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► **To cite this version:**

Cerennaz Bozyiğit, Kürşad Kadir Eriş, Marie-Alexandrine Sicre, Memet Namik Çağatay, Gülsen Uçarkuş, et al.. Middle-late holocene climate and hydrologic changes in the Gulf of Saros (NE Aegean Sea). *Marine Geology*, 2022, 443, pp.106688. 10.1016/j.margeo.2021.106688 . hal-03414955

HAL Id: hal-03414955

<https://hal.science/hal-03414955>

Submitted on 11 Nov 2021

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1 **Middle-late Holocene Climate and Hydrologic Changes in the Gulf of Saros (NE**
2 **Aegean Sea)**

3 Cerennaz Bozyigit^{a, b}, Kürşad Kadir Eriş^{b, c}, Marie-Alexandrine Sicre^d, Memet Namık
4 Çağatay^b, Gülsen Uçarkuş^b, Vincent Klein^d, Luca Gasperini^e

5 ^aDepartment of Climate and Marine Science, Eurasia Institute of Earth Science, Istanbul
6 Technical University, Ayazağa, 34469, İstanbul, Turkey

7 ^bDepartment of Geological Engineering, EMCOL Applied Research Center, Istanbul
8 Technical University, Ayazağa, 34469, İstanbul, Turkey

9 ^cDepartment of Ecology and Evolution, Eurasia Institute of Earth Science, Istanbul
10 Technical University, Ayazağa, 34469, İstanbul, Turkey

11 ^dLOCEAN, Sorbonne Université, CNRS, Campus Pierre et Marie Curie, 4 place Jussieu,
12 75252 Paris Cedex 05, France

13 ^eInstitute of Marine Sciences, ISMAR-Bo, CNR, Bologna, Italy

14 **Abstract**

15 A multi-proxy analyses was applied on the sediment core from the Gulf of Saros (GoS)
16 to identify and characterize climate and hydrological changes during the middle-to-late
17 Holocene. The formation of two discrete Holocene sapropel layers in the GoS sediments was
18 documented for the first time in the sediment core based on total organic carbon analysis.
19 According to our paleo-proxy records, the lower Holocene sapropel was deposited under
20 warm and humid climate conditions that gave rise to high delivery of terrestrial organic
21 matter by numerous rivers in the northern catchment of the GoS. Biomarker and μ -XRF data
22 were used to decipher climate variations during the middle to late Holocene. The general
23 trends of sea-surface temperature records from the GoS and Sea of Marmara (SoM) at the

24 beginning of late Holocene are in good agreement, underlying the influence of the Black Sea
25 inflow. A relatively warm and wet climate together with a high sedimentation rate during
26 the mid-Holocene Climatic Optimum resulted in high organic productivity and ensuing
27 formation of the younger Holocene sapropel between 5.4 and 3.0 cal ka BP. Late Holocene
28 European climate periods are evident in the Saros core records. The Roman Humid Period
29 is represented by high variation in climate, indicating an earlier (2.5-2.3 cal ka BP) dry and
30 a later (2.3-1.55 cal ka BP) wet periods. The abrupt return to drier condition during the Dark
31 Ages Cold Period (1.6-1.3 cal ka BP) was followed by a wetter Medieval Climate Anomaly
32 (1.1-0.7 cal ka BP). The paleo-proxy record of the core indicates a passage from a wetter to
33 drier climate during the cold Little Ice Age period (730-110 cal yr BP), and highlights the
34 influence of deforestation in the catchment of the GoS as a result of human activities during
35 the last three centuries.

36 **Keywords:** North Aegean Sea, Gulf of Saros, Marine sediment core, Geochemical proxies,
37 Paleoclimate, Sapropels

38 **1. Introduction**

39 Palaeoclimatic proxies, including stable isotopes, palynological data, and lake levels,
40 have shown that the Mediterranean region experienced climatic conditions that varied
41 spatially and temporally throughout the Holocene (e.g., Bar-Matthews and Ayalon, 2011;
42 Luterbacher et al., 2012; Lionello, 2012; Triantaphyllou et al., 2014, 2016; Mauri et al.,
43 2015; Caldara and De Santis, 2015; Sadori et al., 2016a; Cheddadi and Khater, 2016). The
44 Aegean Sea is sensitive to climate changes because of the combination of its semi-isolation
45 from the Mediterranean Sea, small size, and proximity to the Çanakkale Strait, linking it to
46 the Black Sea since the early Holocene (Poulos et al., 1996; Rohling et al., 2002) (Fig. 1).
47 Previous studies focusing on the Holocene have shown that the many long and short term
48 changes that are associated with variations in the Siberian High intensity and the Northern

49 Hemisphere Climate (e.g. Rohling et al., 2002; Geraga et al., 2010). Concurrently, Uk 37-
50 SST reconstructions in both the north and south Aegean basins (Gogou et al., 2007; Kotthoff
51 et al., 2008; Triantaphyllou et al., 2009b) reveal a warming trend that reflects the changing
52 regional climatic conditions, coinciding with the Holocene Climatic Optimum.

53 The Holocene period is associated with deposition of the sapropelic sediments in the
54 Aegean Sea as well as in the Mediterranean and Marmara seas due to substantial
55 modifications within the surface and bottom waters. Reduced oxygen supply to bottom
56 waters has been suggested to be a precondition for sapropel formation although increased
57 biological productivity further promoted S1 deposition (Bianchi et al., 2006; Myers et al.,
58 1998; Rohling, 1994; Stratford et al., 2000). The differences in its timing in different parts
59 of the Aegean Sea was explained as a response to distinct changes in the local hydrographic
60 regime and biogeochemical cycling linked to global and regional climatic variations (Aksu
61 et al., 1995; Mercone et al., 2001; De Lange et al., 2008; Kotthoff et al., 2008; Triantaphyllou
62 et al., 2009b; Filippidi, 2013). Moreover, the comparison of the north and south Aegean
63 records shows significant changes in both the productivity and the stratification of the upper
64 water column during S1 deposition. Thus, the difference in TOC concentrations among the
65 different segments of the Aegean Sea likely reflects variation in sedimentation and
66 preservation rates. Two sapropel units timely coincides with the early Holocene S1 sapropel
67 and the middle Holocene sapropel (SMH) in the southern Aegean Sea (Triantaphyllou et al.,
68 2014; Kontakiotis, 2016). The Holocene sapropel (S1) occurs in two parts; S1a and S1b that
69 are well dated from the Aegean Sea (Geraga et al., 2000; Gogou et al., 2007; Triantaphyllou
70 et al., 2009b; Filippidi, 2013).

71 NE-SW-oriented GoS is a westward-widening and -deepening triangular marine
72 embayment that constitutes the eastern extension of the North Aegean Trough (NAT) (Fig.
73 1). The gulf constitutes the eastern extension of the North Aegean Trough (NAT), and is

74 separated from the Marmara Sea by the Thrace Peninsula. The GoS was formed in the course
75 of the postglacial transgression across the Quaternary Saros Basin fill and its basement units
76 (Çağatay et al. 1998; Tüysüz et al. 1998). The hydrological setting of the GoS is controlled
77 mainly by the high river discharges from the northern and southern catchment area (Fig. 2)
78 and the regional climate conditions and notably wind forcing (Aksu et al., 1995). The gulf
79 has been in the focus of various research groups because of its tectonic setting on the North
80 Anatolian Fault Zone (NAFZ) (Çağatay et al., 1998; Tüysüz et al., 1998; Yaltrak et al.,
81 1998; Saatçılar et al., 1999; Kurt et al., 2000; Yaltrak and Alpar, 2002; Ustaömer et al.,
82 2008; Gasperini et al., 2011).

83 The GoS may have been affected by different paleoceanographic changes because of
84 influences of inflows from the Mediterranean and the Black seas (Figs. 1 and 2). The
85 connection with the Black Sea was presumably established between 4 and 5 ka BP (Tolun
86 et al., 2001; Çağatay et al., 2001), slightly after the initiation of the S1 deposition. Although
87 considerable work has been carried out on the Holocene paleoceanography at various sites
88 of the Aegean Sea (e.g. Gogou et al., 2007, 2016; Kotthoff et al., 2008; Triantaphyllou et al.,
89 2009a, b, 2014; Geraga et al., 2010; Kouli et al., 2011; Kontakiotis, 2016), only a few
90 investigations have focussed on the Holocene climate in the gulf (e.g., Çağatay et al., 1998).
91 Small volume and geographical position of the GoS together with the high sedimentation
92 rates heightened sensitivity to climatic variations during the Holocene, and thus, allow us to
93 document immediate and the short term changes for paleoclimate and paleohydrologic
94 reconstructions. Therefore, this study was projected to contribute towards the advance of
95 understanding the causes of paleoceanographic variability and predicting vulnerability to
96 potential climate change in the GoS by using a multi-proxy analyses of a sediment core. The
97 comparison of present oceanographic settings of the GoS with rest of the Aegean Sea may
98 allow us to examine definition of the Holocene sapropel in terms of timing and source for

99 its formation during changing paleoclimate and paleoceanographic situations in the gulf. For
100 this purpose, we also aim to better identify general hydrologic evolution of the gulf with a
101 special interest on formation of sapropels linked to climate variability and water exchanges
102 between the Aegean and Marmara seas.

103 **2. Physiography and Geology**

104 The Aegean Sea is divided into three physiographic regions: the northern Aegean Sea,
105 including the NAT; the central Aegean with islands, shoals, and basins; and the southern
106 Aegean Sea, including the Cretan Trough (Fig.1). The GoS is located between Thrace
107 Peninsula to the north and the Gelibolu Peninsula to the South, and is bounded to the north
108 by the active main strand of the North Anatolian Fault (NAF), named the Ganos-Saros
109 segment (Figs. 1 and 2). The gulf extends for about 60 km westwards, connecting to the
110 main Aegean Sea basin, and reaches a maximum width of about 36 km. The bathymetry of
111 the gulf is rather asymmetric, with a 10-km wide shelf to the north and a deep trough up to
112 15 km to the south. It widens and deepens toward WSW, where it becomes the easternmost
113 part of the NAT. The shelf extends at a water depth of 90–120 m. According to the present
114 bathymetry, the maximum depth is up to about 700 m. The deep trough starts to the southeast
115 of Mecidiye (Fig. 2) as a wedge-shaped depression, deepening and widening toward the
116 west. The slope leading to the deep trough has a mean gradient of 8.5° and displays an
117 irregular morphology caused by active faulting and sediment slumping. The bathymetric
118 map allows us to recognize several canyons cutting the slope, particularly along the northern
119 slope of the gulf. A close-up view of these canyons indicates that it is displaced by a strike-
120 slip fault bounding the northern slope of the gulf (Fig. 2). These canyons were presumably
121 active during the last glacial period, when the sea level was 130 m lower than the present
122 sea level (Çağatay et al., 1998; Gasperini et al., 2011; Vardar, 2019), and the shelf and the
123 uppermost slopes in the gulf were exposed as subaerial.

124 **3. Present Oceanography and Climate**

125 The Aegean Sea is an intermittent source of deep waters for the Eastern Mediterranean
126 (Yüce, 1994; Roether et al., 1996; Lascaratos et al., 1999). Episodic deep water formation
127 during winter cooling known as Eastern Mediterranean Transient (EMT) have been
128 documented (Yüce, 1994; Zervakis et al., 2000), which would have occurred over the past
129 500 years (Incarbona et al., 2016). In terms of hydrological features, it is divided into two
130 sub-basins, namely the north and south Aegean basins (Fig. 1) (Lykousis et al., 2002). The
131 inflow of Black Sea waters and from surrounding rivers together with atmosphere/ocean
132 interactions create complex hydrological conditions within the North Aegean Sea, affecting
133 the water circulation and subsequently the biological, chemical and depositional processes.

134 Climate is relatively humid in the north and semi-arid in the south. In contrast to the
135 south Aegean Sea, the north Aegean Sea is associated with cyclonic circulation (Fig. 1). The
136 present-day oceanographic setting in the Aegean Sea is regulate mainly by the fresh water
137 flowing from the European rivers and the SoM. Eutrophication can take place due to the
138 inflow of the Balkan rivers and freshwater influx from the Black Sea as well as the
139 development of downwelling and upwelling processes caused by the presence of the
140 cyclonic and anticyclonic eddies (Lykousis et al., 2002; Geraga et al., 2010). On the basis of
141 historical hydrographic data, extensive production of dense water in the north Aegean was
142 recorded in the winters of 1987 and 1992-1993 during the last decade (Theocharis and
143 Gerogopoulos, 1993; Zervakis et al., 2000).

144 Lacombe et al. (1958) mentioned for the first time that the Go S is considered as a
145 region where the deep water might be formed and ventilates the NAT and the Lemnos Basin.
146 Pazi (2008) presented the first data suggesting that cold winter of 2002 produced by strong
147 and cold NNE wind have given rise to the formation of this new deep water in the GoS. The
148 dense water formation in the GoS was established by the homogenization of water column

149 between 100 and 600 m afterwards large heat losses in January 2002 (Pazi, 2008). Three
150 main water masses have been documented in the north Aegean: (i) the surface water layer
151 with low salinity (38.85 psu) and temperature (13.8°C) reflecting the influence of Black Sea
152 Water (BSW) in the upper 100-m (ii) the Levantine Intermediate Water (LIW) characterized
153 by higher salinity (39.1 psu) and temperature (14.2°C) lying between 100-400 m, and (iii)
154 the denser and higher salinity (39.1-39.2 psu) North Aegean Deep Water (NADW) with
155 uniform temperature (13-14°C) (Zervakis et al., 2000; Lykousis et al., 2002; Velaoras and
156 Lascaratos, 2005; Geraga et al., 2010; İşler et al., 2016). The warm and saline Levantine
157 Surface Water (16-25°C; 39.2-39.5 psu) and Levantine Intermediate Water are carried by a
158 branch of the northward flow of Asia Minor Current into the Aegean Sea. The circulation in
159 the GoS includes longshore currents flowing over the northern shelf and anticyclonic eddies
160 in the interior basin (Fig. 2) (Sarı and Çağatay, 2001). Nutrient concentrations, abundance
161 of plankton and benthos and fish catch density have been recorded higher in the north-
162 northwest Aegean Sea, including the GoS than in the south-southeast Aegean Sea (Stergiou
163 et al., 1997; Pazi, 2008; Frontalini et al., 2015). Moreover, the Black Sea surface outflow in
164 the NE Aegean Sea has led to the enrichment of dissolved organic carbon and dissolved
165 organic nitrogen in the area (Polat and Tugrul, 1996). Based on historical meteorological
166 data, the GoS has been experienced cold years in 1991, 1992, 1993 and 2002, indicating a
167 sharp decrease of temperature down to 6°C (Pazi, 2008). Low SST data was documented
168 during the winter of 1993 in both North Aegean and the GoS, indicating that the formation
169 processes of dense water are predictable in these areas (Theocharis and Georgopoulos,
170 1993). Mean air temperature of 3.95°C was recorded in the GoS between 30 November and
171 31 December 2001 while low SST values were observed in winter 2002 (Pazi, 2008).

172

173

174 **4. Materials and Methods**

175 4.1. Core Sampling

176 Core SAG-14 (326.5 cm) was recovered from the GoS onboard the R/V Urania during
177 the “MARM11” cruise at 592 m water depth with a 1.2 tons gravity corer (Table 1). The
178 core segments, 1-m-long each, were split into two halves, photographed and described (Fig.
179 3) in the Istanbul Technical University (ITU)-Eastern Mediterranean Centre of
180 Oceanography and Limnology (EMCOL). The working halves were used for
181 lithostratigraphic studies and sampled for total organic carbon (TOC) and radiocarbon
182 dating. The second halves archived at ITU-EMCOL core depository were used for μ -X-ray
183 Fluorescence (XRF) and Multi-Sensor Core Logger (MSCL) analyses.

184 4.2. Multi-Sensor Core Logger (MSCL)

185 Physical properties such as magnetic susceptibility (MS) were measured by using
186 Geotek Multi-Sensor Core Logger Analyzer (0.5 cm resolution) at ITU-EMCOL according
187 to the standard procedures (Weaver and Schultheis, 1994). The data were used to provide
188 information about changes in the relative amount of terrigenous inputs.

189 4.3. μ -XRF analysis

190 Core SAG-14 was analysed for multi-element composition at 0.5 mm resolution using
191 an Itrax XRF Core scanner, equipped with XRF-EDS, X-Ray radiography and RGB colour
192 camera at the ITU-EMCOL Core Analyses Laboratory (Croudace et al., 2006; Thomson et
193 al., 2006). A fine-focus Mo X-ray tube was used as the source. The X-ray generator was
194 operated at 30 kV and 50 mA, and a counting time of 20 s was applied. The relative elemental
195 abundances were recorded as counts per second (cps) and the element-ratio profiles were
196 used as proxies for paleoenvironmental reconstructions.

197 Potassium (K) is a proxy for detrital inputs, in particular illite-mica type of clays that
198 are mainly weathering products of rock-forming minerals (mainly feldspars) during soil
199 formation. Iron (Fe) is a redox-sensitive element, which together with sulphur (S), forms Fe-
200 monosulphides (i.e. greigite, mackinawite) or pyrite by authigenesis or diagenesis in
201 organic-rich anoxic sediments (Leventhal, 1983; Berner, 1985; Lyons, 1997). In oxic
202 sediments, however, Fe occurs mainly as oxyhydroxides and in clay mineral structures, and
203 is hence considered as a detrital mineral proxy, often showing a good correlation with Ti.
204 Therefore, Ti and Fe are generally considered reliable proxy for paleoceanographic and
205 paleoenvironmental reconstructions, providing insights on climate conditions prevailing in
206 the source area as well as on the mechanisms involved in the transport of material from
207 catchment basin to the sea floor (riverine and eolian inputs). Manganese (Mn) is another
208 redox sensitive element. It readily precipitates as Mn-oxides under oxic conditions but
209 remains in a dissolved form and is depleted in anoxic sediments (e.g. Calvert and Pedersen,
210 1993; Thomson et al., 1995). The Fe-Mn ratio was used to assess the level of oxygenation
211 of bottom waters, with higher values of this ratio indicating stronger anoxic conditions.

212 Calcium (Ca) is most likely derived from endogenic calcite and is related to calcifier
213 production (e.g. Cohen, 2003; Çağatay et al., 2014, 2015; Francke et al., 2016). Strontium
214 (Sr) is an indicator of the presence of aragonite. The Ca-Ti ratio provides information on the
215 relative amounts of carbonate or siliciclastic components. The Ca-Sr ratio is also used to
216 discriminate between marine and terrestrial carbonates. This ratio is higher in terrestrial
217 carbonates than marine ones because of the high concentration of Sr in seawater (~70 times
218 the average river water concentration) (Angino et al., 1966; Palmer and Edmond, 1992).

219 4.4. Total Organic Carbon Analysis

220 The total organic carbon (TOC) contents was determined from freeze-dried samples
221 using a Shimadzu TOC/TIC analyser at ITU-EMCOL. First, the total carbon (TC) content

222 of sediments was determined based on the amount of carbon dioxide (CO₂) released by total
223 catalytic combustion at 900°C in the presence of purged oxygen and platinum catalyst. Then,
224 total inorganic carbon (TIC) was obtained from the amount of CO₂ produced after sediment
225 treating with %85 phosphoric and heating at 200°C. The TOC content was calculated by
226 subtracting TIC from TC. TOC content is indicative of organic matter production (Burdige,
227 2007).

228 4.5. Biomarker Analysis

229 Biomarker analyses were performed by using a Thermo Fisher Trace Ultra Gas
230 Chromatograph (GC) Analyzer at LOCEAN, Sorbonne University. The core was sampled
231 at a 1 cm step, and freeze-dried. Lipid extraction was performed in a mixture of
232 dichloromethane/methanol (2:1 v/v) (Sicre et al., 1999). Alkenones were isolated from the
233 total lipid extracted by silica gel chromatography using solvent mixtures of increasing
234 polarity (Sicre et al., 1999). 5 α -cholestane was added as an external standard prior to GC
235 injection to calculate concentration. Detailed analytical procedure can be found in Ternois
236 et al. (1996). Di- and tri- bounds C₃₇ alkenone compounds were used to derive the C₃₇
237 unsaturation index ($U^{K'_{37}} = C_{37:2} / (C_{37:2} + C_{37:3})$). Alkenone-SSTs were calculated using the
238 global calibration of Conte et al. (2006) $T(^{\circ}C) = -0.957 + 54.293 (U^{K'_{37}}) - 52.894 (U^{K'_{37}})^2 +$
239 $28.231 (U^{K'_{37}})^3$.

240 4.6. Radiocarbon analysis and age-depth model

241 Accelerated mass spectroscopy (AMS) radiocarbon ¹⁴C dating of core SAG-14 was
242 performed at TÜBİTAK-MAM facility. Three sediment samples (Table 2) selected for
243 dating were collected from hemipelagic parts of the core, washed under distilled water and
244 dried at 40°C before the analysis. Mostly planktonic foraminifera and small amount of
245 epifaunal benthic and echinoderm spicules were used as dating material. AMS ¹⁴C results

246 were calibrated by using Calib v7.1 software with Marine13 ¹⁴C calibration curve (Reimer
247 et al., 2013). Reservoir ages of 149 ± 30 years were assumed for the sapropel interval
248 (Facorellis et al, 1998), and of 58 ± 85 outside the sapropel (Reimer and McCormac, 2002).
249 The age-depth model of the studied core was constructed using all ages and plotted in Fig.
250 4. Non-Bayesian age-depth model was determined using the script “clam” on R-studio
251 (Blaauw, 2010). The script creates a non-Bayesian, cubic spline age-depth model,
252 calculating the %95 Gaussian confidence interval around the best model.

253 **5. Results**

254 5.1. Core lithology, chronology and sedimentation rates

255 Visual observation of core SAG-14 allowed for compilation of a lithology/texture log,
256 indicating composite fauna of euryhaline molluscs, as well as benthic and pelagic
257 foraminifera (Fig. 3). Although macro fossils are very scarce, rich benthic and planktonic
258 assemblages are present, which include *Ammonia tepida*, *Uvigerina mediterranea*, *Brizalina*
259 *spathulata*, *Elphidium crispum*, *Lobatula lobatula*, *Textularia truncata*, *Textularia bocki*,
260 *Quinqueloculina seminula*, *Hyalina baltica*, *Spiroloculina excavata* and *Cassidulina*
261 *carinata* and *Globigerinoides ruber*, *Globigerinoides seiglei*, *Globigerinoides conglobatus*,
262 *Globigerinoides elongates*, *Globigerinina glutinata*, *Globigerina bulloides*, *Glubigerina*
263 *umbilicata*, *Globorotaloides hexagon*, *Orbulina universa* and *Globorotalia* sp. At the bottom
264 of core SAG-14, the layer until 303 cmbsf is composed of dark olive gray homogenous silty
265 clay with Fe-monosulphide (FeS) patches and silt lense. The layer between 303-274 cmbsf
266 is represented by olive gray homogenous clay. This is sharply overlain by dark olive gray
267 homogenous clay, and is followed by thick layer of olive gray homogenous clay with
268 intercalated light olive brown layer and scattered FeS patches between 196 and 101 cmbsf.
269 The overlying unit between 101 and 29 cmbsf is composed of olive homogeneous clay with
270 rare bioturbation and FeS patches. The uppermost 27 cm-thick layer in the core contains

271 alternations of reddish brown clays with olive gray homogenous clay having abundant FeS
272 patches (Fig. 3).

273 Our age-depth model indicates that the sedimentary sequence encompasses the last 7
274 cal ka BP (Figs. 3 and 4; Table 2). Based mainly on TOC analysis and physical properties,
275 we define two sapropel layers in intervals 326.5-297.4 cmbsf (S1b sapropel) and 277.4-193
276 cmbsf (SMH sapropel) (see section 5.2 below). According to the age-depth models, sapropel
277 S1b and SMH sapropels are dated 7-6.1 cal ka BP and 5.4-3.0 cal ka BP, respectively. The
278 calculated sedimentation rate indicates a high variation between the middle and late
279 Holocene periods. During deposition of the S1b sapropel (7-6.1 cal ka BP), the approximate
280 sedimentation rate is around 0.31 mm/yr, whereas a slight rise to 0.35 mm/yr occurs for the
281 younger SMH sapropel (5.4-3.0 cal ka BP). Following the SMH sapropel deposition, the
282 sedimentation rate increases to 0.64 mm/yr and exceeds 0.75 mm/yr during the last 300 yrs.

283 5.2. Physical and Geochemical analyses

284 In core SAG-14, the two sapropel layers between 326.5 and 194.8 cmbsf interval
285 possesses > 1 wt% TOC (Fig. 5). S1b sapropel (326.5-297.4 cmbsf) contains TOC values,
286 ranging between 0.75 wt% and 1.42 wt%. It is characterized by a relatively high MS value
287 and increasing μ -XRF K, Fe and Ca counts and Ca-Ti ratio in contrast to μ -XRF Ca-Sr and
288 Fe-Mn ratios. The S.R-TOC ratio (sedimentation rate versus total organic carbon) in the S1b
289 is very low at the base but shows a gradual increase. SSTs are relatively warm (19.8 °C) at
290 the base of S1b, but decrease by about 2°C together with the C₃₇ alkenone concentrations
291 (Fig. 5).

292 The intervening layer between S1b and SMH sapropels (297.4-277.4 cmbsf) shows
293 SST increase and slight increase in μ -XRF K and Fe in contrast to μ -XRF Ca, TOC and MS
294 values, together with Ca-Ti and Fe-Mn ratios. The overlying sapropel SMH (277.4-193

295 cmbsf) is distinct from the S1b in having strongly fluctuating and higher TOC values (0.50-
296 1.80 wt%) as well as C₃₇ alkenone concentrations and S.R-TOC ratio, although TOC content
297 drops below %1 between 215 and 231 cmbsf (Fig. 5). The whole layer is also marked by a
298 decreasing trend of μ -XRF Fe-Mn ratio, and points to the almost highest μ -XRF Ca value
299 and Ca-Ti ratio of the core with strong fluctuations that share resemblance with TOC. The
300 elevated SST profile and μ -XRF K, Fe and MS values in the lower part of the SMH sharply
301 decrease above 237.6 cmbsf.

302 Following the SMH, the interval between 193 and 172 cmbsf is marked mainly by a
303 rising trends in μ -XRF K, Fe and MS values that negatively correlate with μ -XRF Ca value
304 and Ca-Ti and Ca-Sr ratios as well as TOC content and μ -XRF Fe-Mn ratio. The abrupt
305 increase in μ -XRF Fe, K and MS values as well as Fe-Mn ratio at the base of the overlying
306 layer (172 cmbsf) is followed by strong fluctuations between 158 cmbsf and 118 cmbsf. μ -
307 XRF Ca value and Ca-Ti and Ca-Sr ratios indicate opposite trends. The overall trend of TOC
308 contents opposite to C₃₇ alkenone concentrations that are recorded the highest between 138-
309 125 cmbsf. The lower SST values above the SMH layer, except sharp increase at 178 cmbsf,
310 display elevated trend until 152 cmbsf and later decline above. The progressive decreases in
311 μ -XRF Fe and K values as well as S.R-TOC ratio between 118.6 and 112 cmbsf in the core
312 is followed by opposite trend until 90.6 cmbsf. This interval is also associated with gradual
313 decreasing μ -XRF Ca value and Ca-Ti and Ca-Sr ratios as well as SSTs, except for slight
314 increases above 106 cmbsf (Fig. 5). μ -XRF Fe-Mn indicates a prominent sharp increase
315 above 110 cmbsf.

316 The interval between 90.6 and 61.6 cmbsf is earmarked by increasing values of μ -XRF
317 K and Fe in comparison to MS values and μ -XRF Fe-Mn ratio. The relatively high values
318 of μ -XRF Ca count and Ca-Ti and Ca-Sr ratios as well as SST record indicate slight
319 decreasing trends. Elevated TOC content in the same interval displays an inverse trend with

320 S.R./TOC ratio. The layer between 61.6-34 cmbsf is marked mainly by lower μ -XRF K and
321 Fe values and Fe-Mn ratio with strong fluctuations, followed by a sharp increase until 14.3
322 cmbsf. The opposite trend is recorded in μ -XRF Ca value and Ca-Ti and Ca-Sr ratios. The
323 overall TOC content in the same layer indicates high values, as opposed to S.R-TOC ratio.
324 The highly-fluctuated relatively low SST values in contrast to C₃₇ alkenone concentrations
325 indicate a prominent increase above 14.3 cmbsf.

326 **6. Discussion**

327 According to the average high sedimentation rate at the core site, paleo-proxy record
328 from core SAG-14 provides a high temporal resolution for reconstructon of paleoclimatic
329 and palaeohydrological changes in the GoS over the last 7 cal ka BP. The overall lack of
330 agreement about the worldwide distribution, precise timing, amplitude or cause of the
331 Holocene climate and hydrologic changes underlines the need for additional paleo-proxy
332 records from different areas such as the GoS. For this purpose, we compare our findings with
333 other high-resolution marine paleo-climate records from the Aegean and Marmara seas and
334 also regional continental records from NW Anatolia (Figs. 6 and 7). These include the
335 palynological records from Core SL-152 in the northern Aegean Sea, located in an area
336 neary region to the study area and the speleothem records from Sofular Cave in NW
337 Anatolia. Even though the former study is low resolution (10 cm), it provides the analysis of
338 multidecadal vegetation changes and moisture anomaly records during the mid-to-late
339 Holocene that are linked with regional climate changes (Kotthoff et al. 2008), while the latter
340 presents high-resolution changes in temperature and precipitation in the region (Fleitmann
341 et al., 2009).

342

343

344 6.1. Hydrology and climate during sapropel formations

345 The TOC data together with chronology obtained from core SAG-14 identify the
346 formation of two discrete Holocene sapropel layers (S1b and SMH) in the gulf, including
347 their timing and source regarding to changing paleohydrologic and paleoclimate conditions.
348 The Holocene S1 sapropel layer in the Aegean Sea is subdivided into two different units, the
349 older S1a and younger S1b, by an intervening low TOC layer (Gogou et al., 2007;
350 Triantaphyllou et al. 2009b). The reason for such interruption in the sapropel deposition is
351 related to increased ventilation of bottom waters (Casford et al., 2003). Our core sampled
352 only the S1b part of S1 sapropel and a younger sapropel layer (SMH) for the first time in the
353 GoS. The sapropelic layers (S1b and SMH) in the core are recognisable as dark olive gray
354 homogenous clay beds, containing mainly planktonic fauna with scarce or no benthic
355 assemblages (Figs. 3 and 5). The scarcity or absence of benthic fauna in the sapropelic layers
356 in our core is similar to the age-correlatable sapropels in the Aegean Sea and the SoM,
357 suggesting persistent sea-floor anoxia/hypoxia in the gulf (Rohling et al., 1993, 1997;
358 Çağatay et al., 1999; Casford et al., 2003).

359 *6.1.1. Sapropel S1b interval (7.0-6.1 cal ka BP)*

360 The timing of sapropel S1b in the Aegean Sea is highly debated based on various core
361 studies, ranging from 7.8 to 6.1 cal ka BP. In this study, the core chronology indicates that
362 termination age of S1b sapropel layer in the core is in agreement with the recent studies in
363 the Aegean Sea, suggesting an age interval between 7.3 and 6.1 cal ka BP (Figs. 4 and 5)
364 (Merccone et al., 2001; De Lange et al., 2008; Triantaphyllou et al., 2009b; Filippidi, 2013).
365 In particular, the termination age of S1b sapropel at ca 6.1 cal ka BP in our core matches
366 well with that of the eastern Mediterranean S1 sapropel (e.g., Rohling, 1994; Aksu et al.,
367 1995; De Lange et al., 2008). Comparison of the north and south Aegean Sea records shows
368 significant changes in both the productivity and stratification of the upper water column

369 during S1 deposition. The TOC average values of the sapropel layer in the north Aegean Sea
370 were previously reported between 0.7 % and 1.9 % (Triantaphyllou et al., 2009b; Filippidi,
371 2013; İşler et al., 2016) that is well in agreement with our TOC values (0.75 wt% and 1.42
372 wt%) for the S1b sapropel in the core.

373 Deposition of the sapropel S1b in the gulf is associated with increasing trends of μ -
374 XRF Ca value and Ca-Ti ratio in the core (Fig. 5), suggesting an increasing production of
375 biogenic carbonate. The general rise of μ -XRF K and Fe values together with high MS are
376 indicative of enhanced detrital inputs during the sapropel deposition (7-6.1 cal ka BP) due
377 to increased precipitation. In this context, the gradual decrease in C₃₇ alkenone
378 concentrations most probably reflect lower preservation or higher dilution of marine organic
379 matter by detrital material supplied by the local delivery of freshwaters to the GoS via the
380 numerous rivers from the northern catchment (Fig. 2). Increased precipitation during this
381 period is also supported by the gradual decrease in $\delta^{13}\text{C}$ values at the Sofular Cave site in
382 NW Anatolia (Fleitmann et al., 2009) and increased Arboreal type (AP) vegetation from the
383 northern Aegean Sea (core SL-152; Fig. 6) (Kotthoff et al., 2008). However, the Holocene
384 summer precession-related insolation maximum in the Northern Hemisphere (Laskar et al.,
385 2004) and the monsoon intensification would have resulted in widespread humidity over the
386 whole Mediterranean region during the S1b deposition (Rossignol-Strick, 1983; Rohling,
387 1994).

388 The relatively high μ -XRF Fe-Mn ratio in the lower part of S1b sapropel (Fig. 5) is
389 followed by concomitant decrease upward, suggesting an increase in deep-water ventilation
390 during the later period of sapropel formation. This may be responsible for the relatively low
391 organic matter and alkenone levels in the core. Similar hydrological conditions during the
392 sapropel deposition has also been reported from the north Aegean Sea by Geraga et al. (2010)
393 who claimed that the water stratification would have been mainly restricted to the surface.

394 Such phenomenon demonstrates that deep water condition in the gulf was not possibly fully-
395 anoxic during the S1b deposition, but still convenient for deposition of sapropelic sediment.
396 Previous studies in the Aegean Sea (Triantaphyllou et al., 2009a, b; Kontakiotis, 2012, 2016)
397 show that warm and humid climate conditions could have triggered water column
398 stratification. Therefore, the paleoceanographic conditions in the gulf as indicated by warm
399 and partly stratified water column together with enhanced detrital organic matter delivery
400 could play the most important role in the deposition of S1b in the GoS. Such environmental
401 conditions would have promoted the Holocene sapropel deposition as proposed for the SoM
402 by earlier core studies (Çağatay et al., 2000, 2015; Tolun et al., 2002).

403 The progressive decrease in SST values during the S1b deposition in core SAG-14
404 (Figs. 5 and 7) reflects sea surface cooling upto ca. 1.5–2 °C due to the increased
405 precipitation that could have led to surface water density decreases in the gulf. Similar SST
406 decline has been reported over the same period in the northern Aegean Sea (Kotthoff et al.,
407 2008) and in the SoM (Sperling et al., 2003) (Fig. 7), suggesting similar hydrologic
408 conditions during the sapropelic deposition. According to previous observations from the
409 Aegean Sea (Geraga et al., 2010), the elevated percentages of the species *G. bulloides* and
410 *T. Quinqueloba* within S1b are more pronounced in the north than in south Aegean, and are
411 related to low salinity and high fertility of the surficial waters, possibly due to the larger river
412 inflows draining the north Aegean catchment area. However, this sea-surface cooling could
413 also be indicative of stronger winds and enhanced mixing that would have provoked
414 intermittent ventilation of deep waters throughout the S1b as inferred from decreasing μ -XRF
415 Fe-Mn ratio in the core (Fig. 5).

416 *6.1.2. Post-sapropel S1b interval (6.1-5.4 cal ka BP)*

417 The paleo-proxy record in the interval between sapropels S1b and SMH is marked by
418 a sharp drop in the TOC content, implying diminished organic matter productivity in the gulf

419 (Fig. 5). This intervening layer is also associated with decreases in μ -XRF Ca value and Ca-
420 Ti ratio, suggesting low biogenic carbonate production and organic productivity. The sharp
421 decreases in μ -XRF Fe-Mn ratio and MS at the termination of S1b, therefore, suggest a
422 return to fully oxygenated bottom waters and lower preservation of organic matter.
423 Therefore, weakened primary productivity and water column oxygenation were the limiting
424 factors for the burial of organic matter and organic-rich layer deposition. During the same
425 interval, the gradual increases in μ -XRF K and Fe values indicate sustained detrital input to
426 the GoS from the catchment area due to a wetter climate. Increased precipitation during this
427 period is also depicted by the abrupt negative excursion in $\delta^{13}\text{C}$ values at the Sofular Cave
428 site in NW Anatolia (Fleitmann et al., 2009) (Fig. 6), whereas the amount of AP in core SL-
429 152 from the northern Aegean Sea (Kotthoff et al., 2008) is commonly high, implying warm
430 and humid conditions. In addition, the progressive increase of SST in core SAG-14
431 coherently indicates warming (Fig. 5).

432 *6.1.3. Sapropel SMH interval (5.4-3.0 cal ka BP)*

433 The existence of a younger sapropel layer (SMH) in core SAG-14 is defined and dated
434 between 5.4 and 3.0 cal ka BP based on the age-depth model together with TOC data (Figs.
435 4 and 5). According to earlier studies (Kouli et al., 2011; Triantaphyllou et al., 2014;
436 Kontakiotis, 2016), deposition of the middle Holocene Sapropel (SMH) has only been
437 recorded from the southern Aegean Sea during 5.4-4.3 cal ka BP. Our termination age for
438 the SMH in the studied core differs by 1.3 ka BP from the southern Aegean Sea, but
439 coincides with the termination of the upper Holocene sapropel reported in the SoM (Çağatay
440 et al., 1999, 2000; Tolun et al., 2002). The differences in the timing of sapropel deposition
441 in the gulf, relative to those of the quasi-equivalent Mediterranean sapropel, can be attributed
442 to the intensity of environmental changes such as timing of establishment of the
443 suboxic/anoxic bottom water conditions, rate of primary productivity, and the size and depth

444 of the basins. The SMH layer in core SAG-14 contains a higher TOC content relative to the
445 S1b, except for lower values between ca 4.5 to 3.5 cal ka BP. The possible reasons must be
446 either greatly enhanced surface productivity, which would suggest a different nutrient
447 regime compared to condition during the formation of S1b sapropel, or to a better
448 preservation of organic matter under higher sedimentation or combination of both. However,
449 the depth of the oxycline and hence the oxygen exposure time of the sinking material would
450 have been variable for each sapropel interval (S1b and SMH) as indicated by a high variation
451 in C₃₇ alkenone concentration in the core (Fig. 5).

452 The 5.4-4.2 cal ka interval during the SMH deposition is accompanied by increases in
453 μ -XRF K, Fe and MS values (Fig. 5), suggesting elevated detrital inputs to the GoS due to
454 either more precipitation or weaker vegetation cover. This circumstance would have induced
455 a relatively higher sedimentation rate (0.35mm/yr) in the GoS based on the core chronology
456 (Fig. 4) that in turn likely provides a better preservation of organic matter. This may in part
457 explain the higher carbonate content of the sediment as inferred from the relatively higher
458 μ -XRF Ca value and Ca/Ti ratio in the core. During the same period, increased $\delta^{13}\text{C}$ values
459 in the Sofular Cave record (Fleitmann et al., 2009) and decrease in AP ratio in core SL-152
460 (Kotthoff et al., 2008) (Fig. 6) suggests relatively drier climate and poor vegetation cover.
461 These overall findings suggest that the warm and humid phase associated with African
462 monsoon forcing apparently was more weakened in the north Aegean Sea and GoS during
463 this period. On the other hand, gradual increase in SST in core SAG-14 until 4.2 cal ka BP
464 supports prevailing warm and dry climate around the GoS during the earlier SMH deposition
465 (5.4-4.2 cal ka). Indeed, our alkenone SST-record suggests significant climatic instability in
466 the GoS during the overall SMH deposition, as implied by 2–3 °C SST variations. These
467 findings are in agreement with SST records from the northern and southern Aegean Sea
468 (Kotthoff et al., 2008) and from the SoM (Sperling et al., 2003), although temporal

469 resolutions are much lower (Fig. 7). The relatively higher μ -XRF Fe-Mn values in
470 comparison to later phase of the SMH deposition indicate a weaker ventilation in response
471 to better water column stratification in the gulf. Therefore, warming of the surface water in
472 the gulf due to a dry climate during this earlier sapropel deposition would have induced a
473 decrease in surface density and, thus, prevented the ventilation of deeper layer, promoting
474 the preservation of organic matter in the sediment. The termination of the humid phase
475 during the earlier SMH deposition suggests a general trend to climatic aridification over the
476 Late Holocene, which is in accordance with the salinity increase and oligotrophic and dry
477 nature of the water column of the modern Aegean Sea.

478 The latest part of the SMH deposition in the GoS between 4.2-3.0 cal ka BP is marked
479 by sharp decreases in μ -XRF K and Fe as well as MS values (Fig. 5), implying reduced
480 detrital inputs. This result can be explained by different factors, including an increased
481 vegetation cover around the catchment area that would have limited erosion and sediment
482 delivery to the gulf. This hypothesis is supported by a prominent decrease in $\delta^{13}\text{C}$ values at
483 the Sofular Cave site (Fleitmann et al., 2009) and increased AP ratio in core SL-152
484 (Kotthoff et al., 2008) (Fig. 6), indicative of the enhanced vegetation cover due to a humid
485 climate phase, timely coinciding with the Mid-Holocene Climatic Optimum. Triantaphyllou
486 et al. (2014) relates its deposition in the southern Aegean Sea to upper water column
487 stratification and enhanced deep chlorophyll maximum under the Mid-Holocene wet and
488 warm conditions resulted from weak Mid-Holocene South Asian monsoon forcing and the
489 Etesian (Meltem) winds.

490 The enrichment of Ca relative to Sr in core SAG-14 indicates that the composition of
491 the SMH sapropel is dominated by carbonate rather than aragonite, implying a lower surface
492 water evaporation in the gulf. This conclusion is in agreement with surface freshening in the
493 GoS after 4.2 cal ka BP due to increased precipitation as inferred from decreasing SSTs,

494 except for a short warming event at around 3.3 cal ka BP (Fig. 5). A similar post-4.2 cal ka
495 BP SST decline has also been documented in northern and southern Aegean Sea (core SL-
496 152) and the SoM (core KL-71) (Fig. 7), reflecting changing regional climatic conditions
497 during the SMH deposition. Such low salinity and high freshening of surface waters in the
498 gulf were possibly driven by enhanced riverine discharges and land runoff from the
499 surrounding the northern and southern catchments of the GoS that was greater towards the
500 end of the SMH sapropel. This rapid surface water cooling during the later phase of
501 sapropelic deposition most likely provoked intermittent ventilation of deep waters
502 throughout the SMH deposition in the gulf. However, we cannot rule out the possible
503 contribution of wind-driven surface water cooling as the μ -XRF Fe-Mn ratio in the upper
504 part of the SMH in the core still shows increasing ventilation. The continuous gradual
505 decrease in μ -XRF Fe-Mn ratio (Fig. 5) underpins better ventilated deep-waters as compared
506 to the S1b sapropel. Such a weak stratification might have been compensated by enhanced
507 productivity and a relatively higher sedimentation rate (0.35mm/yr) to account for the
508 formation of SMH sapropel. The fluctuations in C₃₇ alkenone concentrations together with
509 high S.R-TOC ratios in the SMH layer indicate variable origin of the organic matter.
510 Together with a climatic influence on freshening of the GoS waters, a possible contribution
511 of Black Sea outflow via the SoM after 4.4 cal ka BP (Çağatay et al., 2000; Algan et al.,
512 2001) should also be considered.

513 6.2. Late Holocene climate events

514 The late Holocene period is characterized by several climatic fluctuations that are
515 mainly associated with historical periods documented by Büntgen et al. (2011) in central
516 Europe. These various periods are examined here as the Roman Humid Period (RHP), the
517 Dark Ages Cold Period (DACP), the Medieval Climate Anomaly (MCA) and the Little Ice
518 Age (LIA) are also apparent in the paleo-proxy record from the GoS (Fig. 5). Although the

519 duration of each period is not always synchronous, our multi-proxy approach allows for the
520 identification and discussing the timing and intensity of those climate events in the GoS (Fig.
521 5). The SMH termination in core SAG-14 is marked by decreases in μ -XRF Ca-Ti ratio and
522 Ca value until the onset of the RHP at 2.5 cal ka BP (Fig. 5), likely reflecting diminished
523 marine carbonate production in the GoS. On the other hand, μ -XRF K and Fe values indicate
524 a pronounced increase of detrital delivery due to relatively wetter climate as supported by
525 rising AP ratio in core SL-152 (Kotthoff et al., 2008) even though $\delta^{13}\text{C}$ values at the Sofular
526 Cave site indicate positive excursion (Fig. 6). The lower SST values in core SAG-14 indicate
527 sea surface freshening due to relatively cooler climate and/or continental runoff. The further
528 decrease in μ -XRF Fe-Mn ratio evidences a change in the deep-water stratification in the
529 GoS, suggesting suboxic to oxic conditions. During this period, increased freshwater inputs
530 enhanced ventilation of deepwater masses and diminished organic matter production and
531 preservation. Thus, the deep Saros Gulf basin was fully oxygenated, thus preventing the
532 burial of organic matter and organic-rich deposition.

533 *6.2.1. Roman Humid Period (2.5-1.55 cal ka BP)*

534 The paleo-proxy record in core SAG-14 during the RHP indicates high climate
535 variability in the GoS. The earlier phase of this period (2.5-2.1 cal ka BP) is characterized
536 by decreasing trends in μ -XRF K and Fe values (Fig. 5), marking diminished detrital inputs
537 due to enhanced dryness around the GoS. This is supported by low sedimentation rate (0.42
538 mm/yr) in the core in comparison to the later period as inferred from relatively low S.R/TOC
539 ratio (Figs. 4 and 5). Although the highest AP values are recorded in the northern Aegean
540 Sea (core SL-152 in Fig. 6), the positive excursion in the $\delta^{13}\text{C}$ record at Sofular Cave in the
541 same interval indicates diminished precipitation. This aridification trend coincides with a
542 progressive evolution towards typical Mediterranean climate and aridity, which gradually
543 occurs at the onset of late Holocene (e.g. Wanner et al., 2008). This remarkable dry and

544 warm period at the beginning of the RHP has previously been documented in the Nar and
545 Tecer lakes (Jones et al., 2006; Kuzucuoğlu et al., 2011). On the other hand, higher μ -XRF
546 Ca-Ti ratio and Ca values (Fig. 5) suggest enhanced marine biogenic carbonate production
547 due to possibly SST warming upto 1.5-2 °C until 2.1 cal ka BP. Nevertheless, the relatively
548 high μ -XRF Fe-Mn ratio is followed by sharp decrease that implies a variation in the water
549 stratification, possibly from suboxic to oxic condition. Thus, the GoS was fully oxygenated,
550 preventing the burial of organic matter and organic-rich deposition.

551 A change from warm-dry to warm-wet conditions in the latest part of the RHP (after
552 2.1 cal ka BP) is recorded by higher values of μ -XRF K and Fe, suggesting the higher river
553 discharge to the gulf as a consequence of increased humidity. The higher sedimentation rate
554 (0.64 mm/yr) based on the age-depth model of the core (Fig. 4) together with gradual
555 increase in S.R/TOC ratio (Fig. 5) could be attributed to the increased riverine inputs in the
556 surficial waters in the gulf. A relatively wetter phase is also documented for the same time-
557 interval by depleted $\delta^{13}\text{C}$ values in the Sofular Cave record (Fig. 6). The freshening of sea
558 surface waters driven by enhanced riverine discharges and land runoff was likely responsible
559 for the SST decrease upto 1 °C after 2.1 cal ka BP (Fig. 5). In contrast to the northern Aegean
560 Sea, a similar SST trend has been documented in the SoM, further supporting contribution
561 of surface freshening in the GoS due to the continuous Black Sea outflow. Therefore, the
562 combination of the inflow of water from the Black Sea and abundant rivers and the
563 atmosphere–ocean interactions creates a complex hydrologic system within the GoS that
564 affects its circulation and depositional process. Although decreases in μ -XRF Ca value and
565 Ca-Ti ratio are recorded in the core, the enhanced marine organic productivity is evidenced
566 by higher C_{37} alkenone concentration during this RHP phase (2.1-1.55 cal ka BP). The
567 gradual increase in μ -XRF Fe-Mn ratio during this later part of the RHP (Fig. 5) indicates a
568 change in the deep-water stratification in the GoS, suggesting suboxic to partly anoxic

569 conditions. During this period, enhanced organic matter production together with water-
570 stratification in the gulf water masses are likely responsible for better organic matter
571 preservation as indicated by relatively higher TOC ratio (Fig. 5) in compare to the following
572 climate period after 1.55 cal ka BP.

573 *6.2.2. Dark Ages Cold Period (1.55-1.3 cal ka BP)*

574 The DACP is marked by initial sharp decreases in μ -XRF K and Fe values together
575 with S.R/TOC ratio (Fig. 5), then followed by an increase witnessing drastic changes of
576 detrital inputs. The similar variations are recorded in μ -XRF Ca value and Ca-Ti ratio till
577 1.3 cal ka BP, reflecting the scarcity and the later elevated biogenic carbonate production in
578 the GoS. The overall trends in those elements attest the former dry and later humid climates
579 in the GoS during the DACP. The general trend of AP values in core SL-152 (Kotthoff et
580 al., 2008) (Fig. 6) is agreement with our findings, indicating high climate variability during
581 the same period in the northern Aegean Sea. While the climate was relatively drier until 1.48
582 cal ka BP, SSTs cooling upto ~ 1 °C is consistent with earlier reports in the western and
583 eastern Mediterranean Sea (Sicre et al., 2016; Jalali et al., 2016; 2017) and North Atlantic
584 (Sicre et al., 2008). A progressive dryness during the DACP coincides with decreasing solar
585 insolation in the Northern Hemisphere (Steinilber et al., 2009) and cooler air temperatures
586 in Greenland (Grootes and Stuiver, 1997). Although the later humid phase and the
587 concomitant higher detrital delivery existed in the gulf, the possible reason for gradual
588 increase in SST value could be attributed to enhanced warming of the sea surface waters of
589 the gulf. Although μ -XRF Fe-Mn ratio is low at the beginning of the DACP, the sharp
590 increase in the latter half indicates the poorly oxygenated deep-water condition in the GoS
591 due to enhanced water column stratification.

592

593 *6.2.3. Medieval Climate Anomaly (1.13-0.73 cal ka BP)*

594 The most of MCA period in the GoS is represented by a relatively wetter climate and
595 subsequent higher detrital delivery as inferred from progressive increases in μ -XRF K and
596 Fe values in the core (Fig. 5). Although the Sofular Cave record indicates positive excursion
597 in $\delta^{13}\text{C}$ values since the termination of RHP (Fig. 6), the broad increase in AP values in core
598 SL-152 (Kotthoff et al., 2008) strongly indicates relatively wetter climate around the GoS.
599 Relatively higher organic matter production was likely driven by a warmer climate that
600 possibly gave rise to high marine carbonate production as inferred from general higher μ -
601 XRF Ca value and Ca-Ti and Ca-Sr ratios. While climate became wetter during most of the
602 MCA, the resultant higher river discharge to the gulf gave rise to ~ 2 °C progressive decrease
603 in SST values in core SAG-14, implying continues cooling as documented from the northern
604 Aegean (Kotthoff et al., 2008) and Marmara seas (Sperling et al., 2003; Fig. 7). Another
605 possible reason for decreasing SSTs could be the permanent connection with the Black Sea
606 during the late Holocene. Although the existence of such connection would be expected to
607 create water column stratification, considerable high river runoff would have led to intense
608 mixing of the surface water column that in turn gave rise to weakening of the stratification
609 as inferred from comparatively low μ -XRF Fe-Mn ratio in the core (Fig. 5). Even though,
610 TOC enrichment of the sediment in the core indicates higher preservation of the organic
611 matter that could only be explained by a higher sedimentation rate as documented by
612 elevated S.R/TOC ratio during this period.

613 *6.2.4. Little Ice Age (730-110 cal yr BP)*

614 The LIA period is characterized by high climate variability based on the paleo-proxy
615 record of core SAG-14 (Fig. 5). Lower μ -XRF K and Fe as well as lowest MS until 450 cal
616 yr BP comparable to the DACP imply reduced detrital input to the gulf. The possible reason
617 could be attributed to an extensive vegetation cover in the catchment area based on a higher

618 AP ratio in core SL-152 (Kotthoff et al., 2008) (Fig. 6). Relatively wetter climate in the early
619 LIA is also evidenced by negative excursion of $\delta^{13}\text{C}$ at Sofular as well as the Soreq caves
620 (Bar-Matthews et al., 1998). The lowest SSTs ($\sim 16^\circ\text{C}$) in core SAG-14 during the earlier
621 LIA period documents further sea-surface freshening that could be attributed to a higher
622 winter precipitation and cooler climate during this earlier LIA. Such phenomenon in the gulf
623 is consistent with other paleo-climate records from Europe and Anatolia (Türkeş and Erlat,
624 2005; Roberts et al., 2012; Sicre et al., 2016; Gogou et al., 2016; Oçakoğlu et al., 2016; Jalali
625 et al., 2018). In contrast to detrital input proxies (e.g. K, Fe and Ti), the general high μ -XRF
626 Ca-Ti ratio and Ca value indicate rising carbonate production and organic productivity.
627 Although the water stratification was still weak as documented by a low μ -XRF Fe-Mn ratio
628 in the core, a general high TOC content in this initial phase suggests an enhanced organic
629 matter preservation possibly due to absence of deep water circulation in the gulf (Fig. 5).

630 The higher detrital input to the GoS during the late LIA (350-110 cal yr BP) as inferred
631 from higher values in μ -XRF K and Fe (Fig. 5) could be assigned to low vegetation cover as
632 supported by the sharp decrease of AP ratio in core SL-152 (Kotthoff et al., 2008) (Fig. 6),
633 reflecting a shift to relatively drier climate condition around the GoS. On the other hand, the
634 negative excursion of $\delta^{13}\text{C}$ in the Sofular Cave record at the onset of the LIA is followed by
635 sharp increase, suggesting a shift from wet to drier climate. This is supported by a partly
636 higher SST values in the core, reaching to $\sim 18.5^\circ\text{C}$. Higher soil erosion in the catchment
637 area of the gulf could also be explained by intense deforestation due to human activities over
638 the last 350 cal yr BP. According to Wick et al. (2003), Anatolia undergoes a progressive
639 decrease of forest cover due to human activity during the last 600 yr. Pollen record from the
640 Aegean Sea shows heavy deforestation in southern Europe during the same time-interval
641 (Aksu et al., 1995). The sharp decreases in μ -XRF Ca value and Ca-Sr ratio indicates limited
642 marine organic productivity due to dry and cold climate. The strong fluctuations in μ -XRF

643 Fe-Mn ratio in the core suggests short-term changes in the deep-water condition (Fig. 5),
644 which could not have provided enough time for organic matter preservation into the
645 sediment as documented by the comparatively low TOC content in the late LIA period.

646 **Conclusions**

647 Multi-proxy analyses of core SAG-14 provide detailed insights into the
648 paleoceanographic and paleoenvironmental conditions over the last 7.0 cal ka BP in the GoS.
649 The close match between paleo-proxy data in the GoS with the marine and continental
650 records from the northern Aegean and Black seas indicates that the timing of past climate
651 events are in general agreement with the regional climate patterns. The middle-to-late
652 Holocene sediment in the GoS is characterized by deposition of two discrete sapropel layers
653 with >1% TOC values. The older sapropel layer S1b was deposited under warm and humid
654 conditions, associated with terrestrial inputs from high river discharges from the northern
655 catchment of the GoS. Enhanced productivity under weakly stratified surface waters together
656 with high sedimentation rate played the most important role in the formation of S1b sapropel.
657 The deposition of younger Holocene sapropel (SMH; 5.4-3.0 cal ka BP) initiated under
658 relatively warm and dry conditions till 4.5 ka BP, followed by cold and wet conditions
659 possibly associated with Black Sea outflow to the GoS. The μ -XRF Fe-Mn ratio indicates a
660 progressive ventilation throughout the SMH from suboxic to anoxic bottom water condition
661 and generally high and strong variable TOC content, except for lower values between ca 4.5
662 to 3.5 cal ka BP. Our data also show high variations during the RHP (2.5-1.55 cal ka BP),
663 suggesting a change from dry to wet climate. Dry climate prevailed at the beginning of
664 DACP until 1.3 cal ka BP, but evolved to relatively wetter conditions that persisted during
665 most of the MCA (1.1-0.7 cal ka BP). Similar SST trends in the northern Aegean and
666 Marmara seas pointed out common hydrological conditions in the region suggesting a
667 sustained influence of the Black Sea during the late Holocene. The LIA climate indicates a

668 shift from dry to wet climate between its early and late phases. Higher detrital delivery over
669 the last 350 years emphasize enhanced soil erosion imputable to intense deforestation due to
670 human activities around the GoS. This is consistent with the synchronous progradation of
671 southern Europe delta during the coldest phase of the LIA due to growing population and
672 land use (Maselli and Trincardi, 2013).

673 **Acknowledgements**

674 We would like to thank captains and all crew members of Urania R/V who attended the
675 MARM11 cruise in 2011. We are also thankful to CNRS and the MISTRALS/PALEOMEX
676 program for providing fundings to perform biomarker analyses and for MAS salary support.
677 This study has been supported by the ITU BAP project (42053). We are very grateful to
678 Dursun Acar, Asen Sabuncu and Nurettin Yakupoğlu for their help and advice during the
679 core analyses.

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1037 **Figure Captions**

1038 **Fig. 1 (a)** Simplified inset map showing current tectonic setting of Turkey. **(b)** Location of
1039 the Gulf of Saros (SAG-14) and other sites discussed in the text: SL-152 in north Aegean
1040 Sea (Kotthoff et al., 2008), NS-14 in south Aegean Sea (Triantaphyllou et al., 2014), KL-71
1041 in the Sea of Marmara (Sperling et al., 2003), Sofular cave in Black Sea (Fleitmann et al.,
1042 2009). Red arrows and black lines represent surface water circulation from Kontakiotis
1043 (2016) and the North Anatolian Fault (NAF) according to Gasperini et al. (2011),
1044 respectively.

1045 **Fig. 2** Bathymetric map of the Gulf of Saros and surroundings (yellow dots) showing the
1046 location of core SAG-14 (black dot). White arrows represent longshore currents and
1047 anticyclonic eddies in the gulf (Sarı and Çağatay, 2001). Blue lines and black lines indicate
1048 the drainage system surrounding the Gulf of Saros.

1049 **Fig. 3** Generalized lithologic logs of core SAG-14 obtained from the Gulf of Saros and its
1050 radiographic image, showing the main lithostratigraphy of the Holocene sequence. Red stars
1051 indicate the ^{14}C samples with calibrated ages.

1052 **Fig. 4** Age-depth model of the studied core SAG-14 reconstructed based on AMS ^{14}C ages
1053 and Holocene sapropels layers in the cores by using R-studio and the script “CLAM”
1054 (Blaauw, 2010).

1055 **Fig. 5** Magnetic susceptibility (MS), μ -XRF element and elemental profiles (K, Fe, Fe/Mn,
1056 Ca, Ca/Ti and Ca/Sr) for core SAG-14. Note that total organic carbon analysis and biomarker
1057 analysis were also performed in the core.

1058 **Fig. 6.** Comparison of μ -XRF Ca/Ti data from core SAG-14 in the Gulf of Saros with $\delta^{13}\text{C}$
1059 data from Sofular cave in Black Sea (Fleitmann et al., 2009) and arboreal type (AP)
1060 vegetation from north Aegean Sea core SL-152 (Kotthoff et al., 2008).

1061 **Fig. 7.** Comparison of sea surface temperature (SST) profiles from core SAG-14 in the Gulf
1062 of Saros with core SL-152 in the north Aegean Sea (Kotthoff et al., 2008), core NS-14 in the
1063 south Aegean Sea (Triantaphyllou et al., 2014) and core KL-71 in the Sea of Marmara
1064 (Sperling et al., 2003).

1065 **Table Captions**

1066 **Table 1** Core information for this study.

1067 **Table 2** AMS radiocarbon and calibrated ages of foraminiferal shells samples in the studied
1068 core. Radiocarbon ages are converted into calibrated ages by using Marine13 (Reimer et al.,
1069 2013) calibration curve and Calib v7.1 software (Stuiver and Reimer, 1993) with global
1070 reservoir correction of 400 years. A local reservoir age correction ($\Delta R = 149 \pm 30$ years for
1071 the sapropel interval; Facorellis et al., 1998 and $\Delta R = 58 \pm 85$ outside the sapropel Reimer
1072 and McCormac, 2002) was used for the Aegean Sea.

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