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## **The Miocene – Pliocene boundary and the Messinian Salinity Crisis in the easternmost Mediterranean: insights from the Hatay Graben (Southern Turkey).**

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3

### 4 **Abstract**

5 The Hatay Graben is one of three easternmost basins in the Mediterranean that preserve sediments that  
6 span the Miocene-Pliocene boundary, including gypsums from the Messinian Salinity Crisis (MSC). Here  
7 we integrate existing data and present new sedimentological and micropalaeontological data to investigate  
8 the palaeoenvironments of late Miocene to early Pliocene deposits and place this important area into a  
9 regional stratigraphic framework. Six sections are described along a ~ W – E transect illustrating the key  
10 features of this time period. Late Miocene (Pre-MSC) sediments are characterised by open marine marls  
11 with a benthic foraminiferal fauna suggestive of water depths of 100 – 200 m or less. Primary lower  
12 gypsum deposits are determined to be absent from the graben as sedimentological and strontium isotopes  
13 are characteristic of the resedimented lower gypsums. The intervening Messinian erosion surface is  
14 preserved near the basin margins as an unconformity but appears to be a correlative conformity in the  
15 basin depocentre. No Upper Gypsums or ‘Lago–Mare’ facies have been identified but available data do  
16 tentatively suggest a return to marine conditions in the basin prior to the Zanclean boundary. Sediments  
17 stratigraphically overlying the Messinian gypsums and marls are coarse-grained sandstones from coastal

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18 and Gilbert-type delta depositional environments. The Hatay Graben is not only strikingly similar to  
19 Messinian basins on nearby Cyprus but also to the overall model for the MSC, demonstrating the  
20 remarkable consistency of palaeoenvironments found in marginal basins across the region at this time.  
21 This research also raises questions as to the timing of the Mediterranean reflooding and the significance  
22 of the widespread mega-breccias of the resedimented gypsum deposits.

23

24 *Keywords:* Messinian Salinity Crisis; Turkey; Eastern Mediterranean; Gypsum; foraminifera; Gilbert-type  
25 delta.

26

## 27 **1. Introduction**

28 The Messinian Salinity Crisis (MSC) was a dramatic event (~ 5.9 Ma) that affected the whole  
29 Mediterranean region when the seaways connecting the Mediterranean Basin to the Atlantic Ocean closed  
30 due to uplift in the Betic Arc/Moroccan Rif region (e.g., Duggan et al., 2003; Sierro et al., 2007). Isolation  
31 from the Atlantic Ocean resulted in the deposition of thick evaporite deposits in basin depocentres and  
32 significant erosion around the fringes of the Mediterranean. Studies of onshore Messinian strata preserved in  
33 basins described as either marginal or peripheral (to the deep, central Mediterranean Basins; Fig. 1) have  
34 provided much information on the sedimentology, palaeontology and geochemistry of the period, especially  
35 when combined with recent high-resolution cyclostratigraphic studies (e.g., Hilgen and Krijgsman, 1999;  
36 Sierro et al., 2001; Hilgen et al., 2007; Manzi et al., 2013).

37 The MSC resulted in the deposition of characteristic sedimentary units both in marginal (shallow)  
38 and deep water environments; however, until recently there were a number of contrasting models that  
39 attempted to link marginal and deep basin stratigraphy (Butler et al., 1995; Clauzon et al., 1996; Riding et  
40 al., 1998; Krijgsman et al., 1999; Rouchy and Caruso, 2004; Roveri et al., 2008b). A new scenario proposed  
41 by the CIESM (the Mediterranean Science Commission) consensus report (2008) develops a correlation  
42 scheme that integrates recent sedimentary facies and stratigraphic data from the marginal basins with deep  
43 basin seismostratigraphy in order to try to resolve these correlation problems. Furthermore, Roveri et al.  
44 (2014a, b) demonstrate that strontium isotope ratios ( $^{87}\text{Sr}/^{86}\text{Sr}$ ) provide additional stratigraphic constraints as  
45 distinct populations of  $^{87}\text{Sr}/^{86}\text{Sr}$  values have been documented during the different phases of the MSC event.  
46 This revised Messinian scenario is described within the framework of a 3-stage stratigraphic model

47 constructed mainly with observations from the marginal to intermediate basins exposed onshore in Sicily and  
48 in the Northern Apennines (CIESM, 2008; Roveri et al., 2014a, b).

49 However, despite the extensive ‘back-catalogue’ of work on the Messinian stage (e.g., Roveri et al.,  
50 2014a), many studies from the easternmost extent of the Mediterranean have focussed on Cyprus (e.g.,  
51 Robertson et al., 1995; Krijgsman et al., 2002; Kouwenhoven et al., 2006; Orszag-Sperber et al., 2009;  
52 Manzi et al., 2015) and adjacent IODP data (e.g., Blanc-Valleron et al., 1998; Pierre et al., 1998), with  
53 limited data from southern Turkey (Melinte-Dobrinescu et al., 2009; Darbaş and Nazik, 2010; Poisson et al.,  
54 2011; Cipollari et al., 2013; Faranda et al., 2013; Radeff et al., 2015). In this paper we focus on late Miocene  
55 and early Pliocene sediments of the Hatay Graben (southern Turkey), previously identified by Boulton et al.  
56 (2007, 2008) and Tekin et al. (2010). The Hatay Graben is one of the easternmost marginal basins (the other  
57 being the Syrian Nahir el-Kabir half-graben) that records evidence from this period, and is the ideal location  
58 for investigating the progression of the Messinian salinity crisis and the Zanclean reflooding event in the  
59 most distal part of the Eastern Mediterranean basin (Fig. 2). Here we examine key Tortonian, Messinian and  
60 Zanclean sections, some of which have been previously documented by Boulton et al. (2007) or Tekin et al.  
61 (2010), along with new sedimentological and micropalaeontological data to develop a facies and  
62 palaeoenvironmental model for the Hatay. The aims of the study are to: a) investigate the nature of the  
63 Miocene-Pliocene boundary in this marginal basin, b) place these sediments into the revised stratigraphy of  
64 the MSC (e.g., CIESM, 2008; Roveri et al., 2014a,b), and c) test the applicability of this model in the  
65 easternmost Mediterranean.

66

## 67 **2. Messinian Stratigraphic Framework**

68 The Global Stratotype Section and Point (GSSP) of the base Messinian is defined as the first  
69 occurrence of the planktic foraminifera *Globorotalia miotumida* in the Oued Akrech section (Morocco) at  
70 7.25 Ma (Hilgen et al., 2000). The top of the Messinian is defined by the Zanclean GSSP at Eraclea Minoa  
71 (Sicily), coincident with the base of the Trubi marls and the reflooding of the Mediterranean at 5.33 Ma.

72 The early Messinian (7.25 to 5.97 Ma) is characterised by the change in circulation patterns and  
73 water chemistry caused by progressive restriction of the Atlantic-Mediterranean corridors. Early Messinian  
74 sediments are usually characterized by cyclical stacking pattern, which include diatomites and sapropels

75 (e.g., Kouwenhoven et al., 2006), and show stepwise reductions in the diversity of planktic foraminifera  
76 (Sierro et al., 1999; Blanc-Valleron et al., 2002; Sierro et al., 2003; Kouwenhoven et al., 2006). These  
77 changes in diversity have been interpreted as the effect of 400 kyr orbital forcing superimposed on the  
78 tectonically controlled closure of the connecting oceanic gateway (Kouwenhoven et al., 2006).

79 Stage 1 (5.97 – 5.6 Ma) of the MSC is characterised by the widespread onset of evaporite  
80 precipitation only in the shallow-water marginal basins (Lugli et al., 2010; Manzi et al., 2013); this unit is  
81 termed the *Primary Lower Gypsum* (PLG) (Fig. 1). These deposits typically consist of rhythmically-  
82 deposited gypsum interbedded with shales. Although Vai and Ricchi Lucchi (1977) originally interpreted  
83 these as sabkha deposits with subaerial exposure near the top, the recent work of Lugli et al. (2010)  
84 concludes that deposition was entirely subaqueous. Below ~ 200 m water depth, in intermediate and deep  
85 water basins, lateral facies changes to dolomites and/or barren organic-rich shales have been observed (e.g.,  
86 Manzi et al., 2007; Lugli et al., 2010; Dela Pierre et al., 2011, 2012). A lack of evaporite deposition in deeper  
87 water is possibly due to under-saturation with respect to sulphate in the water column at this time (De Lange  
88 and Krijgsman, 2010). The top of the PLG deposits is normally an unconformity termed the ‘Messinian  
89 Erosional Surface’ (MES), the result of regression during the next stage of the MSC. In some marginal  
90 basins the MES can cut PLG and older deposits and the correlative conformity of the MES can be traced into  
91 deep basins at the base of the RLG unit (Roveri et al., 2008a, b)

92 Stage 2 (5.6 – 5.55 Ma) represents the acme of the MSC when widespread subaerial erosion took  
93 place forming the MES possibly as a result of the high-amplitude base-level fall of the Mediterranean  
94 (CIESM, 2008). In shallow marginal basins, subaerial exposure led to erosion and a hiatus of variable  
95 amplitude. Eroded material was transported offshore and sediment deposition at this time was dominated by  
96 clastic gypsum deposits that form the Resedimented Lower Gypsum unit (RLG; Roveri et al., 2008a, b). A  
97 number of factors (i.e., pressure release and fluid migration - Lazar et al., 2012; crustal loading - Govers et  
98 al., 2009; tectonic instability – Robertson et al., 1995) have been proposed as the cause of slope instability  
99 and gravity failure resulting in mass-wasting deposits and gravity flows of the RLG deposits.

100 Stage 3 (5.55 – 5.33 Ma) is thought to have been a period of complex water exchange between the  
101 Atlantic Ocean and Paratethys (Orszag-Sperber, 2006; Rouchy and Caruso, 2006; Roveri et al., 2008b),  
102 which resulted in selenite and cumulate gypsum deposition in shallow marginal basins in the central and

103 eastern Mediterranean (i.e., Sicily and Cyprus). The Upper Gypsum deposits are superficially similar to the  
104 PLG deposits, yet facies analysis indicates formation in very shallow water (Manzi et al., 2007, 2009; Lugli  
105 et al., 2008; Roveri et al., 2014a). Furthermore, distinctively low Sr isotope values (compared to oceanic  
106 values) have been measured from both the gypsum and fossils of these sections, indicating substantial  
107 freshwater input (Flecker and Ellam, 2006; Roveri et al., 2014a, b). By contrast, in northern and western  
108 marginal basins evaporite-free clastics formed in shallow to deep-water environments with characteristic  
109 brackish to fresh water fauna often referred to as the ‘Lago Mare’ biofacies (Ruggieri, 1967; Bassetti et al.,  
110 2004; Orszag-Sperber, 2006; Grossi et al., 2008, Roveri et al., 2008b; Popescu et al., 2015).

111 The end of the MSC at 5.33 Ma is marked by the return to fully marine conditions and defines the  
112 base of the Pliocene epoch (Van Couvering et al., 2000). The boundary is almost universally recognised as a  
113 near synchronous flooding surface (Iaccarino et al., 1999a; Gennari et al., 2008) as a result of the  
114 catastrophic flood of Atlantic waters into the Mediterranean basin (e.g., Hsu et al., 1973; Blanc, 2002; Meijer  
115 and Krijgsman, 2005; Garcia-Castellanos et al., 2009; Periáñez and Abril, 2015). The re-establishment of this  
116 Atlantic connection is likely the result of retrogressive erosion of the Gibraltar Strait rather than tectonically  
117 driven subsidence (Loget and Van Den Driessche, 2006; Estrada et al., 2011). In many marginal basins, the  
118 Zanclean sediments have been recorded as being relatively deep marine facies overlying Messinian  
119 evaporites or Lago Mare facies sediments. Gilbert-type fan deltas, possibly of Zanclean-age, are also  
120 commonly identified infilling Messinian fluvial canyons cut into underlying deposits (Bache et al., 2012).  
121 However, there are outstanding questions on the nature and progression of the ‘Lago Mare’ event and the  
122 Zanclean reflooding, especially regarding the difference between deep and peripheral basins that require  
123 further investigation (Popescu et al., 2015).

124

125

### 126 **3. Geological setting and stratigraphy**

127

128 The Hatay Graben (also known as the Hatay Basin, the Antakya-Samandag Basin, or the Antakya  
129 Fault Zone) in southern Turkey is a transtensional half-graben that developed during the late Miocene to  
130 Pliocene as a result of the westward extrusion of Anatolia (Boulton et al., 2006; Boulton and Robertson  
131 2008; Boulton and Whittaker, 2009) and the cessation of subduction along the Arabian/Eurasian margin

132 (Robertson et al., 2001). The present day half-graben developed due to the reactivation of basement  
133 structures upon a peripheral foreland basin sequence of Miocene age, consisting of lower Miocene fluvial  
134 conglomerates (Balyatağı Formation), middle Miocene shelf limestones (Sofular Formation) and upper  
135 Miocene (Tortonian) marls and sandy marl (Nurzeytin Formation) (Boulton and Robertson, 2007; Boulton et  
136 al., 2007) (Figs. 3, 4). Several Messinian evaporite locations have been identified in the area (Boulton and  
137 Robertson, 2007; Boulton et al., 2007; Tekin et al., 2010) forming the Vakıflı Formation (Fig. 3). The  
138 Vakıflı Fm. is exposed within the graben margins and also within a perched basin between two normal faults  
139 on the southern basin margin (near Sebenoba; Fig. 3). This uplifted location indicates that the main southern  
140 graben bounding faults did not yet have significant relief prior to and during the deposition of this unit  
141 (Boulton et al., 2006). Therefore, it is likely that during the late Miocene the basin occupied a wider  
142 geographic extent than at the present day and may have been connected to the Iskenderun basin to the north  
143 (Boulton et al., 2006). Overlying the Vakıflı evaporites is a sequence of Pliocene sandstone and marls  
144 (Samandağı Formation) that are exposed only within the margins of the present active graben, suggesting  
145 that the boundary faults had developed sufficiently to influence sediment deposition by early Pliocene time  
146 (Boulton et al., 2006; Boulton and Robertson, 2008). The base of the Samandağı Formation is variably  
147 conformable to unconformable with the underlying Nurzeytin or Vakıflı Formations.

148 The sediments preserved in the Hatay Graben; therefore, allow the investigation into the progression  
149 of palaeoenvironments across the Miocene-Pliocene boundary in the easternmost Mediterranean and provide  
150 a key test to proposals for a universal stratigraphic model of the basin (CIESM, 2008).

151

152

#### 153 **4. Methodology**

154 For micropalaeontological analysis, twenty-six marl samples from the Ortatepe Section (location 2, Fig. 3)  
155 and four marl samples from location 4 (Fig. 3) were disaggregated using the 'solvent method' of Brasier  
156 (1980). The samples were sieved through a 63 µm sieve, dried and benthic foraminifera were picked from  
157 the >63 µm size-fraction. In order to determine the minimum number of specimens to be picked per sample,  
158 rarefaction curves (number of species versus number of specimens) were calculated for a number of samples.  
159 Species-specimen curves become parallel to the species axis at ~150 specimens, so this was considered to be

160 the minimum number of specimens to be picked per sample. In most cases, >200 specimens were picked per  
161 sample, although in one case (sample OR7-33) this was not achieved (total 141 specimens) so this sample  
162 was excluded from the analysis. Benthic foraminiferal species diversity was recorded in terms of the Fisher's  
163 alpha index (Fisher et al., 1943). Alpha index values were read off the base graph in Williams (1964, p. 311)  
164 by plotting the number of species against the number of individuals in a sample. The percentage of planktic  
165 foraminifera relative to the total foraminiferal assemblage (planktic + benthic) in the >63 µm size-fraction  
166 was recorded for each sample. Benthic foraminifera were identified according to Cimerman and Langer  
167 (1991) and Milker and Schmiedl (2012).

168

## 169 **5. Observations**

170 In this section, six representative sections are described from west to east illustrating the stratigraphy of the  
171 late Miocene to Pliocene sediments of the Hatay Graben. Fourteen sedimentary facies (excluding evaporite  
172 facies – see Tekin et al., 2010 for a full description of these) have been identified in exposures attributed to  
173 Miocene-Pliocene age, detailed sedimentary descriptions and interpretation of each facies is given in Table  
174 1. Facies abbreviations follow convention with G for conglomerates, S for sandstones, M for siltstones and  
175 mudstones.

176

177

### 178 **5.1 Mağaracik (Location 1, Fig 3)**

179 Approximately 10 m of cross-bedded, poorly lithified, sandstone is exposed in a strike parallel face in a  
180 quarry to the west of Samandağ (Fig. 5; UTM Zone 35 S; 0765400/4000510). These Samandağı Fm.,  
181 sandstones unconformably overlie the upper surface of the Sofular Formation (middle Miocene limestone),  
182 which is eroded and bored at this location dipping down under the sandstone to the east. At the base of the  
183 outcrop, the litharenite is medium- to coarse-grained to pebbly (Facies Scr; Table 1) sandstone with bi-  
184 directional cross-beds. The outcrop as a whole coarsens upwards with coarse pebbly, cross-bedded  
185 sandstone and lenses of conglomerate (Facies Gm; Table 1) present at the top of the section. There is some  
186 evidence of bioturbation, as rarely vertical burrows are present, and small fragments of bivalves (e.g.,  
187 *Ostrea*, *Cardium*) can be observed.



188 5.1.1 Interpretation

189 The presence of the small bivalve fragments (*Ostrea*, *Cardium*) indicates a marine origin for these sediments.  
190 Coarsening upwards sequences are classic deltaic indicators (Reading and Collinson, 1996), and bi-  
191 directional currents are also very common in such environments, typically the result of tidal influences in a  
192 shoreface depositional setting. The lower cross-bedded sandstones may belong to the distributary mouth-bar  
193 facies, while the conglomerate lenses could be channel-fill deposits as the delta becomes more fluviially  
194 influenced as water depth shallows. Therefore, we interpret this sequence as gravelly-sandy foresets of a  
195 Gilbert-type fan delta (Reading and Collinson, 1996).

196 Currently the age of these deposits is interpreted as Pliocene, in the absence of other data owing to  
197 their stratigraphic position. The basal unconformity is interpreted as the Messinian Erosion Surface that  
198 formed during the acme of the MSC as the underlying middle Miocene limestones are highly eroded at this  
199 horizon, presumably by a high-amplitude base-level fall during the late Miocene. Therefore, the Samandağ  
200 sandstones may have been deposited subsequently possibly during the Zanclean transgression but equally  
201 these sediments could date to later in the Plio-Quaternary or to the latest Messinian.

202

203 5.2 Ortatepe Section (location 2; Fig. 3)

204 Incised Quaternary river terraces near the town of Samandağ expose sections of the Nurzeytin and  
205 Samandağ Formations. On the eastern side of Ortatepe (UTM Zone 36 S; 769653 E; 3998196 N),  
206 excavation to form a field has revealed a exposure ~ 100 m in length and ~ 20 m high, previously described  
207 by Boulton et al. (2007). The lower part of the section exposed to the south, is composed of fossiliferous,  
208 interbedded, thin (< 20 cm) sand beds and interbedded marl of the Nurzeytin Fm., (Facies M and MS; Table  
209 1) gently dipping to the southeast. The fossil content is variable, with macrofossils such as marine  
210 gastropods, including specimens from the Cypraeidae, Ellobiidae and Conidae families, and bivalves  
211 including *Ostrea* sp. and *Corbula* sp., present, while microfossils, including ostracods, such as *Cyprideis*  
212 spp., *Aurila* spp., and *Loxococoncha* spp., and benthic and planktic foraminifera, including *Globigerinoides*  
213 spp., are present near the top of the section (Boulton et al., 2007). Further micropalaeontological analysis  
214 (benthic foraminifera) was undertaken on this section as detailed below.

215 Above the interbedded marl and fine-grained sandstones is an abrupt transision along a gently  
216 dipping planar horizon into medium-grained, massive micaceous sandstone (Facies Sm; Table 1) of the

217 Samandağı Fm., forming moderately dipping (20°) beds that downlap onto the top of the underlying marl  
218 (Fig. 6). Above this interval, the lithology is similar but the bedding is disturbed and contorted (Facies Smc;  
219 Table 1). Rip-up clasts of parallel laminated mud are present along with horizons of shelly conglomerate  
220 containing well-rounded sandstone clasts, bivalves and marine gastropods (e.g., *Neverita josephina*,  
221 *Ringicula* sp., *Demoulia* sp., *Calliostoma* sp., *Turris* sp.). Small (~ 5 m) laterally discontinuous beds (Facies  
222 Sch; Table 1) and further contorted horizons of facies Smc are present nearby (Facies Smc; Table 1).

### 223 5.2.1 *Micropalaeontological results*

224 The preservation of benthic foraminifera is generally moderate to good in the majority of samples.  
225 Some samples contain broken specimens and some contain specimens with iron staining (OR7-20, 24, 26  
226 and 33). Sample OR7-33, which was excluded from the analysis, contained few individuals, which are  
227 poorly preserved and large in size.

228 The top ten ranked species in all samples overall account for 72.7% of the 107 identified species.  
229 The two most abundant species, *Rosalina globularis* and *Asterigerinata mamilla*, occur in every sample and  
230 together account for a mean of 33.5% of all species throughout the studied interval. Their relative  
231 abundances vary throughout the interval and overall show an increase up through the section (Fig. 7). The  
232 percentage of 'high-productivity/low-oxygen species' (sum of % *Bolivina* spp., *Brizalina* spp., *Bulimina*  
233 spp., *Melonis affinis* and *Uvigerina peregrina*) (e.g., Lutze and Coulbourn, 1984; Sen Gupta and Machain-  
234 Castillo, 1993) shows an overall decrease from mean values of 26% to 14% through the section (Fig. 7). The  
235 'high-productivity/low-oxygen' species group is dominated by *Bolivina* spp. and *Brizalina* spp.; whilst  
236 *Bulimina* spp. (0.4% of total), *M. affinis* (0.02%) and *U. peregrina* (0.05%) have very low abundances  
237 throughout the studied interval and only occur sporadically. The percentage of miliolids (*Adenosina* spp.,  
238 *Cornuspira involvens*, *Cycloforina* spp., *Miliolinella* spp., *Pyrgo* spp., *Quinqueloculina* spp., *Spiroloculina*  
239 spp.) fluctuates throughout the interval with lower abundances (<2%) occurring in the middle part of the  
240 section (OR7-18, 4.25 m to OR7-28, 6.75 m) (Fig. 7). The planktic foraminifera are dominated by small,  
241 juvenile specimens in the studied > 63 µm size fraction. Higher percentages of planktic foraminifera occur  
242 in the middle part of the section (mean 40%, OR7-16, 3.75 m to OR7-30, 7.25 m) compared with the interval  
243 before (mean 25%) and after (mean 25%) (Fig. 7). Diversity fluctuated over the studied interval, although  
244 there appears to be a slight temporal trend towards lower values (Fig. 7).

245

246 5.2.2 Interpretation

247 The benthic foraminiferal assemblages (dominated by *Rosalina*, *Asterigerinata*, *Haynesina*,  
248 *Elphidium*, *Ammonia*) indicate that the deposition of the marl succession occurred in an inner shelf  
249 environment (0-100 m water depth) (Murray, 1991, 2006). This is supported by the alpha index values ( $\alpha$  9-  
250 15), which fall within the range typical of inner shelf environments ( $\alpha$  3-19) (Murray, 1991). Barbieri and  
251 Ori (2000) found a similar benthic foraminiferal fauna dominated by ammoniids, elphidiids and epiphytes  
252 from the Neogene of northwest Morocco that they interpreted as indicative of an inner neritic (0-30 m)  
253 environment. The percentage of planktic foraminifera, however, could suggest a middle shelf environment  
254 (Murray, 1976), and water depths of up to 200 m have been proposed by Boulton et al. (2007). However, the  
255 high proportion of juvenile planktic foraminifera supports shallower water depths of middle to inner shelf  
256 environments (Murray, 1976). The apparent contradiction in the palaeoenvironmental reconstruction could  
257 be a function of the size-fraction used in this study compared with other studies. Many studies calculate the  
258 percentage of planktic foraminifera (or P:B ratios) in the  $>125\ \mu\text{m}$  or  $>150\ \mu\text{m}$  size-fraction, but our study of  
259 the  $>63\ \mu\text{m}$  size-fraction would potentially overestimate the proportion of planktic foraminiferal specimens,  
260 particularly if smaller species and/or juveniles are abundant, compared with larger size-fractions. The  
261 increase in the percentage of planktic foraminifera in the middle part of the succession may indicate that the  
262 water depth increased at this time, and the concomitant decrease in the abundance of miliolids, which are  
263 generally more abundant in shallower water (Murray, 1991, 2006), generally supports this observation.

264 In the modern Mediterranean Sea, the two most abundant species, *R. globularis* and *A. mamilla*, are  
265 known to be epiphytic species that are temporarily attached and make up 10-45% of assemblages on  
266 microhabitats with a high sediment content (*Posidonia* rhizomes, algae) (Murray, 2006). It is known that the  
267 distribution of epiphytic foraminiferal assemblages is controlled by substrate, light, availability of plant  
268 substrates and food (Murray, 2006); therefore the observed changes in the abundance of these species are  
269 most likely associated with one or more of these factors. If seagrasses were present, and thus supporting the  
270 epiphytic benthic foraminifera, then the maximum water depths allowing photosynthesis would be 20 m  
271 (Zieman and Zieman, 1989). The increase in abundance of *R. globularis* and *A. mamilla* up through the

272 section is not likely to be associated with increased food fluxes because the percentage of species indicative  
273 of 'high-productivity/low-oxygen' conditions decreases.

274         When combined with the sedimentary data, the majority of the marl facies of the Nurzeytin Fm.  
275 represent background deposition from suspension settling within the basin; the basin floor was possibly  
276 colonised by seagrass (*Posidonia* sp.) supporting a benthic community in water depths of < 100 m and  
277 maybe < 20 m. The layered nature of the shelly material in the lower marls and thin sandstone beds are  
278 suggestive of reworking by high-energy events, possibly storms, turbidity or grain flows, and are  
279 characteristic of downslope transport within the basin and may represent a prodelta environment. Prodelta  
280 facies associations are typically dominated by low-gradient fine-grained deposits from suspension fall-out  
281 and low-density turbidite flows (i.e., Backert et al., 2010), representing the basin environment in front of  
282 deltas. The presence of planktic and benthic foraminifera, marine bivalves and gastropods indicates a marine  
283 setting for the delta; however, some but not all of the ostracods (Boulton et al., 2007) indicate brackish water  
284 conditions (i.e., *Cyprideis* sp). These were likely reworked from the nearshore zone downslope. Evidence  
285 for downslope reworking can also be inferred for some foraminifera due to the presence of abraded and/or  
286 fragmented tests.

287         The decimetre-scale beds of the Samandağı Fm., observed to down-lap onto the lower marl and  
288 sandstones, represent avalanche foresets of a delta that is prograding into relatively deep water with a high  
289 sediment supply from feeder systems (Reading and Collinson, 1996). The disturbed and contorted bedding  
290 observed above the foresets is the result of sediment slumping due to downslope instability as a result of  
291 either oversteepening of the slope close to the angle of repose by bedload deposition or tectonic activity  
292 within the basin. The channelised sands above may represent the lowest-most beds of the subaerial topset of  
293 the deltaic system. This facies association is characteristic of a Gilbert-type delta and is remarkably similar  
294 to the Gilbert-type deltas described elsewhere in the Mediterranean during the Zanclean (i.e., Melinte-  
295 Dobrinescu et al., 2009).

296         Boulton et al. (2007) identified the Messinian-Zanclean boundary within the marls due to first  
297 occurrence of *Globorotalia margaritae* near the top of the section; however, we have found no further age-  
298 diagnostic fauna in this study to corroborate this interpretation. The biota of the marl and sandstone do  
299 indicate fully marine conditions, this is supported by the presence upper Miocene ostracods *Cyprideis*

300 *anatolica* and *C. torosa* and the absence of the post-MSC ostracod *C. agrigentina* (Boulton et al., 2007)  
301 used to indicate Lago Mare facies (Faranda et al., 2013). The Ortatepe location is also stratigraphically  
302 higher than nearby gypsum outcrops, which all suggests that these marls represent latest Messinian to earliest  
303 Pliocene marine conditions in the Hatay Graben. The overlying Gilbert-type delta was therefore deposited  
304 subsequently, perhaps during the Zanclean, although we note that the presence of specific facies is not age-  
305 diagnostic *per se*.

306

### 307 **5.3 Mizrakli (location 3; Fig. 3)**

308 To the east of the villages of Nurzeytin and Mizrakli there is a well-exposed sedimentary succession (as  
309 measured from UTM Zone 35 S; 0769443/4003491 to 0230000/4002521) (Boulton et al., 2007). Boulton et  
310 al. (2007) report Sr isotope measurements in the range of 0.708878 – 0.708925, confirming a Tortonian age  
311 of 8.7 – 9 Ma for the lower to intermediate part of the section. The presence of gypsum deposits at the top of  
312 the succession indicates a Messinian age for the end of the section. Although the base of the Nurzeytin  
313 Formation is not exposed, the lowermost sediments observed are interbedded grey marl and grey lime  
314 mudstone (Facies MS; Table 1). Beds are 30-130 cm thick and fine upwards. The beds are bioturbated and  
315 horizontal (to bedding) burrows were observed; fragments of body fossils are also present and include  
316 bivalve, gastropod and plant fragments as well as planktic foraminifera. These mudstones are replaced  
317 upwards after 10-15 m by a dominantly marl lithology (Facies M; Table 1) with only occasional sandstone  
318 interbeds (Facies Ss; Table 1), which occur singly or in packages. Isolated interbeds, often calcarenites <1 m  
319 thick, exhibit sharp bases and tops but lack sedimentary structures. Interbeds occurring in packages tend also  
320 to be calcarenites, <50 cm thick, with sharp bases, that then fine upwards and grade into a marl bed above.  
321 Sedimentary structures such as parallel laminations, cross-laminations, ripple marks, flute casts and rip-up  
322 clasts are present. Additionally slumped horizons are present (Facies Smc; Table 1). The top of the logged  
323 sequence is capped by ~ 10 m of gypsum following a poor exposed interval of marl. The lower part of the  
324 gypsum sequence is formed of 5 m of *in situ* bedded selenite, overlain by a gypsrudite formed of large  
325 angular blocks (>2 m) laminated alabastrine and selenite gypsum supported in a matrix of gypsiferous sandy  
326 marl.

327

328 **5.3.1 Interpretation**

329 The Tortonian marls represent settling from suspension within a basinal setting. The water depth is  
330 difficult to calculate but probably initially exceeded 100 m in depth (Boulton et al., 2006). The interbeds of  
331 calcarenite observed likely represent low density turbidite deposits based upon the range of sedimentary  
332 structures present and the overall fining-upward nature of the beds. The presence of turbidity currents along  
333 with slumped beds is indicative of down-slope transport of sediments that would have reworked material  
334 from the near-shore environment into deeper water. Unfortunately, the lack of palaeocurrent indicators does  
335 not allow discrimination between transport offshore into the Levant Basin or into the local basinal  
336 depocentre.

337 The marls pass apparently conformably upwards to Messinian gypsum deposits, although the  
338 boundary is not exposed. The gypsum rudite beds, composed of broken selenite crystals, are interpreted as  
339 the result of mass flows in a slope setting. Tekin et al. (2010) suggested that tectonic activity at the basin  
340 margin could have initiated these flows but did not rule out climatic or water level fluctuations leading to  
341 slope instability. The upper chaotic unit is interpreted by Tekin et al. (2010) as the result of active tectonics,  
342 by comparison to similar facies reported by Robertson et al. (1995) from southern Cyprus and by Manzi et  
343 al. (2011) in Sicily.

344

345 **5.4 Main Road Quarry Section (location 4; Fig. 3)**

346 On the main Antakya-Samandağ road, a small quarry (UTM Zone 35 S; 0237433/4004350) reveals the  
347 contact between the Nurzeytin and Samandağ Formations (Fig. 8). The base of the quarry is composed of  
348 blue-grey marls (Facies M; Table 1) and fining upwards beds of very fine-grained sandstone 20 – 50 cm  
349 thick (Facies MS; Table 1). Fragmented woody material is common within these sandstone beds but  
350 sedimentary structures are lacking. The boundary between the Nurzeytin Fm., marls and the overlying  
351 orange-weathering sandstones of the Samandağ Formation is erosive with a slight angular discordance. The  
352 coarse-grained sandstones are up to 30 cm thick, dip towards the southwest, and are laterally discontinuous.

353 Micropalaeontological analyses of the benthic foraminifera on four samples (MBP 1-4) from the  
354 underlying Nurzeytin Formation show generally poor preservation with high number of undetermined and  
355 reworked (as determined due to abrasion and/or fragmentation) specimens (about 22%). The assemblages are  
356 dominated by *Bolivina* spp. and *Brizalina* spp. (together about 30%), where *B. spathulata* (13%) and *B.*

357 *dilatata* (5%) are the most abundant species. Other relatively common species are *Bulimina* spp. (5.2%,  
358 dominated by *B. aculeata* and *B. elongata*), together with *Cibicides lobatulus* (5%), *Cibicidoides* spp.  
359 (4.7%), *Cassidulina obtusa* (3.7%), *Eponides* spp. (3.2%), *Rosalina* spp. (3.1%), *Gyroidinoides* spp. (2.2%),  
360 *Valvulineria* spp. (2.2%), *Anomalinoidea* spp. (2.1%), and *Globocassidulina subglobosa* (2.0%). Others  
361 species with abundances of between 1 and 2% are *Fursenkoina* spp., *Ammonia* spp., *Epistominella vitrea*,  
362 and *Melonis affinis*, whereas miliolids, *Elphidium* spp., *A. mamilla* and *Uvigerina* spp. are less than 1%.  
363 Planktic foraminifera (including *Turborotalita multiloba* and *Neogloboquadrina acostaensis*) are quite  
364 abundant, comprising about 50% of the total foraminifera.

#### 365 5.4.1 Interpretation

366 The benthic foraminiferal assemblages (dominated by *Bolivina* and *Brizalina*, together with  
367 *Bulimina*, *Cibicides*, *Cibicidoides* and *Cassidulina*) indicate that the deposition of the lower part of the  
368 succession occurred in an outer shelf-upper slope environment (100-200 m water depth) (Murray, 1991,  
369 2006). This is supported by a high abundance of planktic foraminifera (about 50%), which is typical for this  
370 environment. The high percentage of 'high-productivity/low-oxygen species' (especially *Bolivina* spp.,  
371 *Brizalina* spp., and *Bulimina* spp.), clearly indicate a low oxygen environment with high flux of organic  
372 matter (e.g., Lutze and Coulbourn, 1984; Sen Gupta and Machain-Castillo, 1993). The planktic foraminifera  
373 *Turborotalita multiloba* is probably an ecophenotypic of *Turborotalita quinqueloba*, and according to  
374 Krijgsman et al. (1999), Sierro et al. (2001) and Lourens et al. (2004) its first influx occurs at 6.42 Ma,  
375 predating the *Neogloboquadrina acostaensis* sinistral to dextral coiling change at 6.35 Ma. The presence of  
376 *N. acostaensis* dextral in the samples confirms that the studied interval belong to the MMi 13c *T. multiloba*  
377 Interval Zone spanning from 6.35 Ma to 5.96 Ma (Lourens et al., 2004), which is the last Mediterranean  
378 Biozone in the Messinian before the non-distinctive zone corresponding to the MSC.

379 The overlying sandstone beds of the, presumably Zanclean, Samandağ Formation cut  
380 stratigraphically downwards to the southwest (seawards) and are lacking in fossil material. The similarity of  
381 these sandstones to the upper sands present in the other described localities implies that these could be the  
382 topset beds of a fan-delta system.

383

384 **5.5 Sutası Section (location 5; Fig. 3)**

385 A well-exposed section of the Samandağ Formation dating to the latest Miocene to earliest Pliocene  
386 (Boulton et al., 2007) is exposed near Sutası (Fig. 3, location 2) along a road cutting ~ 650 m long and ~ 10  
387 m high. The base of the section is dominated by fossiliferous, orange-coloured, lithic calcarenite (Facies Ss;  
388 Table 1), with bedding thickness 0.25-3.00 m thick (Fig. 9a). Shell fragments are common and are mostly  
389 composed of bivalve and gastropod fragments with occasional articulated bivalves forming shell and pebble  
390 lags. Preliminary analyses of the benthic foraminifera from the Sutası section show that the assemblages are  
391 dominated by *Ammonia* spp., together with *Nonionellina* spp., *Elphidium* spp., *Cibicides refulgens*,  
392 *Asterigerinata mamilla*, *Rosalina globularis*, and others. Planktic foraminifera are also present, comprising  
393 <25% of the total foraminifera. Ostracods are also represented by *Cyprideis torosa*, *C. anatolica*, *Aurila*  
394 *convexa*, *A. speyeri*, *Ruggieria tetraptera* and other long lasting species (Boulton et al., 2007). Fragmentary  
395 plant material is also present. Interbedded with these sands are thin mud and limestone layers < 25 cm thick.

396 There is a change in the character of the sediments at ~ 30 m up the section (Fig. 9a); the lithic  
397 calcarenite becomes coarser-grained with common trough and planar cross-bedding (Facies Scr; Table 1),  
398 yet the thickness of the bedding decreases with many beds <10 cm thick. The overlying beds exhibit planar  
399 cross-bedding, parallel-laminations and ripple cross-lamination. These are interbedded with two lenticular  
400 polymict clast-supported conglomerates up to 75 cm thick (Facies Gc; Table 1) with coarse sandstone and a  
401 1 m thick mottled pink mudstone above (Table 1). Bioturbation is generally absent in this interval and, as a  
402 result, sedimentary structures are well preserved. Macrofossil and microfossil material is very rare and  
403 fragmented when present.

404 Above this interval of diverse structures, the uppermost part of the sequence is composed of > 15 m  
405 of medium-grained sandstone with little or no fossiliferous material and mostly lacking in sedimentary  
406 structures, although low-angle cross-bedding can be observed in some horizons (Facies Sb; Table 1). This  
407 massive sandstone characterises the majority of the Pliocene succession in many outcrops and is generally  
408 variably cemented with nodules (similar to doggers) present throughout.

409



### 410 5.5.1 *Interpretation*

411 The lower part of the section is composed of medium-grained sandstones with sharp, often erosional,  
412 bases that fine upwards, with parallel-laminations and planar cross-lamination in some horizons. Pebble and  
413 fossils lags are also commonly present formed as a result of low-relief scours and currents. The bioturbation  
414 suggests that between phases of rapid deposition sedimentation was relatively slow allowing colonisation of  
415 the substrate. This facies association is typical of coarse-grained lower shoreface environments (Reading  
416 and Collinson, 1996; Clifton, 2006).

417 The lower shoreface passes vertically upwards into the upper shoreface facies association with  
418 trough cross-bedded sandstones, the result of oscillatory motion related to the primary onshore waves and  
419 secondary back-flow or the result of tidal influences (Dashtgard et al., 2012). The observed increase in  
420 grain-size is also common from the lower to the upper shore face (Reading and Collinson, 1996).

421 The progradational nature of this sequence suggests that the stratigraphically higher sediments would  
422 be representative of the foreshore and beach. This interpretation is supported by the presence of horizontal  
423 laminations, developed by wave swash and low-angle tabular cross-bedding. The conglomerate lenses could  
424 represent the plunge step marking the transition from the top of the shoreface to the base of the foreshore  
425 (i.e., Sanders, 2000) but the association of the conglomerate with the pink mudstone suggests that these are  
426 more likely to represent small channel fills with an associated palaeosol (as indicated by the mottled colour)  
427 indicating a period of subaerial emergence with fluvial erosion and sedimentation. The lack of sedimentary  
428 structures resulting from the intense bioturbation in the overlying lithic calcarenite makes the environment of  
429 deposition difficult to infer; however, given the overall shallowing upwards sequence these may represent  
430 deltaic or fluvial facies. Therefore, the section as a whole would represent a prograding shoreline.

431

### 432 5.6 *Location 6 (Fig. 3)*

433

434 Location 6 is a mixed clastic and carbonate sequence at the base of the Samandağ Formation, the top  
435 of this section has been dated using strontium isotopic ratios from benthic foraminifera to  $5.35 \pm 0.1$  Ma (Sr  
436 measurement = 0.709023; Age range = 5.2 – 5.41 Ma; Boulton et al., 2007), placing this section within error  
437 of the Miocene-Pliocene boundary. However, caution must be applied with strontium ages from the  
438 Messinian as a wide range of values occur due to a lack of connection with the global ocean, but by the early

439 Zanclean the return to fully marine conditions results in more robust dates (Flecker and Ellam, 2006). Here,  
440 marine conditions are indicated by the presence of a mixed benthic foraminiferal assemblage used to derive  
441 the strontium measurement but the marginal setting could still influence the Sr values and thus the derived  
442 age.

443 The basal part of the section is composed of interbedded calcarenite, chalk and marl (Fig. 9b)  
444 forming a conformable transitional boundary with the underlying Nurzeytin Formation (Facies M,C, Ss;  
445 Table 1). The sandstone is medium-grained and unlithified. Bedding thickness is 0.3-3.0 m. Sedimentary  
446 structures are rare, but parallel laminations and rip-up clasts are present, especially near the base of sandstone  
447 beds. The chalk horizons are very thin (5-15 cm). The marl is burrowed and forms the lowermost bed of the  
448 section.

449 The upper part of the section consists of interbedded marl, sandstone and conglomerate. The  
450 conglomerates are irregular with erosive bases and are laterally discontinuous. The conglomerates are clast  
451 supported and clasts are sub-angular to sub-rounded. Above the conglomerates there are fine-grained  
452 micaceous lithic greywacke beds with parallel laminations. The bases of these beds are sharp and  
453 occasionally erosional; the beds often fine upwards and are generally laterally discontinuous on an outcrop  
454 scale. These are capped by marl beds, containing planktic foraminifera (Boulton et al., 2007), completing a  
455 upwards fining unit.

456

#### 457 5.6.1 Interpretation

458 These sandstone beds are interpreted as redeposited material. In the lower part of the section, these may be  
459 grain-flow and turbidite deposits (Stow et al., 1996), whereas in the upper part of the section the sands may  
460 represent channel-fill deposits with basal conglomerate lags. This suggests an increase in energy upwards  
461 possibly due to shallowing of the water column. This is in agreement with a decrease in marl up the section  
462 that would represent background basin sedimentation (Stow et al., 1996). The Sr isotope value (Boulton et  
463 al., 2007) derived from marls near the top of the exposure indicate marine deposition in the basin after the  
464 end of the MSC. The mean age places the section just prior to the Messinian-Zanclean boundary, but the  
465 error in the measurement does not rule out deposition in the earliest Pliocene.

466

## 467 **5.7 Messinian Gypsums**

468 In addition, to the selenite and gypsum breccia observed capping the top of the Mizrakli sequence  
469 (section 5.3), gypsum outcrops at a number of other localities in the Hatay Graben (Fig. 3) mainly along  
470 strike between the villages of Mizrakli and Vakıflı, and can reach 30 – 40 m in thickness. Typically the  
471 sequence consists of a lower alabastrine gypsum with laminations and thin interbedded marl horizons. Often  
472 the alabastrine gypsum can be observed to be interbedded with *in situ* selenite. These alabastrine gypsums  
473 are normally overlain by gypsum breccias and blocks of gypsum in a gypsiferous marl matrix (Fig. 10). On  
474 the southern margin of the graben near Sebenoba (Fig. 3) only the gypsum breccias were observed, consisting  
475 of clast-supported blades of selenite with minor gypsiferous marl matrix. Tekin et al. (2010) undertook  
476 detailed facies analyses of the evaporites of the Hatay Graben. Their analysis is consistent with our  
477 observations and shows that the gypsum deposits in the Hatay Graben can be divided into two sequences; a  
478 lower interbedded unit and an upper chaotic unit. The lower sequence is formed of interbedded laminated  
479 gypsum, selenite and bedded clastic gypsum facies (Tekin et al. 2010). The laminated gypsum facies is  
480 composed of eroded and resedimented gypsum crystals with slumps, normal and reverse grading present.  
481 Tekin et al. (2010) interpret these laminate deposits as having been deposited by turbidity or gravity flows in  
482 the central part of a density stratified basin (Warren et al., 2006). The bedded gypsum facies are composed  
483 of poorly sorted, massive gypsarenites and gypsrudies with broken selenite crystals up to 4 cm in length, and  
484 are also interpreted as having been deposited by mass flows (Tekin et al., 2010). By contrast, the selenite  
485 facies is interpreted to have grown *in situ* water depths of > 10 m (Tekin et al., 2010). The upper chaotic  
486 unit, as observed at Mizrakli (Fig. 10), is composed of large blocks of selenitic gypsum in a gypsiferous marl  
487 matrix with evidence of slumping indicative of down slope transport, which Tekin et al. (2010) attribute to  
488 intense tectonism during deposition.

489

490

## 491 **6. Discussion**

### 492 **6.1 Timing of deposition of the Vakıflı Formation**

493

494 A key issue when interpreting the sedimentary succession regards timing of gypsum deposition. Do these  
495 sediments represent facies of the Primary Lower Gypsum (PLG), the Resedimented Lower Gypsum (RLG)

496 or the Upper Gypsum (UG)? Lugli et al. (2010; p. 84) state that the PLG and RLG deposits of Sicily ‘are  
497 *never associated laterally or vertically*’, and therefore the gypsums must represent one or other situation and  
498 not both in this model, if it is correct.

499 Tekin et al. (2010) report two  $^{87}\text{Sr}/^{86}\text{Sr}$  values for the Hatay Graben gypsums:  $0.708954 \pm 4 \times 10^{-6}$  and  
500  $0.708946 \pm 4 \times 10^{-6}$ ; although the vertical position within the sections was not stated, it appears that both  
501 samples were from the lower interbedded gypsum deposits based upon facies descriptions. These values are  
502 entirely consistent with values for the PLG and RLG derived from elsewhere in the Mediterranean that span  
503 the range 0.708893 – 0.709024 (Lugli et al., 2010; Roveri et al., 2014b). These data are distinct from values  
504 derived from the later UG deposits (typically  $^{87}\text{Sr}/^{86}\text{Sr} = 0.708750\text{--}0.708800$ ; Roveri et al., 2014b). These  
505 data strongly suggest that the lower interbedded unit in the Hatay basin correlates to the PLG or RLG and  
506 therefore the overlying gypsum mega-blocks will also belong to the same unit. The sedimentary  
507 characteristics of UG deposits are also distinct from those of the Vakıflı Fm, and therefore can be ruled out.

508 Tekin et al. (2010) describe two main gypsiferous facies associations in the Hatay Graben. The lower  
509 facies association is interpreted as part of a ‘sulphate platform’; the upper facies association as an ‘evaporitic  
510 slope-platform’. However, the sedimentology of both of these facies associations indicates downslope  
511 transport of material, initially by grain flows and turbidity currents in the lower bedded units and then by  
512 debris flows in the upper unit forming the ‘mega-blocks’. This evidence points towards the reworking of the  
513 gypsum characteristic of the RLG facies. These facies are strikingly similar to those described on Cyprus as  
514 the lower and intermediate gypsum unit, recently reinterpreted by Manzi et al. (2015) as belonging to the  
515 RLG deposits.

516 Typically, this observation would place the Hatay Graben into the ‘marginal’ basin class of deeper  
517 water basins where RLG facies have been observed (i.e., Sicily: Roveri et al., 2008a; Manzi et al., 2011).  
518 However, these observations are at odds with the presence of an unconformity (i.e., location 1) and the  
519 microfossil data indicating water depths of < 200 m prior to and in the early Messinian. These features are  
520 characteristic of shallow water ‘peripheral’ basins where PLG typically would have accumulated.

521 This contradiction may be resolved by considering the tectonic controls on basin formation. Boulton  
522 et al. (2006) demonstrated that high-angle oblique normal faulting initiated during the latest Miocene to  
523 Pliocene. As a result, footwall uplift and hangingwall subsidence would have (relatively) rapidly produced

524 areas of varying water depth and new depocentres during the Messinian. Therefore, it is possible that in a  
525 relatively short time the basin could have deepened sufficiently, combined with seismicity, to rework  
526 shallow gypsum facies into basin depocentre to form RLG facies. While on the flanks of the graben PLG  
527 facies would have accumulated. In the Hatay basin, shallow water gypsiferous sediments are not preserved  
528 but they are farther to the north (Tekin et al., 2010). Therefore, on balance, the Vakıflı evaporites can be  
529 considered as RLG deposits but further research into field relationships and strontium isotopes is required to  
530 confirm this hypothesis.

531

## 532 **6.2 Comparison to other eastern Mediterranean marginal basins**

533 Although the Hatay Graben is located in the easternmost Mediterranean, a number of other basins  
534 nearby expose Messinian-aged strata that can increase the understanding of the regional palaeoenvironments  
535 of the MSC in the easternmost Mediterranean and aid in the interpretation of the Hatay Graben facies.

536 Almost due south of the Hatay Graben lies the Nahr El-Kabir half graben in present-day Syria where  
537 outcrops of Messinian evaporites up to 100 m thick have been documented (Hardenberg and Robertson,  
538 2007). Underlying Tortonian sediments are generally absent or very thin, suggesting limited accommodation  
539 space in this region prior to the onset of the MSC; this is somewhat different to the shallowing but significant  
540 water depth in the Hatay. Hardenberg and Robertson (2007) describe the Messinian gypsums as having a  
541 tripartite subdivision with a lower unit comprising mainly alabastrine-type gypsum with marl laminations, a  
542 middle selenitic division, and an upper matrix-supported conglomerate. These deposits are interpreted as  
543 deposition in local depocentres with the uppermost unit the result of tectonic instability (Hardenberg and  
544 Robertson, 2007). Indeed, to generate the required accommodation space to accumulate these thick  
545 evaporite deposits, tectonic subsidence needs to be invoked given regional base-level fall. Although,  
546 strontium data are lacking for this area, the stratigraphy is similar to that described for the Hatay Graben,  
547 indicating that the gypsums in the Nahir El-Kabi half-graben could belong to the RLG.

548 Similar successions have been described for the Messinian evaporites in a number of sub-basins on  
549 Cyprus – the Polemi and Pissori sub-basins in the west and the Maroni sub-basin in the south (e.g., Eaton,  
550 1987; Follows, 1992; Payne and Robertson, 1995; Robertson et al., 1995; Rouchy et al., 2001; Krijgsman et  
551 al., 2002; Manzi et al., 2015). In the western Polemi and Pissouri Basins, Tortonian marl successions reflect

552 the progressive shallowing from ~ 500 m at Tortonian/Messinian boundary to < 100 m water depth and  
553 marine isolation leading up to the onset of evaporite deposition during the MSC (Kouwenhoven et al., 2006).  
554 The gypsum deposits are divided into a lower and upper unit by an intervening breccia horizon (Robertson et  
555 al., 1995). The lower unit is predominantly composed of finely-laminated gypsum with evidence for  
556 turbidity currents, slumping and debris flows, indicative of sediment reworking down a slope into deeper  
557 water. The mega-rudite breccia is formed of metre-scale blocks of fine-grained gypsum in a gypsiferous  
558 marl matrix, which Robertson et al. (1995) interpret as large-scale tectonically induced slumping but  
559 Rouchy et al. (2001) interpret as the result of karstic dissolution. The overlying upper unit is composed of  
560 selenitic gypsum and marl, interpreted as having formed in relatively shallow water. The deposition of these  
561 Upper Gypsums is followed by typical Lago Mare facies sediments, which include palaeosols indicating  
562 subaerial exposure during this period (Rouchy et al., 2001). The overlying Zanclean transgressive sediments  
563 were deposited in a well-oxygenated deep marine setting (Robertson et al., 1995). Therefore, the Polemi and  
564 Pissouri Basins have been traditionally considered to have PLG and Upper Gypsum deposits, based upon the  
565 stratigraphic facies constraints. Krijgsman et al. (2002) dated the onset of evaporite formation in the Pissouri  
566 Basin at 5.96 Ma using magnetostratigraphy, apparently confirming the synchronous onset of evaporite  
567 formation across the Mediterranean. However, recent work by Manzi et al. (2015) concludes that the lower  
568 and intermediate units are both the RLG, due to the overall clastic and reworked nature of the facies and that  
569 the base of the evaporites dated by Krijgsman et al. (2002) is in fact the MES. In the Maroni sub-basin, the  
570 evaporites consist of two distinct units (Robertson et al., 1995) but there is no evidence for late Messinian  
571 sediments and the mega-rudite is directly overlain by Pliocene marine marls (Robertson et al., 1995).  
572 Therefore, the overall stratigraphy from these basins is very similar to the Hatay Graben, although the Hatay  
573 Graben lacks the Upper Gypsum deposits possibly as a result of its more landward position.

574 Interestingly, directly to the north of the Hatay Graben in the Iskenderun Basin, onshore exposures  
575 of gypsum described by Tekin et al. (2010) lack this ‘mega-rudite’ conglomeratic unit. Instead, the gypsum  
576 facies that overly upper Tortonian marls are dominated by laminated gypsums, gypsiferous marls and  
577 sandstones, which Tekin et al. (2010) interpret as typical of very shallow water accumulation in lagoons and  
578 sabkhas. There are minor selenite accumulations thought to represent slighter deeper water conditions, but  
579 overall the Iskenderun basin appears to have had shallower water depths during the MSC than the Hatay

580 Graben. This area could represent the source area for the RLG of the Vakifli Fm., as younger tectonics have  
581 dissected the region since deposition (Boulton et al., 2006). Overlying Pliocene deposits are not well  
582 described but a thin Lago Mare succession appears to transition upwards into fluvial and coastal  
583 environments (Tekin et al., 2010).

584 Similarly, the Messinian succession in the Adana Basin indicates shallow water or continental  
585 conditions. Darbaş and Nazik (2010) and Faranda et al. (2013) describe planktic foraminifera and ostracods  
586 from late Miocene sections in the Adana Basin demonstrating that in the early Messinian the area was  
587 characterised by shallow coastal environments such as marshes, lagoons and estuaries. Cosentino et al.  
588 (2010) recognised a succession of rhythmically bedded anhydrites and black shales that they correlate to the  
589 PLG, whereas the outcropping gypsum deposits consist of gypsarenite and gypsrudite containing large  
590 blocks of selenite pertaining to the RLG (Radeff et al., 2015). Interestingly, Consentino et al. (2010) also  
591 recognise two Messinian erosion surfaces in the Adana Basin; one correlating to the wider MES cutting the  
592 lower evaporites, and the other at the base of the overlying continental sequence.

593 Burton-Ferguson et al. (2005) thought that these continental sediments were Pliocene in age;  
594 however, Ilgar et al. (2012) have identified Gilbert-type deltas that are laterally equivalent to the gypsum  
595 deposits, and microfossil analysis by Cipollari et al. (2013) and Faranda et al. (2013) showed that these  
596 sediments were deposited in brackish water environments of the latest Messinian Lago Mare event.  
597 Cipollari et al. (2013) also showed that subsequent Zanclean reflooding resulted in the deposition of deep  
598 marine marls in water depths of 200 – 500 m.

### 600 **6.3 Late Tortonian to Zanclean Palaeoenvironments of the Hatay Graben**

601

602 It is now possible to synthesise field observations, palaeontological and strontium data with regional trends  
603 to develop a model for the late Miocene of the Hatay Graben, which can then be used to test models for the  
604 wider Mediterranean at this time.

#### 605 **6.3.1 Late Tortonian to early Messinian**

606 The late Tortonian and earliest Messinian in the Hatay Graben are represented by the Nurzeytin  
607 Formation, composed mainly of marl with interbeds of sandstones, from locations 3 and 4 (Figs. 3, 11).

608 These sediments are interpreted as basinal deposition from suspension settling with reworking of material  
609 downslope through the action of slumps, turbidity currents and rare debris flows (Boulton and Robertson,  
610 2007). Boulton et al. (2006) suggested maximum water depths of up to 700 m for this unit; however, our  
611 new foraminiferal analysis indicates that by the early Messinian water depths had shallowed to < 200 m in  
612 some places and the seabed may have been carpeted in seagrass. This shallowing is potentially due to  
613 regional tectonic uplift, sea level fall or to the initiation of local faulting (Boulton et al., 2006), but similar  
614 trends have also been recorded in Cypriot basins (Kouwenhoven et al., 2006) resulting from the increasing  
615 isolation of the basin. The pre-MSC section on the main road (section 4) is of limited extent so that any  
616 changes to planktic foraminifera assemblages prior to the onset of the MSC might not have been identified in  
617 this study. Furthermore, it is possible that these pre-MSC sediments have been truncated by an unconformity  
618 and younger sediments have been eroded, as indicated by the angular unconformity observed at section 4.  
619 The section investigated at Mizrakli (section 3) appears continuous through the Tortonian – Messinian  
620 boundary, suggesting that this area may have been protected from later erosion potentially due to a location  
621 more proximal to the basin depocentre (Fig. 11).

### 622 6.3.2 Stage 1 of the MSC

623 No gypsum from this period appears to have been preserved *in situ* in the Hatay Graben. Shallow  
624 water and sub-aerial gypsum facies have been described north of the Hatay Graben near Iskenderun (Tekin  
625 et al., 2010) that are typical of the PLG deposits. PLG and associated deposits have also been described  
626 from the Adana Basin, suggesting that PLG could have been deposited if there were suitable conditions at  
627 that time. Therefore, it is possible that shallow water deposits were present on the edges of the basin feeding  
628 the resedimented gypsum that is observed in the Hatay Graben at the present day, but these deposits have  
629 subsequently been eroded. The Plio-Quaternary faulting that has formed the present topographic graben  
630 (Boulton and Robertson, 2007) has also dissected the region and previously the Hatay basin may have been  
631 part of a wider depositional system that at present.

### 632 6.3.3 Stage 2 of the MSC

633 During Stage 2 of the MSC, it is hypothesised that widespread subaerial erosion took place forming  
634 the MES and rivers cut canyons as the fluvial systems adjusted to base level (CIESM, 2008). In the Hatay



635 Graben subaerial exposure led to the erosion of underlying strata (as observed at location 1) in marginal  
636 locations at the edge of the basin. Despite this, subsequent deposition of Pliocene sediments and tectonic  
637 tilting of the basin makes an evaluation of the lateral extent of the MES difficult due to a lack of exposure  
638 (Fig. 11).

639 In the basin depocentre, formed as a result of active faulting along the southern basin margin, gravity  
640 reworking of previously crystallised gypsum led to the formation of the resedimented lower gypsums (RLG).  
641 In the Hatay, these deposits consist of two distinct facies associations indicating that a change in the nature  
642 of the gravity reworking took place later in this period, resulting in the deposition of the ‘mega-clasts’ at the  
643 end of the RLG period (as indicated by Sr ratios; Tekin et al., 2010). Similar facies are recorded in many  
644 locations around the Mediterranean (Sicily, Cyprus, Turkey), which have commonly been attributed to  
645 tectonic forcing (i.e., Robertson et al., 1995; Tekin et al., 2010). However, the RLG facies in the Adana  
646 Basin have been dated to the early post-evaporitic stage of the MSC (5.55 – 5.45 Ma) owing to the presence  
647 of brackish Paratethyan ostracods (Faranda et al., 2013) in the fine-grained interbedded sediments suggesting  
648 that downslope transport of clastic gypsum material may have taken place at different times in different  
649 basins.

650

#### 651 6.3.4 Stage 3 of the MSC

652 Although Stage 3 gypsums have been recognised from other eastern Mediterranean basins, the  
653 available data suggest that these are lacking in the Hatay Graben owing to either, or probably a combination  
654 of: a) later erosion; b) subaerial exposure resulting in a hiatus, or c) water chemistry or other local conditions  
655 being uncondusive to gypsum formation at this time.

656 Furthermore, several of the sections studied here have stratigraphic constraints indicating that during  
657 the latest Miocene (sections 2, 5, 6) marine conditions may have been present within the Hatay Graben, prior  
658 to the Zanclean reflooding. This is in clear contrast to nearby basins on Cyprus and elsewhere in Turkey  
659 where UG and/or Lago Mare biofacies deposits have been identified (i.e., Rouchy et al., 2001; Faranda et al.,  
660 2013; Manzi et al., 2015; Radeff et al., 2015). Yet Popescu et al. (2009, 2015) and Carnevale et al. (2006)  
661 have recorded fossil evidence indicating marine conditions during this period from deep and peripheral  
662 basins in the western Mediterranean, supporting the idea that marine conditions returned to the

663 Mediterranean prior to the Pliocene (e.g., Butler et al., 1995; Riding et al., 1998; Bache et al., 2012).  
664 Although further stratigraphic and palaeoenvironmental constraints would be desirable, our available data  
665 tentatively support a pre-Pliocene return to marine conditions even in the easternmost Mediterranean.

666

### 667 6.3.5 *Zanclean*

668 The base of the Zanclean is normally recognised as the return to marine conditions across the  
669 Mediterranean, although as stated above this may not be strictly correct. Available stratigraphic constraints  
670 indicate that the earliest Zanclean deposits are composed of interbedded marls and sandstones characteristic  
671 of marine conditions and probably represent the deepest water facies in the basin depocentre. Elsewhere, a  
672 slightly irregular to planar surface truncates the earlier Miocene marls, forming the base of the sandstone-  
673 dominated Samandağ Formation. Coarse-grained sandstones exhibiting a range of facies typical of Gilbert-  
674 type deltas or coastal environments generally outcrop stratigraphically above presumably deposited later in  
675 the Zanclean (Fig. 11). This dramatic change in facies suggests that although subaerial conditions returned  
676 initially in the late Miocene/Pliocene, water depth had shallowed considerably in most of the basin compared  
677 to before the MSC. The presence of Gilbert-type fan deltas is indicative of narrow and steep-gradient  
678 shelves, possibly infilling the incision developed during stage 2 of the MSC.

679

680

## 681 7. Conclusions

682 The available data indicate that the pre-MSC succession of the Hatay Graben is very similar to sequences on  
683 Cyprus, and to some extent in Syria, where water depths were likely  $\geq 100$  m at the onset of the MSC. PLG  
684 facies are generally poorly exposed in the eastern Mediterranean and the Hatay Graben is no exception and the  
685 gypsum deposits in the Hatay Graben are interpreted as RLG deposits. These lower RLG are often observed  
686 to be overlain by a chaotic unit composed of large gypsum ‘mega-clasts’ observed throughout the eastern  
687 Mediterranean and overlying the MES. Robertson et al. (1995), Boulton et al. (2006), and Hardenberg and  
688 Robertson (2007) have all previously interpreted these deposits as debris flows potentially triggered by  
689 tectonic activity. Although similar central Mediterranean deposits have been classically thought of as having  
690 been caused by dissolution collapse, Manzi et al. (2011) has also recently reinterpreted the central

691 Mediterranean breccias as syn-tectonic deposits. This apparent synchronicity raises the question as to how  
692 these basins all experienced sediment instability at the same time. If this is the case then the Mediterranean  
693 apparently underwent widespread and intense tectonic activity ~ 5.5 Ma. Interestingly, this correlates with  
694 proposals that the Arabia-Eurasia collision underwent a period of reorganisation ~ 5 Ma (Allen et al., 2004).  
695 However, the Adana Basin also contains evidence for a younger Messinian unconformity, and the RLG  
696 deposits in this basin are younger (dating to the Lago Mare biofacies event) than those described elsewhere  
697 (Radeff et al., 2015), suggesting that the mega-breccias might span a longer timespan than hitherto  
698 recognised. These stratigraphic differences observed north of the Hatay may reflect the proximity of the  
699 Adana and Iskenderun basins to the collisional zone between the Arabian and Anatolian micro-plates (the  
700 Bitlis-Zagros Suture). Although continental collision was well advanced by the Messinian (e.g., Robertson  
701 et al., 2015), the Adana and Iskenderun basins north of the suture zone, would have experienced different  
702 uplift and subsidence trajectories than areas to the south (i.e., Hatay and Syria) and the west of the collisional  
703 front.

704         Following deposition of the RLG, the Hatay Graben apparently records evidence for marine  
705 conditions at this time in contrast to the other regional basins where Lago Mare facies have been recorded.  
706 Although we cannot rule out the presence of typical Lago Mare facies elsewhere in the basin, the apparent  
707 presence of marine fauna supports the work of Popescu et al. (2009, 2015) and Carnevale et al. (2006) and  
708 others who have proposed a return to marine conditions prior to the Zanclean, though this interpretation  
709 needs further corroboration. Finally, regional and local tectonic uplift (e.g., Boulton and Robertson, 2008;  
710 Boulton and Whittaker, 2009) meant that the Hatay Graben rapidly shallowed during the Pliocene resulting  
711 in continental or coastal sediments and the deposition of Gilbert-type deltas and associated coastal and  
712 fluvial systems.

713         Therefore, this examination of the Miocene to Pliocene transition outcropping in the Hatay Graben  
714 shows that the proposed stratigraphic framework for the whole Mediterranean region is broadly consistent in  
715 this easternmost basin. However, questions still remain regarding the timing of the return to marine  
716 conditions and the possibility that the refilling of the Mediterranean had commenced by the Zanclean, as well  
717 as to the significance of the ‘mega-breccias’ seen in many regions and their possible connection to the Lago  
718 Mare event.

719

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726

727

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## 1135 **Figures**

1136 Table 1. Sedimentological and facies data for the Nurzeytin and Samandağ formations.

1137

1138 Figure 1. (A) Summary stratigraphic model for the three stages of deposition characteristic of the MSC Crisis  
1139 in the Mediterranean; PLG - Primary Lower Gypsum, RLG – Resedimented lower Gypsum (modified from

1140 CIESM, 2008; Roveri et al., 2014). Note: the numbers (1, 2, 3.1, 3.2) refer to stages of the MSC. (B)

1141 Schematic classification of Messinian sub-basins in the Mediterranean (modified from Roveri et al., 2014)

1142 showing shallow, intermediate (these basin are also known as peripheral/marginal) and deep water basins.

1143

1144 Figure 2. Plate tectonic overview of the Eastern Mediterranean showing the location of key Messinian to

1145 Zanclean deposits in the Eastern Mediterranean; EAFZ – East Anatolian Fault Zone; DSZF – Dead Sea Fault

1146 Zone; M-AL – Misis-Andirin lineament; FBFZ – Fethiye-Burdur Fault Zone: 1) Dardanelles (Melinte-

1147 Dobrinescu et al., 2009); 2) Asparta (Flecker et al., 1998); 3 and 4) Polemi, Pissouri, Maroni and Mesaoria

1148 Basins of Cyprus (e.g., Robertson et al., 1995); 5) IODP leg 161 (Iaccarino et al., 1999a, b); 6) Adana Basin,

1149 (Darbas and Nazik, 2010; Ilgar et al., 2012); 7) Iskenderun Basin (Tekin et al., 2010); 8) Hatay Graben (this

1150 paper; Boulton et al., 2006, 2007; Tekin et al., 2010); 9) Latakia Graben (Hardenberg and Robertson, 2009,

1151 2012).

1152

1153 Figure 3. Geological map of the study area showing the location of places and sections described in the text,

1154 modified from Boulton et al. (2006) and Tekin et al. (2010). ① Mağaracik Section; ② Ortatepe; ③ Mizrakli

1155 – Nurzeytin Fm., type section; ④ Quarry; ⑤ Sutası Log, ⑥ Road cutting.

1156

1157 Figure 4. Stratigraphic column for the Cenozoic strata of the Hatay Graben (modified from Boulton et al.,

1158 2007).

1159

1160 Figure 5. Photograph and sketch of Samandağı Formation sediments of presumed Pliocene age exposed

1161 north of Mağaracik (UTM Zone 35 S; 0765400/4000510)

1162

1163 Figure 6. Photograph and field sketch of the downlap surface observed along the terrace at Ortatepe Tepe,

1164 where the Pliocene (?) Samandağı Formation overlies the upper Miocene Nurzeytin Fm., (Grid Ref:

1165 0769750/3998399).

1166

1167 Figure 7. Micropalaeontological results from Ortatepe (location 2; Fig. 3) plus log from the Nurzeytin  
1168 Formation below the downlap surface seen in figure 6. The key for the log is shown on figure 9.

1169

1170 Figure 8. (A) Sedimentary log of Miocene-Pliocene boundary section observed on the Antakya-Samandağ  
1171 road (location 4; Fig. 3) showing location of samples taken for microfossil analysis – key is shown on figure  
1172 9. (B) Photograph of the section, with location of the logged section indicated with the arrow, Note: the  
1173 slight angular discordance between the lower Nurzeytin Formation and the overlying Samandağı Formation

1174

1175 Figure 9. Two sedimentary logs of the Samandağı Formation (A) Log of the Sutası Section (modified from  
1176 Boulton et al., 2007). (B) Log of location 6 (Fig. 3), showing the stratigraphic position of the Sr  
1177 measurement reported by Boulton et al. (2007). Key shown is for all logs.

1178

1179 Figure 10. Photographs illustrating gypsum facies of the Hatay Graben. A) Coarse-grained *in situ* selenite  
1180 crystals up to 4 cm long and B) laminated and interbedded *in situ* selenite and alabastrine from near Mizrakli,  
1181 C) Fine-grained reworked selenite crystals from Sebenoba. Note the lens cap (5 cm diameter) for scale on  
1182 each photograph. D) large alabastrine blocks in a gypsiferous marl matrix forming the ‘mega-breccia’ as  
1183 observed near Vaklıflı.

1184

1185 Figure 11. Sketch stratigraphic correlation (horizontal spacing is not to scale) approximately west to east  
1186 between the key sections (indicated by numbers) discussed in the text and locations shown on figure 3. Note  
1187 the similarity of the facies described here to the idealised model of the Messinian deposits for a peripheral  
1188 basin shown in figure 1a. MES – Messinian Erosion Surface, RLG – Resedimented Lower Gypsum.

1189