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5 **Dolomitization of Triassic microbial mat deposits (Hungary): Origin of microcrystalline dolomite**

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18 inclusions

20 **Abstract**

21 Dolomite most commonly forms via replacement of precursor carbonate minerals. For this reason,
22 diagnosing primarily precipitated organogenic dolomite in microbial mat deposits from the rock record is
23 not straightforward, even though the deposits exhibit microbial fabric. Single and multiple dolomite crusts
24 exhibiting microbial fabric occur in a pervasively dolomitized Middle Triassic platform succession. Two
25 sections were studied in the Transdanubian Range. In both sections, two fabric types occur in the upper
26 part of the metre-scale cycles. One of that is microbial boundstone (fabric type 1)—characterised by
27 clusters of dolomite microcrystals which display diagnostic microbial features, such as calcimicrobes,
28 clotted–spherular aggregates and globules. The other one is different in the two studied sections. In
29 Section 1, it is micritic dolomite (fabric type 2) that is characterised by predominantly fine crystals and
30 contains obscured microbial components. In Section 2, it is bioclastic dolomite (fabric type 3) that is rich
31 in reworked dasycladalean alga fragments and consists of dolomite crystals of wide size-range from fine
32 to coarse. The precipitation of the microcrystalline dolomite phase is interpreted as being facilitated by
33 mats and biofilms favouring/tolerating an increasing frequency of subaerial conditions in the upper
34 intertidal setting. Petrographic analyses revealed that organogenic calcite was also precipitated, especially

35 in mat deposits rich in bioclasts. Synsedimentary dolomitization, resulting in fine crystals, was coupled
36 with aragonite dissolution and it postdated the organogenic precipitation. It took place only in the peritidal
37 caps of the shallowing-upward depositional units. Petrographic analyses provide circumstantial evidence
38 constraining that microcrystalline dolomite did not form via mimetic replacement. Accordingly the
39 microcrystalline dolomite, which shows microbial microfabrics in the studied samples, is interpreted as an
40 organogenic primary precipitate. Both peritidal processes, dolomite precipitation and replacement, were
41 likely controlled by the environmental factors in a semi-arid climate. Those components of the platform
42 succession that were not dolomitized in the peritidal environment were replaced and cemented by medium
43 and coarsely crystalline dolomite during further burial at elevated temperature, as shown by fluid inclusion
44 homogenisation temperature (62 to 83 °C) and negative stable oxygen isotope values. Thus, the majority
45 of the studied formation consists of fabric-destructive dolomite (fabric type 4).

46

47 **1. Introduction**

48

49 Currently, one of the most challenging issues in the field of carbonate sedimentology and geochemistry is
50 to understand the processes responsible for the mineralisation of a microbial mat. Studies in the last 30
51 years have elucidated the precipitation processes and suggested genetic models for microbially mediated
52 dolomites (e.g. McKenzie, 1981; Baltzer et al., 1994; Vasconcelos et al., 1995; Wright 1999; Mazzullo
53 2000; van Lith et al., 2003; Bontognali et al., 2010). These studies emphasise that bacterial degradation of
54 EPS organic matter is of major importance in the formations of microbial carbonates (e.g. Défarge et al.,
55 1996; Decho et al., 2005; Braissant et al., 2007; Gallagher et al., 2012; Krause et al., 2012). Recently,
56 Roberts et al. (2013) found a common association of dolomite precipitation with templates rich in certain
57 organic matter, which are carboxyl-groups, whose high densities occur in degraded natural organic matter
58 and in certain microbial surfaces.

59 Dolomite is a common diagenetic mineral in ancient carbonate rocks and reviews on its genesis are
60 numerous (Land, 1985; Tucker and Wright, 1990; Budd, 1997; Warren, 2000; Machel, 2004; among
61 others). It is generally recognised that dolomite forms via a dissolution–precipitation reaction in which a
62 calcium carbonate precursor is replaced by calcium–magnesium carbonate through interaction with
63 magnesium-rich fluids. Recently, studies in the Coorong lakes, Australia (Wright, 1999, 2000; Wright and
64 Wacey, 2005) and in the coastal environments of the Persian Gulf, United Arab Emirates (Bontognali et
65 al., 2010) suggested an organogenic precipitation model for that type of dolomite instead of the
66 conventional models of replacive dolomite formation by evaporative seepage and/or hydraulic pumping.
67 Diagnosing these genetic types, (1) organogenic dolomite or (2) organogenic calcium carbonate that was
68 subsequently mimetically replaced by dolomite, in ancient microbial mat deposits, is extremely difficult.

69 Theoretically, a petrographic distinction between the two alternatives is not possible in the rock record
70 because of the highly similar precipitation patterns of different carbonate minerals and resulting
71 morphologies documented from recent microbial mat deposits (e.g. Monty, 1976; Dupraz et al., 2004;
72 Bontognali et al., 2010; Couradeau et al., 2013). However, microbially mediated dolomite genesis was
73 interpreted by recognising sub-micrometre-sized spheroids in a few papers on ancient Phanerozoic
74 stromatolite, for example Mastandrea et al. (2006), Perri and Tucker (2007) and You et al. (2013).

75 Dolomites are characterised by a wide range in stable carbon isotope ratio. Mazzullo (2000) reported
76 that dolomite crystals associated to sulphate reduction or methanogenesis typically are ^{13}C -depleted and
77 ^{13}C -enriched, respectively. Many of the ancient shallow platform carbonates are evidently dolomitized in
78 multiple phases. Although the dolomites of different genesis have distinct stable isotope signals, as a rule,
79 the dolomite crystal phases—especially in the case of submicron-sized crystals—cannot be sampled and
80 measured separately via routine analysis from ancient rock samples. Moreover, in the case of ancient
81 rocks, it is not possible to determine whether disordered protodolomite or ordered dolomite precipitated
82 primarily (cf. Wenk et al., 1993).

83 Petrographic analyses revealed that Middle Triassic shallow-marine carbonates of the Budaörs
84 Dolomite contain peritidal deposits. Microfabrics indicate syndepositional organogenic carbonate
85 precipitation in microbial mat deposits and syndimentary dolomitization. The main objective of this
86 study is to analyse the facies-relationship of the diagenetic components which reveals the spatial and
87 temporal succession of diagenetic processes including organogenic precipitations and syndimentary
88 dolomitization.

89 Early stage syndimentary dolomite occurs only in relatively thin intervals of the studied Middle
90 Triassic carbonates. Subsequent diagenetic overprint by additional dolomitization and dolomite cement
91 precipitation at intermediate burial depth is indicated by petrographic, geochemical and fluid inclusion
92 data. These later processes resulted in pervasive dolomitization of a thick succession.

93

94 **2. Geologic setting**

95

96 The study area is located in north-central Hungary where the Triassic platform carbonates are exposed at
97 the surface in the north-eastern part of the Transdanubian Range (in the Buda Hills and Zsámbék Basin;
98 Figs. 1, 2). The present-day structural position of these rocks is the result of a large-scale displacement
99 during the Tertiary (Csontos and Vörös, 2004). In the Middle Triassic, rift tectonics formed topographic
100 highs with thick carbonate platform succession and coeval basins with limestone and tuff layers (Haas and
101 Budai, 1995; Budai, 2004). Based on the relatively well-preserved dasycladalean algae (Kutassy, 1927;
102 Piros unpublished data in Haas and Budai, 2004) the studied Budaörs Dolomite was assigned to the

103 Middle Triassic (uppermost Anisian–Ladinian) and was correlated with the South Alpine Schlern
104 Formations 1–2 (Masetti and Neri, 1980; Ruffer and Zühlke, 1995; Haas and Budai, 1999). The thickness
105 of the extensively dolomitized Middle Triassic platform carbonates is at least 1000–1200 m that was
106 estimated through compiling the geological map. In spite of the relatively great thickness the formation is
107 only discontinuously exposed. Based on a geological mapping of the region (Wein, 1977) two lithofacies
108 were recognised. (1) Massive dolomite generally occurs in great thickness and occasionally contains
109 dasycladalean algae. (2) Laminated dolomite contains cement-filled fenestral pores that define this
110 lithofacies as stromatolite. The latter lithofacies is exposed only at some locations. Facies succession
111 within the formation has not been studied yet.

112 Successively, in the course of the Late Triassic spreading stage, fault-controlled extensional basins
113 were developed via segmentation of the Middle Triassic carbonate platform in a number of locations
114 (Haas, 2002) whereas development of a shallow platform continued on vast areas (Main
115 Dolomite/Hauptdolomit Formation). The rocks of the Buda Hills were subjected to moderate deformation
116 during the Cretaceous. Thrust faults, folds and associated strong brecciation were observed in the studied
117 dolomite (Fodor et al., 1994). Due to tectonically-induced uplifting and intense denudation in the Late
118 Cretaceous and Early Neogene, post-Triassic Mesozoic formations are absent from the Buda Hills.
119 Intensive tectonic activity in the Late Eocene led to block faulting and brecciation. The fractures formed
120 due to the tectonic activity and the connected open pores of dolomite are filled with sediment consisting of
121 silt-sized dolomite clasts, which was later silicified (Magyari, 1994). In the study area, eroded surfaces of
122 dolomite are overlain by breccia and conglomerate that were formed in the surroundings of coastal cliffs
123 via abrasion during the Late Eocene (Magyari, 1999). Calcite, barite, fluorite and associated sulphide
124 minerals were precipitated along fractures within dolomite from hydrothermal basin-derived fluid expelled
125 along Middle Miocene faults (Márton and Fodor, 2003; Györi et al., 2011; Poros et al., 2012). The
126 inversion of the Neogene Pannonian Basin began in the latest Miocene and resulted in the uplift of certain
127 basement blocks, among them the Mesozoic–Palaeogene block of the Buda Hills.

128

129 **3. Materials and methods**

130

131 Sampling of the Middle Triassic Budaörs Dolomite was concentrated on specific lithofacies; accordingly
132 two short sections were studied where stromatolite and dasycladalean alga-rich lithofacies occur (Figs 1,
133 2). Section 1 is located at the cliffs north of Budaörs where two metre-scale intervals were sampled: three
134 samples were taken from the lower 5.8 m (samples 1–3) and eight from the uppermost 2.1 m (samples 4–
135 11). There is an approximately 8-m-thick covered interval between them. Section 2 is located in an

136 abandoned quarry west of Zsámbék from where four samples (12–15) were collected from a 1-m-thick
137 interval.

138 Twenty-two thin sections were examined by conventional petrographic microscopy. The potential
139 presence of organic matter was evaluated using a microscope equipped with an Hg vapour lamp and filters
140 for blue light excitation (450–490 nm). The filter set was composed of a diachromatic beam splitter (510
141 nm) and a barrier filter (515 nm). Cathodoluminescence (CL) petrography was carried out using a Nuclide
142 ELM–3R cold cathodoluminescence device operating at 10 kV. In order to distinguish between calcite,
143 dolomite, and their ferroan variants, all the thin sections were stained with Alizarin Red-S and potassium
144 ferricyanide as described by Dickson (1966).

145 Electron microprobe analyses (EPMA) were carried out for quantitative geochemical analysis of one
146 selected, polished, carbon-coated sample (No. 10). The measurements were done by a JEOL JXA-8600
147 Superprobe (upgraded with SAMX control). The operational parameters and standards were the following:
148 15 kV acceleration voltage, 20 nA probe current, 5x4 μm defocused beam, PAP correction, dolomite
149 USNM 10057 (for Ca and Mg), siderite USNM R 2460 (for Fe) and strontianite NMNH R 10065 (for Sr),
150 and CO_2 was fixed to 46,90 wt. %. Distinct dolomite fabrics were sampled for stable carbon and oxygen
151 isotope analyses, using a hand-held microdrill with a 0.5 mm bit-head. The carbonate powders were
152 divided into two subsamples that were measured separately. The powders were analysed using the
153 continuous flow technique with the H_3PO_4 digestion method (Rosenbaum and Sheppard 1986; Spötl and
154 Vennemann, 2003). $^{13}\text{C}/^{12}\text{C}$ and $^{18}\text{O}/^{16}\text{O}$ ratios of CO_2 generated by acid reaction were measured using a
155 Thermo Finnigan Delta Plus XP continuous flow mass spectrometer equipped with an automated
156 GasBench II. The results are expressed in the δ -notation [$\delta = (R_1/R_2 - 1) \times 1000$] where R_1 is the $^{13}\text{C}/^{12}\text{C}$ or
157 $^{18}\text{O}/^{16}\text{O}$ ratio in the sample and R_2 the corresponding ratio of the Vienna Pee Dee Belemnite (V-PDB)
158 standard, in parts per thousand (‰). Duplicates of standards and samples were reproduced to better than
159 ± 0.15 and ± 0.1 ‰, for oxygen and carbon isotopes, respectively.

160 For fluid inclusion studies, 80–100- μm -thick, doubly polished thin sections were prepared. In order to
161 minimise sample heating and fluid inclusion stretching during sample preparation, a low-speed saw was
162 used for cutting the rock samples. Conventional fluid inclusion petrography and microthermometry were
163 performed on a Linkam FTIR 600 heating–cooling stage mounted on a polarisation microscope.
164 Standardisation was carried out at temperatures of -56.6 , 0 and 385 °C using quartz wafers containing
165 synthetic H_2O and $\text{H}_2\text{O}-\text{CO}_2$ fluid inclusions. The accuracy of the measurements during freezing
166 experiments and heating up to 150 °C was 0.1 °C.

167

168 4. Petrography

169

170 The studied pervasively dolomitized successions consist of two alternating lithofacies occurring in metre-
171 scale cyclic packages. The light grey massive lithofacies is overlain by a white thin-bedded/laminated one
172 (Fig. 2). Their transition is gradual. Both lithofacies could be subdivided into more fabric types by
173 microscopic study. Accordingly, all together four fabric types are distinguished, which occur
174 systematically in accordance with the depositional succession (from bottom to top): fabric-destructive
175 dolomite and bioclastic dolomite are found in massive lithofacies whereas micritic dolomite and microbial
176 boundstone characterise the thin-bedded/laminated one. Dolomite veinlets cut across both lithofacies and
177 detrital dolomite occurs as internal sediment in pores and fractures.

178 Altogether four non-ferroan dolomite crystal phases (DOL-1–4) were identified and classified
179 according to the crystal-size and textural relationships. These are as follows; DOL-1 is microcrystalline
180 dolomite which is defined by submicron-sized crystals; DOL-2 is finely crystalline replacive dolomite
181 which is typified by micrite and microspar-sized crystals; DOL-3 is medium and coarsely crystalline
182 replacive dolomite which is characterised mostly by 70–350- μm -sized turbid crystals; and DOL-4 is
183 dolomite cement which forms a limpid overgrowth zone on DOL-3 phase maximum at a thickness of 150
184 μm .

185

186 *4.1. Thin-bedded/laminated lithofacies*

187

188 In Section 1, thin-bedded/laminated lithofacies is characterised by regular repetitions of thin layer-
189 couplets consisting of micritic dolomite in the lower part and microbial boundstone above (Fig. 3). In
190 Section 2, only thin microbial boundstone layers were observed. In micritic dolomite layers, characteristic
191 components are the flat-pebbles at the base, ripped up from the underlying lithified layer. The size of
192 reworked and rounded lithoclasts decreases upward and they disappear at the upper part of the layer. A
193 dispersed brownish-colouring around the rip-up clasts is common. Burrow-mottled fabric is typical.
194 Occluded laminoid fenestral pores occur more frequently in the microbial boundstone layer. The upper
195 surface of microbial boundstone is often sharp, uneven and occasionally brecciated. Only a few skeletal
196 fragments were encountered, mainly articulated and disarticulated ostracod valves, subordinately
197 dasycladalean algae, foraminifera and gastropods.

198

199 *4.1.1. Microbial boundstone, fabric type 1 (FT1)*

200 Microbial boundstone fabric type is characterised by the ubiquitous presence of microcrystalline dolomite
201 (DOL-1). These dense submicron-sized crystals form clotted–spherular aggregates and bundles of
202 prostrate threads. The textural relationship of solid clots and chambered spheroids exhibits a size-related
203 hierarchy such as aggregates of tiny clots gradually developing into aggregates of larger spheroids. Along

204 with an increasing size of the spheroids their inner part is less densely filled with microcrystals (Fig. 4).
205 One of the most common microfabric types, where attached clots and associated hollow spheroids form
206 closely packed, upward-expanding bushy aggregates, resembles *Renalcis* and *Angusticellularia*
207 calcimicrobes (Pratt, 1984; Riding, 2000; Stephens and Sumner, 2002). Additionally, tufted filamentous
208 calcimicrobes, which resemble *Cayeuxia* (Riding, 1991), are occasionally preserved and surrounded by
209 clotted aggregates. Uniform, oval-shaped globules of *ca* 10–20- μm -size are closely packed and embedded
210 in dense microcrystalline groundmass (Fig. 4).

211 Finely crystalline dolomite (DOL-2) is characterised by micrite and microspar-sized crystals and is
212 present in certain intervals where their amount decreases upward. The mixture of microcrystalline
213 aggregates and fine crystals occurs as a massive groundmass or forming a nodular and/or reticulate
214 microfabric (Fig. 5A, B). The DOL-1 shows bright green fluorescence whereas a mixture of DOL-1 and
215 DOL-2 exhibits dull green fluorescence under blue light (Fig. 5C). Both dolomite types display dull red
216 luminescence under CL but the luminescence of the mixture of DOL-1 and DOL-2 is less intense (Fig. 6).

217 The microstructure of FT1 is typified by cyclic packages of lamina-couplets, which consist of lower
218 nodular/reticulate lamina and overlying dense-clotted microcrystalline lamina (Fig. 5A).

219

220 4.1.2. Micritic dolomite, fabric type 2 (FT2)

221 Micritic dolomite is characterised by the abundance of fine crystals of mostly micrite-size (up to *ca* 20
222 μm ; DOL-2). The fabric additionally includes faint microcrystalline clot clusters (DOL-1) and medium-
223 sized subhedral crystals (DOL-3) in a heterogeneous mixture with fine crystals (Fig. 7). In the upper part
224 of the layer, poorly-defined calcimicrobes (DOL-1) are definitely present and become well-defined and
225 abundant in the overlying microbial boundstone. Under blue light, this fabric exhibits a heterogeneous dull
226 green and non-fluorescent groundmass with dispersed brighter clots and spots (Fig. 7B, D).

227 In the lower part of Section 1, the thin-bedded/laminated lithofacies are the thinnest. Here the upper
228 surface of the micritic dolomite layers is commonly uneven, brecciated and the microbial boundstone
229 (FT1) is absent. Upsection the microbial microfabric progressively becomes obvious through the
230 increasing abundance of DOL-1 crystals among the DOL-2 micrite crystals but FT1 is commonly very
231 thin. In the upper part of Section 1, at microscopic scale, the upward transition from micritic dolomite to
232 microbial boundstone is manifested either in lamina-scale alternations or in a patchy microstructure. In the
233 latter case, the nodular microfabric type (FT1) forms irregular and discontinuous patches within the
234 micritic dolomite (FT2).

235

236 4.1.3. Pore types and infilling phases

237 Open porosity is absent or very minor in the studied samples, but several different occluded pore types
238 were observed. Arborescent clumps are ubiquitous in the microbial boundstone (FT1) and among them
239 occluded pores of a wide size-range occur. The pores formed a complex network system that was
240 occluded by internal sediment and cement infillings (Figs 4, 5). The volume of the original pore space was
241 significant and thus, pores were major components of the microfabric. Three pore types can be
242 distinguished. The most common type included fenestral and/or desiccation pores of variable sizes and
243 shapes, from some tens of micrometres up to several millimetres. The relatively large pores were typically,
244 but not exclusively, elongate and sub-parallel to the bedding. The second type of original pores was a
245 biomould, and the third includes thin fractures and breccia pores.

246 In certain intervals in Section 2, two dolomite phases occluded the fenestral/desiccation pores, which
247 are fine crystals in micrite size (DOL-2) and brownish anhedral crystals in medium size (*ca* up to 130 μm ;
248 DOL-3). The DOL-3 phase appears as sets of elongate crystals. They exhibit sweeping extinction that
249 moves into one direction. Fine crystals (DOL-2) densely or dispersedly surround the brownish crystals in
250 a heterogeneous pattern. These DOL-2 and DOL-3 crystals are mostly non-fluorescent but exhibit faint
251 green mottles (Fig. 4C, D). Locally, brownish-coloured pendant linings are observed at the roof of larger
252 pores where fine crystals show dull green fluorescence in bands (Fig. 4C, D).

253 Variable amounts of fine crystal silt (DOL-2) cover the bottom of voids in microbial boundstone both
254 in Section 1 and 2. The dolomite silt contains lithoclasts originated from the roof of the cavity. In certain
255 beds, a larger amount of dolomite silt completely occluded most of the pores or formed cap-lamina
256 containing floating tattered microbial boundstone fragments. Under blue light, the crystal silt exhibits dull
257 green fluorescence (Fig. 5). Fractures cut across all the above described components.

258 The rest of the pore space in microbial boundstone was occluded by two generations of dolomite
259 crystals, DOL-3 and DOL-4 (Fig. 5). The first phase (DOL-3) is characterised by blocky crystals that are
260 slightly turbid because of the inclusions (Figs 5B, 6C). Many of the solid inclusions are brownish in
261 colour. The crystals show mottled fluorescence and luminescence, i.e. scattered, bright spots appear in a
262 non-fluorescent/non-luminescent background (Figs 5C, 6D). Except for the largest pores, the first
263 generation crystals generally occlude the entire pore space. Where it does not, the second generation of
264 crystals (DOL-4) appears as limpid optical overgrowth on the DOL-3 phase that is typified by
265 rhombohedral termination (Fig. 5B). These crystals have straight or very faint undulose extinction. Two
266 growth bands in the DOL-4 cement phase are visible under blue light as well as under CL. The first one
267 exhibits dull fluorescence with fine subzones and is non-luminescent whereas the second one shows the
268 opposite pattern, being non-fluorescent and of dull luminescence with fine subzones (Figs 5B–C, 6C–D).

269 In micritic dolomite (FT2), limpid subhedral crystals fill a pore network system that is more obvious in
270 the upper part of the layers (Fig. 7C). The crystal size is *ca* 70–90 μm in the lower part of the layers and

271 gradually increases upsection up to 250 μm (DOL-3 and DOL-4). These crystals show growth bands under
272 blue light that are similar to the pattern of the two generations of pore-filling crystals in the FT1 (Fig. 7).

273 Fractures cut across the above-described components, which are filled with silicified internal sediment
274 consisting of silt-sized dolomite (Fig. 3). Silicified dolomite silt is the last pore-occluding phase, post-
275 dating the DOL-4 dolomite cement phase, in the largest, bedding-parallel pores (Fig. 5B, C). The internal
276 sediment often includes detrital fragments of the DOL-4 cement crystals.

277

278 *4.2. Massive lithofacies*

279

280 Massive lithofacies is characterised by a lack of any macroscopic sedimentary structure and it includes
281 bioclastic dolomite and fabric-destructive dolomite. In Section 1, sets of lamina-couplets, identical to
282 those observed in the thin-bedded/laminated lithofacies (FT1–FT2), are present but they are only
283 detectable in thin sections. Bioclastic dolomite occurs only in Section 2 where it overlays fabric-
284 destructive dolomite. Dasycladalean alga fragments are abundant in the bioclastic dolomite; in addition,
285 foraminifers, gastropods and crinoid ossicles are also present, scarcely. Lithoclasts of microcrystalline
286 dolomite are locally very common in both fabric types, especially in those samples, where this lithofacies
287 overlays microbial boundstone.

288

289 *4.2.1. Bioclastic dolomite, fabric type 3 (FT3)*

290 Although the bioclastic dolomite can be classified as fabric-destructive dolomite the depositional texture is
291 partly reflected by the variable crystal size as well as the brownish-staining (due to the abundance of solid
292 inclusions) of the coarser crystals (Fig. 8A). Bioclastic dolomite is characterised by medium and coarsely
293 crystalline dolomite consisting of closely packed subhedral–anhedral crystals up to *ca* 400 μm in size
294 (DOL-3 and DOL-4). These crystals show undulose extinction under crossed polars. Additionally, finely
295 crystalline dolomite (DOL-2), consisting of micrite-sized crystals, appears at the bottom of various pores
296 and otherwise heterogeneously dispersed in the fabric. The DOL-3 phase exhibits a distinct appearance
297 and it is present in three forms, such as dark brownish mosaics, lighter brownish, isopachous elongate
298 crystals and turbid mosaics. The dark brownish mosaic crystals are distributed in irregular patches and
299 they cover the bioclasts. Elongate crystals occur in certain patches—among the ghosts of bioclasts and in
300 the primary intraparticle pores of skeletal fragments—in the lower part of FT3 intervals (Fig. 8B). The sets
301 of elongate crystals exhibit sweeping extinction, which moves into one direction. The various crystal
302 phases display distinct fluorescence under blue light (Fig. 8C, D).

303 The bioclastic dolomite shows gradual transition to microbial boundstone (FT1). In the transitional
304 interval, microcrystalline clot clusters (DOL-1) occur together with dark brownish-stained, fine to

305 medium-sized anhedral mosaic crystals (DOL-3; Fig. 9). The amounts of both the brownish mosaic
306 crystals and the bioclasts decrease upward; thereafter, they became subordinate or disappear in the
307 overlying microbial boundstone along with the increasing amounts of clot clusters of DOL-1
308 microcrystals.

309

310 *4.2.2. Fabric-destructive dolomite, fabric type 4 (FT4)*

311 Fabric-destructive dolomite is predominantly characterised by a medium crystalline (70–250 μm)
312 dolomite consisting of closely packed subhedral–anhedral crystals (Fig. 10). Locally, fine and coarser
313 crystals may also co-occur (*ca* up to 350 μm). The crystals are variably inclusion-rich, or characterised by
314 turbid core and limpid rim (DOL-3 and DOL-4, respectively). The majority of crystals show undulose
315 extinction. Under blue light, the limpid crystals exhibit a fluorescence pattern that is highly similar to that
316 of the pore-filling cement phase (DOL-4) within FT1; otherwise, this fabric typically shows blotchy
317 fluorescence.

318 In the lower part of Section 1, the fabric-destructive dolomite shows a gradual transition upsection to
319 the lamina of the microbial boundstone that was observed in thin sections. At the lower part of the
320 transitional interval, DOL-2 fine crystals appear and they become more common upwards. Under blue
321 light, the fabric-destructive dolomite exhibits a spongy fabric that is very similar to the fabric of the
322 overlying microbial boundstone, but along with an increasing crystal size the fabric progressively becomes
323 obscured (Fig. 10B, C).

324

325 **5. Geochemical data**

326

327 *5.1. Major and trace element compositions*

328

329 Although the back-scattered electron image of dolomite crystals displayed grey-scale heterogeneity there
330 are no measurable differences in chemical compositions between contrasting areas of any of the dolomite
331 types. Concentrations of trace elements were below the detection limit of the EPMA.

332

333 *5.2. Stable carbon and oxygen isotopes*

334

335 The heterogeneity of the dolomite crystals inhibited their separate sampling. The DOL-3 and DOL-4
336 crystal phases were sampled and measured together from the largest pores. In all other samples, bulk rock
337 powders containing multiple dolomite crystals were analysed (Table 1; Fig. 11). The $\delta^{13}\text{C}_{\text{V-PDB}}$ values of
338 all analyses are similar, ranging between 2.2‰ and 3.9‰. In contrast, the $\delta^{18}\text{O}_{\text{V-PDB}}$ values of DOL-3 and

339 DOL-4 phases from large pores (-4.3‰ to -1.7‰) are depleted in ^{18}O relative to those of bulk samples.
340 Bulk samples from microbial boundstone, micritic dolomite and bioclastic dolomite (potential mixtures of
341 all types of dolomite crystals) yielded $\delta^{18}\text{O}_{\text{V-PDB}}$ values of 0.2‰ to 1.2‰ . The fabric-destructive dolomite
342 has values in a wide range. The sample (No. 3; Table 1) containing predominantly DOL-3 crystals has
343 values -1.9 and -0.1‰ . The other sample (No. 10; Fig. 10A) containing lithoclasts has a value 0.3‰ . The
344 sample (No. 1; Fig. 10B, C), which contains a microcrystalline microbial boundstone lamina in thin-
345 section-scale and a relatively large amount of fine crystals among the medium-sized crystals, has a value
346 of 1.6‰ at 1 cm below the nodular part of the lamina, and 0.7‰ and 1.1‰ at 2 cm and 3 cm below that
347 one, respectively.

348

349 **6. Fluid inclusion petrography and microthermometry**

350

351 In order to obtain information on the temperature and the composition of dolomitizing fluid(s), fluid
352 inclusions of two samples from microbial boundstone layers (Section 1) were studied in detail (Table 2;
353 Fig. 12). Fluid inclusion data from fabric-destructive dolomite samples of the same formation from a
354 nearby outcrop were also available (Poros, 2011). In one of the microbial boundstone samples, only the
355 pore-filling DOL-3 and DOL-4 crystals contained measurable aqueous fluid inclusions. In the other
356 sample, measurable fluid inclusions were found also in the fabric-destructive dolomite that occurs in
357 centimetre-scale patches at the crossing of dolomite veinlets in the upper part of the microbial boundstone
358 layer.

359 In the pore-filling crystals, the appearance of fluid inclusions in the turbid DOL-3 generation and the
360 limpid DOL-4 is identical; however, the abundance of inclusions in the limpid rim is significantly lower
361 compared to the other phase. Most of the primary aqueous inclusions contained both liquid (L) and vapour
362 (V) phases (Fig. 12A). In the turbid crystals, the primary inclusions were found along growth zones,
363 whereas primary inclusions of the limpid rim are randomly distributed. At room temperature, the vapour
364 bubble was moving in the case of most inclusions. All-liquid inclusions were also rarely observed;
365 however, where seen they were close to two-phase inclusions, presenting evidence for necking-down after
366 phase separation. Those two-phase inclusions, located next to single-phase ones, were not measured. The
367 size of the measured inclusions ranges between 5 and 15 μm . The visually-determined liquid–vapour ratio
368 ranges between 95:5 and 90:10; and no gas-rich inclusions occur.

369 Euhedral crystals of the fabric-destructive dolomite sample contain similar fluid inclusions to the
370 previous sample. However, in this sample, not only the euhedral crystals but also the replacive finer,
371 anhedral and subhedral crystals (DOL-3) contain small (*ca* 3 μm), but still measurable, randomly
372 distributed two-phase (L–V), aqueous fluid inclusions.

373 All the measured DOL-3 and DOL-4 crystal phases exhibited similar ranges in homogenisation
374 temperatures (T_h) of the primary two-phase inclusions. Homogenisation temperature in turbid DOL-3 and
375 limpid DOL-4 crystals ranged between 72 and 79 °C, with one outlying value at 90 °C which is
376 considered to be invalid due to the supposed volume change after entrapment. Anhedral and subhedral
377 DOL-3 finer crystals of the latter sample show a slightly wider and lower temperature range from 62 up to
378 83 °C (Fig. 12B). 90% of all the data from the different crystals fall into a narrow range from 73 to 83°C.
379 Entrapment temperatures of the fluid could not be calculated (i.e. no pressure correction was applied), but
380 the homogenisation temperature values still provide a valid measure of the minimum entrapment
381 temperature (Goldstein and Reynolds, 1994).

382 Cryoscopic and heating measurements could not be carried out on the same inclusion, because the
383 vapour bubble disappeared during homogenisation and never reappeared. Cryoscopic measurements were
384 not successful on other inclusions because the vapour phase became metastable during freezing, probably
385 because of the very low vapour/liquid ratio. It was also not possible to detect the eutectic temperature
386 because of the very small size of the inclusions. Only, three inclusions were appropriate for salinity
387 measurements. All of them were hosted by the turbid core of the dolomite crystals (DOL-3). The salinity
388 values calculated from the final melting temperatures, assuming a NaCl–H₂O system, are 3.4, 3.8, and 6.4
389 NaCl equ. wt. %.

390

391 **7. Discussion**

392

393 *7.1. Interpretation of sedimentary features*

394

395 *7.1.1. Microbial boundstone (FT1)*

396 The ubiquitous presence of microcrystalline dolomite (DOL-1) in the form of dense groundmass, clots,
397 clot clusters and calcimicrobes suggests that the FT1 was derived from the mineralisation of bacterial EPS
398 (Riding, 2000). Observations on modern and recent benthic microbial mat communities indicate that
399 micrite nucleation is initiated within the EPS and its alveolar organic network is progressively replaced by
400 mineral precipitation, high-Mg calcite (e.g. Défarge et al., 1996; Gautret et al., 2004; Dupraz et al., 2004),
401 aragonite (e.g. Monty, 1976; Couradeau et al., 2013), or dolomite (e.g. Bontognali et al., 2010).
402 Accordingly, a wide variety of precipitation patterns may be developed, which is reflected by the
403 microfabric. These patterns include dispersed smaller and larger spheroids (Bontognali et al., 2010),
404 spherular aggregates (Wright, 1999), clots and clot clusters progressing into a massively mineralised
405 groundmass (Dupraz et al., 2004), dissected or continuous laminae (Visscher et al., 2000) and
406 calcimicrobes (e.g. Riding, 2000).

407 In the studied samples, the microfabric components are consistent with carbonate mineral precipitate
408 morphologies that have been recorded from modern EPS substrate. The spongy fabric is due to the
409 fenestral pores, formed penecontemporaneously within the EPS during the degradation of organic matter
410 (Défarge et al., 1996) and desiccation pores, formed successively during ephemeral subaerial exposure
411 (Shinn, 1983). Additionally, tiny globules accompany the most abundant microcrystalline precipitates.
412 Their size and shape resemble empty ghost remnants of coccoid cyanobacterial sheaths (cf. Dupraz et al.,
413 2004; Golubic and Abed, 2010). The fluorescence in the microcrystalline components is most likely
414 caused by the associated diffuse organic matter (e.g. Dravies and Yurewicz, 1985; Bertrand et al., 1986).
415 Coexistence of carbonate mineral precipitation and organic matter requires anoxic and slightly alkaline
416 conditions (Krumbein and Garrels, 1952; Visscher and Stolz, 2005), and under such conditions, Mn^{2+}
417 could build in the carbonate lattice and that Mn^{2+} even in small amounts could have activated the
418 fluorescence (Gaft et al., 2005). Fine detrital sediment (replaced by DOL-2) was incorporated into the mat
419 and amalgamated with microcrystalline aggregates, forming reticulate or nodular microfabrics in the
420 course of diagenesis. However, an alternative, such as microcrystalline aggregates merging through
421 further precipitation (cf. Dupraz et al., 2004; Bontognali et al., 2010; Spadafora et al., 2012), cannot be
422 excluded.

423
424 *7.1.2. Micritic dolomite (FT2)*
425 Obscured microbial components indicate that the buried microbial mat was definitively present in the
426 precursor of the micritic dolomite but its preservation was limited. The downward decreasing size of the
427 antecedent pores (later filled by cement crystals) is attributed to physical compaction, also implying
428 diffuse mineralisation of the mat (cf. Dupraz et al., 2009). The abundance of sand-sized detrital
429 sedimentary grains, together with the bioturbational mottles, suggests that detrital micrite was also present
430 as sedimentary components. The detrital carbonate mud was replaced by the fine dolomite crystals (DOL-
431 2).

432
433 *7.1.3. Bioclastic dolomite (FT3)*
434 The bioclasts are embedded within patches of dark brownish-stained DOL-3 mosaic crystals that are
435 interpreted, on the basis of the clotted fluorescence pattern, as replaced organogenic precipitate. The
436 spongy pore network within the precursor, filled by DOL-2–4 phases, is interpreted as having a
437 fenestral/desiccation origin. This also implies the predominance of microbial precipitation within this
438 fabric type. Accordingly, the precursor deposits of the bioclastic dolomite were determined by a microbial
439 mat. The mineral phase of the organogenic precipitate (precursor of brownish-stained DOL-3 mosaic
440 crystals exhibiting bright green fluorescence) is interpreted as having been high-Mg calcite (HMC) since

441 this is typical in a marine setting (Dupraz et al., 2009). Aragonite is excluded since aragonite bioclasts
442 were not dolomitized in this fabric type; they were selectively dissolved leaving mouldic porosity (Figs 8,
443 9). A microcrystalline precursor is assumed because this is the documented size of crystals forming clotted
444 fabric (Riding, 2000; Dupraz et al., 2009). The meteoric diagenetic origin of the observed features is not
445 supported by the measured oxygen isotope data (discussed below in the paragenetic sequence).

447 *7.1.4. Fabric-destructive dolomite (FT4)*

448 The medium crystalline dolomite marks pervasive alteration since no original sedimentary texture has
449 been preserved, except for the reworked lithoclasts. Dolomitization resulted in medium crystals and
450 obliterated all features of the precursors, which is clearly reflected in the fluorescence properties. The
451 turbid crystals or cores of individual crystals (DOL-3), which likely preserve solid relics of the precursor
452 carbonate, were altered via replacement (Land et al., 1975). The overgrowth limpid rim and limpid
453 crystals exhibit similar fluorescence to that of the DOL-4 cement generation within the microbial
454 boundstone; thus, they were also precipitated as cement (Choquette and Hiatt, 2008). Spongy network
455 revealed by the fluorescence pattern in the transitional interval toward the microbial boundstone (Fig. 10B,
456 C) suggests a microbial deposit precursor including organogenic precipitates. Similarly to the bioclastic
457 dolomite the organogenic mineral phase, which was replaced partly by DOL-2 fine crystals and partly by
458 DOL-3 medium-sized crystals, is interpreted as having been HMC.

460 *7.2. Depositional environment of microbial mats and biofilm*

461
462 In the studied samples, the observations imply the following constraints on microbial deposits. A
463 decreasing energy of tidal current across the tidal flat is reflected in the composition and size of trapped
464 sedimentary particles. The overall presence of reworked bioclasts in microbial mat deposits (bioclastic
465 dolomite, FT3) indicates permanent connection to a normal marine subtidal zone. The coarser sediment
466 influx was likely controlled by storms and provides evidence for a higher-energy setting and more
467 frequent inundation. Fine dolomite crystals (DOL-2) in the microbial mat deposits appear to have
468 originated from dolomitized fine marine sediment and partly likely from trapped aeolian dolomite silt. The
469 latter could have been transported from the exposed supratidal zone (e.g. Shinn, 1983). Fine sediment was
470 supplied by tidal currents to the lower intertidal zone (micritic dolomite, FT2 and bioclastic dolomite,
471 FT3) but the mat was less frequently inundated in the upper intertidal zone (microbial boundstone, FT1),
472 which is reflected by the upward decreasing amount of fine crystals. In the supratidal zone, a thin layer of
473 dolomite silt covered the mat surface.

474 The reticulate/nodular laminae (FT1, lower part) resemble the features of a pustular mat, such as the
475 way in which the interior heterogeneity reflects the irregularity of the surface by high porosity and internal
476 sediment infilling (Monty, 1976; Halley, 1976). This mat type is mostly reported from the hypersaline
477 upper intertidal zone (e.g. Allen et al., 2009; Abed et al., 2010). Smooth laminae of dense microcrystals
478 with abundant globules (FT1, upper part) likely record the uppermost intertidal thin biofilm composed
479 predominantly of coccoid cyanobacteria (e.g. Jahnert and Collins, 2013). Therefore, the presence of
480 microcrystalline dolomite (DOL-1) appears not only to be related to periods of ephemeral subaerial
481 exposure but facilitated by a mat and biofilm favouring/tolerating an increasing frequency of
482 mesohaline/hypersaline conditions.

483

484 *7.3. Paragenetic sequence*

485

486 *7.3.1. Relative timing of dolomitization processes*

487 The basic question is whether the wide variety of dolomite in the studied samples was created either by a
488 single dolomitization process and thus, the formation of the distinct dolomite types was controlled by the
489 heterogeneity in the precursor deposits, or by successive processes in various diagenetic environments.
490 The combination of the succession of diagenetic events observed, the isotopic data and the fluid inclusion
491 data indicates two stages of dolomitization (Fig. 13). The measured 62–83 °C homogenisation temperature
492 on DOL-3 and DOL-4 implies their intermediate burial origin (Morrow, 1990; Machel, 2004). Burial
493 dolomite can be expected to have negative $\delta^{18}\text{O}$ values due to precipitation at higher temperature at greater
494 burial depth (e.g. Machel, 2004). The $\delta^{18}\text{O}$ values of the pore-filling DOL-3 and DOL-4 phases (between –
495 4.3‰ and –1.7‰) correspond to the reported values of burial dolomite (e.g. Warren 2000). The bulk rock
496 $\delta^{18}\text{O}$ values of the microbial boundstone, micritic dolomite and bioclastic dolomite (from 0.2‰ to 1.2‰)
497 represent a rather distinct population. The difference between these two groups implies dolomitization by
498 various fluids of different compositions rather than by the same fluid at a different temperature. Moreover,
499 the first-stage dolomite crystal-association (DOL-1 and DOL-2) must be more enriched in heavier isotopes
500 than the solid phase mixture since the latter (=bulk rock of FT1, FT2 and FT3) includes the DOL-3 and
501 DOL-4 phases, too (cf. Banner and Hanson 1990). The estimated range of DOL-1 and DOL-2 together is
502 approximately consistent with precipitates occurring under synsedimentary mesohaline condition (Land,
503 1983; Simms 1984). The completely fabric-destructive dolomite (FT4; with DOL-3 and DOL-4) is of
504 relatively negative value (–1.9‰)—that is comparable with values of pore-filling DOL-3 and DOL-4
505 phases—which implies that the precursor was not affected by synsedimentary dolomitization. The
506 presence of reworked microcrystalline to finely crystalline lithoclasts (DOL-1 and DOL-2 phases) and
507 some fine DOL-2 replacive crystals within FT4 shifted the bulk rock isotope values (–0.1 and 0.3‰) that

508 also suggests two different dolomitization processes regarding the lithoclasts and the host deposits. The
509 more positive values (0.7‰ to 1.6‰) are due to the common presence of replacive DOL-2 crystals within
510 the transitional interval of FT4 below the microbial boundstone lamina (Fig. 10B, C). The measured $\delta^{13}\text{C}$
511 values point to seawater-derived pore-fluids.

512 Mineralisation of microbial biofilms by dolomite (DOL-1) and dolomitization of the associated
513 sediments (DOL-2) were early synsedimentary diagenetic processes (Fig. 14). This is constrained by
514 reworked detrital fragments of lithified microbial boundstone which occur in the overlying deposits
515 represented by bioclastic and fabric-destructive dolomites. The internal sediment, encountered in mat
516 deposits, was composed of dolomite silt and/or micrite sediment replaced by fine crystals (DOL-2). All
517 these observations suggest that synsedimentary dolomitization occurred only in the peritidal caps of
518 shallowing-upward cycles; otherwise, the majority of the deposit consisted of CaCO_3 before it was buried
519 (Fig. 14).

520

521 *7.3.2. Synsedimentary and near-surface processes*

522 The sweeping extinction of elongate crystals (DOL-3 in FT1 and FT3) suggests that their precursor
523 precipitated as a radiaxial fibrous calcite cement (RFC; *sensu* Kendall, 1985). Pendant and fibrous calcite
524 (RFC) precipitated from a marine pore fluid (Frank and Lohmann, 1996) as a first phase cement that was
525 observed in certain intervals of microbial boundstone and bioclastic dolomite in Section 2. Dolomitization
526 of the RFC crystals resulted in fine (DOL-2) and coarser (DOL-3) crystals (Fig. 8B, C D). Fine dolomite
527 (DOL-2), replacing partly the RFC crystals, suggests that dolomitization post-dated the cement
528 precipitation. The dissolution of aragonite bioclasts post-dated the precipitation of RFC but predated the
529 dolomite silt infilling (DOL-2). Thus, synsedimentary dolomitization (DOL-2) and the selective
530 dissolution of aragonite took place penecontemporaneously.

531 The internal sediment, that infiltrated into the mouldic pores left behind after selective dissolution of
532 aragonite, indicates that synsedimentary alteration of the deposits took place while the host sediment was
533 in its original depositional setting. The alteration of sediments by selective/non-selective dissolution,
534 cementation and mineral stabilisation via microspar replacement in a meteoric diagenetic environment is a
535 characteristic feature in many cyclic shallow platform carbonates (e.g. Read and Horbury, 1993). In the
536 studied samples, these features are not considered as indicators of intermittent diagenesis in freshwater
537 because of the lack of remnants of calcrete components (such as joint occurrence of hairline
538 circumgranular cracks and pendant cement, glaeboles with composite inner fabric and pisoids) and the
539 lack of inherited negative carbon isotope value. Theoretically fabric heterogeneity resulting from
540 synsedimentary meteoric alterations might have been preserved during later-stage mimetic dolomitization

541 but the measured oxygen isotope values from the studied samples do not support single dolomitization
542 event.

543

544 *7.3.3. Shallow-burial and intermediate-burial processes*

545 Fluid inclusion microthermometry analyses yielded the similar homogenisation temperature ranges for
546 DOL-3 and DOL-4 crystals. In microbial boundstone, where they are pore-filling phases, the DOL-4
547 phase shows fluorescence zonation and makes up the optical overgrowth on the DOL-3 phase, which
548 shows mottled luminescence. On the bases of crystal habits, the DOL-4 was interpreted as having been
549 precipitated as cement after the DOL-3 replacement of the precursor calcite cement (CAL). The inherited
550 blocky crystal form of the DOL-3 suggests a blocky calcite precursor. Brownish, solid inclusions, which
551 show bright green fluorescence, are exclusively present in the CAL/DOL-3 replaced cement generation.
552 Petrographic characteristics suggest that these inclusions consist of organic matter. Remnants of organic
553 matter were likely dispersed within or among the precursor cement crystals and were preserved during the
554 replacement process. Considering the cyclic depositional setting the ubiquitous pore-filling cement phase
555 is thought to have been precipitated from marine-derived pore-water as calcite (CAL), post-dating the
556 synsedimentary replacive dolomitization (DOL-2), when peritidal mat deposits were submerged during
557 subsequent sea-level cycles. DOL-3 subsequently replaced any remnants of calcium carbonate, which
558 include calcareous deposit and calcite cement (remnants of RFC and CAL). The latest stage of
559 dolomitization was dolomite cement precipitation (DOL-4).

560 An intermediate burial dolomitization (summary in Machel, 2004), with seawater-derived fluids
561 circulated by thermal convection, would have had the capacity to drive DOL-3 and DOL-4 formation.
562 Numerical modelling of open half-cell thermal convection shows that convection can drive dolomitization,
563 mostly at temperatures greater than 50 °C and over time scales of millions to a few tens of millions of
564 years (Wilson et al., 2001; Whitaker et al., 2004; Whitaker and Xiao, 2010). Fluid supply was crucial in an
565 open circulation setting where the half-cell discharged toward the top of the thick, porous platform
566 carbonate, which does not contain any aquitards. Seawater is Mg²⁺-rich and supersaturated with respect to
567 dolomite; thus it is an obvious source for dolomitization (e.g. Purser et al., 1994). However, this model
568 alone cannot explain the origin of a slightly hypersaline pore-fluid. Accordingly, it is presumed that an
569 exotic fluid was channelled through a deeply penetrating, low-angle master fault activated in an
570 extensional tectonic setting (cf. Doglioni, 1992; Bertotti et al., 1993) and injected into the fluid of
571 convection cells.

572

573 *7.4. Discussion on synsedimentary dolomitization: organogenic microcrystals and replacive fine crystals*

574

575 Petrographic comparison of the two dolomite fabric types, which exhibit microbial microfabric, suggests
576 circumstantial evidence constraining that microcrystalline dolomite (DOL-1) did not form via a mimetic
577 replacement process. Microbial boundstone (FT1) includes both syngenic dolomite phases, such as
578 microcrystalline and finely crystalline ones, whereas bioclastic dolomite (FT3) does not involve the
579 microcrystalline phase. In the latter case, the organogenic calcite precursor (HMC) was partly replaced by
580 DOL-2 fine crystals in a peritidal environment and partly by brownish-stained DOL-3 medium-sized
581 mosaic crystals during further burial. The DOL-3 crystals display bright fluorescence and reveal a clotted
582 microbial fabric under blue light. Consequently, in a fabric where the DOL-1 microcrystals are present,
583 this phase precipitated primarily (either in the form of protodolomite or as ordered dolomite; cf. Wenk et
584 al., 1993). Moreover, replacement of organogenic HMC by DOL-2 in micritic dolomite (FT2) and in the
585 transitional interval of fabric-destructive dolomite (FT4; Fig. 10B, C), where spongy fabric occurs, is also
586 very probable. This interpretation is supported by the oxygen isotope data from the transitional interval of
587 fabric-destructive dolomite showing a spongy pattern, where DOL-2 fine crystals co-occur with DOL-3
588 medium-sized crystals (sample 1; Table 1).

589 In the studied sections, the presence of the microcrystalline dolomite is facies-dependent. It appears
590 and becomes abundant upwards within the thin layer-couplets where the fabric exhibits gradual transition
591 either from micritic dolomite (FT2) to microbial boundstone (FT1) or from bioclastic dolomite (FT3) to
592 microbial boundstone (FT1). According to the proposed model (Fig. 14), in the first step, organogenic
593 high-Mg calcite was precipitated in the lower intertidal mat deposits whereas organogenic dolomite
594 progressively took over its place in the buried upper intertidal mat and biofilm in the course of increasing
595 frequency of subaerial exposure and under mesosaline conditions. In the transitional phase, dolomite co-
596 precipitated together with high-Mg calcite. The microcrystallinity of primarily precipitated dolomite
597 indicates rapid and multi-site nucleation on EPS organic substrate (cf. Kandianis et al., 2007; Krause et al.,
598 2012). The results of Roberts et al. (2013) revealed that natural surfaces, including organic matter and
599 microbial biomass, having a high density of carboxyl groups promoted the formation of ordered dolomite
600 nuclei. Under a semi-arid climate, environmental conditions commonly show extreme annual variability in
601 the salinity of the water and the extent of water coverage. Facies reconstruction of the studied formation
602 revealed that the environmental factors likely controlled both the spatial distribution of microbial
603 communities (cf. Yannarell et al., 2006; Paerl and Yannarell, 2010) and the precipitated minerals.

604 In cyclic peritidal succession, as a rule, the syngenic dolomitization correlates with prolonged
605 periods of subaerial exposure during regressions (e.g. Mutti and Simo, 1994). The sabkha model of
606 replacive dolomite formation (examples in Purser et al., 1994; Budd, 1997; Warren, 2000) cannot be
607 applied to the studied formation because no traces of associated evaporite minerals were found. Under
608 prolonged subaerial conditions, the shifting of the saturation state of the pore-fluid with respect to

609 aragonite and HMC is interpreted to have led to replacive dolomite formation. The reaction rate of organic
610 matter degradation controls the carbonate precipitation and dissolution by influencing the pH (Ben-
611 Yaakov, 1973). For example high rates of bacterial sulphate reduction near the surface coincide with
612 microcrystalline carbonate precipitation (Visscher and Stolz, 2005), whereas the low reaction rate in other
613 cases often leads to enhanced carbonates corrosion (e.g. Canfield & Raiswell, 1991). A number of studies
614 (e.g. Pinckney et al., 1995a,b; Abed et al., 2007) documented that rates of microbial processes are greatly
615 reduced under harsh hypersaline conditions, which is consistent for primary production, methanogenesis
616 and sulphate reduction. This supports the interpretation of the studied samples that a facies shift as well as
617 seasonal variation, through affecting the mat community, influenced the saturation state of the pore fluid
618 with respect to various carbonate minerals and, thus, the alteration processes of calcium-carbonate within
619 the mat deposits. Accordingly, the presence of large amounts of organic matter and evaporation in the
620 upper intertidal–lower supratidal zone under a semi-arid climate were the most important controlling
621 factors of both dolomite precipitation (DOL-1) and peritidal replacive dolomitization (DOL-2).

622 Infiltration of detrital dolomite silt into the pore system after the dissolution of the aragonite grains
623 implies high permeability and dynamic pore-fluid circulation. In Section 2, marine RFC cement
624 precipitation also indicates dynamic environmental conditions prior to the syndimentary dolomitization.
625 Tidal pumping across the tidal flat is relatively short-lived but it is regarded as an efficient flow
626 mechanism for dolomitization of surficial intertidal sedimentary veneers (Carballo et al., 1987; Mazullo et
627 al., 1987; Gregg et al., 1992; Teal et al., 2000).

628

629 **8. Conclusions**

630

631 Petrographic analyses revealed a microbial origin for peritidal deposits of the Middle Triassic Budaörs
632 Dolomite. Initial organogenic precipitates within the microbial mat deposits included microcrystalline
633 dolomite as well as microcrystalline high-Mg calcite (HMC). Clusters of microcrystalline dolomite (DOL-
634 1) occur as calcimicrobes, clotted–spherular aggregates and globules in microbial boundstone whereas the
635 microcrystalline HMC was replaced by syndimentary fine crystalline dolomite (DOL-2) and by
636 brownish-stained medium crystalline dolomite (DOL-3) in the bioclastic dolomite. In this latter case, the
637 fluorescence pattern revealed fenestral/desiccation pores and a clotted microfabric. The petrographic
638 features show that these two dolomite fabric types (FT1 and FT3) were formed under different
639 environmental conditions and consequently likely in different mat types, such as microbial boundstone in
640 the upper intertidal setting and bioclastic dolomite in the lower intertidal setting. Estimated oxygen isotope
641 values of syndimentary dolomites (DOL-1 organogenic microcrystals and DOL-2 replacive fine
642 crystals) are consistent with dolomitization under mesohaline conditions, indicating a semi-arid climate.

643 All these data suggest that the diagenetic processes within peritidal deposits were likely controlled by
644 organic matter degradation together with environmental factors.

645 Synsedimentary dolomites were overprinted by a higher temperature (*ca* 65 to 80 °C) dolomitization in
646 intermediate burial depth that resulted in replacive and cement dolomite. This process was not restricted to
647 a particular depositional environment but affected the entire platform carbonate succession. An open half-
648 cell thermal convection system would explain the intermediate burial dolomitization.

649

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651

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661

662 **References**

663

664 Abed, R.M.M., Kohls, K., De Beer, K., 2007. Effect of salinity changes on the bacterial diversity,
665 photosynthesis and oxygen consumption of cyanobacterial mats from an intertidal flat of the Arabian
666 Gulf. *Environmental Microbiology* 9/6, 1384–1392.

667 Abed, R.M.M., Kohls, K., Palinska, K.A., Golubic, S., 2010. Diversity and role of cyanobacteria and
668 aerobic heterotrophic microorganisms in carbon cycling in arid cyanobacterial mats. In: Seckbach, J.,
669 and Oren, A. (Eds), *Microbial Mats, Modern and Ancient Microorganism in Stratified Systems*.
670 Cellular Origin, Life in Extreme Habitats and Astrobiology Series 14, 255–276.

671 Allen, M.A., Goh, F., Burns, B.P., Neilan, B.A., 2009. Bacterial, archaeal and eukaryotic diversity of
672 smooth and pustular microbial mat communities in the hypersaline lagoon of Shark Bay. *Geobiology*
673 7, 82–96.

674 Banner, J.L., Hanson, G.N., 1990. Calculation of simultaneous isotopic and trace element variations
675 during water-rock interaction with applications to carbonate diagenesis. *Geochemica et*
676 *Cosmochimica Acta* 54, 3123–3137.

- 677 Baltzer, F., Kenig, F., Boichard, R., Plaziat, J.C., Purser, B.H., 1994. Organic matter distribution, water
678 circulation and dolomitisation beneath the Abu Dhabi Sabkha (UAE). In: Purser, B., Tucker, M.,
679 Zenger, D. (Eds), Dolomites. IAS Special Publication 21, 409–428.
- 680 Ben-Yaakov, S., 1973. pH buffering of pore water of recent anoxic marine sediments. *Limnology and*
681 *Oceanography* 18, 86–94.
- 682 Bertotti, G., Picotti, V., Bernoulli, D., Castellarin, A., 1993. From rifting to drifting: tectonic evolution of
683 the South-Alpine upper crust from the Triassic to the Early Cretaceous. *Sedimentary Geology* 86,
684 53–76.
- 685 Bertrand, P., Piton, J-L., Bernaud, C., 1986. Fluorescence of sedimentary organic matter in relation to its
686 chemical composition. *Organic Geochemistry* 10, 641–647.
- 687 Bontognali, T.R.R., Vasconcelos, C., Warthmann, R.J., Bernasconi, S.M., Dupraz, C., Strohmenger, C.J.,
688 McKenzie, J.A., 2010. Dolomite formation within microbial mats in the coastal sabkha of Abu
689 Dhabi (United Arab Emirates). *Sedimentology* 57, 824–844.
- 690 Braissant, O., Decho, A.W., Dupraz, C., Glunk, C., Przekop, K.M., Visscher, P.T., 2007. Exopolymeric
691 substances of sulfate-reducing bacteria: interactions with calcium at alkaline pH and implication for
692 formation of carbonate minerals. *Geobiology* 5, 401–411.
- 693 Budai, T., 2004. Middle Triassic basin facies and volcanites in the Zsámbék basin, Transdanubian Range,
694 Hungary. *MÁFI Évi Jelentés 2002-ről* 189–194 (in Hungarian with English summary).
- 695 Budd, D.A., 1997. Cenozoic dolomites of carbonate islands: their attributes and origin. *Earth-Science*
696 *Review* 42, 1–47.
- 697 Canfield, D.E., Raiswell, R., 1991. Carbonate precipitation and dissolution. Its relevance to fossil
698 preservation. In: Alison, P.A., Briggs, D.E.G. (Eds), *Taphonomy. Releasing the Data Locked in the*
699 *Fossil Record*. Plenum Press, New York, 411–453.
- 700 Carballo, J.D., Land, L.S., Miser, D.E., 1987. Holocene dolomitization of supratidal sediments by active
701 tidal pumping, Sugarloaf Key, Florida. *Journal of Sedimentary Petrology* 57, 153–165.
- 702 Choquette, P.W., Hiatt, E.E., 2008. Shallow-burial dolomite cement: a major component of many ancient
703 sucrosic dolomites. *Sedimentology* 55, 423–460.
- 704 Couradeau, E., Benzerara, K., Gérard, E., Estève, I., Moreira, D., Tavera, R., López-García, P., 2013.
705 Cyanobacterial calcification in modern microbialites at the submicrometer scale. *Biogeosciences*
706 10, 5255–5266.
- 707 Csontos, L., Vörös, A., 2004. Mesozoic plate tectonic reconstruction of the Carpathian region.
708 *Palaeogeography Palaeoclimatology Palaeoecology* 210, 1–56.
- 709 Decho, A.W., Visscher, P.T., Reid, P., 2005. Production and cycling of natural microbial exopolymers
710 (EPS) within a marine stromatolite. *Palaeogeography Palaeoclimatology Palaeoecology* 219, 71–86.

- 711 Défarge, C., Trichet, J., Jaunet, A-M., Robert, M., Tribble, J., Sansone, F.J., 1996. Texture of microbial
712 sediments revealed by cryo-scanning electron microscopy. *Journal of Sedimentary Research* 66/5,
713 935–947.
- 714 Dickson, J.A.D., 1966. Carbonate identification and genesis as revealed by staining. *Journal of*
715 *Sedimentary Petrology* 36, 491–505.
- 716 Doglioni, C., 1992. Relationships between Mesozoic extensional tectonics, stratigraphy and Alpine
717 inversion in the Southern Alps. *Eclogae Geologicae Helvetiae* 85/1, 105–126.
- 718 Dravies, J.J., Yurewicz, D.A., 1985. Enhanced carbonate petrography using fluorescence microscopy.
719 *Journal of Sedimentary Petrology* 55, 795–804.
- 720 Dupraz, C., Vischer, P.T., Baumgartner, L.K., Reid, P., 2004. Microbe–mineral interactions: early
721 carbonate precipitation in a hypersaline lake (Eleuthera Island, Bahamas). *Sedimentology* 51, 745–
722 765.
- 723 Dupraz, C., Reid, P.R., Braissant, O., Decho, A., Norman, R.S., Visscher, P.T., 2009. Processes of
724 carbonate precipitation in modern microbial mats. *Earth-Science Review* 96, 141–162.
- 725 Fodor, L., Magyari, Á., Fogaras, A., Palotás, K., 1994. Tertiary tectonics and Late Paleogene
726 sedimentation in the Buda Hills, Hungary. A new interpretation of the Buda Line. *Földtani Közlöny*
727 124/2, 129–305.
- 728 Frank, T.D. and Lohmann, K.C., 1996. Diagenesis of fibrous magnesian calcite marine cement:
729 implications for the interpretation of $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ values of ancient equivalents. *Geochimica et*
730 *Cosmochimica Acta* 60/13, 2427–2436.
- 731 Gaft, M., Reisfeld, R., Panczer, G., 2005. *Modern Luminescence Spectroscopy of Minerals and Materials*.
732 Springer, Berlin Heidelberg.
- 733 Gallagher, K.L., Kading, T.J., Braissant, O., Dupraz, C., Visscher, P.T., 2012. Inside the alkalinity engine:
734 the role of electron donors in the organomineralization potential of sulphate-reducing bacteria.
735 *Geobiology* 10/6, 518–530.
- 736 Gautret, P., Camoin, G., Golubic, S., Sprachta, S., 2004. Biochemical Control of Calcium Carbonate
737 Precipitation in Modern Lagoonal Microbialites, Tikehau Atoll, French Polynesia. *Journal of*
738 *Sedimentary Research* 74/4, 462–478.
- 739 Goldstein, R.H., Reynolds, T.J., 1994. Systematics of Fluid Inclusions in Diagenetic Minerals. *SEPM*
740 *Short Course* 31.
- 741 Golubic, S., Abed, R.M.M., 2010. Entophysalis mats as environmental regulators. In: Seckbach, J., Oren,
742 A. (Eds), *Microbial Mats, Modern and Ancient Microorganism in Stratified Systems Cellular*
743 *Orogin. Life in Extreme Habitats and Astrobiology Series* 14, 239–254.

- 744 Gregg, J.M., Howard, S.A., Mazzulo, S.J., 1992. Early diagenetic recrystallization of Holocene (<3000
745 years old) peritidal dolomites, Ambergris Cay, Belize. *Sedimentology* 39, 143–160.
- 746 Győri, O., Poros, Zs., Mindszenty, A., Molnár, F., Fodor, L., Szabó, R., 2011. Diagenetic history of the
747 Palaeogene carbonates, Buda Hills, Hungary. *Földtani Közlöny* 141/4, 341–361 (in Hungarian with
748 English summary).
- 749 Haas, J., 2002. Origin and evolution of Late Triassic backplatform and intraplateform basins in the
750 Transdanubian Range, Hungary. *Geologica Carpathica* 53/3, 159–178.
- 751 Haas, J., Budai, T., 1995. Upper Permian-Triassic facies zones in the Transdanubian Range. *Rivista*
752 *Italiana di Paleontologia e Stratigrafia* 101/3, 249–266.
- 753 Haas, J., Budai, T., 1999. Triassic sequence stratigraphy of the Transdanubian Range, Hungary. *Geologica*
754 *Carpathica* 50/6, 459–475.
- 755 Haas, J. and Budai, T. 2004. Dunántúli –középhegységi egység. In: Haas, J. (Ed.), Magyarország
756 geológiája, triász, ELTE Eötvös Kiadó, Budapest, 25–124.
- 757 Halley, R.B., 1976. Textural variation within Great Salt Lake algal mounds. In: Walter, M.R. (Ed.),
758 *Stromatolites*. Elsevier, Amsterdam, 436–445.
- 759 Jahnert, R.J., Collins, L.B., 2013. Controls on microbial activity and tidal flat evolution in Shark Bay,
760 Western Australia. *Sedimentology* 60/4, 1071–1099.
- 761 Kandianis, M.T., Fouke, B.W., Johnson, R.W., Veysey II, J., Inskeep, W.P., 2007. Microbial biomass: A
762 catalyst for CaCO₃ precipitation in advection-dominated transport regimes. *Bulletin of Geological*
763 *Society of America* 120/3–4, 442–450.
- 764 Kendall, A.C., 1985. Radial-fibrous calcite: a reappraisal. In: Schneidermann, N., Harris, P.M. (Eds),
765 *Carbonate Cements*. SEPM Special Publication 36, 59–77.
- 766 Krause, S., Liebetrau, V., Gorb, S., Sánchez-Román, M., McKenzie, J.A., Treude, T., 2012. Microbial
767 nucleation of Mg-rich dolomite in exopolymeric substances under anoxic modern seawater salinity:
768 New insight into an old enigma. *Geology* 40, 587–590.
- 769 Krumbein, W.C., Garrels, R.M., 1952. Origin and classification of chemical sediments in terms of pH and
770 oxidation-reduction potentials. *Journal of Geology* 60/1, 1–33.
- 771 Kutassy, E., 1927. Beiträge zur Stratigraphie und Paläontologie der Alpenen Triassschichten in der
772 Umgebung von Budapest. *Jahrbuch Königlichen Ungarischen Geologischen Anstalt* 27/2, 105–175.
- 773 Land, L.S., 1983. The application of stable isotopes to studies of the origin of dolomite and to problems of
774 diagenesis of clastic sediments. In: Arthur, M.A., Anderson, T.F., Kaplan, I.R., Veizer, J., Land,
775 L.S. (Eds), *Stable Isotopes in Sedimentary Geology*. Society of Sedimentary Geology, Short Course
776 10, 4.1–4.22.
- 777 Land, L.S., 1985. The origin of massive dolomite. *Journal of Geological Education* 33, 112–125.

- 778 Land, L.S., Salem, M.R.I., Morrow, D.W., 1975. Paleohydrology of ancient dolomites: geochemical
779 evidence. *AAPG Bulletin* 59, 1602–1625.
- 780 Machel, H.G., 2004. Concepts and models of dolomitization: a critical reappraisal. In: Braithwaite, C.J.R.,
781 Rizzi, G., Darke, G. (Eds), *The Geometry and Petrogenesis of Dolomite Hydrocarbon Reservoirs*.
782 Geological Society of London, Special Publication 235, 7–63.
- 783 Magyari, Á., 1994. Late Eocene hydraulic rebrecciation in the Southern Buda Mountains, Hungary.
784 *Földtani Közlöny* 124/1, 89–107 (in Hungarian with English summary).
- 785 Magyari, Á., 1999. Törökugrató Hill: Late Eocene positive flower structure on the southwestern part of
786 the Buda Mountains, Budapest. *Földtani Közlöny* 128/4, 555–572 (in Hungarian with English
787 summary).
- 788 Márton, E., Fodor, L., 2003. Tertiary paleomagnetic results and structural analyses from the
789 Transdanubian Range (Hungary): rotational disintegration of the ALCAPA unit. *Tectonophysics*
790 363, 201–224.
- 791 Masetti, D., Neri, C., 1980. L'Anisico della Val di Fassa (Dolomiti occidentali): sedimentologia e
792 paleografia. *Annales University of Ferrara* 7/1, 1–19.
- 793 Mastandrea, A., Perri, E., Russo, F., Spadafora, A., Tucker, M.E., 2006. Microbial primary dolomite from
794 a Norian carbonate platform, northern Calabria, southern Italy. *Sedimentology* 53, 465–480.
- 795 Mazullo, S.J., 2000. Organogenic dolomitization in peritidal to deep-sea sediments. *Journal of*
796 *Sedimentary Research* 70/1, 10–23.
- 797 Mazullo, S.J., Reid, A.M., Gregg, J.M., 1987. Dolomitization of Holocene Mg-calcite supratidal deposits,
798 Ambergris Cay, Belize. *Geological Society of America Bulletin* 98, 224–231.
- 799 McKenzie, J.A., 1981. Holocene dolomitization of calcium carbonate sediments from the coastal sabkhas
800 of Abu Dhabi, U.A.E.: A stable isotope study. *Journal of Geology* 89, 185–198.
- 801 Monty, C.L.V., 1976. The origin and development of cryptalgal fabrics. In: Walter, M.R., (Ed.),
802 *Stromatolites*. Elsevier, Amsterdam, 193–249.
- 803 Morrow, D.W., 1990. Dolomite – Part 2: Dolomitization models and ancient dolostones. In: McIlreath,
804 I.A., Morrow, D.W. (Eds), *Diagenesis*. Geoscience Canada, Reprint Series 4, 125–139.
- 805 Mutti, M., Simo, J.A. 1994. Distribution, petrography and geochemistry of early dolomite in cyclic shelf
806 facies, Yates Formation (Guadalupian), Capitanian Reef Complex, USA. In: Purser, B.H., Tucker,
807 M.E., Zenger, D.H. (Eds), *Dolomites, a Volume in Honour of Dolomieu*. IAS Special Publication
808 21, 91–107.
- 809 Paerl, H.W., Yannarell, A.C., 2010. Environmental dynamics, community structure and function in a
810 hypersaline microbial mat. In: Seckbach, J., Oren, A. (Eds), *Microbial Mats, Modern and Ancient*

- 811 Microorganism in Stratified Systems Cellular Origin. Life in Extreme Habitats and Astrobiology
812 Series 14, 423–444.
- 813 Perri, E., Tucker, M., 2007. Bacterial fossils and microbial dolomite in Triassic stromatolite. *Geology*
814 35/3, 207–210.
- 815 Pinckney, J., Paerl, H.W., Bebout, B.M., 1995a. Salinity control of benthic microbial mat community
816 production in a Bahamian hypersaline lagoon. *Journal of Experimental Marine Biology and*
817 *Ecology* 187, 223–237.
- 818 Pinckney, J., Paerl, H.W., Fitzpatrick, M., 1995b. Impacts of seasonality and nutrients on microbial mat
819 community structure and function. *Marine Ecology Progress Series* 123, 207–216.
- 820 Poros, Zs., 2011. Fluid migration and porosity evolution in the Buda Hills, Hungary – selected examples
821 from Triassic and Paleogene carbonate rocks. PhD dissertation, Eötvös University, Budapest.
- 822 Poros, Zs., Mindszenty A., Molnár, F., Pironon, J., Györi, O., Ronchi, P., Szekeres, Z., 2012. Imprints of
823 hydrocarbon-bearing basinal fluids on a karst system: mineralogical and fluid inclusion studies from
824 the Buda Hills, Hungary. *International Journal of Earth Sciences* 101, 429–452.
- 825 Pratt, B.R., 1984. *Epiphyton* and *Renalcis*—diagenetic microfossils from calcification of coccoid blue-
826 green algae. *Journal of Sedimentary Petrology* 54/3, 948–971.
- 827 Purser, B.H., Tucker, M.E., Zenger, D.H., 1994. Summary. In: Purser, B.H., Tucker, M.E., Zenger, D.H.
828 (Eds), *Dolomites, a Volume in Honour of Dolomieu*. IAS Special Publication 21, 29–33.
- 829 Read, J.F., Horbury, A.D., 1993. Eustatic and tectonic controls on porosity evolution beneath sequence-
830 bounding unconformities and parasequence disconformities on carbonate platforms. In: Horbury,
831 A.D., Robinson, A.G. (Eds), *Diagenesis and Basin Development*. AAPG Studies in Geology, Tulsa,
832 Oklahoma 36, 155–197.
- 833 Riding, R., 1991. Calcified Cyanobacteria. In: Riding, R. (Ed.), *Calcareous Algae and Stomatolites*.
834 Springer, Berlin, 55–87.
- 835 Riding, R., 2000. Microbial carbonates: the geological records of calcified bacterial–algal mats and
836 biofilms. *Sedimentology* 47/Suppl 1, 179–214.
- 837 Roberts, J.A., Kenward, P.A., Fowle, D.A., Goldstein, R.H., González, L.A., Moore, D.S., 2013. Surface
838 chemistry allows for abiotic precipitation of dolomite at low temperature. *Proceedings of the national*
839 *Academy of Sciences of the USA* 110/36, 14540–14545.
- 840 Rosenbaum, J., Sheppard, S.M.F., 1986. An isotopic study of siderites, dolomites and ankerites at high
841 temperatures. *Geochemica et Cosmochimica Acta* 50, 1147–1150.
- 842 Ruffer, T., Zühlke, R., 1995. Sequence stratigraphy and sea-level change in the Early to Middle Triassic of
843 the Alps: a global comparison. In: Haq, B.U. (Ed.), *Sequence Stratigraphy and Depositional*

- 844 Response to Eustatic, Tectonic and Climatic Forcing. Kluwer Academic Publishers, the Netherlands,
845 161–207.
- 846 Shinn, E., 1983. Tidal flat environment. In: Scholle, P.A., Bebout, D.G., Moore, C.H. (Eds), Carbonate
847 Depositional Environments. AAPG Memoir 33 171–210.
- 848 Simms, M.A., 1984. Dolomitization by groundwater flow systems in carbonate platforms. Transactions of
849 the Gulf Coast Association of Geological Sciences 24, 411–420.
- 850 Spadafora, A., Perri, E., McKenzie, J.A., Vasconcelos, C., 2010. Microbial biomineralization processes
851 forming modern Ca:Mg carbonate stromatolites. *Sedimentology* 57, 27–40.
- 852 Spötl, C., Vennemann, T.W., 2003. Continuous-flow isotope ratio mass spectrometric analysis of
853 carbonate minerals. *Rapid Communication in Mass Spectrometry* 17, 1004–1006.
- 854 Stephens, N.P., Sumner, D.Y., 2002. Renalcids as fossilized biofilm clusters. *Palaios* 17, 225–236.
- 855 Teal, C.S., Mazzulo, S.J., Bischoff, W.D., 2000. Dolomitization of Holocene shallow-marine deposits
856 mediated by sulphate reduction and methanogenesis in normal-salinity seawater, Northern Belize.
857 *Journal of Sedimentary Research* 70/3, 649–663.
- 858 Tucker, M.E., Wright, V.P., 1990. *Carbonate Sedimentology*. Blackwell Science, Oxford.
- 859 van Lith, Y., Warthmann, R., Vasconcelos, C., McKenzie, J.A., 2003. Sulphate-reducing bacteria induce
860 low-temperature Ca-dolomite and high Mg-calcite formation. *Geobiology* 1, 71–79.
- 861 Vasconcelos, C., McKenzie, J.A., Bernasconi, S., Grujic, D., Tiens, A.J., 1995. Microbial mediation as a
862 possible mechanism for natural dolomite formation at low temperatures. *Nature* 377, 220–222.
- 863 Visscher, P.Z., Stolz, J.F., 2005. Microbial mats as bioreactors: populations, processes, products.
864 *Palaeogeography Palaeoclimatology Palaeoecology* 219, 87–100.
- 865 Visscher, P.T., Reid, R.P., Bebout, B.M., 2000. Microscale observations of sulfate reduction: correlation
866 of microbial activity with lithified micritic laminae in modern marine stromatolites. *Geology* 28,
867 919–922.
- 868 Warren, J., 2000. Dolomite: occurrence, evolution and economically important associations. *Earth-*
869 *Science Review* 52, 1–81.
- 870 Wein, Gy., 1977. *A Budai-hegység tektonikája (Tectonics of the Buda Hills)*. Hungarian Geological
871 Institute, Special Publication Budapest (in Hungarian).
- 872 Wenk, H.R., Hu, M., Frisia, S., 1993. Partially disordered dolomite: microstructural characterization of
873 Abu Dhabi sabkha carbonates. *American Mineralogist* 78/7–8, 769–774.
- 874 Whitaker, F.F., Xiao, Y., 2010. Reactive transport modelling of early burial dolomitization of carbonate
875 platforms by geothermal convection. *AAPG Bulletin* 94, 889–917.

- 876 Whitaker, F.F., Smart, P.L., Jones, G.D., 2004. Dolomitization: from conceptual to numerical models. In:
 877 Braithwaite, C.J.R., Rizzi, G., Darke, G. (Eds), *The Geometry and Petrogenesis of Dolomite*
 878 *Hydrocarbon Reservoirs*. Geological Society of London, Special Publication 235, 99–139.
- 879 Wilson, E.N., Hardie, L.A., Phillips, O.M., 1990. Dolomitization front geometry, fluid flow patterns, and
 880 the origin of massive dolomite: the Triassic Latemar buildup, northern Italy. *American Journal of*
 881 *Science* 290, 741–796.
- 882 Wilson, A.M., Sanford, W.E., Whitaker, F.F., Smart, P.L., 2001. Spatial patterns of diagenesis during
 883 geothermal circulation in carbonate platforms. *American Journal of Science* 301, 727–752.
- 884 Wright, D.T., 1999. The role of sulphate-reducing bacteria and cyanobacteria in dolomite formation in
 885 distal ephemeral lakes of the Coorong region, South Australia. *Sedimentary Geology* 126, 147–157.
- 886 Wright, D.T., Wacey, D., 2004. Sedimentary dolomite: a reality check. In: Braithwaite, C.J.R., Rizzi, G.,
 887 Darke, G. (Eds), *The Geometry and Petrogenesis of Dolomite Hydrocarbon Reservoirs*. Geological
 888 Society of London, Special Publication 235, 65–74.
- 889 Wright, D.T., Wacey, D., 2005. Precipitation of dolomite using sulfate-reducing bacteria from the
 890 Coorong Region, South Australia: Significance and implication. *Sedimentology* 52, 987–1008.
- 891 Yannarell, A.C., Steppe, T.F., Paerl, H.W., 2006. Genetic variance in the composition of two functional
 892 groups (diazotrophs and cyanobacteria) from a hypersaline microbial mat. *Applied and*
 893 *Environmental Microbiology* 72/2, 1207–1217.
- 894 You, X., Sun, S., Zhu, J., Li, Q., Hu, W., Dong, H., 2013. Microbially mediated dolomite in Cambrian
 895 stromatolites from the Tarim Basin, north-west China: implications for the role of organic substrate
 896 on dolomite precipitation. *Terra Nova* 25/5, 387–395.

897

898 **Figure captions**

899 **Fig. 1. A)** Locations of the two studied sections (map by Haas, 2002). **B)** Pre-Quaternary geologic map of
 900 the westernmost part of Buda Hills with the location of Section 1 (Fodor unpublished map 2000, modified
 901 after Wein, 1977). Inset map showing Europe and Hungary with the location of map A. TransD.R.=
 902 Transdanubian Range. Budapest and Budaörs are cities (grey).

903

904 **Fig. 2. A–B)** Panoramic view showing the position of the two studied sections and the logs of the sections.
 905 C) A typical exposure of Section 1, showing the alternation of lithofacies. The thin-bedded/laminated
 906 lithofacies (lower two-thirds) is overlain by the massive lithofacies (upper one-third).

907

908

909

910 **Fig. 3.** A typical micritic dolomite and microbial boundstone couplet of thin bedded/laminated lithofacies,
 911 as exposed in Section 1, showing diagnostic features indicative of a peritidal environment (sample 4).

912
 913 **Fig. 4.** Photomicrographs of microfabrics in the microbial boundstone. **A)** Framework structure,
 914 composed of dense microcrystals (DOL-1), with bushy clot clusters (1). Dense micrite involves abundant
 915 uniform, oval-shaped globules (scattered light dots, 2). The pore network is filled with coarser cement
 916 crystals (lighter areas, 3). **B)** Higher magnification of densely-packed oval-shaped globules, interpreted as
 917 ghosts of coccoid sheaths (red arrows). **C)** Clot clusters (1) and occluded various-sized
 918 fenestral/desiccation pores (2). Brownish-coloured pendant cement (3) and pore-occluding fibrous cement
 919 (4) are replaced by predominantly DOL-2 fine crystals and subordinately slightly brownish-coloured
 920 DOL-3 medium-sized crystals. **D)** Fluorescence image of the field of view shown in C. DOL-1
 921 microcrystals are bright green; the replaced pendant cement is dull green with growth zones; the fibrous
 922 cement replaced by DOL-2 and DOL-3 exhibits very faint green mottles along growth bands in the non-
 923 fluorescent groundmass. A–B: sample 6; C–D: sample 12.

924
 925 **Fig. 5.** Photomicrographs of the microstructure and dolomite phases in the microbial boundstone. **A)** A
 926 typical vertical pattern in the microstructure: the lower, thicker lamina is developed from the underlying
 927 micritic dolomite (FT2) and gradually evolves into the upper lamina consisting of dense microcrystals
 928 (DOL-1). The various-sized amalgamated nodules (1) are composed of microcrystalline aggregates (DOL-
 929 1) and fine crystals (DOL-2). Large, quasi layer-parallel pores (2) and fractures (3) are occluded by DOL-
 930 3 and DOL-4 phases. **B)** Distribution of crystal phases within the host deposits (dark grey areas on top and
 931 bottom) and in a layer-parallel pore (middle). **C)** Fluorescence image of the field of view shown in C.
 932 DOL-1 is bright green; DOL-2 is dull green; internal sediment 1 is dull; CAL/DOL-3 has bright spots in a
 933 non-fluorescent groundmass; DOL-4 has two growth bands that are dull with fine subzones and non-
 934 fluorescent; internal sediment 2 is blotchy. A: sample 4; B–C: sample 6.

935
 936 **Fig. 6.** Photomicrographs of the microbial boundstone showing the CL pattern of the crystal phases. **A)**
 937 Nodular/reticulate lamina with DOL-1 microcrystals (white arrow), a mixture of DOL-1 microcrystals and
 938 DOL-2 fine crystals (yellow arrow) and DOL-3 medium-sized crystals (red arrow). **B)** CL image of the
 939 field of view shown in A. Dull red microcrystalline clotted groundmass (white arrow) involves nodules
 940 exhibiting either less intense dull red luminescence (yellow arrow) or brighter spots in a non-luminescent
 941 background (red arrow). **C)** Microbial boundstone (1) and pore-filling crystal phases, such as inclusion-
 942 rich turbid DOL-3 coarse crystals (2) and limpid DOL-4 cement overgrowth (3). Surface of the DOL-4
 943 crystals are serrated, corroded and occasionally covered by calcite microspars (4). Corrosion and calcite

944 precipitation was a result of recent meteoric alteration of the dolomite rocks. **D)** CL image of the field of
 945 view shown in C. Crystal phases have specific CL patterns, such as: DOL-1 microcrystals are dull red (1),
 946 DOL-3 crystals are mottled (2) and DOL-4 crystals have growth zones: non-luminescent and dull red (3).
 947 Blotchy pattern characterises the corroded area of the DOL-4 crystals. Sample 4.

948
 949 **Fig. 7.** Photomicrographs of the micritic dolomite. **A)** Tufted filamentous calcimicrobe (red arrow) and
 950 microcrystalline lithoclasts with a sharp outline (yellow arrow) besides abundant DOL-2 fine crystals. **B)**
 951 Fluorescence image of the field of view shown in A. Dull green groundmass of DOL-2 fine crystals
 952 involves non-fluorescent and brighter mottles as well as bright green components consisting of DOL-1.
 953 Pores (arrows) are filled by cement exhibiting two growth bands: mottled (scattered, brighter green dots in
 954 a non-fluorescent background) and dull green–non-fluorescent. **C)** Mixture of crystal phases. Groundmass
 955 of DOL-2 fine crystals (1) includes microcrystalline clot clusters (DOL-1; 2), microcrystalline lithoclasts
 956 (DOL-1; 3) and DOL-3 medium-sized crystals (4). **D)** Fluorescence image of the field of view shown in
 957 C. Bright microcrystalline clot clusters (top) have a gradual transition from the underlying dull green
 958 finely crystalline spongy fabric. The fabric additionally consists of a few lithoclasts, having a sharp
 959 boundary (yellow arrow), and pores are filled by mottled cement (red arrow). Sample 2.

960
 961 **Fig. 8.** Photomicrographs of the bioclastic dolomite with ghosts of dasycladalean algae. **A)** Bands of
 962 limpid crystals and/or dolomite silt fill biomoulds (yellow arrow). Brownish-coloured DOL-3 mosaic
 963 crystals delineate the moulds (white arrow) and form patches (red arrow) in the finely crystalline
 964 groundmass (DOL-2). **B)** Dolomite silt (DOL-2) and less inclusion-rich crystals (DOL-3 and DOL-4) fill
 965 the biomoulds (yellow arrow), dark brown DOL-3 mosaic crystals delineate dasycladalean alga fragments
 966 (white arrows) and dark brown DOL-3 mosaic crystals form patches (red arrow). Between them, brown
 967 elongate crystals occur (blue arrow) which are surrounded by fine DOL-2 crystals. **C)** A dasycladalean
 968 alga biomould and a primary intraparticle pore with dolomite phases. Dark brown DOL-3 mosaics cover
 969 the bioclast surface both at the inner (white arrow, bottom left) and outer side (white arrow, top right);
 970 lighter brown elongate DOL-3 crystals occur in the primary intraparticle pore (blue arrows); grey fine
 971 DOL-2 crystals are dispersed among both types of DOL-3 crystals; the biomould pore is filled by
 972 dolomite silt (yellow arrow on top) and less turbid DOL-3 and DOL-4 crystals (yellow arrows on left).—
 973 Arrangement of brownish elongate crystals (RFC replaced by DOL-3) in the primary intraparticle pore
 974 space indicates that they composed the first pore-filling cement phase. Fine crystals (DOL-2) partially
 975 replaced the RFC and presumably the HMC crystals as well. **D)** Fluorescence image of the field of view
 976 shown in C. Dark brownish-stained turbid DOL-3 mosaic crystals (which replaced the bioclast-covering
 977 HMC) are bright green revealing the clot-clustered microfabric of the precursor carbonate (white arrows).

978 Micrite and elongate crystals (RFC replaced by DOL-2 and DOL-3, respectively) are mottled bright–dull
 979 (blue arrows); cement in biomould has two growth bands: CAL replaced by turbid DOL-3 is mottled and
 980 limpid DOL-4 overgrowth is non-fluorescent (yellow arrows on left); DOL-2 fine crystals, e.g. dolomite
 981 silt (yellow arrow on top), are dull green. Sample 15.

982
 983 **Fig. 9.** Photomicrograph of the transitional interval between the underlying bioclastic dolomite (FT3) and
 984 the overlying microbial boundstone (FT1) where DOL-1 microcrystalline clot clusters appear in the fabric.
 985 **A)** Microcrystalline DOL-1 (black) together with dark brown DOL-3 mosaic crystals (replaced HMC
 986 precursor) form patches (red arrows) and cover the bioclasts (white arrows). Pore network is occluded by
 987 crystal silt (DOL-2; grey) and/or two generations of cement (turbid CAL/DOL-3 and limpid DOL-4;
 988 yellow arrow). **B)** Fluorescence image of the field of view shown in A. Both DOL-1 and dark brown
 989 DOL-3 mosaic crystals show bright green fluorescence; DOL-2 fine crystals are dull green; and in the
 990 pores, turbid DOL-3 crystals are mottled and limpid DOL-4 cement crystals are non-fluorescent. Sample
 991 13.

992
 993 **Fig. 10.** Photomicrographs of fabric-destructive dolomite in Section 1. **A)** Finely to medium crystalline
 994 dolomite of massive lithofacies with lithoclast of microcrystalline to finely crystalline dolomite, having a
 995 sharp outline (yellow arrow). Many reworked lithoclasts are typified by microcrystalline clot-clusters (red
 996 arrow). **B)** Transitional interval (TR) between the underlying fabric-destructive dolomite and the overlying
 997 microbial boundstone. In the fabric, DOL-1 microcrystalline clot clusters (black; white arrow) appear
 998 among the DOL-2 fine (yellow arrow) and DOL-3 medium-sized (red arrow) crystals and thus, the
 999 nodular/reticulate (NOD) microfabric becomes more obvious upwards. **C)** Fluorescence image of the field
 1000 of view shown in B. The spongy fabric of precursor deposits is visible due to the different fluorescence of
 1001 the components. A: sample 10; B–C: sample 1.

1002
 1003 **Fig. 11.** Stable carbon and oxygen isotope data for whole rock fabrics (FT1, FT2, FT3 and FT4) and for
 1004 pore-filling DOL-3, which replaced the CAL cement, and DOL-4 phases. Because of the small-scale
 1005 heterogeneity of the studied rocks, the calculated mean values (larger dots) of the multiple analyses (small
 1006 dots) of one sample are shown for the case where deviation of the oxygen isotope is larger than $\pm 0.15\%$.

1007
 1008 **Fig. 12.** Fluid inclusion data from Section 1. **A)** An example of primary, two-phase (liquid–vapour)
 1009 aqueous inclusion (arrow) hosted by the turbid DOL-3 phase which in turn replaced CAL cement; sample
 1010 6). **B)** Homogenisation temperatures (T_h) measured from DOL-3 and DOL-4 phases (samples 6 and 7).

1011

1012 **Fig. 13.** Paragenetic sequence showing the successive diagenetic events that occurred up to the completion
1013 of dolomitization.

1014
1015 **Fig. 14.** Diagenetic alteration stages of the fabric of the studied cyclic deposits from deposition to
1016 intermediate burial realm up to the completion of dolomitization.

1017

1018 **Table 1** Stable isotope values (V-PDB)

1019 *Calculated mean values where deviation is larger than $\pm 0.15\%$.

1020 Sample 3, fabric-destructive dolomite: -2.2 and -1.6% .

1021 Sample 9, pore-filling DOL-3 and DOL-4 phases: -2.6 and -2.0% .

1022 Sample 2, pore-filling DOL-3 and DOL-4 phases: -3.3 and -2.0% .

1023

1024 **Table 2** Fluid inclusion homogenisation temperature values

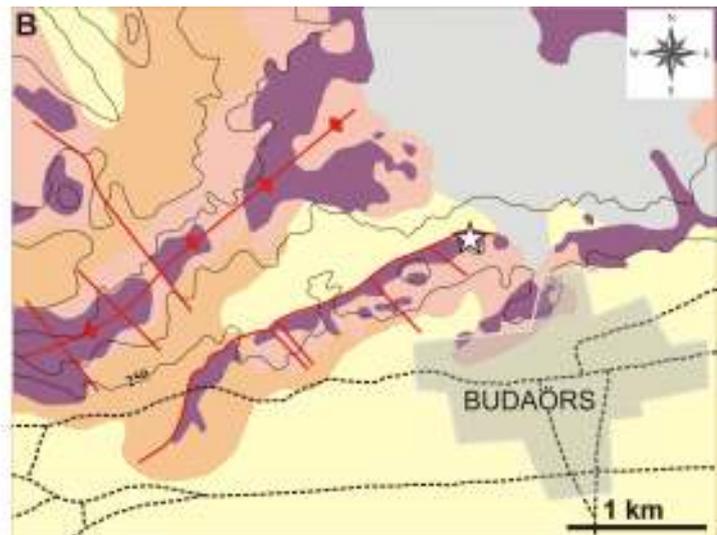
1025



■ surface outcrops of Triassic formations

★ Section 1: N47°28.102 E18°56.909

★ Section 2: N47°32.583 E19°42.055



■ Miocene pebble, sand, freshwater limestone

■ Oligocene clay

■ Eocene–Oligocene marl

■ Eocene conglomerate

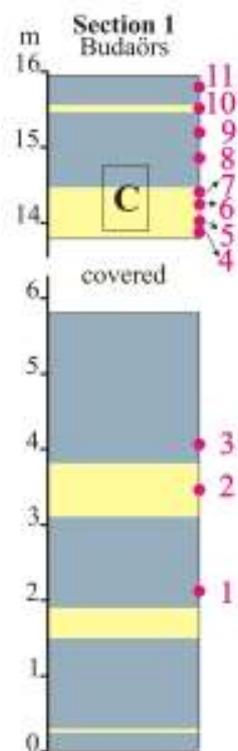
■ Triassic dolomite

— faults

◆ anticline structure

☆ location of Section 1

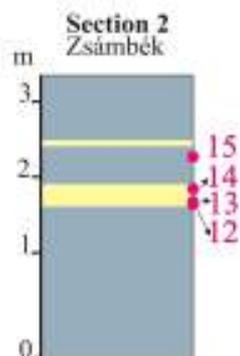
- - - roads



A view of the Odvas Hill at Budaörs from south



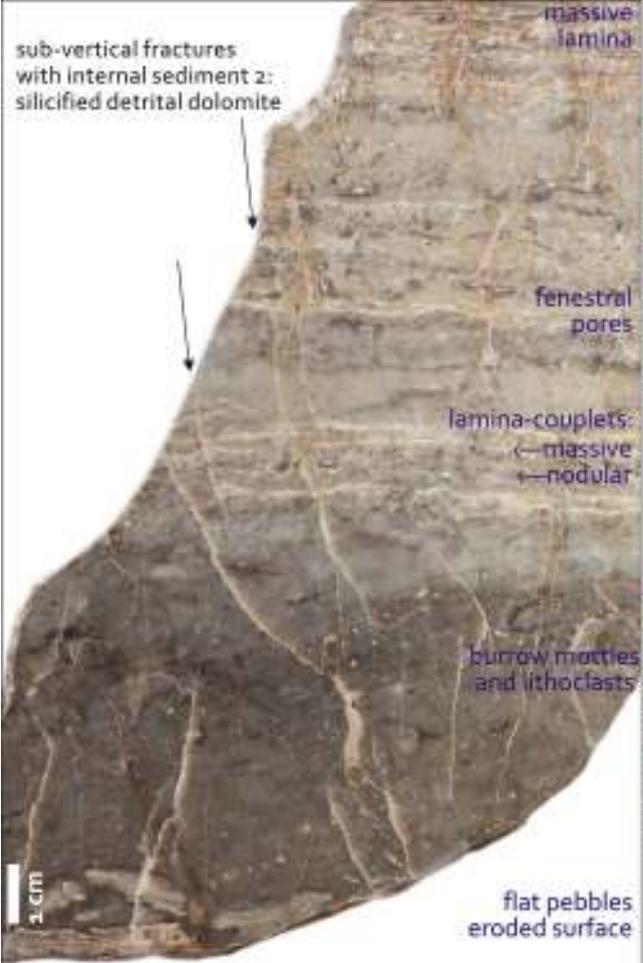
B view of the Zsámbék quarry from west

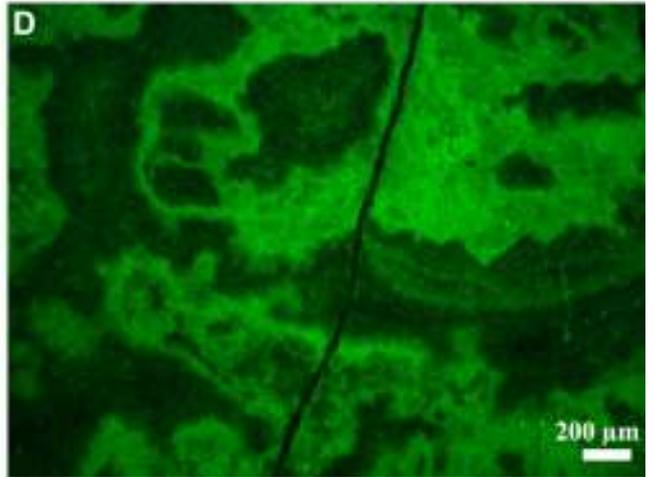
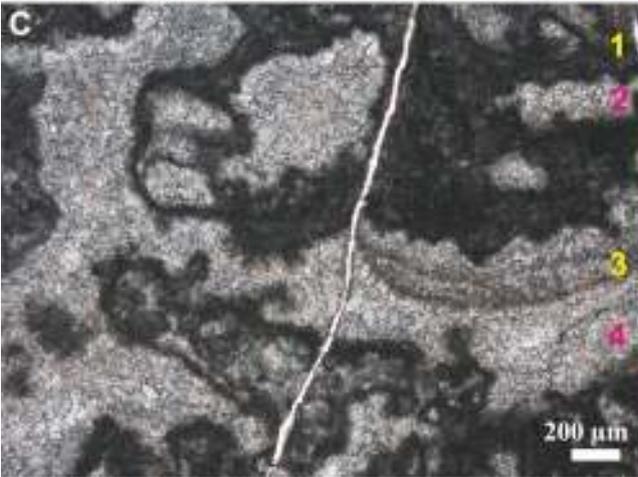
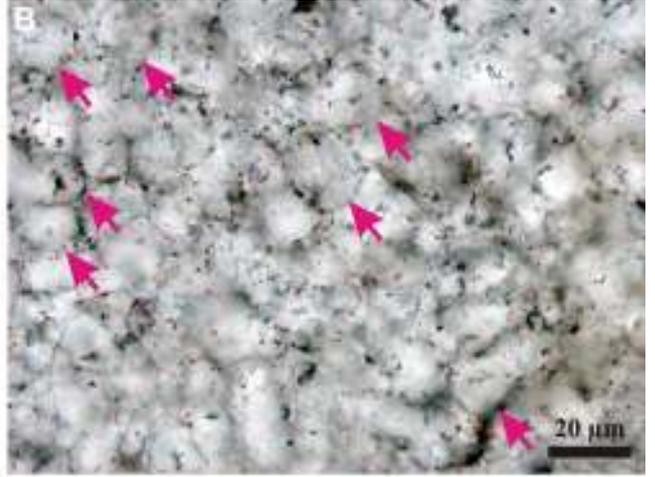
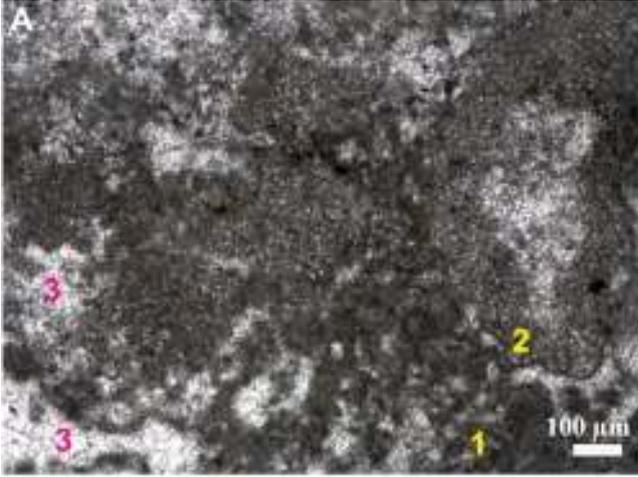


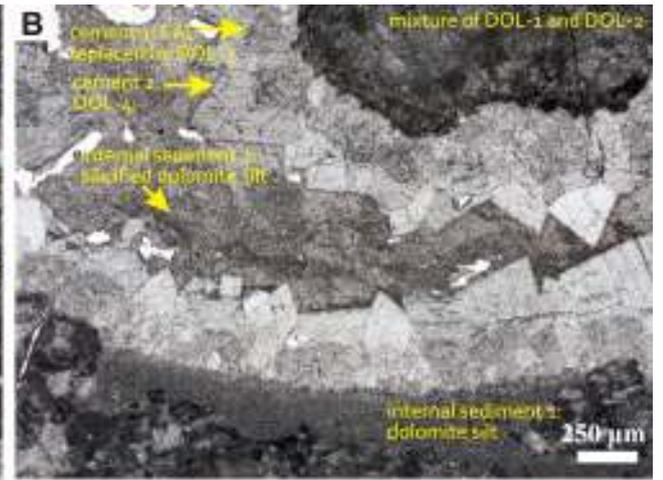
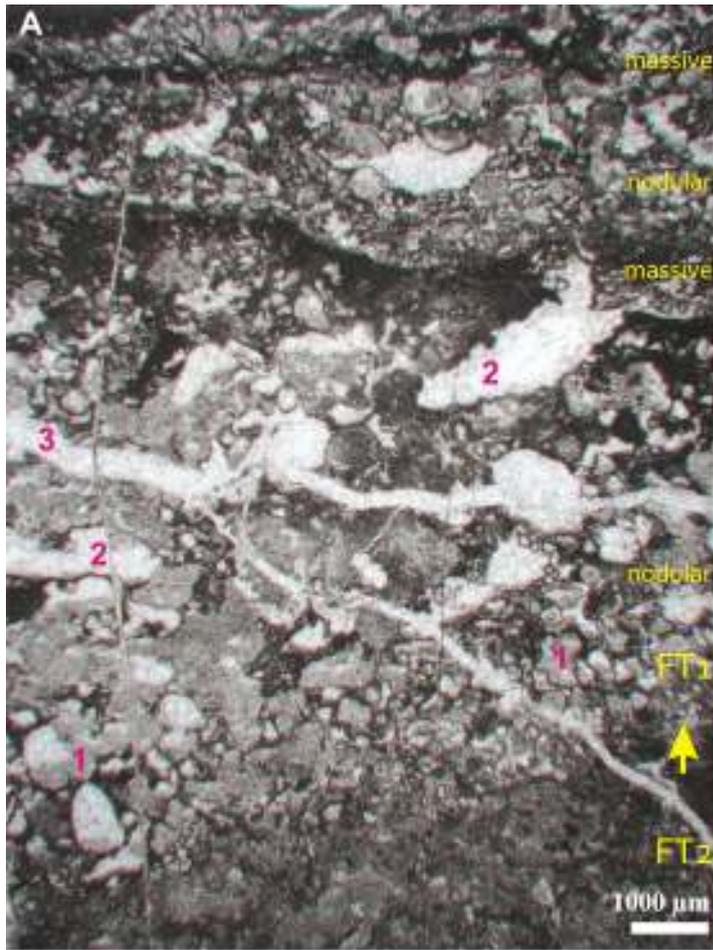
Legend

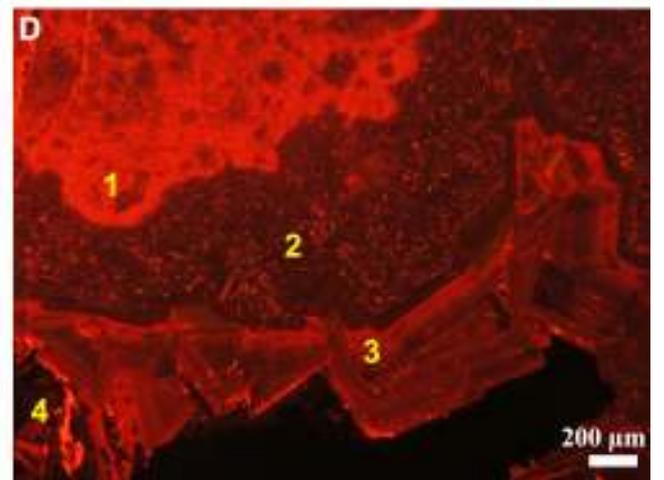
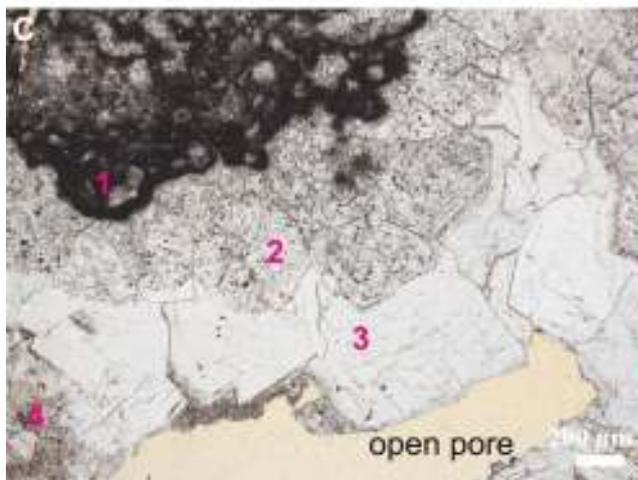
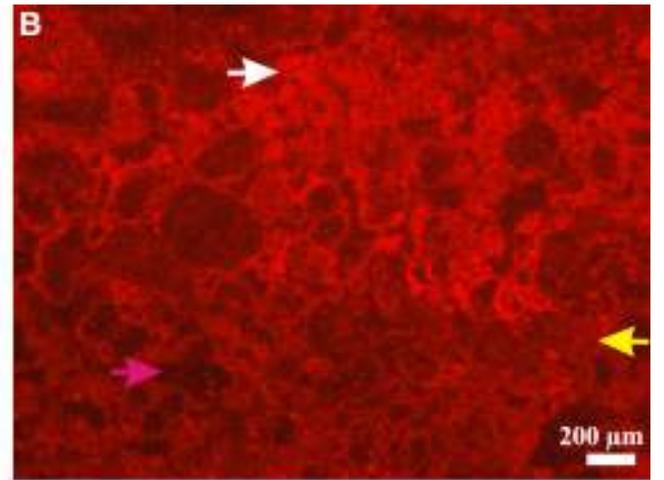
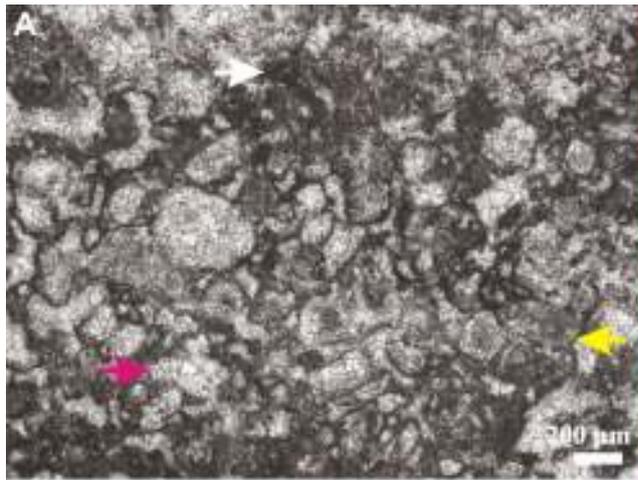
- thin-bedded/laminated lithofacies
- massive lithofacies

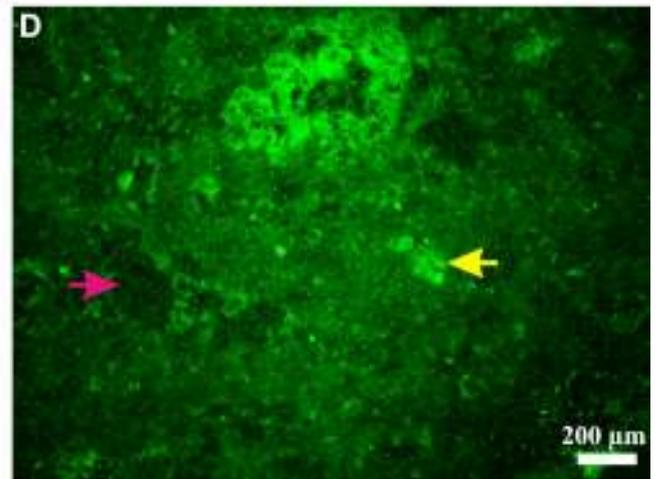
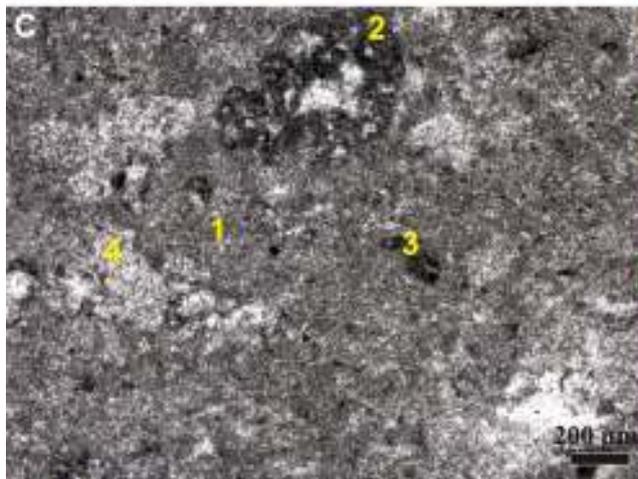
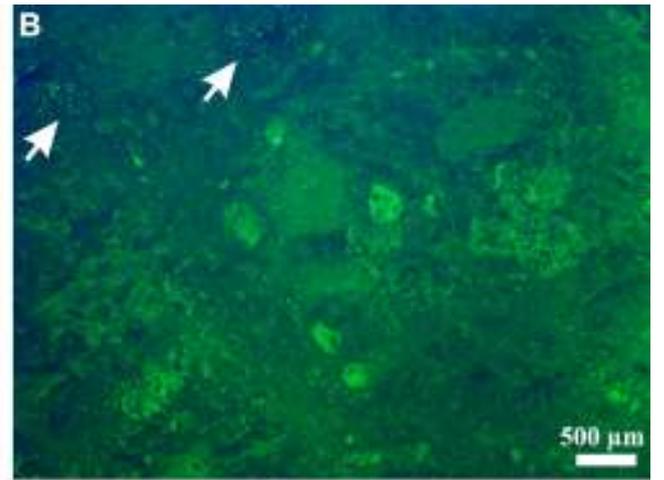
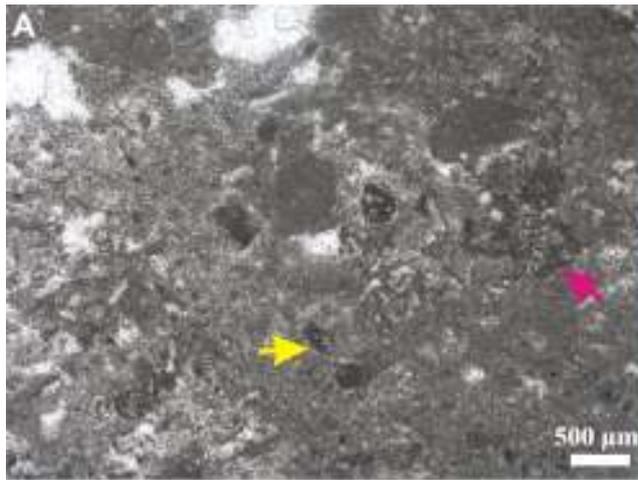
- 1 samples
- locations of studied sections

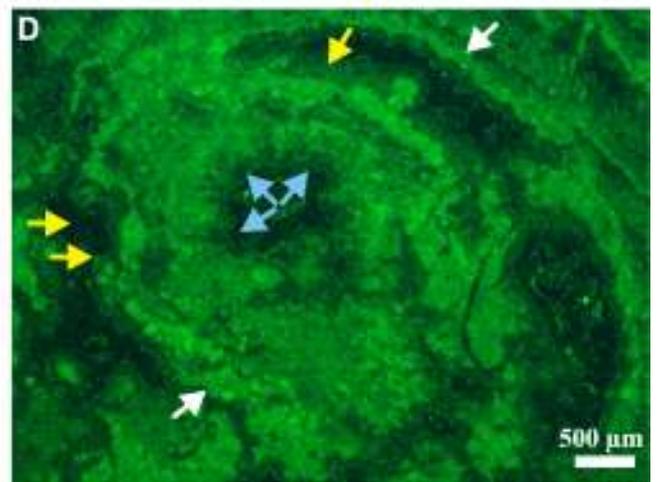
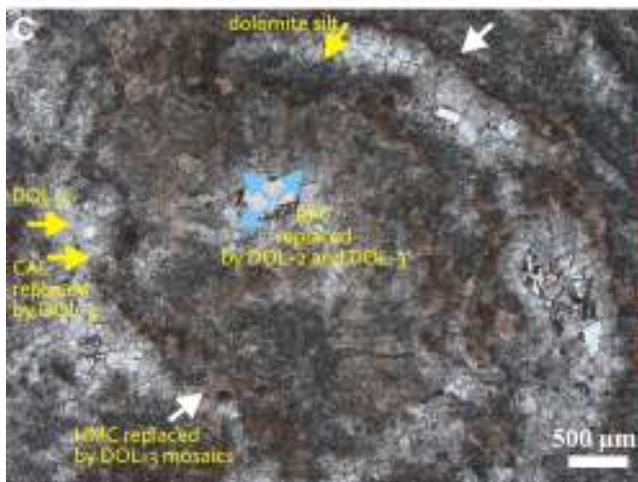
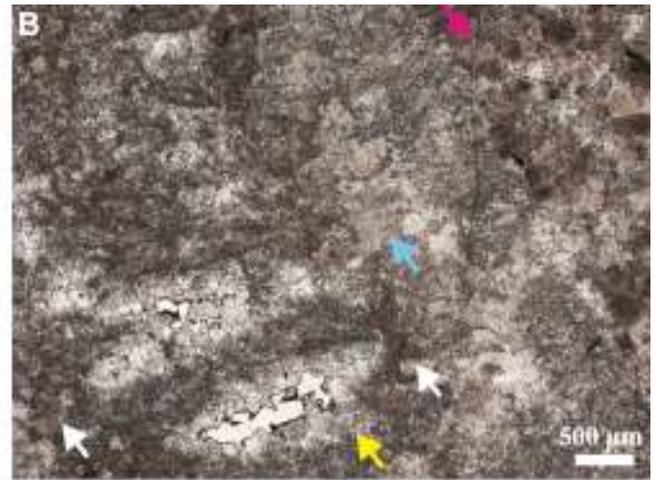
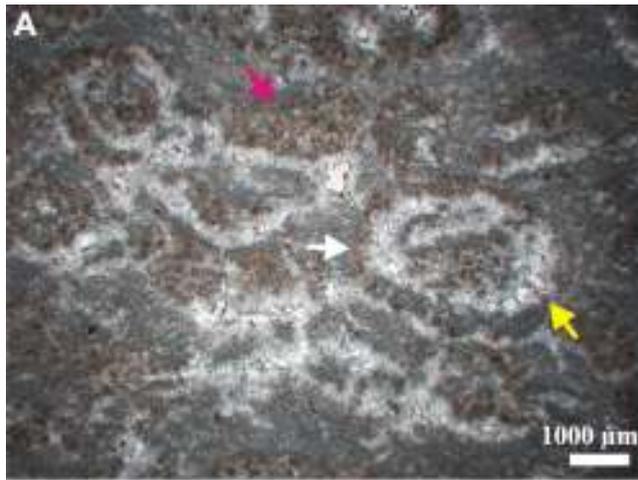


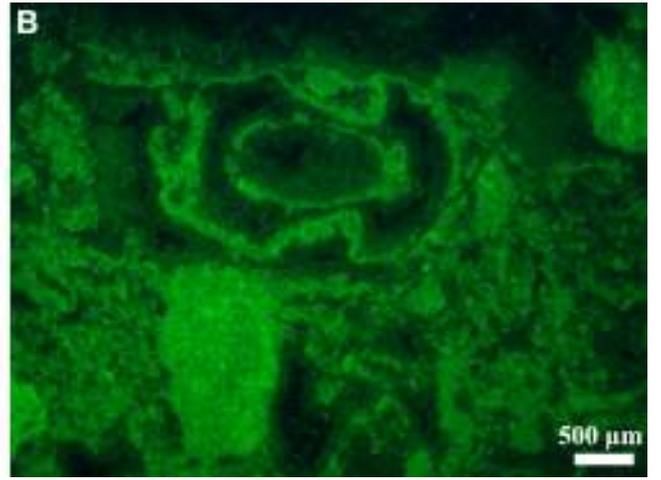
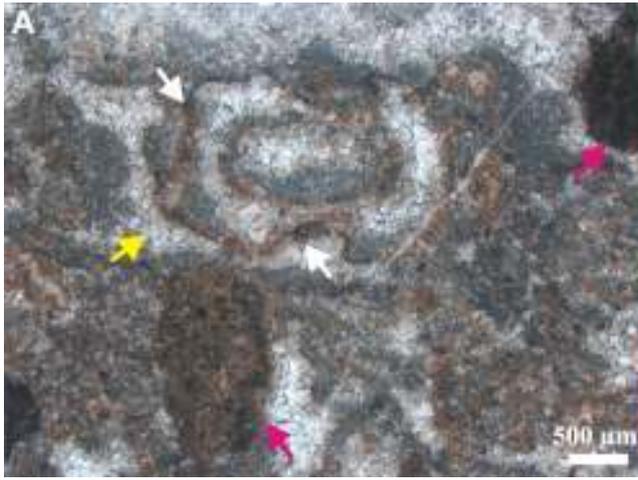


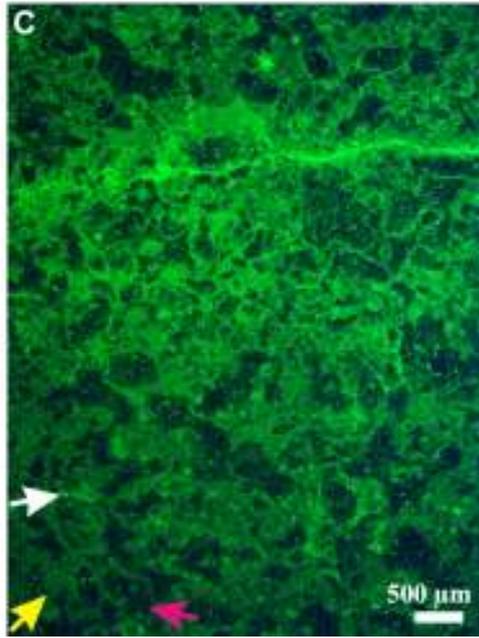
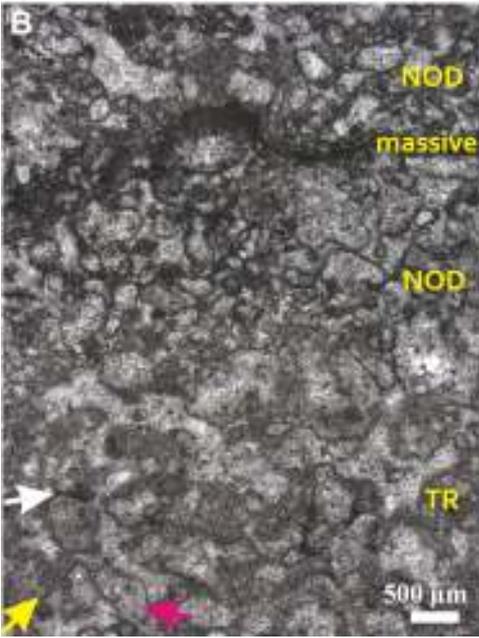
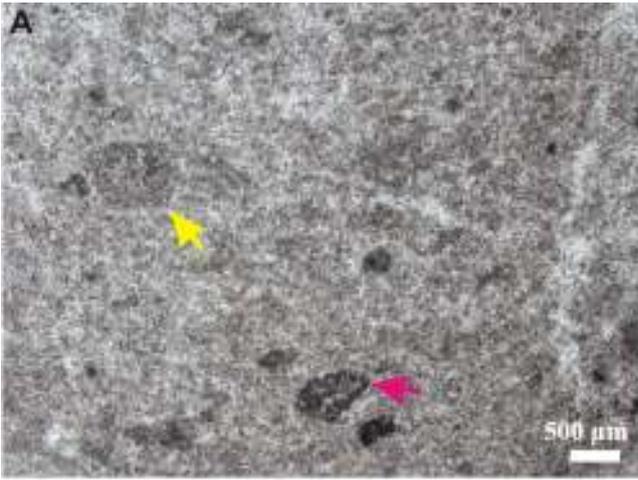


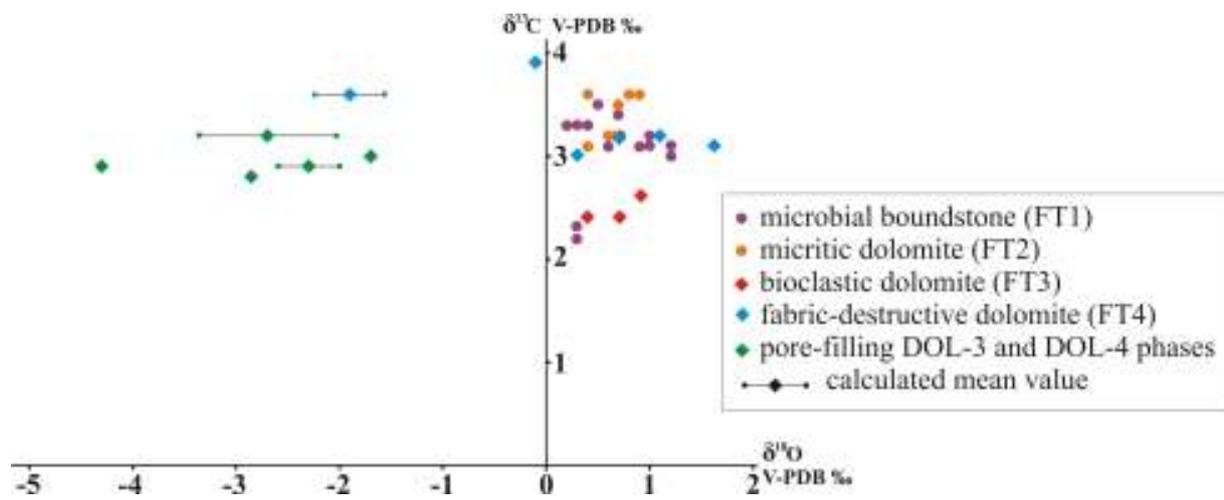


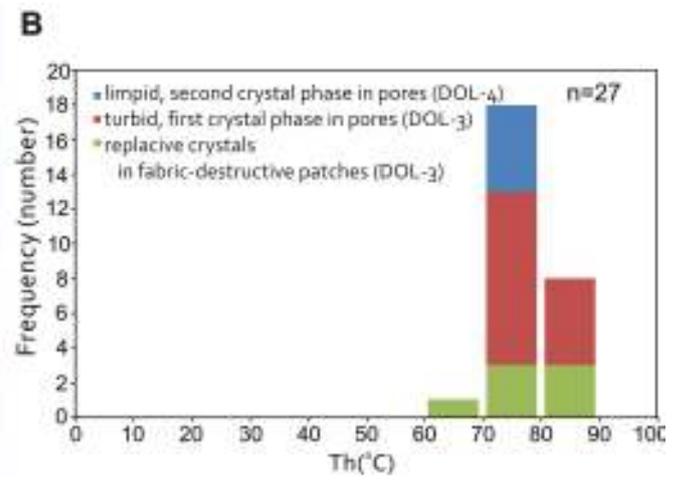


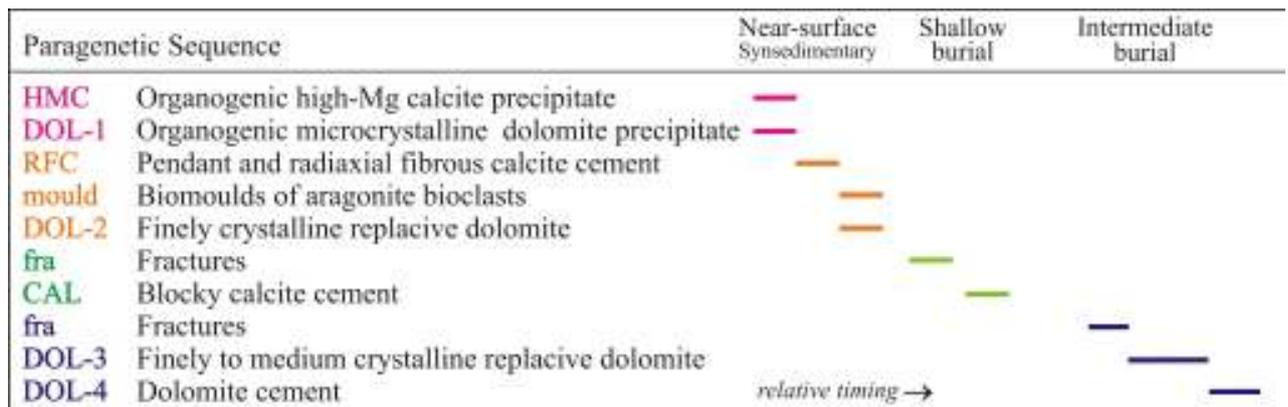


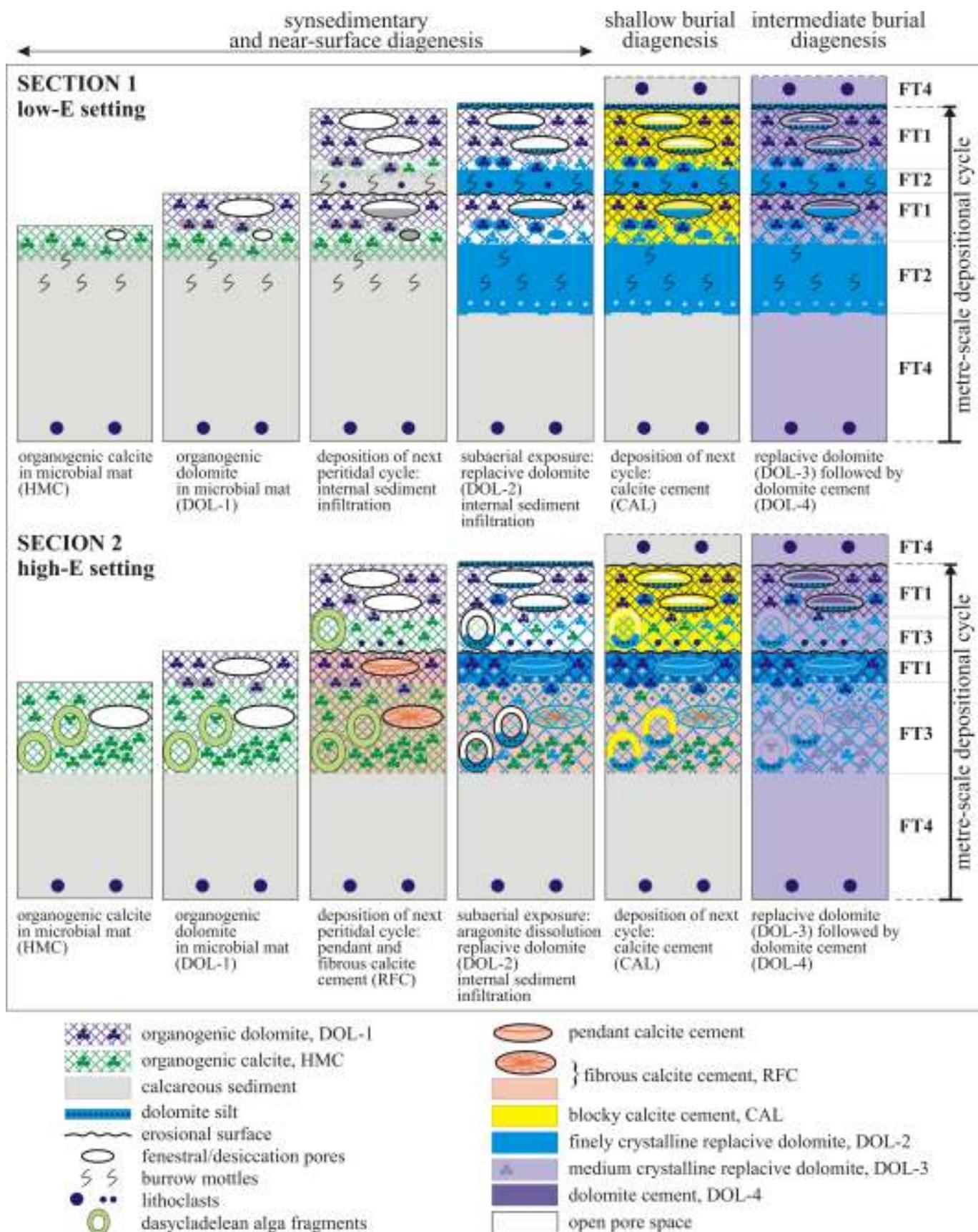












Sample	Dolomite fabric	$\delta^{13}\text{C}$ (‰)	$\delta^{18}\text{O}$ (‰)
4	microbial boundstone	3.1	1.2
1	microbial boundstone	3.0	1.2
1	microbial boundstone	3.1	1.0
4	microbial boundstone	3.2	1.0
4	microbial boundstone	3.2	1.0
4	microbial boundstone	3.1	0.9
2	microbial boundstone	3.4	0.7
6	microbial boundstone	3.2	0.7
4	microbial boundstone	3.1	0.6
2	microbial boundstone	3.5	0.5
2	microbial boundstone	3.3	0.4
2	microbial boundstone	3.3	0.3
12	microbial boundstone	2.2	0.3
12	microbial boundstone	2.3	0.3
2	microbial boundstone	3.3	0.2
2	micritic dolomite	3.6	0.9
2	micritic dolomite	3.6	0.8
2	micritic dolomite	3.5	0.7
4	micritic dolomite	3.2	0.7
4	micritic dolomite	3.2	0.6
2	micritic dolomite	3.6	0.4
4	micritic dolomite	3.1	0.4
15	bioclastic dolomite	2.6	0.9
15	bioclastic dolomite	2.4	0.7
15	bioclastic dolomite	2.4	0.4
1	fabric-destructive dolomite	3.1	1.6
1	fabric-destructive dolomite	3.1	1.6
1	fabric-destructive dolomite	3.2	1.1
1	fabric-destructive dolomite	3.2	0.7
10	fabric-destructive dolomite with lithoclasts	3.0	0.3
3	fabric-destructive dolomite	3.9	-0.1
3	fabric-destructive dolomite	3.6	-1.9*
4	pore-filling DOL-3 and DOL-4 phases	3.0	-1.7
9	pore-filling DOL-3 and DOL-4 phases	2.9	-2.3*
2	pore-filling DOL-3 and DOL-4 phases	3.2	-2.7*
10	pore-filling DOL-3 and DOL-4 phases	2.8	-2.8
8	pore-filling DOL-3 and DOL-4 phases	2.9	-4.3

Sample	Dolomite fabric	Th (°C)
6	turbid crystal phase in pores, DOL-3	81.7
6	turbid crystal phase in pores, DOL-3	75
6	turbid crystal phase in pores, DOL-3	79
6	turbid crystal phase in pores, DOL-3	82
6	turbid crystal phase in pores, DOL-3	85
6	turbid crystal phase in pores, DOL-3	77
6	turbid crystal phase in pores, DOL-3	87
6	limpid crystal phase in pores, DOL-4	75
7	turbid crystal phase in pores, DOL-3	74
7	turbid crystal phase in pores, DOL-3	73
7	turbid crystal phase in pores, DOL-3	73.8
7	turbid crystal phase in pores, DOL-3	73
7	turbid crystal phase in pores, DOL-3	78
7	turbid crystal phase in pores, DOL-3	90
7	turbid crystal phase in pores, DOL-3	79
7	turbid crystal phase in pores, DOL-3	72
7	replacive crystal in fabric-destructive patches, DOL-3	74
7	replacive crystal in fabric-destructive patches, DOL-3	83
7	replacive crystal in fabric-destructive patches, DOL-3	62
7	replacive crystal in fabric-destructive patches, DOL-3	80
7	replacive crystal in fabric-destructive patches, DOL-3	78
7	replacive crystal in fabric-destructive patches, DOL-3	82
7	replacive crystal in fabric-destructive patches, DOL-3	82
7	limpid crystal phase in pores, DOL-4	75
7	limpid crystal phase in pores, DOL-4	79
7	limpid crystal phase in pores, DOL-4	79
7	limpid crystal phase in pores, DOL-4	78