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1 Dinocyst assemblages and water surface
2 conditions in the Sea of Marmara during
3 MIS 6 and 5 from two long cores

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17

18 **Abstract**

19 The Sea of Marmara is the connection between the vast Black Sea-Caspian Sea
20 basin (Pontocaspian) and the Global Ocean via the Mediterranean Sea. Its water
21 levels and water conditions has widely varied over times. Combining two cores in the

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22 Sea of Marmara (Turkey) and using organic-walled dinoflagellate cyst assemblages
23 as the main proxy (combined with alkenones, diatoms and benthic foraminifera),
24 allow qualitatively reconstructing water conditions during Marine Isotopic Stage (MIS)
25 6 and 5, such as salinity and oxygen level. A clear main marine phase is illustrated in
26 MIS 5e. A minor marine incursion occurred during MIS 5c, mostly supported by
27 alkenone data. The rest of the record indicates brackish Pontocaspian conditions,
28 with more *Spiniferites inaequalis* in MIS 6 and more *S. cruciformis* in the non-marine
29 parts of MIS 5.

30 At the MIS 6/MIS 5 transition, an earlier initial marine flooding in the Marmara Sea
31 (dinocyst assemblages) in comparison to the Black Sea was highlighted. The marine
32 reconnection occurred at different moments as seen in the terrestrial vegetation
33 reconstructed from pollen analysis linking the two seas.

34 The sapropels of the Marmara Sea form when marine water penetrates at depth
35 from the Aegean Sea beneath a layer of lower salinity water. Variations of the
36 residence time of the marine deep water in a stratified Sea of Marmara are
37 potentially an important factor underlying hypoxia/anoxia and sapropel deposition.

38 When combining surface water proxies with benthic foraminifera (test assemblages
39 and presence of inner organic lining), it appears that the successive MIS 5 sapropels
40 formed under decreasing salinity and oxygen availability conditions as the marine
41 inflow was increasingly restricted. Understanding the hydrologic evolution of the Sea
42 of Marmara during MIS 5 implies taking into account that the Bosphorus and
43 Dardanelle straits are hydrodynamically coupled and may not be simplified as on/off
44 switches based on present day sill depths.

45 ~~A comparison between the two cores suggests water above 290 m never became~~
46 ~~fully anoxic.~~

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47

48 **Key words**

49 Dinocysts, MIS 6 and 5, Marmara Sea, connection, sea level, sapropel

50

51 **1 Introduction**

52 The Sea of Marmara (SoM) is a gateway between the Mediterranean Sea and
53 the Black Sea in Turkey. It has experienced water-level fluctuations and hydrological
54 variations during the late Quaternary. They were mainly controlled by bedrock sills at
55 the Marmara Sea exit of Çanakkale (Dardanelles) Strait and the Black Sea entrance
56 of the Istanbul (Bosphorus) Strait. The variations in water depths of these sills were
57 possibly governed by longer-term vertical displacements related to tectonics and
58 erosion/sedimentation balance due to sea-level changes and water exchanges. This
59 occurred not only between the Black and Aegean seas, but also, at a larger scale,
60 between the Ponto-Caspian ensemble and the Global Ocean ([Yaltrak, 2002](#);
61 [Gökaşan et al., 2008](#); [Çağatay et al., 2009, 2019](#); [Vidal et al., 2010](#); [Eriş et al., 2011](#))
62 ([Fig. 1](#)). The SoM is indeed a highly seismic area as it is located on the western end
63 of the North Anatolian Fault; however, tectonics are mainly significantly affecting sea
64 levels only in the long term. At the scale of the last climatic cycle, global ocean level
65 change, and thus climate, is the main factor controlling the water level in the SoM.
66 For the last interglacial (Marine Isotope Stage 5, MIS 5), decreasing water level
67 maxima are noted along the successive three warm stages: MIS 5e, 5c and 5a
68 ([Rohling et al., 2021](#)). Moreover, the Black Sea level should be considered as
69 another crucial influence on water level in the SoM. Indeed, the melting of the
70 Eurasian icecap and the subsequent evacuation of the freshwater along large rivers
71 flowing into the Black Sea ([Soulet et al., 2013](#); [Tudryn et al., 2016](#)) and, at times, the
72 outflow from the Caspian Sea via the Manych Passage ([Chepalyga, 2007](#)), affected
73 the amount of water transiting through the SoM. The SoM is the outlet of all the
74 Pontocaspian basins (as far as the Tien-Shan via the Syr-Darya and the Chu River
75 and the Pamir-Indu-Kush via the Amu-Darya). Hence this inland sea is the

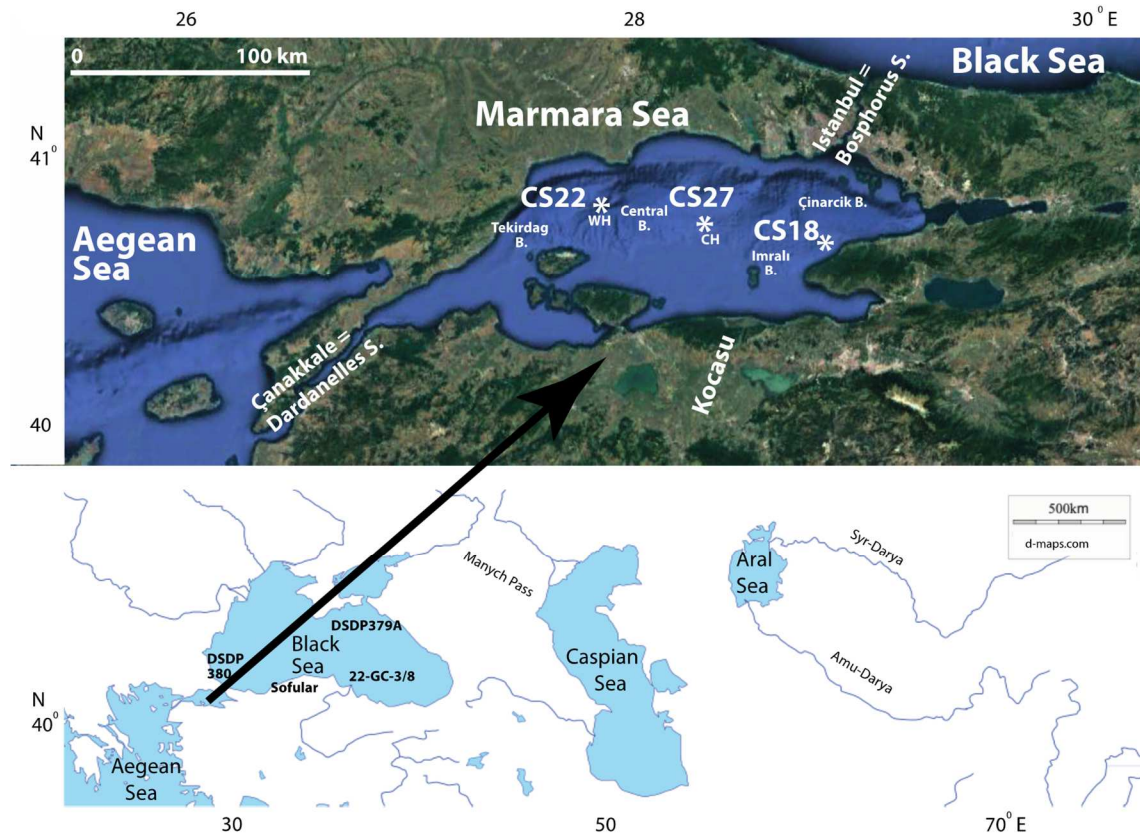
76 receptacle of the history of a large geographical area (Fig. 1). Thus, the water level
77 and water composition history of the SoM is quite complex (local, regional and global
78 forcing) and its insight clearly has an extra-regional relevance. Its history starts to be
79 well-known for the Last Glacial Maximum and the Holocene, but it remains much
80 less clear before (Leroy et al., 2020; Eriş et al., 2007, 2011; Çağatay et al., 2009,
81 2015a, 2019). Moreover, the SoM forms sapropels under certain conditions (Tolun et
82 al., 2002; Sperling, 2003; Vidal et al., 2010). They are dark and finely laminated
83 sediment rich in organic matter and are important to study as they consist of layers
84 where carbon can be locked away from oxidation, thus contributing to the C cycle.
85 The formation of sapropels is diagnostic of low oxygen availability or absence of
86 oxygen at the seafloor, which has been attributed to deep-water stagnation,
87 enhanced biological production or both. The relative importance of variations of
88 these two factors has been widely debated in the Mediterranean Sea (De Lange et
89 al., 2008; Rohling et al., 2015; Zwiép et al., 2018; and references therein). Oxygen
90 isotope records from planktonic foraminifers and other proxies point to a hydrologic
91 driver: surface water freshening causing water column stratification, thus promoting
92 the development of anoxia in the deep basins, and possibly also influencing
93 productivity (Zwiép et al., 2018; Rohling et al., 2015; Grant et al., 2016; Grimm et al.,
94 2013). In the Sea of Marmara (as in the Black Sea) the water body is not stratified
95 during lacustrine stages (Aloisi et al., 2015) and ingress of seawater is considered
96 as the primary cause of stratification and a requirement for sapropel deposition
97 (Çağatay et al., 2009, 2019). However, at the present day, although the Sea of
98 Marmara has a stable stratification (Beşiktepe et al., 1994), the sediment being
99 deposited is not a sapropel. Somehow a paradox, the Holocene sapropel (MSAP1)
100 in the Sea of Marmara occurred at a time when surface water was inferred to be

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101 more saline than at present, although this sapropel is in large part synchronous with
102 Mediterranean sapropel S1 (Sperling, 2003; Vidal et al., 2010).

103 Knowledge of the variations of Black Sea salinity beyond the last deglaciation
104 also remains sketchy. Black Sea fauna were dominantly stenohaline during the
105 peak of the Eemian (or equivalently MIS 5e), bottom water was anoxic, and surface
106 salinity probably higher than in the present day (Zubakov, 1988; Shumilovskikh et al.,
107 2013; Wegwerth et al., 2018). Marine incursions may also have occurred during MIS
108 5a and 5c high stands according to oxygen isotope records obtained on Sofular cave
109 speleothems (Badertscher et al., 2011). These speleothems may reliably inform on
110 the isotopic composition of Black Sea surface waters but are not producing reliable
111 salinity proxies. For instance, during the deglaciation the Black Sea water $\delta^{18}\text{O}$
112 increased from about -10‰ to close to its present-day surface water value (-2.8‰)
113 before reconnection and salinization (Bahr et al., 2006; Soulet et al., 2011).

114



115

116 **Figure 1:** Map of Marmara Sea with core locations (top panel) and map of the seas
 117 that were potentially connected at times (lower panel). CH: Central High, WH:

118 Western High, S: Strait, B: Basin. Cores CS22 and CS18 are the focus of this paper.

119

120 Owing to a major European research programme, long sediment cores (up to
 121 30 m) were taken in the SoM during the MARSITE cruise (Géli et al., 2014) in 2014
 122 to address some of those key problems (<https://cordis.europa.eu/project/id/308417>).

123 The aim of the current investigation is, by using organic-walled dinoflagellate
 124 cyst (herein called dinocyst) analysis combined to other proxies such as sedimentary
 125 content of C37 alkenone, diatoms and benthic foraminifera, to establish the
 126 conditions at the water surface within MIS 5 (130-71 ka ago, Lisiecki and Raymo,
 127 2005) and a large part of MIS 6 (> 130 ka). Thus our results contribute to
 128 reconstruct water-level and surface-water-condition history in the SoM. A sister

129 paper addresses past vegetation, palaeoclimates and chronology on the same
130 samples ([Leroy et al., 2023](#)).

131 **2 Regional setting**

132 The SoM (275 km W-E by 80 km N-S) is located at the western end of the
133 North Anatolian Fault where it splits into several strands; as a result, the sea
134 comprises three deep basins (Tekirdag, Central and Çınarcık) separated by two
135 highs: Western High and Central High ([Fig. 1](#)). Its deepest location is in the Çınarcık
136 and Central Basins at 1270 m. The water depth of the Dardanelles (Çanakkale) Sill
137 is nowadays 65 m, and that of the Bosphorus (Istanbul) Sill is 35 m ([Çağatay et al.,](#)
138 [2009](#)). Rivers are mostly inflowing from the south ([Kazancı et al., 2004](#)). The Imralı
139 Basin in the SE SoM is nowadays a flat area of a water depth only around 300-400
140 m ([Yaltırak, 2002](#)). The Kocasu (or River Koca), the main river flowing into the SoM,
141 exits now from the continent, south of the Imralı Island. At lower water levels, it has
142 formed deltas prograding into the Imralı Basin and has largely contributed to fill it;
143 later on, the sediment was fed into the Çınarcık Basin along a now-submerged
144 canyon ([Kazancı et al., 2004](#); [Sorlien et al., 2012](#)) ([Fig. 1](#)). The region immediately
145 around the SoM is relatively flat, although mountains up to 2500 m are found in the S
146 and SE and 1000 m in the N.

147 Water salinity is 22-24 psu at surface and up to 37-39 psu at depth ([Beşiktepe](#)
148 [et al., 1994](#); [Aydoğdu et al., 2018](#)). The brackish water of the Black Sea (18 psu),
149 being less dense, outflows through the SoM above the deeper denser Mediterranean
150 saline water moving eastwards ([Aydoğdu et al., 2018](#)). The pycnocline is at around
151 25 m water depth. This delicate two-way setting is easily disturbed by water-level
152 changes leading to deep salinity and oxygenation modifications crucial for the

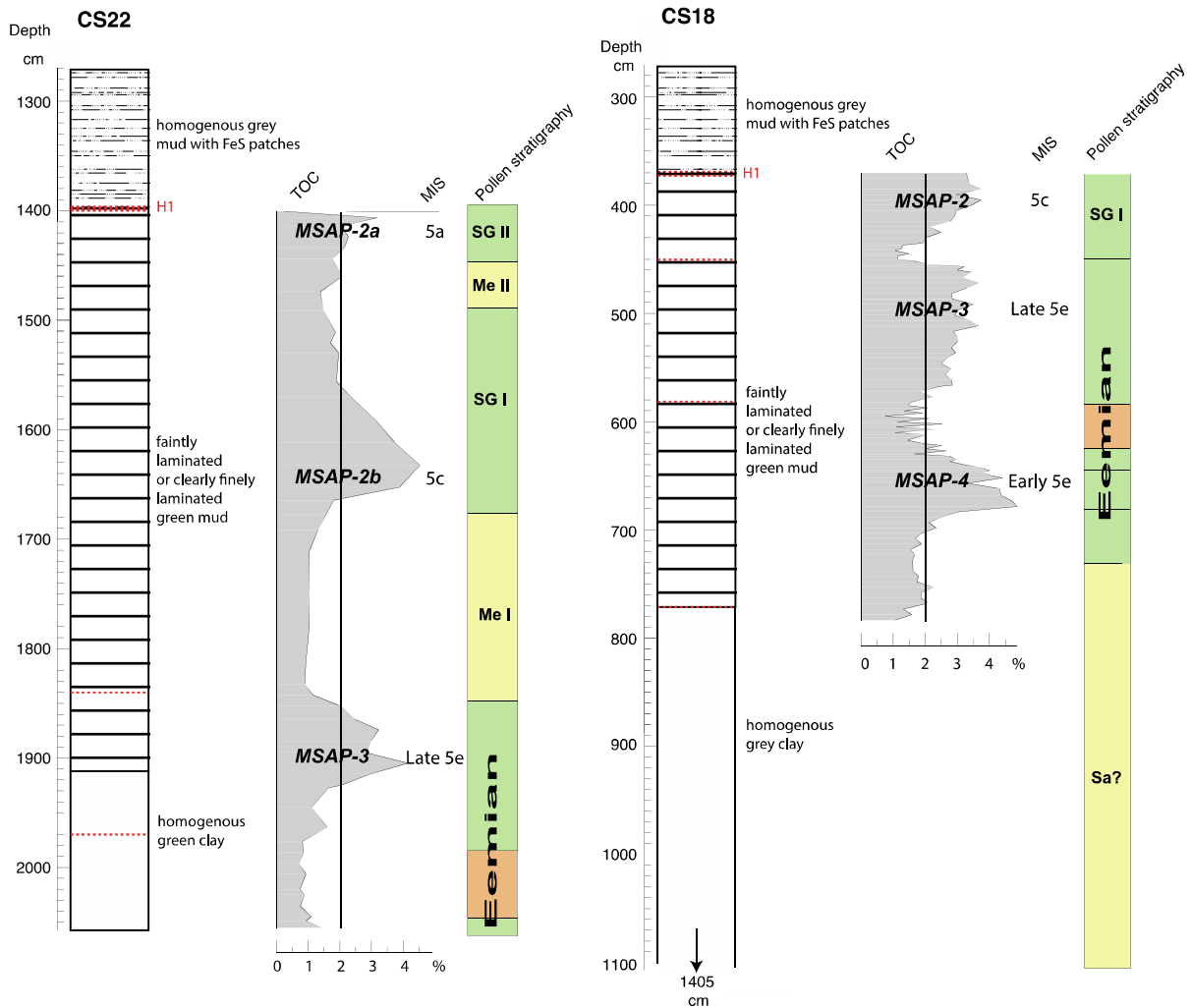
153 survival of aquatic ecosystems. Water temperature fluctuates at most from 4 to 28 °C
154 annually. It displays a slight warming gradient from the NE to the SW: for example,
155 surface water usually varies from 21 (NE) to 26 (SW) °C in summer. All year round,
156 bottom water is relatively stable at c.15 °C (Aydoğdu et al., 2018).

157 The dinocyst assemblages of MIS 6 and 5 have, so far, not been studied in
158 the SoM. However, three dinocyst analyses for these time intervals are available for
159 the Black Sea. In cores DSDP 380 in the SW Black Sea and DSDP Site 379A in the
160 E Black Sea, MIS 5 is covered by a few samples (Ferguson et al., 2018; Hoyle et al.,
161 2021). In the SE Black Sea, a detailed dinocyst diagram of core 22-GC3 includes the
162 period from 129 to 119 ka, with a transition from brackish Pontocaspian to marine
163 conditions at 128 ka (Shumilovskikh et al., 2013) (Fig. 1). More dinocyst records are
164 available for the Late Glacial and Holocene in the SoM (Mudie et al., 2001, 2002,
165 2004; Londeix et al., 2009) where a change from brackish Pontocaspian to marine
166 conditions was observed at c. 9.5-9 ka. A sapropelic layer, Marmara Sapropel 1 or
167 MSAP-1, was identified at 11.5-7 ka ago (Çağatay et al., 2000; Tolun et al., 2002;
168 Vidal et al., 2010). It is worth mentioning the recent study of dinocyst assemblages
169 on a long core in the Gulf of Corinth, covering the MIS 6 and MIS 5, although further
170 afield, due to the similarity of isolation-connection history of its basin (Fatourou et al.,
171 2023).

172 **3 Previous investigations on cores CS22 and CS18**

173 Calypso cores were taken by the vessel “Pourquoi Pas?” in November 2014.
174 Core MRS-**CS22** (20.42 m long) was obtained close to the top of the Western High
175 at 40,83863 latitude N, 27,79906 longitude E and 551 m water depth (Fig. 1). Core
176 MRS-**CS18** (14.25 m long) was acquired from a slope of the Imralı Basin at

177 40,66206 latitude N and 28,87958 longitude E by 291 m water depth. The core
 178 depths of interest are located below a major seismic reflector, i.e. the red-H1
 179 reflector ([Sorlien et al., 2012](#); [Grall et al., 2013](#)).
 180



181
 182 **Figure 2:** Lithology of cores CS22 and CS18. Hiatus shown as red dotted lines. Total
 183 organic carbon in percentage (TOC) with vertical line at 2 % delimiting the sapropels.
 184 MIS: Marine Isotopic Stages. Me I = Melisey I Stadial, SG I = Saint-Germain I
 185 Interstadial, Me II = Melisey II Stadial, SG II = Saint-Germain II Interstadial, Eem =
 186 Eemian Interglacial, according to [Leroy et al. \(2023\)](#).
 187

188 Detailed lithological descriptions for cores CS22 and CS18 and the
189 methodology for Total Organic Carbon (TOC) measurements in core CS22 can be
190 found respectively in [Kende \(2018\)](#) and [Çağatay et al. \(2019\)](#). The sediment is a
191 grey or green clay to silty clay at times clearly finely laminated ([Fig. 2](#)). In core CS18,
192 only rare benthic foraminifera are present in the 770-700 cm interval, but benthic,
193 with some pelagic foraminifera, become abundant in 700-670 cm and 630-580 cm
194 intervals. However, they are absent in the interval 580-430 cm ([Çağatay et al.,](#)
195 [2019](#)).

196 Below the red-H1 reflector, the three sapropels (> 2% TOC) were attributed to
197 MIS 5e, 5c and 5a based on correlation of the μ -XRF Ca with the NGRIP $\delta^{18}\text{O}$, the
198 TOC correlation with regional pollen records (outside of the SoM), oxygen isotope
199 records in the Mediterranean and global sea level. These are for core CS18: MSAP-
200 4 at 699-623 cm, MSAP-3 at 568-455 cm and MSAP-2 at 430-370 cm depth
201 (sapropel numbering proposed by [Çağatay et al., 2009, 2019](#)). These are for core
202 CS22, MSAP-3 at 1920-1853 cm and MSAP-2 in two parts: MSAP-2b at 1653-1570
203 cm and to a lesser extent MSAP-2a at 1460-1407 cm (their naming follows what was
204 proposed in [Çağatay et al. \(2009\)](#)). These authors have proposed a correlation of the
205 MSAP-4, 3 and 2 to the Mediterranean sapropels S5, S4 and S3. A revision of the
206 two Marmara sequence ages, based on geochemistry and pollen analysis, indicates
207 that for core CS22, MSAP-3 is Late Eemian Interglacial, MSAP-2b is Saint-Germain I
208 Interstadial, and MSAP-2a is Saint-Germain II Interstadial, or approximately MIS late
209 5e, 5c and 5a ([Fig. 2; Leroy et al., 2023](#)). In core CS18, MSAP-4 is early Eemian
210 Interglacial, MSAP-3 is late Eemian and MSAP-2 is Saint-Germain I, or
211 approximately MIS early 5e, late 5e and 5c ([Fig. 2; Leroy et al., 2023](#)).

212 Minor and major hiati are defined and discussed in [Leroy et al. \(2023\)](#) and
213 shown as red lines on [figure 2 and following](#).

214 **4 Material and Methods**

215 4.1 Dinocyst analysis

216 Dinoflagellates, a major component of the marine phytoplankton, have a
217 complex life cycle involving a theca that lives in the photic zone and, for about 13 to
218 16% of the species, an organic-walled cyst is produced in the water column, and
219 settling at the bottom of the sea floor for an obligatory dormancy period (e.g. [de](#)
220 [Vernal and Marret, 2007](#); [Bravo and Figueroa, 2014](#); [Mudie et al., 2021](#)). Therefore,
221 the distribution of dinocysts in recent sediments has been linked to surface
222 environmental conditions and this relationship has enabled robust quantitative
223 reconstructions of past sea-surface temperature, salinity, productivity and sea-ice
224 cover duration (e.g. [de Vernal et al., 2020](#)). A study of 181 surface samples in the
225 Pontocaspian region has clearly highlighted that the dinocyst assemblages reflect
226 surface conditions such as salinity, especially from January to March ([Mudie et al.,](#)
227 [2017](#)). Fifty-one sediment samples were treated in core CS22 and 79 in core CS18
228 mostly below the red-H1 reflector for their palynomorph content. The sampling
229 resolution varied between 10 and 20 cm. After measurement of the sediment volume
230 (between 0.5 and 2.5 ml), one tablet of *Lycopodium clavatum* was added. The
231 sediment was then soaked in tetrasodium pyrophosphate, followed by acid attacks:
232 cold HCl (10% dilution), cold HF (between 40 and 60%) and again cold HCl.
233 Samples were then rinsed with distilled water and sieved on nylon meshes at 125
234 and 10 µm. The residues were transferred to vials and slides mounted in glycerol. At
235 least 60 dinocysts were counted in each sample, except in core CS22 above 1400

236 cm (24 to 1124 dinocysts), and in core CS18 above 369 cm (39 to 90 dinocysts) and
237 below 786.5 cm (22 to 274 dinocysts). The sums are given in the diagrams. Some
238 gritty samples in core CS18 were also centrifuged in Sodium Polytungstate at a
239 density of 2.4. All samples were rich in dinocysts, except one at 364 cm depth in
240 core CS18 and one at 2026.7 cm in core CS22, which were barren. All percentages
241 are calculated based on the sum of all dinocysts, except varia (indeterminate,
242 indeterminate and reworked specimens) and foraminiferal organic linings.
243 Reworked specimens are identified by their dark colour, reduced relief and stronger
244 opacity. Concentration of palynomorphs is calculated in ml of wet sediment. The P/D
245 ratio is the ratio of the concentration of pollen on dinocysts. It reflects the
246 continentality of the sample (McCarthy and Mudie, 1998). The H/A ratio is the ratio of
247 heterotrophic (dependent on nutrients synthesised by other organisms) taxa on
248 autotrophic (whose energy source is the light) ones. Zonation (on 23 and 22 dinocyst
249 taxa respectively in cores CS22 and CS18) was made by CONISS after square-root
250 transformation, available in the psimpoll software (Bennett, 2007), which was also
251 used for plotting dinocyst diagrams. The diversity was calculated using the
252 Margalef's species richness (Margalef, 1958) and an ordination analysis with non-
253 Metric Multidimensional Scaling (NMDS) (Primer 7; Clarke and Gorley, 2015)
254 allowing distinguishing groups of samples of similar composition.

255 Identifications were made with the support of the atlas of Mudie et al. (2017),
256 and plates in Marret et al. (2004), Sorrel et al. (2006), Londeix et al. (2009) and
257 Leroy (2010). *Spiniferites cruciformis* specimens were grouped in three different
258 forms (A – standard form, B – pointed apex and C – short and folded processes) as
259 in Marret et al. (2004). *Lingulodinium machaerophorum* ss indicates the sensu stricto
260 form (Leroy et al., 2006; Mertens et al., 2009; Mudie et al., 2017). The “L.

261 *machaerophorum* bulbous” form has short but clearly bulbous (as in [Mudie et al.](#)
262 [\(2017\)](#)) and non-clavate processes. *L. machaerophorum* B is described in [Leroy et](#)
263 [al. \(2006\)](#), as having short microgranulate processes with a large striated conical
264 base. *Impagidinium* sp. A is a form defined in [Londeix et al. \(2009\)](#). The name
265 *Spiniferites inaequalis* is conservatively used, although some ongoing discussions
266 propose to move the species to the genus *Impagidinium* ([Mertens et al., 2018](#)). In
267 the cumulative diagram, the marine taxa are on the left and the brackish
268 Pontocaspian ones on the right. In the middle, are the heterotrophic taxa and the
269 various forms of *Lingulodinium machaerophorum*. Warm-loving dinocysts are defined
270 by the presence of *Operculodinium israelianum*, *Spiniferites pachydermus*,
271 *Polysphaeridinium zoharyi*, *Tuberculodinium vancampoae* and *Spiniferites mirabilis*
272 as they reflect temperatures higher than nowadays ([Marret and Zonneveld, 2003](#);
273 [Londeix et al., 2009](#); [Shumilovskikh et al., 2013](#); [de Vernal et al., 2020](#); [Marret et al.,](#)
274 [2020](#)). Seasonal contrast in sea-surface temperature (SST) is reflected by
275 *Bitectatodinium tepikiense*, *Pentapharsodinium dalei* and *Scrippsiella trifida* ([Head et](#)
276 [al., 2006](#); [Mudie et al., 2017](#)).

277 Although not dinocysts, coiled organic linings of foraminifera were found and
278 added to this investigation, rather than to the pollen diagram ([Leroy et al., 2023](#)), as
279 they usually belong to the marine environments. The inner organic lining is often
280 produced by benthic foraminifera ([de Vernal, 2009](#); [Mudie et al., 2021](#)).

281 4.2 Other proxies in core CS22

282 **C37 Alkenones** were measured in 26 sediment samples from core CS22
283 between 2050 and 1370 cm. After freeze-drying and grinding, samples were
284 extracted using an accelerated solvent extraction system (ASE350, Thermo
285 Scientific). After a clean-up procedure using silica gel column chromatography with a

286 mixture of dichloromethane: hexane (1:1), the total lipid extract was analysed by gas
287 chromatography and flame ionization detection (GC-FID, Trace GC, Thermo
288 Scientific). Identification and purity of alkenones were checked in several samples by
289 gas chromatography coupled to quadrupole mass spectrometry (Trace GC DSQII,
290 Thermo Scientific). GC conditions were similar to those described by [Sonzogni et al.](#)
291 [\(1997\)](#). Long-chain alkenones are mainly synthesized by marine haptophyte algae
292 [\(Marlowe et al., 1984\)](#) and can be used as an indicator of marine environment in
293 the Sea of Marmara. In the last glacial cycle, no alkenones were detected in
294 sediments of the Sea of Marmara before its connection with the Mediterranean Sea
295 [\(Vidal et al., 2010\)](#).

296 **Benthic foraminiferal faunas** were quantitatively investigated in the 125 µm
297 fraction from nine samples. At first a qualitative downcore check of foraminiferal
298 content was performed in the coarse fraction at a 1 m sampling resolution, then
299 refined to 20 cm between 2035 and 1915 cm where very rich benthic foraminifera
300 assemblages were found. Two additional samples were investigated at 1632 and
301 1631 cm. When possible, 300 specimens were counted in sediment aliquots, and
302 identified to the lowest taxonomic level possible, mainly at the species level. To
303 describe major faunal trends, the species Benthic Foraminiferal Number (BFN,
304 expressed as individuals per g or ind.g⁻¹) and frequency data of main species (>1%)
305 were used.

306 Sediment samples for **diatom and chrysophyte** analyses were collected at
307 20 cm intervals in each section of core CS22. Smear slides were prepared with a
308 suspension made of known amounts of dry sediment and distilled water. 200 µl of
309 this suspension were spread onto a cover slip, air-dried and mounted on a glass
310 slide using Naphrax mounting medium. Diatom valves were counted along transects

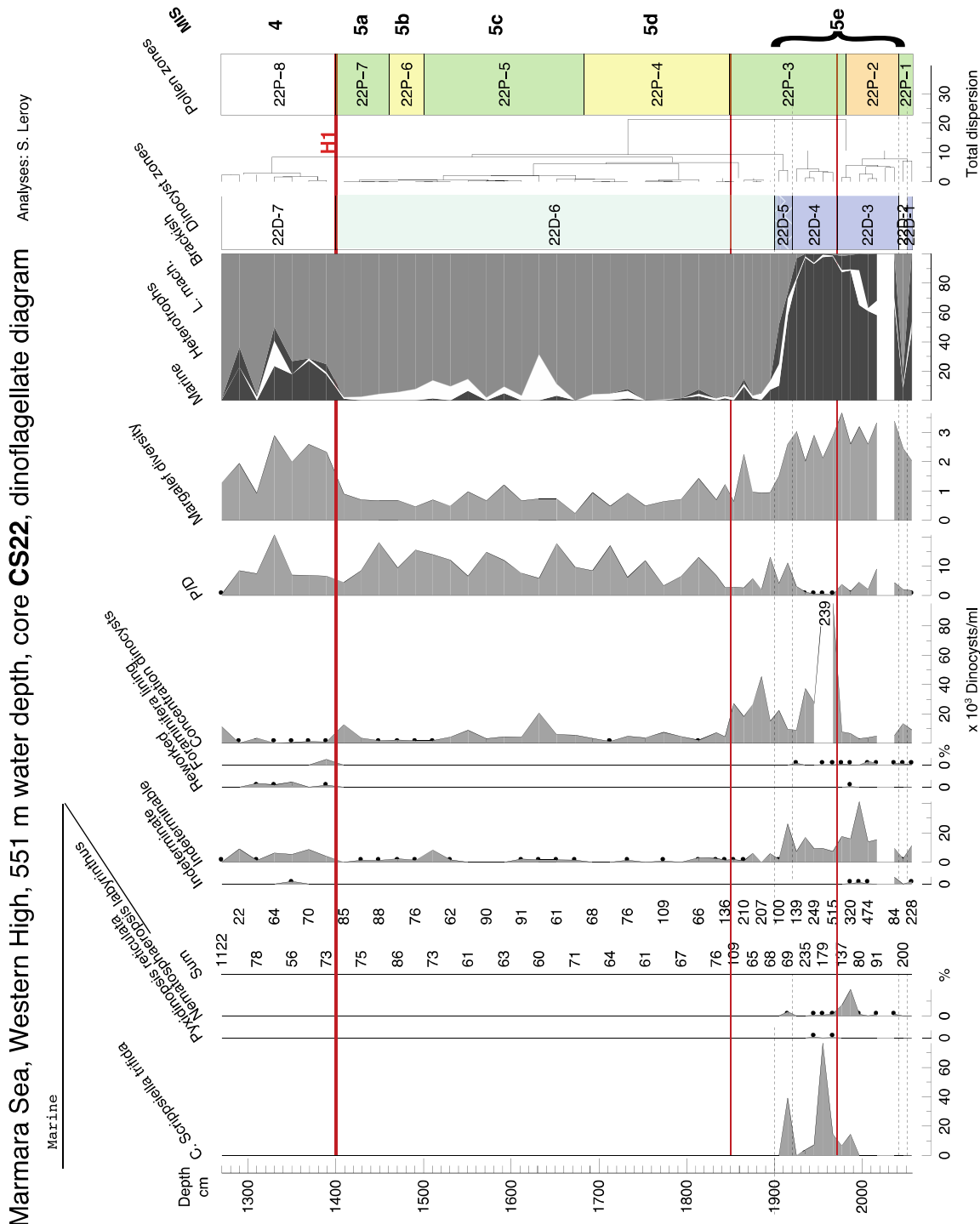
311 and expressed as number of valves per gram of dry sediment. Chrysophycean cysts
312 were also counted; but morphotypes were not identified.

313 **5 Results**

314 Two main types of dinocyst assemblages were observed ([Fig. 3 and 4](#)). The
315 first one is largely dominated by *Spiniferites cruciformis* and *Pyxidinosia psilata*. By
316 comparison to Black, Marmara and Caspian seas Lateglacial-early Holocene
317 assemblages, they are attributed to brackish Pontocaspian conditions, perhaps in
318 the range of ~7-12 psu, thus clearly lower than the surface conditions of the SoM
319 ([Bradley et al., 2012](#); [Marret et al., 2004, 2009, 2019](#)). The second assemblage,
320 called here marine, is an assemblage with diverse taxa also with no direct equivalent
321 in the present. It represents salinities probably close to the surface waters of the
322 modern-day Marmara ([Balkis et al., 2016](#)).

323 5.1 Dinocysts in core CS22

324



327

328

329 **Figure 3 a and b:** Percentage dinocyst diagram of core CS22. Black dots for < 5%.
 330 Pollen zones to the right with chronological attributions. Horizontal brackets indicate
 331 the warm indicators in the marine group. In the cumulative diagram, the marine taxa

332 are on the left and the brackish Pontocaspian ones on the right. In the middle, are
333 the heterotrophic taxa and the various forms of *Lingulodinium machaerophorum*.

334

335 The dinocyst assemblages show a diversity from 3 to 19 taxa, with notably
336 higher numbers of taxa at the bottom of the core, i.e. below 1920 cm (Fig. 3).

337 Broadly, the record shows a succession of two main assemblages, one mostly
338 dominated by marine taxa, and the following one by brackish Pontocaspian taxa.

339 The main change in the diagram according to the CONISS analysis is by far between
340 zones 22D-4 and 5 at 1920 cm depth (Fig. 3). However, it is not sharp.

341

342 **Zones 22D-1 and D-2** each consist of one sample (at 2057 and 2047 cm depth). In
343 rapid alternation, the first assemblage is marine with a dominance of *L.*

344 *machaerophorum* alongside: *Spiniferites* sp., *S. mirabilis*, cysts of

345 *Pentapharsodinium dalei* and *Tuberculodinium vancampoeae*. The second sample

346 has a brackish Pontocaspian assemblage, with the dominance of *Spiniferites*

347 *cruciformis* A and *Pyxidinoopsis psilata* and low occurrence of *L. machaerophorum*.

348 Foraminifera linings are observed in zones D-1 to D-4.

349

350 **Zone 22D-3** (2042-1972 cm depth): Besides zone 22D-1, *L. machaerophorum*

351 displays its highest values here, declining only in the last third of this zone. The

352 percentages of *Operculodinium centrocarpum* increase progressively to a maximum

353 at the end of this zone. *Spiniferites* sp. also has maximal values. The other taxa,

354 mainly marine, are rather diverse (high Margalef diversity index). A high number of

355 indeterminate cysts occur. The sample at 2027 cm depth is rich in pollen, but hardly

356 contains any dinocysts.

357

358 **Zone 22D-4** (1972-1920 cm depth): The assemblage is still marine and diverse, but
359 rather different from the previous zone. The spectra are alternatively dominated by
360 cysts of *P. dalei* and cysts of *Scrippsiella trifida*. *L. machaerophorum* is present but in
361 small amounts. The P/D ratio is at its minimum, largely because the concentration in
362 dinocyst is very high, i.e. >230,000 cysts/ml. The number of indeterminable cysts is
363 decreasing but still significant.

364

365 **Zone 22D-5** (1920-1900 cm depth): This zone is a period of transition with the
366 percentages of *S. cruciformis* A progressively increasing, and those of *P. psilata*
367 starting a continuous curve, while the marine forms are less diverse (decreasing
368 Margalef diversity index). *L. machaerophorum* makes a brief return. Brown baggy
369 cysts form a bell-shape curve. *Operculodinium centrocarpum* has nearly
370 disappeared. Cysts of *S. trifida* peak in one of the samples.

371

372 **Zone 22D-6** (1900-1399 cm depth): This very long (5 m) and rather homogenous
373 (minimal Margalef diversity index) zone is largely dominated by brackish
374 Pontocaspian taxa such as *S. cruciformis* and *P. psilata*. Brown baggy cysts are
375 occasionally more frequent, especially at 1632 cm depth (32%). *Bitectatodinium*
376 *tepinkense* is only sub-continuously present in the middle of this zone between 1652
377 and 1551 cm depth. *L. machaerophorum* is occasionally present at the beginning of
378 this zone and soon disappears. All through this zone, the P/D ratio is relatively high.
379 Cyst concentration drops drastically at the very beginning of this zone and remains
380 low for the rest of the sequence.

381

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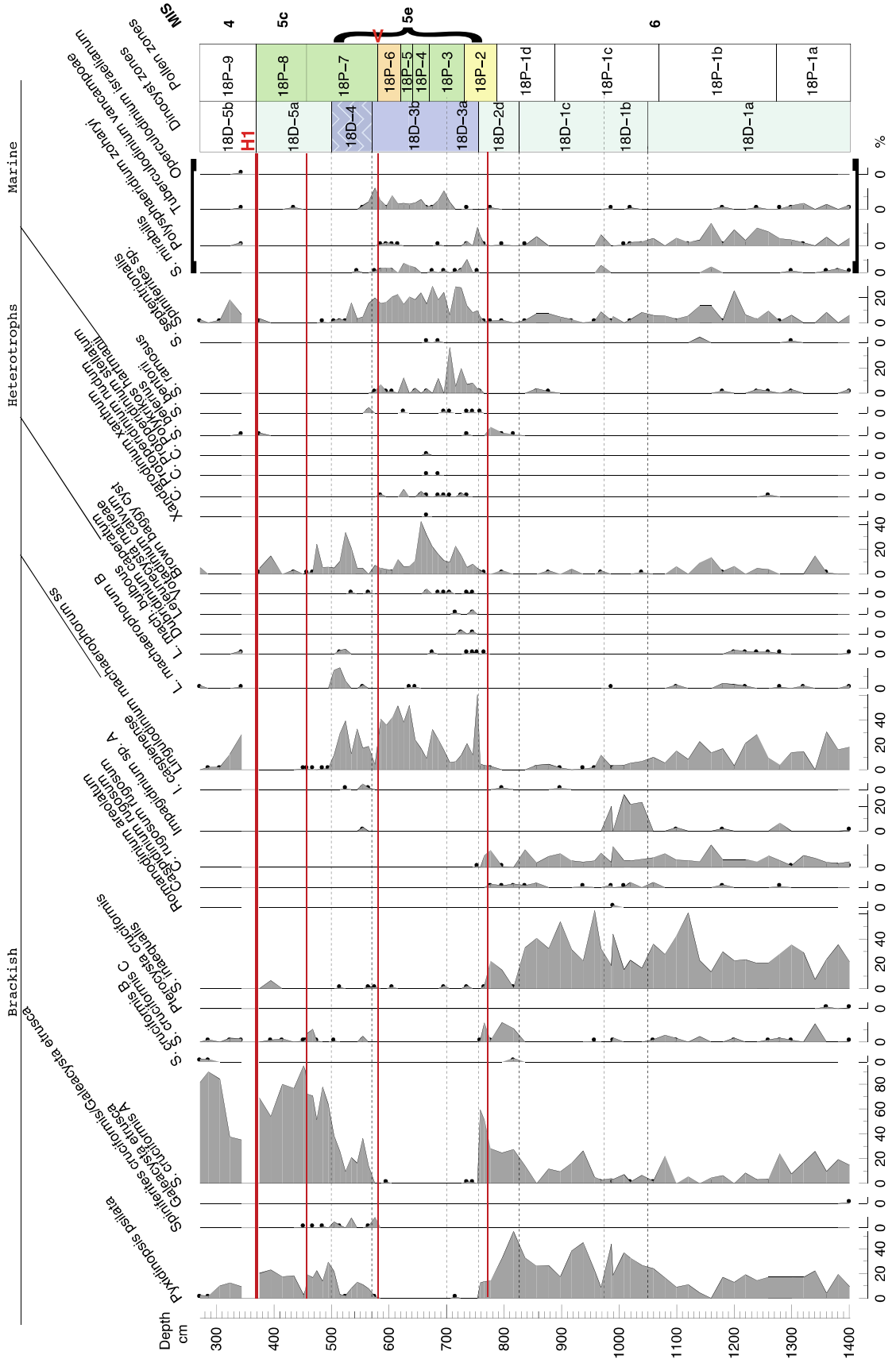
382 **Zone 22D-7** (1399-1270 cm depth): The values of *S. cruciformis* remain the same as
383 in the preceding zone, but those of *P. psilata* significantly drop. The assemblage is
384 diverse with many marine forms but with a low occurrence. *L. machaerophorum* is
385 back with irregular presence between 0.2 and 14 %. Three foraminifera linings (4%)
386 were found at the base of this zone. Concentration is varying, at times below 500
387 dinocysts/ml. Reworked elements are frequent.

388 5.2 Dinocysts in core CS18

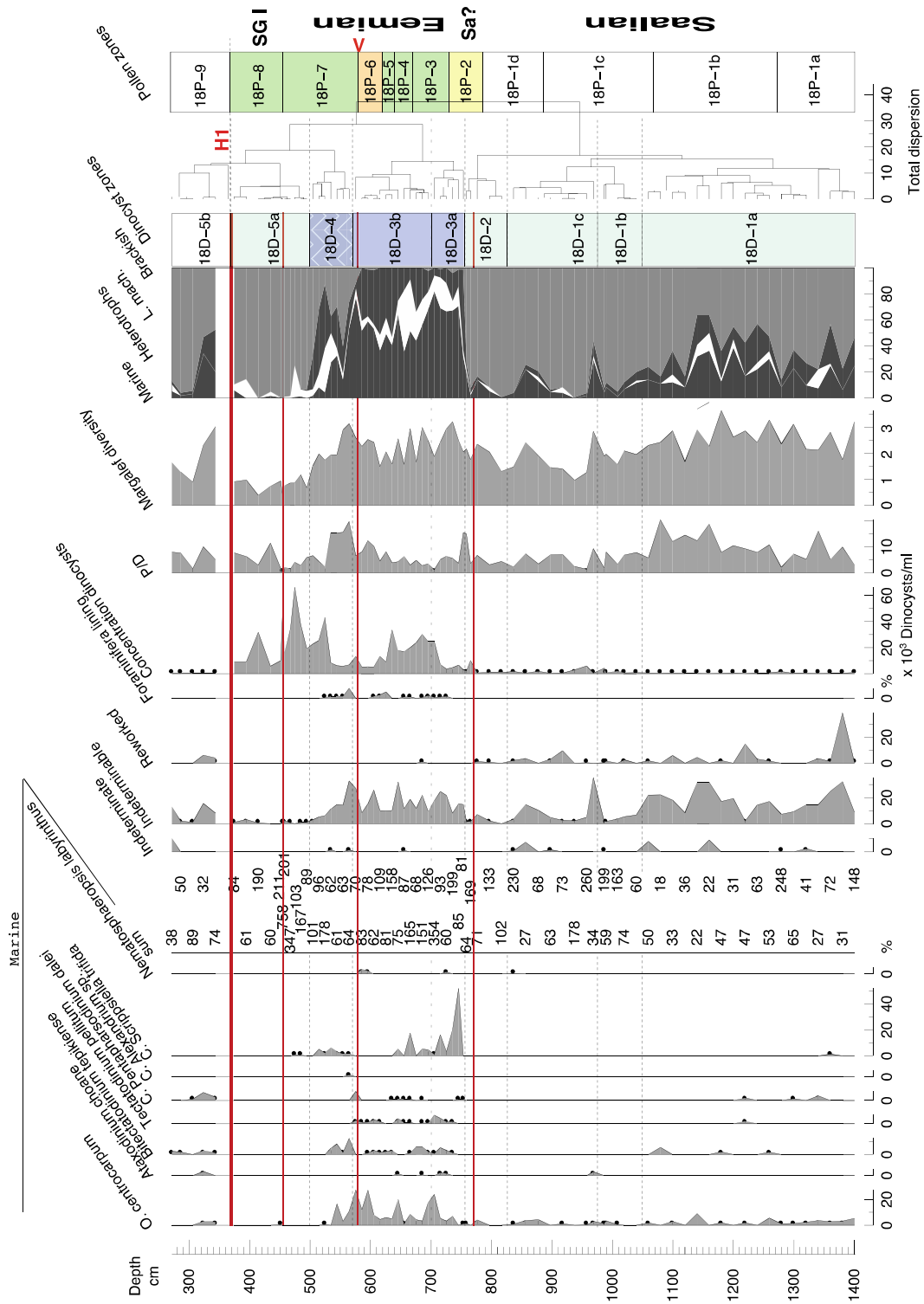
389

Marmara Sea, Imrali basin, 291 m water depth, core CS18, dinoflagellate diagram

Analyses: S. Leroy



Marmara Sea, Imrali basin, 291 m water depth, core CS18, dinoflagellate diagram
Analyses: S. Leroy



391

392

393 **Figure 4 a and b:** Percentage dinocyst diagram of core CS18. Black dots for < 5%.

394 Pollen zones to the right with chronological attributions. Horizontal brackets indicate

395 the warm indicators in the marine group. In the cumulative diagram, the marine taxa

396 are on the left and the brackish Pontocaspian ones on the right. In the middle, are
397 the heterotrophic taxa and the various forms of *Lingulodinium machaerophorum*.

398

399 The dinocyst assemblages show a diversity from 3 to 17 taxa, with notably higher
400 numbers of taxa below 1180 cm, and at 735 and 565 cm (Fig. 4). Broadly, the record
401 shows the succession of three main assemblages, one mostly dominated by marine
402 taxa in the middle, with below and above it two brackish Pontocaspian assemblages
403 of two different types. The two main changes in this diagram are at 756 and 570 cm
404 depth (Fig. 4).

405

406 **Zone 18D-1**, 1400.5 – 826.5 cm depth: This is a very long zone of nearly 6 m. It is
407 dominated by brackish Pontocaspian taxa such as *P. psilata*, *S. cruciformis* A, *S.*
408 *inaequalis* and *Caspidinium rugosum rugosum*, with a significant occurrence of *L.*
409 *machaerophorum* ss. In general, concentrations are relatively low, mostly below
410 1000 cysts/ml. Reworked cysts are frequent. This zone has been subdivided into
411 three subzones, with subdivisions as followed. Subzone 18D-1a (1400.5-1049.5 cm)
412 is dominated by *S. inaequalis*. *L. machaerophorum* ss, *P. psilata* and *S. cruciformis*
413 A are abundant. A range of marine and brackish Pontocaspian taxa occur side by
414 side. Indeterminable dinocysts are abundant. A large peak of reworked cysts occurs
415 at 1381 cm depth. The P/D ratio remains high all through this zone, with maximal
416 values towards the end, i.e. a maximum of up to above 20. Concentrations are very
417 low, often lower than 500 cysts per ml. Subzone 18D-1b (1049.5-977 cm): This zone
418 is characterised by a bell-shaped curve of *Impagidinium* sp. A. *S. inaequalis* is still
419 abundant. *P. psilata* displays increasing values across this zone. In subzone 18D-1c

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420 (977-826.5 cm), *P. psilata* reaches maximal values, i.e. 54%. The percentages of *S.*
421 *inaequalis* are still high. Less diverse marine forms are observed.

422

423 **Zone 18D-2**, 826.5-756 cm depth: This zone is characterised by many changes. *P.*
424 *psilata* decreases and *S. cruciformis* A increases up to a brief maximum of 59%. *S.*
425 *cruciformis* C is very abundant. A small bell-shaped curve of *S. belerius* occurs. *S.*
426 *inaequalis* values drop. Very few marine taxa subsist. Concentrations are slightly
427 increasing. Before the end of this zone a hiatus occurs at 772 cm. Finally, the limit
428 18D-2 and 18D-3 at 756 cm is recognised as one of the strongest statistical changes
429 according to CONISS.

430

431 **Zone 18D-3**, 756-570 cm depth, is a totally different zone, with *L. machaerophorum*
432 (up to 52%) and *Spiniferites* sp. (up to 30%) widely dominant. Brown baggy cysts are
433 very abundant in this zone and the following one, i.e. up to 30%. Many marine taxa
434 are present. *S. ramosus*, *O. centrocarpum* and *Tuberculodinium vancampoae* are
435 well represented. Concentrations are high, often around 10,000-20,000 dinocysts/ml.
436 A split into subzones 3a and 3b at 700.5 cm depth is based on a maximum of *S.*
437 *ramosus*, very large peak of cysts of *S. trifida* (52%), medium-high concentration and
438 maximum percentage of marine taxa in subzone 3a; whilst in subzone 3b high *T.*
439 *vancampoae* and high concentration occur. This zone ends with the second
440 strongest diagram change (as calculated by CONISS). Foraminiferal linings are only
441 present in this zone and in the following zone.

442

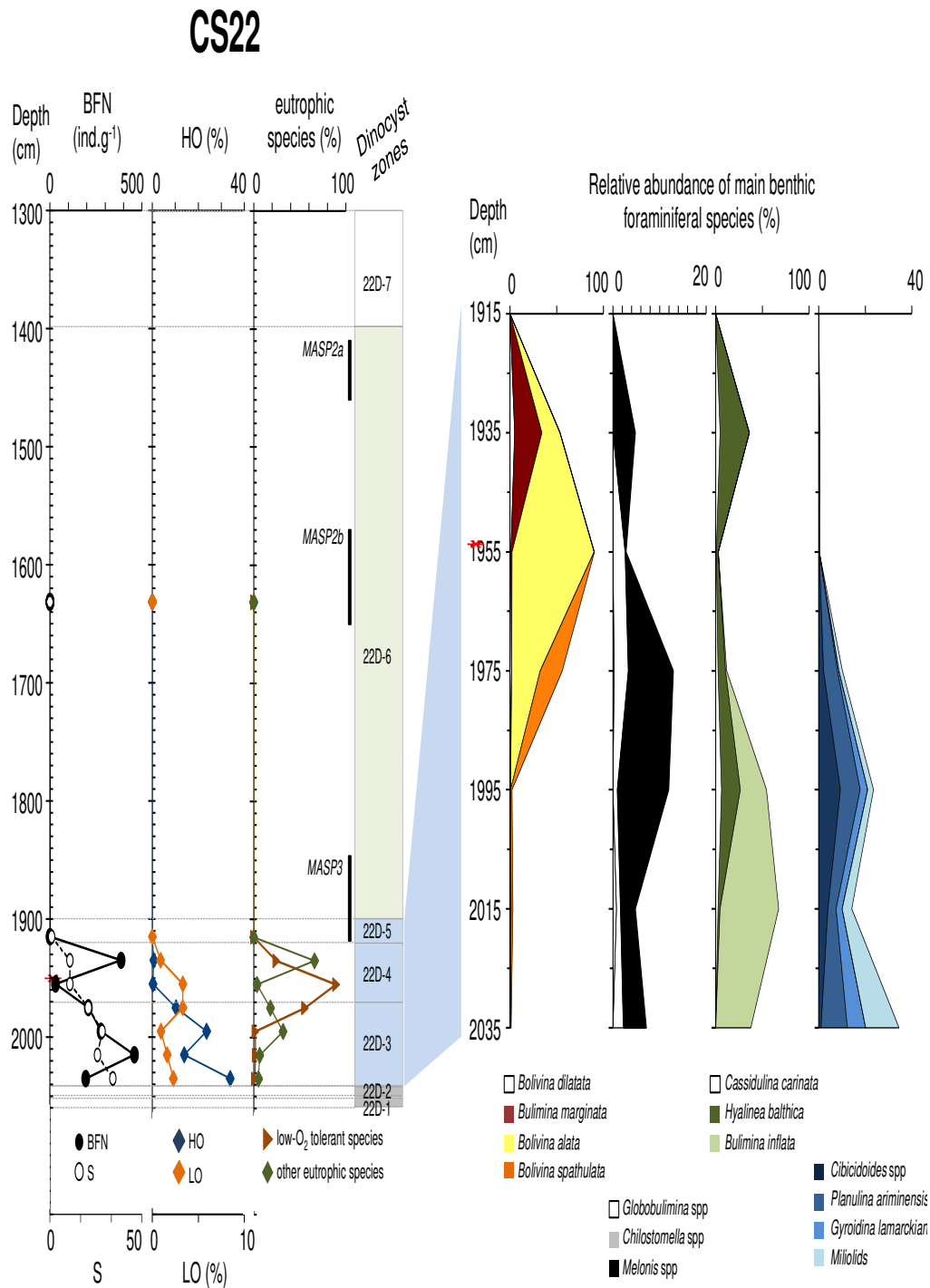
443 **Zone 18D-4**, 570–499.5 cm depth, appears as a transition, with mixed marine and
444 brackish Pontocaspian assemblages. *P. psilata* and *S. cruciformis*, quasi absent in

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445 zone 18D-3, reappear. *L. machaerophorum* (with a peak of other forms of *L.*
446 *machaerophorum* towards the end of this zone) and brown baggy cysts are still
447 abundant. Marine taxa remain present, but their values decrease. Concentrations,
448 after moderate values, become high again. P/D is maximal, i.e. up to just below 20.
449

450 **Zone 18D-5**, 499.5-271.5 cm, is split in two by the seismic red-H1 reflector;
451 moreover, a barren sample occurs a few cm above this stratigraphic hiatus. In
452 subzone 18D-5a, 499.5-369 cm depth, *S. cruciformis* A very largely dominates with
453 percentages up to 95%. Besides *P. psilata* and brown baggy cysts, hardly any other
454 taxa are present. It is noteworthy that *L. machaerophorum* has only rare occurrences
455 in this subzone, and only at its base. This is a very low diversity subzone (minimum
456 in the Margalef index). Concentrations, after being very high, become high, i.e.
457 66,000 to 10,000 dinocysts/ml. Subzone 18D-5b, 369-271.5 cm depth: The
458 assemblages are similar to subzone 18D-5a (dominance of brackish Pontocaspian
459 taxa), but brown baggy cysts have quasi disappeared and are replaced by *L.*
460 *machaerophorum* that have re-appeared. A moderate range of marine taxa is
461 present. Concentrations are very low as in subzones 18D-1a. Reworked elements
462 are common in the lower half of this subzone, just above the barren sample.

463 5.3 Other proxies in core CS22



464

465 **Figure 5:** Benthic foraminiferal record of core CS22 with dinozones and Marmara
 466 sapropels (MSAP). Right panel: relative abundance of main species (>3%) between
 467 2035 and 1915 cm depth. Left panel: Benthic Foraminiferal Number, BFN (ind.g⁻¹),

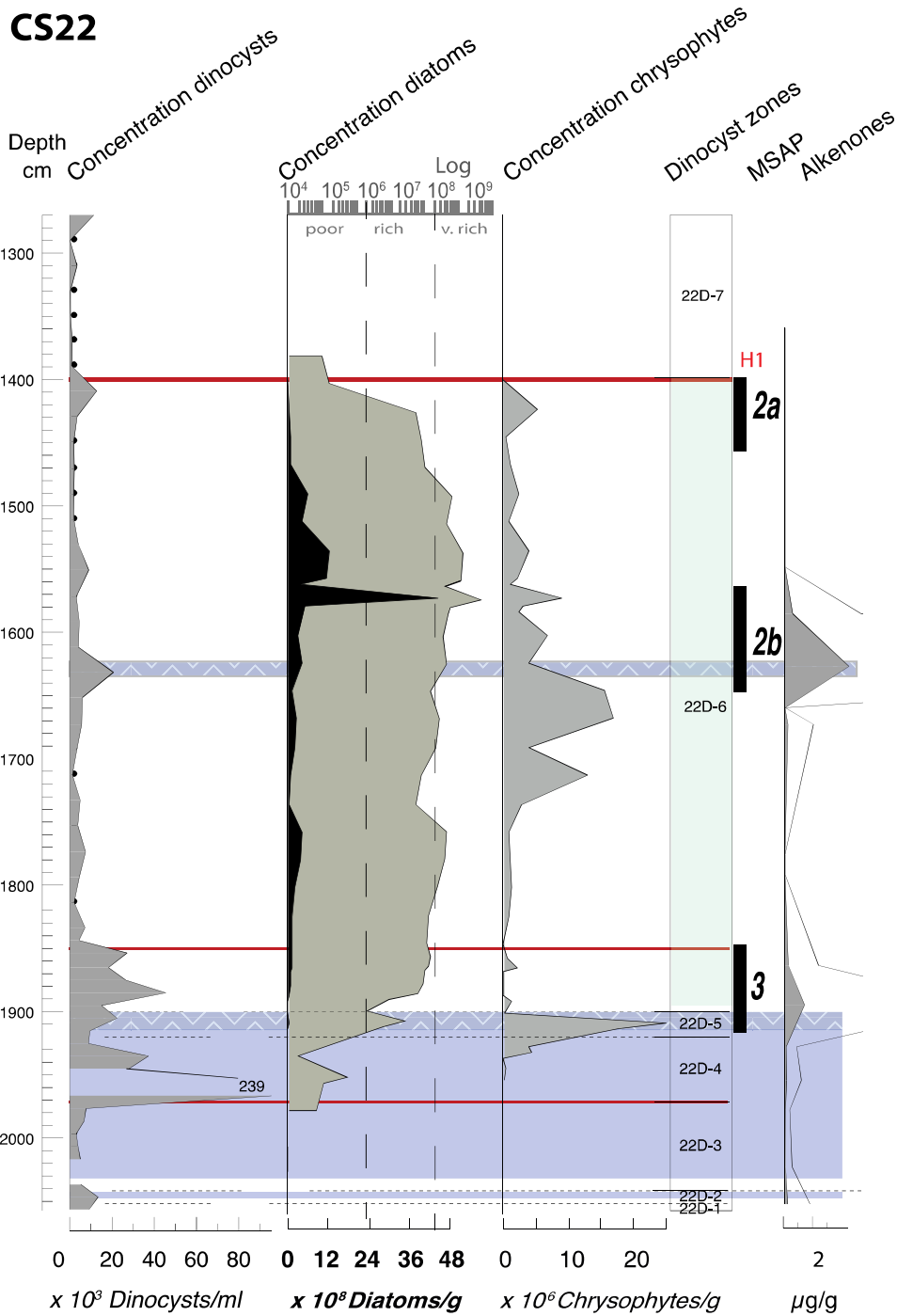
468 species richness S, relative abundance of ecological groups of species. HO =
469 species adapted to high-oxygen level, LO = low-oxygen indicators, eutrophic species
470 = species adapted to high organic carbon fluxes. For further explanation on
471 ecological assignment of species, the reader is referred to the discussion and [Table](#)
472 [SI2](#).

473

474 Quantitative investigation of benthic foraminiferal assemblages relies on the
475 recovery of a well-preserved, abundant (164 to 361 specimens) and diverse (>60
476 species overall) faunal content between 2035 and 1915 cm depth ([Table SI1](#)). Apart
477 from a few individuals (< 10) belonging to neritic taxa (e.g. *Ammonia*, *Discorbinella*,
478 *Elphidium*), assumed to be post-depositional, foraminiferal assemblages
479 overwhelmingly consist of species from marginal to bathyal environments. The BFN
480 ranges from 30 to 459 ind.g⁻¹, with lowest values recorded at 1955 cm. Dominant
481 species (>5%) belong to *Bolivina* spp, *Bulimina* spp, *Cassidulina carinata*,
482 *Cibicidoides* spp, *Hyalinea balthica* and *Melonis* spp and indicate fully marine
483 conditions at the seafloor at the base of the core. The foraminiferal record, though
484 short, exhibits a marked species shift between 1995 and 1975 cm ([Fig. 5](#)). From
485 2035 to 1995 cm, species number is highest, ranging from 31 to 41. Faunas are
486 strongly dominated by *Bulimina inflata* (28-63%), together with various associated
487 species including *Cibicidoides* spp (1-9%), *Gyroidina lamarckiana* (3-8%), *Planulina*
488 *ariminensis* (3-11%) and miliolids (3-14%). Minor species (<3%) include *Bulimina*
489 *aculeata*, *Uvigerina mediterranea* and *U. peregrina*. From 1975 up to 1935 cm, most
490 of the afore-mentioned species disappear from the record. Dominant species consist
491 mainly in *Bolivina alata*, *B. spathulata* and *Bulimina marginata*, which collectively
492 account for 49 to 90% of the assemblage. Concomitant to this faunal shift, species

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493 number drops from 25 to 13 (Fig. 5). By contrast, several other species such as
494 *Cassidulina carinata*, *Hyalinea balthica*, *Melonis affinis* and *M. pompiloides* exhibit no
495 clear upcore trend. Present along the whole record, these species show punctual
496 increases, as especially marked for *H. balthica* at 1995 (32%) and 1935 (20%) cm.
497 Sample at 1915 cm is considered barren of benthic foraminifera as fourteen
498 reworked individuals only were found. Samples at 1632 and 1630 cm were devoid of
499 any foraminifera.



500

501 **Figure 6:** Palaeoproductivity: concentrations of dinocysts, diatoms (in decimal and
 502 log scales) and chrysophytes, with dinozones, C37 alkenones (thin black line curve
 503 for 10 x exaggeration) and Marmara sapropels (MSAP) in core CS22. The
 504 concentration of C37 alkenones in $\mu\text{g/g}$ are based on dry weight sediment. The
 505 second marine phase at 1633-1632 cm depth is indicated by a box with a blue
 506 zigzag highlight.

507

508 Diatom preservation is becoming better from bottom to the red-H1 reflector.
509 Hardly any diatoms were found below 1915 cm (Fig. 6). Then only some
510 occurrences are noted between 1915 and 1866 cm. From 1854 cm upwards and
511 until 1470 m, the diatom concentrations are high to very high. Diatom concentrations
512 are highest between 1591 and 1491 cm depth. Chrysophycean cysts are abundant
513 when diatom concentrations decrease. The diatom assemblages are composed of a
514 mixture of fresh and brackish-water species. Dominant *Stephanodiscus* species (*S.*
515 *minutulus*, *S. parvus*, *S. medius*, *S. neoastrea*, *S. hantzschii*) occurred with
516 numerous fragments of *Entomoneis calixasini*, a fossil species so far only found in
517 core sediments of the SoM (Pailles et al., 2014). The first occurrence of *E. calixasini*
518 in core CS22 is recorded at 1834 cm depth. Spores of a marine planktonic diatom,
519 *Chaetoceros*, are observed from 1850 to 1830 cm.

520 The values of alkenones are usually below detection limit, except at 2050-
521 1865 cm and 1633-1592 cm, reaching respective maxima of 1.3 µg/g at 1896 cm
522 and 4.4 µg/g at 1633 cm depth (Fig. 6).

523 **6 Interpretation and discussion**

524 6.1 Core CS22: taphonomy, surface water condition and chronology

525 6.1.1 Taphonomy

526 *Below the sapropel*

527 At 2027 cm in zone 22D-3, the sample is unexpectedly barren in dinocysts. It
528 is one of the samples of zone 22P-2 with abundant *Pinus* pollen (76%) (Leroy et al.,
529 2023). The lithology shows a bioturbated sediment. Sediment disturbance may also

530 contribute to explain the rapidly changing spectra in the short zones 22-D1 and D2.
531 The sharp zone limit at 1920 cm (between zones 22D-4 and 5) corresponds to a very
532 thin silt layer and a change from bioturbated to unbioturbated sediment, that is the
533 start of the sapropel attributed to MSAP-3 (1920-1853 cm).

534

535 *In the sapropel and until the red-H1 seismic reflector*

536 Dinocysts (as well as pollen) are well preserved. Their concentrations are high
537 in all sapropels as observed in other sapropels ([Cheddadi and Rossignol, 1995](#);
538 [Kotthoff et al., 2008](#)) ([Fig. 6](#)).

539

540 *Impact of red-H1 seismic reflector*

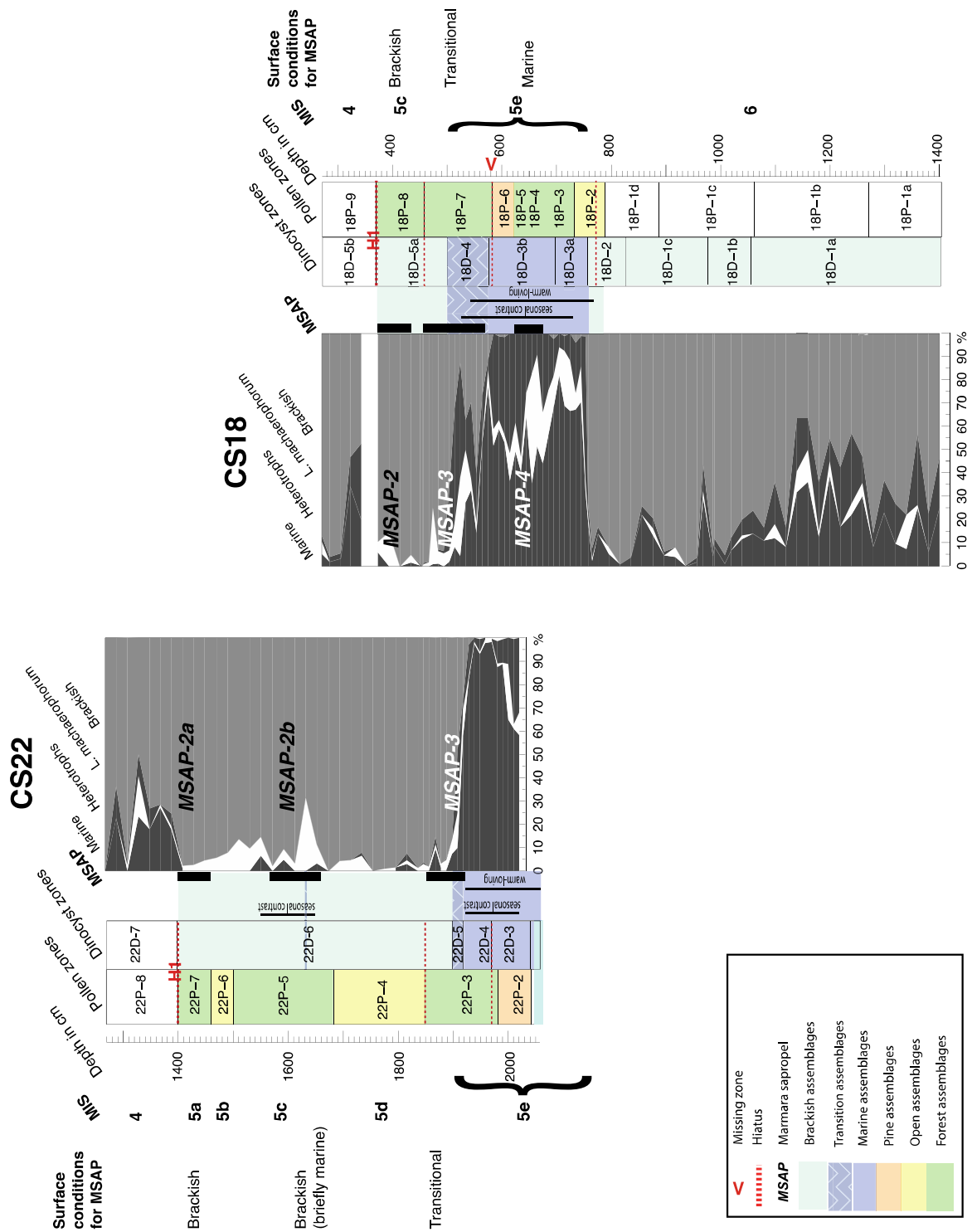
541 The red-H1 reflector corresponds to a widespread erosion event in the Sea of
542 Marmara that causes hiatuses of spatially variable duration across the transition from
543 MIS 5 to 4 ([Grall et al., 2014](#); [Çağatay et al., 2019](#)). The sediment deposited after the
544 hiatus is dated as MIS 4 based on magnetic paleo-intensities (in core CS 22; [Kende,](#)
545 [2018](#)) and on tephra correlations (in cores CS18 and CS 27; [Wulf et al., 2021](#)). With
546 the red-H1 reflector, a long-term change in taphonomy is initiated in the SoM. It is
547 seen in the reworking of older marine layers and of soils. Firstly, some foraminifera
548 (organic linings) are brought in the sediment. Secondly, the non-pollen
549 palynomorphs *Glomus* (a fungal spore) is present, due to inwash of soil erosion
550 products ([Leroy et al., 2023](#)). This reflects a turning point in the aquatic conditions of
551 the SoM.

552 6.1.2 Sea-surface and bottom conditions

553 The depths of 2042-1920 cm (dinocyst zones 22D-3 and 4) are fully marine.

554 This is clearly seen not only in the dinocyst assemblages, but also in the presence of

555 alkenone and in the benthic foraminiferal content (Fig. 3, 5 and 7).



556

557 **Figure 7:** The cumulative diagrams sorting dinocyst groups/taxa by salinity. The four
558 groups are the marine ones (black) to the left, the brackish Pontocaspian dinocysts
559 (grey) to the right), and in the middle the heterotrophic cysts (white) and the
560 *Lingulodinium machaerophorum* forms (black). Dinocyst zones: blue highlight for
561 marine, light blue for brackish Pontocaspian. The distribution of warm-loving and
562 seasonal contrast indicators as vertical black lines. Pollen zones: green highlight for
563 high arboreal taxa (minus *Pinus*) zones and yellow for low arboreal taxa % and high
564 *Pinus* % zones.

565

566 Five indicators of warm sea-surface temperature (SST) are recorded: *O.*
567 *israelianum*, *S. pachydermus*, *P. zoharyi*, *T. vancampoae* and *S. mirabilis* (Fig. 3 and
568 7). They are especially abundant in zone 22-D3. Their occurrence suggests
569 temperatures higher than nowadays. *B. tepikiense*, occurring in zones 22-D3 and 4,
570 reflects marine conditions with seasonal amplitudes of SST. Thus, conditions
571 recorded in zone 22-D3 suggest a warm climate with seasonal contrast.

572 In zone 22D-4 at 1955 cm, high values of cysts of *P. dalei* and *S. trifida*
573 suggest an increase in temperature seasonality and lower salinity (e.g. Head et al.,
574 2006; Mudie et al., 2017). In contrast, very high concentrations point at surface
575 waters rich in nutrients and possibly bottom waters with low oxygen (Mudie et al.,
576 2017). This may be due to alternating deep-water upwelling - when the heavy saline
577 marine waters penetrate the bottom of the basin from the Aegean Sea - and
578 stratification. This occurs below the horizon with high TOC (sapropel MSAP-3 at
579 1920-1853 cm depth and bottom anoxia, Fig. 2).

580 To tackle environmental changes at the seafloor, the approach of Cornuault et
581 al. (2016, and references therein) was followed and foraminiferal species were

582 gathered into three functional groups assumed to be representative of contrasted
583 oxygen and trophic conditions at the seafloor: (1) species adapted to high oxygen
584 conditions (HO), (2) low-oxygen indicators (LO), and (3) species adapted to
585 eutrophic conditions ([Fig. 5](#)). Ecological assignment is given in [Table SI2](#). The major
586 faunal shift observed at 1995 -1955 cm suggests contrasted oxygen conditions
587 between zones 22D-3 and 22D-4, possibly related to the establishment of eutrophic
588 conditions. In zone 22D-3, benthic foraminiferal assemblages are ecologically
589 diverse and consist of a mix of epibenthic and various endobenthic species. This
590 sequence and the high abundance of HO species suggest a mesotrophic
591 environment where well-oxygenated conditions prevailed at the seafloor. By contrast,
592 HO indicators disappearance in zone 22D-4 together with high abundance of species
593 adapted to high organic fluxes points to reduced oxygen level during this phase.
594 Assemblages from this zone are successively dominated by species adapted to
595 contrasted oxygen conditions, suggesting variable oxygenation level during zone
596 22D-4. In particular, the BFN decrease, high dominance of eutrophic species tolerant
597 to low-oxygenation and slight increase of LO species suggest dysoxic conditions at
598 the beginning of this period at 1935 cm, high dominance of species adapted to
599 moderate oxygen depletion strongly suggest environmental changes recorded at the
600 seafloor directly are related to eutrophication rather than ventilation changes.
601 Overall, foraminiferal faunas preceding MSAP-3 deposition exhibit strong similarity
602 with assemblages reported in organic-rich layers from Holocene and Late Glacial
603 archives in the SoM, similarly dominated by various Bolivids and Buliminids, among
604 other species (e.g. [Çağatay et al., 2000](#); [Kaminski et al., 2002](#); [Kirci-Elmas, 2008](#)).
605 Our benthic record thus matches well the establishment of water column stratification

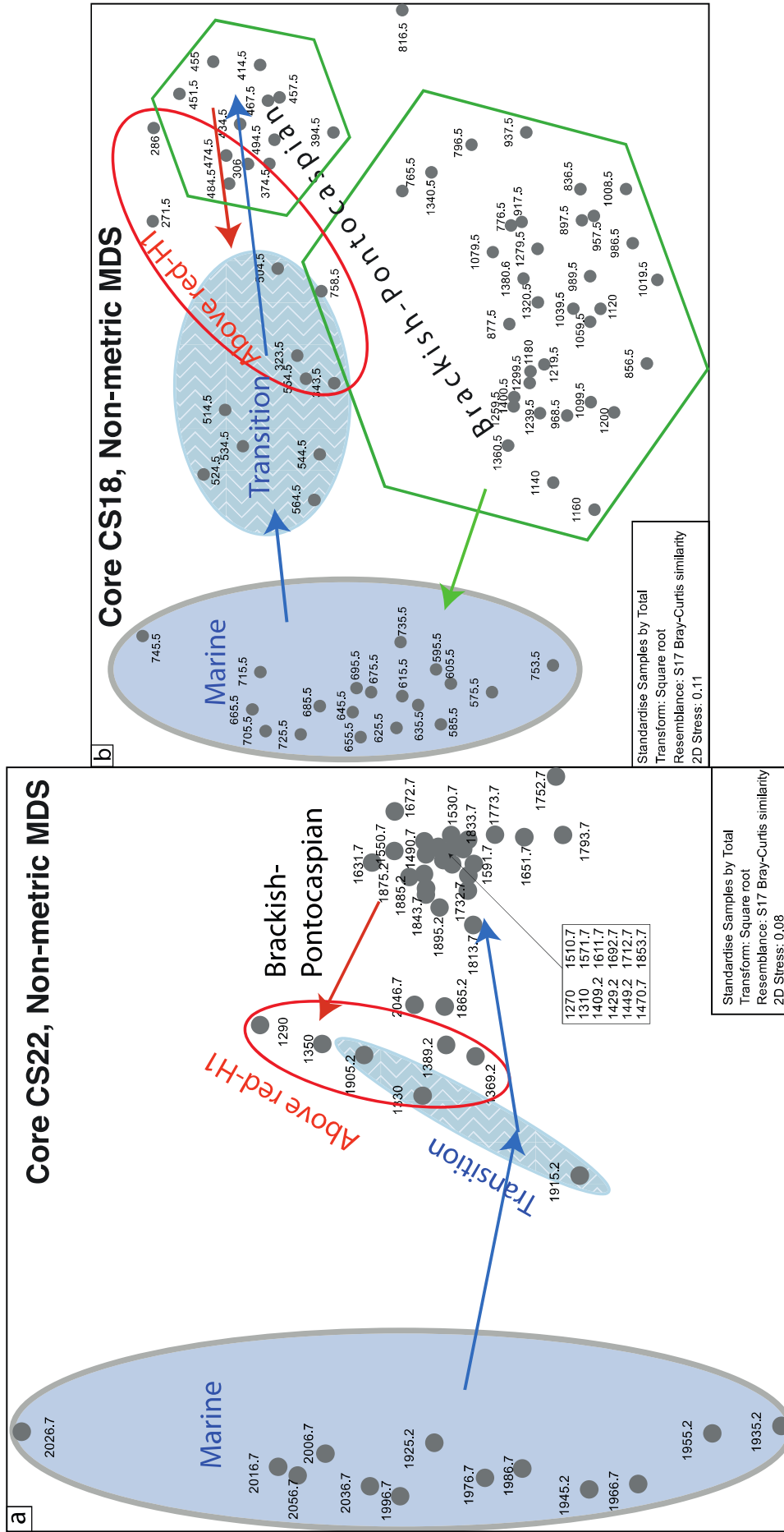
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606 and bottom water hypoxia prior to sapropelic deposition, and indicates anoxic
607 conditions do not prevail at the seafloor in zones 22D-3 and 22D-4.

608 In the dinocyst diagram (Fig. 3), the P/D ratio, informing on the intensity of
609 terrestrial influence on the aquatic realm, is extremely low in zone 22-D4. It is low in
610 general below the hiatus at 1845-1842 cm depth. It is relatively high above it with
611 values frequently higher than 10. It is therefore suggested that the water level was
612 the highest in zone 22-D4 (1972-1920 cm). This is just before the high TOC values
613 linked to the MSAP-3 sapropel.

614

615



617 **Figure 8:** Results of the NMDS analysis on dinocyst percentages from cores CS22
618 (a) and CS18 (b). Depth in cm.

619

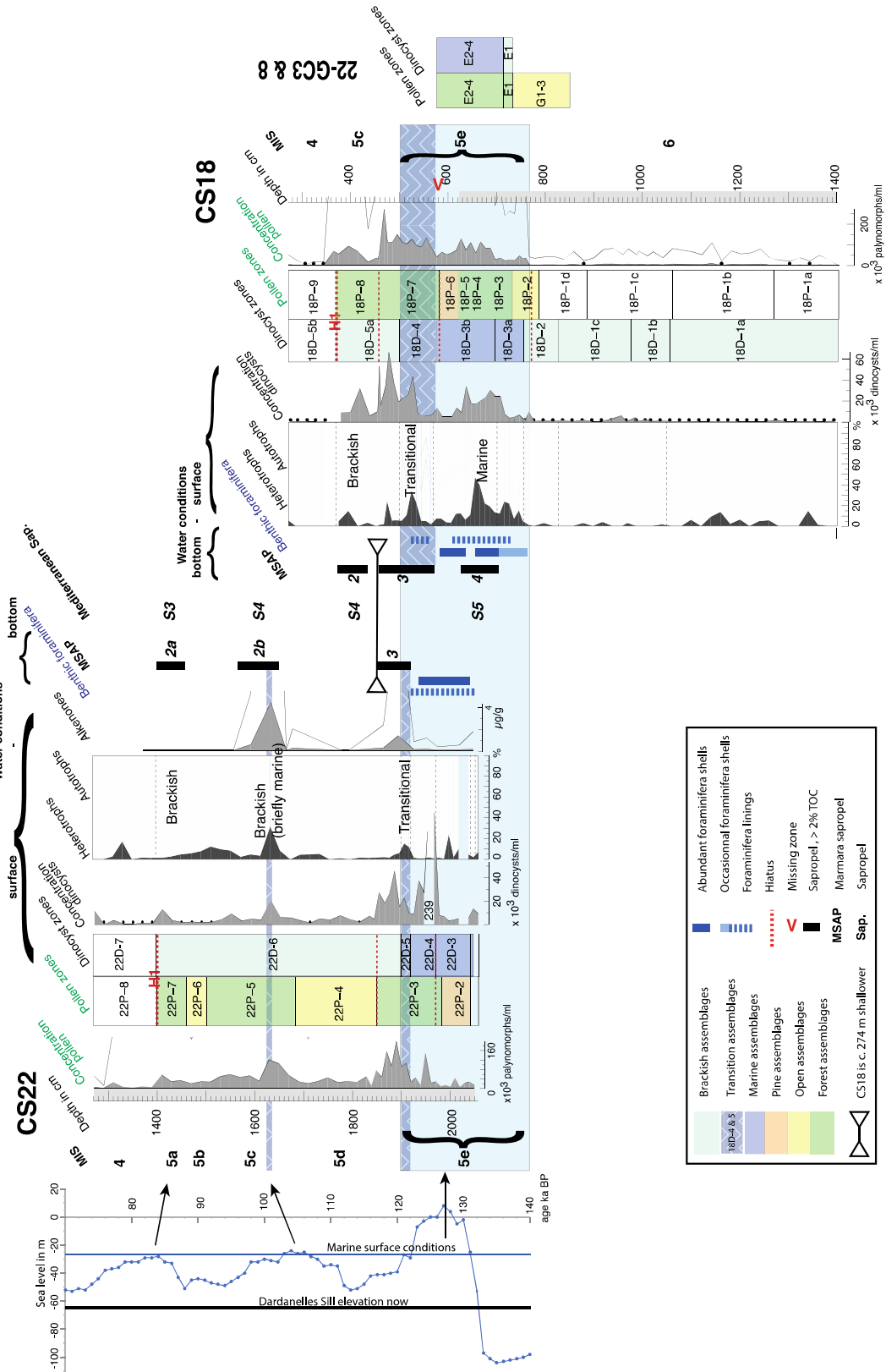
620 Zone 22D-5 (1920-1900 cm) is a transition zone towards brackish
621 Pontocaspian conditions (also seen in the NMDS results in [figure 8](#)). The
622 disappearance of foraminiferal shells at 1915 cm is attributed to bottom anoxic
623 conditions in the subsequent MSAP-3 (1920-1853 cm).

624 Then zone 22D-6 (1900-1399 cm) is clearly brackish Pontocaspian, as seen
625 in dinocysts and in the diatom species that indicate fresh to brackish conditions.
626 Interestingly, marine *Chaetoceros* spores occur at 1850-1830 cm. The highest
627 concentrations of Chrysophycean cysts indicating fresher / more oligotrophic
628 conditions occurred between 1680 and 1611 cm depth when diatom concentration
629 decreases. Later, between 1591 and 1491 cm, the highest diatom concentrations
630 recorded suggest increased seasonal productivity.

Black Sea

Sea of Marmara

Global Ocean level



631

632

633

634

Figure 9: Water conditions. At the surface, dinocyst concentration and the ratio heterotrophic and autotrophic dinocysts (H/A). In the middle for core CS22, also the alkenones. For the bottom: Marmara sapropels (MSAP, >2% TOC), benthic

635 foraminifera shells and benthic foraminifera organic lining. The top of the marine
636 horizon in core CS18 (hiatus between zones 18P-7 and 8) is c. 274 m deeper than
637 that in core CS22 (hiatus between zones 22P-3 and 4). Horizontal bold V for pollen
638 and dinocyst zones absent in core CS18. For linking with the sister paper on pollen
639 ([Leroy et al., 2023](#)), pollen zones and pollen concentrations are also shown. By
640 combining the two light grey vertical boxes, the most complete record is obtained.
641 Sea-level curve from [Rohling et al. \(2021\)](#) with the position of the Dardanelles Sill
642 elevation now (-65 m) and the sea-level threshold where transition from brackish to
643 marine conditions occurred. The pollen and dinocyst records of the Black Sea cores
644 22-GC3 & 8 are shown in a simplified way. The records of these cores and of core
645 CS18 are aligned in two ways. First the beginnings of the interglacial (transition
646 yellow to green box infills) should be synchronous and secondly the Black Sea
647 dinocyst sequence should stop before the return of brackish-Pontocaspian dinocysts
648 in Marmara core CS18. However this long 22D-6 zone is not homogenous. At 1632
649 cm, a peak of heterotrophic dinocyst is observed. It seems to correspond to the
650 second peak in alkenones, indicating marine influence, at 1633 cm ([Fig. 9](#)).
651 According to pollen analysis, it is in an interstadial (Saint-Germain I). Thus, a brief
652 marine incursion is probable. It is located at the beginning of sapropel MSAP-2b at
653 1653-1570 cm.

654 Zone 22D-7 (above 1399 cm, above the red-H1 seismic reflector) suffers from
655 taphonomical problems ([Leroy et al., 2023](#)). Nevertheless, it is mainly brackish-
656 Pontocaspian, but the occurrence of marine taxa can be either attributed to limited
657 marine influences or more likely to reworking (see above).

6.1.3 Elements of chronology

It is suggested that the high sea level of MIS 5e (Bard et al., 1996; Koeling et al., 2009) is represented here by zones 22-D3 to 5. In the whole of MIS 5, it is mainly in the maximum sea level of MIS 5e that the Black Sea would have been connected to the Global Ocean. At that time, the two Marmara straits would have the two-way exchange of water as today. It was established earlier that stratigraphic succession immediately above the red-H1 reflector belongs to MIS 4 (Grall et al., 2014; Kende, 2018; Çağatay et al., 2019), which can now be shown to reflect a brackish-Pontocaspian environment. During MIS 4, the SoM was an oligotrophic lake, with no evidence of water body stratification, and was almost certainly disconnected. During the late phases of MIS 5, the inflow of Mediterranean water was restricted and this probably required a Dardanelles Sill depth higher than the present sill depth. We will come back to this point later in the discussion and here only consider the phasing of the marine to brackish transition. In the age model proposed in Çağatay et al. (2019), this transition occurs after MIS 5d (in the 110 to 100 ka interval) at a time of sea-level rise, and that may be considered unlikely. This transition more probably occurred concurrently with sea-level drop either later in MIS 5c or during the transition from MIS 5e to 5d. Assigning this transition to later in MIS 5c would keep the sapropel correlation proposed by Çağatay et al. (2019) but imply that MIS 5d is not observed in the pollen record in either core (Leroy et al., 2023).

6.2 Core CS18: taphonomy, surface water condition and chronology

6.2.1 Taphonomy

Before the transition 18D-2 to 3 (756 cm) towards marine assemblages, thus already at the beginning of pollen zone 18P-2 (786.5-730.5 cm), one may see

682 sudden changes. This is firstly a sediment change from homogenous to finely
683 laminated around 772 cm. Two round pebbles (of river origin?) are found at 775 and
684 758.5 cm. In between them, one finds a 2 mm-thick silt layer (773 cm). Moreover, at
685 771 cm a sharp increase of the concentration in pollen and spores (from <5000 to >
686 25,000 palynomorphs/ml) and at 762 cm a sharp and long-lasting reduction of the
687 reworked pollen. The sharp statistical limit at 756 cm in the dinocyst diagram is 17
688 cm above the 2 mm-thick silt layer. These observations are in support of an increase
689 of water level with the influx of marine water from the Aegean Sea ([Çağatay et al.,
690 2009](#)) leading to sediment disturbance and taphonomical changes.

691 The sharp dinocyst limit at 570 cm is slightly above the hiatus at 580 cm (one
692 sample). It is interesting to note here the absence of the equivalent to zone 22D-4
693 with its very high percentages of cysts of *P. dalei* and *S. trifida*. The sediment in core
694 CS 18 was probably lost by erosion linked to this hiatus.

695 In general, dinocysts and pollen are well preserved in the three sapropels and
696 their concentrations are maximal, as observed in core CS22. Minimal concentrations
697 are reached in MIS6 and 4.

698 After the red-H1 hiatus, a barren sample (neither pollen, nor dinocysts, but
699 many *Dreissena* shell debris) is noted. In the following sample, a peak of *L.*
700 *machaerophorum*, a large peak of reworked pollen, significant presence of reworked
701 dinocysts and abundance of *Glomus* co-occur. Therefore, no marine invasion can be
702 reconstructed; to the contrary a strong coastal, even soil, input should be
703 recognised. The sediment taphonomy changes completely.

704 6.2.2 Sea-surface conditions

705 The transition from brackish Pontocaspian to fully marine conditions (zones
706 18D-2 to 3, main change highlighted by CONISS) seems to occur very abruptly at

707 756 cm depth (Fig. 4 and 7). Then in zone 18D-3, the conditions are only fully
708 marine. Zone 18D-4 is a transitional period, back towards a brackish-Pontocaspian
709 environment. This transition zone is also well marked in the NMDS results (see
710 green arrow in Fig. 8). The surface waters of subzone 18D-5a are clearly brackish-
711 Pontocaspian. Subzone 18D-5b is mostly brackish-Pontocaspian but some
712 influences from the sea are felt, although as in core CS22 a taphonomical impact is
713 very important. Below 756 cm, the assemblages of zones 18D-1 and 2 are brackish-
714 Pontocaspian, with significant reworking in zone 18D-1 both in dinocysts and in
715 pollen.

716 The group of dinocysts reflecting warm SST (Fig. 4 and 7) is especially
717 abundant in zone 18-D3. The presence of *B. tepikiense*, as in the CS22 core,
718 indicates seasonal temperature contrast at that time.

719 Two large peaks of *L. machaerophorum* are noted just above a hiatus, i.e. at
720 the base of zone 18D-3 and at the base of zone 18D-5b. After the transition from
721 zone 18D-2 to 3, a peak of cysts of *S. trifida* is also present. Both taxa are signs of
722 nutrient enhancement.

723 The P/D ratio fluctuates throughout the record. One may note however high
724 values in subzone 18-D1a and at the base of zone 18-D4. The values are especially
725 low in the middle of zone 18D-3, suggesting high water levels.

726 6.2.3 Elements of chronology

727 Zones 18D-3 (marine taxa and low P/D) and 18D-4 (transition) reflect the MIS
728 5e high sea levels. The warm-loving taxa are present in the same zone and confirm
729 interglacial conditions as deduced from pollen analysis. However, it is likely that this
730 zone is not complete as a characteristic dinozone in core CS22 (high percentages of
731 cysts of *P. dalei* and cysts of *S. trifida*) is missing (Leroy et al., 2023).

732 Zones D18-1 and 2 are suggested to represent low water levels and, if the
733 sequence is in continuity, glacial stage MIS 6. As a reminder, above the red-H1
734 hiatus, brackish conditions have previously been related to MIS 4 ([Kende, 2018](#);
735 [Çağatay et al., 2019](#)).

736 In conclusion, the chronology proposed here represents an in-depth revision
737 of that in [Çağatay et al. \(2019\)](#), with especially the occurrence of a larger part of MIS
738 5e and of MIS 5c and fits the chronology based on geochemistry and pollen analysis
739 proposed in [Leroy et al. \(2023\)](#).

740 6.3 Comparison of the dinocyst records within the Sea of Marmara and to the 741 Black Sea

742 6.3.1 Diversity

743 The dinocyst assemblages are similar in core CS22 and core CS18, with
744 slightly more marine taxa in CS22 and brackish-Pontocaspian taxa in CS18,
745 reflecting the difference in the extension of glacial versus interglacial periods covered
746 in the two sequences ([Table SI3](#)). The total number of taxa is larger in core CS18;
747 the latter because of the long MIS 6 record. Taxa at play in core 22-GC3 in the SE
748 Black Sea ([Shumilovskikh et al., 2013](#)) are similar to those in the two Marmara
749 cores, although the Black Sea seems to have more heterotrophic cyst diversity and
750 the SoM more brackish-Pontocaspian taxon diversity ([Table SI3](#)). Modern and
751 Holocene records across the Black Sea show geographic differences in the
752 distribution of heterotrophic taxa ([Mudie et al., 2017](#)). This may be due to variations
753 in water turbidity, limitations to photosynthesis, or possibly competition with other
754 phytoplankton groups. Remarkably, *S. inaequalis* and *S. trifida* are two particularities
755 of the SoM versus the Black Sea.

756 6.3.2 MIS 5e

757 In core 22-GC3 (SE Black Sea), the transition from brackish Pontocaspian to
758 marine conditions during MIS 5e is dated at 128 ka ([Shumilovskikh et al., 2013](#)). It is
759 marked by an increase in productivity due to rising temperature (increased in
760 dinocyst concentration) and by the inflow of marine waters from the Mediterranean
761 Sea (increase in the taxon number, especially those from typical marine salinities).
762 The change is progressive contrary to that in core CS18, underlining the possibility of
763 a hiatus in core CS18 at the beginning of the marine phase (see the earlier
764 taphonomy section) linked to flooding by marine waters. The subsequent peak of
765 nutrients reconstructed from the dinocyst analysis in core CS18 is possibly due to a
766 combination of shore erosion as sea level rose and increasing temperature. The very
767 high percentages of cysts of *P. dalei* and *S. trifida* in zone 22D-4 find no equivalent
768 in core CS18. The transition from marine to brackish (mixed assemblages) is made
769 over 20 cm only within zone 22D-5, whereas it is much longer in core CS18 (zone
770 18D-4, 70.5 cm long). These two observations suggest other significant hiatus and
771 deeper waters in the west that were invaded quickly by saline waters.

772 In the middle of the marine phase, a *Pinus* zone (22P-2, 2042-1982 cm depth,
773 and 18P-6, 621-581 cm depth) has been recognised in both cores and quantitative
774 climatic reconstruction based on pollen proposed a cool-humid climate not
775 suggestive of a stadial steppe ([Leroy et al., 2023](#)). Due to the presence of the warm-
776 loving dinocysts before, during and after, this mid-Eemian phase thus has an
777 ambiguous interpretation. A continental climate in the *Pinus* zone underlined by the
778 presence of seasonal contrast dinocyst assemblages is nevertheless likely.

779 The end of the marine phase is not presented in core 22-GC3 from the Black
780 Sea due to a turbidite, although the authors suggest that the Eemian is almost

781 complete(Wegwerth et al., 2018). The end of the marine phase may however be
782 studied in the two Marmara cores here and seems to have been progressive with a
783 period when the two types of dinoflagellate assemblages co-existed (zones 22D-5
784 and 18D-4), i.e. marine and brackish.

785 It is noteworthy to underline the peaks of *S. trifida* both at the beginning of the
786 marine phase (zone 18D-3a) and at the end of it (zone 22D-4), in both cases
787 attributed to fluctuating salinities between marine and brackish-Pontocaspian. At the
788 end of the marine period in core CS22, in addition, the high values of cysts of *P.*
789 *dalei* may be attributed to terrigenous input, probably related to water level lowering,
790 as proposed for the Holocene (Verleye et al., 2009).

791 In the Black Sea, the widely-spaced analyses on core DSDP Site 380
792 sampled a weakly marine phase as the occurrence of brackish dinocyst taxa (58%)
793 is quite substantial (Ferguson et al., 2018). A diagram from DSDP Site 379A has one
794 sample in MIS 5e (Hoyle et al., 2021). This unique spectrum shows abundance of *L.*
795 *machaerophorum* with long processes, *Spiniferites* sp. and some marine taxa. The
796 $^{87}\text{Sr}/^{86}\text{Sr}$ ratio supports a marine connection at that time (Wegwerth et al., 2014;
797 Hoyle et al., 2021). The interglacial periods in the Gulf of Corinth were generally
798 marine, as Global Ocean level is high. In the lithostratigraphic unit correlated to MIS
799 5, the surface water conditions were generally warmer as established by warm-
800 loving species similar to those found in the SoM (Fatourou et al., 2023).

801 6.3.3 MIS 6 and 4

802 The NMDS analysis shows that MIS 6 and MIS 4 dinocyst assemblages are
803 clearly different, although both brackish-Pontocaspian (Fig. 8), i.e. more *P. psilata*
804 and clear presence of *S. inaequalis* and *Caspidinium* in MIS 6, contrasted with more
805 *S. cruciformis* in MIS 4. Some common points are noted between the MIS 6 and MIS

806 4 assemblages of the current study and the MIS 3 assemblages of other Marmara
807 cores ([Mudie et al., 2004](#)): all show a mix of dominant brackish-Pontocaspian
808 dinocysts with some marine ones and all have low to very low cyst concentrations. In
809 MIS 4 in the Black Sea ([Shumilovskikh et al., 2014](#)), cyst concentration is also lower
810 than in MIS 5.

811 The very low cyst concentration in the SoM may be due to high sedimentation
812 rates diluting the fossil content. Erosion is active in glacial stages, as seen by the
813 percentages of reworked elements that are very high in the MIS 6 of core CS18 and
814 in the MIS 4 of cores CS22 and CS18. Other investigators often do not count them,
815 neither published them, but they are a significant indicator of erosion and inwash.
816 Often reworking also shows proximity to shores due to low water level. The water
817 level must have been relatively low, most likely lower than at any time in MIS 5. The
818 brackish-Pontocaspian periods during MIS 5 are clearly different from those of MIS 6
819 and 4, in that the latter ones have a lot of reworking and contain abundant signs of
820 erosion ([Fig. 3, 4 and 8](#)).

821 Core CS18 in the Imralı basin is not far from the North Imralı canyon that the
822 Kocasu carved during lowstands ([Çağatay et al., 2015a and b](#)). This relative
823 proximity to a river (if not to the shore) and submarine canyon may also explain the
824 high amounts of reworked elements and the different taphonomy during the glacial
825 times ([Pope et al., 2022](#)).

826 It is very likely that, as suggested in [Hoyle et al. \(2021\)](#) based on Sr ratio in
827 the Black Sea, no marine connection occurred during the glacial periods for the SoM.
828 Several indicators (oxygen isotopes of the Sofular Cave along the Turkish coast of
829 the Black Sea ([Badertscher et al., 2011](#)) and Sr isotope ratio ([Hoyle et al., 2021](#)))
830 suggest an overflow from the Caspian Sea to the Black Sea during MIS 6.

831 Unfortunately, dinocyst assemblages cannot contribute evidence to this debate on
832 water source.

833 For MIS 6 in the SoM, a progressive, even slow, freshening after marine
834 conditions in the previous interglacial (MIS 7) was suggested on the basis of
835 divergent trends in isotopic values, i.e. $\delta^{18}\text{O}$ becomes more negative while $\delta^{13}\text{C}$
836 becomes more positive ([Çağatay et al., 2019](#)). This may explain why here the marine
837 dinocyst group is better represented in subzone 18D-1a than subzones 18D-1b and
838 c, even 18D-2. This also suggests that the base of the core may have just missed
839 the top of MIS 7. As a significant contribution of reworking of marine palynomorphs
840 (as shown here) is revealed, no marine connection would have occurred during MIS
841 6 (zones 18D-1 and 2). For that matter, at the transition from the marine phase to the
842 lake phase at MIS 5 to 4, [Grall et al. \(2014\)](#) and [Aziz et al. \(2019\)](#) suggest slope
843 failure and strong turbidity currents entraining sediment that was stored on the shelf
844 to deeper areas of the SoM, as the sea level fell. Recent investigations (modern
845 discharge and geological data) aiming at disentangling the origin of the sediment in
846 the SoM during low levels have however demonstrated that southern river input was
847 more important than suspended load through one or both straits ([Hiscott et al.,](#)
848 [2021](#)). Moreover, the latter study further estimates that river input was higher than
849 mass-wasting events and shoreface erosion. Nowadays, and most likely in the past,
850 the Kocasu is the largest southern river and it drains a large basin in NW Anatolia
851 ([Kazancı et al., 2004](#)).

852 The water levels in the Black Sea were most likely low during the cold parts of
853 MIS 6. However, three major meltwater pulses of the Eurasian icesheets occurred;
854 but it is not known if the lake level rise of the Black Sea was sufficient to cause an
855 overspill to the SoM although it is likely ([Wegwerth et al., 2019](#)). So far these pulses

856 cannot be seen in our record. For completeness of the record, it is noteworthy to
857 underline that the glacials (MIS 6 and 4) in the Gulf of Corinth were generally
858 brackish ([Fatourou et al., 2023](#)).

859 6.4 Sapropels

860 *6.4.1 Surface conditions and sapropel formation*

861 The successive sapropels deposited in SoM during MIS 5 clearly formed
862 under three different surface water conditions ([Fig. 7](#)). MSAP-2 formed under
863 brackish surface conditions, except briefly on a few cm within MSAP 2b. MSAP-4
864 developed under clearly marine surface conditions. Finally, MSAP-3 built up in a
865 period of transition from marine to brackish surface conditions.

866 Although not a perfect fit, the dinocyst concentration increases in each
867 sapropel. Admittedly dinocysts are only one part of the bloom-forming phytoplankton,
868 diatoms being the main group. In order to appreciate the amount of other
869 phytoplankton such as diatoms, the H/A ratio is used as an indicator of primary
870 productivity, as heterotrophic dinoflagellate prey on diatoms (see [Triantaphyllou et](#)
871 [al., 2009](#); [Sala-Perez et al., 2020](#)). The results show higher H/A ratio, thus higher
872 palaeoproductivity, during the marine phase in MIS 5e in core CS18, especially in
873 MSAP-4 ([Fig. 9](#)). The diatom assemblages (only available for MSAP-2 in core CS22)
874 are freshwater-brackish. This H/A ratio is low in MSAP-2b, suggesting low
875 productivity.

876 The various origins of the successive sapropel formation in the SoM, the
877 decreasing surface salinities and the decrease of the oxic condition at bottom over
878 MIS 5 may be the reason why reconstructed palaeoproductivity based on the H/A

879 ratio is clearly only higher in MSAP-4 corresponding to the warm conditions of the
880 early MIS 5e).

881 *6.4.2 Bottom conditions and sapropel formation*

882 In core CS22 below MSAP-3, the gradual shift of foraminifera assemblages,
883 starting with genera needing oxic conditions towards suboxic-tolerant genera, is an
884 effect of oxygen loss caused by the formation of water column stratification and
885 organic matter mineralisation (Fig. 9). It is worth noting that the occurrence of
886 foraminifera linings is only in the marine period. Accordingly in the marine phase of
887 core CS22, the linings occur from the core base and stop at the beginning of the
888 sapropel MSAP-3 at 1920 cm. Their occurrence is similar to that of the foraminifera
889 shells (Fig. 9).

890 In contrast, in core CS18, foraminifer linings are present at 725.5-605.5 and
891 564.5-524.5 cm depth, thus in sapropels MSAP-4 and 3. Moreover foraminifer shell
892 distribution (Çağatay et al., 2019) (Fig. 9) is like that for the linings, although
893 somewhat shorter. In core CS18, shells are found in the first part of MSAP-4 and
894 between MSAP-4 and 3.

895 This foraminifera distribution in the sapropels is therefore questioning the
896 existence of full anoxia in MSAP-4 (core CS18) and in part of MSAP-3 (mostly in
897 core CS18, no foraminifera in core CS22). For MSAP-2, it is unclear whether the
898 water remained sufficiently high saline for benthic foraminifera to live regardless of
899 oxygen availability.

900 *6.4.3 Sapropel trigger in the Sea of Marmara*

901 Combining the information from bottom and surface conditions (especially
902 dinocysts, foraminifera shells and organic linings), it may be shown that the four

903 Marmara sapropels (4, 3, 2b and 2a) clearly formed over times with decreasing
904 marine influence and increasing bottom water stagnation. Moreover, they are
905 different from each other. Only MSAP-2 may have had fully anoxic conditions during
906 the whole sapropel time.

907 The difference between the two cores may be possibly explained by the
908 shallower position of core CS18 in comparison to core CS22. The water depth
909 difference is 260 m, adding another c. 14 m (at the end of MSAP-3) due to core
910 depth difference, thus a total of c. 274 m difference (Fig. 9). The location of the
911 shallower core CS 18 would make it harder to fall within the depths affected by full
912 anoxia and would justify the occurrence of foraminifera in MSAP-4 and 3 in a dysoxic
913 area. Our results also show that the anoxia did not rise up to 291 m water depth. In
914 the east Mediterranean Sea, sporadic occurrences of foraminifers have been
915 reported in sapropels S5 (MIS5e) and S6 (MIS 6e) and, more importantly, benthic
916 foraminifers show a progressive development of dysoxic conditions before sapropel
917 deposition (Schmiedl et al., 2013), that we also observe before MSAP-3. It is also
918 interesting to note that, during MSAP-1 (Holocene), suboxic conditions extended
919 over the shelf to depths of -75 m (Çağatay et al., 2009).

920 The Marmara sapropels occur during warm periods, evidenced in the current
921 joint pollen-dinocyst investigation i.e. Eemian, Saint-Germain I and II, which more
922 likely facilitated algal blooms. Consequently, the proposed correlation between the
923 Marmara sapropels and the Mediterranean sapropels becomes: MSAP-4 – S5,
924 MSAP-3 – not correlatable, MSAP 2b – S4 and MSAP-2a – S3 (Fig. 9), based on
925 ages (Grant et al., 2016). The Marmara sapropels do not seem to be preconditioned
926 by astronomical forcing the same way as for the Mediterranean ones (Rohling et al.,
927 2015; Grant et al., 2016). A specific trigger mechanism in the SoM, as in the Black

928 Sea, is an inflow of marine water causing water column stratification ([Çağatay et al.,](#)
929 [2009](#)).

930 6.5 Terrestrial and marine transitions in the Marmara and the Black seas

931 Understanding the facies transitions in the SoM or Lake requires taking into
932 account the hydrodynamic processes and how they control sediment deposition and
933 erosion in the Bosphorus and the Dardanelles Straits. The net outflow from the Black
934 Sea to the Sea of Marmara is governed by river inputs to the Black Sea and
935 evaporation/precipitation ([Dubinin and Dubinina, 2014](#)). When the water budget of
936 the Black Sea is positive, it will inevitably overflow into the SoM and then most likely
937 into the Mediterranean Sea as the SoM has a comparatively small surface area for
938 evaporation. On the other hand, the occurrence of a counter current of sea water
939 from the Mediterranean Sea to the SoM when sea level is above the Dardanelles
940 Strait and to the SoM to the Black Sea when sea level is above the Bosphorus Sill
941 cannot be taken for granted. For instance, in the present day with a sill depth of -35
942 m in the Bosphorus Strait, the seawater bottom current is blocked when the surface
943 current exceeds $30,000 \text{ m}^3 \cdot \text{s}^{-1}$ ([Sannino et al., 2017](#)). This situation occurs during
944 transient events in the present conditions, but more generally the hydrodynamics of
945 the straits implies that the influx of seawater is a function of water depth at the sill
946 and of the freshwater outflow, which is in turn governed by river inputs and
947 evaporation/precipitation. Moreover, in steady state, the only way out for seawater
948 entering the Dardanelles Strait is by mixing with the fresher surface water layer in the
949 SoM or in the Black Sea. In the present day, and considering yearly averages, the
950 bottom current flux through the Dardanelles Strait is only about 12 % larger than the
951 bottom current flux through the Bosphorus Strait ([Beşiktepe et al., 1994](#)). It follows
952 that the halt of northward flux through the Bosphorus Strait would also greatly reduce

953 the seawater flux into the SoM and thus increase the residence time of the SoM
954 bottom water. The residence time of deep water in the SoM varies from two years in
955 the Tekirdag and the Central Basin to 10-20 years in Çınarcık Basin ([Beşiktepe et al., 1994](#)), and this causes an eastward decrease of oxygen concentration from 60 to
956 $8 \mu\text{mol l}^{-1}$ (Henry et al., 2007). Increasing the residence time will further reduce
957 oxygen content, which may in turn favour sapropel deposition.

959 Variations of sill elevation depend not only on sea level, but also on tectonic
960 movements, erosion and sedimentation. The shores of the Dardanelles Strait were
961 uplifting at a rate estimated in the $0.25\text{-}0.75 \text{ mm yr}^{-1}$ range ([Yaltırak et al., 2002](#)).
962 This may suggest that the level of the sill has risen with time. However, the
963 Dardanelles Strait is at least in part structurally controlled ([Yaltırak et al., 2000](#);
964 [Gökaşan et al., 2008](#)). Active faults, along part of its shore, may locally cause
965 subsidence of the channel floor. Two sedimentary units were deposited on an
966 erosional surface cutting through Miocene and Pliocene sediments. Their geometry
967 gives clear indications on the erosion and sedimentation processes ([Gökaşan et al., 2008](#)). The lower unit is thickest (up to 80 m) at the mouths of rivers discharging in
968 the Dardanelles Strait where they formed deltas that were subsequently incised. The
969 upper unit is composed of Holocene sediment drifts that are thickest where the
970 channel is wide and are incised by the active bottom current in the narrow parts. It
971 thus appears that the strength of the bottom current in the Dardanelles Strait, which
972 is in the present-day conditions controlled by the dynamics of the Bosphorus Strait,
973 redistributes sediments and prevents clogging of the channel. All the above
974 observations and reasoning leads to hypothesize that the interruption of the
975 northward flux through the Bosphorus Strait leads to the restriction of the
976 Dardanelles channel by deltas forming at the mouths of local rivers. In the
977

978 Dardanelles Strait, the current sill depth is defined by the top of Unit I at -65 m while
979 the basement sill is at -85 m. In the Bosphorus Strait, the present-day sill depth is at
980 -35 m while the basement sill is at -70 m ([Çağatay et al., 2009](#)).

981 In core CS18, the brackish to marine transition occurs still within the Saalian
982 Glacial (c. MIS 6), i.e. before the development of the thermophilic vegetation of the
983 Eemian Interglacial. Moreover, the end of the marine phase occurs before the end of
984 Eemian Interglacial in both cores. This offset is in line with the observation that the
985 Eemian Interglacial and the MIS 5e are not exactly synchronous, with the Global
986 Ocean level rising first, followed by vegetation changes ([Shackleton et al., 2003](#)).

987 In the SE Black Sea (core 22-GC3), the marine influence however starts only
988 at 128 ka (or 815.5 cm depth) between the *Quercus-robur*-type zone (22GC-E1) and
989 the *Juniperus* zone (22GC-E2), i.e. well into the interglacial ([Shumilovskikh et al.,](#)
990 [2013](#)); whereas here in core CS18 it starts earlier at 756 cm depth, i.e. in the middle
991 of zone 18P-2 dominated by *Artemisia* and that is probably still part of the Saalian
992 Glacial ([Fig. 9](#)). Although a large difference between the SoM and the Black Sea
993 palynological diagrams, this can be explained by the marine inflow that has first to
994 penetrate the SoM from the Aegean Sea and cross two sills, before reaching the
995 Black Sea. For comparison, a delay of 4-6 ka has been observed for the Lateglacial-
996 Holocene, with a marine incursion in the SoM starting at 14.7 ka followed by
997 progressive salinisation with a facies transition defined at 12.6 ka ([Vidal et al., 2010](#);
998 [Çağatay et al., 2015](#)) and in the Black Sea at 8.9-8.3 ka ([Shumilovskikh et al., 2013](#);
999 [Mudie et al., 2014](#); [Yakupoğlu et al., 2022](#)). This lag has been explained by the
1000 difference in sill depth between the Bosphorus and the Dardanelles Straights, the
1001 Bosphorus one being close to its present-day level (-35 m) at the time of
1002 reconnection and the Dardanelles one being somewhat deeper (-75 ± 5 m) than its

1003 present-day level. The situation during the MIS 6 to 5 transition may have been
1004 similar with a Bosphorus Strait at about -40 m ([Çağatay et al., 2019](#)). The early
1005 marine incursion in the SoM at the end of MIS stage 6 indicates that the Dardanelles
1006 sill depth was also deeper at that time.

1007 A new result from the present study that remains to be explained is the
1008 prevalence of brackish Ponto-Caspian conditions in surface water during the later
1009 phases of MIS 5 at times when sea level was well above the present-day
1010 Dardanelles sill level ([Fig. 9](#)). The explanation we propose is that the Bosphorus
1011 northward flow was first decreased then interrupted as sea level started decreasing
1012 before the end of MIS 5e, and that resulted in a progressive reduction of seawater
1013 inflow through the Dardanelles Strait and a silting up. This eventually enabled the
1014 sediment filling of the channel by local rivers, further limiting marine input. A
1015 decreased oxygen availability in the marine deep waters, as indicated by foraminifer
1016 assemblages, is observed. Low and/or episodic sea-water inputs during the later part
1017 of MIS 5 may have maintained stratified conditions in the SoM with low salinity
1018 surface waters and nearly anoxic bottom water of intermediate salinity. Sea level
1019 drop during MIS 5d and 5b presumably interrupted the marine inflow, lowering the
1020 vertical salinity contrast and allowing better oxygenation of the deep waters. During
1021 the following glacial lowstands (MIS 4-3-2), the Marmara lake remained isolated and
1022 outflow (either permanent or episodic) may have incised the sedimentary deposits to
1023 allow reconnection at \approx -75 m at the end of the glaciation.

1024 The $\delta^{18}\text{O}$ of the well-dated Sofular stalagmite record reflect the composition
1025 of the Black Sea surface waters by way of evaporation and moisture availability
1026 ([Badertscher et al., 2011](#)) ([Fig. 1](#)). These authors propose one clear period of
1027 connection between the Black Sea and the Mediterranean Sea at roughly MIS 5e

1028 based on a threshold of $-8.5 \pm 1\text{‰}$ $\delta^{18}\text{O}$ values. The next period below that threshold
1029 is at MIS 5c and the data also suggest connection but for a shorter period. The last
1030 period at MIS 5a would not have led to sufficiently high water levels to allow flowing
1031 of the Mediterranean marine waters into the Black Sea. This fits rather well our
1032 results in the SoM, with a full surface water connection at MIS 5e when global sea
1033 level reached 8 m above sea level, a marine incursion affecting surface waters
1034 during MIS 5c when global sea level reached 24 m below sea level and no incursion
1035 during MIS 5a when global sea level reached only 28 m below sea level ([Rohling et](#)
1036 [al., 2021](#)).

1037 **7 Conclusions**

1038 The results of this investigation have shown the widely changing states of the
1039 Sea of Marmara during the MIS 6 and 5, implying vastly changing connexions with
1040 the Pontocaspian basin and the Global Ocean.

1041 From the methodological point of view, as for the Caspian Sea ([Leroy et al.,](#)
1042 [2013, 2014](#)) when using the palynological method at large, it is essential to study
1043 separately and objectively the terrestrial conditions and the aquatic ones. A
1044 comparison of the two approaches may then clearly demonstrate if the changes are
1045 synchronous or not, and if different forcing factors are at play.

1046 As demonstrated for the Iberian margin by pollen and isotopic analyses
1047 ([Shackleton et al., 2003](#)), the Eemian starts and finishes respectively after the start
1048 and after the end of MIS 5e high levels. The same offset can be demonstrated here
1049 in this joint pollen - dinocyst study.

1050 The combination of the two sequences, CS22 and CS18, following the
1051 chronology proposed in [Leroy et al. \(2023\)](#), allows highlighting one main marine

1052 phase corresponding to MIS 5e and a minor one in MIS 5c. The rest of the record is
1053 brackish-Pontocaspian although with some differences across the sequence (e.g.
1054 MIS 6 versus MIS 4). Based on vegetation and climatic data allowing linking the
1055 records, a delay in the marine influence between SoM and SE Black Sea is
1056 observed, perhaps due to the closure of the Bosphorus Sill and its late opening or a
1057 different sill height.

1058 The current water level reconstruction in MIS 5 corresponds well to the Global
1059 Ocean level curve and Black Sea salinity reconstructions, both indicating high stands
1060 that become progressively lower and thus decreasing connections between the
1061 Black Sea and the SoM.

1062 The four sapropels (MSAP-4, 3, 2a and 2b) formed under very different
1063 surface water conditions ranging from marine to brackish-Pontocaspian, but always
1064 under a warm climate. Moreover, the shallower site in the east (Imralı basin) does
1065 not form as often full anoxia than the western and deeper site.

1066 In the future, it would be interesting to apply palynological investigations
1067 (pollen and dinocysts) on long cores on both sides of the Bosphorus Strait to define
1068 better for example the timing difference for the marine reconnection at the boundary
1069 MIS 6-5.

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1082

1083 **Data availability**

1084 The dinocyst data are available in PANGEA.

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