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**New insights on the genesis of the Miocene collapse structures of the Island of Gozo
(Malta, Central Mediterranean Sea)**

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29

30 **Abstract**

31 The large palaeosinkholes located in the NW of Gozo Island (Central Mediterranean Sea, Malta)
32 offer excellent exposures that provide information on the geometry and kinematics of large
33 karst-related collapse structures. Detailed geological analysis of these peculiar palaeosinkholes
34 indicates that deep-seated evaporite dissolution is the most feasible hypothesis to explain their
35 formation, according to the following evidence: (1) several structures have been formed by
36 progressive foundering of cylindrical blocks with limited internal deformation as reveal the
37 synsedimentary subsidence recorded by their Miocene sedimentary fill. This subsidence
38 mechanism is more compatible with interstratal dissolution of evaporites than karstification and
39 cave development in limestone formations. (2) The dimensions and deformation style of the
40 palaeosinkholes are similar to those of other collapse structures related to deep-seated
41 dissolution of salt-bearing evaporites. (3) The arcuate monocline associated with some of these
42 collapse structures is also a characteristic feature of subsidence related to dissolution of
43 evaporites. Notwithstanding, no major evaporite formations have been documented so far in the
44 subsurface of the Malta Platform.

45

46 **Supplementary material:** Detailed descriptions of the collapse structures of the Island of Gozo
47 (Malta, Central Mediterranean Sea) are available at www.geolsoc.org.uk/SUP00000.

48

49 Giant collapse sinkholes are one of the Earth's most dramatic and enigmatic landforms. Well-
50 known examples include the Tiankengs of China and SE Asia, that reach several hundred meters
51 deep and across (Waltham 2006), the Obruks of Central Anatolia (Bayari *et al.*, 2009), the large
52 collapse dolines of the Dinaric karst (e.g. Velika Dolina, Waltham 2006; Crveno Jezero, Garašić
53 2000), and the Zacatón sinkhole in Mexico (Gary & Sharp 2006). Large collapse
54 palaeosinkholes have been preserved in the geological record as gravitational deformation
55 structures (e.g. ring faulting, breccia pipes) and related sedimentary features (e.g. sinkhole fills,
56 local thickness variations, cumulative wedge-outs). These features have been described in the

Colorado Plateau, USA (Weir *et al.* 1961; Cater 1970; Sugiura & Kitcho 1981; Gutiérrez 2004), Southern Saskatchewan, Canada (Christiansen 1971; Gendzwill & Hajnal 1971; Christiansen and Sauer 2001), NE Spain (Gutiérrez 1996, 2014) and Southern Hunan, China (Min *et al.* 1997). Large collapse palaeosinkholes have also been identified through 3D seismic surveys in sedimentary basins of Southern Australia (Brown 1999), Venezuela (Castillo & Mann 2006), USA (McDonnell *et al.* 2007), Mediterranean Sea (Bertoni & Cartwright 2005, Lofi *et al.* 2012) and the South China Sea (Sun *et al.* 2013).

These structures have a multidisciplinary importance because their identification aids for ore and hydrocarbon exploration in sedimentary basins affected by deep-burial dissolution processes. Fossil collapse sinkholes are markers of large scale deep-seated karstification that favoured the migration and/or accumulation of hydrocarbons in the Maracaibo basin, Venezuela (Castillo & Mann 2006), in the Fort Worth basin, Texas, USA (McDonnell *et al.* 2007), and in the South China Sea basin (Sun *et al.* 2013). Palaeosinkhole-hosted uranium deposits of a high economic importance have been exploited in China (Min *et al.* 1997) and Arizona (USA) (Wenrich & Titley 2008). Sun *et al.* (2013) highlights that deep-burial dissolution should be taken into account when assessing reservoir heterogeneity and infrastructure risks in the largest offshore oil fields. This also applies to the evaluation and exploitation of ore deposits associated with palaeosinkholes. However, to our knowledge, studies on palaeosinkholes that may help to characterize deep-seated dissolution environments are very scarce. Detailed analyses on well-exposed large collapse structures, paying attention to their geometry, internal structure, kinematics and origin are needed to better understand similar structures and geomorphic features identified by indirect methods in the subsurface or in the sea floor.

The large palaeocollapse sinkholes of Gozo Island, Malta, central Mediterranean, described by Pedley (1974), are revisited in this paper. They offer excellent exposures that allow obtaining direct information on the internal architecture and kinematics of large karst-related collapse structures developed on the sea floor. In some cases, they display spectacular landforms of high

85 scientific and scenic value, which deserve protection and conservation measures (Coratza *et al.*
86 2012). Previous studies, dating back to the 1970s and 1980s, analyzed these fossil collapses
87 from a sedimentological perspective (e.g. Pedley & Bennett 1985). Their main interest was that
88 these paleosinkholes acted as sediment traps favouring the preferential accumulation of detrital
89 phosphorites. Other aspects, such as their structural style and origin have not been analyzed in
90 depth by previous authors, although Soldati *et al.* (2013) undertook a geomorphological study.
91 Detailed examination and mapping of structural and stratigraphic features documented in this
92 work, contribute to shed some light on their debatable origin. These large collapse features have
93 been previously attributed to the breakdown of a cave system developed in limestone (Pedley
94 1974) or to deep-seated evaporite dissolution (Illies 1980; Pedley *et al.* 2002). The aim of this
95 paper is to discuss these theories and the possible karstification and subsidence mechanisms
96 responsible for their development. We have collected and analysed novel data regarding (1)
97 regional geological and stratigraphic context; (2) stratigraphic and structural relationships
98 observed in detailed geological maps; (3) stratigraphic and sedimentological features of the
99 sediments deposited in the submarine sinkholes; (4) biostratigraphic data inferred from
100 planktonic foraminifera; and (5) bathymetric data and geomorphological evidence of probable
101 residual activity. This information allowed us to infer aspects related to the development of the
102 sinkholes, including the deformation style, the spatial association with other structures of
103 probable gravitational origin, the kinematics of the subsidence and its duration. The discussion
104 on the possible genetic alternatives for the palaeosinkholes takes into account the new
105 knowledge frame and a number of large dissolution-induced collapse structures documented
106 worldwide.

107

108 **Geological setting**

109 The Maltese archipelago is located in the central Mediterranean Sea, around 90 km south of
110 Sicily and 290 km east of Tunisia (Fig. 1A). It consists of three main islands: Malta (246 km²),
111 Gozo (67 km²) and Comino (3.5 km²) (Fig. 1B). From the geotectonic perspective, the
112 archipelago is located in the northern African plate, and more specifically on the Pelagian

Platform, a continental shelf between southern Sicily (Malta-Ragusa platform) and northern Libya (Tripolitanian platform) (Illies 1981) (Fig. 1A). These platforms constitute the foreland of the south-verging Maghrebian Thrust Belt, at the collision zone between the Eurasian and African plates. The Pelagian Platform is cut by the deep NW-SE trending trough of the Pantelleria Rift. The northeastern part of the rift is designated as the Malta Graben (Gardiner *et al.* 1995). The Maltese archipelago is located on the NE shoulder of the Pantelleria Rift, very close to the master fault that controls the graben margin escarpment (Fig. 1A). Around 10 km northwest of the Island of Gozo, there is an ENE-WSW-oriented graben, called the North Gozo Graben. This structure is controlled by oblique dextral-normal faults that connect with the normal faults of the Pantelleria Rift (tear faults). The kinematic evolution of these structures has been inferred by Gardiner *et al.* (1995) on the basis of stratigraphic and structural relationships observed in seismic profiles (e.g., thickness variations and offset sedimentary units). The Pantelleria Rift began to open in late Miocene to mid-Pliocene time under an extensional regime. The formation of the North Gozo Graben started in the late Pliocene, in relation with a change in the tectonic regime, from simple extension to right-lateral transtension. The oblique dextral faults with vertical components that control the North Gozo Graben offset in Pleistocene sediments and the sea floor, strongly suggesting that they correspond to an active fault system (Gardiner *et al.* 1995).

131

The rocks exposed in the islands comprise a Late Oligocene (Chattian) to Late Miocene (Messinian) marine sedimentary succession mostly composed of limestones and marls (Fig. 1B). Five lithostratigraphic units form the outcropping sequence (Oil Exploration Directorate 1993; Pedley *et al.* 2002). The lowermost exposed unit is the *Lower Coralline Limestone Formation* (Chattian) of pale-grey, hard, shallow marine biomicrites and biosparites around 140 m thick. The following units, in ascending order, are: (1) The *Globigerina Limestone Formation* (Aquitania-Langhian), a yellowish, fine-grained, planktonic foraminiferal limestone, 20 m to over 200 m in thickness. This formation has been subdivided into three members by the occurrence of laterally extensive phosphorite conglomerate beds; i.e. *Lower and Upper Main*

141 *Phosphorite Conglomerate Beds* (Pedley & Bennet 1985). (2) The *Blue Clay Formation*
142 (Serravallian-Tortonian) consists of grey, soft marls, clays and silty sands 20 to 70 m thick. (3)
143 The *Greensand Formation* (Tortonian) is brown and greenish glauconite-rich generally less
144 than 1 m thick, but reaching 11 m thick. (4) The *Upper Coralline Limestone Formation*
145 (Tortonian-Messinian) is broadly similar to the Lower Coralline Limestone Formation and up
146 to 160 m thick.

147

148 The unexposed stratigraphic units beneath the Late Oligocene Lower Coralline Limestone
149 Formation, were shown by the BP oil exploration borehole (Naxxar 2) on the Island of Malta
150 (Fig. 1B). It revealed a 3000 m thick sequence of platform carbonates, mainly dolomites, dating
151 back to the Lower Cretaceous in its lowermost part (Pedley 1990). Extensive dissolution-
152 related cavernisation was found within the Tertiary and Mesozoic carbonate strata in this
153 borehole (Pedley 1974). Another borehole drilled offshore by the Shell company (MS-A1
154 borehole; Fig. 1A) 40 km northeast of the Island of Malta shows a similar dolomite succession
155 5000 m thick down to the Lower Jurassic (Jongsma *et al.* 1985). The Madonna Taz-Zejt 1ST1
156 borehole, drilled on land in western Gozo (Fig. 1), proved a 3200 m thick Late Triassic
157 succession from a depth of 4.5 km, consisting of dolomites, shales and meters-thick anhydrite
158 beds, the latter with an aggregate thickness of ca. 230 m (Debono *et al.* 2000). In Zabbar and
159 Aqualta boreholes, drilled in the Malta Platform (Fig. 1A), a 1-m-thick gypsum bed has been
160 recorded at the Eocene-Oligocene boundary, 100-150 m below the base of the Lower Coralline
161 Limestone (Gatt & Gluyas 2012).

162

163 Overall, the geological structure of the Maltese archipelago is characterised by subhorizontal or
164 gently dipping Tertiary successions offset by different high-angle fault systems. The NW-SE-
165 trending Maghlaq Fault, on the southwestern edge of the Island of Malta (Fig. 1B), is the main
166 exposed normal fault related to the Pantelleria Rift, mostly located offshore to the southwest of
167 the archipelago (Illies 1980) (Fig. 1A). The general gentle NE dip of the Tertiary strata in the
168 archipelago is attributed to upwarping and backtilting on the NE shoulder of the Pantelleria Rift

169 (Illies 1980, 1981; Grasso & Pedley 1985). An ENE-WSW trending graben system around 15
170 km wide spans from southeastern Gozo to the central sector of Malta, including Comino Island
171 and the straits between the three islands. This extensional structure, oblique with respect to the
172 adjacent Pantelleria Rift, is bounded by the North Comino Channel Fault (or South Gozo Fault,
173 of Grasso *et al.* 1986) to the north, and the Victoria Lines Fault to the south (Illies 1980) (Fig.
174 1B). The graben shows a concordant topography with a succession of flat-topped ridges (horsts)
175 and valleys (grabens) controlled by ENE-WSW secondary normal faults. According to Gardiner
176 *et al.* (1995), these originally normal faults may have reactivated as dextral strike-slip faults
177 during the Plio-Pleistocene, concurrently with the development of the North Gozo Graben.

178

179 In the western sector of Gozo there is a system of E-W oblique-slip normal faults. Some of these
180 faults offset vertically and laterally the annular collapse structures analysed in this work (Illies
181 1980, Oil Exploration Directorate 1993). Most probably, this fault system is related to the
182 development of the adjacent North Gozo Graben since the late Pliocene, which shows evidence
183 of recent faulting consistent with a right-lateral transtensional regime (Gardiner *et al.* 1995).
184 This is a unique example in which dissolution collapse structures may be used as a structural
185 marker to identify and assess neotectonic deformation.

186

187 **Palaeosinkhole characteristics**

188 The Island of Gozo has 13 mapped collapse palaeosinkholes controlled by subvertical dip-slip
189 faults with circular to elliptical cartographic trace (Pedley 1974; Illies 1980; Oil Exploration
190 Directorate 1993; Soldati *et al.* 2013) (Fig. 1B). We analysed, nine of the largest structures,
191 mostly located in the western coastal sector of the island. Parameters including their size,
192 geometry, amount of vertical displacement, subsurface dissolution and relative spatial
193 distribution are shown in Table 1. Some palaeosinkholes reach exceptionally large dimensions
194 ranging from 65 m to 600 m and most of them have a nearly circular geometry. The sizes of the
195 palaeosinkholes generally decrease in area towards the northeast. The minimum vertical offset
196 values range from >15 m to >60 m. The minimum cumulative dissolution volume for the eight

197 palaeosinkholes with data is 45.25 hm³, roughly equivalent to a cube with an edge of 360 m.
198 The actual subsurface dissolution volumes may be substantially larger than the estimated ones
199 due to the following factors: (1) the structural throw used in the calculations is a minimum
200 value; (2) although strongly affected by dissolution, some stratigraphic levels may have not
201 collapsed; (3) the foundered sediments may have brecciated and increased their volume by
202 gravitational deformation due to the bulking effect (Gutiérrez & Cooper, 2013).

203

204 The Upper Coralline Limestone is the youngest stratigraphic unit affected by the collapse
205 structures. No deformed Quaternary deposits or landforms have been identified on land
206 associated with the faults bounding the palaeosinkholes. Quaternary stratigraphic and
207 geomorphic markers at the margins of the collapse structures are quite scarce. These data
208 suggest that at least some of the structures have been active sometime after the deposition of the
209 Late Miocene (Tortonian-Messinian) Upper Coralline Limestone, and that later they may have
210 been inactive or had a low subsidence rate.

211

212 Some collapses are offset by oblique-slip normal faults with strikes varying from ENE-WSW to
213 ESE-WNW. The subvertical collapse faults have been displaced horizontally by these tectonic
214 faults by as much as 46 m (Qawra palaeosinkhole) (Table 1). This evidence suggests that the
215 oblique-slip faults are younger than the gravitational collapse faults (post-Upper Coralline
216 Limestone). These oblique faults might still be active and are very probably related to the
217 formation of the ENE-WSW North Gozo Graben, which started in the Late Pliocene and shows
218 evidence of very recent activity (Gardiner *et al.* 1995).

219

220 Regarding the spatial distribution, the five palaeosinkholes in the western coastal sector of the
221 Island of Gozo are tightly clustered (Fig. 1B). The Xlendi and Tal Harrax structures, that have
222 elongated geometries and sharp changes in width, may correspond to the coalescence of
223 adjoining palaeosinkholes (Fig. 1B). Hereabouts, the development of the collapse structures
224 seems to have been controlled by NW-SE trending faults. The Dwejra North, Dwejra Bay and

225 Tal Harrax structures form a clear tightly-packed NW-SE alignment, and the major axis of the
226 latter elongated structure is oriented in the same direction (Fig. 2A). The rose diagram in Figure
227 2B shows the azimuth of the lines drawn between the centroid of each palaeosinkhole and its
228 nearest neighbour, plus the strikes of the faults mapped in the area. The lack of coincidence
229 between the prevalent NW-SE orientation inferred for the sinkholes and the strike of the tectonic
230 faults suggest that: (1) the development of the collapse structures may have been controlled by
231 NW-SE trending buried faults or joints with no cartographic expression; (2) the oblique-slip
232 faults that offset the whole Tertiary stratigraphic succession and the collapse structures may
233 have formed after the development of the palaeosinkholes.

234

235 Regarding the internal structure of the collapses, they are essentially downdropped blocks
236 bounded by steeply-dipping annular faults. Some structures seem to be controlled by a single
237 master ring fault (Il-Maxell, Xlendi, Qwara, Tal Harrax, Ghajn Abdul; Figs. 3, 4A, 5A and B,
238 6A and C, and 7) whereas others include nested failure planes with concentric (Qawra, Tal
239 Harrax, Dwejra Bay; Figs. 5A and C, 6A and 8) or exocentric arrangement (Dwejra North Bay;
240 Fig. 8). Small throw secondary faults (e.g., Il-Maxell, Xlendi and Qawra; Figs. 3, 4A and 5A)
241 and inward dips (e.g., Tal Harrax; Fig. 6B) are common in the periphery of the collapses.

242

243 The palaeosinkholes' infill shows the following characteristics (Pedley 1974; Pedley & Bennet
244 1985): (1) Some Miocene units are thicker in the collapsed blocks than in the surrounding areas
245 (Xlendi, Ghajn Abdul), and may also wedge out towards the rim of the palaeosinkholes (Il-
246 Maxell, Xlendi, Tal Harrax; Fig. 4C). (2) The strata in the collapsed plug typically display a
247 basin structure with centripetal dips (e.g., Xlendi, Tal Harrax, Ghajn Abdul; Figs. 4A, 6A and B,
248 and 7A). (3) The infill shows local changes in the depositional processes and facies. This is
249 clear in the cases of Xlendi (Fig. 4B and C) and Ghajn Abdul (Figs. 7C and 9A). (4) The
250 sediments of the palaeosinkhole fills abutting the marginal collapse faults may include
251 allocthonous blocks fallen from an adjacent submarine sinkhole scarp, which may be
252 accompanied by soft sediment deformation (Qawra, Tal Harrax, Ghajn Abdul; Figs. 5D, 6D and

253 9B).

254

255 The Dwejra area in western Gozo has a dramatic coastal landscape related to a cluster of four of
256 the studied palaeosinkholes and represents an area of special interest (Figs. 1B and 10). These
257 collapse structures are spatially associated with a peculiar monoclinal structure, already cited by
258 Pedley & Bennet (1985). A structure-contour map of the top of the Lower Coralline Limestone
259 has been constructed in order to characterise the fold (Fig. 10A). This is a west-facing
260 semicircular monocline with centripetal dips and a maximum radius of about 900 m, which
261 resemble half-a-structural basin. The lack of offshore data precludes elucidating whether the
262 structure corresponds to a complete structural basin, or to a tightly curved down-to-the west
263 monocline. The structural relief of the monocline, as indicated by the top of the Lower Coralline
264 Limestone contour map, is around 60 m. The measured dips on the Tertiary strata are lower than
265 10° (or horizontal) in the upper and lower limbs, and exceed 15° in the dipping limb. Dwejra
266 North palaeosinkhole, currently submerged by the sea, is located in the centre of the arcuate
267 monocline, whereas Qwara, Tal-Harrax and Dwejra Bay palaeosinkholes partially overlap the
268 dipping limb and/or the crest of the monocline (Fig. 10).

269

270 **Discussion**

271 The Island of Gozo shows unique examples of large collapse structures with dramatic
272 geomorphic expression related to differential erosion controlled by the contrasting erodibility
273 between the downdropped sediments and the surrounding bedrock (Soldati *et al.* 2013). These
274 gravitational structures have been interpreted as dissolution-induced submarine palaeosinkholes
275 that were active in the Miocene during the deposition of stratigraphic units overlying the Lower
276 Coralline Limestone. Two hypotheses have been proposed to explain their genesis: (1) collapse
277 of large caves developed within the Lower Coralline Limestone (Pedley 1974), and (2) deep-
278 seated dissolution of halite-bearing evaporites and collapse of the overlying caprock sediments
279 (Illies 1980).

280

281 According to Pedley's (1974) interpretation, the palaeosinkholes result from the collapse of
282 large caverns developed within the Lower Coralline Limestone. This author, inspired by the
283 "blue holes" of the Caribbean region, suggests that the cave systems formed during a probable
284 emergence phase in the Eocene or Oligocene period. Sea-level falls of ~50-60 m are reported in
285 the global sea-level curves at the Eocene-Oligocene boundary and in the Middle Oligocene
286 (Miller *et al.* 2011). Relative sea-level falls of 10 to 20 m are reported over Malta Platform in
287 the lower Chattian period by Gatt & Gluyas (2012). However, none appear to have been
288 sufficient to place western Gozo more than a few metres above sea-level.

289
290 Illies (1980) succinctly presented his evaporite dissolution interpretation without providing
291 much supporting data. Surprisingly, he used the oblique-slip faults that offset the
292 palaeosinkholes as diagnostic indicators of deep-seated salt dissolution and caprock collapse.
293 This highly questionable concept was based on the interpretation by Belderson *et al.* (1978) of
294 similar features recognized in the sea-floor of the eastern Mediterranean. According to these
295 authors, in their study area strike-slip faulting may have allowed sea water access to an
296 underlying salt layer triggering its dissolution. However, in our case study the oblique-slip faults
297 are superimposed on the paleocollapses, and consequently postdate them.

298
299 Below we discuss the origin of the Gozitan palaeosinkholes considering the available geological
300 data and analysing the similarities and differences with other large sinkholes and collapse
301 structures related to evaporite and carbonate dissolution documented worldwide, both onshore
302 and offshore.

303

304 *Subsidence kinematics and duration*

305 Overall, the available data indicate that the collapse process in the study area has operated over
306 a very long period of time in the late Cenozoic. The subsidence kinematics has probably been
307 dominated by progressive deformation, rather than by major episodes of gravitational faulting.
308 The study of the paleosinkholes infill provides multiple evidence of synsedimentary subsidence

309 in the sea-floor (Pedley 1974; Pedley & Bennet 1985): (1) Cumulative wedge-out arrangements
310 in the palaeosinkhole fill (Fig. 4C). (2) The boulders entombed in the latter indicate that there
311 were periods during which subsidence was not counterbalanced by aggradation (Figs. 5D, 6D
312 and 9B). (3) The depressions created by collapse subsidence in the sea floor controlled changes
313 in the depositional processes and facies. Moreover, in some palaeosinkholes synsedimentary
314 subsidence has been active over several periods. For instance, in Tal Harrax the available data
315 indicate that progressive subsidence has been active during deposition of the Globigerina
316 Limestone and the Blue Clay, probably spanning around 10 Ma in the Miocene (Fig. 6).
317 Stratigraphic evidence in Ghajn Abdul suggests that subsidence may have been active during
318 deposition of the Upper Coralline Limestone (Fig 7). Two collapse structures also record that
319 significant subsidence occurred after the sedimentation of the Upper Coralline Limestone
320 (Wardija Point, Wied il-Mielah; Fig. 11). Moreover, the nested enclosed depression 160 m
321 across and 4 m deep identified in the submerged floor of Dwejra North Bay suggests that some
322 collapses have been active in recent times (Fig. 8). Therefore, the analysed subsidence
323 phenomenon can be explained by progressive interstratal dissolution of soluble rocks at depth
324 over a very long time period, and the concomitant gradual collapse of the overlying formations.
325 These features related to the subsidence kinematics and duration seem to be more compatible
326 with interstratal dissolution of evaporites.

327

328 The stratigraphic relationships of the Crater Lake structure, and the Saskatoon low,
329 Saskatchewan, Canada, indicate that deep-seated salt dissolution and synsedimentary
330 subsidence have been active over several periods from the Late Cretaceous to the late
331 Pleistocene (Christiansen 1971; Christiansen & Sauer 2001). In the Delaware Basin, New
332 Mexico and Texas, USA, dissolution of Permian evaporites has generated depositional basins
333 filled with Neogene continental sediments that may reach 50 m in thickness (Hill 1996 and
334 references therein). Synsedimentary subsidence related to interstratal karstification of salt-
335 bearing evaporites has been also documented in a number of fluvial valleys, where Quaternary
336 alluvium shows substantial thickenings in dissolution basins (Guerrero *et al.* 2013 and

337 references therein; Gutiérrez & Cooper 2013). In contrast, the sedimentary fill in large sinkholes
338 related to the collapse of limestone caves commonly shows parallel-layered strata indicating that
339 subsidence has not been active during deposition. A good example of this is the Holocene fill of
340 the Blue Hole of Light House Reef, Belize, where there is no evidence of growth strata
341 (Gischler *et al.* 2013). This steep-sided submarine sinkhole 120 m deep and 320 m wide was
342 formed in the late Pleistocene by the collapse of a water table cave (White 1988). Features
343 indicative of synsedimentary subsidence have also been documented in limestone karst
344 sinkholes, although with notable differences with those observed in the Gozitan collapses. For
345 instance, in lakes of north-central Florida related to the coalescence of sinkholes, high-
346 resolution seismic data reveal geometrical relationships in the deposits indicative of
347 synsedimentary subsidence, but these are spatially and temporally restricted, and related mainly
348 to suffusion processes and dissolution at the rockhead (Kindinger *et al.* 1999).

349

350 *Dimensions and deformation style*

351 A striking feature of the studied palaeosinkholes is their gigantic dimensions. In the Maltese
352 Archipelago, cave systems have been developed in the Lower Coralline Limestone during
353 Quaternary sea-level lowstands, and locally roof breakdown has resulted in the development of
354 collapse sinkholes (Pedley 1974; Pedley *et al.* 2002). There are several significant onshore and
355 offshore examples. Il-Maqluba sinkhole in southern Malta is an elliptical depression 100 m long
356 with vertical cliffs 30 m high. There are also well-known submarine caves (e.g., Blue Dome, N
357 Gozo) and sinkholes (e.g., Blue Hole, Dwejra area, W Gozo, Fig. 10A), the latter up to 25 m
358 across and 15 m deep (Lemon 2012). However, the size of these collapses, as well as their
359 subsidence mechanism and kinematics, are very different from those of the Miocene
360 palaeosinkholes (Tonelli *et al.* 2012). These data indicate that large cavities exist within the
361 Lower Coralline Limestone, but they are most probably much younger than the Miocene
362 palaeosinkholes and not large enough to form tightly clustered collapses several hundred meters
363 across.

364

365 The largest documented caverns and collapse sinkholes associated with carbonate rocks occur in
366 inland mature karsts developed in massive limestones that have a high mechanical strength,
367 usually in humid tropical regions. The largest known cavern is Sarawak Chamber in Gunung
368 Mulu National Park, Malaysia. This underground void is 700 m long, 400 m wide and its vault-
369 shaped ceiling reaches around 100 m in height (Waltham 2004, 2006). The largest collapse
370 sinkholes are the Tiankengs described in China. These are giant sinkholes related to the collapse
371 of large chambers, commonly associated with substantial cave rivers (Waltham *et al.* 2005;
372 Xuewen & Waltham 2006). Xiaozhar Tiankeng, also known as the Heavenly Pit, is 628 m long
373 and 662 m deep. Deep collapse dolines have been also documented in the Dinaric karst, like
374 Crveno Jezero sinkhole, Croatia. This is a vertical-walled collapse around 500 m wide and more
375 than 530 deep, half submerged under a lake (Garašić 2000). However, it is not clear whether this
376 collapse structure is rooted in carbonate or evaporite rocks.

377

378 Although the size of the palaeosinkholes in Gozo is comparable with those of the
379 aforementioned carbonate karst landforms from China, there are significant differences
380 regarding the subsidence mechanism. The roof of large caverns developed in limestone typically
381 propagates upward through successive collapses controlled by arched failure planes. Ground
382 subsidence does not occur until the stopping process reaches the surface, leading to the
383 catastrophic formation of a collapse sinkhole. In contrast, as the stratigraphic record reveals,
384 subsidence has affected the sea-floor over millions of years, indicating long-sustained deep-
385 seated dissolution and sinkhole rejuvenation. In the analysed palaeosinkholes this subsidence
386 has been accommodated by the progressive foundering of cylindrical plugs with limited internal
387 deformation. This is clearly observable in the case of Qawra structure (Fig. 5). The
388 downdropped blocks bounded by steeply-dipping annular faults of the palaeosinkholes are
389 similar to those documented in volcanic calderas related to the slow decline of the supporting
390 magma. According to Roche *et al.* (2001), the foundering of this type of piston-like integral
391 blocks develops when the roof has an aspect ratio (roof thickness/roof width) within a specific
392 range, which depends on the mechanical properties of the rocks. This deformation style of the

393 palaeosinkholes of Gozo is very similar to the subsidence mechanism in other collapse
394 structures related to interstratal dissolution of salt-bearing evaporites described in the literature.
395 The Crater Lake depression, Saskatchewan, Canada, is a circular topographic basin underlain by
396 a downthrown cylindrical block with two concentric fault zones 100 m and 200 m in diameter.
397 This collapse, with a vertical throw of 73 m, is related to interstratal dissolution of the halite-
398 bearing Middle Devonian Prairie Evaporite Formation, located at a depth of 914 m
399 (Christiansen 1971; Gendzwill & Hajnal 1971). The 45 km long structural basin of the
400 Saskatoon Low, Saskatchewan, Canada, is controlled by subvertical faults generated by
401 interstratal dissolution of Devonian salts, whose original thickness coincide with the maximum
402 throw of the structure (Christiansen 1967). This basin includes nested collapse structures
403 bounded by ring faults several kilometres in diameter (Christiansen & Sauer 2001). In the
404 Canyonlands section of the Colorado Plateau, USA, interstratal dissolution of evaporites in salt-
405 cored anticlines has produced collapse structures controlled by vertical failure planes with
406 circular and concentric forms. The downthrown blocks reach more than 500 m in diameter, are
407 rooted in evaporites at a depth of more than 600 m, and show vertical displacements up to 300
408 m (Weir *et al.* 1961; Cater 1970; Sugiura & Kitcho 1981; Gutiérrez 2004). In the Calatayud
409 Neogene Graben, Iberian Chain, NE Spain, interstratal karstification of halite- and glauberite-
410 bearing evaporites has generated collapses several kilometres across with a vertical
411 displacement of more than 200 m. Here the marginal faults of the collapse structures have an
412 irregular cartographic trace (Gutiérrez 1996, 2014).

413

414 Although the structure of the palaeosinkholes of Gozo have characteristics of deep-seated
415 collapses induced by interstratal evaporite dissolution, similar structures identified using 3D
416 seismic data have been attributed to the dissolution of carbonate rocks. Brown (1999)
417 recognized karst pits 200-500 m in diameter in deep burial Miocene carbonate rocks of the
418 Gippsland basin, southeastern Australia. Castillo and Mann (2006) identified subcircular
419 features up to 600 m wide and about 100 m deep in the Lower Cretaceous carbonates of the
420 Cogollo group in the southern Maracaibo Basin, Venezuela. These features were interpreted as

421 sinkholes formed subaerially by tropical weathering during a well-known eustatic drop in the
422 sea level that occurred during the Aptian period in the Maracaibo region. In the northern Fort
423 Worth Basin, Texas, McDonnell *et al.* (2007) identified subcircular collapse and sagging
424 structures 500-2000 m in diameter that extend vertically for 760-1060 m. They explain these
425 subsidence features as the coalescence of linked paleocaves in a limestone formation
426 (Ellenburger Group) with incremental collapse of the overlying strata. In the northern South
427 China Sea, 221 dissolution-collapse pipe structures 100-710 m wide and 134-1010 m deep were
428 identified by Sun *et al.* (2013). In the Gulf of Lions, SW France, Lofi *et al.* (2012) used seismic
429 profiles to image a subcircular collapse structure 2 km wide and 800 m deep with inward
430 dipping concentric faults and an inner sagging structure. The authors attribute the genesis of the
431 subsidence structures described in China and France to the collapse of limestone caves,
432 although they do not have direct data on the soluble rock units at depth.

433

434 *Spatial association with an arcuate monocline*

435 The Dwejra palaeosinkhole cluster is spatially associated with a peculiar west-facing
436 semicircular monocline suggesting a genetic link. The monocline has a radius of around 900 m
437 and a structural relief of approximately 60 m. Monoclines related to the down-dip migration of
438 dissolution fronts in evaporites and the consequent subsidence of the overlying strata (drape
439 gravitational folding) have been documented in numerous regions; Saskatchewan, Canada.
440 (Hopkins 1987; Anderson *et al.* 1988), NE England (Cooper 2002), southwest and central
441 sectors of the USA (De Mille *et al.* 1964; Walters 1978; Anderson *et al.* 1994; Neal and Colpitts
442 1997; Gutiérrez *et al.* 2014), central Saudi Arabia (Powers *et al.* 1966; Memesh *et al.* 2008),
443 Thailand (Supajanya and Friederich 1992), NE Spain (Gutiérrez *et al.* 2012a). Some of these
444 dissolution-induced monoclines are accompanied by large caprock collapse sinkholes. In the
445 Interior Homocline of central Saudi Arabia, dissolution of Late Jurassic anhydrite/gypsum units
446 has generated a sinuous monocline more than 500 km long, locally affected by large caprock
447 collapse sinkholes including Dahl Hit (Memesh *et al.* 2008; Gutiérrez & Cooper 2013). The
448 sinkholes of Bottomless Lakes State Park, New Mexico, USA, are a sinuous chain of collapse

449 sinkholes that traverse a dissolution-induced monoclinical scarp on the eastern margin of the
450 Pecos River valley (Land 2003). The Holbrook Anticline in northern Arizona is a monocline
451 generated by interstratal dissolution of halite- and sylvite-bearing Permian evaporites at more
452 than 200 m depth. A number of active subsidence depressions 2-3 km across with nested
453 sinkholes (e.g., McCauley Sinks, Richard Lake) have been identified linked to this gravitational
454 drape fold (Neal *et al.* 2013; Conway & Cook 2013).

455

456 *Soluble rocks in the subsurface*

457 The available borehole data indicate the presence of very thick carbonate units with dissolution-
458 related cavern formation throughout the Tertiary and Mesozoic succession underlying the
459 Globigerina Limestone (Pedley 1974). In contrast, the reported evaporitic units are very deep or
460 have a limited thickness. The Late Triassic anhydrite beds recorded in Madonna Taz-Zejt 1ST1
461 borehole have a cumulative thickness of around 230 m, but occur below 4.5 km depth (Debono
462 *et al.* 2000). The relative shallow gypsum bed found 100-150 m below the base of the Lower
463 Coralline Limestone in the Malta Platform is just 1 m thick (Gatt & Gluyas 2012). These data
464 challenge the evaporite dissolution hypothesis. Dissolution-induced collapse structures around 1
465 km deep have been reported by several authors (e.g., Lu & Cooper 1996; McDonell *et al.* 2007;
466 Lofi *et al.* 2012), but to our knowledge, never as deep as 4.5 km. However, there is the
467 possibility that during the development of the palaeosinkholes in the Miocene, there were halite-
468 bearing evaporites that have been largely removed by dissolution. As Warren (2006) indicates, a
469 large proportion of the evaporitic formations deposited in the Earth's history have been largely
470 removed from the rock record by subsurface dissolution, and in many cases the only evidence of
471 their previous existence corresponds to frequently overlooked pointers of karstification such as
472 insoluble residues or breccias. The size of the palaeosinkholes in Gozo decreases towards the
473 NE. This spatial trend might be related to some kind of stratigraphic control, such as the
474 wedging out towards the NE of the soluble rock responsible for the collapse structures.

475

476 *Mechanism of karstification*

477 Large collapse sinkholes are usually formed in geological settings where buried evaporites are
478 present (e.g., Paradox Valley and Delaware Basin, USA) or where large scale carbonate
479 dissolution has occurred (e.g., China and Dinaric karst). Two main modes of large-scale
480 carbonate dissolution have been documented in different karst settings: (1) karstification in a
481 subaerial or shallow-burial environment by meteoric waters in humid regions and (2) deep-
482 burial carbonate dissolution due to circulation of deep aggressive fluids. The first model cannot
483 be invoked for the generation of the Maltese paleocollapses because of their deep-seated origin
484 and evolution in a submerged environment. Several palaeosinkholes show syndepositional
485 subsidence during the sedimentation of the Globigerina Limestone Formation. This formation
486 records a long period spanning ca. 10 Ma (Aquitania-Langhian) of sedimentation in the sea
487 floor below the action of waves, interrupted by shallowing phases represented by hardgrounds
488 and phosphorite conglomerate beds also formed below sea level (Pedley *et al* 2002; Föllmi *et al.*
489 2007).

490

491 Deep-burial carbonate dissolution can result from the circulation of corrosive fluids. The
492 aggressiveness of these fluids on carbonate rocks may be enhanced by CO₂ (normally
493 geothermal fluids) or H₂S (basinal fluids, connate waters) enrichments that are characteristics of
494 active volcanic regions or areas adjacent to oil or gas fields (Ford & Williams, 2007).

495

496 Sinkholes and collapse structures related to CO₂ enriched geothermal fluids have been
497 documented in numerous regions including; dissolution-collapse pipes in the South China Sea
498 related to a magma intrusion (Sun *et al.* 2013), sinkholes of the Sistema Zacatón, Mexico (Gary
499 & Sharp 2006), the Obrucks of central Anatolia, Turkey (Bayari *et al.* 2009), and collapse
500 sinkholes of the San Vittorino Plain and Pontina Plain, Italy (Salvati & Sasowsky 2002).
501 However, volcanic activity in the Maltese archipelago was restricted to the Jurassic (Patacca *et*
502 *al.* 1979) and consequently it cannot be responsible for the Gozitan paleocollapses.

503

504 Large scale deep-burial carbonate dissolution may also be caused by acid pore waters resulting

505 from hydrocarbon degradation. Large caves have been formed due to this mechanism, including
506 the famous Carlsbad Caverns and Lechuguilla Cave, New Mexico. However, this alternative can
507 also be ruled out because, among other constraints, this process creates large caves at and
508 around the water table where the H₂S rising from depth mixes with O₂-rich meteoric waters
509 (Ford & Williams 2007). Therefore, it is not consistent with the submerged environment of
510 formation of the Gozitan palaeosinkholes.

511

512 The mixing of two fluids saturated in calcite may change the saturation state of the new
513 solution. According to Corbella & Ayora (2003) the mixing of two fluids in deep aquifers is
514 much more important for carbonate dissolution than the changes induced by the oxidation of
515 H₂S. According to their calculations, this is due purely to the mixing effect. No dissolution at the
516 scale of karst conduits would be expected by mixing two fluids with temperatures that differ
517 less than 10°C. Although these mechanism have been invoked to enhance porosity in deep
518 carbonate reservoirs, the modality and extent of karst features possibly formed by burial
519 carbonate dissolution are not well known and have been recently questioned by Ehrenberg *et al.*
520 (2012).

521

522 Conversely, evaporite dissolution in sub-sea burial environments is still poorly understood, but
523 it has been documented in the Eastern Mediterranean basin (Ross & Uchupi 1973; Kastens &
524 Spiess 1984; Bertoni & Cartwright 2005) and in the North Sea (Lohmann 1972; Jenyon 1983).
525 Thus, the mechanism of karstification associated with the Maltese palaeosinkholes may be
526 related to dissolution and removal of evaporites by sub-sea groundwaters. However, the
527 paleohydrogeological model responsible for the karstification of evaporites is difficult to
528 explain due to the shortage of data. A model involving groundwater discharge into the sea
529 cannot be invoked because the study area was an isolated carbonate platform when the Gozitan
530 palaeosinkholes started to form (Gatt & Gluyas 2012). A possible conceptual model adapted to
531 this situation is a subjacent dissolution related to focused vertical hydrothermal flow favoured
532 by seismic activity (*cf.* Bertoni & Cartwright 2005 for a detailed explanation). According to this

mechanism, the dissolution and subsidence that led to the development of the palaeosinkholes may be correlated to tectonically active periods, as reported by Gardiner *et al.* (1995), and not to sea level changes. This would explain why dissolution-induced subsidence is recorded by the Globigerina Limestone. Deposition of this formation was coeval to periods of activity on the ENE-WSW fault system (Pedley *et al.* 2002). The development of growth faults (Illies 1980) and the opening of Neptunian dykes (cracks formed in the sea floor into which sediment fell; Pedley *et al.* 2002), provide evidence of tectonic deformation affecting the sea floor. The current null, or very low, active subsidence in the palaeosinkholes may be attributed to a tectonic stabilization of the region and/or to the almost total dissolution of the evaporite horizons.

So far, the data available are quite limited to propose a robust hypothesis on the origin of the analysed palaeosinkholes. However, the data integrated in this work, together with the discussed points, provide a basis for future studies on the gigantic Gozitan paleosinkholes and collapse structures in other regions worldwide.

Conclusions

The detailed analysis of the well-exposed palaeosinkholes of Gozo Island provides novel insights into the geometry, internal structure, kinematics and origin of the collapse structures. The new knowledge frame, together with data from a number of large dissolution-induced collapse structures documented worldwide, suggest that deep-seated evaporite dissolution is the most satisfactory genetic hypothesis. This interpretation is supported by the following lines of evidence:

(1) The progressive foundering of cylindrical blocks with limited internal deformation over periods lasting of the order of 10 Ma. This indicates gradual removal of large volumes of evaporite rocks and the concomitant settlement of the overlying strata, rather than the development of large caverns and their upward propagation by stoping, characteristic of limestone karst.

(2) The dimensions and deformation style of the palaeosinkholes are similar to other collapse

561 structures related to interstratal dissolution of salt-bearing evaporites.

562 (3) The arcuate monocline associated with some of the studied paleosinkholes. This spatial

563 association is commonly found in areas underlain by evaporitic formations affected by deep-

564 seated dissolution.

565

566 Notwithstanding, no major evaporite beds have been documented so far in the subsurface

567 stratigraphy of the Malta Platform. Future subsurface exploration (e.g., deep boreholes and

568 geophysics), as well as petrological and geochemical analyses of Oligocene-Eocene carbonate

569 units may provide new data on potential markers of dissolved evaporites or deep-burial

570 carbonate dissolution. These studies and the examination of currently inaccessible exploration

571 drillholes (e.g. Madonna Taz-Zejt well) might shed some light on the origin of these enigmatic

572 structures and test the presented hypothesis.

573

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579

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816 **Tables**
817

818 **Table 1.** *Parameters related to the size, geometry, amount of vertical displacement, subsurface*
819 *dissolution volume, relative spatial distribution and geomorphic expression of the nine selected*
820 *palaeosinkholes of western Gozo (note 1hm³ is 1,000,000 m³).*

Palaeosinkhole name	Major axis (m)	Minor axis (m)	Elongation ratio	Area (m ²)	Throw (m)	Volume of dissolved karst rocks (hm ³)	Youngest sediments displaced	Distance to the nearest collapse (m)	Horizontal offset on cross-cutting tectonic faults (m)	Geomorphic expression
Qawra	368	365	1.01	110,085	>60	>6.61	Blue Clay	75	46	depression
Tal Harrax	596	556	1.07	325,064	~100	>16.25	Blue Clay	30	16	depression
Xlendi	330	278	1.19	112,171	>50	>6.17	Globigerina Limestone	1,600		depression
Dwejra North	337	311	1.08	82,088	>40	>3.28	Globigerina Limestone?	38	/	bay
Dwejra Bay	380	351	1.08	120,762	>40	>4.83	Globigerina Limestone?	30	16	bay
Il-Maxell	>120	>120	?	?	>15	?	Globigerina Limestone	1,290	/	depression
Ghajn Abdul	440	371	1.18	130,575	>50	>6.53	Upper Coralline Limestone	550		mesa
Wardija Point	65	48	1.35	2,497	>50	>0.12	Upper Coralline Limestone	845	/	butte
Wied il-Mielah	181	176	1.03	24,389	>60	>1.46	Upper Coralline Limestone	2,180	/	butte

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823

824 Figure captions

825 **Fig. 1.** (A) Sketch showing the geotectonic setting of the Maltese archipelago, and the
 826 main deep boreholes drilled in and around the islands. (B) Simplified geological map of
 827 the Maltese archipelago showing the distribution of stratigraphic units and faults, as
 828 well as the palaeosinkholes of Gozo Island. On the right, close up of the NW sector of
 829 Gozo indicating the name of the collapse structures analysed in this work.

830
 831 **Fig. 2.** (A) Sketch of the western sector of Gozo Island showing the NW-SE prevalent
 832 distribution of the palaeosinkholes. (B) Rose diagram representing azimuth of the lines
 833 between the centroid of each palaeosinkhole and that of the nearest one (black color)
 834 and the strike of the faults mapped in the western sector of Gozo (grey color). The
 835 dashed circle corresponds to one measurement.

836
 837 **Fig. 3.** The Il-Maxell palaeosinkhole. A: Main NW-SE-oriented exposure showing the
 838 western edge of the collapse structure controlled by arcuate master and secondary faults.
 839 B: Northern side of the exposure and sketch. C: Southern exposure with sketch. The Il-
 840 Maxell structure is the only palaeosinkhole of Gozo in which it is possible to observe
 841 the deformation style in the sediments underlying the depression fill, a mosaic
 842 packbreccia (Warren 2006), largely consisting of meter-sized slab-like blocks of
 843 fragmented and dislocated tabular beds.

844
 845 **Fig. 4.** (A) Detailed geological map and cross-section of the Xlendi palaeosinkhole,
 846 offset by oblique dextral-normal tectonic faults. Several small throw radially disposed
 847 dip-slip faults in the structure affect the sinkhole infill. (B) General view of the Xlendi
 848 collapse from the south. (C) Cumulative wedge-out and progressive upward dip
 849 attenuation in the Globigerina Limestone strata of the palaeosinkhole fill (NE margin).

850
 851 **Fig. 5.** (A) Detailed geological map of the Qawra palaeosinkhole. Most of the collapse
 852 depression is underlain by the Blue Clay, largely masked by vegetation, but where
 853 exposed these sediments show an undisturbed dominant subhorizontal structure. Two
 854 samples were collected from this stratigraphic unit for planktonic foraminifera
 855 assemblage biostratigraphy (the asterisk indicate the location of samples). The lower
 856 sample contained *Paragloborotalia siakensis*, *Dentoglobigerina altispira* and
 857 *Globigerinoides* spp. The upper one had more and better developed *P. siakensis* and
 858 scarce *Paragloborotalia partimlabiata*. According to Abels *et al.* (2005), the first
 859 occurrence of *P. partimlabiata* is found in the upper part of the Blue Clay suggesting
 860 that the exposed sediments correspond to the middle and upper sections of the Blue
 861 Clay (B) View of the NE sector of the erosional depression excavated within the
 862 palaeosinkhole. The steep scarp on Lower Coralline Limestone corresponds to the
 863 exhumed rim fault of the collapse structure. The vegetated slopes at the foot of the scarp
 864 are underlain by softer Blue Clay and Upper Globigerina Limestone. The lagoon is
 865 connected to the sea through a fault controlled cave. (C) Southeast margin of the
 866 collapse, where Upper Globigerina Limestone (vegetated slope) is juxtaposed to Lower
 867 Coralline Limestone (bare rock exposure). (D) Boulder of reworked Lower Globigerina
 868 Limestone incorporated within the Upper Globigerina deposited in the palaeosinkhole.
 869 The boulder is interpreted as a block fallen from the scarped margin of the submarine
 870 sinkhole in Miocene time. LC: Lower Coralline Limestone; LG: Lower Globigerina
 871 Limestone; UG: Upper Globigerina Limestone.

872

873 **Fig. 6.** (A) Detailed geological map of the Tal Harrax palaeosinkhole. (B) Sagging strata
 874 of Upper Globigerina Limestone at the southernmost sector of Tal Harrax depression.
 875 (C) Tal Harrax rim fault outcrop. LC: Lower Coralline Limestone; LG: Lower
 876 Globigerina Limestone. (D) Boulders of Globigerina Limestone (GL) entombed within
 877 Blue Clay (BC) deposits at the centre of Tal Harrax structure. This Blue Clay outcrop
 878 contains a high detrital fraction, including abundant glauconite clasts, suggesting that it
 879 may correspond to the upper part of the formation, close to the contact with the
 880 Greensand Formation (Giannelli & Salvatorini 1975). This interpretation is supported
 881 by the planktonic foraminifera assemblage in two samples collected for this
 882 investigation (*P. siakensis*, *P. partimlabiata* and *Globorotalia menardii*).
 883

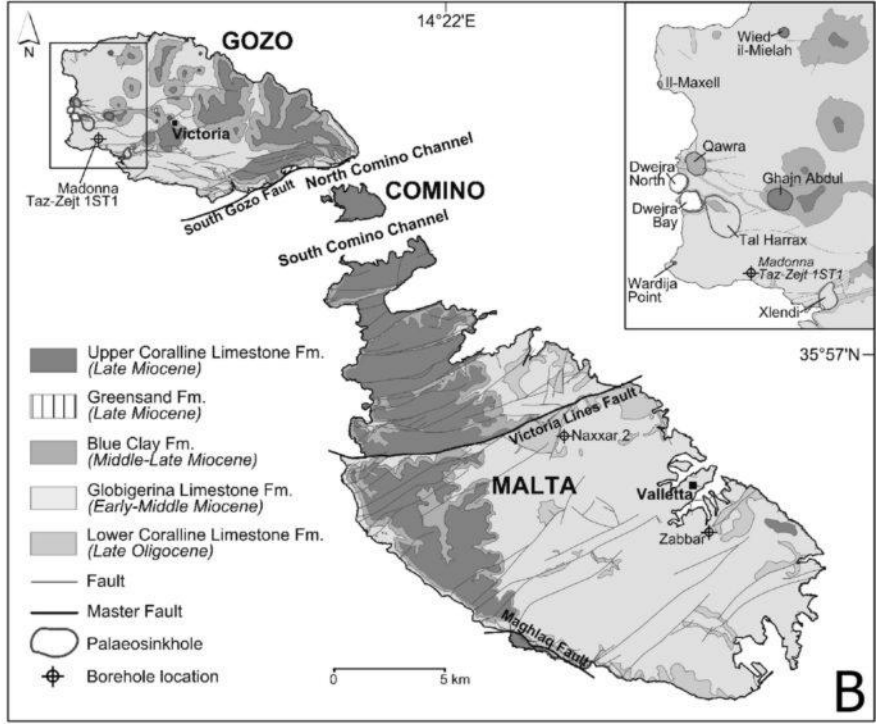
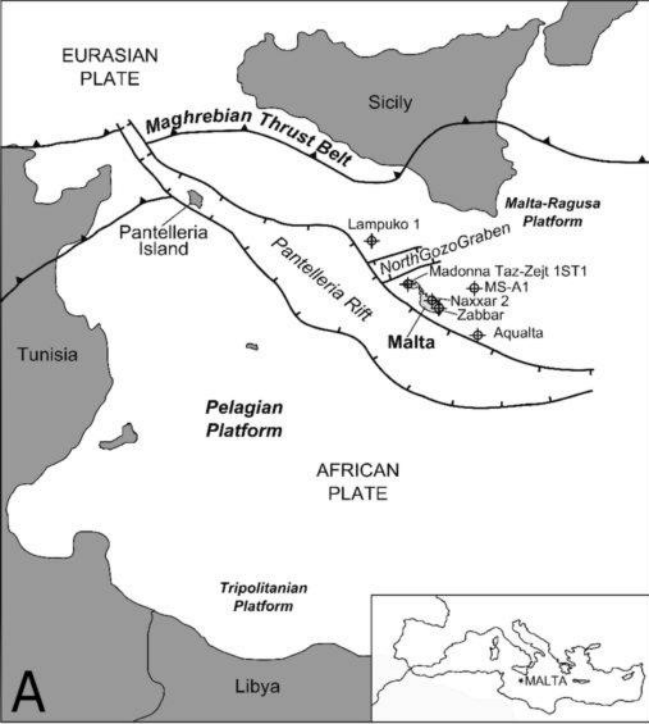
884 **Fig. 7.** Detailed geological map of the Ghajn Abdul palaeosinkhole (A) and cross-
 885 section (B). The numbers in the geological map indicate the stratigraphic sections
 886 shown in C. (C) Stratigraphic sections recorded in Ghajn Abdul palaeosinkhole (1, 2
 887 and 3) and in the adjacent mesa.
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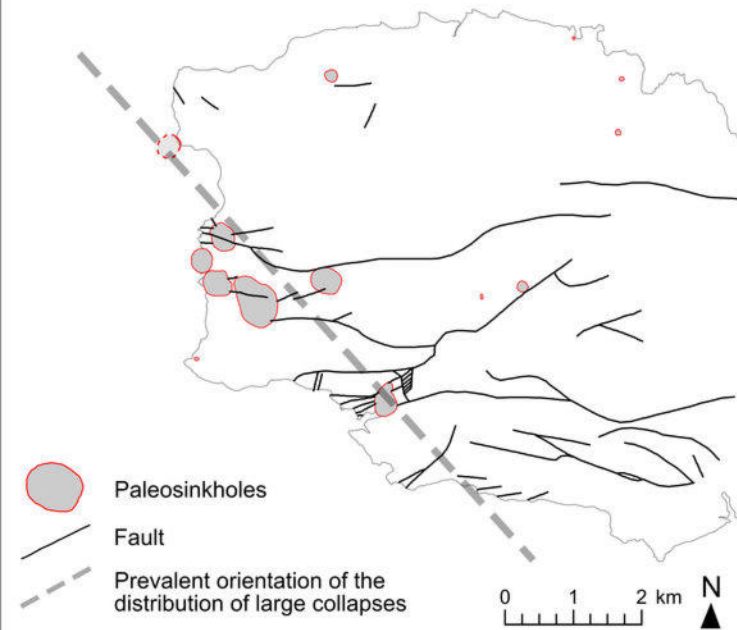
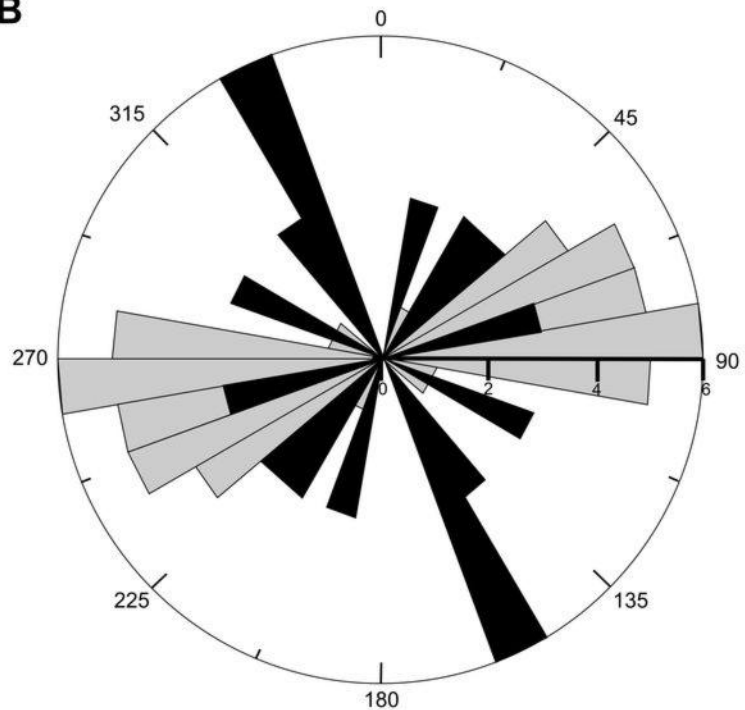
889 **Fig. 8.** Digital elevation model generated from LiDAR data, showing the topography of
 890 the sea floor at Dwejra North Bay and Dwejra Bay. Note the steep-sided enclosed
 891 depression in the NE sector of the bay controlled by the collapse structure. The bottom
 892 of the closed depression has a marginal trough and a protruding central sector with
 893 blocky appearance.
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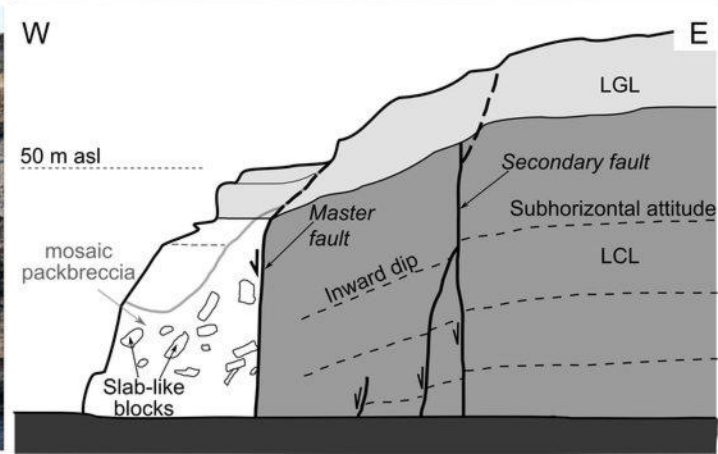
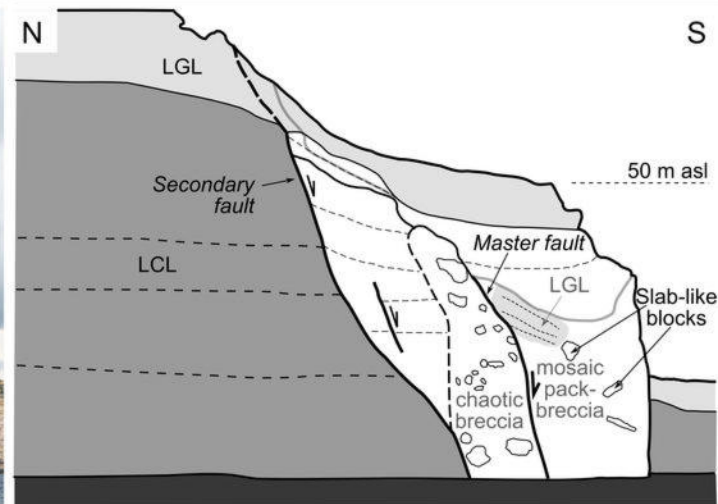
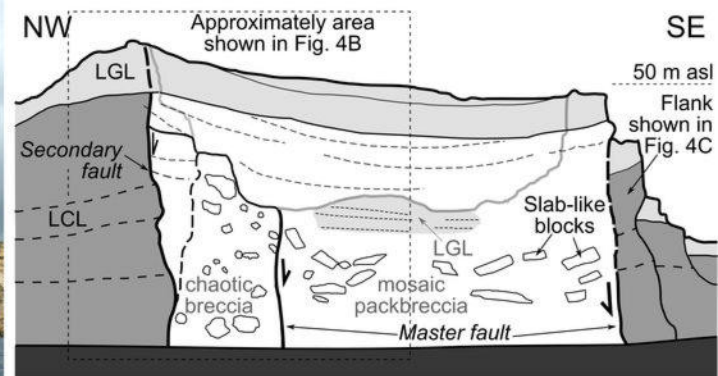
895 **Fig. 9.** (A) Upper part of Ghajn Abdul mesa capped by Upper Coralline Limestone.
 896 Note the low-angle cross stratified packstones and grainstones overlain by bioturbated
 897 massive limestone at the top. (B) Load casts and contorted bedding in the cross
 898 stratified packstones and grainstones associated with large meter-sized allochthonous
 899 boulders of grainstones containing pectinids, echinoderms, oysters and gastropods.
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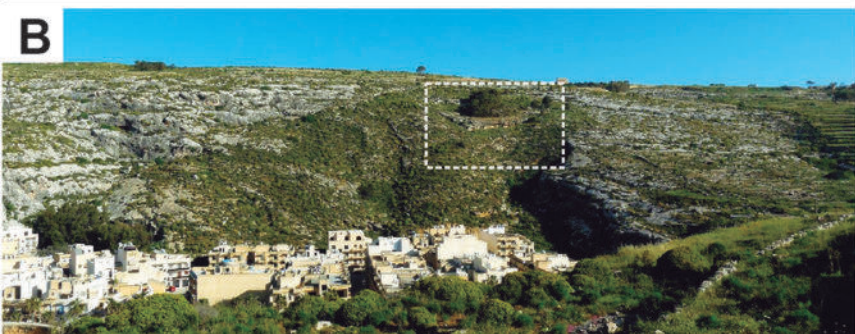
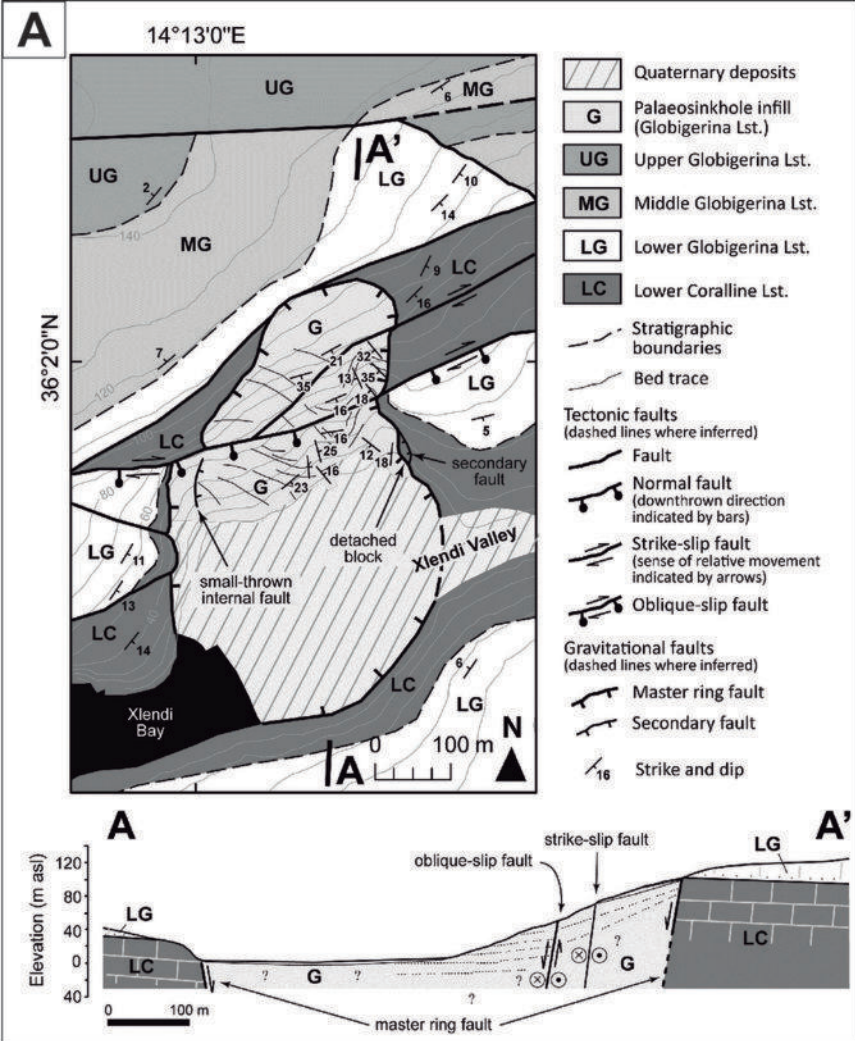
901 **Fig. 10.** (A) Structure contour map of the top of the Lower Coralline Limestone
 902 illustrating the arcuate west-facing monoclinical structure of Dwejra area. (B) General
 903 view of the west sector of the monocline from the sea. (C) Detail of the attitude of the
 904 strata at Fungus Rock.
 905

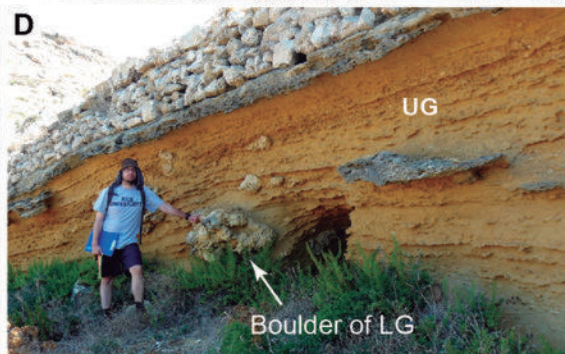
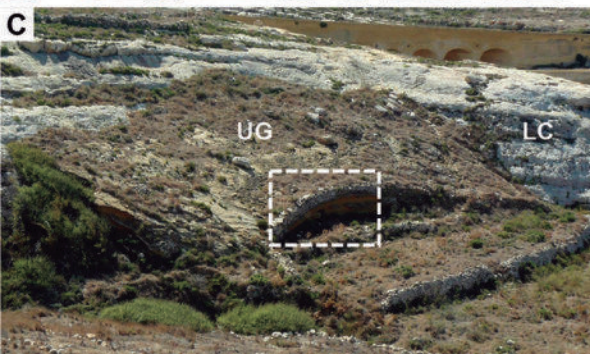
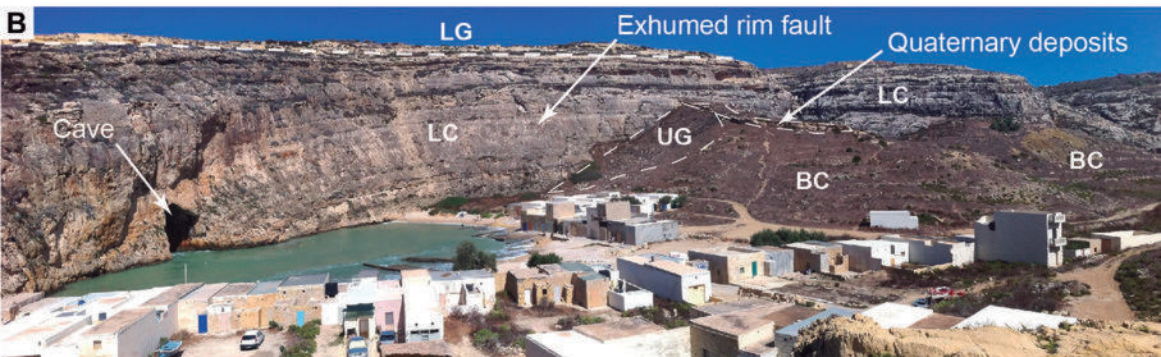
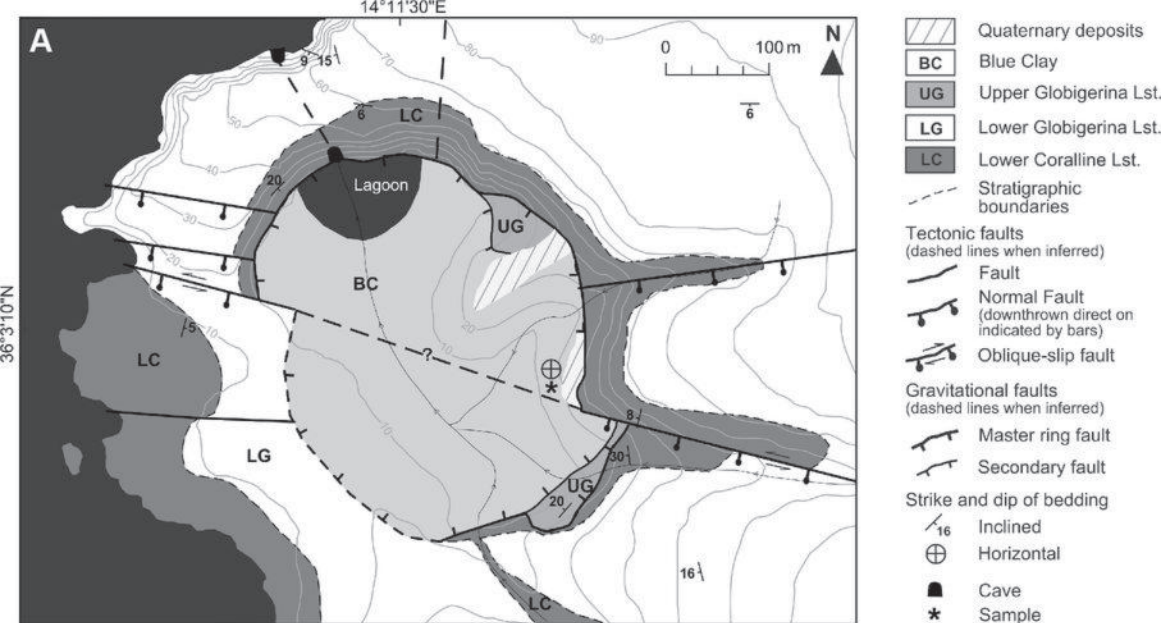
906 **Fig. 11.** Aerial orthoimages of Wied il-Mielah (A) and Wardija Point (B)
 907 palaeosinkholes. The cartographic relationships observed in these structures indicate
 908 that the collapses were active after the deposition of the Upper Coralline Limestone in
 909 Pliocene and probably Quaternary times. UCL: Upper Coralline Limestone; GL:
 910 Globigerina Limestone.
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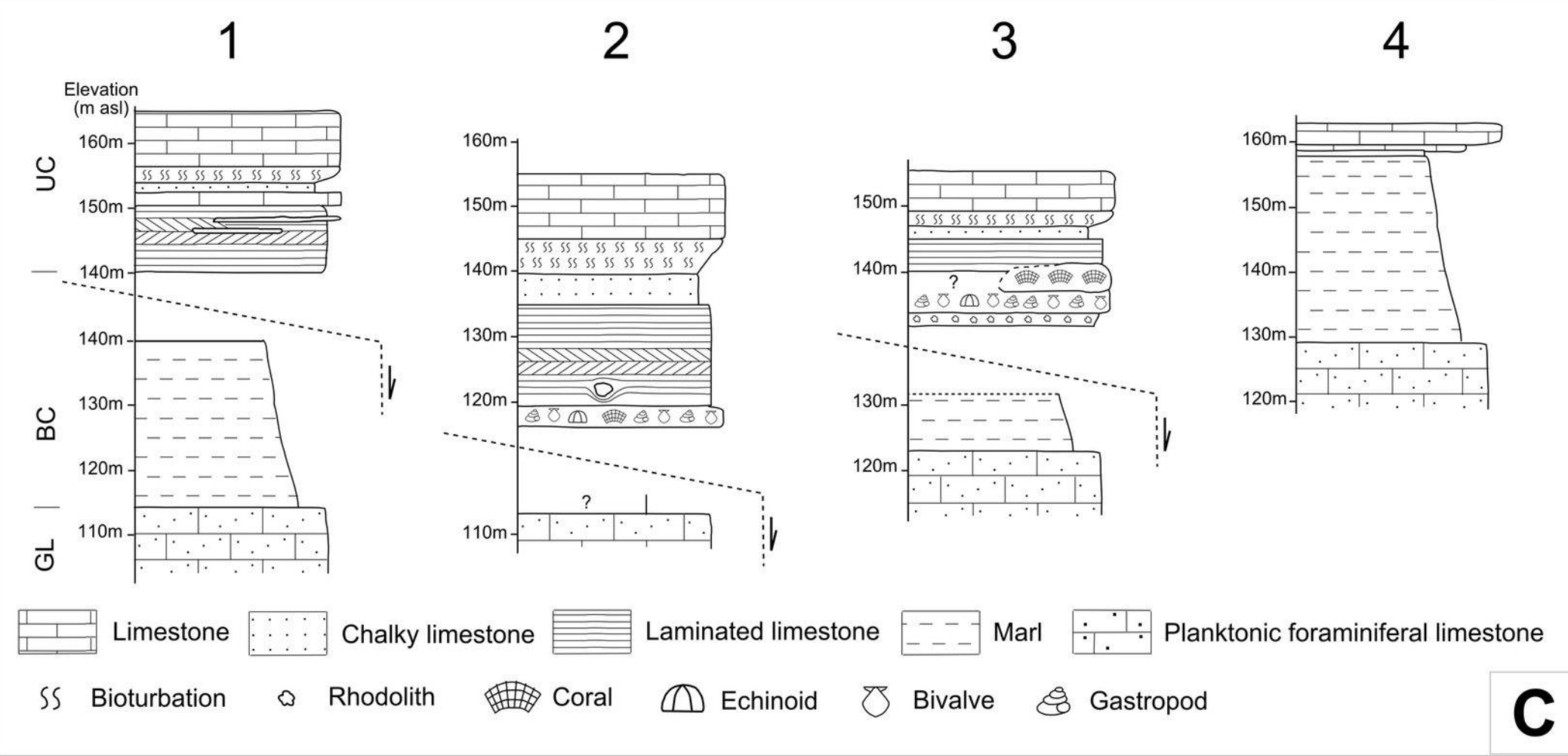
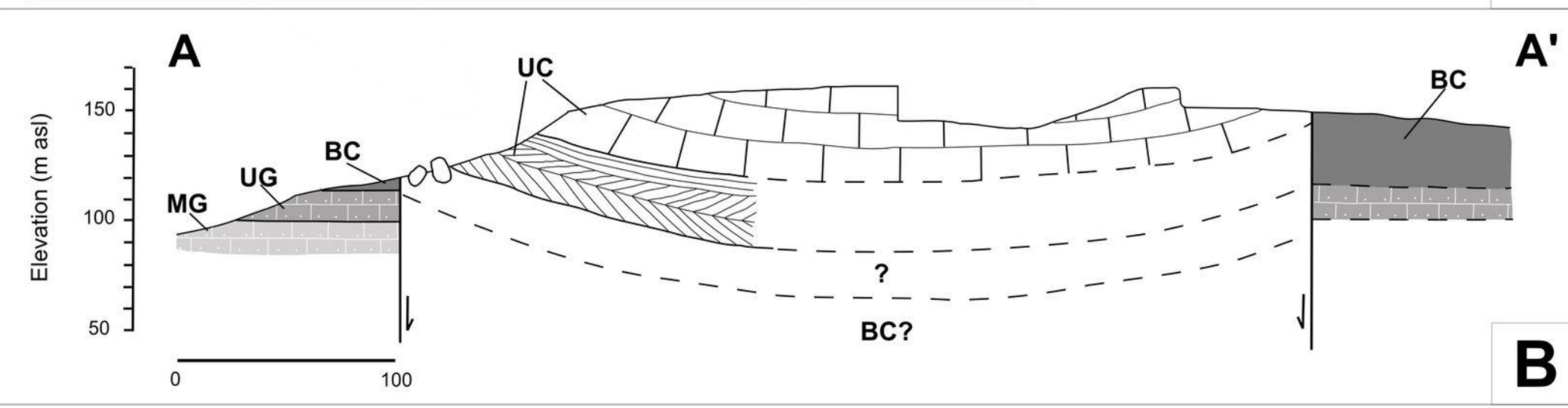
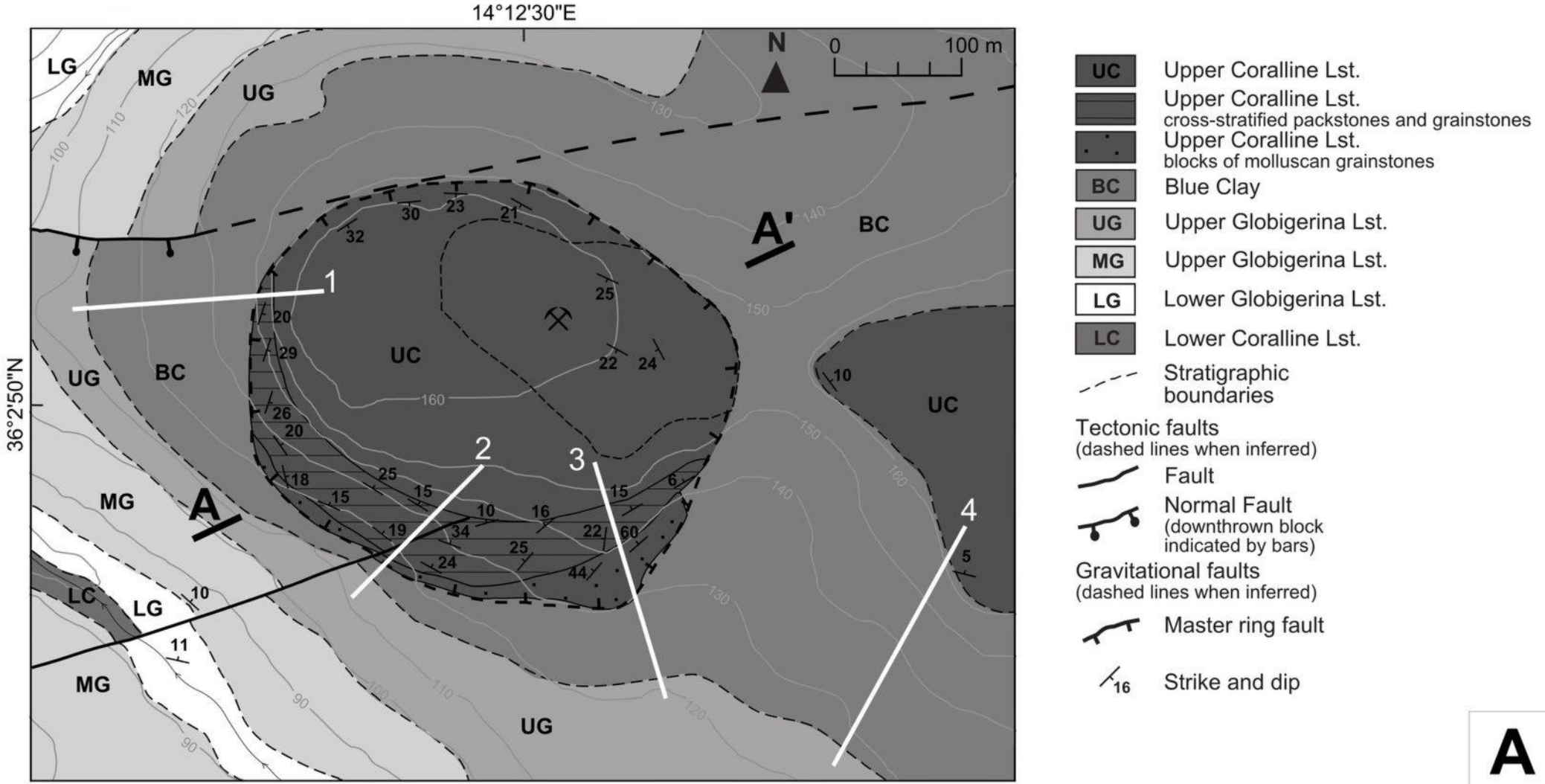


A**B**









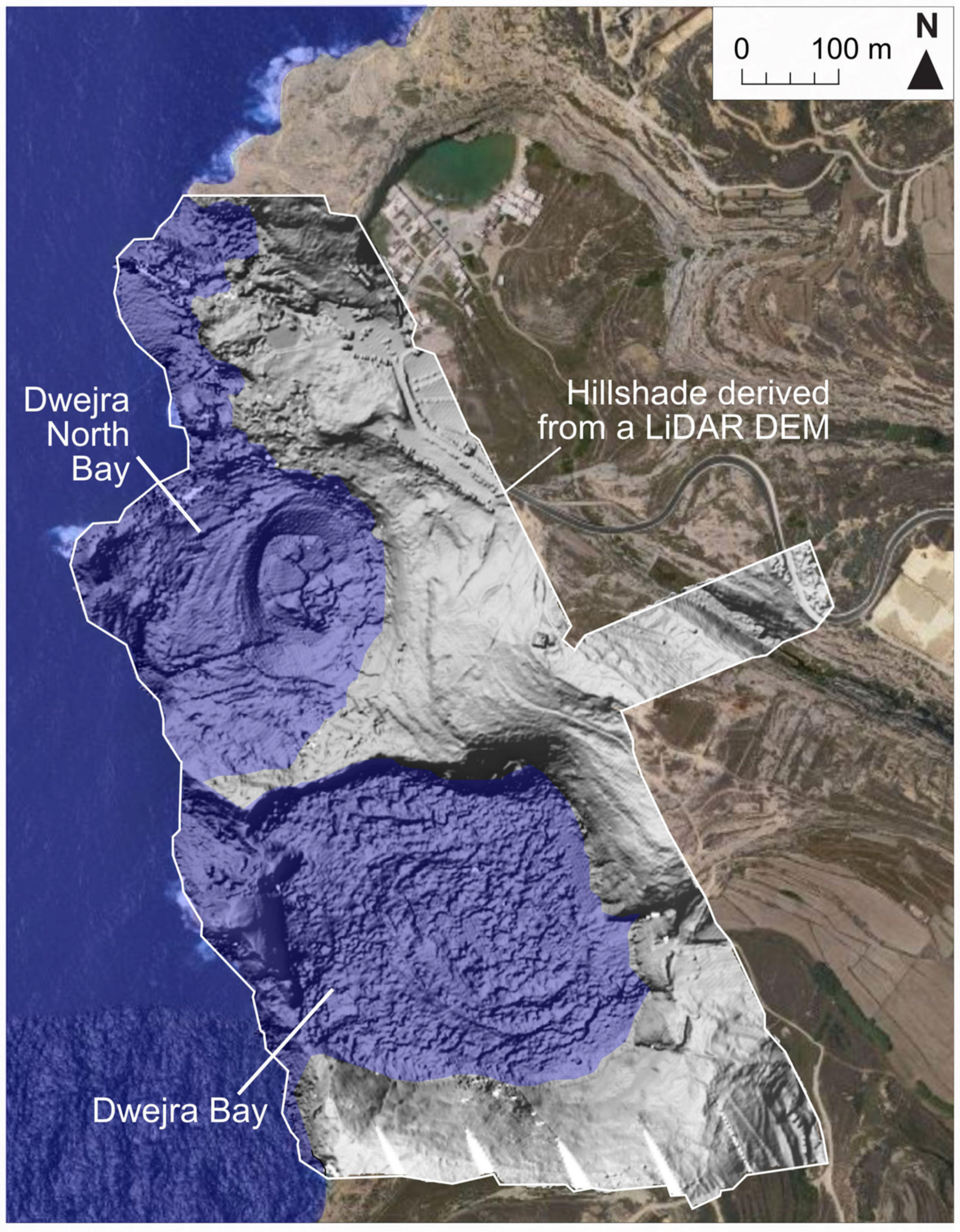
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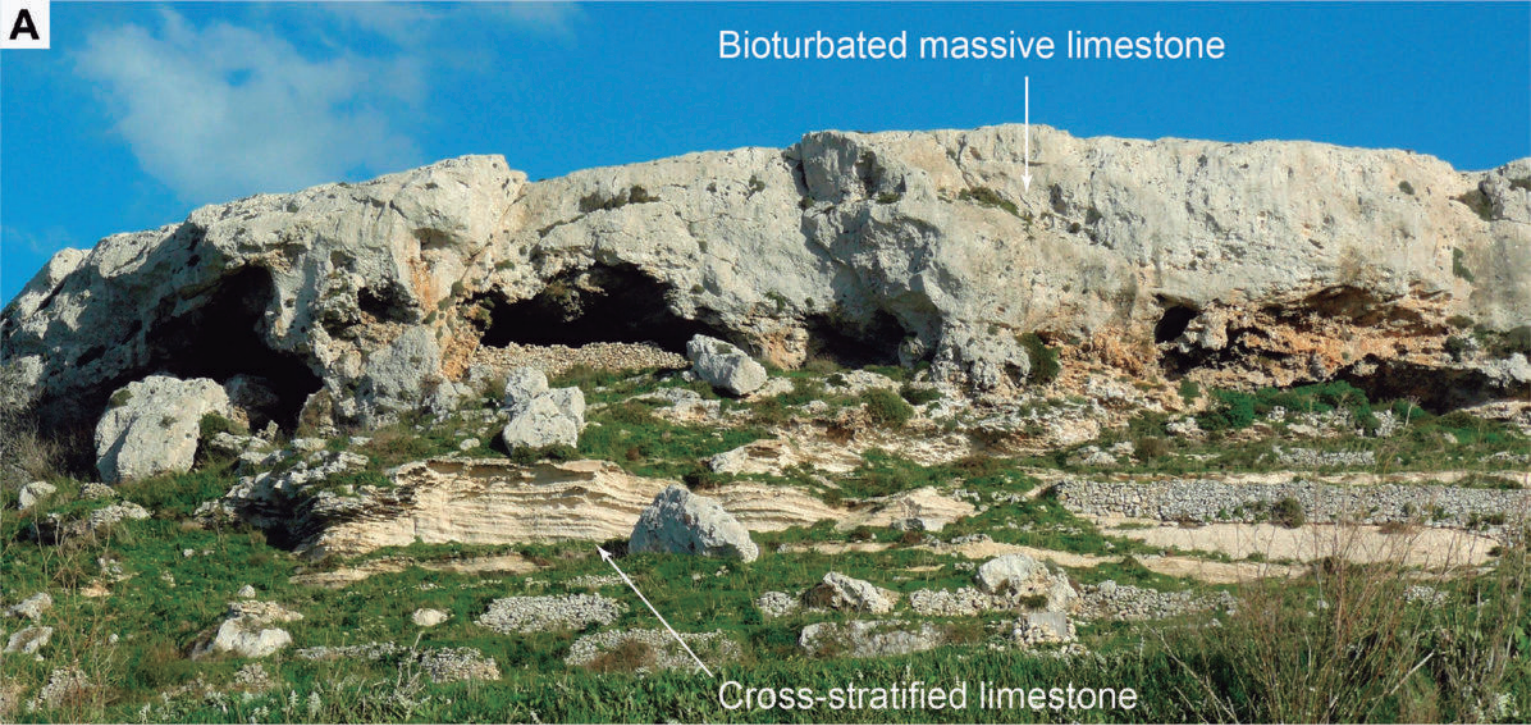


Dwejra
North
Bay

Hillshade derived
from a LiDAR DEM

Dwejra Bay



A**B**

