

Deep-water cusplate stromatolites of the Cryogenian Trezona Formation

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Deep water cusate stromatolites of the Cryogenian Trezona Formation

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ABSTRACT

Stromatolites and microbialites contain a rich repository of environmental and biological information. Despite extensive research, questions remain regarding the biological, chemical, and physical processes that control stromatolite macro, meso, and microstructure. We report unusual deep water cusate stromatolites from the Cryogenian Trezona Formation, South Australia, from a mixed siliciclastic-carbonate open marine ramp setting. Cusate stromatolite horizons develop near the base of decameter-scale transgression-regression cycles and typically overlie decimetre-scale irregular erosion surfaces. The cusate structure within the stromatolites form near vertical, stacked cusp structures in cross section. In plan view, the cusps form cm-scale sharp parallel ridges oriented predominantly perpendicular to the regional down-slope direction and perpendicular to the elongation direction of stromatolites. We postulate that stromatolites colonised topographic highs of the irregular erosion surfaces (often hardgrounds) and grew in carbonate-supersaturated, iron rich marine waters in low turbidity sediment starved settings. Cusate stromatolites are interpreted as forming during maximum transgression in condensed sections within deep water environments largely below storm wave base. Their microbial

33 metabolism may require low light and low oxygen. The deep water origin for the Trezona
34 Formation cusped stromatolites and other Precambrian cusped stromatolites suggests a link
35 between the cusped morphology and physical/chemical(carbonate-supersaturated, low light, and
36 low oxygen) conditions of Precambrian deep water marine settings.

37

38 INTRODUCTION

39 The Precambrian macroscopic fossil record is dominated by stromatolites that represent
40 diverse and complex microbial communities (Grotzinger and Knoll, 1999). Stromatolites are a
41 major feature of Proterozoic carbonate and mixed carbonate-siliciclastic systems (Grotzinger and
42 James, 2000; Hoffman, 1974; Beukes, 1987) and are important indicators of paleoenvironment.
43 The microstructure of stromatolites has been interpreted as being controlled by either biological
44 (Walter, 1976; Bosak et al., 2012), or chemical processes (Grotzinger and Knoll, 1999; Visscher,
45 Reid, and Bebout, 2000). Similarly, the controls on stromatolite macrostructure have been
46 variably interpreted to be the result of either environmental (e.g., Golubic, 1976; Dill, Shinn, and
47 Kendall 1989; Andres and Reid, 2006), or coupled environmental/biological processes
48 (Ginsburg, 1991; Suosaari et al., 2019). Proposed factors include intrinsic biological controls of
49 the microbial community, siliciclastic input, current energy, underlying substrate, light
50 availability/depth of formation, geochemical conditions/redox gradients, degree of
51 environmental-biological variability, and others (Walter, 1976; Horodyski, 1977a; Awramik and
52 Vanyo, 1986; Ginsburg and Planavsky, 2008; Bosak et al., 2013; Mackey et al., 2017; and many
53 others). Detailed sedimentological and petrographic observations of ancient stromatolites—
54 considering the surrounding lithologies and association with cycles—can be used to shed light on
55 the conditions that influence stromatolite occurrence and morphology.

56 In this study we report deep water cusped stromatolites from the Cryogenian Trezona
57 Formation in South Australia. We use the term ‘stromatolite’ to describe all laminated build-ups
58 formed by sediment trapping, binding, and/or carbonate precipitation of microbes (as defined by
59 Awramik, Margulis, and Barghoorn, 1976). The stromatolites described here form elongate
60 domes with unusual cusped laminations. They are developed in cyclic association with
61 calcareous mudstone, hardgrounds, poorly sorted intraclastic packstone, and marine erosion
62 surfaces in deep water settings. These deep water forms were likely deposited in low energy, low
63 sedimentation, and carbonate supersaturated settings. These stromatolites add to a small

64 collection of deep water cusped stromatolites. The deep water origin for the Trezona Formation
65 stromatolites and other Precambrian cusped stromatolites may suggest a link between
66 Precambrian deep water settings (low light, low oxygen, carbonate supersaturated) and the
67 cusped morphology. There is an absence of similar cusped stromatolites from the rock record
68 after the Cryogenian with no known exact modern analogues of these forms. This may suggest
69 the combined physical, biological, and geochemical conditions needed for deep water cusped
70 stromatolite growth did not exist after the Cryogenian.

71

72 **GEOLOGICAL BACKGROUND**

73 The Cryogenian Trezona Formation is part of the Upalina Subgroup of the Umberatana
74 Group, central Flinders Ranges of South Australia (Preiss, 2000; Fig. 1; Adelaide Fold Belt). The
75 Upalina Subgroup includes the upper nonglacial stratigraphy (late Cryogenian) between the
76 Sturtian and Marinoan ‘snowball Earth’ glaciations (Fig. 2; Preiss, 1998). In the central Flinders
77 Ranges, the Enorama Shale, Trezona, and Yaltipena formations (nonglacial strata, Fig. 2)
78 underlie the Elatina Formation (Marinoan glacial sediment, Fig. 2) and thus record the “lead up”
79 to the Marinoan glaciation. The Umberatana Group has few age constraints but includes U-Pb
80 zircon dates from a tuffaceous horizon in the uppermost Sturtian diamictite with an age of 663.03
81 ± 0.11 Ma (Cox et al., 2018; Fig. 2), with no upper geochronologic constraints for the nonglacial
82 sediment or Marinoan diamictite. In Tasmania, Australia, the termination of the Marinoan
83 glaciation has been constrained to 636.41 ± 0.45 Ma, consistent with geochronological
84 constraints that suggest the Marinoan glaciation ended ca. ~ 635 Ma globally (Calver et al., 2013;
85 Rooney, Strauss, Brandon, and Macdonald, 2015). In Namibia, the Marinoan glaciation is
86 constrained by the syn-glacial Ghaub Formation of Namibia with an age of 639.29 ± 0.26 Ma
87 (U-Pb ID-TIMS) (Prave, Condon, Hoffmann, Tapster, and Fallick, 2016) and 635.5 ± 1.2 Ma
88 (Hoffmann et al., 2004). In Death Valley, USA, the upper non-glacial Thorndike Submember is
89 estimated at 651.69 ± 0.64 Ma (Nelson et al., 2020). In the Northern Flinders Ranges, the
90 Balcanoona Formation (equivalent to the Etina Formation underlying the Trezona Formation;
91 Fig. 2) is estimated to be ~ 650 Ma (Wallace et al., 2015). Therefore, based on available
92 constraints and stratigraphic relationships, deposition of the Enorama Shale, Trezona, and
93 Yaltipena formations is estimated to range in age from ~ 650 to 640 Ma.

94 The Trezona Formation has been the topic of considerable study, as it records one of the
95 largest $\delta^{13}\text{C}$ excursions in Earth history, the ‘Trezona Anomaly’ (e.g., McKirdy et al., 2001;
96 Ahm et al., 2021). Despite the importance of the Trezona Formation in understanding oceanic
97 dissolved inorganic carbon leading up to ‘snowball Earth’, the sedimentology of the unit is not
98 well understood. Although some authors have suggested a restricted or lacustrine origin for the
99 Trezona Formation (Klaebe and Kennedy, 2019), the Trezona Formation is more commonly
100 interpreted as marine—either a shallow low-energy lagoon (Preiss, 1987; Lemon, 1988) or open
101 marine shelf to slope deposition (Rose et al., 2012).

102

103

104

105

106 **METHODS and TERMINOLOGY**

107 This study is focused on four locations in the central Flinders Ranges: Bulls Gap,
108 Angorichina Hostel, Angorichina Station, and Glass Gorge Road (Fig. 1). Measured sections
109 were taken ~5–20 km apart arranged proximal to distal, with one lateral/oblique (Figs. 1, 3).
110 Stratigraphic sections were measured using a Jacob’s staff at the centimetre and meter scale,
111 noting bed thickness, lithology, sedimentary structures, and stromatolite morphology. Samples
112 were collected within measured sections and in laterally equivalent beds for thin section analysis
113 using transmitted light microscopy.

114 The terminology of Burchette and Wright (1992) is used to describe this mixed
115 siliciclastic-carbonate ramp, whereby a ramp is defined as having a gentle slope which extends
116 from the shoreline to the basin and is not rimmed nor is there a sharp distinct break in slope. The
117 environments of a ramp setting are divided into inner, mid, and outer ramp environments. Ramp
118 environments reflect deposition above fair weather wave base, between fair weather wave base
119 and storm wave base, and below storm wave base, respectively. Cycle refers to a bounded
120 package of repeated lithofacies with no implications for lateral continuity or a driving
121 mechanism. Transgression, regression, and relative sea level change also do not suppose a
122 driving mechanism (e.g., subsidence, eustasy, sediment supply).

123 Because of possible confusion/complications in describing mixed siliciclastic-carbonate
124 lithologies, the terminology used in this paper is as follows: mudstone refers to a mix of silt and

125 clay with possible minor carbonate; calcareous mudstone refers to a mix of silt, clay, and
126 carbonate. Intraclast refers to a clast of calcareous mudstone or micrite from the same basin
127 (either eroded and redeposited in situ or eroded upslope and transported/deposited downslope);
128 poorly-sorted intraclastic packstone refers to grain supported platy intraclasts with a fine-grained
129 matrix (either micrite, mudstone, or calcareous mudstone, similar to “intraclastic breccia”); and
130 rounded intraclastic packstone-grainstone refers to generally well-sorted grain-supported
131 spherical and rod-like intraclasts with some fine-grained matrix (packstone) or no matrix
132 (grainstone) present. The terms dolomitised mudstone/hardground refers to authigenic
133 precipitation of dolomite on the seafloor (see Wallace et al., 2019); irregular marine erosion
134 surface refers to a bedding surface with irregular erosional topography (up to ± 1 m of relief over
135 a few meters laterally along the bed). The terms ooid and giant ooid (instead of pisoid) refer to
136 coated grains lacking obvious microbial textures.

137

138 **STRATIGRAPHY and SEDIMENTOLOGY**

139 The Trezona Formation conformably overlies the Enorama Shale (Preiss et al., 1987).
140 The two units together make up a thick and laterally extensive shallowing upward succession
141 (Rose et al., 2012) from deep water basinal mudstone, deep to shallow mixed siliciclastic-
142 carbonate ramp facies, to shallow water carbonate ramp facies. The type section of the Trezona
143 Formation is located at Enorama Creek and totals 240 m in thickness (Dalgarno and Johnson,
144 1964). Lemon (1988) noted that the lower and upper stratigraphy of the Trezona Formation is
145 distinctly different in terms of lithology. Therefore, the Trezona Formation is informally divided
146 into two members: a (lower and upper) siliciclastic member and a carbonate member (Table 1).
147 The lower siliciclastic member consists of mudstone, interbedded mudstone and siltstone to very
148 fine-grained sandstone, calcareous mudstone, intraclastic packstone, elongate cusped
149 stromatolites, erosion surfaces, and hardgrounds (these lithologies are the main focus of this
150 study; Table 1). These lithologies transition upwards into the upper siliciclastic member. The
151 upper siliciclastic member consists of mudstone to v.f.g sandstone, wavy, flaser, and lenticular
152 bedded mixed lithologies, domal stromatolites, rare cusped stromatolites, and crossbedded
153 rounded intraclastic packstone-grainstone (Table 1). The siliciclastic member is in turn overlain
154 by the carbonate member. The carbonate member consists of calcareous mudstone, micrite, ooid
155 grainstone, rounded intraclastic grainstone, domal-columnar and longitudinal stromatolite

156 horizons, very rare cusped stromatolites, and stromatolite bioherms (Table 1). The Trezona
157 Formation is either overlain by the muddy tidal Yaltipena Formation (only present at the Bulls
158 Gap locality in this study, Fig. 3) or unconformably overlain by the Marinoan glacial-associated
159 sediment of the Elatina Formation (Angorichina Hostel, Glass Gorge, and Angorichina Station,
160 Fig. 3; Priess, 1987).

161

162 **Enorama Shale**

163

164 The Enorama Shale is a poorly outcropping unit of green and red mudstone and siltstone
165 at the base of the studied succession (Fig. 4; Table 1). Mudstone is finely laminated and can have
166 a carbonate content of 12–30% (Priess, 1987). These fine-grained lithologies form the valley
167 units between the major carbonate units of the nonglacial Cryogenian sediments of the central
168 Flinders Ranges (Figs. 1, 2). In the Blinman Region, the thickness of the Enorama Shale varies
169 from 150–450 m (Priess 1987). The Enorama Shale conformably underlies the Trezona
170 Formation (Figs. 2, 3, 4A; Priess, 1987). The basal third of the unit is dominated by red and red-
171 brown siltstone, the middle third of the unit is dominated by laminated green mudstone with no
172 coarse-grained beds, and the upper third of the unit is dominated by red siltstone and mudstone
173 (Lemon, 1988). Red mudstone gradationally transitions into the overlying Trezona Formation.
174 The Enorama Shale is documented in this study in the Angorichina Hostel, Angorichina Station,
175 and Glass Gorge Road sections (Fig. 3).

176 In the Enorama Shale, silty sand and current ripple laminations in thin beds (10–30 cm)
177 are occasionally present (Figs. 3,4). Some beds are sharp based with parallel lamination overlain
178 by small-scale hummocky bedding (Fig. 4B; Lemon, 1988). At Angorichina Hostel and Glass
179 Gorge, red siltstone and mudstone with sole marks are present and flutes indicate paleocurrent
180 flow N–NW (Figs. 3, 4C). Other distinctive features include sharp-based or scoured beds with
181 climbing ripples (Fig. 4D), and BCE Bouma divisions (Lemon, 1988). One bed with wave ripple
182 cross lamination was noted north of the Glass Gorge section locality. Wave ripple features were
183 not noted in any other localities. An unusual metre-scale red sand lense was noted at Angorichina
184 Station.

185

186 **Trezona Formation**

187 The contact between the Enorama Shale and the Trezona Formation is placed at the first
188 carbonate rich, poorly sorted intraclastic packstone and stromatolitic bed in the succession (Fig.
189 3). Measured sections of the Trezona Formation at Bulls Gap, Angorichina Hostel, Glass Gorge,
190 and Angorichina Station are 90 to ~260 m thick (Fig. 3). The Trezona Formation is informally
191 divided into two members in ascending order: a siliciclastic member (lower and upper) and a
192 carbonate member (Table 1), described below.

193

194 **Siliciclastic member**

195 Lower siliciclastic member

196 The lower siliciclastic member is a mixed unit of fine-grained siliciclastic and carbonate
197 sediment with a dominant siliciclastic component (Fig. 5; Table 1). Facies are non-cyclically and
198 cyclically interbedded. Cycles in this unit are particularly well developed at the Glass Gorge
199 locality (idealized in Fig. 6). At the base of each cycle is an erosion surface in calcareous
200 mudstone with local topographic high and low points (Figs 5A,D, 6). This erosion surface has up
201 to ± 1 metre of topographic variation in calcareous mudstone over a few meters laterally (Figs 5-
202 A,D, 6). Topography on the erosion surface is difficult to measure with 2D exposures but
203 appears to roughly parallel the paleocurrent direction and regional slope direction N–NW (Fig.
204 3). The erosion surface is overgrown by elongate cusped stromatolites and rare columnar
205 stromatolites (Fig. 5A, D), laterally associated with poorly-sorted intraclast packstone (Figs 5-
206 A,D, 6). Stromatolites occupy the topographic highs of the marine erosion surface, while
207 intraclastic packstone lithologies are generally found in the topographic lows of the marine
208 erosion surface (Fig. 5D, 6). Stromatolitic laminae are finely cusped (Fig. 5B, 6) and become
209 progressively more clay-rich toward the uppermost layers.

210

211 The stromatolite beds are in turn overlain by recessively weathered gray-green to red-brown clay
212 and mudstone (Fig. 5A, 6). Mudstone is generally finely laminated at the base, becoming
213 progressively more resistant with interbedded silty and current rippled fine-grained sand beds
214 upwards (Figs. 3, 6). Within the mudstone intervals, flute casts are preserved on the lower bed
215 surfaces of coarser silty to very fine-grained sand beds, with paleocurrents generally oriented N-
216 NW (regional down-slope directed). Current ripples in coarser beds similarly indicate dominant
217 paleoflow N-NW. Purely siliciclastic lithologies are then overlain by interbedded calcareous

218 mudstone (often dolomitized or containing dolomitic nodules) and poorly-sorted intraclastic
219 packstone (where the top few cm of the bed is often dolomitized). Intraclastic packstone beds are
220 typically 10 to 40 cm thick with an erosional base (Figs. 5C, D, 6). Intraclasts are typically 5
221 mm–25 mm in length, with a mixture of iron oxide and carbonate lithologies. Angular elongate-
222 platy intraclasts are generally poorly sorted with chaotic arrangement of clasts (Fig. 5C). In rare
223 intraclast beds (generally not associated with stromatolites), imbricated and edgewise fabrics can
224 be capped by hummocky cross stratified fine-grained carbonate grainstone. Interbedded, poorly-
225 sorted elongate-platy intraclast packstone and calcareous mudstone beds are capped by an
226 erosion surface that starts the next cycle. The erosion surface and underlying ~10 cm of
227 stratigraphy may be dolomitized (Figs 5A, 6).

228

229 Upper siliciclastic member

230 The upper siliciclastic member is a mixed unit of fine-grained siliciclastic and carbonate
231 sediment with a dominant siliciclastic component (Fig. 7; Table 1). The base of the upper
232 siliciclastic member is arbitrarily placed where the following lithologies become common:
233 crossbedded fine-grained grainstone; crossbedded well-sorted intraclastic packstone-grainstone;
234 wavy, flaser, and lenticular bedded mixed lithologies; siltstone–fine-grained sandstone; and
235 columnar and domal stromatolites without cusped structures (Figs. 3, 7). Erosion surfaces and
236 dolomitized beds rarely occur. Ooids are present, but rare. Well-sorted intraclastic packstone-
237 grainstone beds are dominated by rounded intraclasts (40–95% rounded grains) (Fig. 7A).
238 Rounded platy, rod-like, and spherical intraclastic equidimensional grains dominate, but there are
239 also many complex and irregular shaped grains with angular irregular, internal cavities (Fig. 7A).
240 Spherical grains are commonly <1 mm to 1 cm in diameter. Platy and rod-like grains are
241 commonly <1mm to 3 cm. Outsized irregular intraclasts are up to 7 cm long. Intraclast beds tend
242 to be thick (40 cm to 2 m) and commonly contain cross-lamination (several cm to ~30 cm thick)
243 (Figs. 3, 7B). Most intraclasts consist of a mixture of iron oxides, micrite, and calcite (Fig. 7A).
244 Micrite or blue-white marine cements commonly occupy the inter-intraclast space. Intraclastic
245 packstone-grainstone and fine-grained grainstone can have mud drapes (Fig. 7C). Channelization
246 and bidirectional currents in intraclast beds are common (Fig. 3). Wavy, flaser, and lenticular
247 bedding are common in some intervals (Fig. 7E).

248 Cyclicality is also present in this unit (Fig. 7F). Domal (Fig. 7D) and rarer cusplate
249 stromatolites directly overlie many lithology types: mudstone, erosional surfaces in mudstone
250 (no hardground), crossbedded intraclast packstone-grainstone, and non-continuous erosionally-
251 based crossbedded ooid packstone-grainstone (Fig. 3). Columnar stromatolites can also occur on
252 the margins of crossbedded, channelised intraclast packstone-grainstone beds. Where present,
253 cusplate stromatolites are poorly developed when compared with the underlying lower
254 siliciclastic member. Similar to the underlying lower siliciclastic unit, stromatolites are overlain
255 by clay and mudstone.

256

257 **Carbonate member**

258 The carbonate member is a mixed unit of siliciclastic and carbonate sediment dominated
259 by carbonate lithologies (Fig. 8; Table 1). The base of the carbonate member is arbitrarily placed
260 where carbonate strata become more prevalent, and mudstone units (found in the siliciclastic
261 member) transition to more calcareous mudstone (up to 32% carbonate, Preiss, 1987) (Figs. 3,
262 8A; Table 1). Compared with the siliciclastic member, erosion surfaces are absent or less
263 pronounced, there are only rare instances of dolomitized surfaces. Cusplate stromatolites are rare
264 (see few occurrences at Angorichina Station, Fig. 3). At the base of the carbonate member, red-
265 gray calcareous mudstone is interbedded with low relief stromatolites, longitudinal stromatolites
266 (oriented N-NW), domal-columnar stromatolites, and intraclastic grainstone (Fig. 3; Fig. 8A-C).
267 At the top of the carbonate member, this member becomes increasingly more massive and
268 carbonate dominated (Figs 3, 8A). Calcareous mudstone can be desiccation-cracked (Fig. 8D).
269 Thick, resistant massive blue-brown limestone beds of domal and columnar stromatolite
270 bioherms can be up to 6 m thick (Fig. 8A). Oolitic grainstone is common (Fig 8 E,F) and is often
271 associated with stromatolite bioherms (Fig. 3). Micritic ooids can have iron-oxide rich laminae.
272 Ooids can be small (<1mm) to very large (giant ooids up to 2 cm, Fig. 8F). The uppermost
273 resistant and thickest carbonate intervals of the Trezona Formation are ~28–60 m in thickness
274 (Bulls Gap, Angorichina Station: Fig. 3, Fig. 8A).

275

276

277 **Cusplate stromatolite occurrence, morphology, and petrology**

278

279 Cusate stromatolite horizons are commonly developed in cycles of the siliciclastic
280 member (Fig. 6). Biostromes of cusate stromatolites in such cycles are laterally continuous and
281 can be traced for over 5 km. The stromatolite biostromes likely extend further, but the basin is
282 somewhat faulted with many offset beds and faulted intervals (Fig. 1). Elongate cusate
283 stromatolites are found in all studied localities (Fig. 3). Cusate stromatolites are best developed
284 at the base of the Trezona Formation in the lower siliciclastic member. Some cusate intervals
285 are present in the upper siliciclastic member (Glass Gorge Road, Bulls Gap) and lower part of
286 the carbonate member (Angorichina Station, Fig. 3).

287 The cusate stromatolites typically grow on an irregular erosion surface and hardground.
288 In almost all cases, irregular erosion surfaces develop over calcareous mudstone or mudstone
289 (Fig. 5A, 9A,B) and are associated with intraclast-rich lithologies. Along erosional horizons,
290 stromatolites develop on areas of high topography and intraclastic lithologies are found in
291 depressions on the erosional surface (Figs. 5D, 6). Lenses of intraclastic debris are occasionally
292 found in stromatolite depressions (gutters) along the biostrome. The erosion surface is commonly
293 dolomitized, and multiple erosion and dolomitization intervals are sometimes noted. The contact
294 between the stromatolite and underlying substrate is sharp. Elongate cusate stromatolites
295 (described in detail, below) grade upwards into green or red-brown claystone and then mudstone
296 (Figs. 5A; 9 A,B). This upper contact is transitional: the cusate stromatolite laminae become
297 progressively more clay-rich before stromatolites are overlain by clay and mudstone (Fig. 6)

298 At the macroscale, the stromatolite biostromes are ~tabular to irregular. The lower
299 surface is controlled by the irregular erosion surface and the upper surface is flat lying to gently
300 domed and continuous (Fig. 5A). The biostromes crop out conspicuously and are irregular along
301 strike (generally 0.5–1 m thick, Fig. 9). Cusate stromatolites can be gently domed at the base
302 and be associated laterally with columnar stromatolites (where present) at the base of the
303 biostrome (Fig. 6). Gently domed elongate cusate stromatolites and rarer columnar
304 stromatolites can transition upwards to continuous flatter-lying stratiform cusate stromatolites.
305 Stromatolite growth is generally directed upwards and perpendicular to the underlying substrate.
306 In some cases, the associated columnar stromatolites grow in an inclined manner departing from
307 vertical growth.

308 Stromatolites are elongate at all localities (>0.60 m long, ~0.10 to 1 meter thick, ~0.10 to
309 ~2 m wide; Fig. 9A–C,G). In some localities, closely spaced cusate stromatolites consist of only

310 one downslope elongate cusped stromatolite over an erosion surface. These types are best
311 developed at Bulls Gap, where cusped stromatolites have irregular domed macrostructure (~4 m
312 long, 50 cm thick, ~1m wide), with their long axis preferentially oriented NNW-SSE (SSE-
313 oriented tail and a bulbous head- oriented NNW with and an average strike around 339° (n=11),
314 Fig. 9G). Here, the morphology of the stromatolite domal macrostructure conforms to the
315 underlying irregular erosion surface.

316 At the mesoscale, stromatolites are gently convex to stratiform structures with stacked,
317 mm-scale cusped laminae. Moving upwards, laminae appear to flatten as clay increases in
318 stromatolite laminae. In cross section view, cusp peaks are vertically aligned, or slightly drift
319 asymmetrically north (downslope; Fig. 5A and Fig. 9 E,F). In plan view, cusped stromatolite
320 peaks form straight or slightly sinuous cusped ridges oriented perpendicular to the macroscale
321 elongation direction of domes (see also Preiss, 1987) (Fig. 9 CD,G).

322 At the microscale, three main types of laminae are present in thin section: 1) dark micritic
323 laminae forming cusped ridges; 2) light micritic laminae with upward branching shrubby
324 microbial textures; and 3) stylolitized red iron-oxide rich laminae (Fig. 10 A–D). There are
325 distinct boundaries between all types of laminae. Dark micritic cusped laminae and light micritic
326 laminae form couplets (0.1–2 mm thick) (Fig. 10 A–B). Stylolitized dark red cusped laminae are
327 generally <0.2 mm in thickness and largely follow and accentuate the original cusped
328 lamination. Some stylolites also crosscut the original sedimentary lamination.

329 Dark micritic laminae form cusped laminations and are made of homogeneous micrite,
330 with minor siliciclastic and disseminated iron-oxide material (Fig. 10B). Upward tufted cusped
331 ridges can have a relief greater than 0.5 mm and are laterally linked by thinner, concave-up
332 laminae (Fig. 10 A–C). Relief between the lowest and highest point along a single lamination is
333 generally <1mm. Light micritic laminae with irregular upward branching shrubby microbial
334 textures (Fig. 10D) appear to thicken between and drape over the dark micritic cusped laminae
335 to fill available space (Fig. 10B,D). Shrubby microbial types do not appear to construct the
336 cusps. Microspar fills the intergranular space around irregular upward branching microbial
337 textures (Fig. 10D). The cusped laminae have a concave-upward profile on each side of the
338 structure (Fig. 10B). Cusped structures have low synoptic relief $W \gg H$ (Fig 10. A–C). The
339 wavelength of the cusps generally ranges from around 6.5 mm to 8 mm. Cusped features may be

340 linked vertically or can be disrupted by interstitial carbonates or stylolitized red iron-oxide rich
341 laminae with no upward continuation. In gently convex cusped stromatolites, dark micritic
342 laminae are less cusped and thin moving away from the top of the gently domed surface. Fine-
343 grained iron oxides are disseminated in both the light micritic laminae and homogeneous dark
344 micritic laminae (Fig. 10). Stylolitized red iron-oxide rich laminae concentrate very fine-grained
345 iron oxides and clay. Fenestrae are not present. Voids are uncommon. Detrital grains like quartz
346 are uncommon near the cusp ridges and become more common moving away from cusped
347 structures near the edge of the stromatolite.

348

349 **Process interpretation and depositional setting of the Enorama and Trezona formations**

350 The Enorama Shale is interpreted to represent deposition in a quiet, deep water
351 environment (Preiss, 1987). The Enorama Shale shallows upward into outer ramp, middle ramp,
352 and upper ramp mixed carbonate-siliciclastic environments of the Trezona Formation.

353 In the Enorama Shale, there are no mudcracks or indications of subaerial exposure.
354 Siltstone beds with flute casts (Fig. 4C) and current ripple cross lamination are present but
355 uncommon. These lithologies are likely deposited by dilute, low density turbidity currents, which
356 suggests a distal location from sediment input source. Minor sharp-based parallel sandstone beds
357 overlain by hummocky rippled beds are likely formed via oscillatory and combined flow during
358 storm events (Dumas, et al., 2006).

359 The lower siliciclastic member of the Trezona Formation (Table 1) has thick
360 accumulations of mudstone, silty beds with flute casts, calcareous mudstone (with dolomitic
361 nodules), poorly-sorted platy-elongate intraclastic packstone, and irregular erosion surfaces.
362 Thick accumulations of clay and mudstone are interpreted to represent deposition in a quiet, low
363 energy deep water environment (similar to the underlying Enorama Shale). Thin siltstone beds
364 with flute marks are interpreted as low density turbidites. Thin beds of erosively-based elongate-
365 platy intraclast packstone likely represent debris flows deposited below fair weather wave base
366 (Fig. 3). The platy-elongate and angular nature, chaotic arrangement of clasts, and poor sorting
367 of the intraclasts indicate they were not subjected to extensive reworking, consistent with a mass
368 flow origin. Rare occurrences of intraclasts with both imbricated and edgewise fabrics overlain
369 by hummocky rippled very fine-grained grainstone—thought to require oscillatory and combined
370 flow—may indicate some intraclast beds are storm event beds or debris flows reworked by

371 storms (Myrow et al., 2004). Elongate-platy intraclasts were likely sourced from a carbonate-rich
372 environment or marine cemented beds up dip and were redeposited down dip.

373 Laterally continuous irregular erosion surfaces present below stromatolite layers (Fig.
374 5D), with as much as 1m of relief at an outcrop scale, are interpreted as deep water erosion
375 surfaces. These may result from erosion during sediment starved conditions by deep water
376 bottom currents or debris flow events on the ramp. Similar scale longitudinal bedforms and small
377 furrows have been documented from deep water off the Bahama Outer Bank as a result of
378 erosion by strong bottom currents (Hollister et al., 1974). Debris flows or strong bottom currents
379 in deep water eroded down into partially lithified sediment to create the highly irregular
380 topography. Topographic lows of the marine erosion surface are filled with intraclastic debris
381 during mass flow events, or later infilled by downslope transport of intraclastic debris (erosion
382 via filling of low areas may enhance topographic relief).

383 Dolomitized irregular erosion surfaces, dolomitized tops of some intraclastic packstone
384 beds, and dolomitized mudstone beds are interpreted as hardgrounds associated with sediment
385 starvation and authigenic dolomite precipitation, similar to other deep water hiatal surfaces in the
386 Flinders Ranges (Wallace et al, 2019). Hardgrounds (hiatal surfaces) require long exposure to the
387 seafloor under high carbonate saturation in sediment starved intervals, possibly linked with
388 anoxia in Cryogenian deep marine settings (Wallace et al., 2019). Hardgrounds and associated
389 cusped stromatolites are interpreted to record deposition in deep water, low energy conditions
390 completely below fair weather and largely below storm wave base in an outer to mid ramp
391 setting (e.g., Burchette and Wright, 1992). Actual constraints on depth are difficult to estimate
392 because there are many complexities in neritic and coastal settings, and wave base is a function
393 of many factors such as wave height, fetch and period, sediment size, storm intensity (e.g.,
394 Immenhauser, 2009; Plint, James, and Dalrymple, 2010). Despite uncertainty, in modern shallow
395 siliciclastic seas average fair weather wave base is $\sim 10 \pm 5$ m and storm wave base is $\sim 20 \pm 5$ m
396 depth (Walker and Plint, 1992). We therefore place deposition of the lower siliciclastic member
397 in at least ~ 10 s to a couple 10s of meters depth. Such depths may be in the photic zone but in low
398 light. For example, at 20 m depth in clear coastal water there is $\sim 3\%$ of surface irradiance (350–
399 700 nm; Jerlov, 1976).

400 The upper siliciclastic member (Table 1) has laminated mudstone, domal-columnar
401 stromatolites, heterolithic bedding, rounded intraclast packstone-grainstone, and fine-grained

402 grainstone. Spherical and rod-like rounded intraclast packstone-grainstone and fine-grained
403 grainstone (Fig. 6 A,B,C) have large-scale crossbedding (Fig. 3). Less commonly, these
404 lithologies have bidirectional crossbedding with mud drapes (Fig. 6C). Wavy, flaser and
405 lenticular bedded mudstone-sandstone and limestone have variable paleocurrent indicators (no
406 mudcracks) (Fig. 3; 6E). These higher energy lithologies suggest parts of the upper siliciclastic
407 member were exposed to tidal, wave, and current reworking above fair weather wave base in
408 subtidal conditions. Cycles with low and high energy lithologies are suggestive of deposition in a
409 mid to inner mixed siliciclastic-carbonate ramp setting.

410 The carbonate member (Table 1) has calcareous mudstone, crossbedded ooids,
411 grainstone, crossbedded rounded well-sorted intraclast grainstone, and massive domal-columnar
412 stromatolite bioherms. The base of the carbonate member—with interbedded calcareous
413 mudstone and a variety of stromatolite types—is similar to the upper siliciclastic member but
414 more carbonate-rich (Fig. 8A). Calcareous mudstone and stromatolite lithologies may reflect low
415 energy deposition in a carbonate rich mid to inner ramp setting. Moving upwards, the occurrence
416 of ooid grainstone suggests continuous agitation of sediment in shallow, high energy conditions.
417 Desiccation cracked calcareous mudstone beds are tidal flats (Fig. 8D). The carbonate member
418 was influenced by tides and shallowed to intertidal conditions in rare instances (similar lithology
419 to the overlying tidal Yaltipena Formation; Lemon, 1988). Massive crossbedded ooids,
420 crossbedded rounded well-sorted intraclast grainstone, and massive domal-columnar stromatolite
421 bioherms may represent deposition in the inner ramp setting, representing the culmination of the
422 Enorama-Trezona shallowing upwards succession.

423 Palaeogeographically, the Bulls Gap section (south) is likely located more proximal to
424 the shoreline than the Angorichina Hostel and Glass Gorge localities (north). This interpretation
425 is based on the relative abundance of wave and current ripple cross lamination, more variable
426 paleocurrents, more common and thicker crossbedded intraclastic grainstone, and coarser
427 lithologies (Fig. 3; Bulls Gap). Evidence of deepening towards the north is consistent with the
428 basin geometry suggested in previous studies (Preiss, 1987; Rose et al., 2012).

429 Our interpretation of a mixed siliciclastic-carbonate ramp for the Trezona Formation
430 differs from earlier interpretations of a lagoonal-shallow marine system (Preiss et al., 1987;
431 Lemon, 1988) or an isolated lacustrine basin (Klaebe and Kennedy, 2019). The restricted basin-
432 lacustrine interpretation of Klaebe and Kennedy (2019) is not supported by our data, as evidence

433 for tidal currents during deposition of the Trezona Formation indicates a direct marine
434 connection during deposition (Sztanó, 1995). Preiss et al. (1987) and Lemon (1988) noted the
435 low-energy nature of the stromatolite beds and suggested a lagoonal (instead of deep water)
436 setting. No barrier facies were noted in this study, and we suggest a deep water low energy origin
437 for the stromatolites is supported by all available data.

438

439 **Basin redox chemistry**

440 The Trezona Formation has a variety of iron oxide rich lithologies, and they are strongly
441 facies controlled. These include finely disseminated hematite in deep water cusped stromatolite
442 laminae, iron oxide rich intraclasts (up to 17% hematite; Preiss et al., 1987), and iron oxide rich
443 ooids. The ooids have alternating carbonate and iron rich laminae (Fig. 8D). In all iron oxide rich
444 lithologies, iron exists as very fine-grained hematite with no clear evidence for the replacement
445 of ferrous precursor minerals. Such lithologies, and their strong facies/component control,
446 suggest syn-sedimentary iron oxide precipitation within the Trezona Formation rather than later
447 diagenetic iron enrichment. Syn-sedimentary iron oxide rich lithologies may provide clues to
448 basin redox conditions during deposition. Iron oxide rich lithologies are present in all units of the
449 Trezona Formation (from deep water to shallow water), suggesting iron oxidation in both
450 shallow and subtidal conditions. This is consistent with other evidence for ferruginous ocean
451 conditions and fluctuating redox conditions in this Cryogenian basin (Hood and Wallace, 2014;
452 O'Connell et al., 2020) and elsewhere during the Cryogenian (Canfield, Poulton, Narbonne,
453 2007). We tentatively suggest a chemocline largely situated in deep water around the outer ramp
454 setting. Iron oxide rich ooids also suggest incursions of anoxic marine water (delivering
455 dissolved iron) into shallower environments of the basin.

456

457 **Controls on cyclicity**

458 The Trezona Formation is strongly cyclic (Fig. 7F). The repeating packages appear to
459 reflect cyclic fluctuations in sediment delivery and sediment starvation. Such cyclic fluctuations
460 may be allogenic related to subsidence or regression-transgression via eustatic or relative sea
461 level change linked with sediment supply (Goodwin and Anderson, 1985; others). Alternatively,
462 they may be autogenic and related to internal dynamics of the ramp system, or some combination
463 of drivers.

464 Within the Enorama-Trezona shallowing upwards succession, cycles are highly regular
465 (Fig. 7F), laterally extensive, and are often traceable over long distances (>5km along strike).
466 Therefore, allogenic regression-transgression and associated changes in sediment supply likely
467 play a role in deposition of cyclic packages. However, there can also be variability in
468 stratigraphic thickness and lithology between cycles (Fig. 3), perhaps suggestive of a
469 combination of drivers.

470 In the lower siliciclastic member, data suggest the erosion surface-hardground-
471 stromatolite assemblage represents a condensed interval of sedimentation during maximum
472 transgression of each transgression-regression cycle (Fig. 6). There are no shallow water
473 lithologies or sedimentary structures associated with cusped stromatolites. Association of
474 cusped stromatolites with poorly-sorted platy intraclastic packstone, marine erosion surfaces,
475 and numerous hardgrounds (Fig. 6) is suggestive of sediment-starved conditions, elevated
476 carbonate saturation, low energy (except for punctuated debris flow events), and prolonged
477 exposure of the seafloor prior to and during stromatolite growth, consistent with a deep water
478 origin. Carbonate saturation in this deep water setting may be linked with anoxia or low oxygen
479 (Wallace et al., 2019). Intraclastic packstone lithologies are found in topographic lows on the
480 erosion surface and occasionally as lenses in depressions (gutters) along the cusped biostrome.
481 This is unusual, as stromatolites associated with intraclasts are not typically associated with
482 subtidal stromatolites (Kah et al., 2009). The association of stromatolites with poorly-sorted
483 intraclastic packstone (large grain size) may seem, at first glance, hard to reconcile with an
484 interpretation of stromatolite growth in deep water during maximum transgression. We suggest
485 enhanced carbonate production during relative sea level rise resulted in increased carbonate
486 sediment production and oversteepening of the ramp, resulting in instability and initiation of
487 debris flows. This relationship has been demonstrated on modern carbonate platforms, where
488 enhanced carbonate production during relative sea level rise increases mass wasting (Droxler and
489 Schlager, 1985). In a somewhat analogous ancient setting, it is postulated that intraclast
490 deposition resulted from widespread seafloor cementation and increased storm intensity during
491 transgression, and thrombolites formed during max transgression in the absence of turbidity
492 (Myrow et al., 2012). Some combination of factors is possible, whereby storms trigger mass
493 flows on an unstable, oversteepened ramp. Following a large debris flow, deep water conditions
494 would be extremely sediment starved with conditions suitable for cusped stromatolite growth.

495 Transgression-regression cycles occur against a large-scale backdrop of relative sea level
496 fall during the Enorama-Trezona shallowing upward succession. What was the driver of relative
497 sea level fluctuations? Syn-sedimentary diapir movement has been suggested as a relative sea
498 level control during deposition of the Trezona Formation (Lemon, 1988; Klæbe and Kennedy,
499 2019). Syn-sedimentary diapir movement may have caused some variability in stratigraphic
500 thickness, relative depth, and lithology between sites. Another driver may be eustatic changes in
501 sea level, linked with sediment supply. Cycles in coeval strata in the Yukon have been
502 interpreted as cyclic deepening and shallowing during the lead up to the Marinoan glaciation
503 (Day et al., 2004). Given the Trezona cycles are highly regular and do closely precede the
504 Marinoan glaciation, we speculate that the cycles may be tied to eustatic fluctuations caused by
505 the buildup and decay of ice sheets in high latitude regions. In this sense, the Trezona cycles may
506 record the beginnings of the global Marinoan glaciation.

507 Broadly, evidence for condensed stromatolite deposition during maximum transgression
508 of cycle may suggest the need for careful consideration and independent evidence for
509 paleoenvironmental conditions when interpreting stromatolites in ancient mixed siliciclastic-
510 carbonate settings. Carbonate-rich lithologies and stromatolites are commonly interpreted as
511 shallow water intervals during cyclic sedimentation. This may not always apply to mixed settings
512 or deep water cycles (Myrow et al., 2012; this study), especially considering carbonate saturation
513 at depth during the Precambrian (Wallace et al., 2019).

514

515 **Factors influencing the occurrence and morphology of cusped stromatolites**

516 The distinctive cusped morphology of the stromatolites of the base of the siliciclastic
517 member (and their elongate macrostructure) is not typical of the stromatolites from the
518 shallower-water carbonate member of the Trezona Formation. The environmental factors relating
519 to stromatolite occurrence and growth are largely interpreted as the result of transgression-
520 regression cycles in deep water (discussed above). Stromatolite growth appears to be linked with
521 sediment starvation and associated elevated carbonate saturation. Other controlling factors
522 include a hard underlying substrate (hardground formation), topography on irregular marine
523 erosion surfaces, and downslope-oriented currents.

524 Stromatolites are here interpreted as growing in sediment starved conditions during
525 maximum transgression. A resumption of sediment supply and increase in clay delivery at the

526 top of stromatolite biostromes eventually leads to their demise (Fig. 6). Sediment flux also
527 appears to influence stromatolites morphology at the mesostructure scale. Overlying the erosion
528 surface, stromatolites form elongate domes with cuscate laminae. Moving upwards, laminae
529 appear to flatten as clay increases in stromatolite laminae. Other studies have similarly noted a
530 flattening of stromatolite laminae corresponding with increased sediment flux (Mackey et al.,
531 2017).

532 Based on the stromatolite morphology, carbonate composition, condensed nature, and
533 association with hardgrounds, another factor that appears to control stromatolite growth and
534 occurrence is elevated carbonate saturation. Although they contain some detrital material (iron
535 oxides and quartz), stromatolites are dominantly composed of micrite with delicate cuscate and
536 upward branching shrubby microbial textures. The micritic composition and preservation of
537 delicate microbial features is most consistent with direct carbonate precipitation by stromatolite-
538 forming microbial communities. Carbonate saturation may be linked with low sediment flux and
539 anoxia in Precambrian deep water settings (Wallace et al., 2019), explaining why cuscate
540 stromatolites are best developed at the base of the section (deepest water). The association of
541 stromatolites overlying hardgrounds may also suggest stromatolites preferred to colonise hard
542 substrates. A hard substrate, however, may not be strictly necessary as there are a few instances
543 of cuscate stromatolite growth on top of poorly cemented mudstone moving up-section.

544 Stromatolites colonised the topographic highs of the erosion surface and are laterally
545 associated with elongate-platy intraclastic packstone in topographic lows. This may indicate
546 stromatolites preferred to grow in less turbulent conditions. Both the underlying erosion surface
547 and downslope currents appear to have a control on the elongation morphology. At Bulls Gap,
548 stromatolite bioherm domes are oriented ~N-NW, roughly parallel to both the downslope slope
549 direction and the predominant paleo-current direction (Fig. 3). Their morphology with a S-SE
550 tail and a bulbous nose N-NW (downslope) (Fig. 9G) is also likely current controlled.

551 Individual cusps can be slightly asymmetrical leaning N-NW, possibly caused by cuscate
552 growth structures drifting downslope under current flow. Straight-sinuuous ridges are oriented
553 perpendicular to stromatolite elongation (in plan view). Because the parallel-slightly sinuous
554 ridges are present on stromatolites (precipitated carbonate instead of detrital sediment), have a
555 high angle of repose, short wavelength, and are linked with individual cusps, they are interpreted

556 as growth structures. However, their parallel-slightly sinuous ridge growth morphology may be
557 somewhat influenced by microorganism interaction with currents.

558 At the microscale, the interaction of at least two different microorganism communities
559 (cusate and upward branching shrubby lamina types) likely control the unusual morphology of
560 deep water cusate stromatolites. Low synoptic relief of cusate ridges may be the result of
561 similar growth rates for cusate and upward branching shrubby lamina types. Shrubby microbes
562 commonly drape over cusate ridges and thicken in the inter-cusp space, filling available
563 accommodation. Cusate microorganisms must continuously re-establish on the stromatolite
564 surface, which may prevent them from gaining high synoptic relief. It is also possible that low
565 synoptic relief of cusps is a biological feature. The gently domed morphology (when not
566 controlled by the underlying erosion surface) appears to reflect thickening and preferential
567 growth of microorganism types in the middle of the dome. Dark micritic laminae are less cusate
568 and thin moving away from the top of the gently domed surface. Therefore, upward branching
569 shrubby microbes do not have an elevated cusate surface to continuously drape over at the edge
570 of the gently domed structure.

571

572 **Comparison with other Precambrian cusate stromatolites**

573 Stromatolites and fenestrate microbialites with a cusate morphology have been
574 documented from several Archean and Proterozoic successions (Fig. 11; Hoffman, 1974;
575 Beukes, 1987; Sumner, 1997; Flannery and Walter, 2012; Bartley et al., 2015; Kurucz and
576 Fralick, 2018). Similar morphologies have been variously referred to as Thyssagetacean-type
577 stromatolites (Vlasov, 1977, in Hofman and Masson, 1994), thesaurus stromatolites (Hofmann
578 and Masson, 1994), peak or peak-shaped stromatolites (Jackson, 1989), or tufted microbialites
579 (Flannery and Walter, 2012). In addition, cusate stromatolites share some similarities with
580 coniform stromatolites and cusate/tented fenestrate microbialites (Sumner 1997).

581 In the Mesoproterozoic Dismal Lakes Group, cusate stromatolites formed offshore in a
582 sediment-starved outer platform setting, with a minimum depth for cusate growth of 12 m, but
583 perhaps up to ~80 meters depth in low oxygen carbonate supersaturated settings below wave
584 base (Bartley et al., 2015). Bartley et al. (2015) interpreted the vertical and horizontal
585 components as a result of the interplay of different microbial communities, potentially driven by
586 light availability or variable redox conditions. Although the environment of deposition and

587 cuscate morphology is grossly similar, strong vertical supports, pyrite laminae, and substantial
588 primary void space are present in the Dismal Lakes stromatolites (Bartley et al., 2015), unlike the
589 Trezona cuscate stromatolites. It's not clear if the Dismal Lakes Group cuscate stromatolites
590 have linked ridges on the top of the structure (Bartley et al., 2015, but see their Fig. 10). In the
591 Gamohaam and Frisco formations of the Campbellrand Subgroup, South Africa, cuscate
592 fenestrate microbialites have also been interpreted as being of deep subtidal origin (Sumner,
593 1997). These fenestrate microbialites are characterized by a netlike framework of fine, dark
594 carbonate laminae, cemented by fibrous calcite, which form stacked true cuscate microstructures
595 (but lack iron oxides and parallel cusp ridges in plan view). Cuscate microbialites of the Pethei
596 Group are interbedded with turbidites and basin-slope facies and are also interpreted as deep
597 water forms (Fig. 11B; Hoffman, 1974). Perhaps most similar to the Trezona Formation
598 stromatolites are the elongated Boetsap-style giant stromatolite mounds with red tufted
599 laminations in the Archean Campbellrand Subgroup (Beukes, 1987; Fig. 11C). Their cuscate
600 morphology is similar, and the downslope elongation of domes is also similar to the Trezona
601 Formation cuscate stromatolites. The Boetsap cuscate structures also appear to have Trezona-
602 style parallel-slightly sinuous ridges (Alexander, 2020). The elongated mounds at Boetsap have
603 been interpreted as subtidal (Truswell and Eriksson, 1973; Beukes, 1987), but it's not known
604 how deep. Similar cuscate stromatolites have also been found in the Neoproterozoic Callison
605 Lake Formation in the Yukon, Canada (Fig. 11D; Strauss et al., 2015). They are associated with
606 low energy lithologies in likely deep water. Callison Lake Formation cuscate stromatolites are
607 not associated with iron oxides.

608 Precambrian cuscate stromatolites share a similar subtidal deep water origin (Hoffman,
609 1974; Sumner, 2007; Bartley et al., 2015). In at least one case, cuscate stromatolites were also
610 interpreted as being formed in carbonate supersaturated, sediment starved, and low oxygen
611 settings (Bartley et al., 2015). Other insights, such as environmental/sedimentological setting,
612 redox setting (low oxygen, ferruginous, euxinic), low light, carbonate supersaturated, etc were
613 not able to be gleaned from existing work on cuscate stromatolites. The deep water origin of the
614 Trezona Formation cuscate stromatolites and most other Precambrian cuscate stromatolites may
615 suggest: 1) cuscate stromatolites may be deep water indicators if supported by the associated
616 lithologies; and 2) there may be a link between the cuscate morphology and the biological,
617 physical, and chemical conditions of Precambrian deep water marine settings.

618

619 **Modern cusplate microbialities**

620 Modern cusplate forms include mineralised pinnacle cusplate and tufted structures found
621 in Antarctic sub-glacial lakes (Parker et al., 1981; Sumner et al., 2016; Mackey et al., 2017),
622 unmineralized forms in sabkha and salt marsh environments of the Laguna Figueroa, Baja
623 California, Mexico (Horodyski and Bloeser, 1977b), and cusplate mat forms in Shark Bay
624 (Awramik and Vanyo, 1986; Flannery and Walter, 2012). Some cusplate features, although
625 documented from different environments, share a continuously submerged, low energy, low
626 turbulence, and low sedimentation setting (Horodyski and Bloeser, 1977; Parker et al., 1981;
627 Mackey et al., 2017). These environmental variables are similar to the Trezona Formation
628 cusplate stromatolites.

629 Despite some similarities, modern cusplate forms are not robust analogues for the deep
630 water cusplate stromatolites of the Trezona Formation. Modern examples tend to be characterized
631 by isolated tufts/cusps/pinnacles (Parker, 1981; Horodyski and Bloeser, 1977; Sumner et al.,
632 2016) or cusps that have a reticulate, net-like geometry (Flannery and Walter, 2012). Ridges in
633 microbial mats can be oriented perpendicular to tidal currents in the Lyngbya mats of Shark Bay
634 (Awramik and Vanyo, 1986; their Fig. 2B). However, their microstructure is not similar, as tufts
635 are non-laminar with grains and patchy micrite in thin section (Jahnert and Collins, 2013). In
636 contrast to modern examples, in the Trezona Formation each laminated cusp (in cross section
637 view) makes up a straight or slightly sinuous ridge (in plan view) that is perpendicular to the
638 downslope elongation direction of the stromatolite and downslope oriented currents (Fig. 9D).
639 To our knowledge, no similar cusplate forms have been documented in the modern. This
640 difference in mat structure potentially means that modern tufted forms may be only partially
641 related or perhaps even completely unrelated to the Trezona Formation cusplate forms. Some
642 similarities in mat structure may not reflect a true homology.

643 To our knowledge, no cusplate stromatolites similar to the cusplate stromatolites of the
644 Trezona Formation have been documented after the Cryogenian. For example, Late Devonian
645 deep water stromatolites have a large variety of morphologies, but no cusplate structures
646 (Playford, Hocking, Cockbain, 2009). It may be possible that cusplate deep water forms are
647 linked to specific biological, chemical, and physical processes (low light, low oxygen, carbonate

648 supersaturated, and/or possibly fluctuating redox conditions) in Precambrian deep water settings
649 that did not persist after the Cryogenian.

650

651 **Possible cusplate stromatolite metabolism**

652 Deciphering the metabolism of ancient stromatolites is problematic, but some constraints
653 can be placed on the cusplate forms described here. It is unlikely that the Trezona cusplate
654 microorganisms had sulphur-based metabolisms, as neither sulphide minerals (such as pyrite) nor
655 their weathered pseudomorphs are present in the cusplate stromatolite laminae. While there is
656 finely disseminated hematite in the stromatolite laminae, it is unlikely that the Trezona cusplate
657 forms were constructed by chemotropic iron-oxidizing microorganisms. The apices of some
658 cusps do show high concentrations of iron oxide (Fig 10A,C). However, iron oxide is not
659 generally concentrated in specific lamina types, as would be expected for microbially-mediated
660 iron precipitation (Planavsky et al., 2009). Finely disseminated hematite is also present
661 throughout the Trezona Formation and is not restricted to the cusplate stromatolites. There are
662 also no iron oxides associated with similar deep water cusplate stromatolites in the Callison Lake
663 Formation (Fig. 11D). Therefore, there is little evidence to support a dominant role for iron-
664 oxidizing microbes in the formation of cusplate stromatolites.

665 Cusplate stromatolites in the Trezona Formation grew in deep water settings, perhaps
666 suggestive of a microbial metabolism that could persist in low light settings or may have even
667 been independent of photosynthesis. The mesoscale structure of Trezona cusplate stromatolites
668 have a gently domed convex morphology. This morphology is commonly linked with a
669 phototrophic growth response, which may be suggestive of low light dependence (e.g., Awramik
670 and Vanyo, 1986; however, see Petryshyn and Corsetti, 2011).

671 Cusplate stromatolites appear to span the rock record from the Archean through to the
672 Cryogenian mid Neoproterozoic. Cusplate stromatolites are not present after the Cryogenian. This
673 may reflect a relationship with anoxia—their disappearance perhaps related to Neoproterozoic
674 oxygenation. Some research indicates that the Ediacaran late Neoproterozoic is characterized by
675 increasing levels of oxygenation (Fike et al., 2006 and many others). Cusplate stromatolites may
676 have required a very specific set of conditions such as a combination of low light (as their deep-
677 water occurrence would suggest) and anoxia. After the Cryogenian, the oxic/anoxic chemocline
678 may have been too deep to support both low light and anoxia. Hence this ecological niche may

679 have been completely removed from the Earth and Trezona-style cusped stromatolites became
680 extinct.

681

682 **CONCLUSIONS**

683 Unusual cusped stromatolites in the Trezona Formation are interpreted as forming during
684 maximum transgression as condensed sections in deep water settings. Stromatolite occurrence
685 and morphology is related to cyclic processes on the marine shelf (transgression-regression and
686 associated sediment supply, debris flows), chemical processes (carbonate supersaturation and
687 low oxygen), hardground formation, sediment starvation/siliciclastic influx, topography on
688 irregular marine erosion surfaces, and downslope oriented currents. Cusped stromatolites growth
689 during maximum transgression in deep water cycles suggests the need for careful evaluation and
690 independent evidence of paleoenvironmental conditions (relatively deep versus shallow) when
691 interpreting mixed siliciclastic-carbonate cyclic sedimentation. The deep water origin for the
692 Trezona Formation cusped stromatolites and other Precambrian cusped stromatolites may
693 suggest: 1) cusped stromatolites are deep water indicators if supported by the associated
694 lithologies; and, 2) there may be a link between the cusped morphology and the biological,
695 physical, and chemical conditions of Precambrian deep marine settings. After the Cryogenian,
696 the chemocline may have been too deep to support both low light and low oxygen cusped
697 forming microbial metabolisms.

698

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700

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706

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969

970 **FIGURE CAPTIONS**

971

972 Figure 1. Geological map of Neoproterozoic strata near Blinman, Central Flinders Ranges, South
973 Australia. Modified from the Parachilna Region geological map (compiled by Reid and Preiss,
974 1999).

975

976 Figure 2. Generalized stratigraphy of the Neoproterozoic Umberatana Group in the central
977 Flinders Ranges, South Australia. Modified from Preiss et al., 1987.

978

979 Figure 3. Measured sections and interpreted correlations (note: faulting does not allow for
980 precise correlation, see Fig. 1). Note relative abundance of carbonate facies in the south at Bulls
981 Gap (left), and increase in mudstone in the northern sections, Angorichina and Glass Gorge
982 (right). Measured section localities (Bulls Gap, Angorichina Hostel, Angorichina Station, and
983 Glass Gorge Road) are shown in Fig. 1.

984

985 Figure 4. Field photos of the Enorama Shale. A. Red mudstone with thin silty beds transitioning
986 upwards into the Trezona Formation. Angorichina Hostel. B. Mudstone interbedded with sharp-
987 based sands with hummocky cross stratification; a potential storm event bed, Angorichina
988 Hostel. C. Flute casts from the underside of silty beds of the Enorama Shale, Glass Gorge.
989 Paleocurrents (when in place) are oriented north. D. Scoured bed with climbing ripples, possibly
990 a sandy turbidite, Angorichina Hostel.

991

992 Figure 5. Field photographs of the lower siliciclastic member, Trezona Formation. A. Mudstone
993 at base transitions upwards to poorly sorted intraclastic packstone, calcareous mudstone,
994 hardgrounds (arrow), an irregular erosion surface, and then a resistant deep water stromatolite
995 horizon. Stromatolites are overlain by mudstone. B. Cross section view of cusped stromatolites,
996 Angorichina Hostel C. Elongate poorly-sorted intraclastic packstone with an erosional base. Note
997 possible hardground at top of bed, Angorichina Hostel. D. Irregular erosion surface overlain by
998 poorly-sorted intraclastic packstone (in topographic lows, arrow a) and cusped stromatolites
999 (topographic highs, arrow c). Note truncation of red mudstone beds by the erosion surface (arrow
1000 b), Angorichina Hostel.

1001

1002 Figure 6. Idealized deep water cycle in the lower siliciclastic member. Transgression, regression,
1003 and relative sea level change do not suppose a driving mechanism (e.g., subsidence, eustasy,
1004 sediment supply).

1005

1006 Figure 7. Field photographs of the upper siliciclastic member, Trezona Formation. A. Iron-rich
1007 rounded and elongate intraclasts, Angorichina Station. B. Bidirectional crossbedding in intraclast
1008 grainstone, Emu Gap. C. Crossbedded fine-grained grainstone with mud-drapes. Note wavy,
1009 flaser, and lenticular bedding at the base of the crossbedded grainstone, Angorichina Hostel D.
1010 Domal stromatolites in plan view, Glass Gorge. D. Wavy, flaser and lenticular bedding,
1011 Angorichina Hostel. E. Regular cyclicity in the siliciclastic member, north of Bulls Gap.

1012

1013 Figure 8. Upper carbonate member of the Trezona Formation. A. Contact between the
1014 siliciclastic member and the carbonate member (arrow). Note change in vegetation at the contact.

1015 Fine-grained carbonate at base of the carbonate member grades up into massive carbonate beds
1016 at the top of the Trezona Formation. B. Rounded intraclastic grainstone bed overlain by
1017 columnar stromatolites. C. Longitudinal stromatolites oriented N-NW, Bulls Gap. D.
1018 Desiccation-cracked red calcareous mudstone, Bulls Gap. E. Blue oolitic grainstone,
1019 Angorichina Station. F. Thin section photomicrograph of iron-rich giant ooids, Bulls Gap. Note
1020 alternating iron-rich laminae (red) and micritic laminae (grey).

1021
1022 Figure 9. Cuspate stromatolites in the Trezona Formation. A–B. Deep-water cuspate stromatolite
1023 horizons overlying an erosion surface, Angorichina Hostel. (north to the right of the images).
1024 Arrow in B points to intraclasts in topographic lows of the erosion surface. C. Macrostructure of
1025 elongate cuspate stromatolite domes in plan view, Emu Gap. Note cusp ridge lines oriented
1026 perpendicular to elongation direction. D. Plan view of cuspate ridges on stromatolite dome
1027 oriented perpendicular to paleo downslope direction. E. Cross section of cuspate stromatolite
1028 laminae with stacked cusp peaks drifting to the north (right). F. Line drawing of cuspate
1029 stromatolites. G. Field sketch and interpretation of cuspate stromatolites at Bulls Gap.

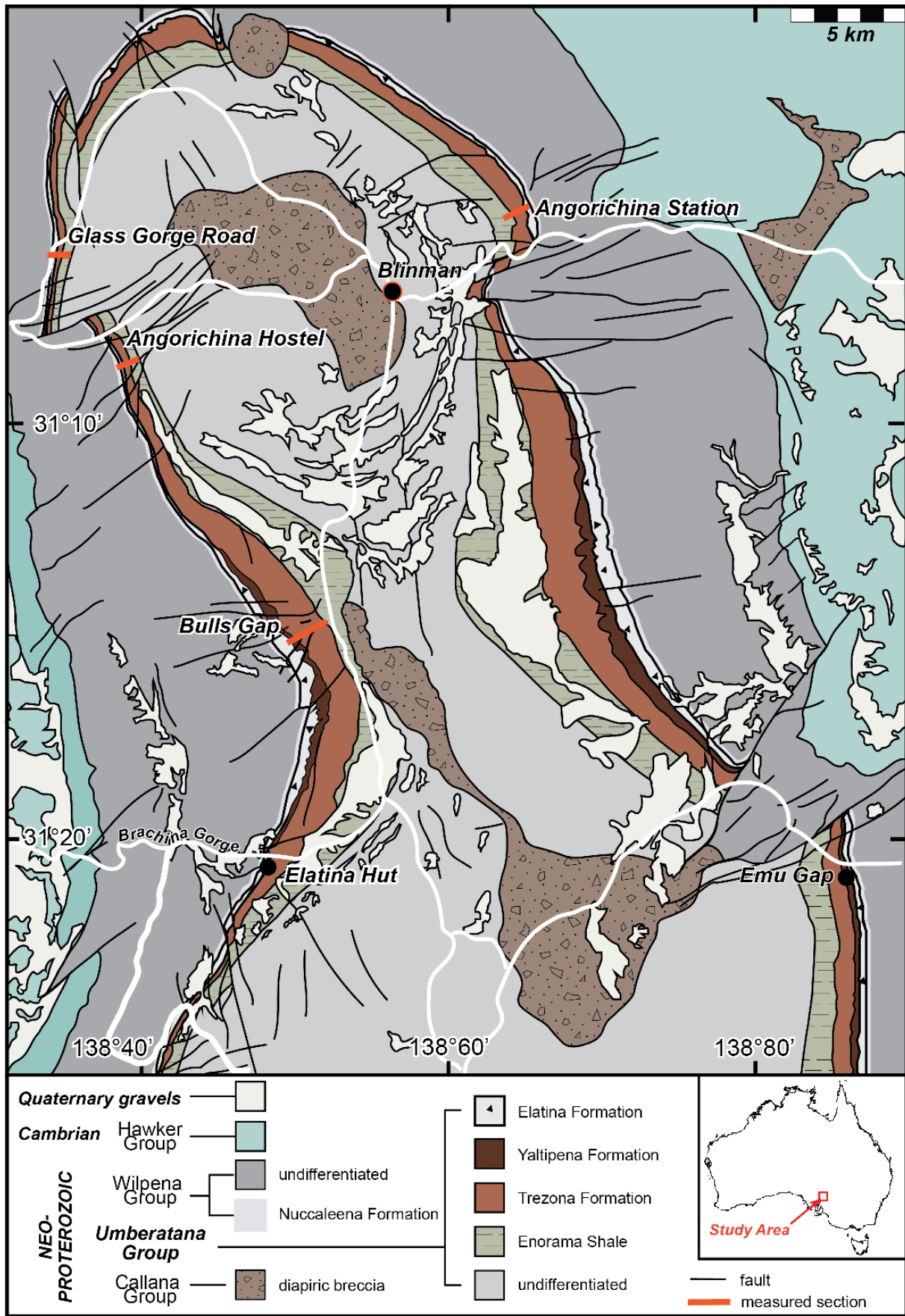
1030
1031 Figure 10. Thin section photo-micrographs of cuspate stromatolite microstructure. A.
1032 Interlaminated dark micritic carbonate, light micritic carbonate, and stylolitized dark red
1033 laminae. B. Dark micritic cuspate lamina (arrow). Note fine grained disseminated hematite. C.
1034 Interlaminated dark micritic and light micritic carbonate, and stylolitized dark red laminae. D.
1035 Upward branching shrubby micritic microbial textures (arrow). Inter-microbial space is filled
1036 with microspar.

1037
1038 Figure 11. A comparison of some Precambrian cuspate stromatolites and microbialites. A.
1039 Cuspate stromatolites from the Cryogenian Umberatana Group, Trezona Formation. B. Cuspate
1040 stromatolites from slope facies of the Paleoproterozoic Pethei Group. C. Cuspate stromatolites
1041 from the Archean Campbellrand Subgroup, Gamohaam Formation, coin is 16 mm diameter. D.
1042 Cuspate stromatolites from the Neoproterozoic Mount Harper Group, Callison Lake Formation.

Table 1. Lithology and Interpretation

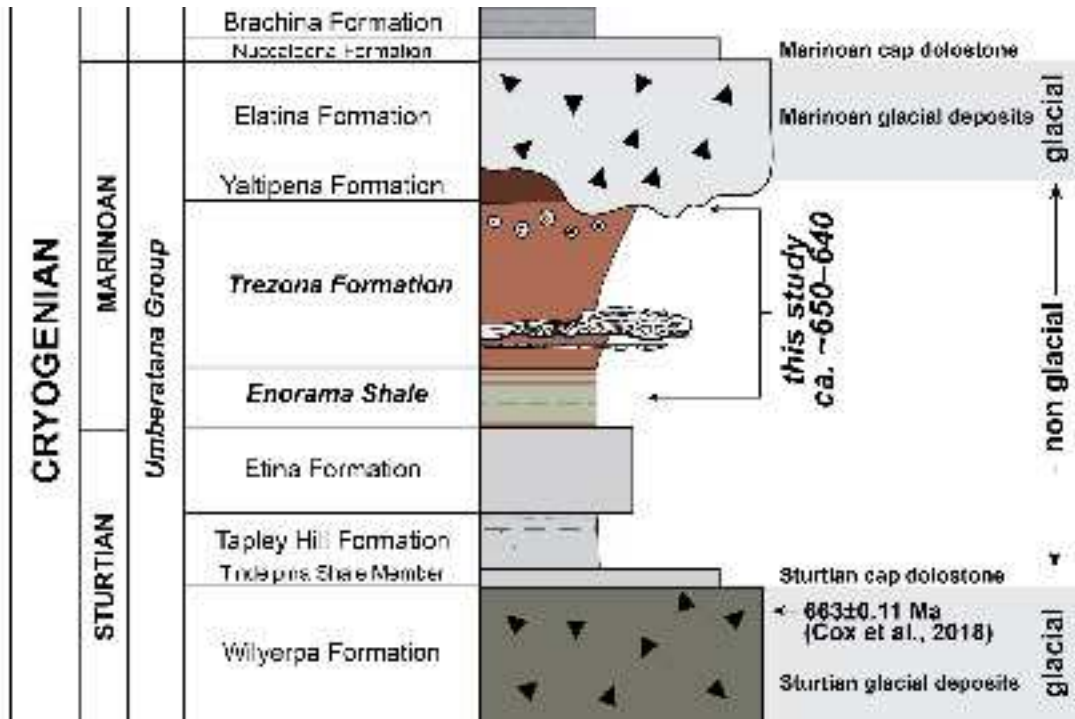
Formation	Member	Lithology/Facies	Process Interpretation	Environmental Interpretation	
Enorama Shale		Mudstone, rare silty-sandy beds with flute casts, rare sharp bedded silty-sandy beds with parallel lamination overlain by hummocky rippled beds	Low energy deposition with rare turbidites and storm beds.	Deep water basinal to outer ramp. Below fwwb, some intervals above swb	
Trezona Formation	siliciclastic member	lower	Calcareous mudstone	Sediment starved low energy deposition. Hardgrounds. Authigenic precipitation of dolomite on the seafloor (Wallace et al., 2019)	outer ramp
			Erosively-based, poorly sorted platy intraclastic packstone with chaotic arrangement of clasts. Can be capped with hardground.	Debris flow	outer ramp
			Erosion surfaces. Often capped by hardground.	Marine erosion, possibly associated with debris flows on ramp and/or strong erosional currents. Associated with extreme sediment starvation.	outer ramp
			Elongate cusped stromatolites	Sediment starved, deep water microbial carbonate deposition. Condensed deposition. Maximum transgression	outer ramp (where best developed)
			Mudstone	Deep water low energy suspension settling	mid-outer ramp
			Mudstone and siltstone with flute casts, unidirectional and climbing ripples	Turbidites, storm event beds(?)	mid-outer ramp
			Erosively-based, poorly-moderately sorted platy intraclastic packstone with edgewise and imbricated fabrics. Can be capped with a hummocky fine-grained grainstone. Rare.	Storm event bed or storm reworking of debris flow	mid ramp
		upper	Domal, columnar, and elongate cusped stromatolites	Microbial carbonate deposition during intervals of sediment starvation	mid ramp
			Rare erosion surfaces. Not capped with hardground.	Marine erosion via strong currents, and/or mass wasting on ramp	mid ramp
			Mudstone	Low energy suspension settling	mid-inner ramp
			Rounded, crossbedded intraclastic packstone-grainstone and fine-grained grainstone. Mud-drapes, bidirectional paleocurrents, some channelization.	Wave and tide-worked bedforms. Some tidal bars, tidal channels. Shallow water equivalents of debris flows offshore(?). Subtidal	inner ramp, tidal channels possibly mid-inner ramp(?)
			Wavy, flaser, and lenticular heterolithic bedding (mud to v.f.g. sand mixed lithologies)	Reworking and deposition by tidal currents. Subtidal	inner ramp
	carbonate member		Calcareous mudstone	Low energy suspension settling	mid-inner ramp(?) (similar environment to upper siliciclastic member but more

		carbonate rich)
Low relief, longitudinal, and columnar stromatolites. Very rare elongate cusped stromatolites	Microbial carbonate deposition during intervals of sediment starvation	mid-inner ramp(?) (similar environment to upper siliciclastic member but more carbonate rich)
Rounded and well-sorted intraclastic grainstone	Wave-worked intraclasts	inner ramp
Ooid grainstone. Giant ooids. Giant ooids with iron oxide rich laminae.	Continuous agitation of sediment in shallow, high energy conditions. Extremely high carbonate supersaturation (giant ooids). Fluctuating redox conditions (micrite and iron oxide rich laminae)	inner ramp
Domal and columnar stromatolite bioherms	Shallow marine microbial carbonate deposition	inner ramp

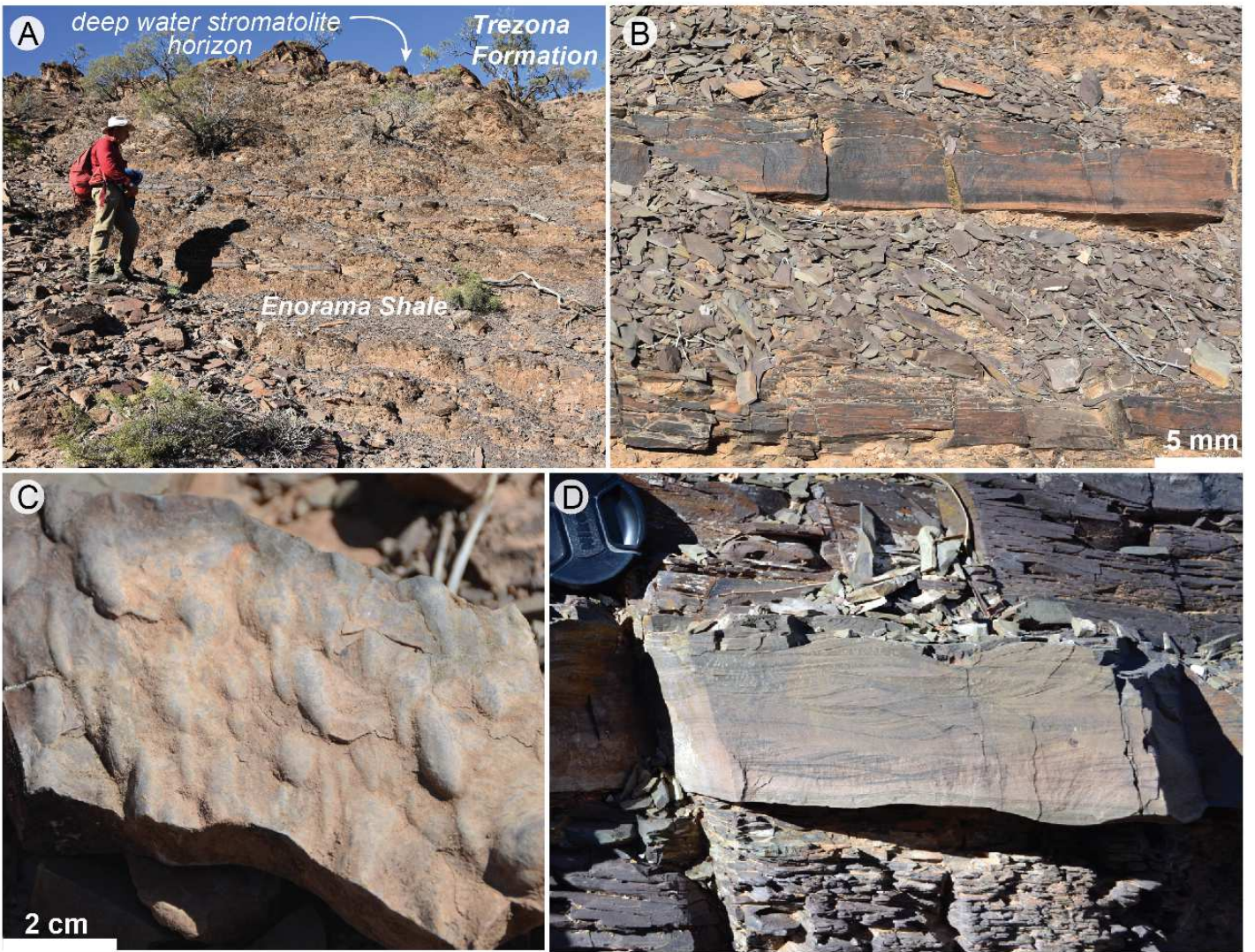


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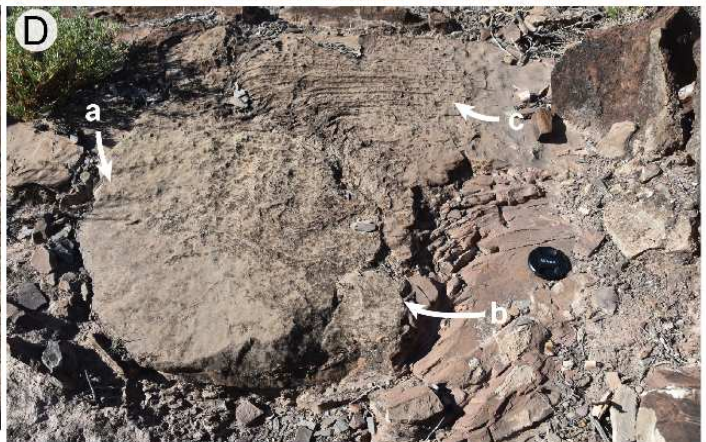
NEOPROTEROZOIC



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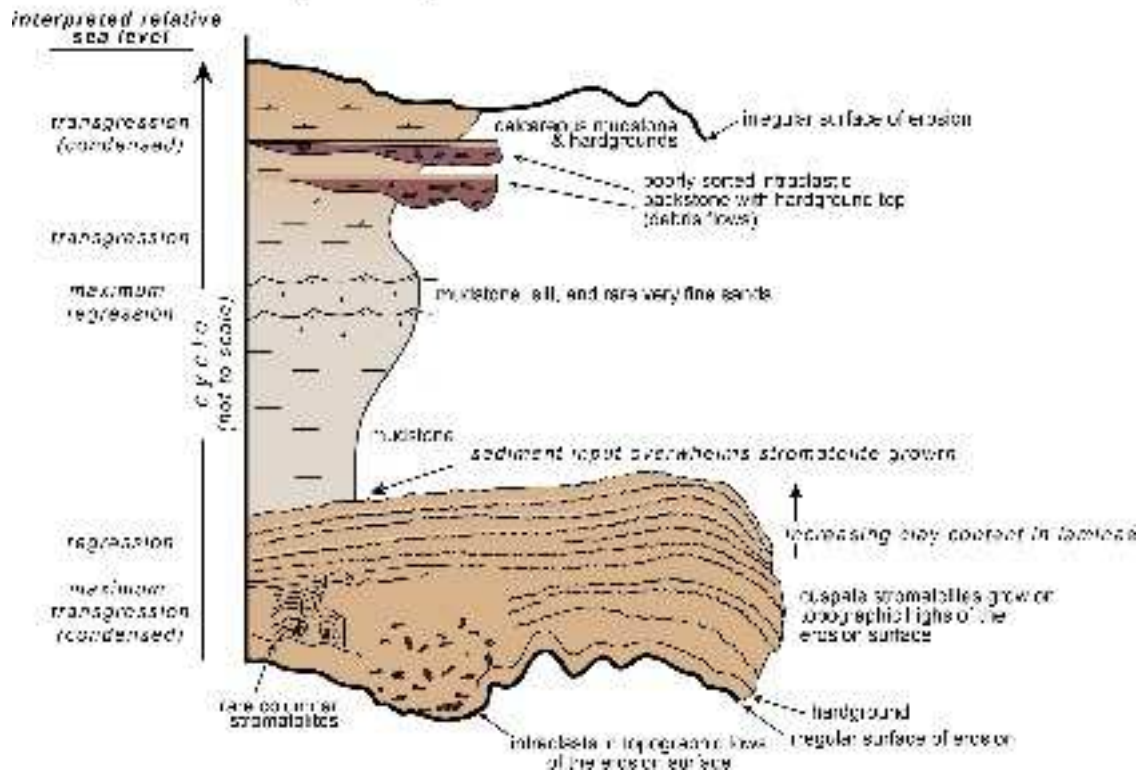


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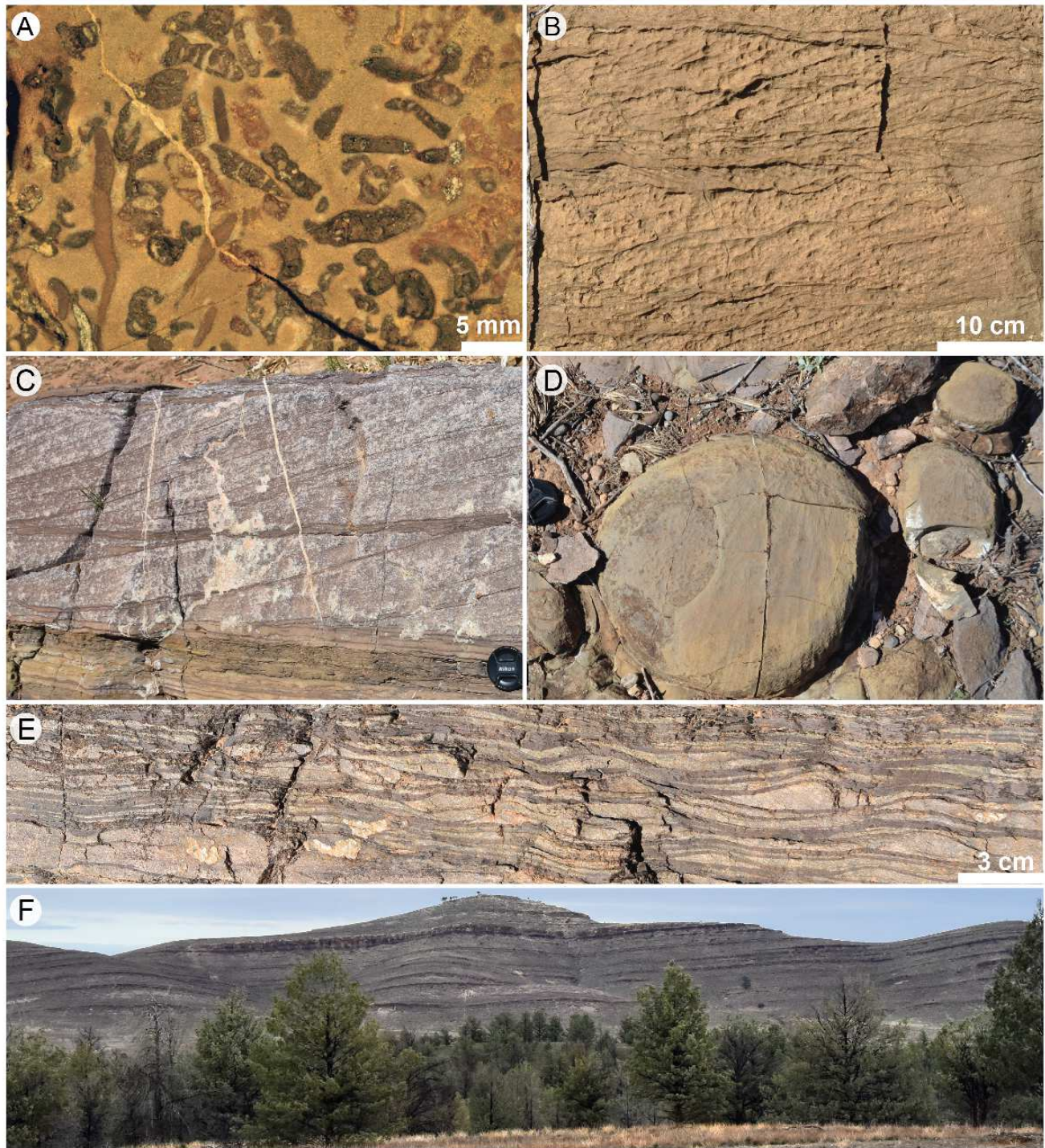


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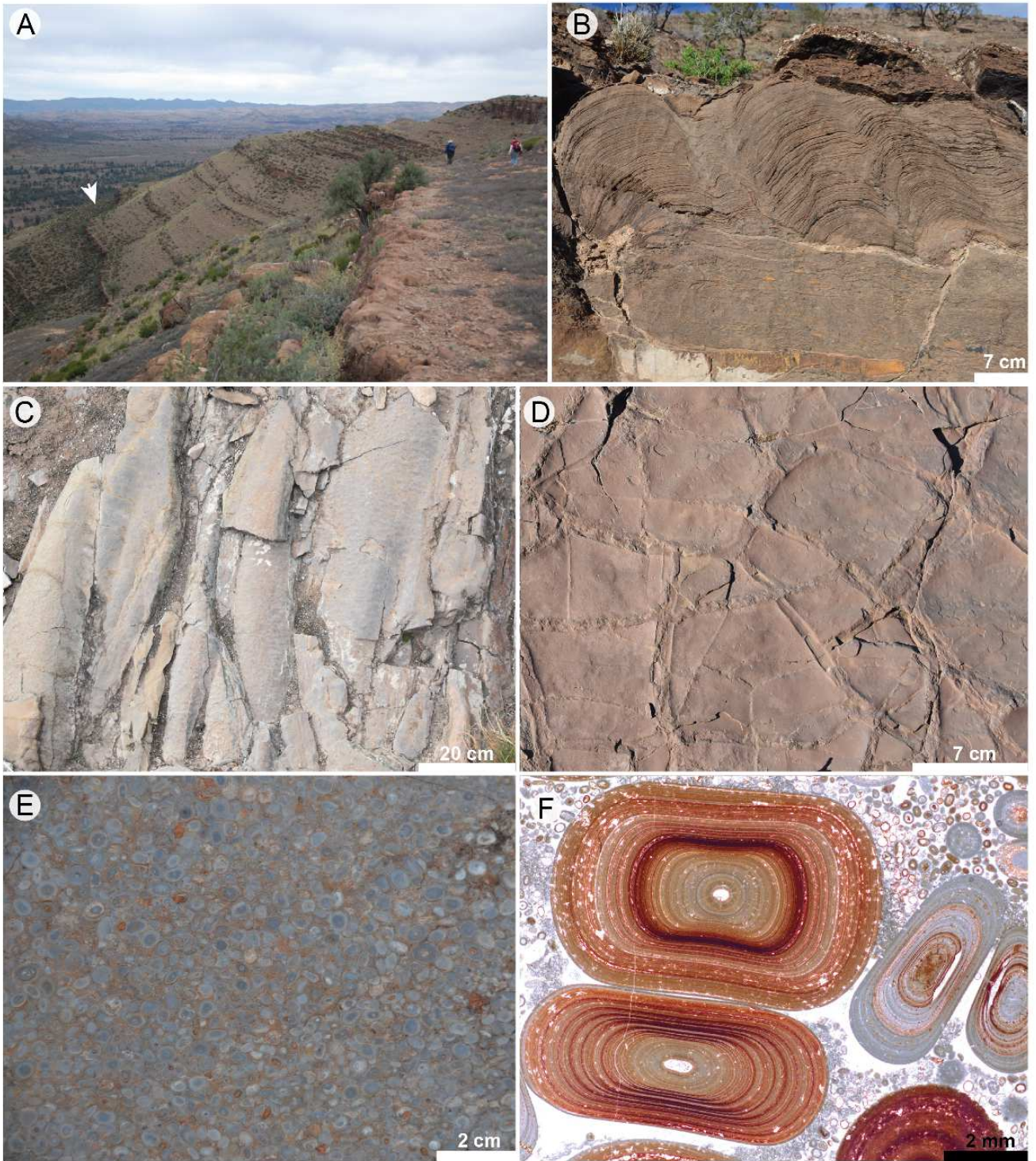
Idealized deep water cycle in the lower siliciclastic member



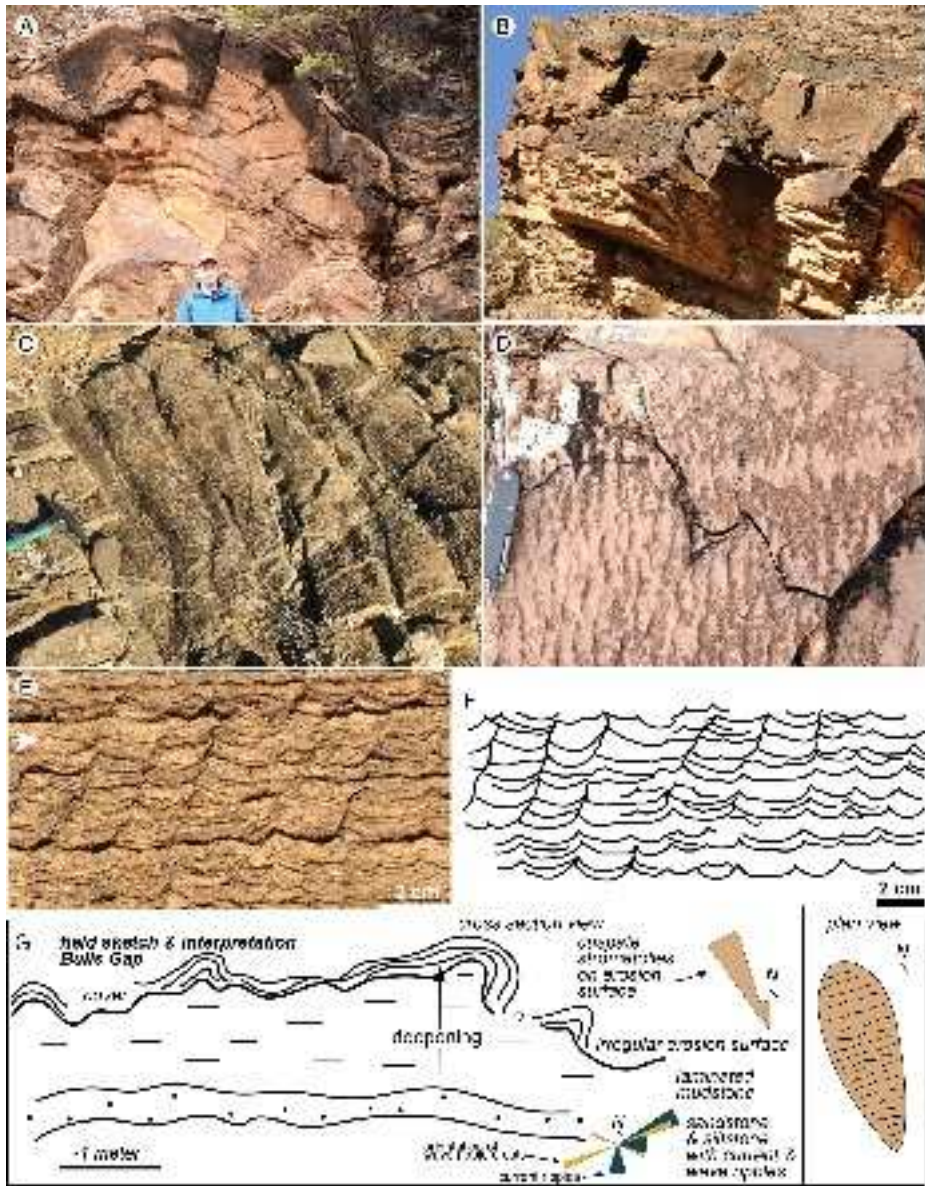
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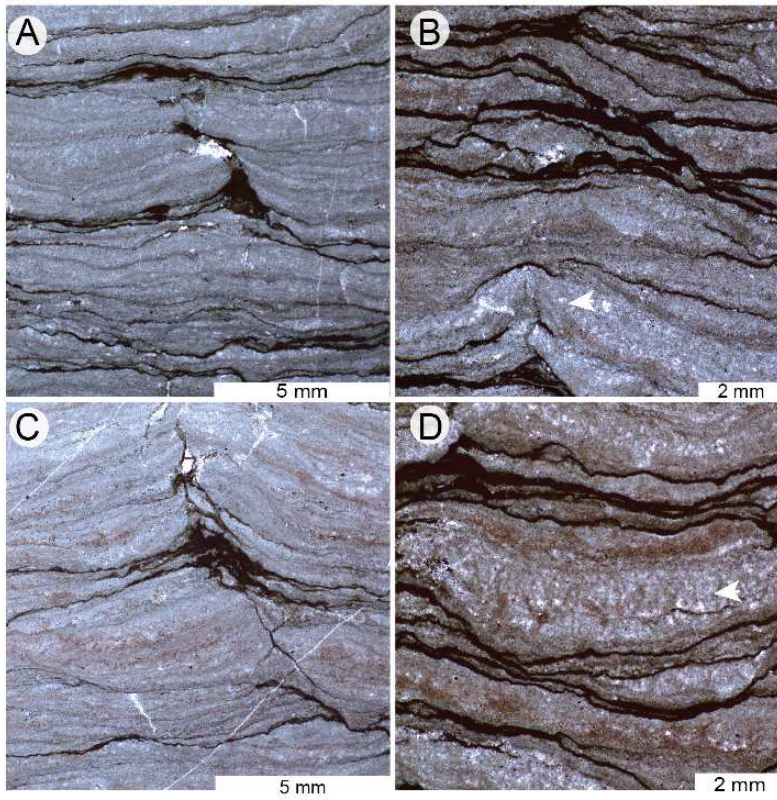
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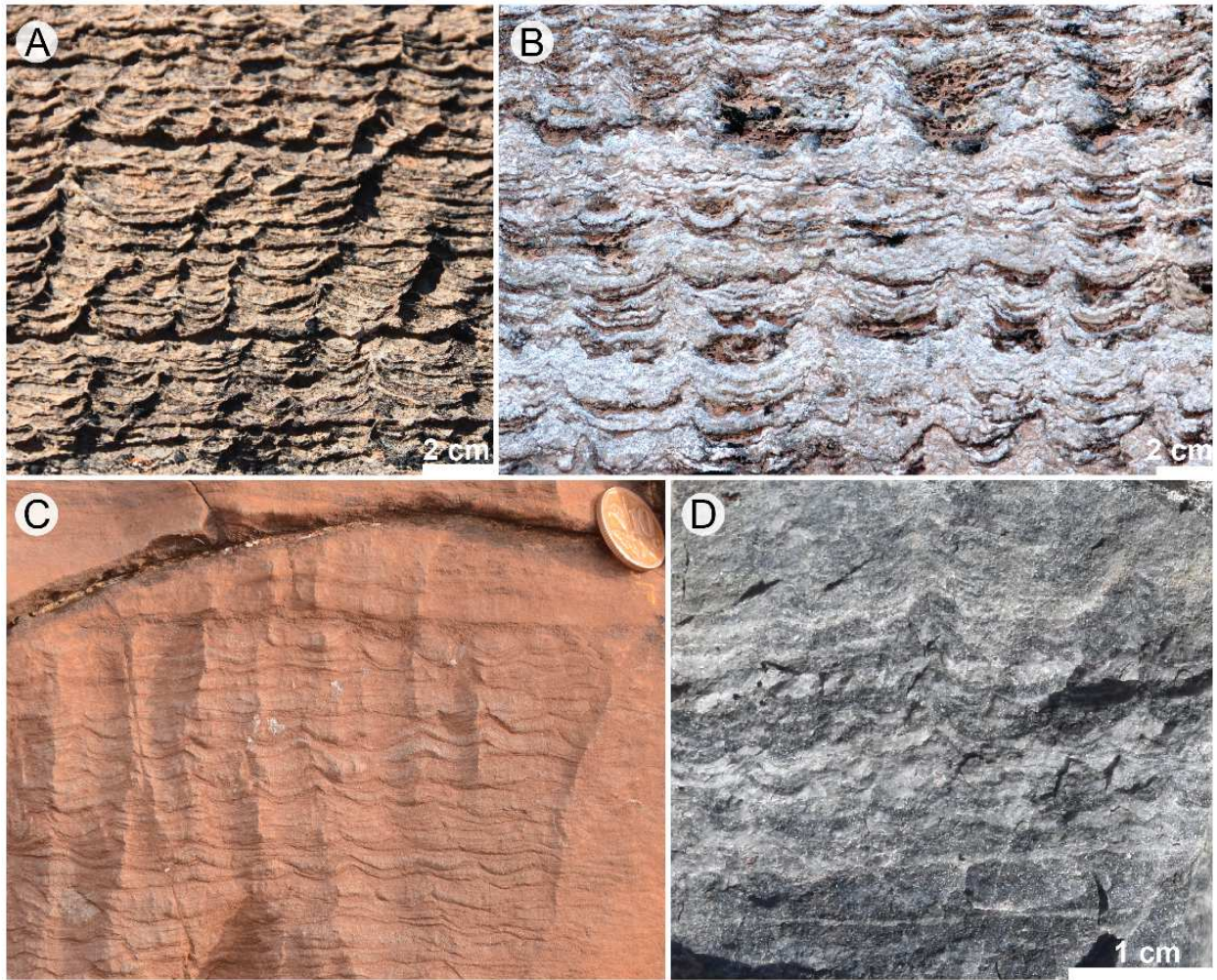
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gbi_12479_f9.png



gbi_12479_f10.png



gbi_12479_f11.png