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1           LARGE-SCALE BOUDINAGE OF LATE MIOCENE PLATFORM SERIES  
2 TRIGGERED BY MARGIN COLLAPSE DURING THE MESSINIAN SALINITY CRISIS  
3 (IBIZA ISLAND, SPAIN)

4  
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13  
14 **ABSTRACT**

15           In Ibiza Island, we show for the first time that the post-tectonic late Miocene deposits are locally  
16 deformed in synsedimentary large scale boudinage structures. The late Miocene series is composed of a  
17 carbonate rich Unit 1 that is overlaid by a clastic Unit 2 and a Unit 3 in which a basal oolite-rich deposit is  
18 distinguished from an upper stromatolite-rich part. The oolite-rich and competent layers of Unit 3 are dissected  
19 into large rectangular to barrel-shape boudins that indicate a multidirectional extension. The separation of the  
20 blocks develops significantly northward revealing a stretching increase in a seaward direction that could have  
21 been favored by sliding of the distal part of the underlying marly unit over the clays. Along and in between the  
22 boudins, the laminated layers form soft-sediment syncline folds. The progressive bending of these laminated  
23 layers created spaces that were filled with calcite crystal wedges, providing evidence for large fluid circulation  
24 during deformation. Microthermometry on fluid inclusions in these calcite crystals revealed growth-  
25 temperatures ranging from 190 to 260°C. This high recrystallization rate is an indication of important fluid  
26 circulation during deformation processes. Besides, the progressive infill of paleosols over some of the  
27 synsedimentary synclines indicates that deformation was a slow process during the emersion of the series. And  
28 the subaerial Messinian channels location in the topographic depressions formed by the boudinage of the late  
29 Miocene series confirms that the deformation and sliding occurred during the relative emersion associated with  
30 the Messinian Salinity Crisis. We thus interpret the boudinage as resulting from the destabilization of a still soft

31 carbonate platform whose sliding has been helped substantially by the injection of relatively hot fluids. This  
32 work provides an onland evidence for extensional collapse during the Messinian sea-level drawdown.

33

34

35 Keywords: Ibiza, late Miocene, Messinian Salinity Crisis, boudinage, soft-sediment deformation.

36

## 37 **1. Introduction**

38 In Ibiza Island (Figs. 1 & 2), the late Miocene series are mostly composed of calcareous rocks  
39 (Rangheard, 1971; Durand-Delga et al., 1993) and affected by synsedimentary structures recognized by Simó  
40 Marfa (1982). A succession of gentle folds and segmented layers (boudinage) deforms the late Miocene series  
41 that can be observed in the north-eastern part of the island, around Portinatx (Fig. 3). These sedimentary rocks  
42 are interpreted in the Messinian Salinity Crisis (MSC) context occurring at the end of the Miocene, and  
43 compared to well-known deposits processes and coeval to the crisis. The Ibiza late Miocene limestones are well  
44 exposed. They were deposited in the platform topographic lows. They appear to be poorly affected by thin-  
45 skinned tectonics compared to the Mesozoic to middle Miocene formations elsewhere on the island that recorded  
46 the Betic orogeny.

47 The structures around Portinatx may result from “soft-sediment deformation”. The soft-sediment  
48 deformation mainly results in the relative displacement of grains in unconsolidated sediments (Allen, 1982;  
49 Maltman, 1984). Nevertheless, recognition of soft-sediment deformation processes and trigger mechanisms is  
50 often a difficult process (Moretti and Sabato, 2007), particularly in carbonate rocks where the eodiagenetic  
51 cementation can occur very early after the period of deposit. In that case, fluid circulations or water escape  
52 structures indicate the occurrence of soft-sediment deformation (Lowe, 1975; Leeder, 1987; Owen, 1987). In the  
53 Miocene sediments, around Portinatx, the widespread recrystallization of the limestones and the great number of  
54 calcite veins in interbedded positions give evidence of such large circulation of fluids but it is not any proof of  
55 cold synsedimentary conditions. The comparison of the converging pictures of metamorphic rock structures and  
56 soft-sediment deformation structures (Ogata et al., 2016) provides substantial help to describe the structures.  
57 The complete lack of cleavage, of veins in the fold hinges and growth of metamorphic minerals, are clear  
58 indications of the occurrence of soft-sediment deformation. It is the case for the late Miocene limestones that  
59 have been gently deformed, but the conditions of the deformation structures have not been clearly defined.

60           The ductile deformation structures of the late Miocene series provide an opportunity to investigate the  
61 potential relationship between early deformation of the platform and the sea-level drawdown during the  
62 Messinian Salinity Crisis. We investigate those structures in order (i) to understand the formation of atypical  
63 boudinage and folding, (ii) to recognize the occurrence of soft-sediment deformation, (iii) to propose triggering  
64 mechanisms, (iv) to possibly relate them to the MSC event.

65

## 66 **2. Geological context and methodology**

### 67 2.1 Regional geodynamics of the Balearic Promontory

68           The Balearic Promontory is a continental rise (500 km-long, 120 km-wide, Fig. 1) that includes the  
69 Balearic Islands, surrounded by narrow platforms and steep slopes toward the basins. Although located between  
70 two extensional basins, the Valencia Basin to the north and the Algerian Basin to the south, the Balearic Islands  
71 result mostly from the compressional deformation associated with the Betic orogenic system that has been active  
72 by parts from Eocene to Miocene (Rangheard, 1971; Ramos Guerrero et al., 1989; Sanz De Galdeano, 1990;  
73 Roca, 2001) and belong to the External zones of the mountain range. The Betic thrusts are well expressed on  
74 Ibiza and Mallorca (Fourcade et al., 1982; Gelabert et al., 1992, 2004; Sàbat et al., 2011) as well as offshore in  
75 the Valencia Basin, in the lower slope domain of the islands (Maillard et al., 1992; Mauffret et al., 1992; Sàbat et  
76 al., 1997, 2011; Ethève et al., 2016). In Ibiza, Ethève et al., (2016) show that the extensional event responsible  
77 for the formation of the surrounding Valencia basin predates the Betic compression (middle Miocene).

78           Actually, the Balearic Promontory underwent an extensional event from the late Serravallian to  
79 Tortonian, notably expressed in the Palma graben in Mallorca (Roca and Guimera, 1992) and in offshore basins,  
80 possibly linked to post-orogenic collapse or / and to the westward translation of the Alboran domain, along the  
81 E. Baudot transform margin (Fig. 1) (e.g. Frizon de Lamotte et al., 1991; Mauffret et al., 2001; Platt et al., 2003;  
82 Platt et al., 2013; Driussi et al., 2015). Moreover, although the borders of the Balearic Promontory are considered  
83 as passive margins, gentle active deformation has been reported in the Alicante shelf, and in the Ibiza and  
84 Mallorca channels (Sabat et al., 1997; Alfaro et al., 2002; Maillard and Mauffret, 2013). In the Ibiza Channel, the  
85 abundance of mass failure structures also supports the occurrence of active tectonic processes and/or volcanism  
86 (Acosta et al., 2002; Lastras et al., 2004; Acosta et al., 2013). Cenozoic succession of compressional and  
87 extensional tectonic events is thus responsible for the complex structural setting of that area.

88

### 89 2.2 Mio-Pliocene context of Ibiza Island



90 Ibiza Island is mainly composed of Mesozoic rocks separated from the overlying marine late Oligocene  
91 conglomerate and Langhian nanno and planctonic rich marly turbidites (Durand-Delga et al., 1993; Rangheard et  
92 al., 2011). These Cenozoic deposits are involved in the orogenic deformation (Fourcade et al.,1982; Fontboté et  
93 al., 1990; Sàbat et al.,1997). In the NE part of Ibiza Island, the postorogenic transgressive deposits are late  
94 Miocene, composed of 3 marine units described by Simó-Marfa (1982) and attributed to the Tortonian with  
95 some uncertainties (Rangheard, 1971; Durand-Delga et al., 1993; Rangheard et al., 2011). It is topped by an  
96 erosion surface covered by marine then aeolian Plio-Pleistocene deposits (Simó-Marfa, 1982; Lezin et al., 2014;  
97 2017). This erosion surface is related to the one affecting most of the Mediterranean margins, called the Margin  
98 Erosion Surface (MES) related to a high-amplitude drop of the sea level during the Messinian Salinity Crisis  
99 (Lofi et al., 2011). This short-term dramatic event occurred between 5.96 and 5.32 Ma (Gautier et al., 1994;  
100 Krijgsman et al., 1999). It resulted first in the deposit of evaporites in “shallow” peripheral basins (Primary  
101 Lower Gypsum); then evaporite precipitation shifted to the abyssal basins (“deep basin evaporites”), while  
102 subaerial erosion affected shallow areas and Mediterranean margins (CIESM, 2008; Ryan, 2008; Roveri et al.,  
103 2014). The MSC sea-level drawdown is thus responsible for incisions of deep fluvial canyons (Hsü et al., 1973;  
104 Clauzon et al., 1996; Rouchy and Caruso, 2006; Lofi et al., 2011), large transfer of sediments and margin  
105 collapse (Bache et al., 2009; Maillard et al., 2014; Cameselle and Urgeles, 2017). Unlike in Mallorca Island  
106 where Primary Gypsum in the Palma Bay reveals a shallow peripheral basin, there are no MSC-related  
107 evaporites in Ibiza Island but the crisis is recorded by the MES, that extends throughout the offshore Balearic  
108 margins (Fig. 2) until it turns to the deep basin evaporites in surrounding Valencia and Algerian abyssal basins  
109 (Driussi et al., 2014).

110

### 111 2.3 Methodology

112 High-resolution stratigraphic sections have been made in the different calas of NE Ibiza. In each  
113 section, the different units have been recognized and correlated over the studied area. A specific paper devoted to  
114 the high-resolution stratigraphy of the Miocene-Pliocene units is in preparation (Lezin C., personal  
115 communication). In the present paper, we show a synthetic stratigraphy of the area in figure 4. At the same time  
116 and over the same outcrops, structural observations and measurements have been performed (fold axes  
117 orientations, small scale fracture measurements) and they have always been related to the corresponding  
118 stratigraphic unit.

119 Numerous calcite veins are observed in each outcrop. They permitted observation of fluid inclusions  
120 trapped in the crystals. Microthermometric measurements were performed to obtain homogenization  
121 temperatures that are considered to be representative of the trapping temperature conditions (Tilhac et al., 2013),  
122 thus giving an estimation of the temperature of the fluid that circulated. Double polished sections (~ 300 µm  
123 thick) of calcite crystals were prepared following Goldstein's (2003) recommendations. Homogenization  
124 temperature (Th) measurements were performed at the Laboratoire de Géologie de Lyon (LGL) on a Linkam  
125 THMS-600 heating-freezing stage connected to a programmable thermal control unit. Pure CO<sub>2</sub> fluid inclusions  
126 in quartz from Camperio, Central Alps (Poty and Stalder, 1974; Mullis et al., 1994) were used for calibration at  
127 CO<sub>2</sub> triple point (-56.6°C) and clathrate fusion (10.0°C) temperatures, and synthetic pure H<sub>2</sub>O fluid inclusion in  
128 quartz for calibration of H<sub>2</sub>O triple point (0.2°C) and critical point (375.13°C) temperatures. The accuracy of the  
129 measurements is estimated to be ± 1.0°C. Homogenization occurred in the liquid phase.

130

### 131 **3. Results**

#### 132 3.1 Late Miocene tectono-sedimentary evolution of the NE of Ibiza Island.

133 Miocene-Pliocene units outcropping on the NE part of the Ibiza Island (Figs 1 & 2) were studied to  
134 establish their sedimentological and stratigraphic characteristics, to propose a tectono-sedimentary evolution and  
135 try to link them with the MSC event. The late Miocene (Fig. 4) is firstly composed (Unit 1), from base to top, of  
136 fossil-rich limestones (benthic foraminifera, corals, rhodoliths) accumulated in shallow marine environments.  
137 This sedimentary Unit presents some similarities with the Heterozoan unit previously described in Sorbas and  
138 therefore could be late Tortonian to lower Messinian in age (Bourrillot et al., 2010). In Portinatx area, it lies over  
139 the Mesozoic series. Some soft-sediment deformation structures have been observed at the top of Unit 1 at Punta  
140 Xarraca, where sandy to pebble rich clastic sediments have been injected to form clastic dykes in Unit 1  
141 laminated limestones. Such soft-sediment deformation has been recognized as being triggered by earthquakes  
142 (Moretti and Sabato, 2007).

143 A detrital unit (Unit 2) formed by gravitational and fluvial processes in the nearshore environment  
144 overlays Unit 1. The edification of the coastal alluvial fans has been controlled by extensional faults that delimit  
145 the late Miocene distribution in the north-eastern Ibiza Island (Fig. 2).

146 Oolite, microbialite (thrombolite, stromatolite) -rich limestones (Unit 3) locally affected by ductile  
147 deformation onlap Unit 2. Usually, oolitic grainstones with herringbone structures and trough cross-  
148 stratifications are attributed to the lower part of Unit 3 (Unit 3-1) while microbialites dominate limestone

149 composition in the upper part (Unit 3-2). Unit 3 is interpreted as the TCC (Terminal Carbonate Complex) that is  
150 well known elsewhere in onland Spain (Bourillot et al., 2010; Clauzon et al., 2015). These three units were  
151 deposited in shallow water or nearshore environment. Locally, the TCC is covered by paleosols, wind dunes and  
152 mud/debris flows recording an onland deposition. A strong erosion of the Miocene and Mesozoic underlying  
153 series occurred after the deposition of the TCC. The important erosion and karstification surface locally shapes  
154 the valleys building the “calas” (local name for long and narrow gulfs) that reflect the present-day landscape  
155 (Fig. 3). This erosion and associated gravity flows are interpreted as evidence of emersion of the series (Lezin et  
156 al., 2017).

157         The erosion surface is filled with shallow marine calcarenites with gastropods, benthic foraminifera, and  
158 red algae that constitutes Unit 4 lying unconformably on Unit 3. It is attributed to the Zanclean Stage (early  
159 Pliocene) (Lezin et al., 2017). Aeolian bioclastic sands dated from Pleistocene (Del Valle et al., 2016) cover  
160 these deposits. Together with emersion evidence, we can undoubtedly relate the erosional surface between Unit 3  
161 and Unit 4 to the MSC relative sea-level fall, the MES.

162

### 163 3.2. Folds and other deformational structures of the Mio-Pliocene series of Ibiza

164         Around Portinatx (Fig. 3) along Aigua Blanca cliff, a large succession of gentle folds can be observed in  
165 Unit 3 (Fig. 5). They have been recognized as synsedimentary structures (Simó Marfa, 1982) though not studied  
166 nor explained. Our measurements of the fold axes orientations permitted to group them in two main trends:  
167 065°N and 140°N (Fig. 3). Due to the low eastward plunge of the series, the lower, middle and upper parts of the  
168 series are successively exposed from west to east. They correspond respectively to the western, the “Moscarter  
169 lighthouse” and the eastern areas presented below.

170

#### 171 3.2.1. The western area

172         On Punta Galera Island, the exposed series are composed of limestones Unit 1 top, white clay Unit 2  
173 and limestones Unit 3 (Fig. 6). The series is folded into a large and open syncline trending N 175° and resting on  
174 flat-lying Unit 1. In between, the argillaceous part of Unit 2 forms a wedge thickening toward the west. On the  
175 west side of the syncline, at the base of Unit 3, numerous soft-sediment deformation structures were observed.  
176 On vertical sections (Fig. 7a and b), they look like load structures with balls of grain rocks sinking into the  
177 argillaceous sediments of Unit 2. Whereas classical load structures show usually a regular and circular  
178 morphology (Moretti and Sabato, 2007), the structures of Punta Galera are linear like long ripples (Fig.7). The  
179 long axis of each load structure is orientated to the north, parallel to the syncline axis, suggesting the

180 development of the soft-sediment deformation structures during folding. Few metres above (Fig. 6c), calcite  
181 geodes, lined up at the bed top and wrapped in reworked carbonate sediments, indicate the occurrence of fluid  
182 circulation in the lower part of Unit 3.

183 In Cala Fondo area (Fig. 8) the exposed series shows the flat top of Unit 1 made of limestones, a  
184 reduced part of Unit 2 composed of white clay and topped by a strongly brecciated and partly recrystallized  
185 (neomorphism) limestone bed that can be attributed to the base of Unit 3. The outcrop outlines a large and open  
186 fold, trending N 25°, formed in the brecciated limestones and marls (Unit 2 and 3). Unit 2 marls are only a few  
187 metres thick near the syncline axis but they are thicker on both the western and eastern edges of the opened  
188 syncline. Some beds of black highly diagenetized limestones (recrystallization) are broken by small-scale  
189 fractures filled with red clays and that are trending either in the N-S or in the E-W directions.

190

### 191 3.2.2. The Moscarter lighthouse area

192 One of the most spectacular folded structure is exposed along the cliff, immediately east of the  
193 Moscarter lighthouse (Fig. 9). At that place, only Unit 3 is exposed, its base being under the present sea level.  
194 The calcareous herringbone beds form large rectangular blocks, about 100 m wide and approximately 15 m  
195 thick. The blocks appear to be little deformed but only moved apart like stretched rectangular boudins  
196 (Wegmann, 1932). Between the blocks, the laminated upper layers of Unit 3 are continuous and form scar folds  
197 made of a tight syncline between two open half anticlines (Hobbs et al., 1976; Goscombe et al., 2004). From  
198 their horizontal position over the blocks, the layers are laterally curved to a steep position between the block  
199 corners where limbs have been tilted up to 60° (Fig. 9). The tight synclines between blocks have developed in  
200 the laminated layers of Unit 3. In numerous places, these laminated layers have been totally transformed by  
201 diagenesis, resulting in the development of mm-scale calcite crystals (Fig. 10). In these specific areas of  
202 recrystallized layers, small fractures are formed perpendicular to bedding and are filled with red argillaceous  
203 sediments (Fig. 9c). These fractures develop close to tight syncline and are orientated from N 125° to N 145°,  
204 parallel to the fold axis. The laminated layers (Unit 3.1) and the oolitic grainstones are partially affected by  
205 diagenetic transformations. Transformations intensity increases towards the syncline axis. Simó-Marfa (1982)  
206 observed and described these transformations in the field and under the microscope. In the field, in the first stage  
207 of transformation, the color of the limestones is little affected but the sedimentary structures and the components  
208 are not visible and pores are filled with large crystals of calcite (Fig. 10 d-e). At an advanced stage, the  
209 limestones acquire a greyish color and a massive appearance. Black colored large crystals (1 to 3 cm) mark the

210 extreme case. Under the microscope, we observe in the first case a neomorphic reorganization allowing  
211 recognizing some grains. In both the advanced and extreme stages, a large amount of poecilitic cement destroys  
212 grains and sedimentary structures.

213 West of Moscarter lighthouse (Fig. 3), the large structure of the west Moscarter syncline (Fig. 11a)  
214 shows that the laminated layers (Unit 3-2), above the herringbone beds (Unit 3-1), are folded into two half  
215 anticlines with horizontal axes orientated N136° and N020° for the western and eastern folds respectively.  
216 Because of limited accessibility, the tight syncline between the two half anticlines cannot be precisely described.  
217 If the tight syncline can link the two non-parallel anticlines, it is probably not a cylindrical fold. Over the  
218 laminated beds, the bending of the layers progressively reduces upwards with the progressive onlap of paleosol  
219 layers that fill the depression (Fig. 11b). These paleosols are composed of reworked marls and are probably late  
220 Miocene in age (Lezin et al., 2017). The deformation is high on the limbs of the west Moscarter syncline (Fig.  
221 11), where extensional deformation can be recognized from pinch and swell structures that have deformed both  
222 the laminated layers and the clastic sediments injected from the top into the layers. The extensional brittle  
223 deformation of the clastic sediments has formed small-scale normal faults with a stretching direction parallel to  
224 the layers (Fig. 11d). Some secondary folds, as shown in Figure 11c, have been estimated to have approximately  
225 N105° axis directions.

226 In the Aigua Blanca area (Figs. 5 and 12), the laminated layers are bent around the corners of the boudins;  
227 the lower layers being highly tilted. As a result, space is opened between the layers and filled with calcite  
228 wedges. The vertical orientation of the crystals shows that they have not grown in a completely open geodic  
229 volume but that they have progressively grown vertically in the space opened between the layers (Fig. 12).

230 In some places, such as over the bowl syncline located east of Portinatx, red clays filling numerous  
231 fractures or sedimentary dykes inject the upper layers (Fig. 12, area located on Figs. 3 & 5). The different  
232 intersections of the dykes are perpendicular or oblique but not any prevailing extensional direction has been  
233 recorded even after top view survey from a drone.

234

### 235 3.2.3. The eastern area

236 Punta des Gat is close to the distal part of the Cala d'en Serra section (Figs. 3 and 13) where the  
237 thickness of Unit 2 is highly reduced due to its facies variations from thick-bedded coarse conglomeratic  
238 deposits to the south (nearshore deposits) to thinner white clay deposits to the north (offshore deposits).  
239 Accordingly, sections in the Punta des Gat area display mainly the flat upper part of Unit 1 and some deformed

240 and strongly stretched layers of Unit 3 (Fig. 13). Coral rich carbonate Unit 1 shows evidence of neomorphism  
241 (recrystallization) and dissolution. Unit 3 is marked by the local presence of dissolution breccias and mineral  
242 floors. Both units are separated by some strongly deformed layers composed of mixed white and green clays,  
243 black recrystallized limestones, reddish limestones blocks, fragmented limestones separated by red clays (Fig.  
244 14). The fractures range in strike from N 162° to N 050° but most of them are approximately N-S. Some of the  
245 clays may have a decalcification origin and the limestones correspond to the deformed and highly stretched part  
246 of Unit 3. On top of these deformed areas, lenses of Unit 4 material have been deposited. They are up to 10 m  
247 thick and they have been locally dated to the Zanclean Stage from microfaunal assemblages (Lezin et al., 2017).

248 In some places from the Punta des Gat area, clastic dykes and numerous veins filled with calcite crystals  
249 are observed within Unit 3. The veins, up to 20 cm thick, are always parallel to the layers and are filled with  
250 vertical calcite crystals. They are very similar to the calcite wedges observed west of the Moscarter lighthouse  
251 (Fig. 12). Fluid inclusions trapped in calcite have been observed in veins at different outcrops. Petrographic  
252 study shows that these inclusions are primary inclusions trapped during crystallization of calcite and growing of  
253 the veins. They present aqueous liquid and vapor phases at room temperature with a homogenous liquid/vapor  
254 ratio indicating the trapping of a single-phase aqueous fluid at trapping conditions. It was not possible to  
255 determine salinity of the fluid because inclusions are too small and calcite not transparent enough.  
256 Homogenization temperatures of the fluid inclusions could be measured. Only samples coming from the Cala de  
257 Ponent give clear results (Table 1). They indicate calcite crystallization between 190 and 260°C, relatively hot  
258 temperature for minerals formed under a relatively thin sedimentary cover. It is an indication of hot fluids  
259 circulation that can facilitate ductile deformation. In the Punta des Gat area, the three Miocene sedimentary units  
260 are very deformed under various processes from early diagenetic conditions to hot fluid ductile mechanisms.  
261 Unit 4 is not affected by these processes, either the deformation or the calcite veins formation.

262

## 263 **4. Interpretation**

### 264 4.1. A synthetic model of large-scale boudinage

265 From field observations, the structures of the Mio-Pliocene series of Ibiza appear to be organized in  
266 three areas. In the western area, folds involve the lower part of the series: they are narrow anticlines between  
267 larger opened synclines (Figs 6 and 8). Soft-sediment deformation observed at Punta Galera is an indication that  
268 folding started as soon as the deposit had occurred (Fig. 7). In the Moscarter area, only the mid and upper parts  
269 of the series can be observed, Unit 3-1 and Unit 3-2 respectively. The Unit 3-1 is mainly broken into large

270 blocks. The folds that develop in Unit 3-2 are narrow synclines between larger half anticlines. In the fold limbs,  
271 extensional deformation is observed from small scale normal faults (Fig. 11d) or tensile fractures formed under  
272 mode 1 opening conditions (Fig. 9c). Compressional structures have never been observed on the folds; it is an  
273 indication that there is not any neutral surface in the mid part of the folded layers and that folds have formed by  
274 bending instead of buckling (Fossen, 2010). In the Moscarter area, the bending of the laminated layers occurred  
275 during the progressive sedimentation of the paleosol layers (Fig. 11) and it stopped by the time of deposition of  
276 the upper carbonate. In the eastern area, the series has been highly stretched (Fig. 13): series thickness has been  
277 strongly reduced and some depressions have formed at the surface of the Miocene series. Marine and/or aeolian  
278 continental post-MSC sediments filled many of these depressions (Unit 4, Fig. 13b).

279 We propose that the late Miocene series of NE Ibiza (mainly Units 2 and 3) underwent extensional  
280 deformation. The thick homogeneous limestone bed with herringbones structures (Unit 3-1) is segmented into  
281 large scale symmetrical rectangular “torn boudins” (classification from Goscombe et al., 2004). The axes of the  
282 scar folds (Hobbs et al., 1976) formed in the inter-boudin (Fig. 15) indicate the extensional directions (Fig. 3). In  
283 analogue experiments of multidirectional extension (Victor and Moretti, 2006; Zulauf et al., 2011), the fractures  
284 formed between the blocks exhibit a large range of orientations but formation of true “chocolate tablet” patterns  
285 (Wegman, 1932; Ramsay, 1967) is not possible during one single phase of pure flattening strain (Zulauf et al.,  
286 2011) but requires two successive boudinage events (Goscombe et al., 2004). Unfortunately, there is a small  
287 number of boudins to be observed and few data are available. Nevertheless, the importance of both the 65°N and  
288 140°N directions of folds (Fig. 3) suggests that extension occurred along roughly two perpendicular directions,  
289 forming badly organized “chocolate tablet” patterns. Large translational slides that have been described by  
290 Huntoon (1982), Huvenne et al. (2002) and Callot et al. (2008) could provide an illustration of large scale sliding  
291 and boudins (Fig. 15) that developed through the inhomogeneous deformation of the different layers during  
292 extension (Wegmann, 1932).

293 We propose a ductile stretching conceptual model of the Miocene series of north Ibiza (Fig. 16). The low  
294 tilting of the series to the east allows an exposition of the base of the boudins to the west, while the upper part is  
295 exposed to the east. Three types of boudinage deformation are observed, depending on their location within the  
296 sliding system:

297 1) On the western side of the studied area, rectangular boudins have been formed in Unit 3 (about 20 m-  
298 thick grainstone bed). During sliding and stretching, the mid part of the boudins sags gently to form the large  
299 syncline observed in Figure 6, while soft clay-rich Unit 2 escapes on the sides of the boudins where domes of

300 Unit 2 are observed. Some recrystallized limestones have been observed locally in Cala Fondo (Fig. 8) while in  
301 Punta Galera there are indications of soft-sediment deformation (Fig. 7).

302 2) Around the Moscarter lighthouse (Fig. 5), the grainstone layers of Unit 3 have been segmented to form  
303 large boudins while the laminated layers have bent to form ductile folds (Figs. 9 to 12). Diagenetic  
304 recrystallization of limestones occurred and large calcite veins formed between the bent layers. To the top, over  
305 the synclines that developed between the boudins, depressions have been formed and progressively filled with  
306 paleosols onlapping the laminated layers. In some places, subaerial erosion is observed. At the final stage, Unit 4  
307 sediments have completed the infill of most of these depressions.

308 3) On the eastern side of the studied area, both Unit 2 and Unit 3 are stretched (Punta des Gat area, Figs. 3  
309 and 13). The grainstone beds with herringbones of Unit 3 have formed irregular pinch and swell structures.  
310 Limestone recrystallization is observed everywhere at the base of stretched Unit 3 and calcite veins are  
311 numerous between the layers. It corresponds therefore to the most stretched domain of the studied area.

312

#### 313 4. 2. Rheological control of soft layers over collapse structures

314 Rectangular boudins are recognized to have form under brittle conditions (Wegmann, 1932; Strömgård,  
315 1973); a strong layer being segmented in relation to the ductile adjacent rocks while an increase of ductility leads  
316 to the formation of fishmouth boudins. In the Miocene series of NE Ibiza, the grainstones beds with  
317 herringbones have low fracture density and form large boudins (about 30 m thick and ca. 100 m wide). This  
318 dependence of fracture density on the textural parameter in carbonate rocks has been established by Ladeira and  
319 Price (1981), Hanks et al. (1997), Eyssautier-Chuine et al. (2002), Lezin et al. (2009), showing that the fracture  
320 density decreases when the grain content increases. The length to thickness ratio of the boudins (L/W according  
321 to Goscombe et al., 2004 classification) ranges from 2 to 4 in most natural examples. This closely fits the  
322 observations made in the Portinatx area but boudins appear to be larger in the WSW-ENE direction (L/W is up to  
323 10) than in the NW-SE direction (Fig. 17).

324 Above and below the segmented competent bed, the incompetent layers deform under nearly ductile  
325 conditions with a low viscosity. Due to friction between the ductile layers and the boudins, most of the  
326 deformation is supposed to occur in between the boudins (Ramberg, 1955) and around the corners of the  
327 rectangular boudins where stress concentration is expected (Strömgård, 1973; Lloyd and Ferguson, 1981). Field  
328 observations of the NE Ibiza Miocene series agree with this theory. The high value of the L/W ratio is an  
329 indication of a high viscosity contrast between the boudins and the surrounding matrix because, in sedimentary



330 rocks, fracture spacing increases when the thickness of the incompetent layers increases or when its viscosity  
331 decreases (Ladeira and Price, 1981; Ji et al., 1998; Bai and Pollard, 2000). Deformation concentration in the  
332 block corners is observed in Figures 9 and 11. Calcite wedges indicating an important fluid circulation often  
333 mark it. Both a ductile layer and a high pore fluid pressure are the required conditions for sliding on a basal  
334 surface (Locat and Lee, 2002; Bryn et al., 2005), here the clay-rich distal part of Unit 2.

335

## 336 **5. Discussion**

### 337 5. 1. Large-scale boudinage versus active deformation

338 The studied deformation in the late Miocene series of northeastern Ibiza is fundamentally after the main  
339 compressional event of the Betic orogenesis (Fontboté et al., 1990; Gelabert et al., 1992; Sàbat et al., 1997). Unit  
340 1 of Punta Xarraca area has been affected by soft-sediment injections probably triggered by earthquakes at the  
341 time of deposition or a short while later when Unit 2 deposition seems to be associated with the formation of  
342 large normal faults, probably during the extensional Serravalian-Tortonian phase. However, recent active  
343 deformation is recorded offshore in the surrounding areas and the Alicante shelf, and can be related to the  
344 general inversion of the Mediterranean domain (Gargani et al., 2010; Maillard and Mauffret, 2013; Gorini et al.,  
345 2015).

346 In some places, the occurrence of crossed fold directions could suggest the formation of superimposed  
347 folding with type 1 interference patterns, according to Ramsay's classification (1967). Nevertheless, both the  
348 occurrence of small-scale stretching deformation structures everywhere in the series and the formation of soft  
349 folds in the lower and upper layers of Unit 3, while the mid-layers remain fragmented but not folded, are  
350 indications of large scale ductile boudinage instead of classic superimposed folding.

351 Raft tectonic due to radial extension has been proposed from field analysis and from analogue  
352 experiments to accommodate large scale extension (Cobbold and Szatmari, 1991; Mauduit et al., 1997; Gaullier  
353 and Vendeville, 2005) but most of the models propose the formation of extensional tilted blocks over a mobile  
354 salt horizon. Such models cannot be exactly used to describe the structures formed in NE Ibiza where there is no  
355 evidence of tilted blocks neither any indication of a salt interval in the Miocene series. At the scale of the Ibiza  
356 outcrops, the geometry of the boudins is controlled by extensional fractures whereas shear fractures are formed  
357 during growth of raft systems on the extensional domain of a passive margin (Rouby et al., 2002).

358

### 359 5. 2. Indications of viscosity ratio

360 Torn boudins are formed by extension in materials with contrasted viscosities (Ramberg, 1955, Price  
361 and Cosgrove, 1990); the most rectangular boudins are observed in materials with the higher viscosity contrast  
362 (Lloyd and Ferguson, 1981). The transition from a rectangular boudin to a barrel-shape boudin is then an  
363 indication of the reduction of the strain rate and of the progression to a more ductile deformation (Goscombe et  
364 al., 2004). Over the west Moscarter syncline (Fig. 11), low strain rate conditions are indicated by the progressive  
365 deposition of marls (paleosol layers) during the deformation process, which could explain the softening of some  
366 boudin corners observed from various localities (Fig. 9). In the same way, the secondary folds, such as observed  
367 at the lower part of the east side of the Figure 11c, have been recognized by Lloyd and Ferguson (1981) as an  
368 early stage in the formation of barrel-shaped boudins.

369 The aspect ratio of the boudins (Fig. 17) is expressed by the L/W parameter (Fig. 15; Goscombe et al.,  
370 2004). As far as we can see, the L/W ratio seems to range from 2 to 4 in the NW-SE direction. The high value of  
371 the L/W ratio and the ductile deformation in the boudin corners could lead to opposite interpretations of the  
372 viscosity contrast but most examples from literature refer to metamorphic conditions while the Messinian series  
373 of NE Ibiza may have been deformed as synsedimentary structures (Simó Marfa, 1982). Ogata et al. (2014;  
374 2016) have pointed out the interest of the analysis of soft-sediment deformation in comparison with  
375 metamorphic micro-tectonics (Passchier and Trouw, 1996). In the same way, the calcareous cross-laminated  
376 grainstones deform as elongated boudins bounded by the ductile laminated layers. The high L/W ratio is an  
377 indication that the boudins are separated by a few vertical fractures, as usually observed in grainstones.  
378 Nevertheless, the ductile behavior of boudin corners is an indication that the viscosity ratio between the boudins  
379 and the folded laminated layers was probably quite lower than what may be expected from the boudins shape.  
380 When only meter scale boudinage is recorded in the literature (Goscombe et al., 2004), the development of such  
381 large-scale boudinage ( $L > 100\text{m}$ ) might be a specificity of soft-sediment deformation in carbonate sedimentary  
382 rocks.

383

### 384 5. 3. Soft-sediment deformation helped by fluid circulations

385 Such large-scale sedimentary boudinage requires stretching and sliding of the sedimentary layers.  
386 Huvenne et al. (2002) described some large translational slides and their occurrence appears to be favored by  
387 elevated pore-pressure. The high level of limestone recrystallization and the numerous occurrences of calcite  
388 veins in Moscarter area are indications of high fluid pressure that can reduce the frictional resistance to sliding  
389 (Locat and Lee, 2002; Sultan et al., 2004; Alves, 2015) and weaken layers within stratified sequences (Masson et

390 al., 2006). Under these conditions, sliding can develop on slopes as low as 1° (Haflidason et al., 2004) and  
391 among the causal factors is the sea level drawdown (Hilbrecht, 1989; Spence and Tucker, 1997; Dreyer et al.,  
392 1999; Sultan et al., 2004; Callot et al., 2009). The early lithification of the carbonate layers compared to the soft-  
393 deformation of the detrital layers (Callot et al., 2008) can lead to stress reduction on the interface. All that  
394 produces an increase in boudins width (Marques et al., 2012) and easy sliding of Unit 3 carbonate layers on Unit  
395 2 clastic layers.

396 Both the deposition of paleosols during the bending of Unit 3 and the formation of calcite wedges on  
397 the fold limbs are indications of a long time process during large fluid circulation. The high values of the fluid  
398 temperatures, up to 250°C, require a deep crustal origin of the fluids that can be explained by the immediate  
399 vicinity of the compressional and extensional structures of Ibiza Island (Etheve et al., 2016) with regard to the  
400 destabilization of the messinian carbonate platform.

401

#### 402 5. 4. Submarine or subaerial sliding conditions and triggering conditions

403 On a Mediterranean scale, the MSC sea-level drawdown is responsible for a large transfer of sediment  
404 from shelves to basin where large clastic layers are observed (Bache et al., 2009; Madof et al., 2019). Comeselle  
405 and Urgeles (2017) have proposed that a large margin collapse in the Valencia basin was triggered by the sea-  
406 level drawdown during the MSC. Offshore, between Ibiza and Mallorca islands, the MSC-related unit displays  
407 200 to 400m-long rafts cut by low-angle normal faults interpreted as thin-skinned deformation due to gliding on  
408 a detachment surface (Maillard et al., 2014). Because the seismic profiles do not reach the upper slope, it is  
409 difficult to correlate this margin collapse with our observations of large-scale boudinage on the upslope of Ibiza  
410 Island. Nevertheless, both appear to have been caused by the same major event of sea-level drawdown, which  
411 can be considered as long-term triggering (Sultan et al., 2004). It produced the translational sliding of soft  
412 carbonate layers of the late Miocene on a gentle slope; the layers already cemented by early diagenetic processes  
413 can break and slide while the softer layers are bent around the blocks.

414 The stretching deformation appears to increase to the NE of Portinatx but boudins are shorter in the  
415 NW-SE direction; all indicates that the sliding has developed mostly northward, towards the Valencia Basin. The  
416 sliding has occurred during the emersion of the MSC, as recorded by the onlap of paleosols over some Unit 3  
417 folds. Finally, MSC erosion took place in the channels created by the synclines depressions and lowered areas  
418 (Figs. 13 and 16), as shown by their progressive filling by Unit 4 shallow marine deposits during the Zanclean  
419 reflooding (Gauillier et al., 2014; Giresse et al., 2015; Maillard et al., 2016).

420 It is difficult to infer the exact chronology between the MES, the gravity flows, and paleosols  
421 expansion, but most are evidence of subaerial events produced after bending of Unit 3. Continental deposits like  
422 fluvial systems were also evidenced onland in Spain, Morocco and Eastern Mediterranean regions, developing  
423 during the peak of the MSC (Do Couto et al., 2014; Cornée et al., 2016, Madof et al., 2019). Bending of Unit 3  
424 under soft conditions, could have begun while still underwater during the beginning of the drawdown, but more  
425 probably in subaerial conditions. If such subaerial destabilization has been evidenced offshore, mostly on the  
426 slopes during the MSC sea-level fall, examples on the shelf remain poorly documented, except where evaporites  
427 are present in the pre-MSB peak succession (Sorbas: Clauzon et al. 2015; Bourrillot et al., 2010; Cyprus:  
428 Orszag-Sperber et al., 2009). Mass flows are documented within fluvio-lacustrine Lago-Mare deposits in Crete,  
429 interpreted as triggered by an early Pliocene tectonic event, though the slope failure is the MES (Zachariasse et  
430 al., 2008; Alves and Lourenço, 2010). Note that in the eastern Mediterranean area, a change of tectonic regime  
431 occurred just at the end of Miocene (Hall et al., 2005) and that thick-skin tectonics could be overprinted by thin-  
432 skin salt tectonics, meaning that MSC (with sea-level drop and collapses) could be responsible for complex  
433 deformation occurring precisely at the Messinian / Pliocene boundary (Maillard et al., 2010).

434

## 435 **7. Conclusions**

436 Most of the deformation of the late Miocene series of north-western Ibiza Island appears as  
437 synsedimentary large scale boudinage. The calcareous herringbone layers of Unit 3 (U3-1) form large  
438 rectangular to barrel-shaped boudins. The overall geometry of these large blocks, reconstructed from mapping  
439 and structural data, corresponds to a badly organized “chocolate tablet” model that likely reveals a  
440 multidirectional extension. The increasing block deformation to the north suggests an increase of stretching, or  
441 translational sliding, in that direction. It could have been favored by the north-eastward rapid increase in clay  
442 content of Unit 2 at the base of boudins and by the large circulation of relatively hot fluids, up to 260°C.

443 Along and between Unit 3 boudins, the laminated layers (U3-2) form ductile scar folds. At the corners  
444 and between the boudins, the progressive bending of these laminated layers of the upper part of Unit 3 manages  
445 spaces filled with calcite crystal wedges that provide an indication for important fluid circulation during  
446 deformation. Unit 3 recrystallization above, below and between the boudins strengthens the indication of large  
447 fluid circulation supporting the conditions of ductile deformation.

448 The topographic depressions formed by the boudinage of the late Miocene series have controlled the  
449 drainage patterns and the coeval development of erosive channels. In Ibiza Island, the occurrence of continental

450 debris flows within these channels testifies that deformation of soft sediments occurred in subaerial conditions,  
451 shortly after their deposition. The progressive infill of paleosols over some of the synsedimentary synclines  
452 confirms that the deformation developed relatively slowly but resulted in an important deformation during the  
453 emersion period. The process of margin collapse during Messinian sea-level drawdown, proposed elsewhere in  
454 the Mediterranean Basin, appears to be responsible for the translational sliding, boudinage and large-scale  
455 collapse of the late Miocene series on the northeastern Ibiza Margin.

456

457

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## Figure captions

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Figure 1. Location map of Ibiza and Balearic islands, south-east of Iberia. The Balearic Promontory is located between extensional basins that opened during early Miocene: the Valencia continental Basin to the north and the Algerian and Ligurian oceanic Basins to the south and East. Contractual deformation of the Betic orogen occurring during mid-Miocene times affected the Promontory at that time. The Messinian Salinity Crisis resulted in an erosion of the margins and deposition of evaporites in the basins that are drawn on the map. The red rectangle shows the location of the map figure 2.

Figure 2. Regional geological sketch of Ibiza Island and surrounding offshore. A) Geological map of Ibiza, with zoom in on the NE part of the Island showing the location of the upper Miocene outcrops around Portinatx that we studied. The offshore area displays the Present-day depth of the base of the Pliocene unit in meters. It corresponds to the MES on the shelf. MSC related deposits (evaporites) are distinguished. B) Schematic cross-section (from seismic data and bibliography) showing the relationship between the onshore study area and the offshore known MSC markers in their geological context.

Figure 3. Map of the fold axes orientations in Unit 3, east of Portinatx; anticlinal and synclinal fold axes are labeled. The fold axes orientations are grouped around 65°N and 140°N directions. Locations of the different figures are indicated. Small stereographic projection of fold axes has been drawn with the *Stereonet* software (Cardozo and Allmendiger, 2013).

Figure 4. Late Miocene series of the Cala de Serra, NE Ibiza. Unit 1 is a carbonate-rich unit assigned to Tortonian - lower Messinian in age. This unit that overlaps mesozoic recrystallized limestones is underlined by eight meters of reworked corals (*Porites* rudstone) with benthic foraminifera and gastropods (section A). Southward, near the main faults, these deposits change into detrital sedimentation composed of Mesozoic blocks. Unit 2 is a detrital unit, with significant thickness decrease to the North. The size of the elements also decreases progressively from very coarse elements (conglomerate) to the South (section A) to marls and clay-rich sediments to the North (section C). This conglomerate-clay unit was set up by gravitational and fluvial processes in the nearshore environment. Unit 2 is overlapped by oolite, microbialite, gastropod-rich limestones of Unit 3. The lower part of Unit 3 (U3-1) is mainly composed of oolitic rich limestones indicating open shallow marine

754 environment dominated by tide dynamics. In the upper part of Unit 3 (U3-2), microbialites (thrombolite,  
755 stromatolite) become predominant and characterize the nearshore environment. The three stratigraphic logs show  
756 the faciological variation in space and time of the tortono-messinian series along a SW-NE transect North of Cala  
757 de Serra. From section A to section C, the top of Unit 1 is located near the current sea surface. Unit 2 shows  
758 significant thickness reduction (25 m to 0.5 m) and lithological change (conglomerate dominant to marls  
759 dominant). Unit 3 reveals no significant lateral faciological change but the top of this unit is eroded northward. At  
760 the SW all the Miocene series is overlap by the Margin Erosion Surface and overcome by Pliocene-Pleistocene  
761 formation.

762

763

764 Figure 5. a) Picture of Unit 3 (late Miocene series) along Aigua Blanca cliff, east of Portinatx. The lower part of  
765 the cliff (pink color, Unit 3-1) is composed of the herringbone layers (oolitic grainstones) that rest in an almost  
766 horizontal position when the upper part of the cliff (Unit 3-2) exhibits gentle folds; b) Line drawing of the  
767 structures. The outcrop length is 1.2 km when the maximum elevation of the cliff is 30 m. The base of Moscarter  
768 lighthouse appears on the left.

769

770 Figure 6. On Punta Galera Island, Unit 3 is folded in a large syncline trending N 175° (small stereographic  
771 projection) resting on the nearly flat Unit 1. A part of Unit 2 forms a wedge thickening to the west. The cliff is  
772 25 m high. Unit 1 and 3 are made of limestones, while Unit 2 of white clay.

773

774 Figure 7. a and b) On the west side of the syncline, at the base of Unit 3, numerous load structures are observed  
775 sinking into the argillaceous sediments of Unit 2. They are elongated soft-sediment deformation structures and  
776 each long axis is orientated N0°, parallel to the syncline axis; c) Calcite geodes indicate the occurrence of fluid  
777 circulation in the lower part of Unit 3.

778

779 Figure 8. Cala Fondo outcrop. a) It is a large and flat syncline made of brecciated limestones resting on the top  
780 surface of Unit 1 (stereographic projection for fold axis); b) Unit 2 is less than one meter thick but is more  
781 developed on both the western and eastern sides of the syncline. The base of the black recrystallized limestones  
782 is broken by small extension fractures filled with red clays. Hikers for scale.

783

784 Figure 9. a) Picture of large-scale boudins formed in Unit 3, few meters east of the Moscarter lighthouse. The  
785 cliff is 26 m high. The calcareous herringbone layers of Unit 3-1 are segmented in large rectangular boudins.  
786 Around the blocks, deformation is limited to the corner parts where Unit 3-2 laminated upper layers form scar  
787 folds composed of a tight syncline between two open anticlines; b) Line drawing of the picture; triangles indicate  
788 the brecciated part of the layers; fold axis is drawn from the stereographic projection of Cardozo and  
789 Allmendiger (2013); c) Brittle deformation of the laminated layers that have been transformed into recrystallized  
790 limestone during diagenetic process. Small mode I fractures have been formed perpendicular to the layers and  
791 they are filled with red argillaceous sediments. Fracture orientations range from N 125° to N 145°, parallel to the  
792 fold axis; pen for scale.

793

794 Figure 10. Facies variations and diagenetic transformations in the upper part of Unit 3 below Moscarter  
795 lighthouse. a-b) Unit 3-1 (lower part) is composed of oolitic grainstones with herringbone structures and trough  
796 cross-stratification, while microbialites dominate limestone composition in the upper part (Unit 3-2).  
797 Unit 3-2 shows the boundary between well preserved and diagenetically transformed (darkened part of figure b)  
798 microbialitic facies. Laminations are visible (white lines). c,d,e) Neomorphic transformations of microbialitic  
799 facies (Unit 3-2, darkened part picture b). c, d) Plain-polarized light (PPL) photomicrograph showing the original  
800 texture and grain composition: microbial mat with abundant peloids. e) Cross-polarized light (CPL)  
801 photomicrographs showing large-scale carbonate cement, a product of neomorphic changes.

802

803 Figure 11. a) West of Moscarter lighthouse, the herringbone layers (Unit 3-1) have been segmented in large  
804 blocks while the microbialitic layers (Unit 3-2) have formed two half anticlines orientated N136° and N20°; b)  
805 Line drawing of the picture showing the progressive onlap of paleosols that have filled the depression on top of  
806 Unit 3; c) The small scale folds observed at the lower part of the eastern side have been estimated to have N105°  
807 orientation; d) Extensive brittle deformation in the tilted layers of a scar fold.

808

809 Figure 12. East of Portinatx (location on Fig. 3 and Fig. 5), the onset of bedding tilt on the border of a large  
810 boudin shows a large calcite wedge. The progressive increase of the layer tilting downward has opened the space  
811 filled with vertical calcite crystals. Small stereographic projection indicates the construction of the fold axis.

812

813 Figure 13. Punta des Gat outcrop, a) Sections are composed of the upper part of Unit 1 and some strongly  
814 stretched layers of Unit 3; b) Depressions have formed at the surface of the late Miocene series (stretched layers  
815 of Unit 3 are in blue). Continental post Messinian sediments fill many of these depressions (Unit 4). The  
816 maximum elevation of the outcrop is 13 m.

817

818 Figure 14. Detail of the lower part of Unit 2 at Punta des Gat outcrop. This deformed formation is composed of  
819 mixed white and green clays, black recrystallized limestones, red blocks, and white clays, fragmented limestones  
820 and injected red clays that may come from the decalcification of limestone layers. The clays are a deformed part  
821 of Unit 2 and the limestones are the deformed and highly stretched part of the base of Unit 3. Notebook for scale.

822

823 Figure 15. From rectangular boudins to fishmouth boudins with an increasing ductility of the layers (a to c),  
824 redrawn from Wegmann (1932) and using parameters of Goscombe et al., (2004): L is long axis of boudin, W is  
825 thickness of boudinaged layer and N is the amount of dilatation.

826

827 Figure 16. Conceptual model of the large-scale boudinage of Unit 3 to form “chocolate tablets” in the late  
828 Miocene series of NE Ibiza. The herringbone layers of Unit 3-1 appear to be more competent and they form  
829 large scale rectangular boudins. On top of the boudins, the laminated layers of Unit 3-2 are bent to form passive  
830 folds. Sliding over the ductile clayey layers of Unit 2 has shifted the boudins. Rectangular boudins have been  
831 formed to the west and progressively a more diffuse stretching of the layers is observed to the east. a) Base of the  
832 boudins as they can be seen in Punta Galera outcrop (Fig. 6); b) Large blocks separated by a tight syncline and  
833 the progressive onlap of paleosols filling the depression (Fig. 11); c) Extensional deformation and progressive  
834 tilting of the laminated layers forming calcite crystal wedges (Fig. 9); d) The strongly stretched part of Unit 3  
835 resting on Unit 1 flat top; there, the two units are separated by clays that may be considered as coming from  
836 decalcified limestones (Fig. 13). The continental post-Messinian sediments (Unit 4) have filled many of the  
837 depressions formed by the stretching of Unit 3. U1, U2, U3, and U4 refer to the different sedimentary units.

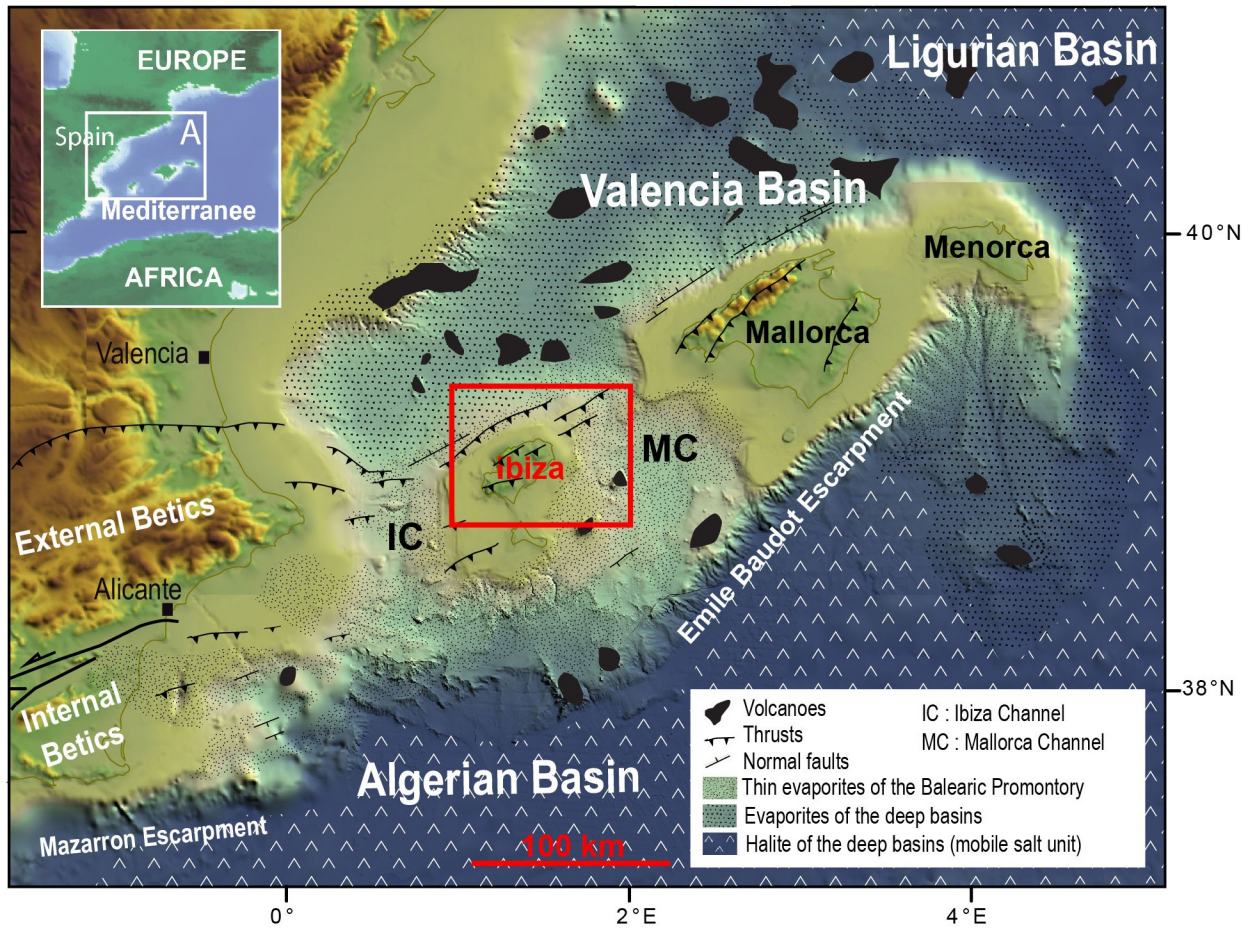
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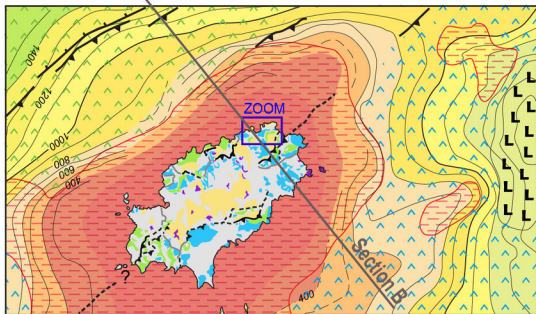
839 Figure 17. Schematic map of Portinatx area. Blocks edges are figured by black segments. The length of the  
840 boudins (L) is drawn in red when measured in the WSW-ENE direction (red circles) and green in the NW-SE  
841 direction (green stars). The boudins are longer in the WSW-ENE direction.

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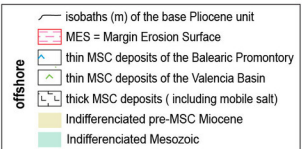
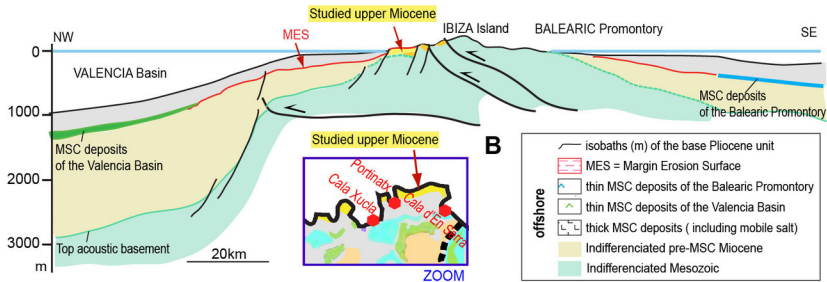
843 Table 1. Microthermometric data of fluid inclusions in calcite crystals. The trapping conditions are reconstructed  
844 from the homogenization temperatures. Samples coming from the Cala de Ponent indicate calcite crystals  
845 formation between 190 and 260°C. Temperatures are given with an accuracy of 1.0°C.  
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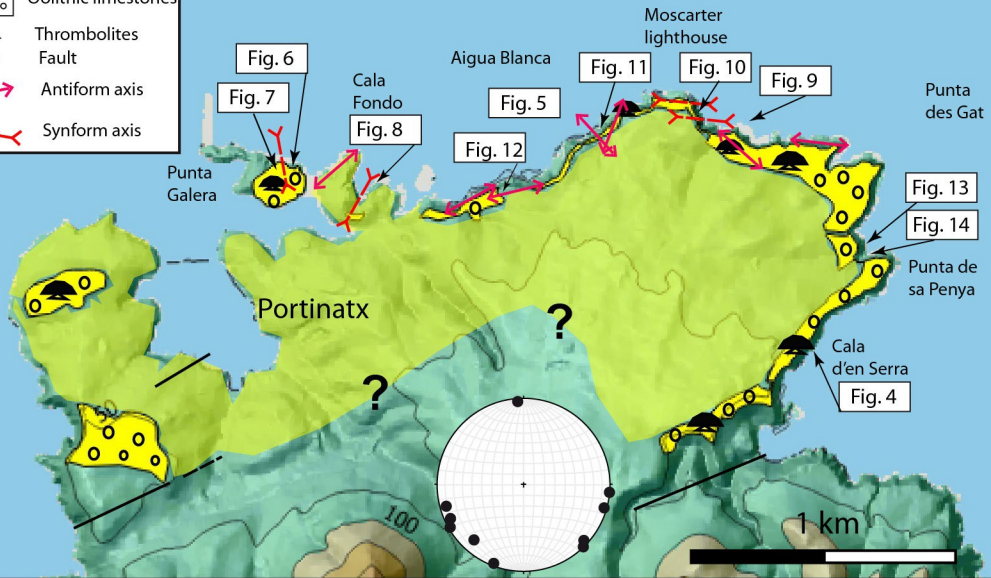
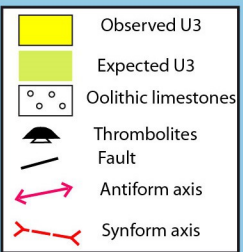


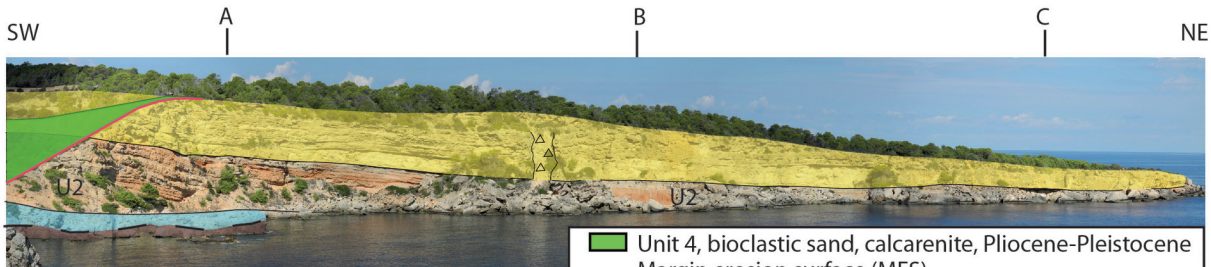


**A**

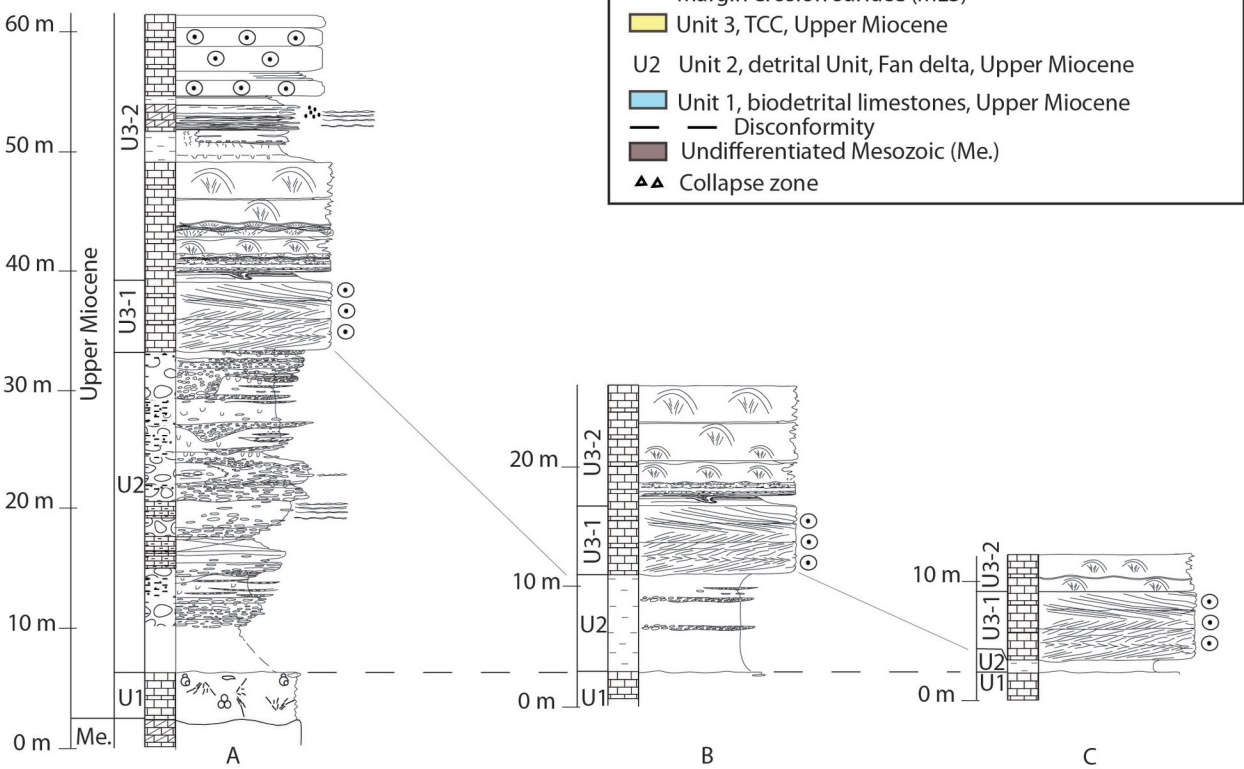


**B**



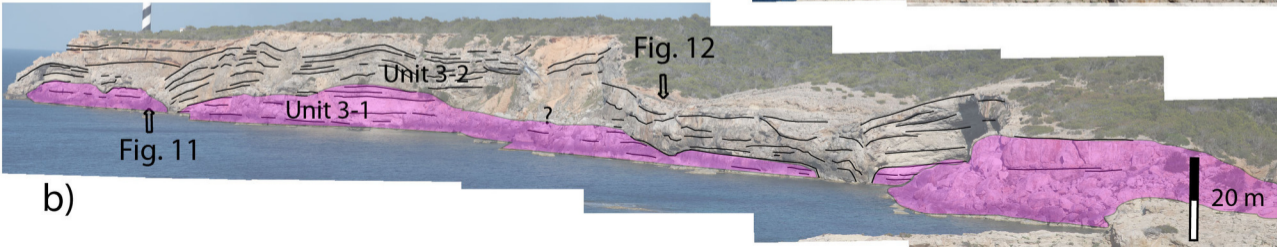


- Unit 4, bioclastic sand, calcarenite, Pliocene-Pleistocene
- Margin erosion surface (MES)
- Unit 3, TCC, Upper Miocene
- U2 Unit 2, detrital Unit, Fan delta, Upper Miocene
- Unit 1, biodetrital limestones, Upper Miocene
- Discontinuity
- Undifferentiated Mesozoic (Me.)
- Collapse zone



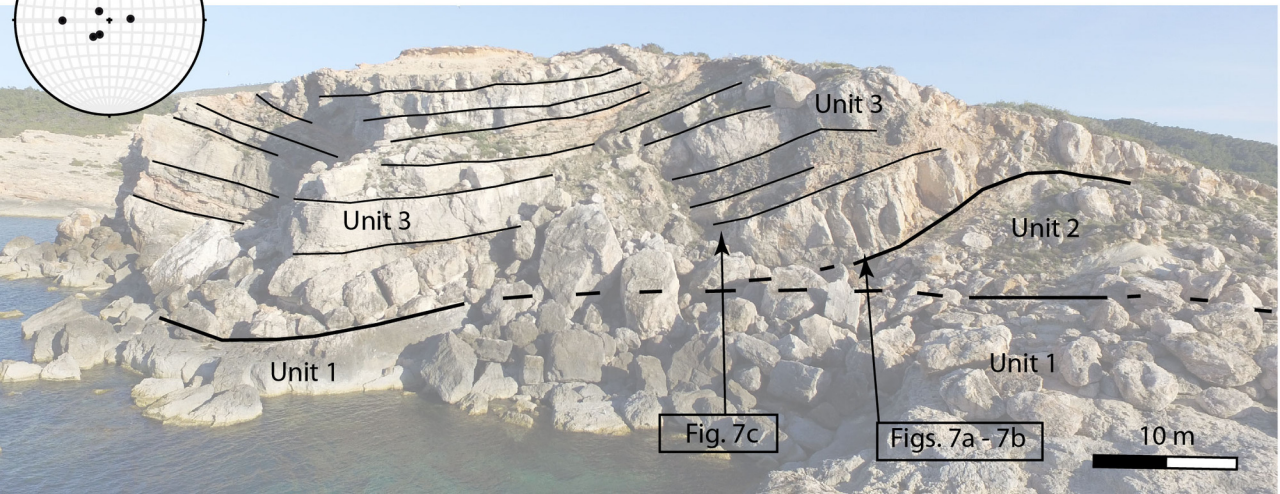
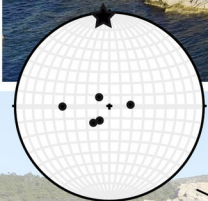
- |  |   |   |
|--|---|---|
| <span style="display: inline-block; width: 15px; height: 10px; border: 1px solid black; background-color: white; margin-right: 5px;"></span> Limestones                | <span style="display: inline-block; width: 15px; height: 10px; border: 1px solid black; background-color: white; margin-right: 5px;"></span> Reworked corals      | <span style="display: inline-block; width: 15px; height: 10px; border: 1px solid black; background-color: white; margin-right: 5px;"></span> Peloids                      |
| <span style="display: inline-block; width: 15px; height: 10px; border: 1px solid black; background-color: white; margin-right: 5px;"></span> Recrystallized Limestones | <span style="display: inline-block; width: 15px; height: 10px; border: 1px solid black; background-color: white; margin-right: 5px;"></span> Thrombolite          | <span style="display: inline-block; width: 15px; height: 10px; border: 1px solid black; background-color: white; margin-right: 5px;"></span> Stromatolites/Algal Mats     |
| <span style="display: inline-block; width: 15px; height: 10px; border: 1px solid black; background-color: white; margin-right: 5px;"></span> Conglomerates             | <span style="display: inline-block; width: 15px; height: 10px; border: 1px solid black; background-color: white; margin-right: 5px;"></span> Gastropods           | <span style="display: inline-block; width: 15px; height: 10px; border: 1px solid black; background-color: white; margin-right: 5px;"></span> Pebbles Gravels              |
| <span style="display: inline-block; width: 15px; height: 10px; border: 1px solid black; background-color: white; margin-right: 5px;"></span> Clay                      | <span style="display: inline-block; width: 15px; height: 10px; border: 1px solid black; background-color: white; margin-right: 5px;"></span> Benthic foraminifera | <span style="display: inline-block; width: 15px; height: 10px; border: 1px solid black; background-color: white; margin-right: 5px;"></span> Roots                        |
| <span style="display: inline-block; width: 15px; height: 10px; border: 1px solid black; background-color: white; margin-right: 5px;"></span> Silty-marls               | <span style="display: inline-block; width: 15px; height: 10px; border: 1px solid black; background-color: white; margin-right: 5px;"></span> Ooids                | <span style="display: inline-block; width: 15px; height: 10px; border: 1px solid black; background-color: white; margin-right: 5px;"></span> Cross bedding stratification |
| <span style="display: inline-block; width: 15px; height: 10px; border: 1px solid black; background-color: white; margin-right: 5px;"></span> Marls                     |   |   |





N-E

S-W



Unit 3

Unit 3

Unit 2

Unit 1

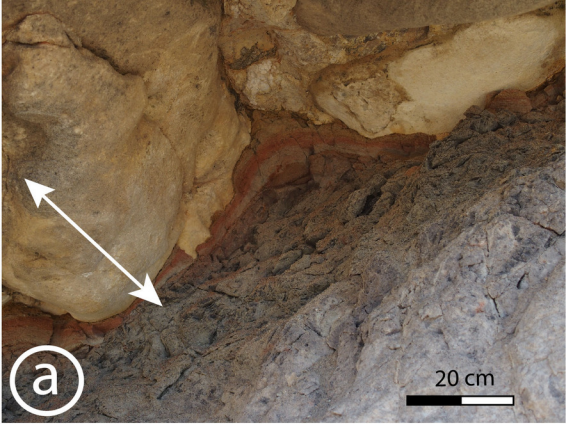
Unit 1

Fig. 7c

Figs. 7a - 7b

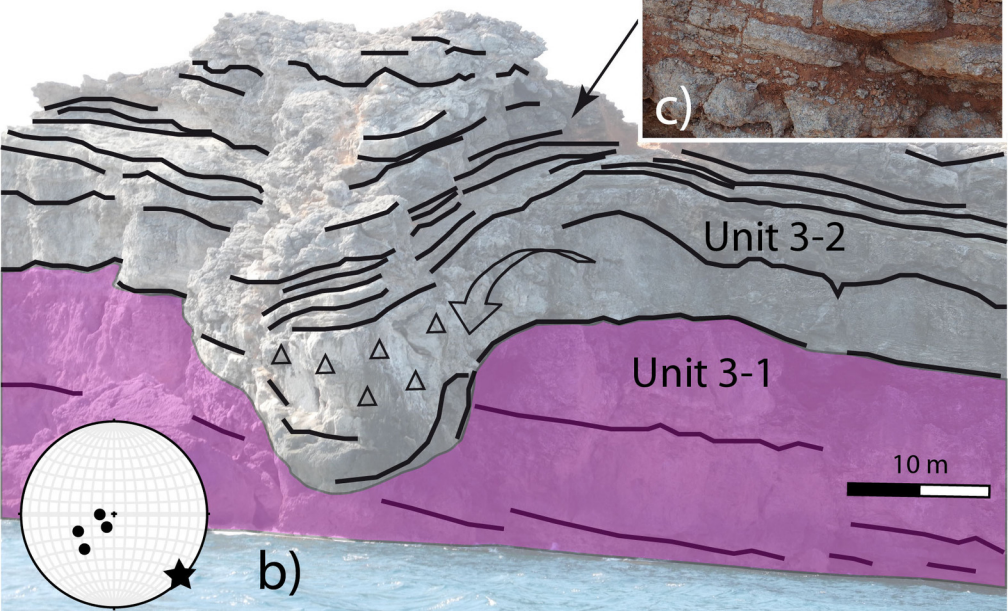
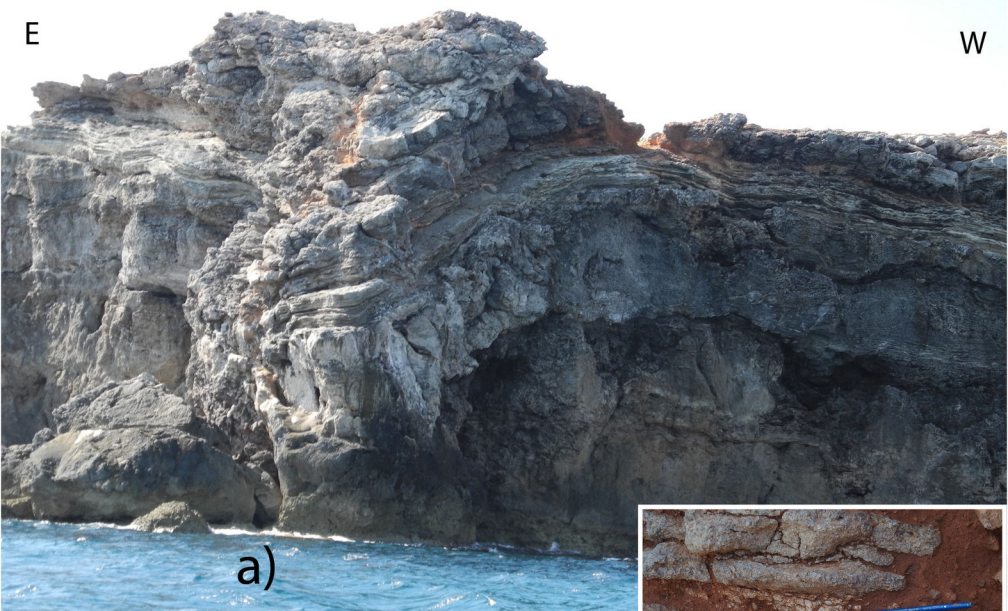
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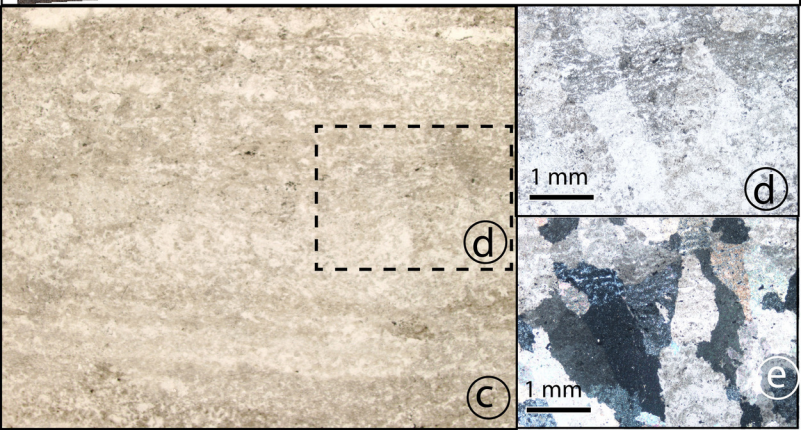
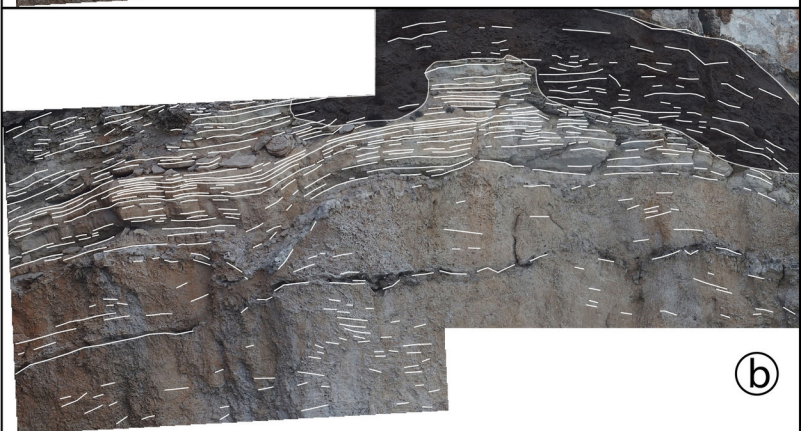




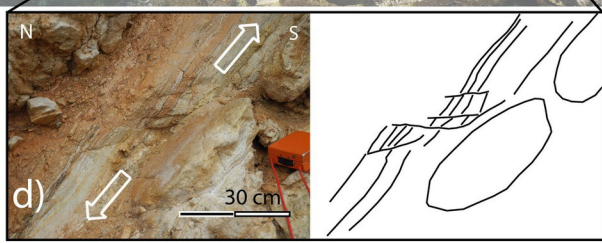
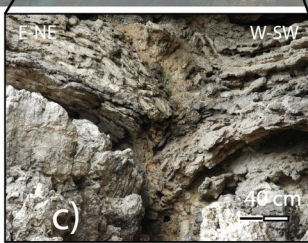
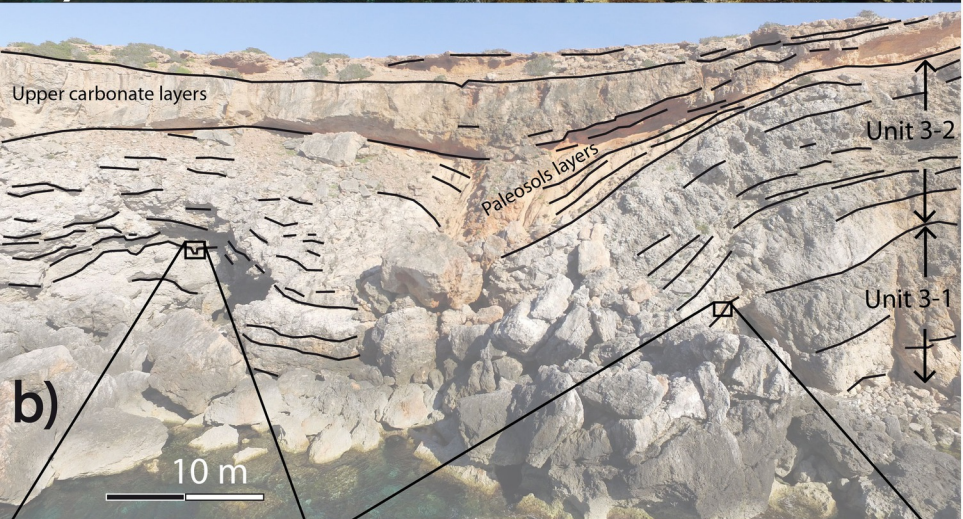




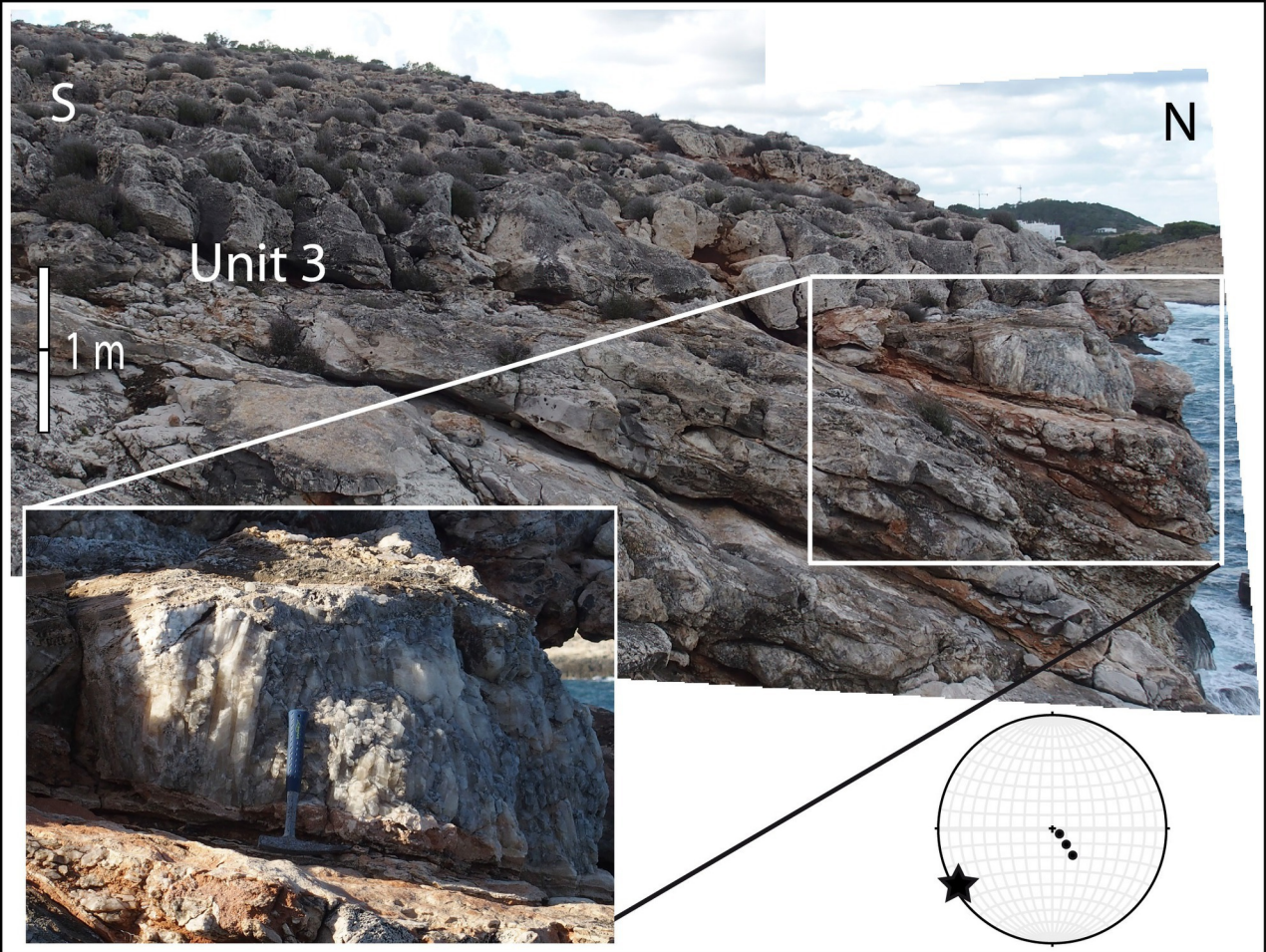


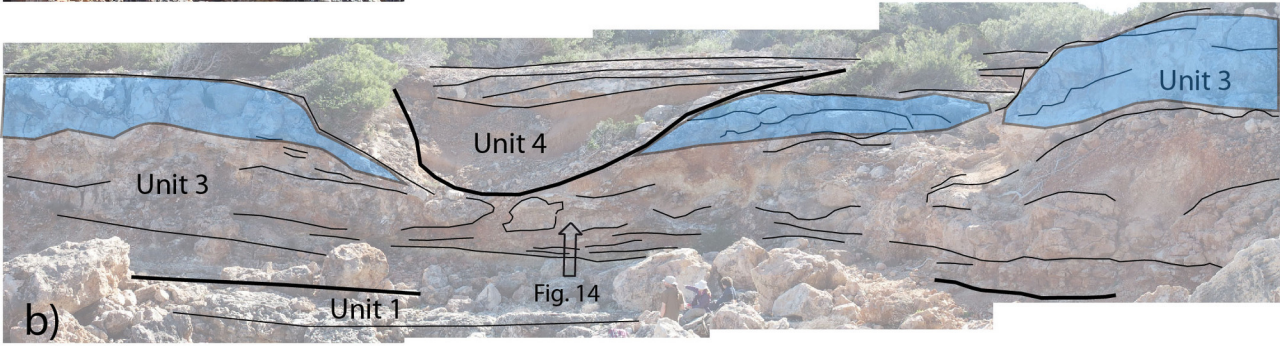














Fragmented limestones  
and injected red clays

SE

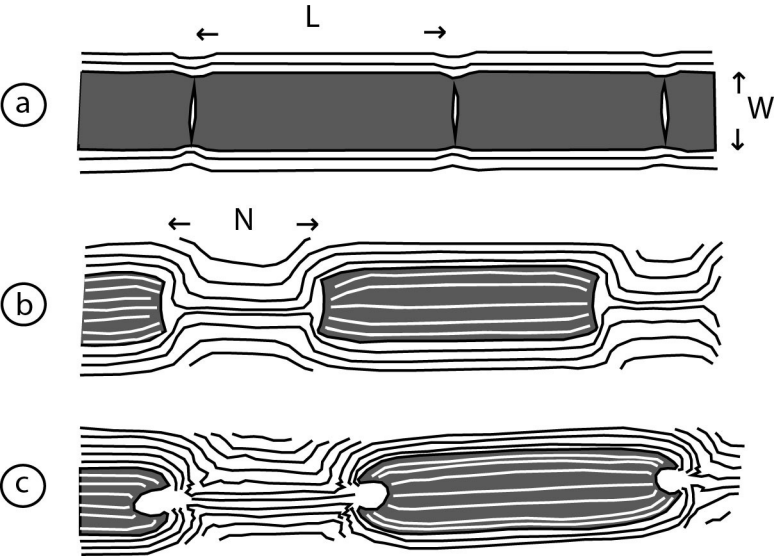
NW

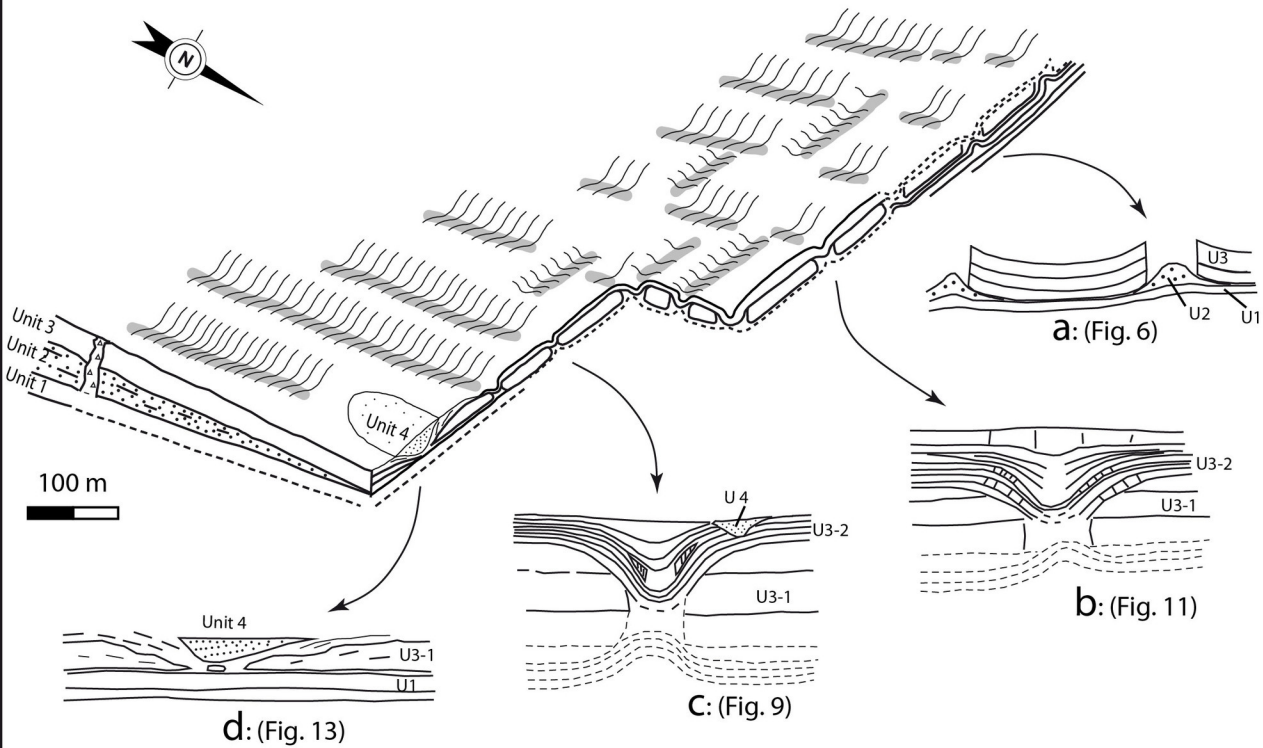
Red blocks and  
white clays

Black recrystallization  
limestones

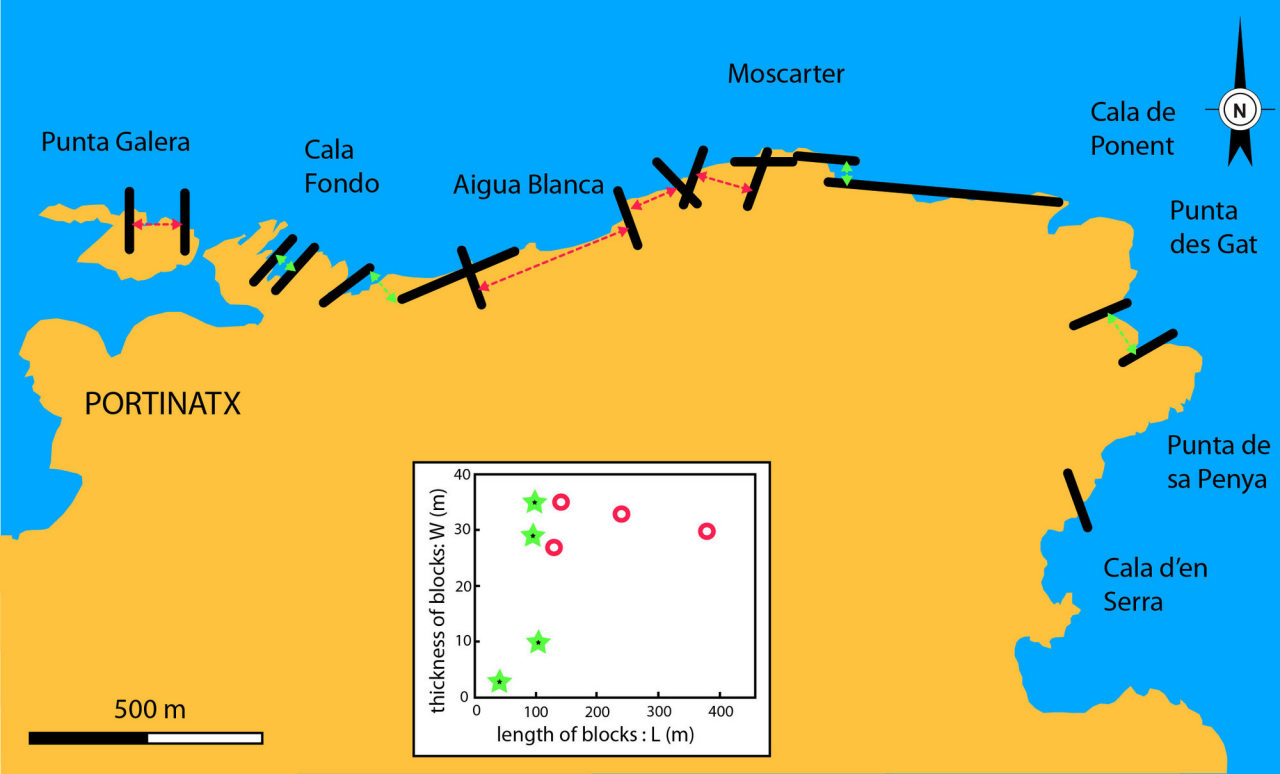
White and  
green clays











samples	<b>PO1</b>	<b>PO2</b>
homogenization temperature	219	188
	242	190
	256	202
	257	205
	260	209
	260	
	260	
	260	