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1 **Cenomanian–Turonian Oceanic Anoxic Event (OAE2) imprint on the northwestern part**
2 **of the Adriatic Carbonate Platform and a coeval intra-platform basin (Istria and**
3 **Premuda Island, Croatia)**

4
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13

14

15 **Abstract**

16

17 The Cenomanian–Turonian boundary (CTB) on the intra-Tethyan Adriatic Carbonate Platform
18 (AdCP) is generally characterised by a transition between microbially laminated and/or
19 bioclastic limestones to calcisphere-rich massive limestone with bioturbated intervals, organic-
20 rich interbeds, firmgrounds, as well as neptunian dikes, carbonate turbidites, tempestites and
21 slumped structures. Compilation of the results from two study sites in the northwestern part of
22 the AdCP and from previous research (on Istria Peninsula and islands in the Adriatic Sea in
23 Croatia) provides a more complete overview of geological events and paleoenvironmental
24 conditions that transformed the formerly contiguous shallow-marine environments during this
25 time period. For the first time, a comparison between protected inner-platform area (Barban

26 section) and a coeval intra-platform basin (Premuda Island section) during the CTB was made.
27 This study utilized a combination of litho-, bio-, and microfacies studies with SEM, EDS, TOC,
28 $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ stable isotope analyses. The stratigraphic successions start with shallow-marine
29 carbonate deposits of the Milna Formation that is conformably overlain by the drowned-
30 platform deposits of the Sveti Duh Formation on the platform and by the Veli Rat Formation in
31 the contemporaneously developed intraplatform basin. These deposits are in turn overlain by
32 the Gornji Humac Formation, which represents re-establishment of shallow-marine
33 depositional systems on the AdCP, whereas the deeper water environment persisted in the intra-
34 platform basin until the Santonian.

35 Despite diagenetic modifications of shallow-marine carbonate deposits, the results of TOC and
36 stable isotope analyses indicate the influence of global Oceanic Anoxic Event 2 (OAE2).
37 Combination of local and regional synsedimentary tectonics and global Late Cretaceous sea-
38 level changes accompanied by anoxic and hypoxic conditions, extinction of numerous benthic
39 foraminifera, diversification and expansion of planktonic foraminifera and calcareous
40 dinoflagellates, provide new insights into the character of the CTB interval in this part of the
41 Tethyan realm.

42

43 Keywords: OAE2, Cenomanian–Turonian, Adriatic Carbonate Platform (AdCP), Sv. Duh
44 Formation, Istria, Kvarner.

45

46

47 **1. Introduction**

48

49 This study aims to define the influence of Oceanic Anoxic Event 2 (OAE2, Jenkyns [1980](#),
50 Scholle et al. [1980](#); Schlanger et al. [1987](#); Arthur et al. [1987](#), [1988](#), [1990](#); Tsikos et al. [2004](#);

51 Keller et al. 2004; Karakitsios et al. 2010; Pearce et al. 2009; Jarvis et al. 2011, 2015; among
52 others) on two different sites of the generally shallow-marine intra-Tethyan Adriatic Carbonate
53 Platform (AdCP) in present-day Croatia. OAE2 was one of the most severe and globally most
54 extensive paleoenvironmental changes. The event occurred at the Cenomanian–Turonian
55 boundary (CTB; about 93.9 My ago; Jarvis et al. 2011; among others). An extinction event,
56 marine anoxia, burial of large amounts of organic carbon (Schlanger and Jenkyns 1976;
57 Schlanger et al. 1987; Parente et al. 2008), and rise in sea level (Hardenbol et al. 1998; Miller
58 et al. 2005; Voigt et al. 2006; Haq 2014; Sames et al. 2016) clearly define the OAE2. Present-
59 day oceans are well oxygenated thanks to a conveyor-belt circulation and distribution of cold,
60 oxygen-rich waters from high latitudes all the way to the abyssal zone. Approximately half of
61 the oxygen injected into the deep-sea regions is consumed by decomposition of organic matter
62 in the water column. Increasing ocean's nutrient content could stimulate biological productivity
63 leading to widespread anoxia (Meyer and Kump 2008, and references therein). Cenomanian
64 submarine volcanism was interpreted as a trigger for a chain reaction that caused OAE2 (Sinton
65 and Duncan 1997; Kerr 1998; Larson and Erba 1999; Bralower 2008; Keller 2008; Turgeon
66 and Creaser 2008; Du Vivier et al. 2014). One of the consequences of OAE2 was enhanced
67 preservation and deposition of organic matter on a global scale recognized as positive shifts of
68 $\delta^{13}\text{C}$ values of carbonate deposits. Peak of the crisis (i.e., disturbed dynamics of the global
69 carbon cycle) lasted between 320 Ky (min.) and 1.04 ± 0.12 My (max.) (Sageman et al. 2006;
70 Strauss 2006; Voigt et al. 2008; Gambacorta et al. 2015; Sullivan et al. 2020). Paleobiological
71 consequences of the crisis included the extinction of approximately 26% of marine animal
72 genera (Raup and Sepkoski 1986; Monnet 2009), and changes in diversity and abundance of
73 planktonic foraminifera and radiolarians (Caron and Homewood 1983; Jarvis et al. 1988; Huber
74 et al. 1999; Culver and Rawson 2004; Erba 2004; Gebhardt et al. 2010), larger foraminifera
75 (Parente et al. 2008; Caus et al. 2009), calcareous nannoplankton (Leckie et al. 2002), as well

76 as rudists (Philip and Airaud-Crumière 1991; Steuber and Löser 2000). Widespread dysoxic to
77 anoxic paleoenvironmental conditions and biotic crises in the world oceans also left a
78 significant mark on shallow-marine carbonate systems (Jenkyns 1991; Gušić and Jelaska 1993;
79 Hilbrecht et al. 1996; Davey and Jenkyns 1999; Parente et al. 2008; Immenhauser et al. 2008;
80 Elrick et al. 2009; Gertsch et al. 2010; Nagm 2015; Brčić et al. 2017).

81 This study focuses on shallow-marine carbonate deposits in the wider area of Istria and Kvarner
82 (Barban and Premuda, Croatia, Fig. 1). Starting in early Cenomanian, the stable and uniform
83 shallow-marine AdCP (Fig. 2a) was impacted by synsedimentary tectonism and sea-level
84 changes, resulting in episodes of pelagic influence. Despite sea-level rise, in some parts of the
85 study area synsedimentary tectonism overprinted eustatic changes and created emerged areas
86 surrounding intra-platform basins (Gušić and Jelaska 1990; Vlahović et al. 1994; Tišljarić et al.
87 1998; Korbar et al. 2012; Brčić et al. 2017; Fig. 2b). One of the main contributions of this study
88 is a comparison between protected inner-platform area (Barban section) and a coeval intra-
89 platform basin (Premuda Island section) during the CTB. This comparison was made possible
90 for the first time in the study area by using an integrated litho-, bio-, and chemostratigraphic
91 approach that allowed the recognition of characteristic signatures of the global OAE2
92 perturbation.

93

94

95 **2. Geological Setting of the Study Area**

96

97 Deposits of the Mesozoic Adriatic Carbonate Platform (AdCP) are currently exposed in an area
98 approximately 800 km long and 200 km wide, from Italy in the northwest, across Croatia, and
99 to Albania in the southeast (Jelaska 2002; Vlahović et al. 2005). These deposits are
100 predominantly shallow-marine carbonate platform successions, typical of the intra-Tethyan

101 realm (Fig. 2). Their maximum thickness reaches 5000-8000 m with a stratigraphic range from
102 the upper part of lower Jurassic (Toarcian) to the Eocene (AdCP *sensu stricto*). The extensive
103 vertical thickness is the result of synsedimentary tectonics and long-term shallow-marine
104 sedimentation, and was also influenced by subsequent post-Cretaceous formation of thrust-
105 nappe structures. In the study area (Istria and Kvarner, Fig. 1), one of the tectonic phases
106 occurred during the early and late Cenomanian. At the Cenomanian–Turonian boundary,
107 eustatic and synsedimentary tectonics locally established drowned platform environments, but
108 at the same time caused uplift and subaerial exposure of the surrounding areas (Davey and
109 Jenkyns 1999; Velić et al. 2002, 2003; Čosović et al. 2004; Vlahović et al. 1994, 2002a, 2003,
110 2005; Korbar et al. 2012; Brčić 2015, 2017; Fig. 2). Eocene ramp-type limestones and
111 synorogenic flysch deposits marked the onset of the Alpine orogenesis in the region (Grandić
112 et al. 1997; Vlahović et al. 2005; Schmid et al. 2008), when the AdCP successions were strongly
113 deformed. The compressional tectonics (Fig. 1b) resulted in the fold-and-thrust structures of
114 the present-day External Dinarides (Tari 2002; Korbar 2009).

115 Paleogeographically the studied deposits belong to the northwestern part of the AdCP (Fig. 2a).
116 During the Cenomanian the larger (mostly western) part of present-day Istria and Kvarner was
117 subaerially exposed (Vlahović et al. 1994, 2002a, b; Brčić 2015, 2017, Fig. 2b). This study
118 focuses on the eastern areas with deeper-marine deposits where sedimentation continued
119 throughout the Turonian (Barban section; Fig. 2b) and until Coniacian–Santonian (Premuda
120 section; Fig. 2c). The term “deeper-marine” is used for paleoenvironments of intraplateau
121 basins with limited pelagic influence and water depth of less than approximately 150 m.
122 Stratigraphic successions in both localities begin with shallow-marine, middle to upper
123 Cenomanian skeletal mudstone-wackestone-packstone and rudist floatstone (the Milna
124 Formation), followed by Cenomanian–Turonian deeper-marine calcisphere limestones (the Sv.
125 Duh and Veli Rat Formations). The Barban succession ends with Turonian to Coniacian

126 shallow-marine skeletal and peloidal mudstone-wackestone, alternating with rudist floatstone
127 (the Gornji Humac Formation; Gušić and Jelaska 1990), while the prolonged deeper-marine
128 sedimentation at Premuda Island resulted in the Veli Rat Formation (Fuček et al. 1991).

129

130

131 **3. Materials and Methods**

132

133 The Barban section, named after a nearby settlement of Barban, is located in the south-eastern
134 part of peninsula Istria in Croatia (Figs. 1 and 2). This section was sampled in road cuts from
135 Barban to the Raša river valley and in the quarry on the eastern side of the valley (Fig. 1e). The
136 Premuda section is located on Premuda Island in the south-eastern part of Kvarner Bay. The
137 exposures at Cape Lopata on the south-eastern tip of Premuda were examined and sampled in
138 detail (Fig. 1d and 1f).

139

140 *3.1. Fieldwork and thin-section microscopy*

141 In the study area, the best quality exposures of continuous limestone successions were sampled
142 (Fig. 1e and 1f). The surrounding area was mapped at a scale of 1:12.500 and 1:25.000. Two
143 detailed stratigraphic sections were measured with a total thickness of 315 m: 175 m of the
144 Barban section and 140 m at Premuda (Fig. 3). Selected lithotypes were sampled and a total of
145 237 polished slabs and 293 thin sections were made for petrographic and micropaleontological
146 analyses in order to define lithotypes and microfossil assemblages, interpret depositional
147 environments, document diagenetic modifications and select the most suitable samples for
148 geochemical analyses. Biostratigraphic analyses focused on numerous species of benthic and
149 planktonic foraminifera (Fig. 3; Croatian Geological Survey repository of the studied material).

150

151 3.2. Stable-isotope Analysis ($\delta^{13}\text{C}_{carb}$, $\delta^{18}\text{O}_{carb}$)

152 Stable isotope analysis was carried out on 122 samples (81 from Barban and 41 from Premuda;
153 Fig. 4; Table 1). Small amounts of carbonate powder (homogeneous, micritic, and non-
154 weathered material, excluding areas with carbonate cement and skeletal fragments) were
155 collected from polished slabs using a microscope-mounted microdrill. Stable-isotope analyses
156 were performed using a DeltaXL mass spectrometer at the University of Massachusetts,
157 Amherst, USA. After heating for an hour at 400°C to remove any volatile organic components,
158 samples were reacted at 70°C with 100% anhydrous phosphoric acid (H_3PO_4) for 10 min.
159 Standard isobaric and phosphoric acid fractionation corrections were applied to all data. Internal
160 analytical precision, monitored through daily analysis of carbonate standards, was better than
161 or equal to 0.1‰ for both carbon and oxygen isotope values. Results are expressed as $\delta^{13}\text{C}$ and
162 $\delta^{18}\text{O}$ values in ‰ relative to the Vienna PeeDee Belemnite standard (VPDB).

165 3.3. TOC and Insoluble Residue Analyses

166 Total organic carbon (TOC; Fig. 5) was measured on mechanically pulverized limestone
167 samples (i.e., bulk powdered micrite), obtained by drilling micritic limestone with a 1 mm
168 diameter drill-bit. A representative weight (10 g) of each sample was treated with hydrochloric
169 acid (4.2M HCl) for 24 h to eliminate carbonate fractions. To dissolve dolomite that could have
170 been present in the samples the dissolution was performed by heating the hydrochloric acid at
171 80°C. The samples were filtered and washed several times in distilled water to remove the
172 remaining acid. The insoluble residue (IR) was weighed in tin capsules and analysed using a
173 Thermo Fisher Scientific Flash 2000 NC Elemental Analyser at the Croatian Geological Survey
174 (HGI-CGS). Assuming a complete elimination of carbonate components during the acid
175 treatment, the percentage of IR was calculated using the equation $\text{IR} = (\text{DM}/\text{TM}) \times 100$, where

176 DM is the weight of the insoluble residue remaining after dissolution of carbonates and TM is
177 the total weight of sample before acid treatment. The amount of TOC_{IR} (%) within IR was
178 determined with elemental analyser and the calculated $\text{TOC}_{\text{sample}}$ (%) for the whole sample was
179 calculated as $\text{TOC}_{\text{sample}} (\%) = (\text{DM}/\text{TM}) \times \text{TOC}_{\text{IR}} (\%)$. The calibration accuracy was verified
180 by measuring samples of certified Soil Reference Material NC (Thermo Scientific), treated in
181 the same way as the samples. Standard quality check analysis of internal standards performed
182 at HGI-CGS yielded a relative standard deviation (RSD) on TOC measurements of 0.4%.

183

184

185 **4. Results**

186

187 The Barban section is composed of three lithostratigraphic units: the Milna (38 m in thickness),
188 Sv. Duh (116 m) and Gornji Humac Formations (21 m; Gušić and Jelaska 1990; Figs. 3 and 4).

189 The Premuda section contains two lithostratigraphic units: the Milna (30 m thick; Gušić and
190 Jelaska 1990) and Veli Rat Formations (110 m; equivalent of the Sv. Duh Formation with
191 prolonged deeper-marine sedimentation; Fuček et al. 1991; Figs. 3 and 4).

192 Well-stratified, shallow-marine, middle to upper Cenomanian, foraminiferal wackestone–
193 packstone, alternating with bioclastic (rudist and chondrodontid bivalves) floatstone and
194 microbial laminites, represent the Milna Formation (Figs. 3, 6, 7 and 8). These deposits underlie
195 a drowned-platform succession of the Sv. Duh Formation. Transition between the Milna and
196 Sv. Duh (or its equivalent Veli Rat) Formations can be gradual (laminites and/or bioclastic
197 limestone replaced with progressively thicker intercalations of calcisphere mudstone-
198 wackestone, e.g. Barban section) or sharp (the well-stratified lithotypes overlain by massive
199 calcisphere-rich limestone, e.g. Ćićarija sections in Brčić et al. 2017). The Sv. Duh Formation
200 consists of 116 m of massive calcisphere wackestone with rare fine-grained bioclastic

201 intercalations and sporadically enriched in organic matter formed in deeper-marine settings of
202 the temporarily submerged/drowned carbonate platform with a significant open-ocean
203 influence. Typical characteristics of these deposits are greyish to light brown erosional surfaces,
204 poorly stratified to massive mudstone-wackestone with calcareous dinoflagellate cysts, fine-
205 grained carbonate bioclasts, planktonic foraminifera, ostracods, pelagic crinoids, sponge
206 spicules, echinoid spines, rare benthic foraminifera, thin-shelled bivalves, and gastropods.
207 Other characteristics include bioturbation, dissolution seams, current microlamination, and
208 undulated upper bedding planes. Stratigraphic range of this unit is late Cenomanian–early
209 Turonian. Unlike the Sv. Duh Formation, the Veli Rat Formation is characterised by carbonate
210 turbidites reflecting a different setting and prolonged deeper-marine sedimentation (late
211 Cenomanian–early Santonian; for a more detailed explanation see Sections 4.2. and 5). The
212 uppermost well-stratified shallow-marine succession belongs to the Gornji Humac Formation
213 characterised by fenestral mudstone, bioclastic-peloid-skeletal wackestone and radiolitid
214 floatstone (with rudist debris, benthic foraminifera, echinoid spines, fine carbonate detritus, and
215 peloids). Transition between the Sv. Duh and Gornji Humac Formations is commonly defined
216 by a shallowing-upward trend (in some cases by oncoid and laminites facies). Deposits in the
217 lower part of Gornji Humac contain shallow-marine bioclastic material mainly composed of
218 rudist debris, benthic foraminifera, crinoids, echinoid spines, fine carbonate detritus, and
219 peloids that infilled intraplatform depressions. Stratigraphic range of the Gornji Humac
220 Formation is middle Turonian-Coniacian (Figs. 3, 6, 7 and 8; for a more detailed explanation
221 see Sections 4.1. and 5).

222

223 *4.1. The Barban section*

224 The Barban section (Figs. 3 and 4) starts with shallow-marine peloid-skeletal-bioclastic
225 wackestone–packstone alternating with microbial laminites and sporadic rudist bioclastic

226 floatstone. Dominant allochems are benthic foraminifera, rudist debris, fine-grained carbonate
227 bioclasts and detritus, ostracods, fragments of dasyclad algal and rare *Decastronema kotori*
228 (Radoičić) and *Thaumatoporella parvovesiculifera* (Raineri). The presence of benthic
229 foraminifera *Pastrickella balcanica* (Cherchi, Radoičić and Schroeder) *Chrysalidina gradata*
230 d'Orbigny, and *Vidalina radoicicae* Cherchi and Schroeder stratigraphically defines these
231 deposits as middle to upper Cenomanian and as the uppermost part of the Milna Formation
232 (Fig. 9).

233 Transition between the Milna and Sv. Duh Formations (42 m from the bottom of Barban
234 section, Fig. 7b) begins with successive intercalations of deeper-marine fine-grained bioclastic
235 calcisphere wackestone within predominantly shallow-marine bioclastic to skeletal packstone.
236 Above these 2 m thick transitional deposits, the bioclastic lithofacies is replaced by thick-
237 bedded to massive calcisphere mudstone to wackestone with planktonic foraminifera
238 (*Rotalipora* sp., *Praeglobotruncana* sp. and *Heterohelix* sp.), ostracods, pelagic crinoids,
239 sponge spicules, echinoid spines, bivalve bioclasts, and fine carbonate detritus. Exposures of
240 this interval are characterised by greyish-light to brown colour, and by their brittle, fractured,
241 partly recrystallized, and thick-bedded to massive appearance. Stratigraphically these 98 m
242 thick deposits belong to the uppermost Cenomanian to middle Turonian and represent the
243 deeper-marine Sv. Duh Formation (Fig. 10).

244 An 18 m thick interval (between 136 and 154 m of the Barban section) is characterised by a
245 shallowing-upward trend represented by cross-bedded and fining upward grainstone followed
246 by calcisphere wackestone. The sharp contact (at 136 m) between the grainstone and
247 wackestone is marked by prominent stylolites (Figs. 3, 4 and 7d). This part of the section
248 represents a transition between the Sv. Duh and Gornji Humac Formation.

249 The uppermost 21 m of the succession consists of shallow-marine bioclastic floatstone of the
250 Gornji Humac Formation with radiolitic rudists *Distefanella* sp. and thin-shelled bivalves

251 (*Exogyra* sp.), alternating with fenestral mudstone and bioclastic-peloid-skeletal wackestone.
252 The stratigraphic range of the Gornji Humac Formation in the wider study area is middle
253 Turonian to Coniacian, based on the presence of *Distefanella* sp. and *Hippurites requieni*
254 Matheron, *Decastronema kotori* (Radoičić), *Thaumatoporella parvovesiculifera*,
255 *Moncharmontia* sp., *Pseudocyclamina sphaeroidea* Gendrot, *Scandonea samnitica* De
256 Castro, *S. mediterranea* (De Castro), *Dicyclina schlumbergeri* Munier-Chalmas and *Murgella*
257 *lata* Luperto-Sinni (Velić 2007; Figs. 3 and 4).

258

259 4.2. The Premuda section

260 The Premuda section (Figs. 3 and 4) starts with the Milna Formation shallow-marine peloidal-
261 skeletal wackestone–packstone, sporadically alternating with slightly undulating microbial
262 laminites, peloidal packstone–grainstone and rudist lithostromes (Fig. 8a). The most common
263 skeletal allochems are benthic miliolid and nezzazatid foraminifera. Index benthic foraminifera
264 *Pastrickella balcanica* (Cherchi, Radoičić and Schroeder), *Chrysalidina gradata* d’Orbigny,
265 *Vidalina radoicicae* Cherchi and Schroeder, *Pseudorhapydionina dubia* De Castro (Fig. 9)
266 clearly indicate middle to late Cenomanian age for the top of Milna Formation. Rudist debris
267 and in places whole radiolitid shells make up the lithostrome interlayers. The remaining
268 allochems include fine-grained carbonate bioclasts and detritus, ostracods, echinoid spines,
269 crinoids, and fragments of dasycladal algae (*Heteroporella lepina* Pratulon). This interval also
270 features bioturbated interlayers, a neptunian dike (at 13 m of the section) and undulated upper
271 bedding planes.

272 Relative to other localities and sections in the surrounding area, the transition between Milna
273 and Veli Rat Formations at Premuda is atypical. At 30 m of the section (Figs. 3 and 4), there is
274 a first 2.5 m thick layer of calcisphere wackestone–packstone with increased proportion of fine
275 bioclasts, overlain by 4 m of shallow-marine lithotypes. This pattern is repeated for the next 24

276 m. The proportion of fine bioclasts in the calcisphere wackestone–packstone decreases upwards
277 and the thickness of the deeper-marine deposits increases. The shallow-marine unit becomes
278 more thinly bedded and with a higher proportion of radiolitid bioclasts upsection. Deeper-
279 marine limestones dominate the stratigraphic interval between 44 and 140 m. Within this
280 interval, from 44 to 60 m, there are two up to a meter-thick lenticular intercalations of coarse-
281 grained bioclastic-lithoclastic floatstone (Figs. 8b, 8c, 8d and 8e). Farther upsection, from 60
282 to 102 m, the intercalations are thinner and characterised by fine-grained bioclastic packstone
283 (up to 0.5 m thick lenses at every 5 to 10 m of the section). From 102 m to the top of the section
284 there are no intercalations of platform-derived bioclastic limestone and the succession contains
285 only pelagic allochems.

286 Based on the first appearance of thick-bedded to massive calcisphere wackestone–packstone,
287 the 30–140 m interval of Premuda section is attributed to the Veli Rat Formation. This interval
288 contains calcareous dinoflagellates, planktonic foraminifera, ostracods, pelagic crinoids,
289 sponge spicules, echinoid spines, fine bivalve bioclasts, and undefined carbonate detritus.
290 Identified planktonic foraminifera include: *Whiteinella archaeocretacea* Pessagno,
291 *Helvetoglobotruncana praehelvetica* (Trujillo), *Whiteinella* cf. *paradubia* (Sigal), *Dicarinella*
292 *primitiva* (Dalbiez), *Dicarinella imbricata* (Monrod), *Praeglobotruncana* cf. *algeriana*
293 (Caron), *Rotalipora* sp., *Praeglobotruncana* sp., *Praeglobotruncana gibba* (Klaus),
294 *Helvetoglobotruncana* cf. *helvetica* (Bolli), *Marginotruncana sigali* (Reichel),
295 *Archaeoglobigerina* cf. *cretacea* (d'Orbigny), *Archaeoglobigerina* cf. *blowi* Pessagno, and
296 *Heterohelix* sp. (Fig. 10). Index planktonic foraminifera indicate the lower Turonian to lower
297 Coniacian stratigraphic range (Caron 1985).

298

299 4.3. Stable-isotope Data ($\delta^{13}C_{carb}$, $\delta^{18}O_{carb}$) of Bulk Micrite

300 Sampling of the stratigraphic sections for carbon and oxygen isotope analysis targeted the CTB
301 interval with the intention of comparing the results with the European carbon-isotope reference
302 curve from Eastbourne (Gun Gardens, England; Paul et al. 1999; Jarvis et al. 2006; Pearce et
303 al. 2009; Figs. 4 and 5). Micritic carbonate components were sampled from the 20 to 135 m
304 section interval at Barban, and from 18 to 97 m at Premuda. The Barban section samples have
305 $\delta^{13}\text{C}$ values ranging between 0.71 and 3.77‰ and $\delta^{18}\text{O}$ between -5.64 and -0.98‰, and the
306 Premuda samples yielded $\delta^{13}\text{C}$ values of -0.96 to +4.13‰ and $\delta^{18}\text{O}$ values of -6.58 to -3.48‰
307 (Fig. 5; Table 1). The Barban section shows a certain degree of covariance between the $\delta^{13}\text{C}$
308 and $\delta^{18}\text{O}$ data (for a more detailed explanation see Section 5). The Premuda section has no
309 significant covariance between the $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ data (Fig. 5).

310

311 *4.4. Carbon isotope values ($\delta^{13}\text{C}_{carb}$) of the Barban section*

312 The first sample for stable isotope analysis was collected in the Milna Formation at 20.5 m of
313 the Barban section. The $\delta^{13}\text{C}$ values start at about 3.0 ‰ and for the next 10 m they gradually
314 decrease to 2.5 ‰ (Fig. 4). The 30-45 m interval (the uppermost Milna Formation) is
315 characterised by fluctuations in $\delta^{13}\text{C}$ values between 2.0 and 4.0 ‰. At 45 m of the Barban
316 section, in the Sv. Duh Formation, there is a steep decrease in $\delta^{13}\text{C}$ values from 3.0 to 0.75 ‰.
317 In the overlying 10 m (45-55 m of the section) the values vary between 0.75 and 2.1‰. At 56
318 m of the section, the values sharply increase to 3.5 ‰. In the following 24 m (56-80 m) the
319 values remain relatively uniform (around 3.0 ‰) and without major oscillations. At 80 m of the
320 section there is another positive shift in $\delta^{13}\text{C}$ values from 3.0 ‰ to 4.0 ‰. The values remain
321 relatively high (between 3.5 and 4.0 ‰) in the next 25 m (105 m of the section). The interval
322 from 105 to 115 m of the section revealed a decline from 4.0 to 2.4 ‰ in $\delta^{13}\text{C}$ values. In the
323 following 2 m the values increase to 3.5 ‰ (at 117 m of the section) and for the rest of the
324 succession (117-135 m) they vary between 3.0 and 3.5 ‰ (Figs. 4 and 5; Table 1).

325

326 *4.5. Oxygen isotope values ($\delta^{18}O_{carb}$) of the Barban section*

327 Starting with -0.1 ‰ at 20.5 m in the Milna Formation of the Barban section, the $\delta^{18}O$ values
328 decrease gradually to -3.9 ‰ for the next 13 m (Fig. 4). The interval from 33 m to 45 m of the
329 section is characterised by significant fluctuations in $\delta^{18}O$ values between -2.7 and -5.4 ‰, with
330 a generally decreasing-upward trend. For the next 10 m (45–55 m of the section) the $\delta^{18}O$ curve
331 oscillates between -5.4 and -4.3 ‰. Above this interval, a general shift towards positive values
332 is maintained for 30 m of the section. Between 55 and 85 m, the values increase upsection and
333 vary from -5.5 to -3.2 ‰. For the rest of the section the values vary slightly between -3.9 and -
334 3.0 ‰ (Figs. 4 and 5; Table 1).

335

336 *4.6. Carbon isotope values ($\delta^{13}C_{carb}$) of the Premuda section*

337 The first sample collected in the Milna Formation at 18 m of the Premuda section has a $\delta^{13}C$
338 value of -0.5 ‰ (Fig. 4). For the next 2 m the $\delta^{13}C$ values increase to 2.2 ‰ followed by a sharp
339 decline to -0.7 ‰ (at 25 m of the section). The next 5 m record an increase to the highest $\delta^{13}C$
340 value of 4.0 ‰ at the transition between Milna and Veli Rat Formations. The 30-42 m interval
341 is characterised by fluctuating, but generally decreasing values, reaching the minimum of -1.0
342 ‰. An upward shift to positive values (up to 4.0 ‰ at 55 m of the section) in the 42-59 m
343 interval is also marked by pronounced fluctuations. For the next 12 m (60–72 m) the carbon-
344 isotope curve follows a decreasing trend from 4.0 to 0.5 ‰. A recovery to a maximum value of
345 4.1 ‰ occurs at 74 m of the Premuda section, and the uppermost 23 m (74-97 m) show the
346 values decline to 2.8 ‰ in an oscillating fashion (Figs. 4 and 5; Table 1).

347

348 *4.7. Oxygen isotope values ($\delta^{18}O_{carb}$) of the Premuda section*

349 The Premuda section samples show a considerable variation in their $\delta^{18}\text{O}$ values from -6.6 to -
350 3.5 ‰. Initially the values show a slight increase from -4.5 ‰ at 18 m of the section to -3.5 ‰
351 at 30 m, followed by a sharp decline to -6.6 ‰ at 34 m at the transition between Milna and Veli
352 Rat Formations (Fig. 4). In the next 18 m (30-48 m) the values fluctuate between -6.6 and -4.6
353 ‰. The fluctuating trend continues upward to the top of the section (48-97 m), with a change
354 in the slope of the curve towards slightly more positive values. The lowest values (-6.0 to -5.5
355 ‰) are recorded at 55, 74, 84 and 92 m, and the highest values of -4.5 to -4.3 ‰ come from 68,
356 80 and 86 m (Figs. 4 and 5; Table 1).

357

358 *4.8. TOC and Insoluble Residue Analysis*

359 The variations in TOC and insoluble residue from the Barban section are shown in Figure 5 and
360 Table 1. To display the relationship of TOC/insoluble residue/stable-isotope data of relatively
361 pure carbonates (shallow-marine limestones) the TOC values are multiplied by 100 (Fig. 5).
362 In the lowermost 23 m of the Barban section (20-43 m) the TOC values oscillate between 0.02
363 and 0.7 ‰. In the next 2 m (43-45 m) the TOC reaches the highest value of 1.8 ‰ at the
364 transition between Milna and Sv. Duh Formations, followed by a sharp decline (45-47 m) to
365 0.2 ‰. For the next 48 m (47-95 m) the TOC curve shows fluctuations with a generally
366 increasing upward trend from 0.2 to 0.8 ‰. For the rest of the section the TOC values show
367 somewhat less variation and a general upward decrease from 0.8 to 0.1 ‰. The insoluble residue
368 curve (clay minerals and quartz) coincides with the TOC one, from a minimum value of -0.13
369 % and the maximum of 2.47 %. Noticeable fluctuations in the insoluble residue values were
370 observed at intervals 37-46 m, 83-96 m and 114-121 m (Fig. 5).

371

372

373 **5. Interpretations and Discussion**

374

375 Despite local and regional mid-Cretaceous synsedimentary tectonism and late Cenomanian
376 global sea-level changes, the paleoenvironmental conditions throughout the AdCP remained
377 relatively stable and highly productive throughout this time (Gušić and Jelaska 1990; Vlahović
378 et al. 2005; Tišljarić et al. 1998; Cvetko Tešović et al. 2011; Korbar et al. 2012; Brčić et al. 2017;
379 Picotti et al. 2019). Post-Cenomanian tectonic changes, however, served as a prelude for a
380 transition in depositional geomorphology from rimmed carbonate platform (Paleocene) to a
381 ramp (Eocene). In the study area during the CTB interval, synsedimentary tectonism locally
382 overprinted eustatic changes and laterally created uplifted areas undergoing subaerial exposure
383 coeval with the existence of intra-platform basins (Brčić et al. 2017; Fig. 6). This caused a
384 pronounced lateral differentiation of the AdCP facies (Tišljarić et al. 1994, 1998, 2002; Vlahović
385 et al. 1994, 2002a, b, 2005; 2011; Velić et al. 2002, 2003). The processes of karstification and
386 the formation of paleorelief affected the emerged parts of western AdCP (upper part of the
387 Milna Formation). Tectonically and eustatically drowned areas (i.e., intra-platform basins)
388 experienced deeper-marine depositional conditions (see Fig. 6) as reflected in the transition into
389 the overlying Sv. Duh and Veli Rat Formations.

390

391

392 *5.1. Shallow-marine facies*

393 The upper parts of the Milna Formation are characterised by intensive carbonate bioproduction
394 (dominant contribution of biomaterial in the form of large benthic foraminifera and rudist
395 bioclasts; Fig. 4 and Fig. 9). Cenomanian foraminiferal assemblages in the study area are
396 probably the richest in the entire succession of shallow-marine carbonates (from the Toarcian
397 to the Santonian; Velić 2007). Cyanobacterial laminites predominate in the top few meters of
398 the Milna Formation, and sporadic fenestral mudstones with laminites indicate the minimum

399 water levels in large parts of the carbonate platform (Raspini 2012). The upper part of the Milna
400 Formation is represented by the shallow-marine carbonate platform facies in the first 42 m of
401 the Barban section and 30 m of the Premuda section (Fig. 4; Gušić and Jelaska 1990). Peloid-
402 skeletal wackestone–packstone alternating with microbial laminites and sporadic rudist
403 bioclastic floatstone with fine-grained carbonate bioclasts, ostracods, fragments of dasycladial
404 algae and crinoids were formed in protected environments, ranging from shallow subtidal,
405 across intertidal to tidal flats (Fig. 6a). Detailed descriptions of this paleoenvironment are
406 provided by Korbar et al. (2001), Steuber et al. (2005), and Korbar and Husinec (2003; Kvarner
407 area; Fig. 1).

408 Late Cenomanian in the study area is characterised by the presence of benthic foraminifera
409 *Pastrickella balcanica* (Cherchi, Radoičić and Schroeder), *Vidalina radoicicae* Cherchi and
410 Schroeder, *Chrysalidina gradata* d'Orbigny, and *Pseudorhapydionina dubia* De Castro (Fig.
411 9). In addition, there are also *Cisalveolina* sp., *Peneroplis planatus* (Fichtel & Moll), *Scandonea*
412 sp., *Cuneolina* cf. *pavonia* (d'Orbigny) *Pseudonummoloculina heimi* (Bonet), *Idalina* cf.
413 *antiqua* (Munier-Chalmas et Schlumberger), *Nezzazata* cf. *gyra* (Smout), and *Nezzazata*
414 *simplex* (Omara). Following the biozonations of Velić (2007), Chiocchini (2008) and Frijia et
415 al. (2015), this interval belongs to the *Chrysalidina gradata* benthic foraminifera biozone.

416 At the Barban section, following the carbonate platform drowning at the CTB, a minor
417 regressive phase and infilling of intraplatform basins with shallow-marine paleoenvironmental
418 conditions were re-established in middle Turonian (Fig. 6c). The stratigraphic interval from 154
419 to 175 m in the Barban section belongs to the Gornji Humac Formation. Its middle Turonian to
420 Coniacian age is defined by the presence of *Distefanella* (Henhöfer et al. 2014) and *Hippurites*
421 *requieni* Matheron rudists, *Moncharmontia* sp., *Pseudocyclamina sphaeroidea* Gendrot,
422 *Scandonea samnitica* De Castro, *S. mediterranea* (De Castro), *Dicyclina schlumbergeri*
423 Munier-Chalmas and *Murgella lata* Luperto-Sinni benthic foraminifera, but also *Decastronema*

424 *kotori* (Radoičić) and *Thaumatoporella parvovesiculifera*. Coniacian limestones of the Gornji
425 Humac Formation at the Barban section represent the last deposits of the AdCP in this area.
426 Unlike the above, the Veli Rat Formation at the Premuda section continued with deeper-marine
427 sedimentation into the Santonian (see Section 5.3. [Figs. 3 and 6](#)).

428

429

430 5.2. Transition from shallow-marine to deeper-marine facies

431 In contrast with the Cenomanian, the Turonian on AdCP (and surrounding shallow-marine
432 carbonate platforms) is characterised by a decrease in the diversity and abundance of benthic
433 foraminiferal assemblages as a consequence of global sea-level rise in the latest Cenomanian
434 and earliest Turonian (Gušić and Jelaska [1990](#); Velić [2007](#); Parente et al. [2008](#)). Transition
435 between the Milna and Sv. Duh Formations at the Barban section is a typical example of
436 oscillatory transgression (series of cyclic steps generally showing a deepening upward trend).
437 The stratigraphic interval between 40 and 44 m contains several successive repetitions of
438 shallow-marine bioclastic material and micritic intercalations with pelagic influence. The same
439 transition in the Premuda section has characteristic features of the proximal part of the carbonate
440 turbidite facies with reworked, poorly sorted bioclastic breccias found inside the deeper-marine
441 calcisphere mudstone-wackestone (Colacicchi and Baldanza [1986](#); Fuček et al. [1991](#); Moro and
442 Čosović [2013](#)). These changes are repeated throughout the 30–54 m interval of the Premuda
443 section. Occasionally there are slope-derived slump structures associated with fine-grained,
444 reworked benthic platform bioclasts and autochthonous pelagic material with the appearance
445 of neptunian dykes. These deposits indicate that Premuda, unlike the Barban section, was
446 located on the intraplatform basin margins during the CTB ([Figs. 2a and 6](#)). Benthic fossils in
447 bioclastic input within the transitional zone are represented with index foraminifera *Pastrickella*
448 *balcanica* (Cherchi, Radoičić and Schroeder), *Vidalina radoicicae* Cherchi and Schroeder,

449 *Chrysalidina gradata* d'Orbigny, *Cisalveolina* sp., and *Pseudorhapydionina dubia* De Castro.
450 The CTB is also characterised by re-sedimented upper Cenomanian index benthic foraminifera
451 within autochthonous pelagic layers of the transitional zone with planktonic foraminifera
452 (*Whiteinella archaeocretacea*, *W. praehelvetica*, *W. cf. paradubia*, *Rotalipora* sp. and
453 *Praeglobotruncana* sp.).

454

455 5.3. Deeper-marine facies

456 The total thickness of the Sv. Duh Formation deposits in the Barban section is 116 m.
457 Stratigraphically these deposits are entirely Turonian, and have a sharp transition into the
458 overlying shallow-marine Gornji Humac Formation (Turonian–Coniacian). Unlike Barban, the
459 Premuda section reveals a prolonged deeper-marine sedimentation (the Veli Rat Formation).
460 The youngest pelagic fossils from the upper part of the Premuda section are of Coniacian age
461 (*Marginotruncana sigali*, *Dicarinella* sp., *Archaeoglobigerina cf. cretacea*, *A. blowi*; Fig. 10).
462 Geological mapping of the wider area of the island also revealed some even younger pelagic
463 deposits (Santonian; Moro and Čosović 2013; Fuček et al. 2018). The difference in
464 paleogeographic location of the individual sections is also evident in the dominance of pelagic
465 components at Premuda (dinoflagellate cysts and planktonic foraminifera; Figs. 4 and 6). The
466 pelagic deposits are about 100 m thick in both stratigraphic sections. At Barban the complete
467 pelagic interval formed during the latest Cenomanian and almost the entire Turonian (in the
468 span of approximately 5 million years, Cohen et al. 2018). At the Premuda section, the same
469 interval was deposited from the latest Cenomanian through the beginning of Coniacian
470 (approximately 6 million-year span, Cohen et al. 2018). The main reason for lower depositional
471 rates and/or smaller thickness of the Premuda deposits (i.e., the Veli Rat Formation) is the
472 difference in their paleoenvironment. The Premuda section was situated on the margin of an
473 intraplateau basin, which resulted in greater accommodation space and stronger open ocean

474 influence, but was characterised by lower depositional rates (see Fig. 3 for the relationship
475 between thickness and stratigraphic intervals). In contrast, the Barban section was likely located
476 in a more protected intraplatform area where high bioclastic production (see Figs. 2 and 6) and
477 small accommodation space resulted in faster infilling and shallowing of the drowned platform
478 facies (i.e., the Sv. Duh Formation). Thus, Barban is characterised by a greater thickness of the
479 CTB drowned platform succession (constant backfill of fine bioclastic material from the
480 surrounding areas resulted in thicker pelagic successions).

481 Early–middle Turonian global eustatic sea-level fall (Hardenbol et al. 1998; Miller et al. 2005;
482 Voigt et al. 2006; Haq 2014; Sames et al. 2016) caused a faster backfill of intraplatform
483 depressions with bioclastic material (mainly fragments of rudist colonies and benthic
484 foraminifera) from the surrounding shallow-marine areas (Gušić and Jelaska 1993; Korbar et
485 al. 2001). This process is clearly visible in the Barban section at the 136–175 m interval
486 (transition between the Sv. Duh and Gornji Humac Formations). The re-establishment of
487 shallow marine depositional environments in the late Turonian was characterised by gradual
488 recolonization of rudists and benthic foraminifera, but with lower richness of taxa and smaller
489 number of specimens relative to the Cenomanian (Velić 2007). At the Premuda section, the
490 shallow-marine deposits of the Gornji Humac Formation (the upper boundary of the Veli rat
491 Formation) are buried under the recent Adriatic Sea deposits (Fig. 1f) and are thus inaccessible
492 for direct observation.

493

494

495 *5.4. Stable isotope, TOC and insoluble residue data*

496 Chronostratigraphic calibration of Upper Cretaceous shallow-marine carbonate platform
497 deposits of the Tethyan area is commonly hindered by low-resolution stratigraphic schemes due
498 the lack of chronostratigraphic markers such as ammonites, planktonic foraminifera and

499 calcareous nannoplankton (Fleury 1980; De Castro 1991; Chiocchini et al. 2008; Velić 2007).
500 For this reason, isotope geochemistry ($\delta^{13}\text{C}$ and $\delta^{18}\text{O}$) is currently the best available
501 stratigraphic tool (Frijia et al. 2015; Brčić et al. 2017). The results of stable isotope analysis
502 (Table 1; Fig. 5) were used to improve stratigraphic interpretations and global correlation of the
503 studied deposits. The challenging aspects of this part of the research were the limited number
504 of samples, tectonically disturbed successions, and diagenetic modifications. Recrystallization
505 of carbonate mud is the main diagenetic process that impacted the carefully selected micritic
506 samples without obvious carbonate cement and skeletal fragments. Nevertheless, the study
507 produced isotope curves that show significant correlation with the informal reference curves of
508 basinal successions (Paul et al. 1999; Jarvis et al. 2006; Pearce et al. 2009; Fig. 5). During the
509 Cenomanian and Turonian there was no terrigenous input to the north-western part of the
510 isolated AdCP (Fig. 2a). Even though the interiors of large carbonate platform are commonly
511 isolated and subject to localized environmental fluctuations, global events such as OAE2
512 (Jenkyns 1980, 2010; Schlanger et al. 1987; Arthur et al. 1987, 1988, 1990; Paul et al. 1999;
513 Tsikos et al. 2004; Keller et al. 2004; Pearce et al. 2009; Jarvis et al. 2011; among others) and
514 sea level maxima (Haq 2014) were large enough in scope to overprint local influences and get
515 recorded and preserved in the depositional succession of the AdCP shallow-marine settings
516 regardless of diagenetic modifications (Jenkyns 1991; Gušić and Jelaska 1993; Hilbrecht et al.
517 1996; Davey and Jenkyns 1999; Parente et al. 2008; Immenhauser et al. 2008; Elrick et al. 2009;
518 Gertsch et al. 2010; Nagm 2015; Brčić et al. 2017). These influences are evident in the drowned
519 platform facies of the northwestern part of the AdCP (Fig. 6b), oxygen-restricted
520 paleoenvironmental conditions and intervals of carbonate factory crisis reflected in bioturbation
521 (Fig. 3), sulphate reduction, syndimentary pyrite and greenish glauconite (other sections in
522 Istria, Ćićarija, Brčić et al. 2017). The nearest previously examined OAE2 sedimentary
523 successions are carbonate-free black shales in Italy (Gubbio; Coccioni and Luciani 2005),

524 Austria (Rehkogelgraben; Wagreich et al. 2008) and Greece (Ionian zone; Karakitsios et al.
525 2007, 2010). However, the results of carbon and oxygen isotope analyses were here compared
526 to chalk deposits from Eastbourne Gun Gardens (Fig. 5) because they currently represent the
527 best reference curve for this stratigraphic interval. This reference curve records the pre-
528 excursion levels, the first build up (or peak *a*), the trough, the second build-up (peak *b*), and the
529 plateau (ending with peak *c*; Paul et al. 1999; Tsikos et al. 2004).

530 Facies variation and diagenesis may limit the reliability of $\delta^{13}\text{C}$ data from the same interval
531 within a single section (Immenhauser et al. 2008; Wendler 2013; Jarvis et al. 2015). The
532 covariance between $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ data (correlation coefficient is 0.74) observed in the Barban
533 section (Fig. 5) suggests that carbon isotope values here may have been affected by diagenetic
534 modifications (such as recrystallization in the presence of meteoric fluids; Swart and Oehlert
535 2018). Despite this diagenetic potential, the similarity with Premuda carbon-isotope record,
536 which shows less covariance with $\delta^{18}\text{O}$ values (correlation coefficient is 0.19), and the ability
537 to correlate these local sections with the global reference curve, regardless of major differences
538 in their thickness and lithology (Figs. 3 and 5), supports the application of carbon-isotope
539 stratigraphy in this research regardless of its limitations. Such application was made possible
540 by careful integration of chemostratigraphy with detailed litho- and biostratigraphic data (Figs.
541 3-5 and 7-10), and was also aided by TOC and insoluble residue data.

542 The $\delta^{18}\text{O}$ values of diagenetically modified carbonates are mainly controlled by fluid
543 composition, temperature and water/rock ratios (Brand and Veizer 1981). In most Cretaceous
544 carbonates (Scholle and Arthur 1980), the $\delta^{18}\text{O}$ data show significant depletion at CTB,
545 suggesting meteoric water influence (Hajikazemi et al. 2010). Similarly, the $\delta^{18}\text{O}$ values of the
546 Barban section are diagenetically modified, but the $\delta^{18}\text{O}$ values of Premuda section generally
547 coincide with the global trend interpreted to reflect the warmest conditions at end of
548 Cenomanian (Jarvis et al. 2011). Despite diagenetic modifications of oxygen-isotope values

549 that limit their potential as paleoenvironmental proxies, the carbon-isotope ratios of the same
550 samples are expected to be more resilient to diagenetic resetting and to more closely resemble
551 the original depositional signatures (Marshall 1992; Parente et al. 2007). Departures towards
552 more negative $\delta^{13}\text{C}$ values are interpreted as a consequence of interaction with fluids enriched
553 in ^{12}C derived from organic-matter degradation (Irwin et al. 1977). Compared to Eastbourne
554 $\delta^{13}\text{C}$ curve, both research sections show large amplitude in $\delta^{13}\text{C}$ variations. The end of peak *a*
555 in both sections is marked by a very abrupt, rapid shift to low $\delta^{13}\text{C}$ values, followed by very
556 low values between *a* and *b* peaks. A possible explanation for the observed trends is condensed
557 sedimentation. The rich carbonate production reflected in bioclastic lithotypes of the Milna
558 Formation was abruptly replaced with drowned platform facies (Sv. Duh and Veli Rat
559 Formations) of mudstones with rare dinoflagellate cysts and planktonic foraminifera. These
560 events are closely related to the sharp rise in sea level at the very end of the Cenomanian (Haq
561 2014). Similar examples of condensed sedimentation at CTB can be found locally (Brčić et al.
562 2017), but also in other parts of Tethys realm (Gambacorta et al. 2015; Wohlwend et al. 2015).
563 Furthermore, deposition on carbonate platform shoals and intraplatform basins (depths up to
564 150 m) is highly sensitive to eustatic and tectonic events as reflected, for example, in carbonate
565 turbidite facies of the Veli Rat Formation at Premuda section and bioclastic intercalations of
566 the Sv. Duh Formation at Barban section. This may account for some of the observed
567 fluctuations in $\delta^{13}\text{C}$ values (e.g., due to variable rates of organic matter respiration; Patterson
568 and Walter 1994).

569 The results of TOC and insoluble residue analyses (Fig. 5) of shallow-marine deposits from the
570 Barban section indicate low amounts of non-carbonate components (less than 1% on average
571 of clay minerals and quartz), as expected for shallow parts of isolated carbonate platforms. The
572 slightly elevated TOC values at 45 m of the Barban section correlate with the transition from
573 the first build up to the second build-up on the isotope curves (transition Milna–Sv. Duh

574 Formation; Figs. 3 and 5). A similar, but less pronounced change occurred at 81 m of the section
575 and correlates with the transition from the second isotope build-up to the plateau.
576 Stratigraphically, this transition also closely corresponds with the Cenomanian–Turonian
577 boundary and can help determine and position the CTB in the study area (Fig. 5). Correlation
578 coefficient between the TOC and carbon-isotope record at the Barban section is 0.27 (Fig. 5),
579 which suggests that post-depositional alteration in the presence of organic matter played a role
580 in diagenetic history of these deposits (Irwin et al. 1977; Oehlert and Swart 2014). Despite these
581 challenges, the integration of detailed biostratigraphy with TOC/insoluble residue and carbon-
582 isotope records helped determine and position the CTB in the study area, as well as improve
583 stratigraphic resolution through placing stage boundaries and global correlation of Barban and
584 Premuda sections. The first OAE2 geochemical imprint (positive carbon isotope values in
585 response to enhanced organic carbon burial) was detected in uppermost Milna Formation
586 deposits (i.e., the first build up, see Section 5.4 and Fig. 4), preceded by a global rise in sea
587 level at the end of the Cenomanian (Haq 2014). The strongest OAE2 geochemical imprint is
588 recorded in transition from shallow-marine to deeper-marine facies (uppermost Milna
589 Formation deposits to Sv. Duh/Veli rat Formation; peak *b*; see Section 5.4 and Figs. 4 and 5).
590 The OAE2 geochemical imprint within deeper-marine facies interval (Sv. Duh/Veli rat
591 Formation) is represented with the carbon-isotope *c* plateau phase (see Section 5.4 and Figs. 4
592 and 5).

593

594

595 6. Conclusions

596

597 1) The Milna, Sv. Duh, Veli Rat and Gornji Humac Formations reflect distinct
598 paleoenvironments that existed at the Cenomanian–Turonian boundary (CTB) in

599 western part of the Adriatic Carbonate Platform (AdCP; Barban and Premuda sections
600 in present-day Croatia). Their deposition was impacted by eustatic sea level changes
601 and syndimentary tectonics (folding and faulting) resulted in facies differentiation and
602 karstification through uplifts and lowering of individual local platform areas. The
603 Premuda section reveals such tectonic influence (initial forming of intraplatform basins)
604 in combination with sea level rise (pelagic influence).

605 2) The late Cenomanian sea-level rise lead the AdCP into a carbonate factory crisis and
606 condensed sedimentation. Bottom and top layers of the intraplatform basin facies of the
607 Veli Rat Formation and the inner platform drowning facies of the Sv. Duh Formation
608 reflect the interplay between the platform drowning and growth (accumulation and
609 aggradation of lateral shallow-marine sediment). Variations in subsidence and
610 accommodation space fine-tuned the depositional processes and stratigraphic record
611 within the research area.

612 3) Global oceanic anoxic event (OAE2) at the CTB left its mark starting from the shallow-
613 marine facies (SMF, Milna Formation), through transitional (TF), and ending in deeper-
614 marine facies (DMF, Sv. Duh and Veli Rat Formations) of the study area. The $\delta^{13}\text{C}$
615 values of Barban and Premuda sections indicate that OAE2 impacted the north-western
616 part of AdCP almost entirely in the latest Cenomanian. The observed fluctuations in
617 $\delta^{13}\text{C}$ values are evidence of condensed sedimentation and shallow-marine influence.
618 The covariance between $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ values at Barban section indicate meteoric
619 diagenesis, but $\delta^{18}\text{O}$ values of Premuda section coincide with the global trends
620 interpreted to represent the warmest conditions at the end of Cenomanian. The elevated
621 TOC values at 45 m of the Barban section correlate with the AdCP drowning and a shift
622 to very low $\delta^{13}\text{C}$ values. Integration of litho-, bio-, microfacies, and TOC, insoluble
623 residue and stable isotope data indicated a carbonate factory crisis (low sedimentation

624 rate, drowned platform facies) and geochemical OAE2 imprint at the boundary between
625 the Milna and Sv. Duh/Veli Rat Formations (the latest Cenomanian) in the eastern Istria
626 and southern Kvarner area.

627 4) Carbon-isotope values from the CTB interval of Barban and Premuda sections in
628 western AdCP represent a combination of global paleoceanographic effects, local
629 environmental factors and diagenetic alteration, and are correlative with the reference
630 curve from English Chalk (Eastbourne, Gun Gardens, England). These correlations
631 contribute towards fine-tuning and calibration of biostratigraphy based on benthic and
632 planktic foraminifera in the study area.

633 5) Detailed facies interpretations concluded that the Premuda section was located on the
634 margin of an intra-platform basin during the CTB, and the Barban section was in the
635 inner protected areas of the north-western parts of the AdCP. This research contributes
636 an example of integrating the evidence for the influence of global paleoceanographic
637 perturbations on two different (protected platform interior and intra-platform basin)
638 relatively restricted shallow-marine environments with detailed paleogeographic
639 information (emerged, shallow-marine, and drowned platform area) during the CTB in
640 the Tethyan realm.

641

642

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644

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656

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658

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991 Fig. 1: Study area. a) Global position of the study area in Croatia; b) Regional setting including
992 location of previous OAE2-related research areas; c) Local position of the Barban section; d)
993 Local position of the Premuda section; e) Orthophoto (with Digital Relief Model - DRM) of the
994 Barban section exposure (A – starting point, D – end point); f) Orthophoto (with DRM) of the
995 Premuda section exposure (A – starting point, B – end point).

996

997 Fig. 2: Paleogeographic setting of the study area. a) The wider Perimediterranean region (prior
998 to approximately 100 Mya) with paleogeographically reconstructed locations of the Barban
999 (BS) and Premuda (PS) sections within the Adriatic Carbonate Platform (AdCP) (modified after
1000 Blakey 2010, and references therein); b) Peninsula Istria during the late Cenomanian to early
1001 Turonian (Brčić et al. 2017) with location of the Barban section; c) Premuda island during the
1002 late Cenomanian to Santonian with location of the study section.

1003

1004 Fig. 3: Detailed measured stratigraphic sections at the Barban and Premuda sites, indicating
1005 rock formations names (lithostratigraphy), age, and lithologies and fossils present.

1006

1007 Fig. 4: Detailed stratigraphic sections measured at Barban and Premuda with photomicrographs
1008 of the typical facies present (SMF – shallow-marine facies, TF – transition shallow-marine to
1009 deeper-marine facies, DMF – deeper-marine facies; see Section 5), correlated (grey interval)
1010 on the basis of their carbon isotope compositions. Isotope curves were constructed by
1011 connecting all individual data points. Photomicrograph scale = 1 mm.

1012

1013 Fig. 5: Correlation of the Barban and Premuda sections with the reference Eastbourne section
1014 (Pearce et al. 2009) using stable-isotope data and benthic/planktonic foraminifera biozones.

1015 Green dotted line represents CT boundary. TOC and insoluble residue data for the Barban
1016 section are also shown.

1017

1018 Fig. 6: Block diagrams of paleoenvironmental conditions from late Cenomanian to early
1019 Coniacian in the wider research area: NWI – North-western Istria; BS – paleogeographic
1020 location of the Barban section; PS – paleogeographic location of the Premuda section (see text
1021 for details).

1022

1023 Fig. 7: Field photographs of the Barban section: **a)** Thin-bedded upper Cenomanian deposits at
1024 the western part of the Barban section (see Fig. 1e); **b)** Thin-layered transition between the
1025 Milna (shallow-marine) and Sv. Duh (with pelagic influence) Formations (oscillating
1026 transgression, see Section 5.2.; hammer for scale is 32 cm long); **c)** Radiolitid rudist biostrome
1027 of the Milna Formation (upper Cenomanian); **d)** Transition shallow-marine to deeper-marine
1028 facies with a stylolite (black dotted line) between the Sv. Duh (SD, calcisphere wackestone)
1029 and Gornji Humac Formations (GH, peloidal grainstone).

1030

1031 Fig. 8: Field photographs of the Premuda section: **a)** Radiolitid (Ra) and chondrodontid (Ch)
1032 lithostrome floatstone of the Milna Formation; **b)** Coarse lithoclasts (lth) and other shallow-
1033 marine material re-deposited in bioclastic-lithoclastic lithosome intercalated within the
1034 calcisphere wackestones in the lower part of the Veli Rat Formation; **c)** and **d)** Lithoclasts (lth)
1035 of the shallow-marine Milna Formation deposits within the deeper-marine Veli Rat Formation
1036 deposits; **e)** subvertical layers (interval between 44 and 60 m) of the Premuda section (view to
1037 the east; see Figs. 1 and 3).

1038

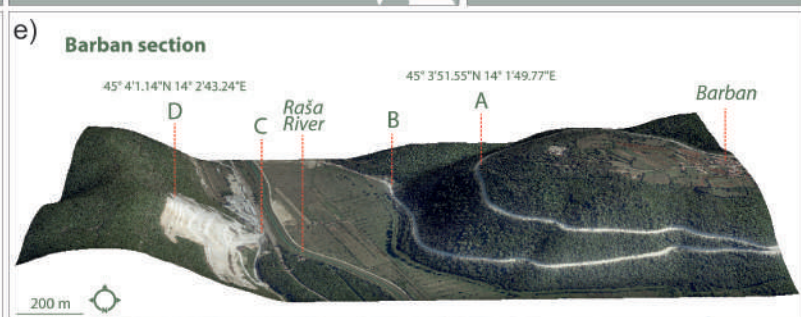
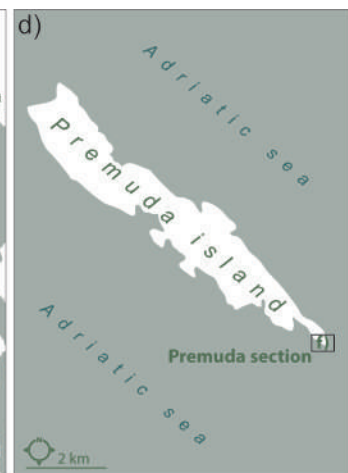
1039 Fig. 9: Photomicrographs of benthic foraminifera from the Barban and Premuda sections (scale
1040 = 0.5 mm): **a**) and **b**) *Pastrickella balcanica* (Cherchi, Radoičić and Schroeder), samples Plo-
1041 26b and BB-6; **c**) and **d**) *Heteroporella lepina* Pratulon, sample BB-32; **e**) *Heteroporella*
1042 *lepina* Pratulon, sample BB-32; **f**) *Thaumatoporella parvovesiculifera* (Raineri), sample Plo-
1043 37a; **g**) and **h**) *Cisalveolina* sp., samples Plo-16b and Plo-24 ; **i**) *Peneroplis planatus* (Fichtel
1044 and Moll), sample Plo-34; **j**) *Scandonea* sp., sample BB-11; **k**), **l**) and **m**) *Cuneolina* cf. *pavonia*
1045 (d'Orbigny), samples BB-06 and BB-04; **n**) *Vidalina radoicicae* Cherchi and Schroeder, sample
1046 BB-11; **o**) *Pseudonummoloculina heimi* (Bonet), sample BB-06; **p**), **q**) and **r**) *Chrysalidina*
1047 *gradata* d'Orbigny, samples BB-06 and Plo-03; **s**) *Pseudorhapydionina dubia* De Castro,
1048 sample BB-06; **t**) *Idalina* cf. *antiqua* (Munier-Chalmas et Schlumberger), sample BB-06; **u**)
1049 *Nezzazata* cf. *gyra* (Smout), sample BB-06; **v**) *Nezzazata simplex* Omara, sample BB-04.

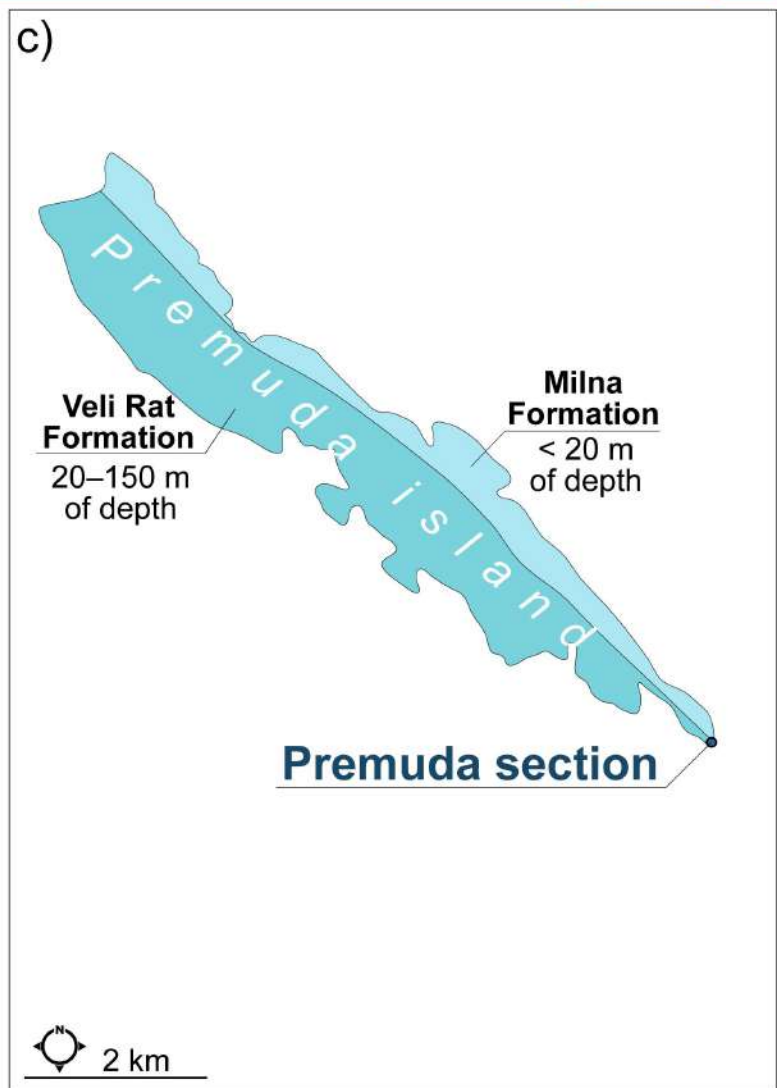
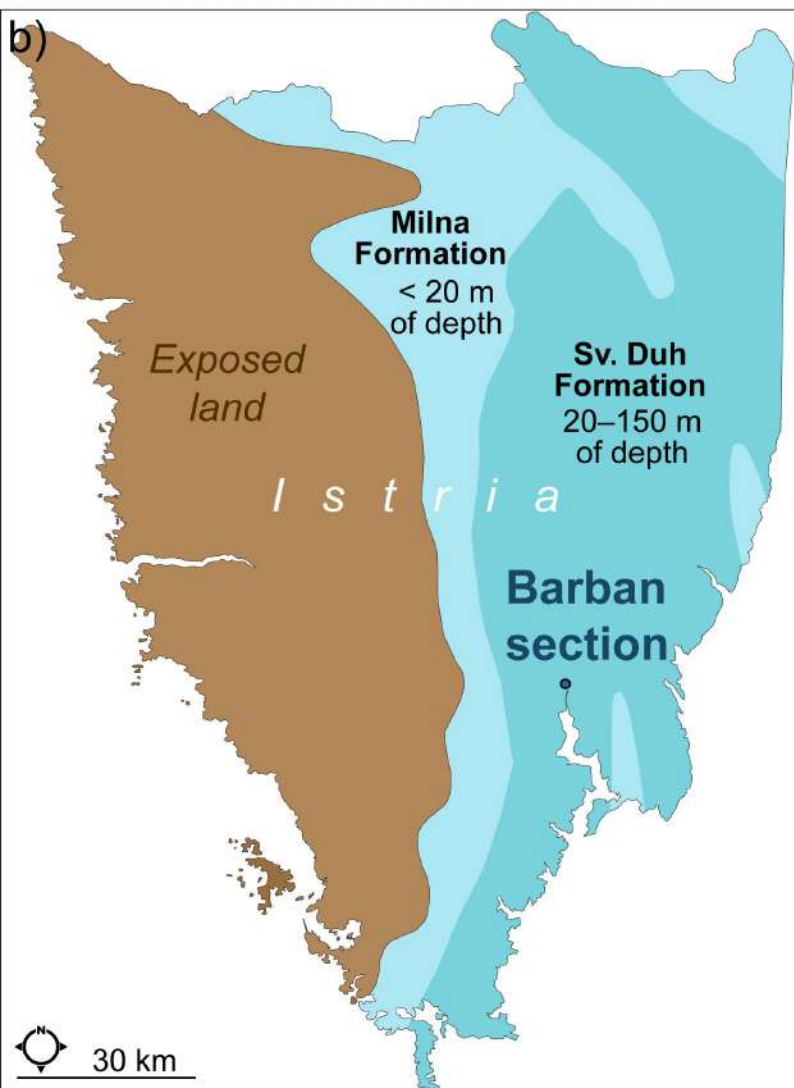
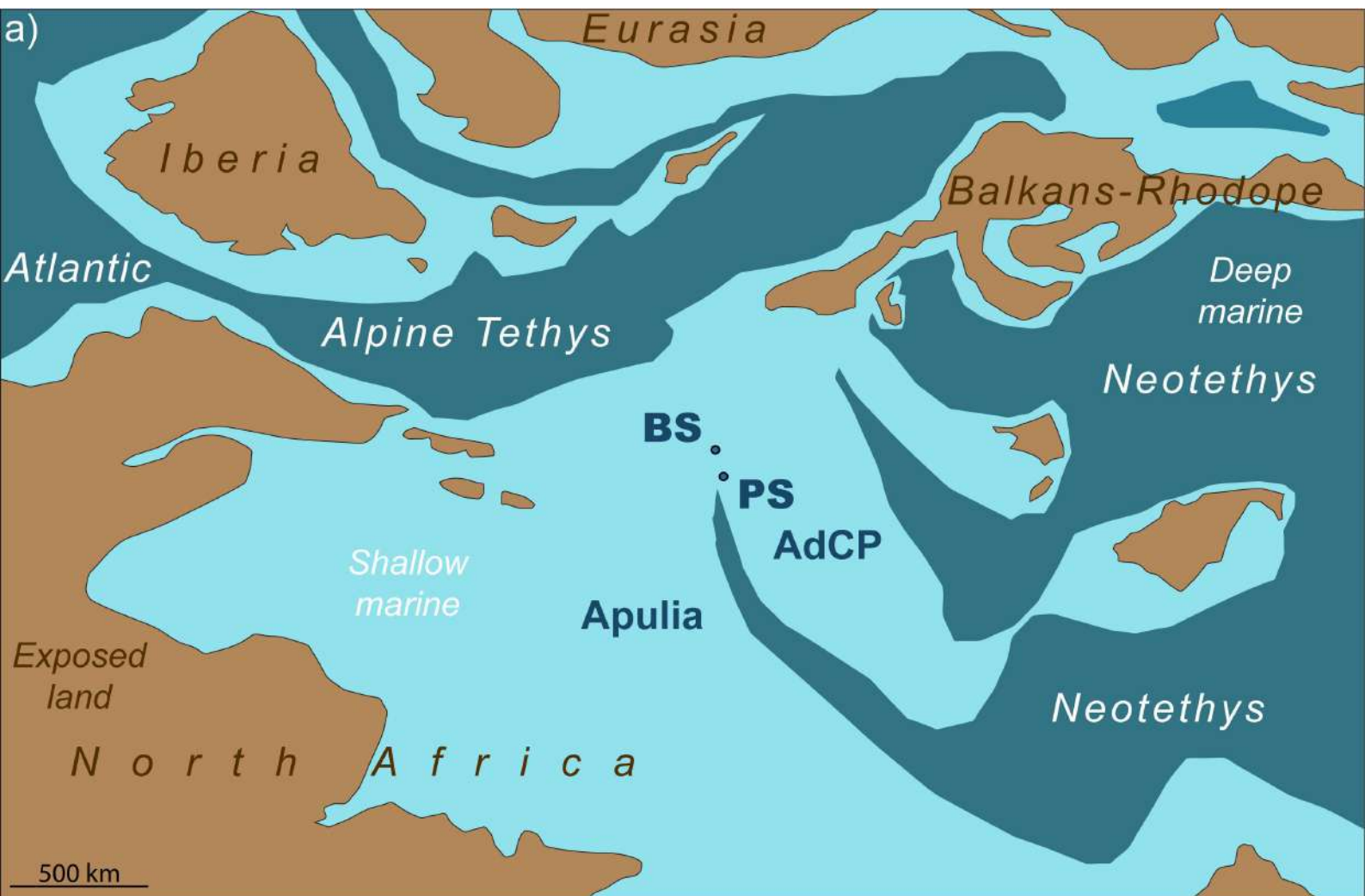
1050

1051 Fig. 10: Photomicrographs of planktonic foraminifera from the Barban and Premuda sections
1052 (scale = 200 µm): **a**) *Marginotruncana* cf. *renzi* (Gandolfi), sample Plo-72; **b**) *Heterohelix* sp.,
1053 sample Plo-71; **c**) *Marginotruncana sigali* (Reichel), sample Plo-70; **d**) *Dicarinella* sp., sample
1054 Plo-70; **e**) *Marginotruncana schneegansi* (Sigal), sample Plo-70; **f**) *Archaeoglobigerina* cf.
1055 *blowi* Pessagno, sample Plo-65; **g**) *Praeglobotruncana gibba* (Klaus), sample Plo-61; **h**)
1056 *Helvetoglobotruncana* cf. *helvetica* (Bolli), sample Plo-59; **i**) *Helvetoglobotruncana*
1057 *praehelvetica* (Trujillo), sample Plo-53; **j**) *Whiteinella* cf. *paradubia* (Sigal), sample Plo-52a;
1058 **k**) *Helvetoglobotruncana praehelvetica* (Trujillo), sample Plo-52a; **l**) *Dicarinella imbricata*
1059 (Monrod), sample Plo-51; **m**) *Whiteinella* cf. *archaeocretacea* Pessagno, sample Plo-42b; **n**)
1060 *Helvetoglobotruncana praehelvetica* (Trujillo), sample Plo-37b; **o**) *Whiteinella* cf. *paradubia*
1061 (Sigal), sample Plo-22.

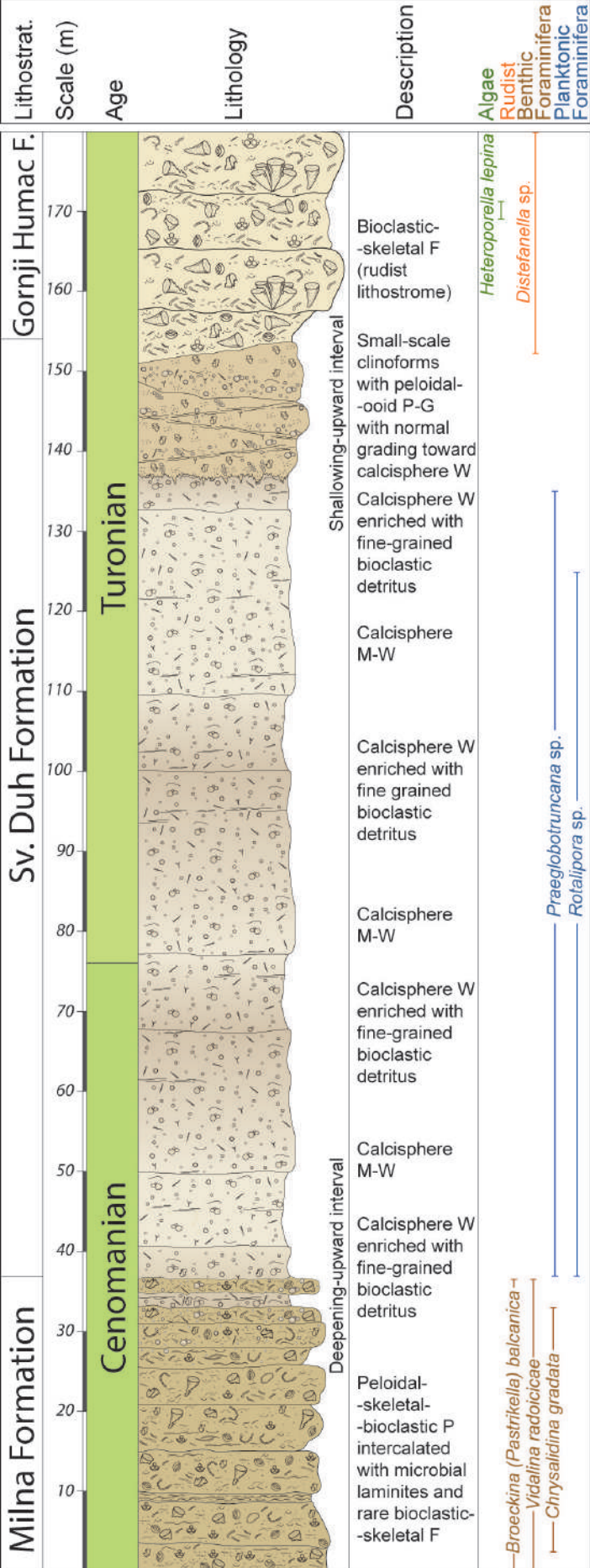
1062

1063 Table 1: Results (values) of stable isotope analysis of 122 samples (81 from Barban and 41
1064 from Premuda section) and TOC and Insoluble Residue Analyses (81 samples from Barban
1065 section). - *Appendix/supplement*

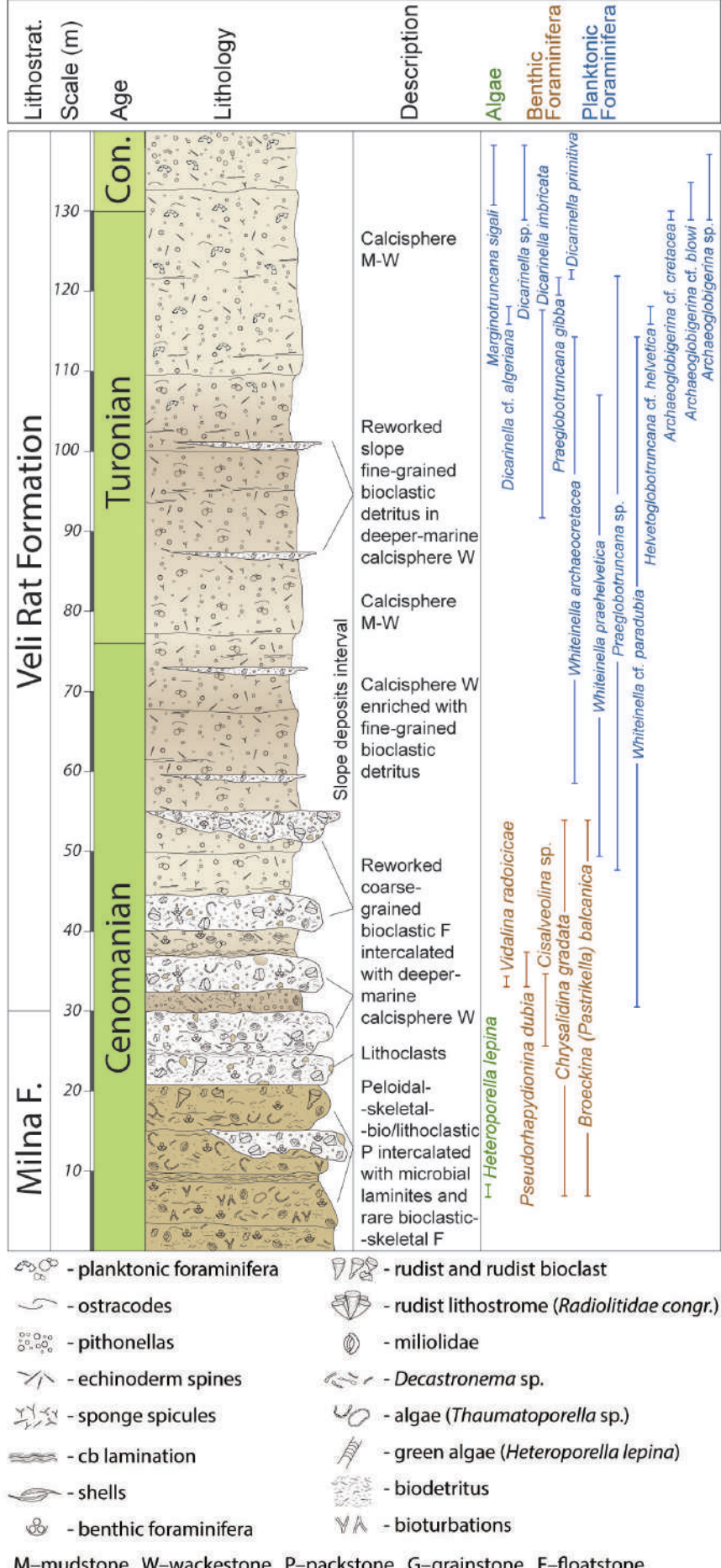




Barban section

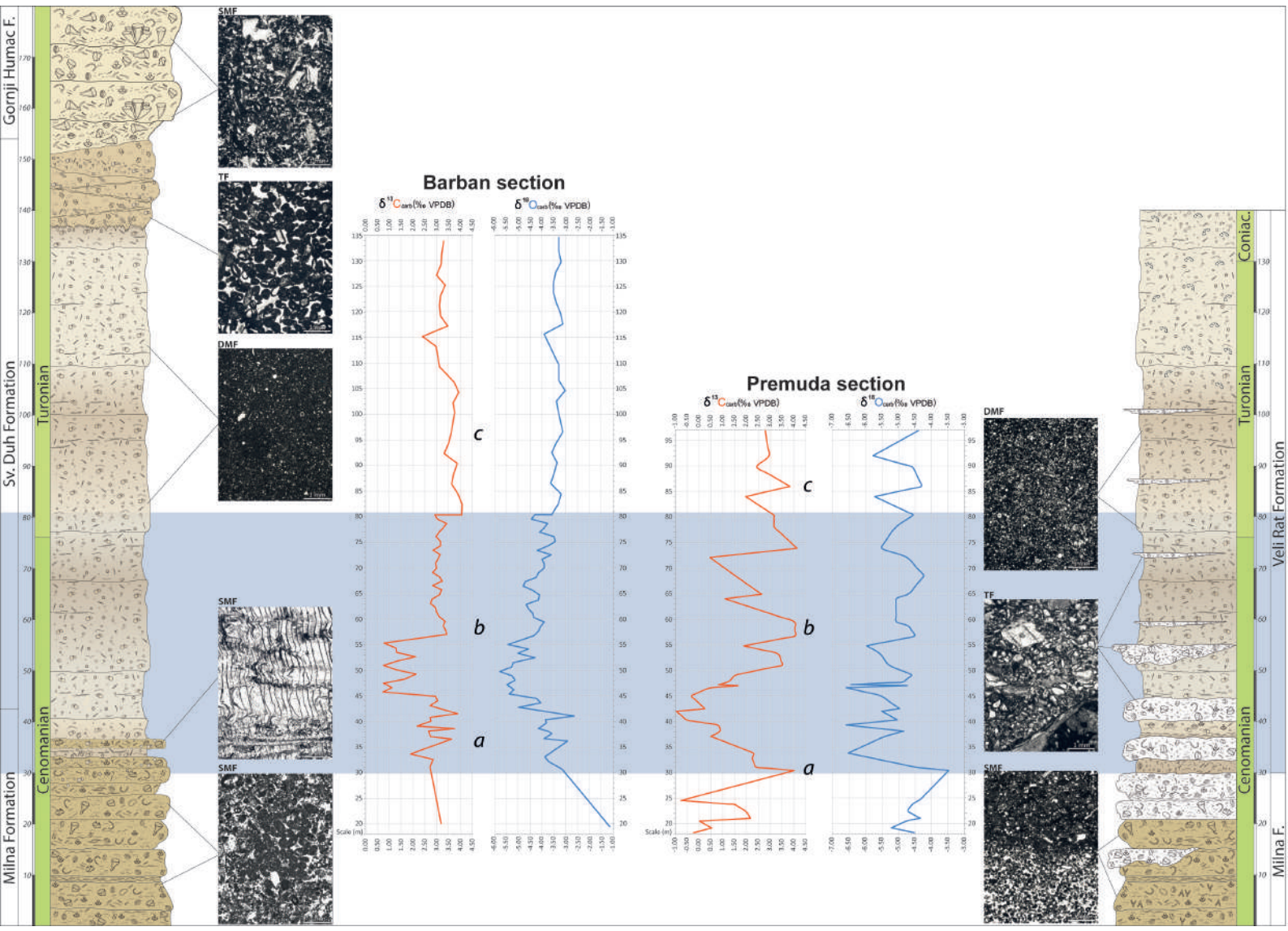


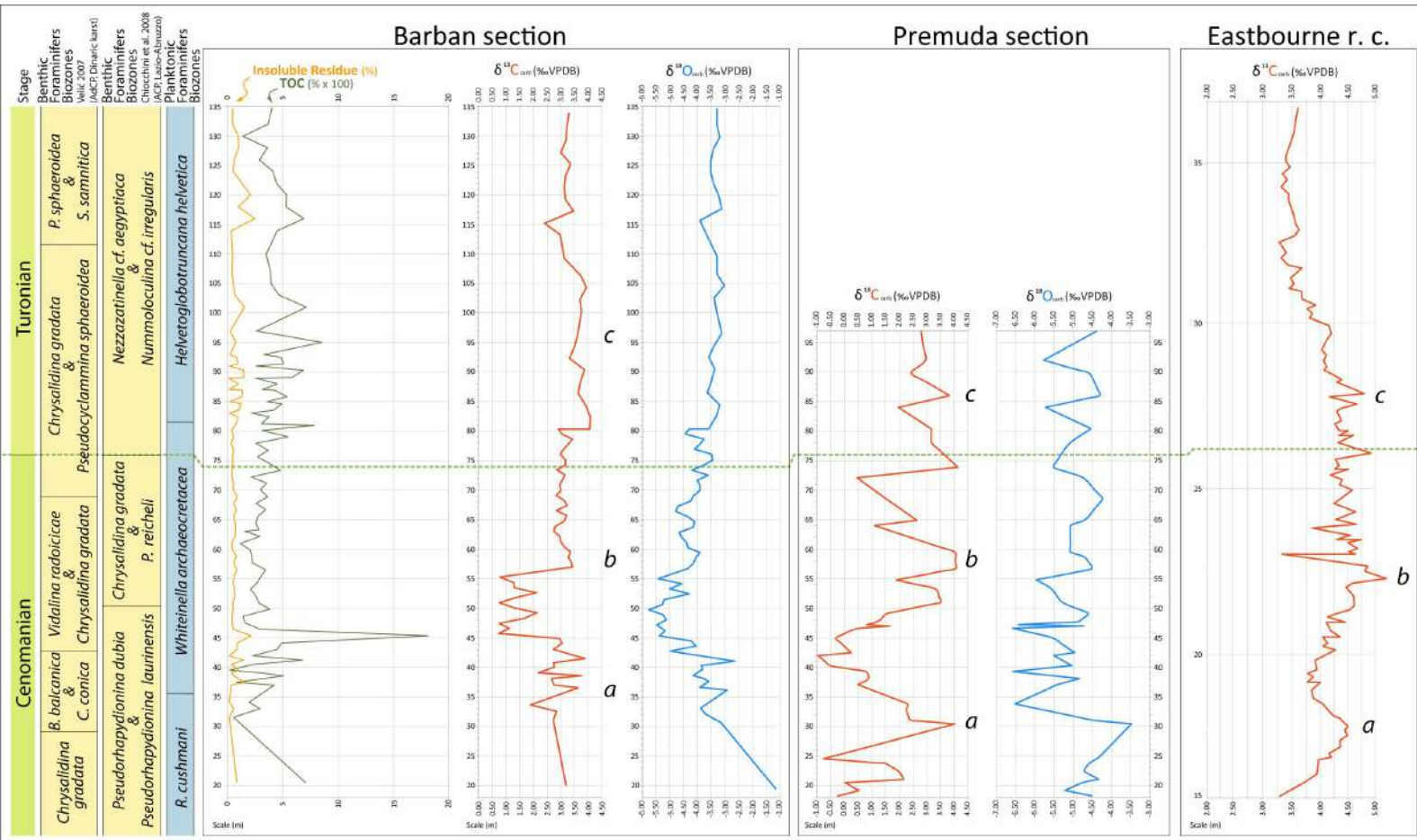
Premuda section



- planktonic foraminifera
- ostracodes
- pithonellas
- echinoderm spines
- sponge spicules
- cb lamination
- shells
- benthic foraminifera
- rudist and rudist bioclast
- rudist lithostrome (*Radiolitidae* congr.)
- miliolidae
- *Decastronema* sp.
- algae (*Thaumatoporella* sp.)
- green algae (*Heteroporella lepina*)
- biodetritrus
- bioturbations

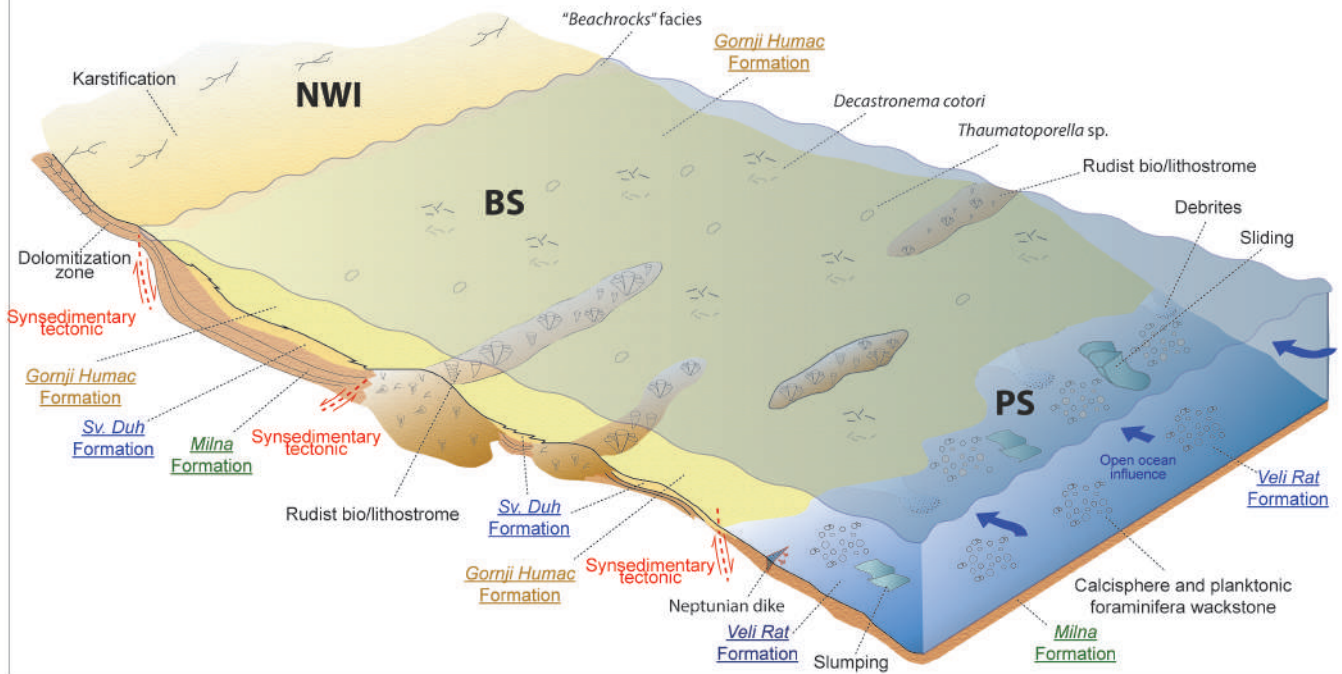
M-mudstone W-wackestone P-packstone G-grainstone F-floatstone





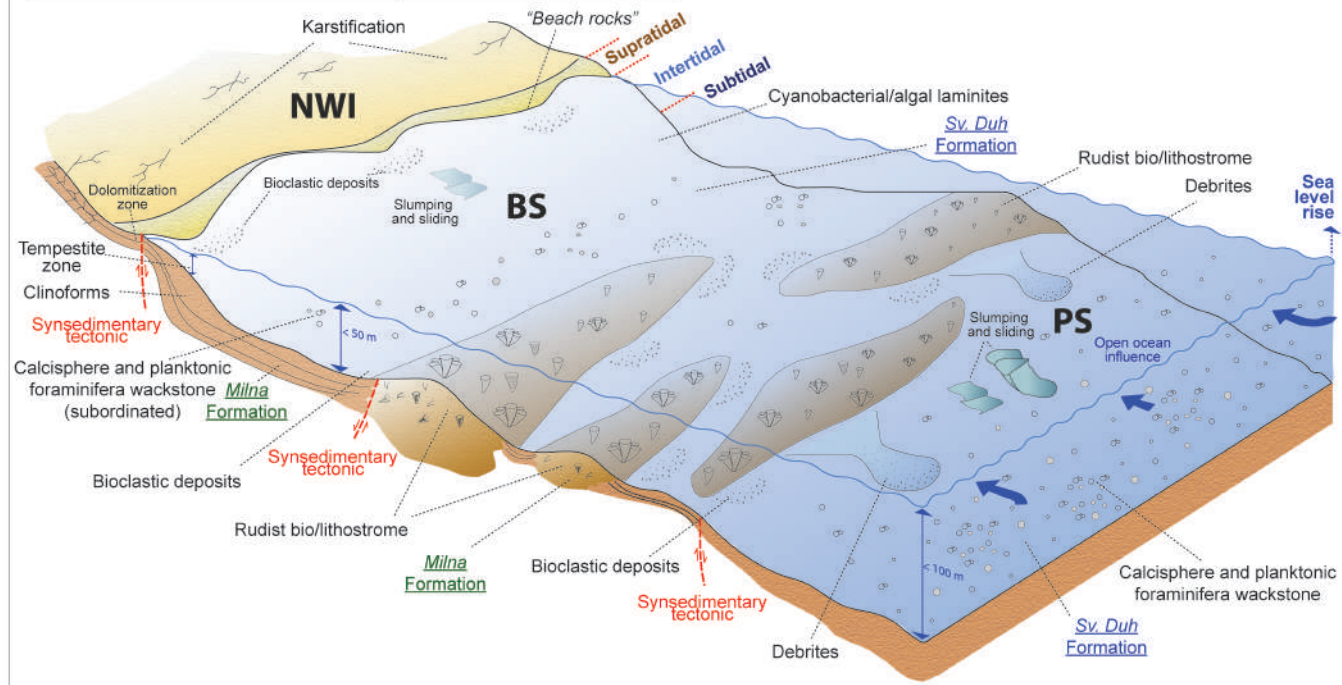
Turonian to Coniacian (90-85 Ma)

c)



Late Cenomanian to Early Turonian (94 Ma)

b)



Middle to Late Cenomanian (96 Ma)

a)

