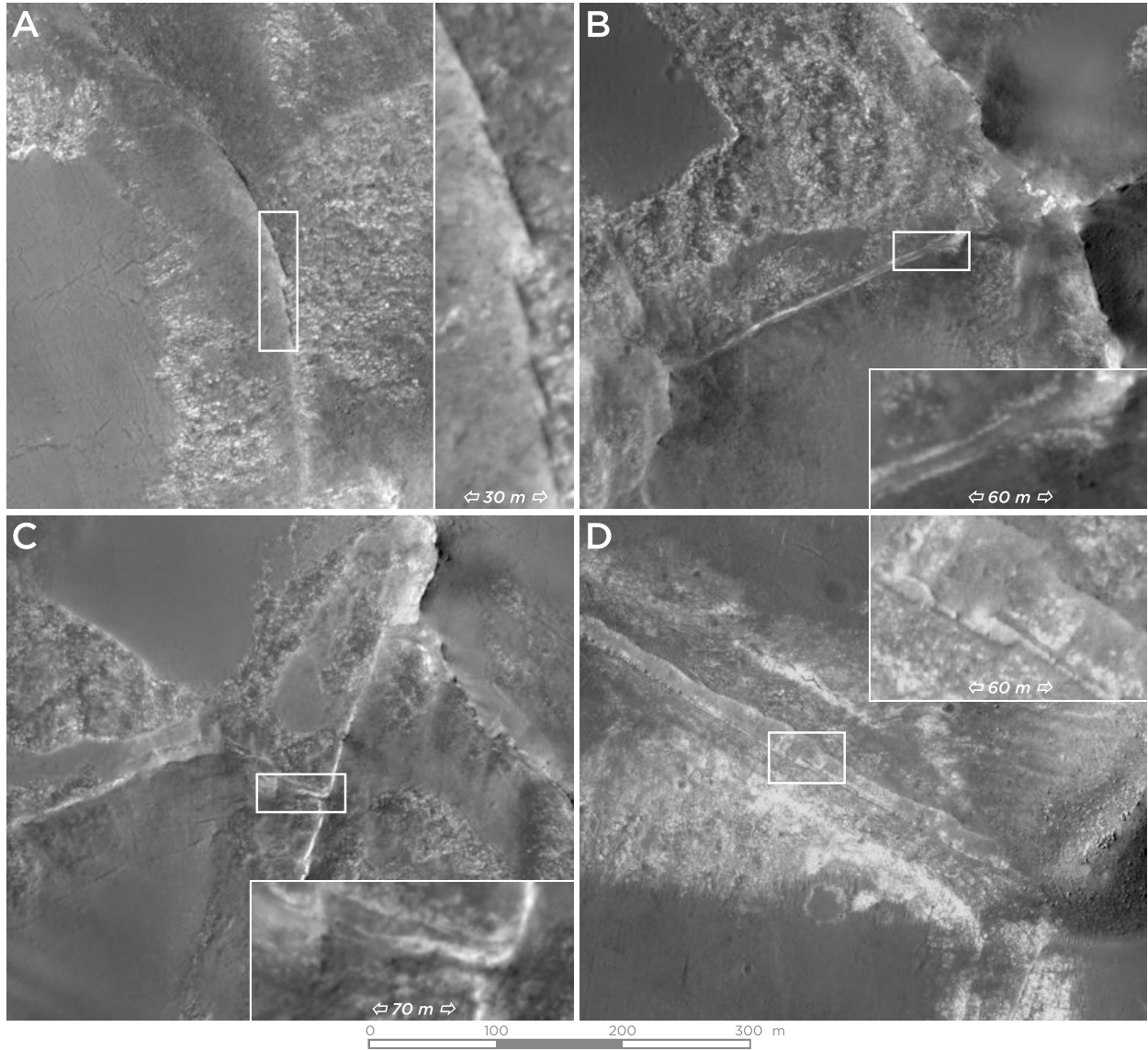


Figure 12: Examples of boxwork fractures from HiRISE images within the study area. **(a)** A minimally altered fracture <1 m thick (nf3). **(b)** A slightly thicker fracture showing a characteristic double-walled morphology of channelized fluid along the fracture (nf2). **(c)** A fracture junction showing a dark-toned halo around the fracture (nf2). **(d)** A wide fracture zone showing a light-toned fracture surrounded by a halo, with the entire alteration zone > 100 m wide (HiRISE ESP_026069_1970).

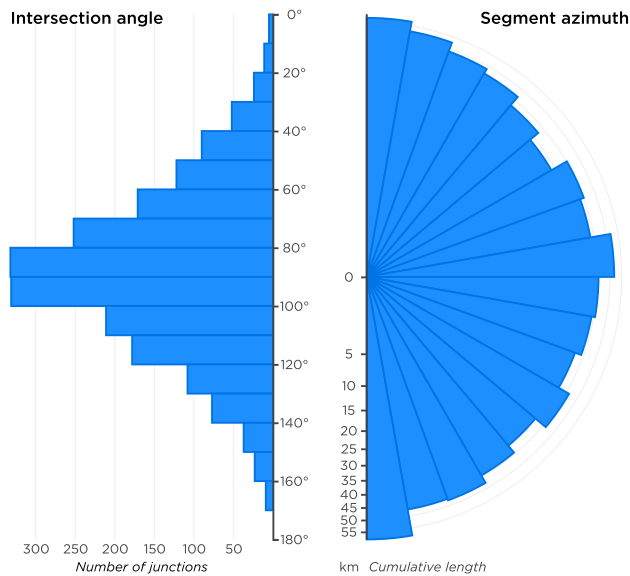


Figure 13: Summary parameters for boxwork fractures measured across the study area [Figure 3]. (a) Frequency distribution of boxwork fracture intersection angles. The largest angle is redundant and is excluded (all angles must sum to 360°). Thus, “T-junctions” between fractures are plotted as two ~90° angles with an implicit 180° angle between parallel segments. (b) Frequency diagram of boxwork fracture azimuth, demonstrating a lack of preferred orientation for the fractures.

435 **4.2.3 Dip changes at small spatial scale** Changes in local bedding orientations of 2-3° are resolv-
 436 able in many areas of the layered sulfates with boxwork fractures. nf3 and nf5 contain a 200 m
 437 thick exposure of sulfates cut by boxwork fractures which range from near-flat to ~2° south- or
 438 northeast-dipping. nf7 contains planar-dipping boxwork domains with southeast dips steepen-
 439 ing from ~1° to 3° over a few hundred meters in successive fracture polygons. The best-exposed
 440 example of localized bedding variation associated with boxwork contains more dramatic changes
 441 of bedding orientation (up to 10° over a kilometer) on a west-facing slope in nf4, within the
 442 southern part of the *Erosional Window* [Figure 14]. The slope contains sulfate material layered at
 443 meter-scale and buttressed by resistant polygonal ridges. In the northern part of this exposure,
 444 dips are variable but generally southward. Scatter to the northwest represents high uncertainties
 445 aligned with the local hillslope. The southern part of this exposure contains dips that are generally
 446 northwestward. The opposing dips occur within a 2-km section over several boxwork domains. A
 447 projected cross-section [Figure 14d] shows dip changes both gradually within boxwork domains
 448 and abruptly between them. Many of the individual measured bedding traces are curved relative
 449 to their best-fitting plane, suggesting that individual layers are not planar and leading to large
 450 error magnitudes for orientation data.

451 The association of bedding-orientation changes at small spatial scale with the boxwork ridges
 452 suggests that the formation of the ridges may have modified dips of the nearby beds. The pro-
 453 posed mechanism for creating these boxwork fractures can lead to localized bed orientation changes
 454 at sub-kilometer scale [Section 5.2.1].

455 **4.2.4 Fracture mineralogy** Some of the boxwork polygons, particularly in nf2, are within a
 456 CRISM scene and wide enough to be covered by a single pixel. *Ehlmann and Mustard* (2012) pre-
 457 sented measurements showing spectral signatures of jarosite-family minerals ($(\text{K, Na, H}_3\text{O})\text{Fe}_3^{3+}(\text{OH})_6(\text{SO}_4)_2$)

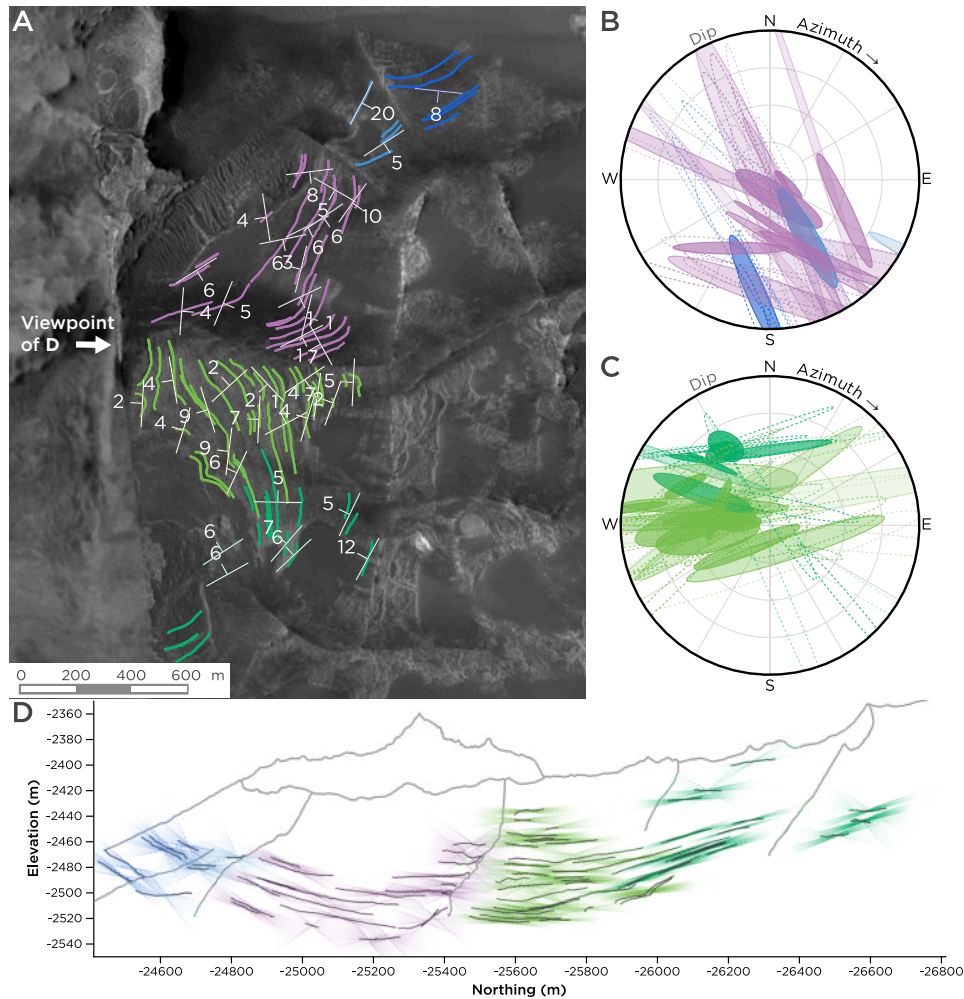


Figure 14: Attitude measurements of layered sulfates in boxwork fracture domains within HiRISE scene nf 4 in the southern *erosional window*. The measurements show a change in bedding dips from dominantly south to north-dipping over ~2 km north to south. **(a)** Map view of bedding orientation measurements. **(b)** and **(c)** Upper-hemisphere stereonets showing poles to bedding for the north and south parts of **(a)**. Measurements are color-coded by boxwork domain and shaded by confidence. Dotted, unfilled ellipses represent components of grouped measurements. **(d)** a N–S cross section within the measurement domain, showing the apparent dip of bedding traces and hyperbolic error bounds to planar fits. The poles to bedding and projected cross-section show a change in apparent dip from north to south along the cross section, and sudden small-scale dip changes across boxwork fractures.

458 in the boxwork polygonal ridges. Applying the same methods to other ridged areas within the
459 layered sulfates, we have found several other examples of polygonal ridges enriched in jarosite
460 [Figure 5]. Given the similar morphology and erosional resistance of ridges across the study area,
461 jarosite is likely a key feature of the boxwork fractures across the entire study area.

462 **4.3 Penetrative alteration**

463 Some areas of the layered sulfates are characterized by pervasive mineralization not channelized
464 along boxwork fractures. In all cases with CRISM coverage, the “altered sulfate” shows spectral
465 signatures of jarosite. This light-toned, erosionally resistant “altered sulfate” is notably present
466 in the north part of nf1, the *Erosional Window* (nf4 and nf8), and the central part of nf6 [Fig-
467 ure 3]. In the *Erosional Window*, several sub-kilometer scale flat-topped outcrops of the olivine-
468 carbonate unit (*Ehlmann and Mustard, 2012*) are surrounded by erosionally resistant, massive,
469 and light-toned altered sulfates [Figure 15]. The valley containing these materials is ringed by
470 eroded layered sulfates with raised boxwork ridges. The margins of the altered sulfates show
471 finely patterned fractures at the boundary and wider, linear fractures extending ~200 m into the
472 sulfates [Figure 15b]. This pattern of fracturing is much denser than the boxwork domains, sug-
473 gesting a localized and intense fluid interaction at the basal contact of the sulfates in this location.

474 Light-toned, mineralized zones are also associated with the upper surface of layered sulfates,
475 just below the smooth capping surface [e.g. Figure 6c]. Unlike the more intense fluid alteration
476 overprint described above, these zones show no contrast in erosional resistance relative to the
477 groundmass of the sulfates. However, they do have jarosite spectral signatures [251C0 – Jarosite
478 in Figure 5]. In the *Erosional Window*, this light-toned material includes a collection of coarsely
479 patterned, rounded alteration domains, with a ~10 m characteristic scale [Figure 15c]. The pat-
480 terned material grades laterally eastward into boxwork ridges which cut visibly layered material,
481 and the light-toned material in these domains appears to be channelized along boxwork ridges.
482 The mineralogical parallels between massive, altered domains and the boxwork ridges suggests
483 that these features were formed in the same or a similar episode of fluid interaction.

484 **4.4 Capping units**

485 The layered sulfates are exposed at the boundary of the Syrtis Major volcanic province, and studies
486 to date have identified its capping surface as the Syrtis Major lava throughout the region (*Ehlmann
487 and Mustard, 2012; Bramble et al., 2017*). We find that the capping surface in the central part of the
488 study area is distinct from the Syrtis Major lavas, and we interpret this “smooth capping surface”
489 to have formed by a fundamentally different process.

490 **4.4.1 Morphology** Two distinct types of unit locally overlie the layered sulfates within the study
491 area [Figure 16]. The “smooth capping surface” is uniformly dark and relatively featureless. Pre-
492 served craters are small and fresh, with a relative lack of intermediate-sized craters; larger craters
493 that are preserved appear as poorly defined “ghost” features [Figure 16a]. The “hummocky cap-

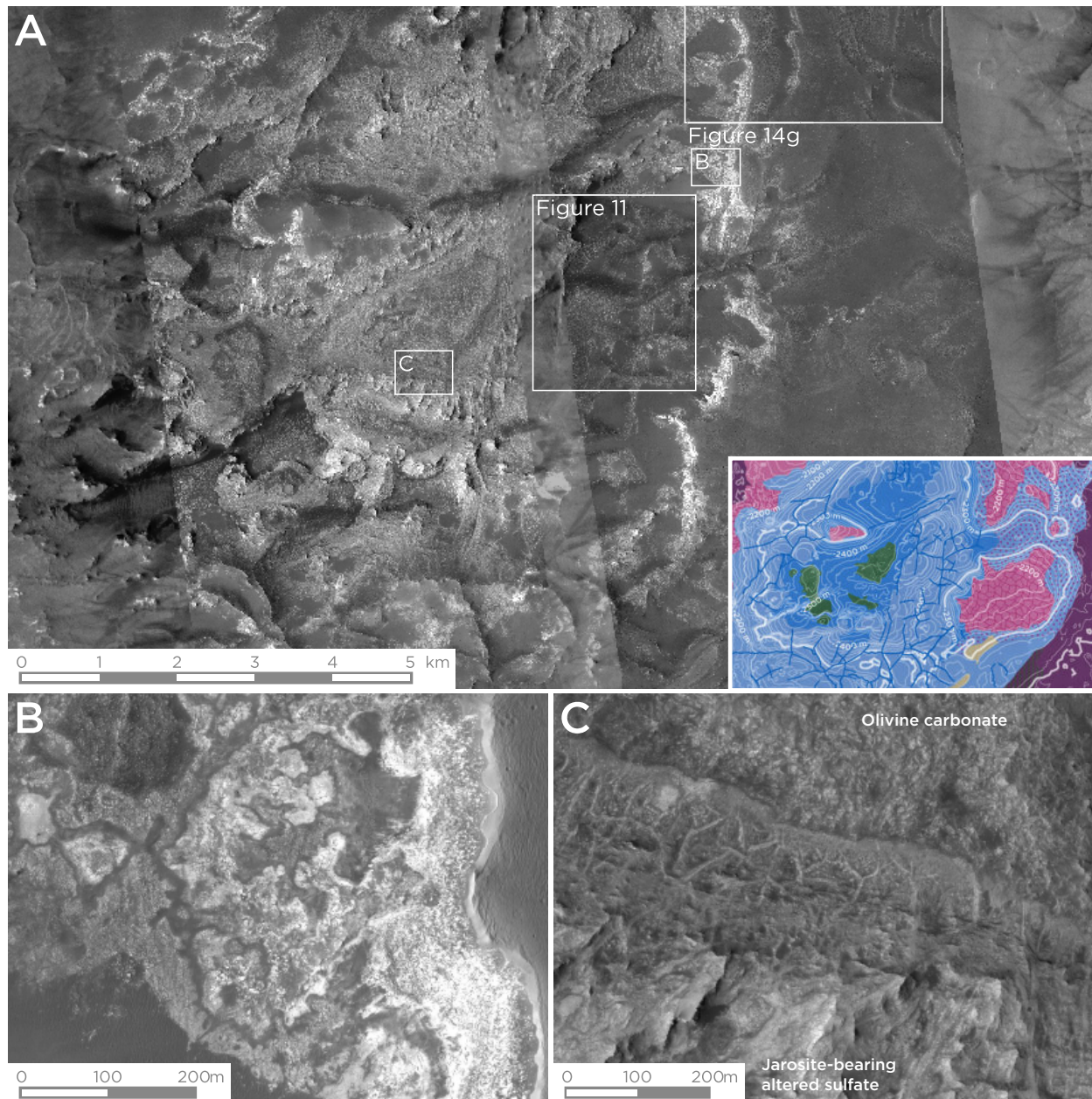


Figure 15: (a) The *Erosional Window* at the southwest margin of the study area, showing parts of HiRISE scenes nf4 and nf8, where the layered sulfates are erosionally resistant and carry strong jarosite signatures. The inset geologic map (from Figure 3) covers the same area. (b) Mottled light-toned altered domains associated with the contact between the “hummocky capping surface” and unaltered exposures of layered sulfate. Light-toned material is partially channelized along boxwork fractures, suggesting that the two features are linked. (c) Jarosite-bearing, erosionally resistant altered sulfates adjacent to and stratigraphically above an exposure of olivine-carbonate, with dense fractures indicative of fluid flux at the interface.

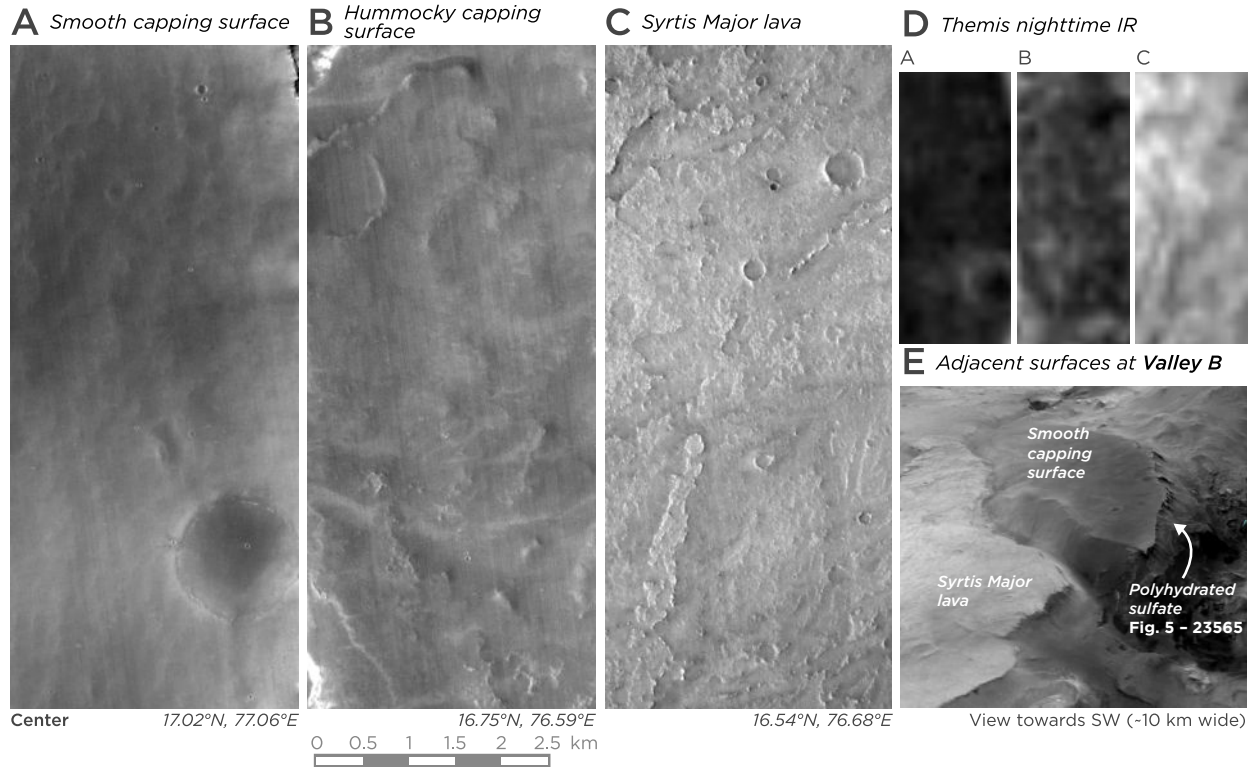


Figure 16: Same-scale views atop a CTX orthomosaic of morphological characteristics of (a) the smooth capping surface, (b) the hummocky capping surface, and (c) the Syrtis Major lavas. (d) THEMIS nighttime temperature corresponding to panels (a-c). (e) an example of the smooth capping surface and the Syrtis Major lavas in contact.

494 ping surface,” a subtype of the smooth capping surface, has increased apparent roughness at 100-
 495 m lateral scale [Figure 16b], created by localized topographic variations of ~10-50 m in elevation
 496 [Figure 17g]. The smooth and hummocky capping surfaces are predominantly found in the central
 497 and southeast parts of the study area.

498 The Syrtis Major lavas mapped within the study area are continuous with a regionally eastward-
 499 sloping surface of the Syrtis Major volcanic province (*Hiesinger and Head, 2004; Ivanov and Head,*
 500 *2003*). This surface has a notably different character than the other capping units: it preserves
 501 small-scale features (e.g. low-relief benches and scarps), retains craters well [Figure 16c], and pre-
 502 serves evidence of fluvial incision on its surface and edges [Figure 19b; see also Section 4.5].

503 **4.4.2 Structural characteristics** The range of surface orientations of the smooth capping sur-
 504 face is more restricted than the range of bedding dips within the layered sulfates [Figure 10]. The
 505 smooth capping surface dominantly slopes southeast, with surface slopes of up to 5°. The differ-
 506 ence in orientation distributions of the smooth cap and the underlying layered sulfates suggests
 507 a low-angle unconformity, with the smooth capping surface emplaced atop the layered sulfates
 508 after a period of erosion.

509 The surfaces of the smooth capping surface are likely to have formed in-situ atop dipping

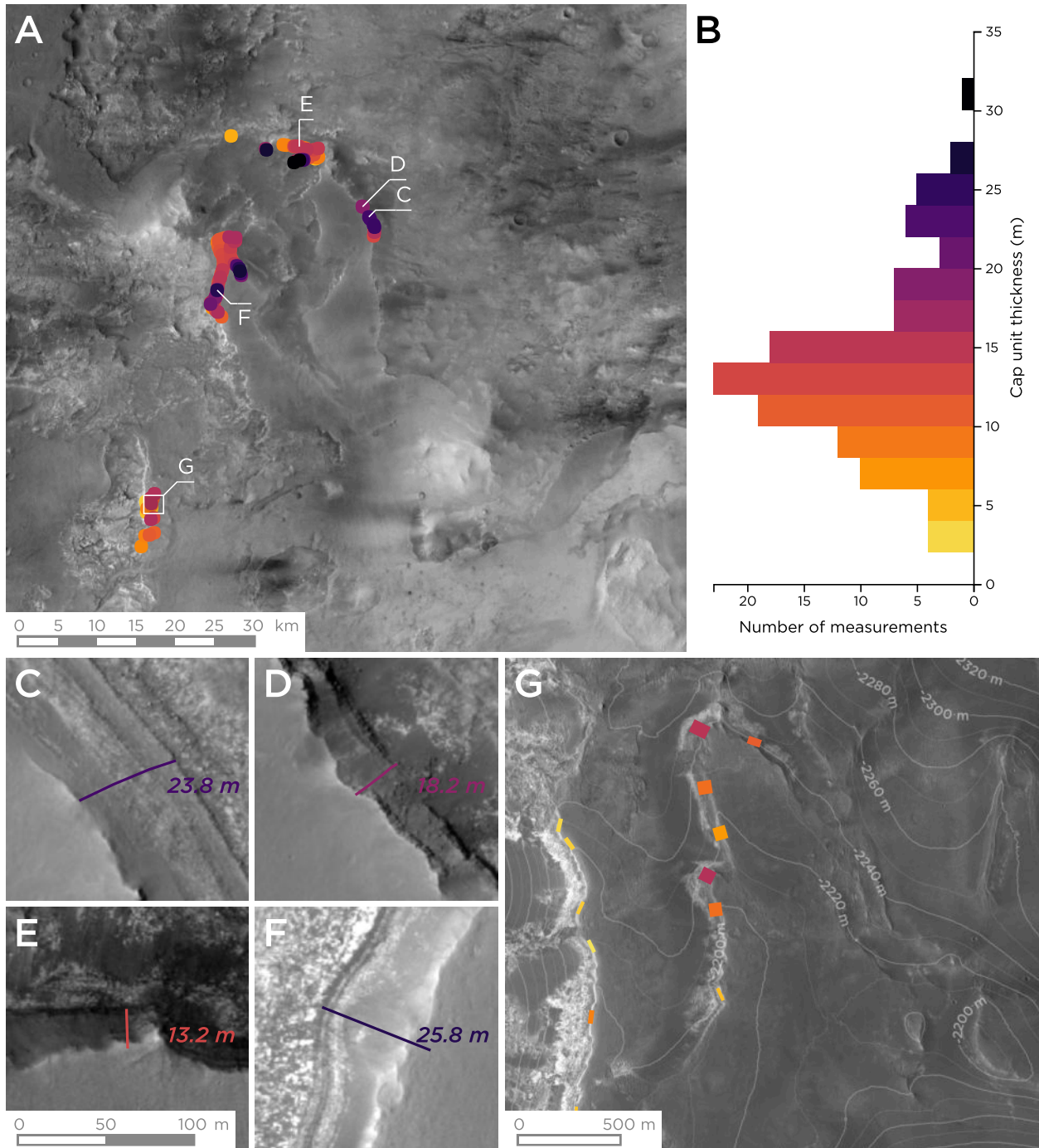


Figure 17: Thickness and morphology of the smooth capping surface in HiRISE scenes within the study area. (a) A summary of sampling, with measured locations color-coded by thickness. (b) Histogram of thicknesses sampled throughout the study area. The coloration of histogram bands corresponds to the points on a. Thickness ranges from 5 to 25 m across the sampled HiRISE scenes. (c-f) Close-ups of capping unit margin showing thickness and morphology at a single location: (c) coarse internal layers within the capping surface; (d) poorly resolved layers and shedding of boulders downslope; (e) poorly resolved internal structure; (f) downwarping of the basal contact by cratering. (g) The “hummocky cap surface” at the eastern edge of the *Erosional Window*. This area hosts the thinnest recorded examples of the capping unit and two parallel curved scarps (left side of image) doubling the edge of the capping surface.

510 exposures of the layered sulfates. In contrast, the Syrtis Major lavas are low-dipping and generally
511 sloped to the east with dips of less than 2° [Figure 10]. Broadly, the lavas form surfaces with
512 regionally consistent dips at 10-20 km scale, oriented with the topographic gradient into Isidis
513 Basin.

514 **4.4.3 Spectral and thermophysical characteristics** Both the Syrtis Major lavas and smooth
515 capping surface have an indistinctly mafic mineralogy with olivine and pyroxene absorptions in
516 CRISM data. The smooth capping unit has generally lower thermal inertia than the Syrtis Ma-
517 jor lava flow [Figure 18]. Low thermal inertia is typically the result of lesser induration, higher
518 porosity, or mantling fine-grained debris.

519 **4.4.4 Thickness of the smooth capping surface** The low thermal-inertia character of the smooth
520 capping surface is coupled with a resistant erosional style. The measured thicknesses of the cap
521 unit [Figure 17b] are largely between 10 and 20 m with several outliers in nf3 and nf4. Close-up
522 views of the internal structure of this interval show coarse internal layering and boulder-shedding
523 scarps [Figure 17c-f]. Figure 17f shows warping of the lower contact by an impact and continuous
524 dark bands within the light-toned material at the base of the cap surface.

525 The “hummocky capping surface” subtype of the smooth capping surface is topographically
526 rough and may be affected by dislocations in the underlying layered sulfates. Figure 17g shows 10-
527 20 m elevation steps separating differently-dipping “plates” of capping material at sub-kilometer
528 scale. In the center, the capping surface is resolved into two distinct surfaces, the lower of which
529 is thinner (~5 m) and slopes eastward and slightly away from the scarp defining the edge of the
530 window. The two scarps are at nearly the same elevation, and may be the result of small-offset
531 (~10 m) normal faulting within the underlying layered sulfates, propagated upwards to cause dis-
532 locations in the hummocky capping surface.

533 **4.4.5 Relationship with the Syrtis Major lavas** The local relationships between the smooth
534 capping surface and the Syrtis Major lavas are key to separating the two units. In the central
535 portion of the study area, the distal Syrtis Major lavas flow eastwards from the outlet of the *I-*
536 *80* valley, capping a mesa of the layered sulfates (the *Causeway*) and embaying a major basement
537 peak [Figure 19a]. The lava flow terminates at an indistinct point in the upper part of *Valley A*,
538 with morphologically similar surfaces forming “steps” at progressively lower elevation. Below this
539 transition zone, *Valley A* is mantled by the “draping valley fill” [Section 4.5.4] in its lower reaches.
540 On the valley’s southwest flank, the layered sulfates form the bulk of *Mesa B*, which is capped by
541 the smooth capping surface at ~-1600 m. The smooth capping unit is ~200 m higher than adjacent
542 Syrtis Major flows. This elevation relationship suggests that the smooth capping surface was
543 formed atop the sulfates prior to and at significantly higher elevations than the Syrtis Major lavas.
544 The Syrtis Major lavas appear to have flowed down a valley that was eroded through the layered
545 sulfates and cap and into the upper reaches of *Valley A*. The topography is now inverted to form

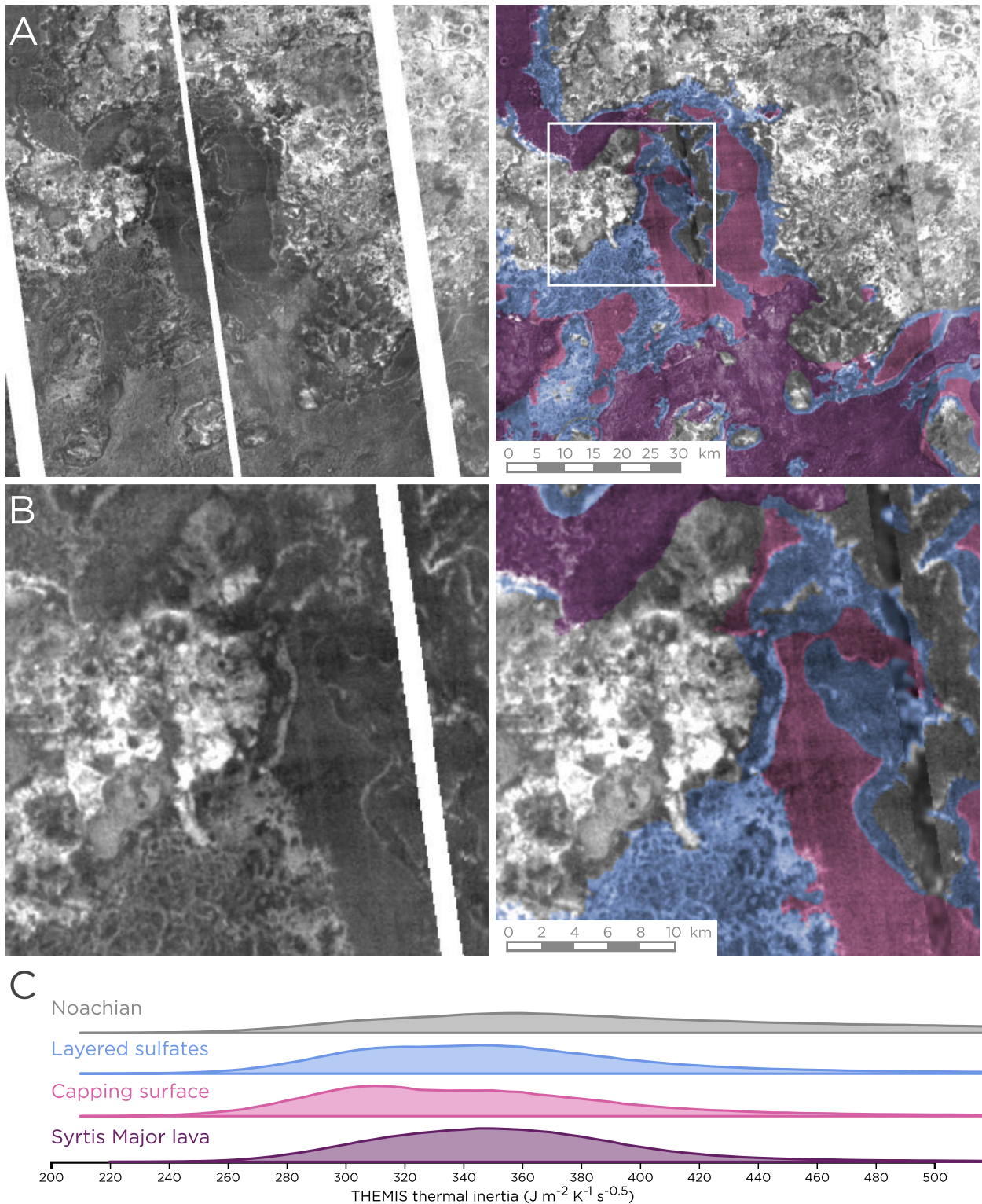


Figure 18: (a) THEMIS quantitative thermal inertia images, selected for capture during the same season (autumn) for consistency based on the guidelines established in *Ferguson et al.* (2006); in the right panel, gaps are filled with mosaicked THEMIS nighttime IR, and unit boundaries are overlaid. This dataset shows the low thermal retentivity of the smooth capping surface. (b) Zoom highlighting the central part of the study area. (c) A histogram of thermal inertia subset by unit over the study area, showing that the capping surface has the lowest average thermal inertia of all mapped units.

546 the *Causeway*, capped by lavas. The presence of an erosional unconformity at the sulfate–smooth
547 capping surface contact implies significant erosion before the formation of the capping surface.
548 The mesas flanking *Valley A* were formed by another episode of fluvial erosion that postdates the
549 formation of the smooth capping unit.

550 *Valley B* at the southeastern margin of the study area contains a similar relationship [Fig-
551 ure 19b]. The northwest flank of the valley slopes inward at $\sim 5^\circ$, and is composed of CRISM-
552 verified layered sulfates capped by the smooth capping surface; the southwest side does not have
553 CRISM coverage but is morphologically similar. The floor of the valley contains a lobe of Syrtis Ma-
554 jor lava, which has a rougher surface with an indurated, crater-retentive character. On both sides
555 of the valley, the contact between the lavas and smooth capping surface is erosionally modified
556 with channels cut into the boundary (discussed in Section 4.5.1). This relationship demonstrates
557 that the Syrtis Major lavas flowed through significant pre-existing relief of paleovalleys formed
558 in capped layered sulfates.

559 **4.5 Late fluvial history**

560 **4.5.1 Channels atop the lava flows** Fluvial activity continued after sulfate and smooth cap unit
561 erosion and Syrtis Major lava emplacement, substantially eroding parts of the Syrtis Major lava
562 flows. The Syrtis Major lavas are modified by inscribed channels across the surface and at the
563 edges of the lava throughout the study area, in agreement with the outflow channels reported to
564 the south of the study area by *Mangold et al.* (2008a). The down gradient and incised nature of
565 the inscribed channels distinguishes them from lava channel features.

566 The inscribed fluvial features are paired with deeper (~ 100 m) canyons cut into the surface
567 of the lava flow. In Figure 9b, these features separate parts of the lava flow at discrete elevation
568 steps, suggesting that different flow bodies formed at different times, during progressive erosion
569 of the layered sulfates.

570 These channels and canyons cut into the lava flow surface postdate the sulfate-hosted paleo-
571 valleys described in Section 4.4.5. After *Valley B* in the southern part of the study area was filled
572 with a lobe of the Syrtis Major lava flows, smaller canyons were inscribed at the contact between
573 the lavas and the capping surface forming the slopes of the valley [Figure 9a and Figure 19b].
574 *Valley A* did not experience similar reoccupation, which we discuss in Section 4.5.4.

575 **4.5.2 Fluvial and lacustrine deposits** The north margin of *Deep Basin* contains an integrated
576 fluvial system, with an amphitheatre-shaped canyon incised into the edge of the Syrtis Major lavas
577 connected by a preserved channel to a small delta (elevation -2320 m) [Figure 20a and b]. This
578 channel formed atop Noachian basement, layered sulfates, and the Syrtis Major lavas. Distal
579 to this delta, a flat-lying surface of presumably lacustrine origin is preserved at -2340 m. These
580 features are perched ~ 500 m above the floor of *Deep Basin*, which has its deepest point at -2800 m
581 only 3.5 km to the southwest. The canyon cut into the Syrtis Major lavas is tied to lightly incised

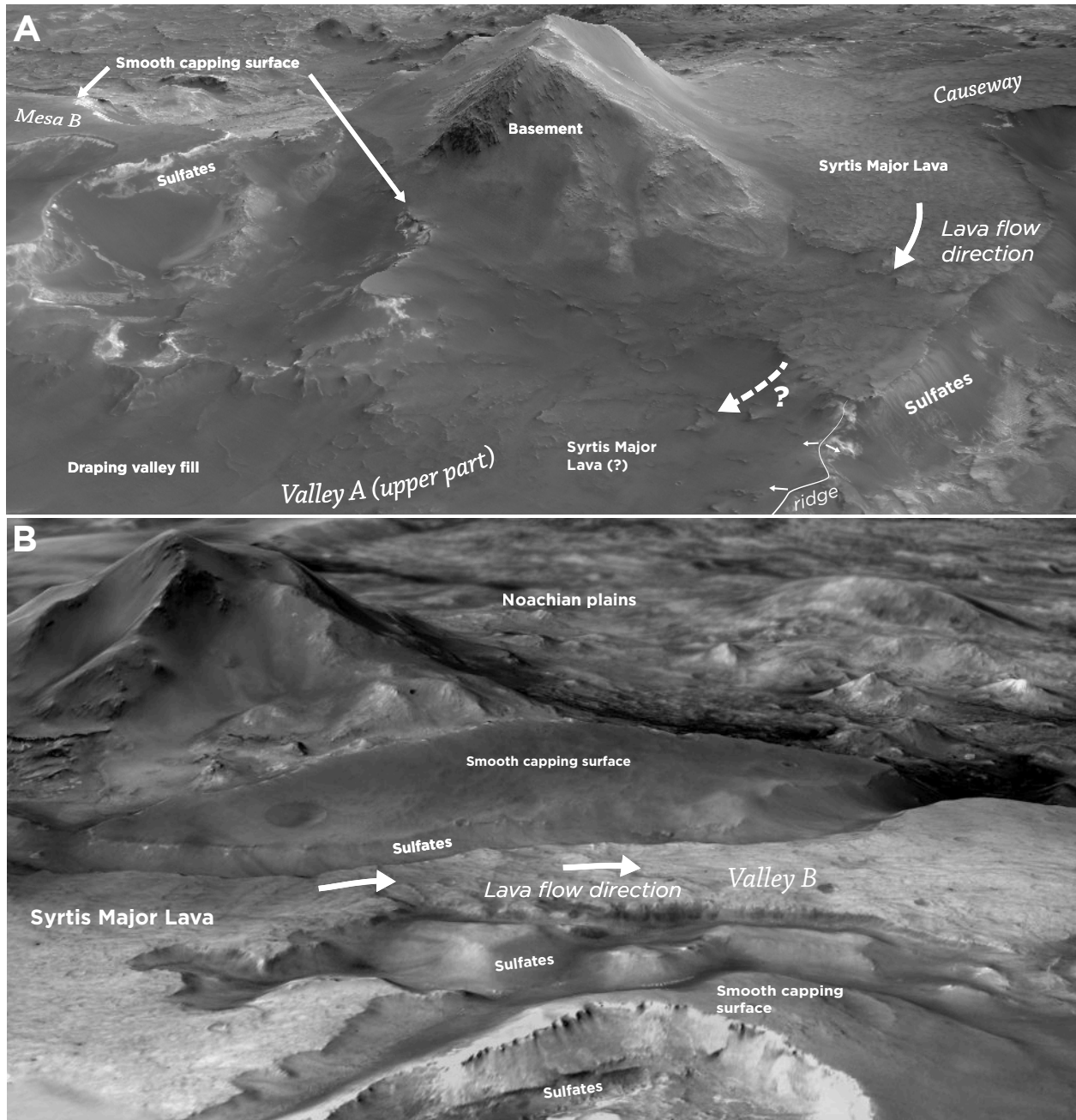


Figure 19: Key examples of the relationship between the smooth capping surface and the Syrtis Major Lavas, with orthoimagery atop CTX and HiRISE elevation models. (a) view southwestward towards *Mesa B*. The Syrtis Major lava flow lobe of the *Causeway* ends in the foreground, embaing the basement exposures and flowing down a gradient towards *Valley A*. The smooth capping surface crops out ~200 m higher on the south flank of this basement exposure. (b) View of *Valley B* with the smooth capping surface mantling layered sulfates on the flanks of the valley. The Syrtis Major lavas flows are channelized between these elevated exposures, and secondary channels have eroded the edge of the lava flow.

582 channels on the flow surface, and to the similar canyon at the flow scarp in nf4 [Figure 6c], which
583 feeds into *North Basin*.

584 An inverted channel cuts fractured exposures of the layered sulfates within the *Erosional Win-*
585 *dow* at the southern part of the study area [Figure 20c]. Layers within this channel body are nearly
586 flat, dipping at most 1° to the east. The channel flows north of a Noachian basement peak, over-
587 tops and cuts boxwork ridges within the layered sulfates, and is confined to a narrow belt roughly
588 300 m wide. The channel deposit aligns with valleys cut into the edge of the Syrtis Major lavas to
589 the west [Figure 3], and a valley at the northern edge of the Syrtis Major lavas that results from
590 focused erosion of the layered sulfates along this boundary.

591 **4.5.3 Basin-floor deposits** Parts of *North Basin* and *Deep Basin* are floored with flat, smooth sur-
592 faces that suggest fluvial or lacustrine deposition. These low-elevation exposures form nearly flat
593 surfaces (<1° east dips) within an interconnected network of basins; dipping sediments suggestive
594 of alluvial fans are sometimes found at the margin of these surfaces [Figure 21]. In *North Basin*,
595 the fluvial system discussed above feeds into these basins, and an outlet channel is preserved to
596 the east [Figure 3]. The sourcing of associated channels atop the Syrtis Major lavas suggests Hes-
597 perian or later deposition, and crater-counting of *North Basin* surfaces (*Skok and Mustard, 2014*)
598 yields an Amazonian age of 1.29 Ga. These surfaces were mapped as “Capping unit” by *Bramble*
599 *et al.* (2017) but appear to be fundamentally different than the exposures of the same morpholog-
600 ical unit at higher elevations atop the *NE Syrtis Plains*. *Bramble et al.* (2017) grouped these units
601 based primarily on their crater-retentive character, but they are likely much younger than these
602 cratered plains. We reinterpret these surfaces as fluvial/lacustrine sediment sheets, and suggest
603 that they are late deposits, based on their association with clearly post-sulfate fluvial systems.
604 We map these deposits as “Basin Floor” on Figure 3.

605 Low-relief layered scarps in HiRISE scenes nf1, nf3, and nf5 in *North Basin* are mapped as
606 undifferentiated sedimentary fill, closely associated with the Basin Floor deposits. These features
607 are associated with topographic scarps ~10 to 20 m high and are flat-lying to ~2° east-dipping,
608 similar to the adjacent Basin Floor surfaces. In nf3, they increase in dip to ~10° over ~100 m of
609 elevation, grading into thin fan deposits mantling the Noachian basement [Figure 21]. In nf2
610 and nf4, similar layered scarps at somewhat higher elevation are associated with the “Channel
611 Fill” deposits discussed above.

612 In nf2 and nf3, these scarps contain planar, bright-toned layers with CRISM signatures in-
613 dicative of phyllosilicates. In *Bramble et al.* (2017), these exposures are mapped as “Undifferen-
614 tiated” and are typically adjacent and slightly above exposures of the “Capping Unit”. The thin
615 packaging of beds, confinement to deep basins in present-day topographic lows, and formation
616 of crater-retaining flat floors contrast with the thickly packaged layered sulfates. We interpret
617 these features as representing late fluvial and lacustrine deposits, and map them as “Late basin
618 sediments” [Figure 3].

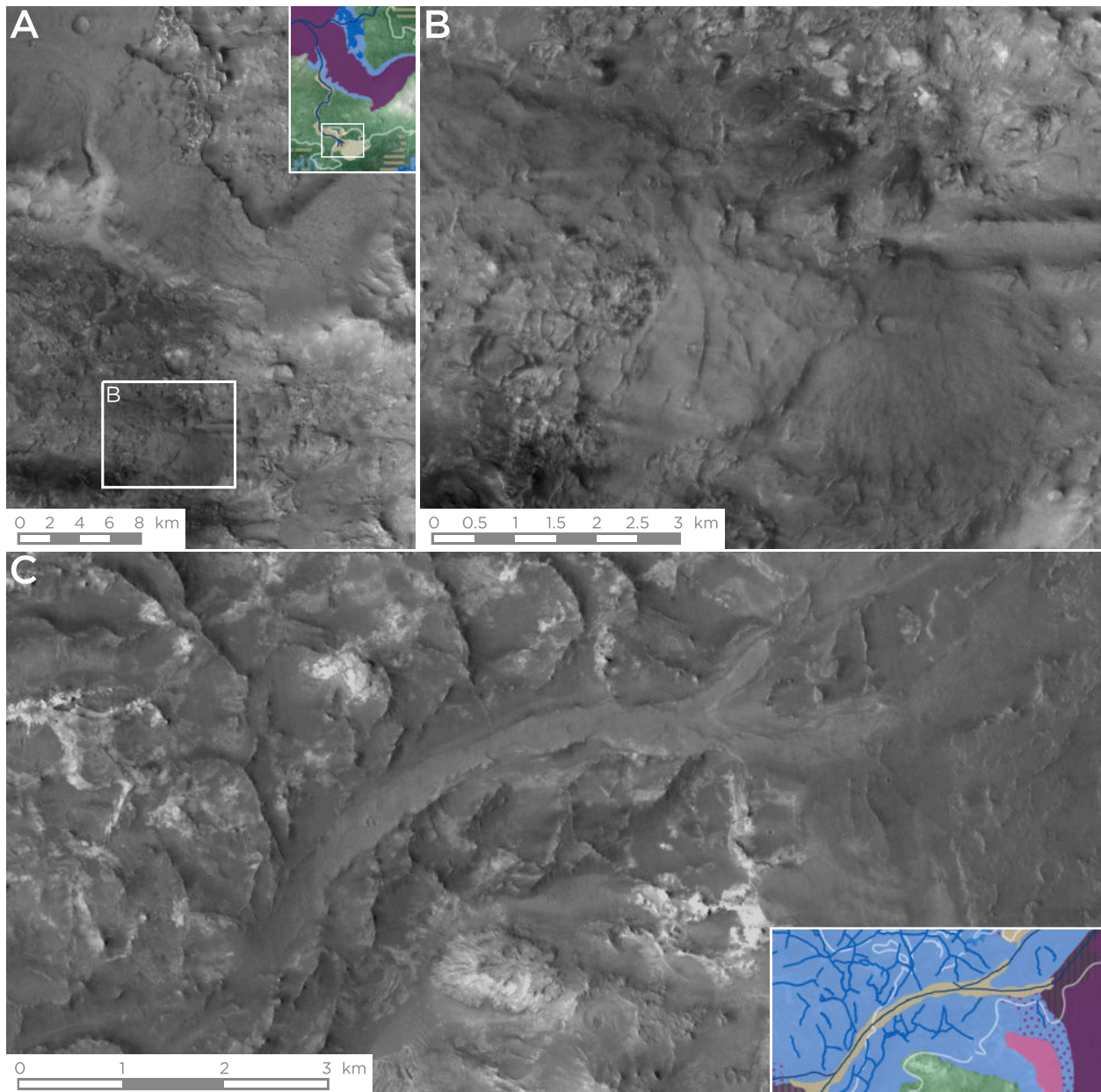


Figure 20: (a) Canyons cut into the edge of the Syrtis Major lava flows in the northwest of the study area, with a preserved downstream channel system and a 1.5 km-wide delta deposit at roughly -2300m elevation at its terminus. Inset geologic map covers the same area. (b) Close-up of the delta deposit, with basement megabreccia in the upper right. (c) An inverted fluvial deposit ~10-20 m thick in the *Erosional Window* in the southwestern part of the study area. The fluvial deposit is sourced from atop the Syrtis Major lavas. Inset geologic map is of the same area.

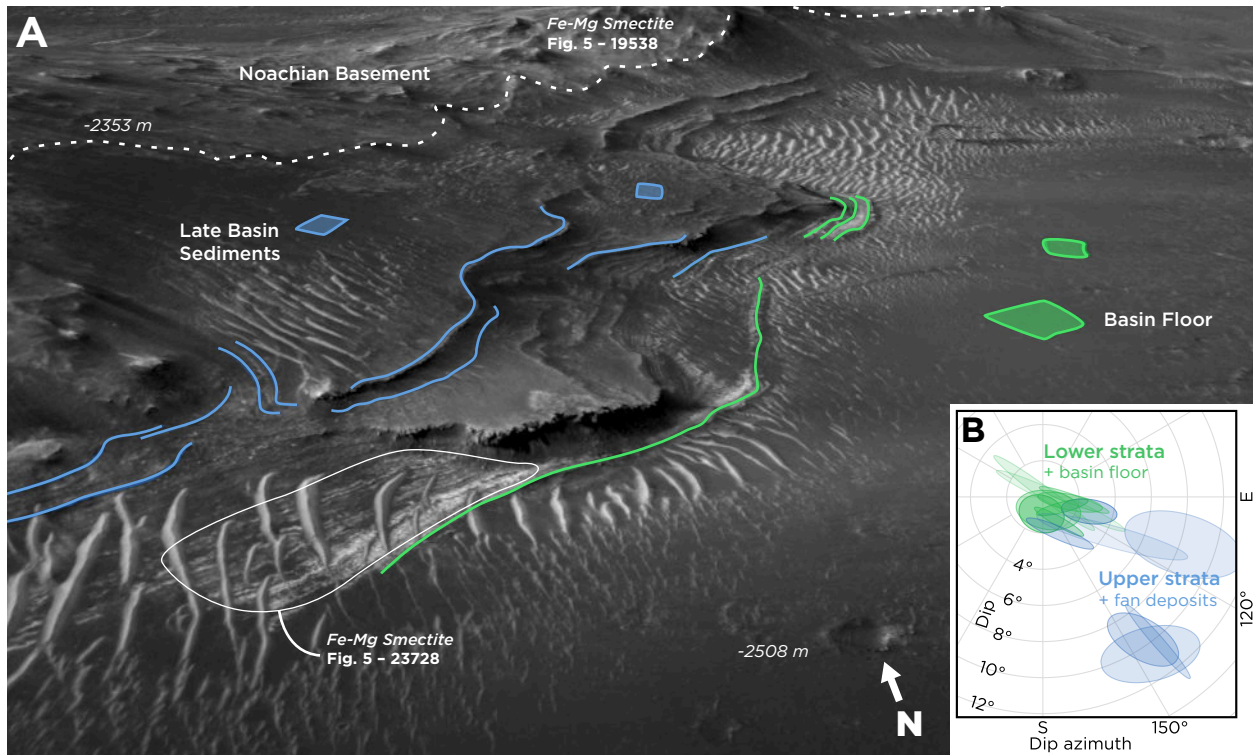


Figure 21: N-looking oblique view in nf3 (no vertical exaggeration) showing the Basin Floor units and layered scarps and fans of the Sedimentary Fill deposits. Measured bedding traces (green for the lower scarp and basin floor, blue for the upper fan) are shown. Calculated orientations for these traces (bottom right) show upper fan deposits dipping up to 10° into the basin. Approximate locations for CRISM spectra [Figure 5] showing smectite clay signatures within the basement and detrital sediment are shown. View is ~1 km wide.

619 **4.5.4 Draping valley fill** *Valley A* does not contain a clear channel system, despite the significant
620 erosion of *Mesas A* and *B* on its flanks and its overall southward slope. It is instead floored by the
621 “Draping Valley Fill”, a unit unique to the interior of *Valley A* [Figure 22]. Within the footprint
622 of nf7, this unit is characterized by a flat surface with $<2^\circ$ slopes to the southeast (truncated
623 at a 40 m layered scarp in the foreground of Figure 22a). Continuous with the surface capping
624 the valley floor, resistant thin surfaces dip steeply (15-20°) into the valley, mantling the layered
625 sulfates below and forming distinctive, sinuous hogback ridges up to 40 m high at its erosional
626 boundary [Figure 22b]. The valley floor surface hosts fine cracks near its edge, just inside of the
627 eastern slope of the valley [Figure 22c]. We interpret these as tension cracks caused by differential
628 deflation of the unit after emplacement. The draping valley fill has low thermal inertia and shows
629 indistinct mafic infrared signatures. The restricted elevation range and draping sedimentary style
630 of this unit suggests that it was formed during partial inundation of a previously-existing valley.

631 The draping valley fill mantles the entire bottom of *Valley A*, and extends north to an uncertain
632 boundary with a lobe of the Syrtis Major lava flows that extends southeast towards the head of
633 the valley. This zone (the foreground of Figure 19a) suggests that the lava flow terminates at
634 the northwestern margin of the draping valley fill. It is also possible that the Syrtis Major flows
635 continue down-gradient into *Valley A* beneath a thin veneer of draping valley fill; however, lavas
636 are not exposed at the surface anywhere further down the valley.

637 **4.5.5 Elevation alignment of fluvial features** With the exception of the inscribed channels,
638 the late fluvial-lacustrine features discussed above occur at or below -2300 m across the entire
639 study area. This elevation (highlighted on Figure 2, Figure 3, and Figure 9) is an open contour
640 connecting all the major valleys in the study area, including the closed *Deep Basin* and *Erosional*
641 *Window*. The delta shown in Figure 20 has topsets just below an elevation of -2300 m, and the
642 “basin floor” surfaces and associated draping sedimentary deposits within *North Basin* and *Deep*
643 *Basin* [Figure 21] are at or below this level (as low as -2800 m in *Deep Basin*). In *North Basin*,
644 flat basin-floor surfaces cover most of the area within this contour. The Draping Valley Fill is
645 also consistently associated with this contour, draping the interior of *Valley A* to an elevation
646 of -2300 m over the entire length of the valley; -2300 m is also the transition elevation between
647 the Syrtis Major lavas and the draping valley fill [Figure 19a]; this transition zone at the upstream
648 end of *Valley A* is separated from *North Basin* by the narrow *Saddle Ridge*, which has elevations just
649 over -2300 m. The alignment of post-sulfate fluvial and lacustrine deposits at a single elevation
650 suggests that this stage of deposition involved inundation of the entire study area to a single base
651 level. The level is similar to the -2320 m pre-incision elevation of Jezero crater’s eastern outflow
652 channel (Fassett and Head, 2005).

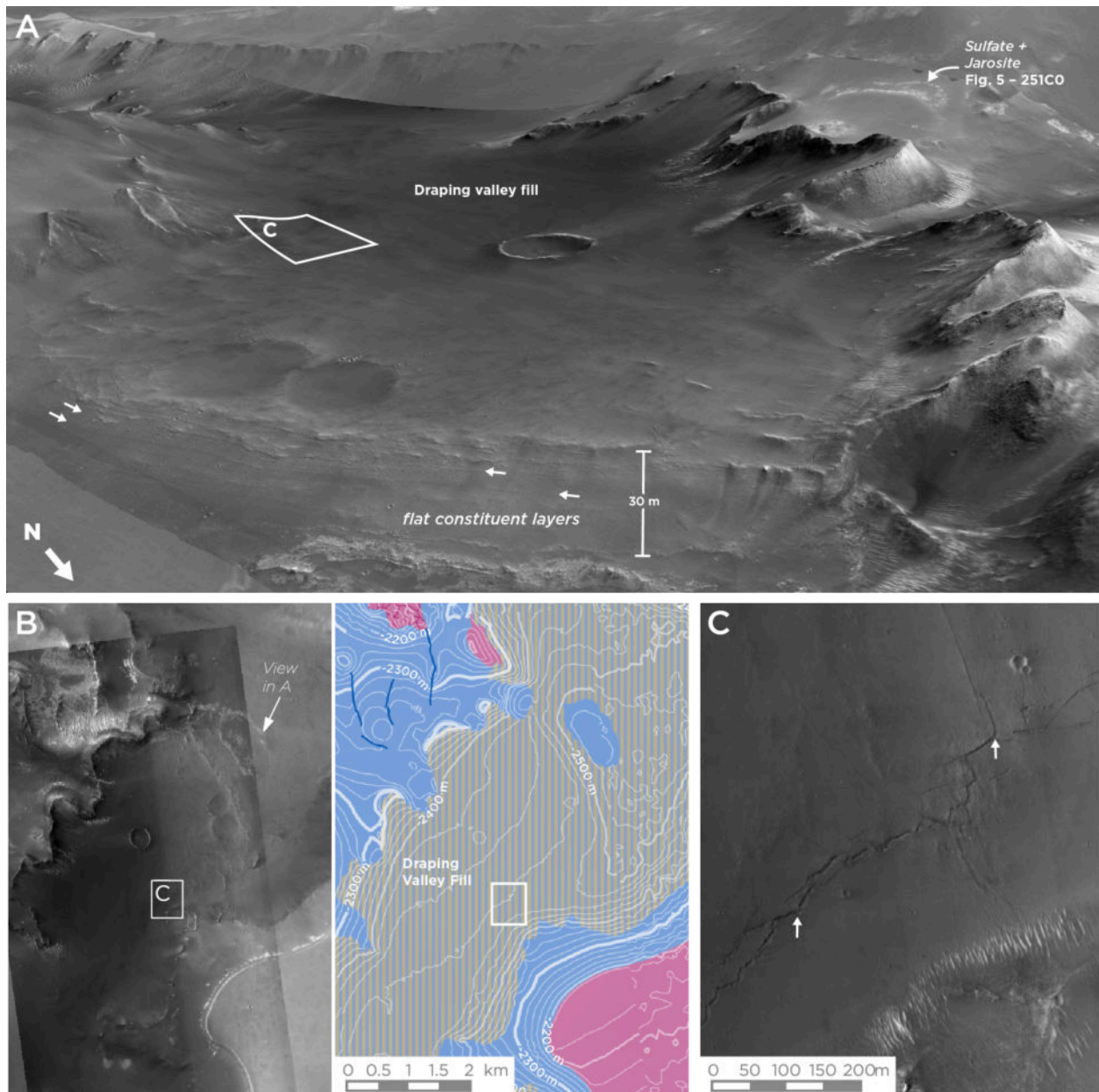


Figure 22: The draping valley fill in Valley A. (a) Southwest-looking HiRISE (nf7) oblique view of the unit, showing its nearly-flat base and a rim that extends up the side of the valley to roughly the -2300 m contour. The raised edges of the unit appear thin and buttress the material beneath them, and the inconsistent erosion creates a sinuous map pattern on the edge. (b) Paired map views of corresponding to a, showing imagery datasets and elevation overlaid on geology. (c) Cracks in the unit located just inside the slope break at the edge of the valley floor.

653 5 Discussion

654 5.1 Deposition of the layered sulfates

655 The sulfates have five key characteristics: (1) parallel, closely-spaced bedding, which indicates sed-
656 imentary accumulation; (2) poor induration and susceptibility to erosion; (3) dips always $<10^\circ$ and
657 mostly $<7^\circ$ with regionally variable orientations; (4) variable unit thickness, ranging up to 600 m;
658 (5) unconformable emplacement on and thinning up to basin highs. Gentle structural folding can
659 be excluded as a cause for dipping layers in the layered sulfates. The generation of folds requires
660 basement-involved tectonic shortening that is not evident in the NE Syrtis region. Additionally,
661 dip directions within the sulfates do not conform to a consistent fold pattern, and no tectonic
662 stresses were present during diagenetic fracturing of the layered sulfates [Section 4.2.2]. Conse-
663 quently, shallow but non-zero dips within the layered sulfates are best explained as depositional
664 dips; we seek a sedimentary mechanism that can form thickly packaged, regionally extensive,
665 shallowly dipping strata.

666 Depositional mechanisms that operate by sediment traction [Figure 4] are a poor fit for the
667 layered sulfates. Shallow lake, evaporite playa, or eolian sedimentation could form the observed
668 laterally continuous, meter-scale layering, but these mechanisms typically form deposits that on-
669 lap pre-existing topography and fill localized, low-lying basins, with near-zero depositional dips
670 (the “Basin Floor” unit is interpreted to represent this type of environment) rather than thinning
671 against pre-existing topographic highs. Deposits formed by fluvial networks would additionally
672 be limited in extent and associated with clear erosive and constructional features such as valley
673 networks and inverted channel casts (*Fassett and Head, 2008; DiBiase et al., 2013*). The sulfates
674 extend at least 50 km east of the Isidis inner rim, too far to have been formed from proximal allu-
675 vial sediments shed from local topographic highs, as has been proposed for some layered deposits
676 in Valles Marineris (*Fueteu et al., 2011*). Thick eolian sedimentary deposits generally have cross-
677 bedding, which is observable at orbital HiRISE scale in Gale Crater (*Milliken et al., 2014*) but not
678 at Meridiani Planum (e.g. *Grotzinger et al., 2005; Hayes et al., 2011*); eolian deposits also tend to
679 form horizontal bedsets due to water-table driven cementation (*Kocurek, 1981*), which are not
680 observed. We consider three sedimentation mechanisms viable, all of which entail fallout from
681 suspension: distal ash fall, dust deposition from ices, and deposition in a deep lake (or lakes).

682 There is significant global evidence of explosive volcanism on Mars (e.g. *Brož and Hauber,*
683 *2012*), and ash falls have been suggested as a likely depositional mechanism for other layered
684 deposits on Mars (*Kerber et al., 2011, 2012*), including in the eastern Medusae Fossae Formation,
685 which hosts polygonal ridges interpreted as filled fractures (*Kerber et al., 2017*) that have simi-
686 larities to those within the NE Syrtis layered sulfates. While Syrtis Major is mostly an effusive
687 basaltic province (*Hiesinger and Head, 2004*), there is significant evidence that Nili Patera in its
688 center hosted major pyroclastic eruptions (e.g. *Fawdon et al., 2015*). *Ghent et al. (2012)* finds sig-

689 natures of devolatilization of a substantial amount of pyroclastic material within Isidis Planitia,
690 but these postdate the Syrtis Major lava flows.

691 The layered sulfates show no structures associated with chaotic pyroclastic emplacement or
692 devolatilization that would be expected for large-volume ash flow deposits (e.g. *Ghent et al.*, 2012).
693 Instead, the dozens of similarly-thick (0.5-2.5 m) beds with no evidence of unconformities inter-
694 nal to the deposit would require airfall ash emplacement from consistently small or distant erup-
695 tions, regularly paced and close in time. While ash fall cannot be excluded based on the structural
696 form of the layered sulfates, the regularly-spaced bedding lacking internal unconformities does
697 not easily match a stochastic process such as pyroclastic volcanism.

698 Loess deposition, in which obliquity-mediated climate cycles drive the accumulation of air-
699 fall dust, likely explains some Martian layered sedimentary rocks (*Lewis et al.*, 2008; *Bridges and*
700 *Muhs*, 2012; *Lewis and Aharonson*, 2014). *Niles and Michalski* (2009) suggested that layered sul-
701 fates might occur as sublimation residues of ice-rich sediments. Crudely layered dust mounds left
702 over from progressive sublimation of these deposits could exhibit shallow dips and drape prior
703 topography (*Michalski and Niles*, 2012; *Michalski et al.*, 2013). Again, the number of layers, their
704 regularity in size, and most importantly the lack of internal unconformities is difficult to explain
705 by any obliquity-driven sedimentation process. Additionally, sediment volume loss under air-
706 fall accumulation would entail gradual dehydration, dewatering, or sublimation rather than the
707 single-stage, throughgoing episode of volume loss that impacted the layered sulfates [see Sec-
708 tion 4.2].

709 If the layered sulfates were formed by airfall or ice deposition, correlative deposits should be
710 found elsewhere in the study area, including at higher elevations; none are evident at NE Syrtis.
711 Like airfall mechanisms, deep lacustrine sedimentation operates by fallout from suspension; it
712 additionally forms deposits with regional architecture similar to the NE Syrtis layered sulfates.
713 Terrestrial deepwater sedimentation occurs at large scale along passive continental margins, en-
714 vironments with both a steady supply of sediment and steep underwater topography (e.g. *Stuart*
715 *and Caughey*, 1977). Deepwater sedimentary packages can dip relatively steeply ($\sim 5^\circ$ depositional
716 dips are common) while maintaining internally parallel geometries; bedding both onlaps and dips
717 concordantly with pre-existing topography. Depositional sequences are limited in thickness by
718 available accommodation space (i.e. the water depth). Prograding sediments accumulate outward
719 from the basin margin, and deposits thin and decrease in elevation into the basin (*Mitchum et al.*,
720 1977). The geometry of pre-existing topography, along with relationships between individual
721 stratal sequences, leads to a diversity of bedding orientations with dominantly but not exclusively
722 basinward dips (*Mitchum et al.*, 1977). Deepwater sedimentation operated on Mars in crater lakes
723 (*Grotzinger et al.*, 2015), and deepwater deposition without regional topographic confinement has
724 been proposed to explain large-scale sedimentary features observed from orbit in Valles Marineris
725 (e.g. *Dromart et al.*, 2007) and in the Northern Plains (*Oehler and Allen*, 2012).

726 Major bedding features of the NE Syrtis layered sulfates, such as variable bedding dips (up to
727 7-10°) and parallel-bedded stratal packages, are typical of deepwater sediments. Regionally, the
728 extent, lack of confinement to localized basins, and onlapping and embayment of pre-existing
729 highs also suggest this type of sedimentation. Both the preserved thickness and overall elevation
730 of the layered sulfates decrease eastward into Isidis Basin [Figure 9d], although the original thick-
731 ness may be masked by erosion. Collectively, the structural characteristics of the layered sulfates
732 are typical of deposits at the margin of a deep, water-filled basin. However, deposits formed by
733 fallout from atmospheric suspension cannot be fully excluded based on layer orientations and
734 deposit thickness alone.

735 **5.2 Post-depositional alteration of the layered sulfates**

736 If the layered sulfates formed by deepwater sedimentation or any non-evaporative process, their
737 sulfate mineral signature must arise from interaction with groundwater after deposition; ample
738 morphologic evidence of such groundwater activity exists.

739 **5.2.1 Interpretation of boxwork fractures** The large scale and continuity, throughgoing nature,
740 and positive relief of boxwork polygonal ridges are typical of injective dikes; however, the detailed
741 morphology, mineralogy, and structural form of these ridges [Section 4.2] instead clearly indicate
742 a two-stage formation history: fracturing of the layered sulfates in a polygonal pattern followed
743 by later mineralization channelized along fracture surfaces.

744 Tectonically controlled ridges are pervasive in the phyllosilicate-bearing Noachian basement
745 of the NE Syrtis region. These features, classified as “Nili-type” ridges by *Kerber et al.* (2017),
746 follow the NE-SW regional trend of the Nili Fossae circum-Isidis ring graben (*Saper and Mustard,*
747 2013). The filled fractures in the overlying layered sulfates demonstrate a fundamentally different
748 structural form: the absence of directional bias in ridge-orientation measurements implies that
749 they were not tectonically controlled. Instead, fracturing occurred in an isotropic regional stress
750 field, under volume loss and internal contraction.

751 Several mechanisms are known to generate polygonal fractures through volume loss in sedi-
752 mentary material. Melting is a common process forming polygonal “patterned ground” on Earth
753 (e.g. *Kocurek and Hunter,* 1986) and on Mars (*El-Maarry et al.,* 2010). However, it generally occurs
754 at near the free surface and involves sagging and large volumetric reductions that can disrupt
755 or destroy internal layering (*Soare et al.,* 2017). Typical dessication polygons such as mud cracks
756 are generally vertical and tied to the free surface; they tend to form hexagonal patterns due to
757 progressive annealing. By contrast, fractures in the sulfate unit (1) have 90° preferred intersec-
758 tion angles and “T-junctions”, (2) are both straight and curvilinear, (3) penetrate the full exposed
759 thickness of the layered sulfates, and (4) are steeply-dipping as well as vertical [Figure 14]. These
760 characteristics show that the fractures resulted from a single stage of volume loss (*Goehring et al.,*
761 2010) and were not formed at the free surface.

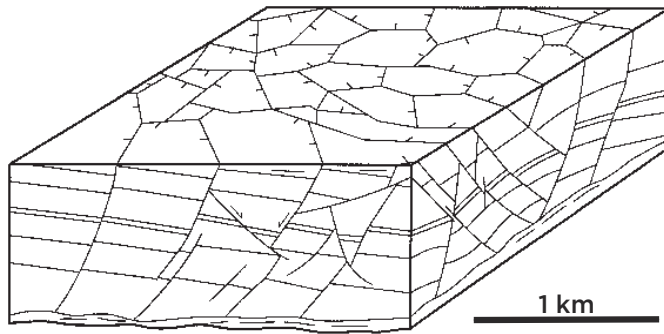


Figure 23: Schematic view of boxwork fracturing compiled from seismic well logs in the North Sea after *Cartwright and Lonergan* (1996). The polygonal geometry forms in the subsurface due to compaction during diagenesis, and is found at similar scales in offshore basins on Earth as the exposures in Syrtis Major.

762 One type of fracture that matches these characteristics is three-dimensional “polygonal fault-
 763 ing” (*Goultly*, 2008), which forms during diagenesis and dewatering of cohesive, clay-rich or chalk-
 764 rich sediments. These features are often found in shallow offshore sedimentary basins on Earth
 765 (*Cartwright and Lonergan*, 1996). Most examples of polygonal faulting on Earth are at similar
 766 scales to that examined in this study with 500 m polygon domains typical [Figure 23]. The scale
 767 of boxwork domains is controlled by the strength and cohesion of sediments undergoing diagen-
 768 esis (*Cartwright and Lonergan*, 1996; *Goultly*, 2008). Most examples of polygonal fractures occur
 769 a few hundred meters below the seafloor in continental-margin sedimentary basins; these have
 770 been investigated solely with seismic imaging (e.g. *Gay et al.*, 2004). *Tewksbury et al.* (2014) doc-
 771 uments a rare surface-exposed polygonal fault network in Egypt.

772 Key to polygonal faulting is that during compaction-driven dewatering and diagenesis, vol-
 773 ume loss creates localized extensional stresses. The exact mechanism causing the initiation of
 774 polygonal faults is unclear, but synaeresis (e.g. *Siebach et al.*, 2014), overburden, and density in-
 775 version are several possibilities. The depth of burial required for faulting to initiate is unclear,
 776 but depths of 100 m or more are typical in submarine settings (*Goultly*, 2008; *Cartwright*, 2011).
 777 Once faults are initiated, the numerous “T-junction” and right-angle crossing fractures charac-
 778 teristic of this mechanism are due to a preference for linear defect propagation as dewatering
 779 progresses. Three-dimensional material shrinkage is converted to one-dimensional compaction
 780 by small-magnitude slip along fault surfaces. This suggests small bedding offsets and dip changes
 781 between boxwork domains, matching the character of boxwork in the layered sulfates.

782 On Mars, *Oehler and Allen* (2012) proposed polygonal faulting as a mechanism to explain ~2-
 783 10 km polygons expressed on the surface of Acidalia and Utopia Planitia, using this to argue for
 784 their formation in a subaqueous setting. Similar features have also been attributed to periglacial
 785 processes (e.g. *Haltigin et al.*, 2014). Unlike these examples, the northeast Syrtis layered sulfates
 786 show the full 3D geometry of the fault network, which allows a much clearer identification of
 787 polygonal faulting, since the ridge characteristics match the scale, morphology, and penetrative
 788 nature of the fractures.

789 **5.2.2 Implications for sediment size and amount of burial** Based on the character of Earth
 790 analogs, polygonal faulting has a particular set of implications for sediment characteristics and di-

791 agenetic environment. Polygonal faults dependably involve dewatering of fine grained (silt/clay),
792 water-rich sediments under three-dimensional compaction in the subsurface. Dewatering solely
793 by compaction might require a few hundred meters of sediment above presently exposed ridges
794 in the layered sulfates. However, dewatering may require less overburden in cases where drying
795 conditions are favored by thermal/climatic factors (e.g. arid, evaporative conditions). Boxwork
796 fracturing is chiefly found in the western part of the study area, and most areas of boxwork cluster
797 in basin lows, which are predicted to be sufficiently buried based on projection of current
798 exposures [Figure 9]. A clear exception is the boxwork fractures in nf4, which occur within some
799 of the highest-elevation exposures of the sulfates in the study area [Figure 6c]. The presence of
800 boxwork at these high elevations requires either substantial overburden at levels above the currently
801 preserved sulfates or volume loss without significant burial, possibly by changes in the
802 hydration state of water-bearing minerals.

803 The chief factor permitting polygonal faulting for fine-grained sediments is the internal friction
804 angle, which sets the stress threshold to initiate fracturing. For silt- and mud-sized sediment,
805 low internal friction allows defects to form and propagate under small, localized stresses
806 (*Cartwright and Lonergan, 1996; Goult, 2008*). Polyhydrated sulfates often have >10 wt. % water,
807 and fine-grained polyhydrated sulfate sediments would have characteristics conducive to polygonal
808 faulting. Fracture mineralization is also indicative of fine-grained sediments: channelization
809 of altering fluids along fractures suggests that the bulk of the unit has low permeability. In contrast,
810 sand-sized sediments have open pore spaces and easily permit fluid migration.

811 The formation of boxwork polygonal ridges by the mechanism outlined above requires thick,
812 initially water-saturated, fine grained sediments that undergo single-phase volume loss after deposition
813 of the entire package. As such, boxwork fractures align with the unit's structural geometry and layer
814 orientations to suggest a deep lacustrine sedimentary origin for the layered sulfates.

815 **5.2.3 Mineralization of fractures and implications for water volume** The parallel-sided, or isopachous,
816 geometry of boxwork fractures, their current existence as resistant ridges, and a mineralogy distinct
817 from the rest of the unit indicates the mineralization of pre-existing fractures. The diffuse "halo"
818 around some fractures [Figure 12] represents an interaction between fracture-filling material and a
819 relatively impermeable groundmass. These morphologies can be formed by either fracture-filling
820 cements that close an open fracture inwards or an outward-propagating zone of alteration around a
821 channel carrying reactive fluid (e.g. *Nelson et al., 1999*). Such a fluid would either chemically alter
822 the groundmass of the unit or fill pores with light-toned cement.

823 The margin of Isidis basin is modeled as an area of groundwater upwelling (*Andrews-Hanna and Lewis, 2011*)
824 and the deep basins in NE Syrtis would have the strongest topographic gradient in the region. This may
825 suggest the focusing of abundant groundwater into the base of the layered sulfates to drive fracture
826 mineralization. The presence of jarosite mineral detections on the ridges indicates infiltration of
827 pH<4 fluids (e.g. *Ehlmann and Mustard, 2012; McCollom et al., 2013*). Fluid

828 leaching combined with induration has been proposed as a mechanism to develop haloed fractures
829 in Candor Chasma (*Okubo and McEwen, 2007*). Double-walled fractures similar to those seen here
830 have been identified from orbit in smaller (decameter-scale) boxwork fractures in the upper Gale
831 Crater mound (*Siebach and Grotzinger, 2014*). For the Gale Crater system, a mass-balance given
832 plausible limits on pore-water volume suggested that the formation of $1.75 \times 10^6 \text{ m}^3$ of cemented
833 fractures required the evaporation of at least $\sim 0.4 \text{ km}^3$ of water (*Siebach and Grotzinger, 2014*).
834 Using the same 30% porosity and mineral-precipitation assumptions and a fracture volume of
835 0.86 km^3 (860 km of fractures mapped, an average vertical penetration of 200 m, and a fracture
836 width of 5 m), we estimate that the evaporation of $\sim 515 \text{ km}^3$ of water was required to mineralize
837 the fractures within the layered sulfates.

838 In a few locations, localized, pervasive jarosite mineralization within the groundmass of the
839 sulfate unit is unchannelized and at the high end of alteration intensity observed [Figure 15]. In-
840 tricately patterned exposures in the *Erosional Window* could be driven by surface water as well
841 as groundwater with mineralization occurring in a transient, evaporitive lake fed by periodic
842 outflow-channel inundation (*Mangold et al., 2008a*).

843 **5.3 Comparison with other Martian layered deposits**

844 The layered sulfates at NE Syrtis share some features with other layered deposits in the martian
845 mid-latitudes, many of which show sulfate mineral signatures, but range in local characteristics
846 and structural context.

847 Many thick layered deposits on Mars have been attributed to deepwater sedimentation. The
848 sulfate-bearing interior layered deposits (ILDs) in Valles Marineris occur in steep-walled, con-
849 nected basins that could host deepwater deposition (e.g. *Lucchitta et al., 1994; Warner et al., 2013*);
850 sedimentary structures typical of deepwater, prograding sediments have been identified in the
851 walls of Melas Chasma (*Dromart et al., 2007*) at elevations of $\sim 1700 \text{ m}$. In other areas, warped,
852 folded and locally dislocated strata suggest dipping sedimentation and soft-sediment slumping
853 typical of deepwater environments (*Okubo et al., 2008; Roach et al., 2009; Metz et al., 2010*).

854 Evaporite or evaporite-mediated eolian sedimentation has also been suggested to form some
855 Valles Marineris ILDs (*Roach et al., 2009; Murchie et al., 2009; Flahaut et al., 2010*). Possibly re-
856 lated sulfate-bearing layered deposits on the adjacent highland plains (*Fueteu et al., 2011*) could
857 not have formed in any conceivable deepwater basin. The deposits on the highland plains co-
858 occur with preserved channel networks (*Mangold et al., 2008b*) and were likely formed by fluvial
859 or lacustrine processes, although ash fall and aeolian deposition with later weathering are not
860 excluded (*Weitz et al., 2008*).

861 Sulfate-bearing sedimentary rocks are widespread at Meridiani Planum; those investigated
862 by the Opportunity rover contain aeolian sediments reworked by fluvial processes (*McLennan*
863 *et al., 2005; Grotzinger et al., 2005*). These deposits may be related to layered, sulfate-bearing de-
864 posits that mantle the adjacent highlands at Arabia Terra. These deposits are relatively thick (up

865 to 200 m), internally layered and capped by a thin, resistant unit (*Fassett and Head, 2007*). Like
866 the Valles Marineris rim deposits, sulfate-bearing sediments at Arabia Terra are perched on the
867 highlands and cannot be related to deepwater deposition. Instead, hydrological modeling sug-
868 gests formation due to a perched water table mediating detrital sedimentation and groundwater
869 upwelling and evaporation in the highlands during the Noachian-Hesperian transition (*Andrews-*
870 *Hanna et al., 2007, 2010; Zabrusky et al., 2012*).

871 At Sinus Meridiani, thick, enigmatic sulfate-bearing “etched terrain” units are stratigraphi-
872 cally beneath the units investigated by the Opportunity rover (*Wiseman et al., 2010; Michalski*
873 *et al., 2013*). Layered deposits of uncertain origin form a thick, partially incised bedrock sequence
874 at Aeolis Dorsa (*Kite et al., 2015*). Thick layered deposits also form the substrate for valley net-
875 works and paleolake deposits within Melas Chasma (*Quantin et al., 2005*). These sulfate-bearing
876 layered deposits are unconfined by topography but otherwise similar to sequences that occur in
877 bounded crater basins, such as Aram Chaos (*Lichtenberg et al., 2010*).

878 The construction of thick sedimentary sequences, including the NE Syrtis sulfates, in either
879 sediment-filled or open lacustrine basins could straightforwardly relate to the global groundwater
880 system modeled to drive aggradation of the perched Arabia Terra deposits (*Andrews-Hanna et al.,*
881 *2007, 2010*). These models imply that low-elevation basins would contain either thick evaporite
882 sequences supported by high water tables or open-basin bodies of water fed at least partially by
883 upwelling groundwater.

884 Understanding of formation mechanisms of thick, low-elevation layered deposits and their
885 potential links with NE Syrtis sulfates is inhibited by poorly-known layer orientations. The Meridi-
886 iani deposits have low dips ($<1^\circ$) based on regionally-traced layers (*Hynek and Phillips, 2008*) but
887 no localized bedding information is captured. Many deposits have rigorously described mineral-
888 ogy but more ambiguous internal structure (*Roach et al., 2009; Wiseman et al., 2010*). Future work
889 on these deposits using the structural approach taken by this study will illuminate any structural
890 similarities to the NE Syrtis sulfates, and allow our proposed deepwater sedimentation mecha-
891 nism to be tested in more places. Alternative hypotheses of airfall or ice-mediated deposition can
892 also be evaluated.

893 **5.4 Containing basin for the layered sulfates**

894 A key question for the deep subaqueous sedimentation hypothesis is the geometry and confine-
895 ment of the containing basin. Deposition of the sulfates in a deep subaqueous setting would
896 require a basin filled to about -1600 m, the maximum elevation of the layered sulfates in the re-
897 gion [Figure 9d]. Given the location at the edge of Isidis Basin, the basin is not confined by current
898 topography. Two possible mechanisms to inundate the NE Syrtis region deeply enough to deposit
899 the layered sulfates include a global ocean or a ice-dammed lake marginal to an ice sheet within
900 Isidis basin.

901 A basin at the margin of a global ocean would provide a straightforward analog to Earth. How-

902 ever, evidence for a global ocean is uncertain (e.g. *Ghatan and Zimbelman*, 2006), and the required
903 topographic level is hundreds of meters higher than putative ocean deposits (*Perron et al.*, 2007;
904 *DiBiase et al.*, 2013). An alternative is confinement of a large lake by a basin-filling ice sheet,
905 which could provide the regional topographic confinement and be a source of abundant meltwa-
906 ter. *Ivanov and Head* (2003) and *Ivanov et al.* (2012) suggested that ice sheets covered parts of the
907 Isidis basin rings and floor during the early to late Hesperian (~3.5-3.1 Ga) and eroded previously-
908 existing layered sediments, based on morphologic similarities of sinuous ridges to terrestrial,
909 subglacially-formed eskers. *Souček et al.* (2015) modeled the potential extent of a crater-filling
910 ice sheet given expected precipitation and climate, finding that an ice sheet would preferentially
911 mantle the NE Syrtis region relative to other parts of the Isidis rim. Such an ice sheet could dam
912 a regional lake including NE Syrtis, creating the conditions for deposition of the layered sulfates.

913 A crucial question associated with both the ice-dammed regional lake and hemispheric ocean
914 hypotheses is whether the climatic state supporting either of these phenomena would support a
915 lake for a sufficient cumulative lifetime to produce the observed sedimentation. If its meter-thick
916 beds were laid in uninterrupted annual cycles, the layered sulfates could have formed in as little
917 as ~1000 years. However, with reasonable constraints on sediment supply and intermittency,
918 deposition rates would likely be orders of magnitude slower. On Earth, lakes and small basinal
919 seas commonly show sedimentation rates of 0.1 to 10 m per kyr (*Sadler*, 1981), corresponding to
920 100 kyr to 10 Myr to accumulate the thickness of the layered sulfates. Abundant glacially-eroded
921 sediment at the margin of wet-based ice sheets can drive aggradation of ~60 m per ka (*Dowdeswell*
922 *and Siegert*, 1999). However, extrapolation from terrestrial sediment fluxes and environments
923 is of limited value for forecasting cyclical sedimentation rates in an unknown Martian surface
924 aqueous system.

925 The regional scale of proposed basin confinement, by either an Isidis-filling ice sheet or a hemi-
926 spheric ocean, also raises a key question of preservation: why are the layered sulfates only found at
927 NE Syrtis? No comparable deposits have been found elsewhere on the rim of Isidis Basin, though
928 the Libya Montes to the south are otherwise similar (*Bishop et al.*, 2013). Erosion clearly played a
929 major role: the layered sulfates are only preserved in association with more durable units such as
930 the Syrtis Major lavas and mineralized boxwork fractures. In the absence of the fortuitous cap-
931 ping or groundwater interaction seen at NE Syrtis, any sulfates deposited elsewhere could easily
932 be stripped away by wind erosion or later fluvial incision.

933 **5.5 Modification by fluvial erosion, lavas, and late lake deposits**

934 Another major finding of this study is continued erosion and fluvial activity postdating the layered
935 sulfates. The long sedimentary history after layered sulfate deposition includes several phases
936 of erosion, cap unit emplacement, further fluvial erosion, lava emplacement, and then still-later
937 fluvial-lacustrine erosion and deposition. No matter how the layered sulfates formed, the smooth
938 capping unit, Syrtis Major lavas, and late fluvial features formed significantly afterwards. This

939 history implies substantial episodic interaction with surface water significantly postdating the
940 formation of the layered sulfates.

941 Angular differences between the smooth capping surfaces and the underlying sulfates indi-
942 cate erosional truncation of the sulfates [Figure 6-19]. This unconformity suggests the erosion
943 of a significant volume of sulfates prior to cap unit formation. The observed smooth, featureless
944 surface, low thermal inertia, and low crater retentivity are at odds with the observed resistant
945 nature of the cap surface. A partly cemented sandstone, welded or later-indurated ash fall, highly
946 degraded lava flow, or capping “duststone” (*Malin and Edgett, 2000; Bridges and Muhs, 2012*) could
947 potentially generate the characteristics of the deposit. The jarosite-containing, light-toned “halo”
948 extending up to 30 m beneath the capping unit [Figure 6a] could indicate interaction with the un-
949 derlying sulfates during cementation of the capping surface or simply a weathered surface covered
950 by the capping material.

951 *Valley A* in the center of the study area is cut between *Mesas A* and *B*, which are both topped
952 by the smooth capping surface. *Valley A* and its upstream extension may be remnants of a major
953 fluvial system that flowed from the northern margins of the Syrtis Major volcanic province [Fig-
954 ure 2b]. Beginning in the early Hesperian, effusive Syrtis Major lava flows flowed through these
955 pre-existing fluvial channels and embayed the partially eroded layered sulfates at their southern
956 margin in multiple locations, notably *Valleys A* and *B* [Figure 19b]. Cap unit formation, paleo-
957 valley erosion, and embayment by lava flows did not occur as a single event, and these processes
958 may have been interleaved and closely spaced in time during a geologically active late Noachian
959 to Hesperian transition.

960 The latest fluvial systems within the study area start atop the Hesperian Syrtis Major flows
961 and erode the capped sulfates and Noachian basement, forming deltas and inverted channels.
962 The preserved fluvial and lacustrine deposits are relatively small, with sedimentary deposits at
963 most ~20 m thick atop the basement and layered sulfates, and amphitheater canyons cut back at
964 most 1 km into the Syrtis Major lavas. The basin-fill and associated layered scarps in *North Basin*
965 and *Deep Basin* show lacustrine deposition and phyllosilicate-bearing sediments; they indicate
966 inundation of substantial portions of the study area after erosion of the layered sulfates.

967 The deltas, basin-filling deposits, and draping valley fill described above occur at similar topo-
968 graphic levels within the study area. This open-contour alignment of late sedimentary deposits
969 suggests that they were deposited at a single base level, marginal to a open-basin lacustrine sys-
970 tem not bounded within the study area. The ~2300 m elevation of these features is similar to
971 that of the Jezero delta and outflow channel [Figure 2] and is near the elevation of the various
972 deltas and coastline features making up the proposed coastline of a hemisphere-spanning ocean
973 (*Di Achille and Hynek, 2010*). The presence of phyllosilicates in layered scarps of the late sedi-
974 mentary deposits at NE Syrtis [Figure 21] suggests that these deposits contain detrital material
975 similar to that in the Jezero delta (*Ehlmann et al., 2008b; Goudge et al., 2015*).

976 The final resolvable phase of lacustrine activity in the area built a series of lakes in intercon-
977 nected topographic lows within *North Basin*. The outlet channel leading eastward from *North*
978 *Basin* shows that a lake filled this basin to -2400 m and drained to the east. That this outlet
979 channel does not exit the basin at its current lowest-elevation location (-2550 m in nf 1) is consis-
980 tent with a prior blockage of this exit or later erosion of the outlet (*Skok and Mustard, 2014*). We
981 propose that the current lowest exit in nf 1 was blocked by layered sulfates prior to Amazonian
982 wind-driven erosion of the layered sulfates.

983 Outflow features continuous with those mapped across the Syrtis Major lava plains to the
984 south of the study area by *Mangold et al. (2008a)* suggest episodic, powerfully erosive flows across
985 the Syrtis Major lava plains. Flow across the lava plains may have also caused episodic inundation
986 of the *Erosional window* in the southwest part of the study area, driving the intense and localized
987 acid-sulfate alteration seen solely in this basin [Figure 15].

988 **6 Summary, Conclusions, and Future Work**

989 The layered sulfates at NE Syrtis Major form a thick (up to 600 m) sedimentary package uncon-
990 formable with the underlying Noachian basement and olivine-carbonate units. The polyhydrated-
991 sulfate enriched, basaltic-composition sedimentary rock contains parallel, meter-scale beds that
992 dip up to 7-10° with no preferred direction. The unit thins against and embays basement highs, to
993 a maximum elevation of -1600 m. Boxwork polygonal ridges in the layered sulfates at hundreds
994 of meters scale record fracture-generating volume loss during diagenesis, followed by the pre-
995 cipitation of jarosite-bearing material from fluids circulating through the fractures. Subsequent
996 erosion of the surrounding material exposed the mineralized fractures as raised ridges.

997 Overall, the upper stratigraphy at NE Syrtis Major was built by a multistage history of water-
998 related processes, likely spanning much of the Hesperian and perhaps into the Amazonian [Fig-
999 ure 24]: The sulfates record (1) deposition atop the olivine-carbonate and Fe/Mg phyllosilicate
1000 Noachian plains units, likely in a deepwater, subaqueous setting, (2) diagenetic dewatering, volume-
1001 loss and fracturing, (3) partial erosion to form paleotopography, (4) mineralization of fractures
1002 and capping of partially-eroded sulfates with the “smooth capping surface.” The extended history
1003 of these deposits then includes later (5) paleovalley incision, (6) capping and embayment by Hes-
1004 perian Syrtis Major lavas, (7) differential erosion of sulfates and lavas to uncover adjacent deep
1005 basins, and (8) the construction of small fluvial-deltaic and lacustrine sedimentary deposits in
1006 these basins.

1007 There is no evidence of either local topographic confinement of the sulfates or the existence
1008 of higher-elevation sulfates on the adjacent *Nili Fossae Plains*. Their meter-scale layered character
1009 lacking internal unconformities, fine grained nature, and thinning while onlapping topographic
1010 highs favors detrital sedimentation. If correct, this implies a deepwater setting in a lake system
1011 confined by an Isidis Basin-filling ice sheet or deposition at the margin of northern hemisphere

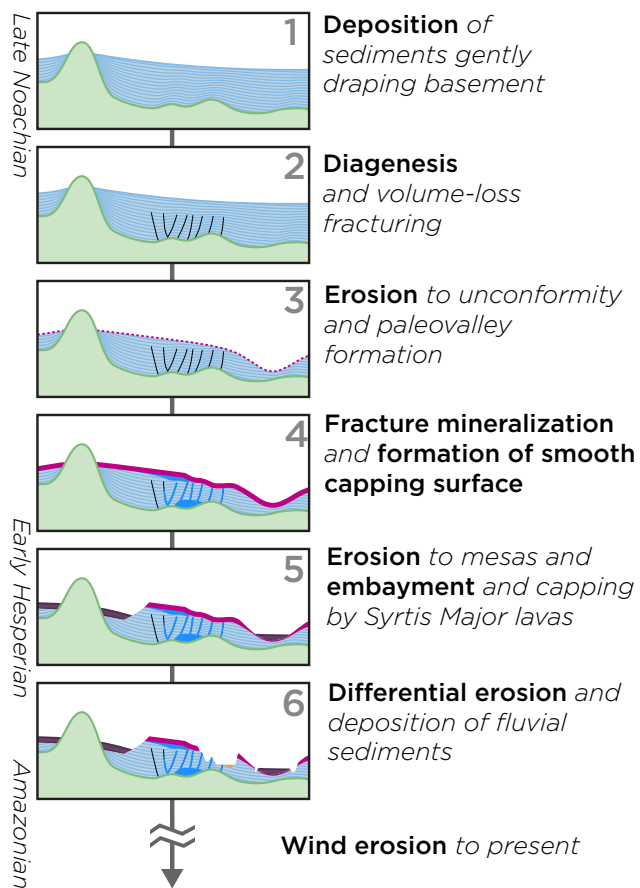


Figure 24: Model emplacement and alteration history of the layered sulfates. The relative timing of steps 3 and 4 is uncertain, and the genesis of the smooth capping surface is unknown.

1012 ocean with a shoreline at a level much deeper than previously proposed. Neither explanation
1013 is fully satisfying, yet the specific characteristics of the deposit and its later fracturing are even
1014 less consistent with other lacustrine or aeolian mechanisms, airfall deposition (e.g. from ash) or
1015 deposition from ices emplaced by obliquity cycling.

1016 Regardless of the depositional setting, volume-loss fracturing involves water-saturated or
1017 very hydrated deposits. Subsequent fracture mineralization requires abundant sulfate-bearing
1018 groundwater over hundreds of meters of depth. Both dewatering and jarosite mineralization oc-
1019 curred during the Noachian-Hesperian transition, well after clay and carbonate formation in the
1020 Noachian highlands. Later extensive erosion of the layered sulfates and superposed fluvial fea-
1021 tures on the Hesperian Syrtis Major lavas demonstrate that surface water was at least episodically
1022 present and shaped the landscape into the Late Hesperian and Early Amazonian. The pattern seen
1023 at NE Syrtis, of aggradation of thick layered deposits during the Noachian–Hesperian transition,
1024 followed by significant erosion and superposed fluviodeltaic deposits from the Late Hesperian–
1025 Early Amazonian, is thus the result of a Martian surface water cycle at least episodically active for
1026 much of the planet’s history.

1027 This study uses the maximum available resolution imagery across a wide area and integrated
1028 new techniques for DEM creation, error analysis and visualization; the level of detail presented
1029 here will be difficult to surpass using orbital data. The key unsolved questions of this study are
1030 most productively assessed at rover scale. Outcrop observation of sedimentary bedforms and
1031 grainsize within the layered sulfates would decisively confirm or refute our deep-basin sedimen-
1032 tation hypothesis. This result will inform future work on other layered sedimentary deposits
1033 globally. If deepwater sedimentation is confirmed, the variation in sedimentary textures within
1034 the sequence could indicate the size of the basin (ice-sheet confined vs. open-ocean). If airfall or
1035 ice-enriched sedimentation is confirmed, the sequence could indicate the timing of volcanic pro-
1036 cesses or climate cycles, respectively. Detailed chemical analysis of the filled volume-loss fractures
1037 can distinguish between a top-down weathering and upwelling groundwater source of fracture-
1038 filling fluids and place firm bounds on the type and scale of groundwater interaction within the
1039 layered sulfates. Sufficient potassium might be present to place K-Ar dates to constrain absolute
1040 timing of alteration by isotopic analysis of the jarosite-filled fractures.

1041 Additionally, detailed examination of relationships between the units described in this study
1042 would substantially clarify the sequence of geologic events affecting NE Syrtis and the surround-
1043 ing region including the Jezero crater watershed. The origin of the smooth capping surface that
1044 unconformably overlies the layered sulfates is unknown, and the contact between the draping
1045 valley fill and the Syrtis Major lavas in *Valley A* may record the interaction of lavas with water-
1046 lain sedimentary deposits. Investigation of both of these features would clarify the timing and
1047 interplay of sulfate erosion, lava embayment, and the late fluvial-lacustrine history of the region
1048 (including a potential Late Hesperian or Early Amazonian inundation). These key features of the

1049 layered sulfates and their context can be evaluated *in situ*, within a 5 km-wide area in nf3, near the
1050 point in the layered sulfates closest to the NE Syrtis Mars 2020 landing ellipse (as of early 2018).
1051 Along the way, the rover would investigate the Hesperian–Amazonian fluviodelatic basin floor
1052 deposits near nf3 and nf5. These deposits are significant because they are at a nearly identical
1053 base level to those found in Jezero crater. Such a 25-30 km traverse with observation and sam-
1054 pling campaign has the potential to greatly illuminate the multistage history of aqueous activity
1055 captured in the upper stratigraphy of northern Syrtis Major and provide new insights into the
1056 Mars surface environment, its climate, and its habitability over a period spanning the Noachian
1057 to Amazonian.

1058 **7 Acknowledgements**

1059 We would like to thank NASA for the Earth and Space Science Fellowship (NNX14AO61H, to D.P.
1060 Quinn) that funded this work, as well as a gift to Caltech by F. Stanback that supported early
1061 efforts in large 3D dataset creation and visualization. We also thank Nicholas Mangold and Paul
1062 Niles for their thoughtful reviews of this manuscript. A supplementary data table containing the
1063 highest-quality layer orientation measurements is available with this publication. HiRISE eleva-
1064 tion datasets created for this study are archived in the Planetary Data System, and the regional
1065 CTX elevation model is archived at CaltechData (DOI: 10.22002/D1.1087). High-resolution, pro-
1066 jected versions of Figure 2 and Figure 3 are archived at CaltechData (DOI: 10.22002/D1.1085),
1067 along with a PostGIS database containing all vector data, including all measured bedding traces,
1068 layer orientation fit data, structural control points, and geologic map data (DOI: 10.22002/D1.1086).
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