



HAL
open science

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

S.A.G. Leroy, P. Henry, F. Marret, Christine Paillès, L. Licari, J. Kende, F. Rostek, E. Bard

► To cite this version:

S.A.G. Leroy, P. Henry, F. Marret, Christine Paillès, L. Licari, et al.. Dinocyst assemblages and water surface conditions in the Sea of Marmara during MIS 6 and 5 from two long cores. *Quaternary Science Reviews*, 2023, 314, pp.108229. 10.1016/j.quascirev.2023.108229 . hal-04227584

HAL Id: hal-04227584

<https://hal.science/hal-04227584>

Submitted on 3 Oct 2023

HAL is a multi-disciplinary open access archive for the deposit and dissemination of scientific research documents, whether they are published or not. The documents may come from teaching and research institutions in France or abroad, or from public or private research centers.

L'archive ouverte pluridisciplinaire **HAL**, est destinée au dépôt et à la diffusion de documents scientifiques de niveau recherche, publiés ou non, émanant des établissements d'enseignement et de recherche français ou étrangers, des laboratoires publics ou privés.

Quaternary Science Reviews

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

--Manuscript Draft--

Manuscript Number:	
Article Type:	Research Paper
Keywords:	Dinocysts, MIS 6 and 5, Marmara Sea, connection, sea level, sapropel
Corresponding Author:	Suzanne A.G. Leroy FRANCE
First Author:	Suzanne A.G. Leroy
Order of Authors:	Suzanne A.G. Leroy Pierre Henry fabienne Marret-Davies Christine Pailles Laetitia Licari julia kende frauke rostek edouard bard
Abstract:	<p>The Sea of Marmara is the connection between the vast Black Sea-Caspian Sea basin (Pontocaspian) and the Global Ocean via the Mediterranean Sea. Its water levels and water conditions has widely varied over times. Combining two cores in the Sea of Marmara (Turkey) and using organic-walled dinoflagellate cysts as the main proxy (allied to alkenones, diatoms and benthic foraminifera), allow qualitatively reconstructing water conditions during Marine Isotopic Stage (MIS) 6 and 5. A clear main marine phase is illustrated in MIS 5e. A minor marine phase occurred during MIS 5c, mostly supported by alkenone data. The rest of the record indicates brackish Pontocaspian conditions, with more <i>Spiniferites inaequalis</i> in MIS 6 and more <i>S. cruciformis</i> in MIS 5.</p> <p>An earlier initial marine flooding in the Marmara Sea (MIS 6/MIS 5) in comparison to the Black Sea was highlighted on the base of difference in terrestrial vegetation reconstructed from pollen analyses in the same Marmara cores. Understanding the hydrologic evolution of the Sea of Marmara during MIS 5 implies taking into account the consequences of the disconnection of the Black Sea at the Bosphorus on the hydrodynamics and sediment transport processes in the Dardanelles Strait.</p> <p>The sapropels of the Marmara Sea form when marine water penetrates at depth from the Aegean Sea. Combining surface water proxies with benthic foraminifera (shells and inner organic lining), it suggests that the sapropels over time formed under decreasing anoxic conditions. A comparison between the two cores suggests anoxia did not rise up to 291 m water depth.</p>
Suggested Reviewers:	<p>Lyudmila Shumilovskikh University of Göttingen lshumil@gwdg.de she has published a dinocyst diagram in the Black Sea covering part of MIS5e</p> <p>Katarina Kouli National and Kapodistrian University of Athens akouli@geol.uoa.gr she has published a dino diagram from the G of Corinth with similar assemblages to the Marmara Sea</p> <p>Laurent Londeix University of Bordeaux l.londeix@epoc.u-bordeaux1.fr</p>

22/03/2023

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

4

5

6 Leroy S.A.G. ^{1-2-3*}, Henry P. ¹, Marret F. ³, Pailles C. ¹, Licari L. ¹, Kende J. ¹, Rostek
7 F. ¹, Bard E. ¹

8

9 *1 Aix Marseille Univ, CNRS, IRD, INRAE, Coll France, CEREGE, Aix-en-Provence,*
10 *France, suzleroy@hotmail.com, henry@cerege.fr, pailles@cerege.fr,*
11 *licari@cerege.fr, jj.kende@gmail.com, rostek@cerege.fr, bard@cerege.fr*

12 *2 Aix Marseille Univ, CNRS, Minist. Culture & Com., LAMPEA, 13094 Aix-en-*
13 *Provence, France*

14 *3 School of Environmental Sciences, University of Liverpool, L69 7ZT Liverpool, UK,*
15 *fmarret@liverpool.ac.uk*

16 ** Corresponding author*

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

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

29 An earlier initial marine flooding in the Marmara Sea (MIS 6/MIS 5) in comparison to
30 the Black Sea was highlighted on the base of difference in terrestrial vegetation
31 reconstructed from pollen analyses in the same Marmara cores. Understanding the
32 hydrologic evolution of the Sea of Marmara during MIS 5 implies taking into account
33 the consequences of the disconnection of the Black Sea at the Bosphorus on the
34 hydrodynamics and sediment transport processes in the Dardanelles Strait.

35 The sapropels of the Marmara Sea form when marine water penetrates at depth
36 from the Aegean Sea. Combining surface water proxies with benthic foraminifera
37 (shells and inner organic lining), it suggests that the sapropels over time formed
38 under decreasing anoxic conditions. A comparison between the two cores suggests
39 anoxia did not rise up to 291 m water depth.

40 **Key words**

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

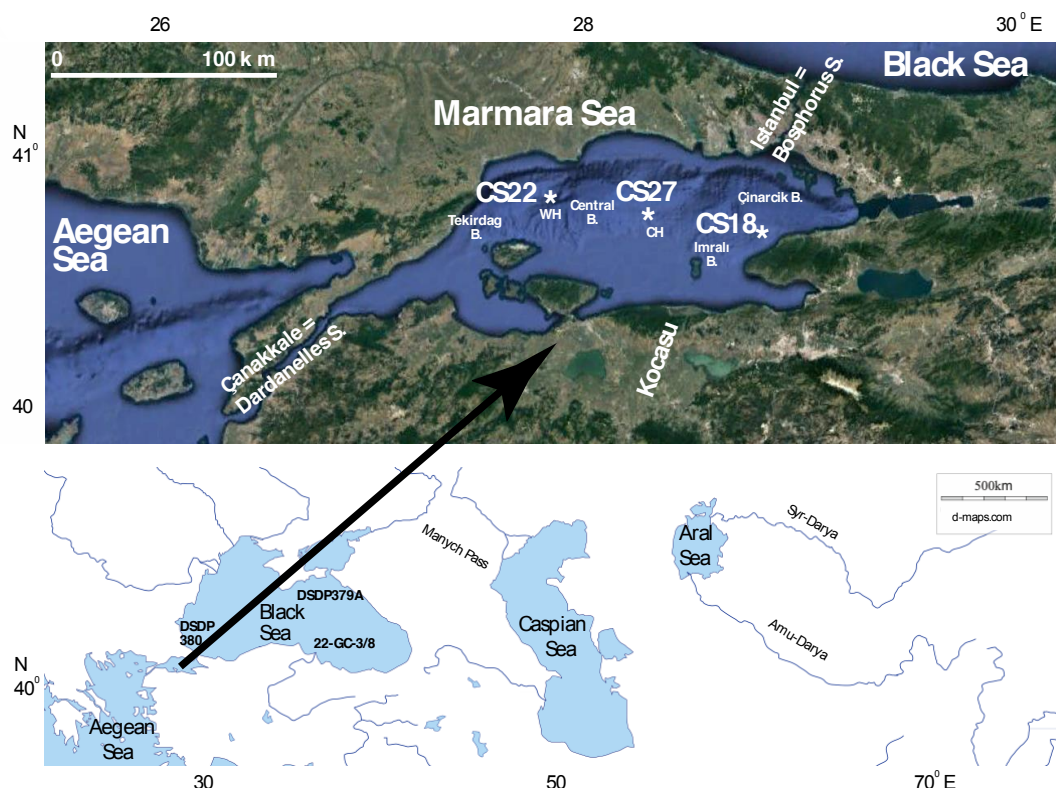
42

43 **1 Introduction**

44 The Sea of Marmara (SoM) is a gateway between the Mediterranean Sea and
45 the Black Sea in Turkey, and has experienced water-level fluctuations and
46 hydrological variations during the late Quaternary, which were mainly controlled by
47 bedrock sills at the Marmara Sea exit of Çanakkale Strait (Dardanelles) and the
48 Black Sea entrance of the Istanbul (Bosphorus) Strait, respectively. The variations in
49 water depths of these sills were possibly governed by longer-term vertical
50 displacements related to tectonics and erosion/sedimentation balance due to sea-
51 level changes and water exchanges between the Black and Aegean seas, or
52 between the Ponto-Caspian ensemble and the Global Ocean ([Yaltrak, 2002](#);
53 [Gökaşan et al., 2008](#), [Çağatay et al., 2009, 2019](#); [Vidal et al., 2010](#); [Eriş et al., 2011](#))
54 ([Fig. 1](#)). The SoM is indeed a highly seismic area as it is located on the western end
55 of the North Anatolian Fault; however, tectonics are mainly significantly affecting sea
56 levels only in the long term. At the scale of the last climatic cycle, global ocean level
57 change is the main factor controlling the water level in the SoM. For the last
58 interglacial (Marine Isotope Stage 5, MIS 5), decreasing water level maxima are
59 noted along the successive three warm stages: MIS 5e, 5c and 5a ([Rohling et al.,](#)
60 [2021](#)). Moreover, the Black Sea level should be considered as another crucial
61 influence on water level in the SoM, evidenced as the melting of the Eurasian icecap
62 and the subsequent evacuation of the freshwater along large rivers flowing into the
63 Black Sea ([Tudryn et al., 2016](#)) and, at times, the outflow from the Caspian Sea via
64 the Manych Passage ([Chepalyga, 2007](#)). The SoM is the outlet of all the
65 Pontocaspian basins (as far as the Tien-Shan via the Syr-Darya and the Chu River
66 and the Pamir-Indu-Kush via the Amu-Darya) and hence this inland sea is the
67 receptacle of the history of a large geographical area ([Fig. 1](#)). Thus, the water level

68 history of the SoM is quite complex (local, regional and global forcing) and its insight
 69 clearly has an extra-regional relevance. Its history starts to be well-known for the
 70 Last Glacial Maximum and the Holocene, but it remains much less clear before
 71 (Leroy et al., 2020; Eriş et al., 2007, 2011; Çağatay et al., 2009, 2015a). Moreover,
 72 the SoM forms sapropels under certain conditions. They are dark and finely
 73 laminated sediment rich in organic matter and are important to study as they consist
 74 of layers where carbon can be locked away from oxidation, thus contributing to the C
 75 cycle. In the Mediterranean Sea, enhanced marine productivity increases organic
 76 matter degradation and anoxia (Rohling et al., 2015; Zwiép et al., 2018).

77



78

79 **Figure 1:** Map of Marmara Sea with core locations (top panel) and map of the seas
 80 that were potentially connected at times (lower panel). CH: Central High, WH:
 81 Western High, S: Strait, B: Basin.

82

83 Owing to a major European research programme, long sediment cores (up to
84 30 m) were taken in the SoM during the MARSITE cruise in 2014 to address some of
85 those key questions (<https://cordis.europa.eu/project/id/308417>).

86 The aim of the current investigation is, by using organic-walled dinoflagellate
87 cyst (herein called dinocyst) analysis (combined to other proxies such as alkenone,
88 diatoms and benthic foraminifera), to establish the conditions at the surface of the
89 water within MIS 5 (130-71 ka ago, [Lisiecki and Raymo, 2005](#)) and part of MIS 6 (>
90 130 ka), and thus to contribute to water-level and surface-water-condition history in
91 the SoM. A sister paper addresses past vegetation, palaeoclimates and chronology
92 on the same samples ([Leroy et al., submitted](#)).

93 **2 Regional setting**

94 The SoM (275 km W-E by 80 km N-S) is located at the western end of the
95 North Anatolian Fault where it splits into several strands; as a result, the sea
96 comprises three deep basins (Tekirdag, Central and Çınarcık) separated by two
97 highs: Western High and Central High ([Fig. 1](#)). Its deepest location is in the Çınarcık
98 and Central Basins at 1270 m. Rivers are mostly inflowing from the south ([Kazancı
99 et al., 2004](#)). The Imralı Basin in the SE SoM is nowadays a flat area of a water
100 depth only around 300-400 m ([Yaltırak, 2002](#)). The Kocasu (or River Koca), the main
101 river flowing into the SoM, exits now from the continent, south of the Imralı Island. At
102 lower water levels, it has formed deltas prograding into the Imralı Basin and has
103 largely contributed to fill it; later on, the sediment was fed into the Çınarcık Basin
104 along a now-submerged canyon ([Kazancı et al., 2004](#); [Sorlien et al., 2012](#)) ([Fig. 1](#)).

105 The region immediately around the SoM is relatively flat, although mountains up to
106 2500 m are found in the S and SE and 1000 m in the N.

107 Water salinity is 22-24 psu at surface and up to 37-39 psu at depth ([Beşiktepe](#)
108 [et al., 1994](#); [Aydoğdu et al., 2018](#)). The brackish water of the Black Sea (18 psu),
109 being less dense, outflows through the SoM above the deeper denser Mediterranean
110 saline water moving eastwards ([Aydoğdu et al., 2018](#)). The pycnocline is at around
111 25 m water depth. This delicate two-way setting is easily disturbed by water-level
112 changes leading to deep salinity and oxygenation modifications crucial for the
113 survival of aquatic ecosystems. Water temperature fluctuates at most from 4 to 28 °C
114 annually. It displays a slight warming gradient from the NE to the SW: for example,
115 surface water usually varies from 21 (NE) to 26 (SW) °C in summer. All year round,
116 bottom water is relatively stable at c.15 °C ([Aydoğdu et al., 2018](#)). The water depth
117 of the Dardanelles (Çanakkale) Sill is nowadays 65 m, and that of the Bosphorus
118 (Istanbul) Sill is 35 m ([Çağatay et al., 2009](#)).

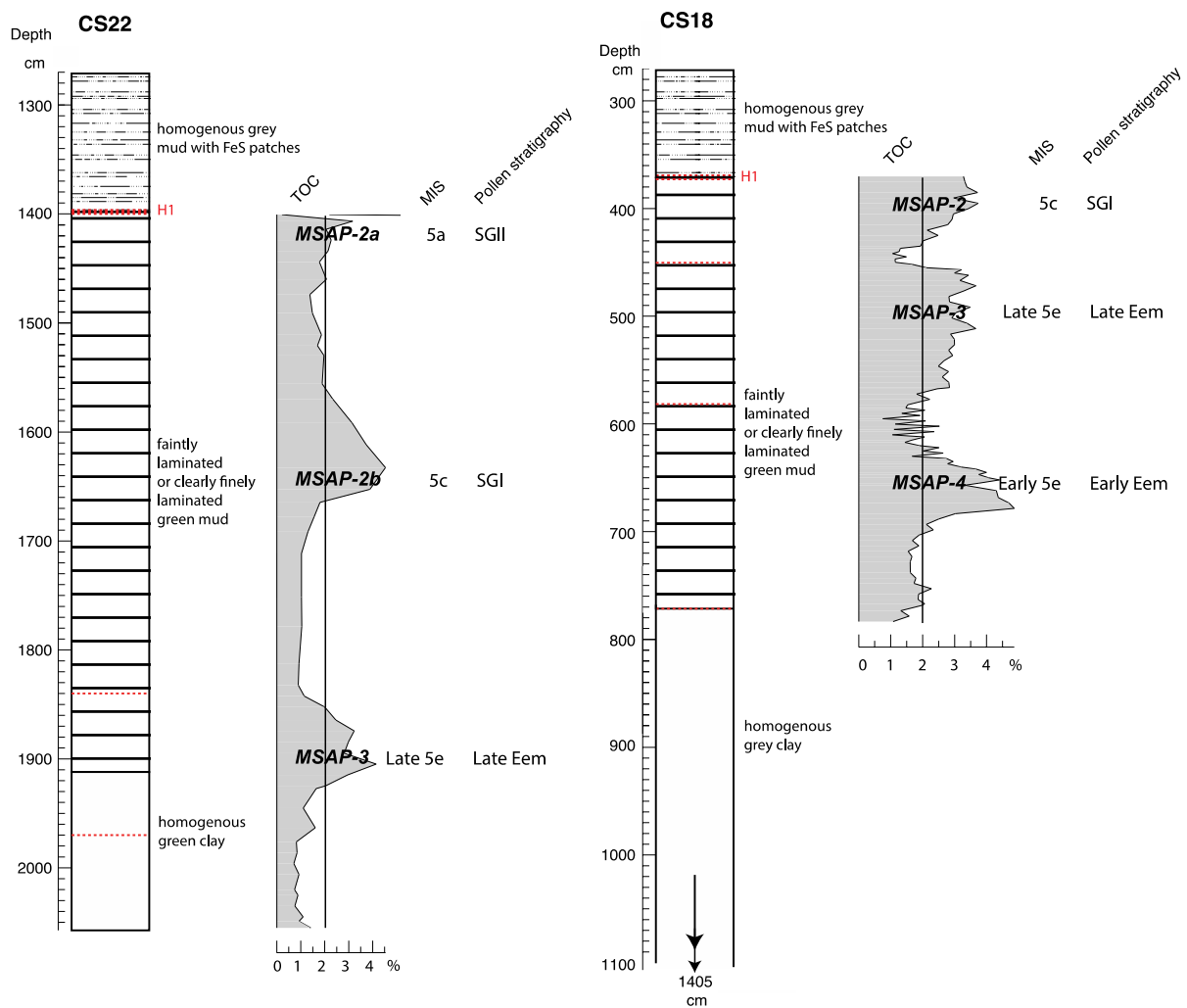
119 The dinocyst assemblages of MIS 6 and 5 have, so far, not been studied in
120 the SoM. However, three dinocyst analyses for these time intervals are available for
121 the Black Sea. In cores DSDP 380 in the SW Black Sea and DSDP Site 379A in the
122 E Black Sea, MIS 5 is covered by a few samples ([Ferguson et al., 2018](#); [Hoyle et al.,](#)
123 [2021](#)). In the SE Black Sea, a detailed dinocyst diagram of core 22-GC3 includes the
124 period from 129 to 119 ka, with a transition from brackish Pontocaspian to marine
125 conditions at 128 ka ([Shumilovskikh et al., 2013](#)) ([Fig. 1](#)). More dinocyst records are
126 available for the Late Glacial and Holocene in the SoM ([Mudie et al., 2001, 2002,](#)
127 [2004](#); [Londeix et al., 2009](#)) where a change from brackish Pontocaspian to marine
128 conditions was observed at c. 9.5-9 ka. A sapropelic layer, MSAP-1, was identified at
129 11.5-7 ka ago ([Çağatay et al., 2000](#); [Tolun et al., 2002](#); [Vidal et al., 2010](#)). It is worth

130 mentioning the recent study of dinocyst assemblages on a long core in the Gulf of
131 Corinth, covering the MIS 6 and MIS 5, although further afield, due to the similarity of
132 isolation-connection history of its basin ([Fatourou et al., 2023](#)).

133 Dinoflagellates, a major component of the marine phytoplankton, have a
134 complex life cycle involving a theca that lives in the photic zone and, for about 13 to
135 16% of the species, an organic-walled cyst is produced in the water column, and
136 settling at the bottom of the sea floor for an obligatory dormancy period (e.g. [de](#)
137 [Vernal and Marret, 2007](#); [Bravo and Figueroa, 2014](#); [Mudie et al., 2021](#)). Therefore,
138 the distribution of dinocysts in recent sediments has been linked to surface
139 environmental conditions and this relationship has enabled robust quantitative
140 reconstructions of past sea-surface temperature, salinity, productivity and sea-ice
141 cover duration (e.g. [de Vernal et al., 2020](#)). A study of 181 surface samples in the
142 Pontocaspian region has clearly highlighted that the dinocyst assemblages reflect
143 surface conditions such as salinity, especially from January to March ([Mudie et al.,](#)
144 [2017](#)).

145 **3 Previous investigations on cores CS22 and CS18**

146 Calypso cores were taken by the vessel “Pourquoi Pas?” in November 2014.
147 Core MRS-**CS22** (20.42 m long) was obtained close to the top of the Western High
148 at 40,83863 latitude N, 27,79906 longitude E and 551 m water depth ([Fig. 1](#)). Core
149 MRS-**CS18** (14.25 m long) was acquired from a slope of the Imralı Basin at
150 40,66206 latitude N and 28,87958 longitude E by 291 m water depth. The core
151 depths of interest are located below a major seismic reflector, i.e. the red-H1
152 reflector.



153

154

155 **Figure 2:** Lithology of cores CS22 and CS18, total organic carbon in percentage
 156 (TOC). Vertical line at 2 % delimiting the sapropels. MIS: Marine Isotopic Stages. Me
 157 I = Melisey I Stadial, SG I = Saint-Germain I Interstadial, Me II = Melisey II Stadial,
 158 SG II = Saint-Germain II Interstadial, Eem = Eemian Interglacial.

159

160 Detailed lithological descriptions can be found for cores CS22 and CS18
 161 respectively in [Kende \(2018\)](#) and [Çağatay et al. \(2019\)](#). The sediment is a grey or
 162 green clay to silty clay at times clearly finely laminated ([Fig. 2](#)). Total Organic Carbon
 163 (TOC) values in core CS22 were obtained by using an *Fisons* NA1500 elemental
 164 analyser and for core CS18 by the catalytic combustion method using a Shimadzu

165 TOC analyzer (Kende, 2018; Çağatay et al., 2009; Leroy et al., submitted). In core
166 CS18, foraminifera were counted in the sand size fraction and their abundance were
167 qualitatively recorded. Only rare benthic foraminifera are present in the 770-700 cm
168 interval, but benthic, with some pelagic foraminifera, become abundant in 700-670
169 cm and 630-580 cm intervals. However, they are absent in the interval 580-430 cm
170 (Çağatay et al., 2019).

171 Below the red-H1 reflector, the three sapropels (here > 2%) were attributed to
172 MIS 5e, 5c and 5a based on correlation of the μ -XRF Ca with the NGRIP $\delta^{18}\text{O}$, the
173 TOC correlation with regional pollen records (outside of the SoM), oxygen isotope
174 records in the Mediterranean and global sea level. These are for core CS18: MSAP-
175 4 at 699-623 cm, MSAP-3 at 568-455 cm and MSAP-2 at 430-370 cm depth
176 (sapropel numbering proposed by Çağatay et al., 2009, 2019). These are for core
177 CS22, MSAP-3 at 1920-1853 cm and MSAP-2 in two parts: MSAP-2b at 1653-1570
178 cm and to a lesser extent MSAP-2a at 1460-1407 cm (their naming follows what was
179 proposed in Çağatay et al. (2009)). The latter authors have proposed a correlation of
180 the MSAP-4, 3 and 2 to the Mediterranean sapropels S5, S4 and S3. A revision of
181 the two Marmara sequence ages, based on geochemistry and pollen analysis,
182 indicates that for core CS22, MSAP-3 is Late Eemian Interglacial, MSAP-2b is Saint-
183 Germain I Interstadial, and MSAP-2a is Saint-Germain II Interstadial, or
184 approximately MIS late 5e, 5c and 5a (Fig. 2; Leroy et al., submitted). In core CS18,
185 MSAP-4 is early Eemian Interglacial, MSAP-3 is late Eemian and MSAP-2 is Saint-
186 Germain I, or approximately MIS early 5e, late 5e and 5c (Fig. 2; Leroy et al.,
187 submitted).

188 Minor and major hiati are defined and discussed in Leroy et al. (submitted)
189 and shown as red lines on figure 2 and following.

190 **4 Material and Methods**

191 4.1 Dinocyst analysis

192 Fifty-one sediment samples were treated in core CS22 and 79 in core CS18
193 mostly below the red-H1 reflector for their palynomorph content. The sampling
194 resolution varied between 10 and 20 cm. After measurement of the sediment volume
195 (between 0.5 and 2.5 ml), one tablet of *Lycopodium clavatum* was added. The
196 sediment was then soaked in tetrasodium pyrophosphate, followed by acid attacks:
197 cold HCl (10% dilution), cold HF (between 40 and 60%) and again cold HCl.
198 Samples were then rinsed with distilled water and sieved on nylon meshes at 125
199 and 10 µm. The residues were transferred to vials and slides mounted in glycerol. At
200 least 60 dinocysts were counted in each sample, except in core CS22 above 1400
201 cm, and in core CS18 above 369 cm and below 786.5 cm. Some gritty samples in
202 core CS18 were also centrifuged in Sodium Polytungstate at a density of 2.4. All
203 samples were rich in dinocysts except one at 364 cm depth in core CS18 and one at
204 2026.7 cm in core CS22. All percentages are calculated based on the sum of all
205 dinocysts, except varia (indeterminate, indeterminable and reworked specimens) and
206 foraminiferal organic linings. Concentration of palynomorphs is calculated in ml of
207 wet sediment. The P/D ratio is the ratio of the concentration of pollen on dinocysts. It
208 reflects the continentality of the sample ([McCarthy and Mudie, 1998](#)). The H/A ratio
209 is the ratio of heterotrophic (dependent on nutrients synthesised by other organisms)
210 taxa on autotrophic (whose energy source is the light) ones. Zonation (on 23 and 22
211 dinocyst taxa respectively in cores CS22 and CS18) was made by CONISS after
212 square-root transformation, available in the psimpoll software ([Bennett, 2007](#)), which
213 was also used for plotting dinocyst diagrams. The diversity was calculated using the

214 Margalef's species richness ([Margalef, 1958](#)) and an ordination analysis with non-
215 Metric Multidimensional Scaling (NMDS) (Primer 7; [Clarke and Gorley, 2015](#))
216 allowing distinguishing groups of samples of similar composition.

217 Identifications were made with the support of the atlas of [Mudie et al. \(2017\)](#),
218 and plates in [Marret et al. \(2004\)](#), [Sorrel et al. \(2006\)](#), [Londeix et al. \(2009\)](#) and
219 [Leroy \(2010\)](#). *Spiniferites cruciformis* specimens were grouped in three different
220 forms (A – standard form, B – pointed apex and C – short and folded processes) as
221 in [Marret et al. \(2004\)](#). *Lingulodinium machaerophorum* ss indicates the sensu stricto
222 form ([Leroy et al., 2006](#); [Mertens et al., 2009](#); [Mudie et al., 2017](#)). The “*L.*
223 *machaerophorum* bulbous” form has short but clearly bulbous (as in [Mudie et al.](#)
224 [\(2017\)](#)) and non-clavate processes. *L. machaerophorum* B is described in [Leroy et](#)
225 [al. \(2006\)](#), as having short microgranulate processes with a large striated conical
226 base. *Impagidinium* sp. A is a form defined in [Londeix et al. \(2009\)](#). The name
227 *Spiniferites inaequalis* is conservatively used, although some ongoing discussions
228 propose to move the species to the genus *Impagidinium* ([Mertens et al., 2018](#)). In
229 the cumulative diagram, the marine taxa are on the left and the brackish
230 Pontocaspian ones on the right. In the middle, are the heterotrophic taxa and the
231 various forms of *Lingulodinium machaerophorum*. Warm-loving dinocysts are defined
232 by the presence of *Operculodinium israelianum*, *Spiniferites pachydermus*,
233 *Polysphaeridinium zoharyi*, *Tuberculodinium vancampoae* and *Spiniferites mirabilis*
234 as they reflect temperatures higher than nowadays ([Marret and Zonneveld, 2003](#);
235 [Londeix et al., 2009](#); [Shumilovskikh et al., 2013](#); [de Vernal et al., 2020](#); [Marret et al.,](#)
236 [2020](#)). Seasonal contrast in sea-surface temperature (SST) is reflected by
237 *Bitectatodinium tepikiense*, *Pentapharsodinium dalei* and *Scrippsiella trifida* ([Head et](#)
238 [al., 2006](#); [Mudie et al., 2017](#)).

239 Although not dinocysts, coiled organic linings of foraminifera were found and
240 added to this investigation, rather than to the pollen diagram (Leroy et al.,
241 submitted), as they usually belong to the marine environments. The inner organic
242 lining is often produced by benthic foraminifera (de Vernal, 2009; Mudie et al., 2021).

243 4.2 Other proxies in core CS22

244 **Alkenones** were measured in sediment samples from core CS22 between
245 2050 and 1370 cm. After freeze-drying and grinding, samples were extracted using
246 an accelerated solvent extraction system (ASE350, Thermo Scientific). After a clean-
247 up procedure using silica gel column chromatography with a mixture of
248 dichloromethane: hexane (1:1), the total lipid extract was analysed by gas
249 chromatography and flame ionization detection (GC-FID, Trace GC, Thermo
250 Scientific). Identification and purity of alkenones were checked in several samples by
251 gas chromatography coupled to quadrupole mass spectrometry (Trace GC DSQII,
252 Thermo Scientific). GC conditions were similar to those described by Sonzogni et al.
253 (1997). When present, they are considered as marine indicators.

254 **Benthic foraminiferal faunas** were quantitatively investigated in the 125 µm
255 fraction from nine samples. At first a qualitative downcore check of foraminiferal
256 content was performed in the coarse fraction at a 1 m sampling resolution, then
257 refined to 20 cm between 2035 and 1915 cm where very rich benthic foraminifera
258 assemblages were found. Two additional samples were investigated at 1632 and
259 1631 cm. When possible, 300 specimens were counted in sediment aliquots, and
260 identified to the lowest taxonomic level possible, mainly at the species level. To
261 describe major faunal trends, the species Benthic Foraminiferal Number (BFN,
262 expressed as individuals per g or ind.g⁻¹) and frequency data of main species (>1%)
263 were used.

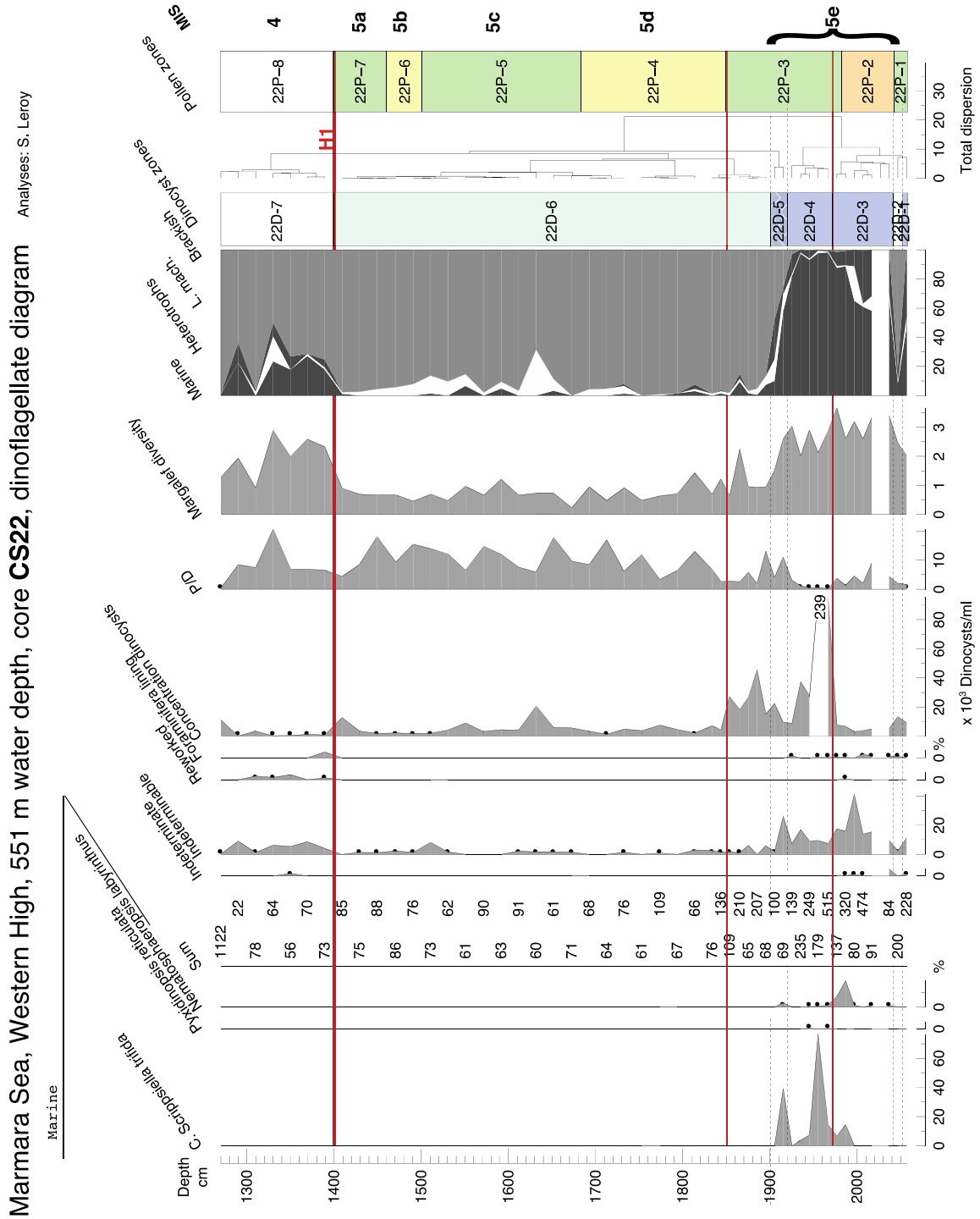
264 Sediment samples for **diatom and chrysophyte** analyses were collected at
265 20 cm intervals in each section of core CS22. Smear slides were prepared with a
266 suspension made of known amounts of dry sediment and distilled water. 200 µl of
267 this suspension were spread onto a cover slip, air-dried and mounted on a glass
268 slide using Naphrax mounting medium. Diatom valves were counted along transects
269 and expressed as number of valves per gram of dry sediment. Chrysophycean cysts
270 were also counted; but morphotypes were not identified.

271 **5 Results**

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

281 5.1 Dinocysts in core CS22

282



285

286

287 **Figure 3 a and b:** Percentage dinocyst diagram of core CS22. Black dots for < 5%.

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

289 the warm indicators in the marine group.

290

291 The dinocyst assemblages show a diversity from 3 to 19 taxa, with notably
292 higher numbers of taxa at the bottom of the core, i.e. below 1920 cm (Fig. 3).
293 Broadly, the record shows a succession of two main assemblages, one mostly
294 dominated by marine taxa, and the following one by brackish Pontocaspian taxa.
295 The main change in the diagram according to the CONISS analysis is by far between
296 zones 22D-4 and 5 at 1920 cm depth (Fig. 3). However, it is not sharp.
297
298 Zones 22D-1 and D-2 each consist of one sample (at 2057 and 2047 cm depth). In
299 rapid alternation, the first assemblage is marine with a dominance of *L.*
300 *machaerophorum* alongside: *Spiniferites* sp., *S. mirabilis*, cysts of
301 *Pentapharsodinium dalei* and *Tuberculodinium vancampoeae*. The second sample
302 has a brackish Pontocaspian assemblage, with the dominance of *Spiniferites*
303 *cruciformis* A and *Pyxidinoopsis psilata* and low occurrence of *L. machaerophorum*.
304 Foraminifera linings are observed in zones D-1 to D-4.
305
306 Zone 22D-3 (2042-1972 cm depth): Besides zone 22D-1, *L. machaerophorum*
307 displays its highest values here, declining only in the last third of this zone. The
308 percentages of *Operculodinium centrocarpum* increase progressively to a maximum
309 at the end of this zone. *Spiniferites* sp. also has maximal values. The other taxa,
310 mainly marine, are rather diverse (high Margalef diversity index). A high number of
311 indeterminable cysts occur. The sample at 2027 cm depth is rich in pollen, but hardly
312 contains any dinocysts.
313
314 Zone 22D-4 (1972-1920 cm depth): The assemblage is still marine and diverse, but
315 rather different from the previous zone. The spectra are alternatively dominated by

316 cysts of *P. dalei* and cysts of *Scrippsiella trifida*. *L. machaerophorum* is present but in
317 small amounts. The P/D ratio is at its minimum, largely because the concentration in
318 dinocyst is very high, i.e. >230,000 cysts/ml. The number of indeterminable cysts is
319 decreasing but still significant.

320

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

327

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

337

338 Zone 22D-7 (1399-1270 cm depth): The values of *S. cruciformis* remain the same as
339 in the preceding zone, but those of *P. psilata* significantly drop. The assemblage is
340 diverse with many marine forms but with a low occurrence. *L. machaerophorum* is

22/03/2023

341 back with irregular presence between 0.2 and 14 %. Three foraminifera linings (4%)

342 were found at the base of this zone. Concentration is varying, at times below 500

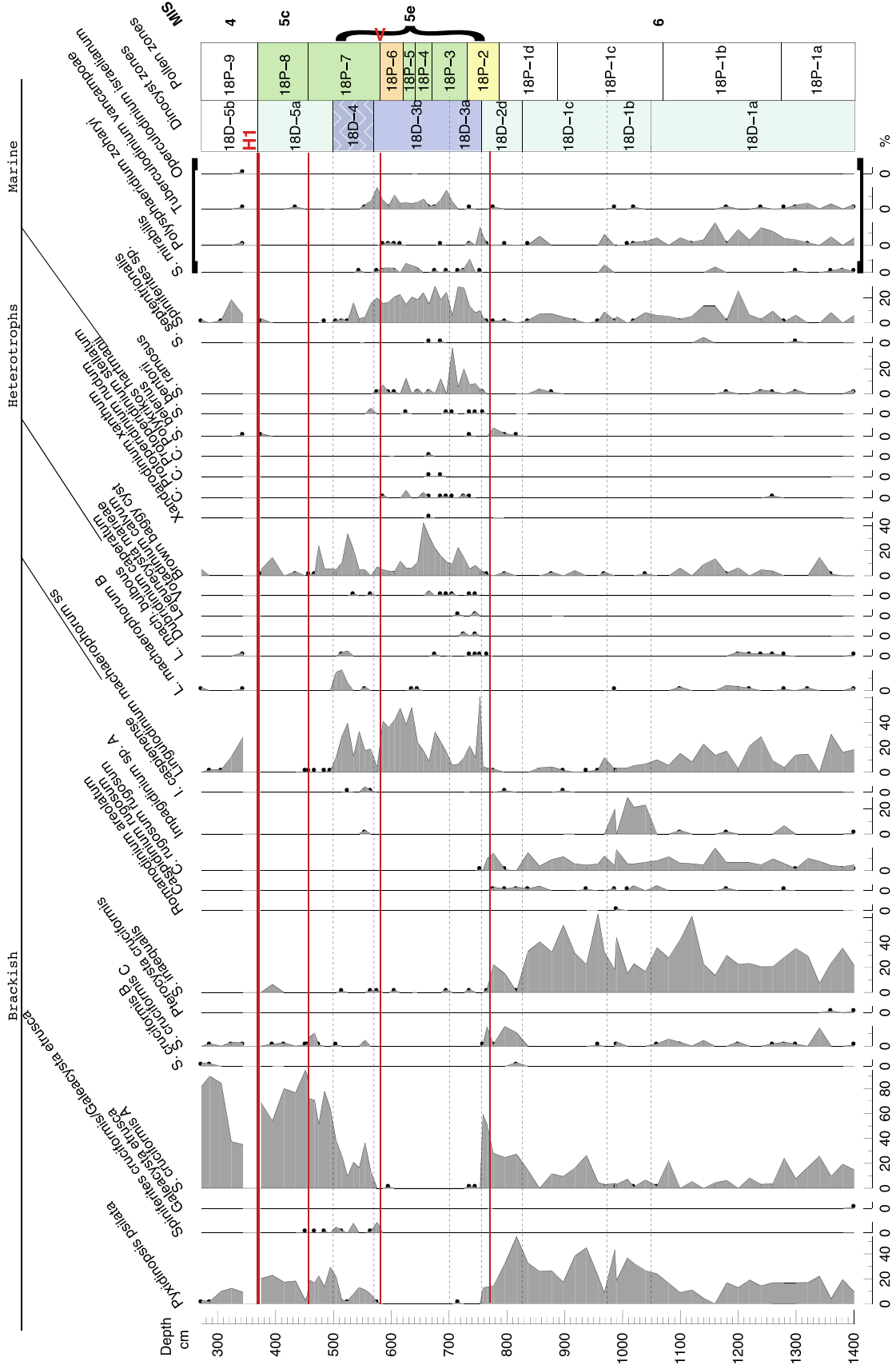
343 dinocysts/ml. Reworked elements are frequent.

344 5.2 Dinocysts in core CS18

345

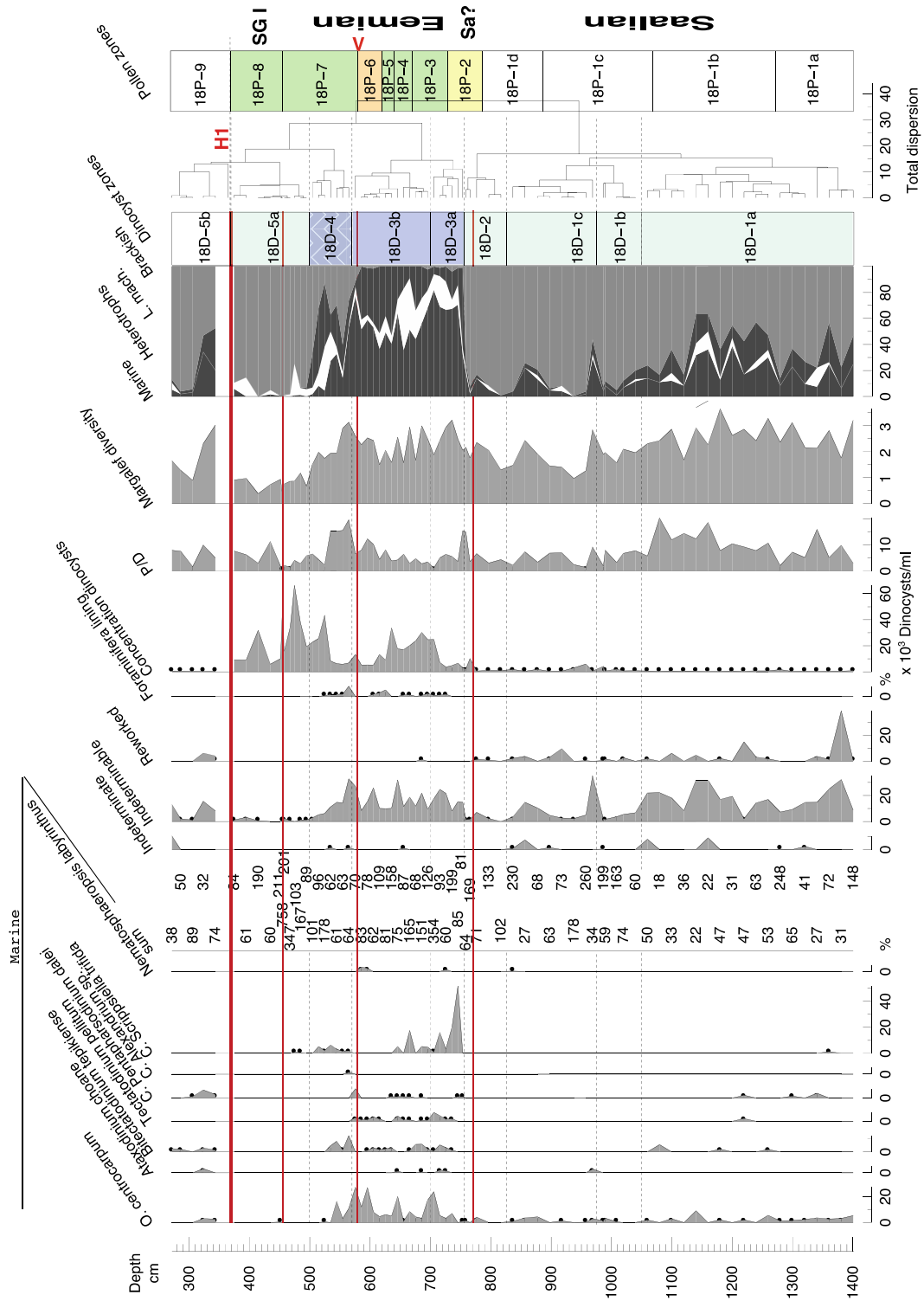
Marmara Sea, Imrali basin, 291 m water depth, core CS18, dinoflagellate diagram

Analyses: S. Leroy



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

Analyses: S. Leroy



347

348

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

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

351 the warm indicators in the marine group.

352

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

359

360 Zone 18D-1, 1400.5 – 826.5 cm depth: This is a very long zone of nearly 6 m. It is
361 dominated by brackish Pontocaspian taxa such as *P. psilata*, *S. cruciformis* A, *S.*
362 *inaequalis* and *Caspidinium rugosum rugosum*, with a significant occurrence of *L.*
363 *machaerophorum* ss. In general, concentrations are relatively low, mostly below
364 1000 cysts/ml. Reworked cysts are frequent. This zone has been subdivided into
365 three subzones, with subdivisions as followed. Subzone 18D-1a (1400.5-1049.5 cm)
366 is dominated by *S. inaequalis*. *L. machaerophorum* ss, *P. psilata* and *S. cruciformis*
367 A are abundant. A range of marine and brackish Pontocaspian taxa occur side by
368 side. Indeterminable dinocysts are abundant. A large peak of reworked cysts occurs
369 at 1381 cm depth. The P/D ratio remains high all through this zone, with maximal
370 values towards the end, i.e. a maximum of up to above 20. Concentrations are very
371 low, often lower than 500 cysts per ml. Subzone 18D-1b (1049.5-977 cm): This zone
372 is characterised by a bell-shaped curve of *Impagidinium* sp. A. *S. inaequalis* is still
373 abundant. *P. psilata* displays increasing values across this zone. In subzone 18D-1c
374 (977-826.5 cm), *P. psilata* reaches maximal values, i.e. 54%. The percentages of *S.*
375 *inaequalis* are still high. Less diverse marine forms are observed.

376

377 Zone 18D-2, 826.5-756 cm depth: This zone is characterised by many changes. *P.*
378 *psilata* decreases and *S. cruciformis* A increases up to a brief maximum of 59%. *S.*
379 *cruciformis* C is very abundant. A small bell-shaped curve of *S. belevius* occurs. *S.*
380 *inaequalis* values drop. Very few marine taxa subsist. Concentrations are slightly
381 increasing. This zone represents a transition that is truncated by a hiatus, recognised
382 as one the strongest statistical changes according to CONISS.

383

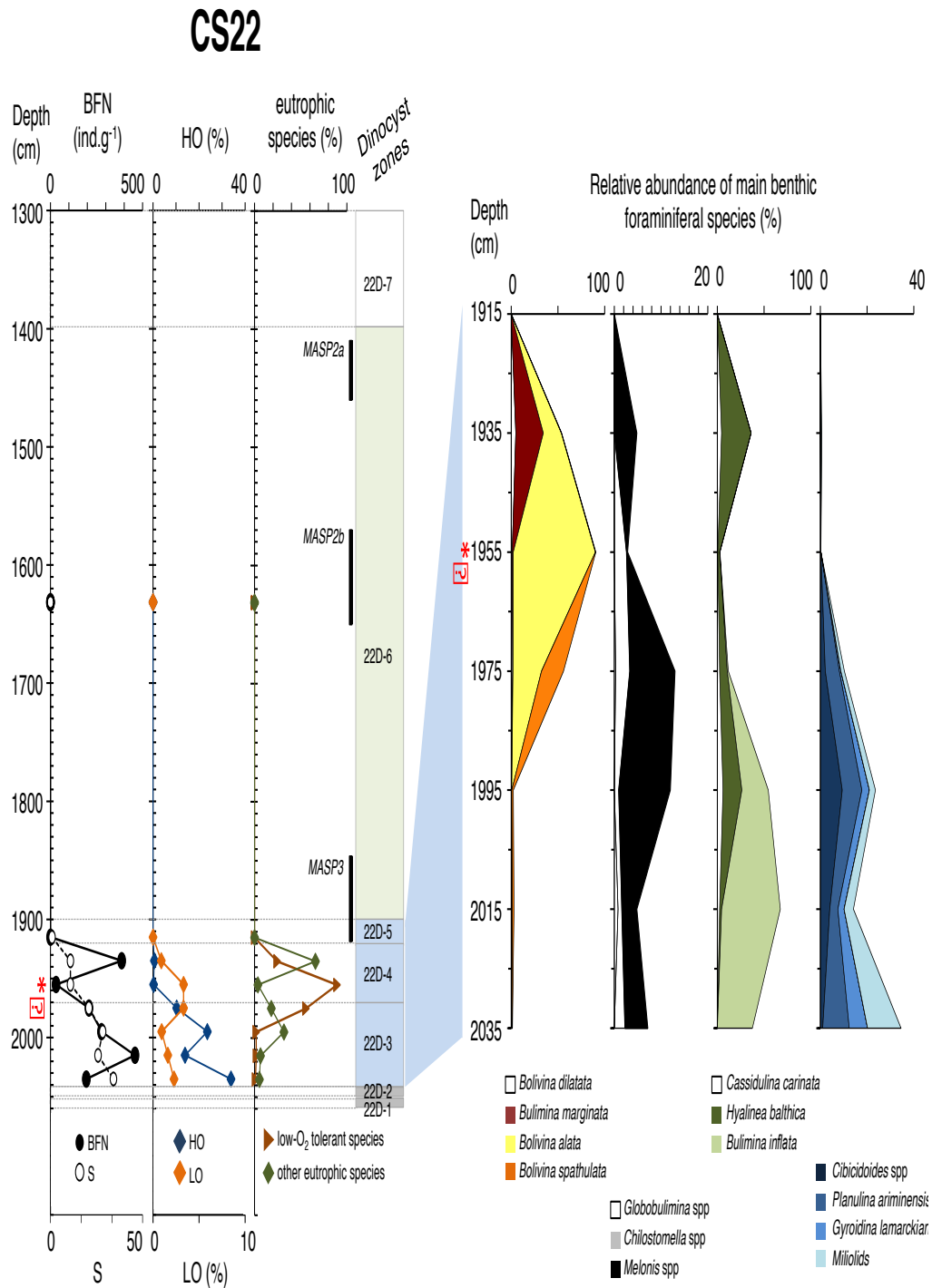
384 Zone 18D-3, 756-570 cm depth, is a totally different zone, with *L. machaerophorum*
385 (up to 52%) and *Spiniferites* sp. (up to 30%) widely dominant. Brown baggy cysts are
386 very abundant in this zone and the following one, i.e. up to 30%. Many marine taxa
387 are present. *S. ramosus*, *O. centrocarpum*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

395

396 Zone 18D-4, 570–499.5 cm depth, appears as a transition, with mixed marine and
397 brackish Pontocaspian assemblages. *P. psilata* and *S. cruciformis*, quasi absent in
398 zone 18D-3, reappear. *L. machaerophorum* (with a peak of other forms of *L.*
399 *machaerophorum* towards the end of this zone) and brown baggy cysts are still

400 abundant. Marine taxa remain present, but their values decrease. Concentrations,
401 after moderate values, become high again. P/D is maximal, i.e. up to just below 20.
402
403 Zone 18D-5, 499.5-271.5 cm, is split in two by the seismic red-H1 reflector;
404 moreover, a barren sample occurs a few cm above this stratigraphic hiatus. In
405 subzone 18D-5a, 499.5-369 cm depth, *S. cruciformis* A very largely dominates with
406 percentages up to 95%. Besides *P. psilata* and brown baggy cysts, hardly any other
407 taxa are present. It is noteworthy that *L. machaerophorum* has only rare occurrences
408 in this subzone, and only at its base. This is a very low diversity subzone (minimum
409 in the Margalef index). Concentrations, after being very high, become high, i.e.
410 66,000 to 10,000 dinocysts/ml. Subzone 18D-5b, 369-271.5 cm depth: The
411 assemblages are similar to subzone 18D-5a (dominance of brackish Pontocaspian
412 taxa), but brown baggy cysts have quasi disappeared and are replaced by *L.*
413 *machaerophorum* that have re-appeared. A moderate range of marine taxa is
414 present. Concentrations are very low as in subzones 18D-1a. Reworked elements
415 are common in the lower half of this subzone, just above the barren sample.

416 5.3 Other proxies in core CS22



417

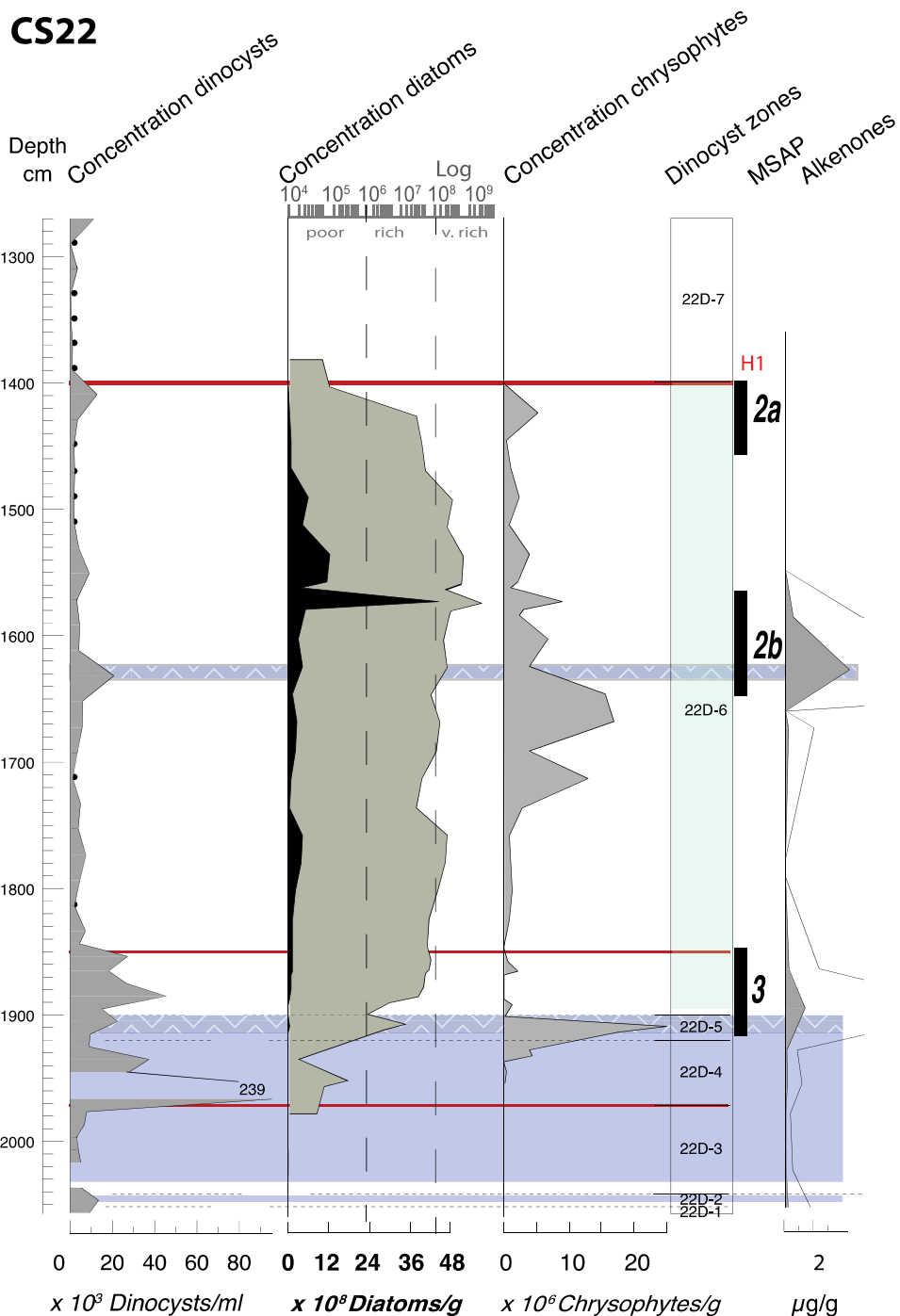
418 **Figure 5:** Benthic foraminiferal record of core CS22 with dinozones and Marmara
 419 sapropels (MSAP). Right panel: relative abundance of main species (>3%) between
 420 2035 and 1915 cm depth. Left panel: Benthic Foraminiferal Number, BFN (ind.g⁻¹),

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

426

427 Quantitative investigation of benthic foraminiferal assemblages relies on the
428 recovery of a well-preserved, abundant (164 to 361 specimens) and diverse (>60
429 species overall) faunal content between 2035 and 1915 cm depth ([Table SI1](#)). Apart
430 from a few individuals (< 10) belonging to neritic taxa (e.g. *Ammonia*, *Discorbinella*,
431 *Elphidium*), assumed to be post-depositional, foraminiferal assemblages
432 overwhelmingly consist of species from marginal to bathyal environments. The BFN
433 ranges from 30 to 459 ind.g⁻¹, with lowest values recorded at 1955 cm. Dominant
434 species (>5%) belong to *Bolivina* spp, *Bulimina* spp, *Cassidulina carinata*,
435 *Cibicidoides* spp, *Hyalinea balthica* and *Melonis* spp and indicate fully marine
436 conditions at the seafloor at the base of the core. The foraminiferal record, though
437 short, exhibits a marked species shift between 1995 and 1975 cm ([Fig. 5](#)). From
438 2035 to 1995 cm, species number is highest, ranging from 31 to 41. Faunas are
439 strongly dominated by *Bulimina inflata* (28-63%), together with various associated
440 species including *Cibicidoides* spp (1-9%), *Gyroidina lamarckiana* (3-8%), *Planulina*
441 *ariminensis* (3-11%) and miliolids (3-14%). Minor species (<3%) include *Bulimina*
442 *aculeata*, *Uvigerina mediterranea* and *U. peregrina*. From 1975 up to 1935 cm, most
443 of the afore-mentioned species disappear from the record. Dominant species consist
444 mainly in *Bolivina alata*, *B. spathulata* and *Bulimina marginata*, which collectively
445 account for 49 to 90% of the assemblage. Concomitant to this faunal shift, species

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



453

454 **Figure 6:** Palaeoproductivity: concentrations of dinocysts, diatoms (in decimal and
 455 log scales) and chrysophytes, with dinozones, alkenones and Marmara sapropels
 456 (MSAP) in core CS22. The second marine phase at 1633-1632 cm depth is indicated
 457 by a box with a blue zigzag highlight.

458

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

471 The values of alkenones are usually below detection limit, except at 2050-
472 1865 cm and 1633-1592 cm, reaching respective maxima of 1.3 µg/g at 1896 cm
473 and 4.4 µg/g at 1633 cm depth (Fig. 6).

474 **6 Interpretation and discussion**

475 6.1 Core CS22: taphonomy, surface water condition and chronology

476 6.1.1 Taphonomy

477 *Below the sapropel*

478 At 2027 cm in zone 22D-3, the sample is unexpectedly barren in dinocysts. It
479 is one of the samples of zone 22P-2 with abundant *Pinus* pollen (76%) (Leroy et al.,
480 submitted). The lithology shows a bioturbated sediment. Sediment disturbance may
481 also contribute to explain the rapidly changing spectra in the short zones 22-D1 and

482 D2. The sharp zone limit at 1920 cm (between zones 22D-4 and 5) corresponds to a
483 very thin silt layer and a change from bioturbated to unbioturbated sediment, that is
484 the start of the sapropel attributed to MSAP-3 (1920-1853 cm).

485

486 *In the sapropel and above*

487 Dinocysts (as well as pollen) are well preserved and their concentrations are
488 maximal as observed in other sapropels ([Cheddadi and Rossignol, 1995](#); [Kotthoff et](#)
489 [al., 2008](#)) ([Fig. 6](#)).

490

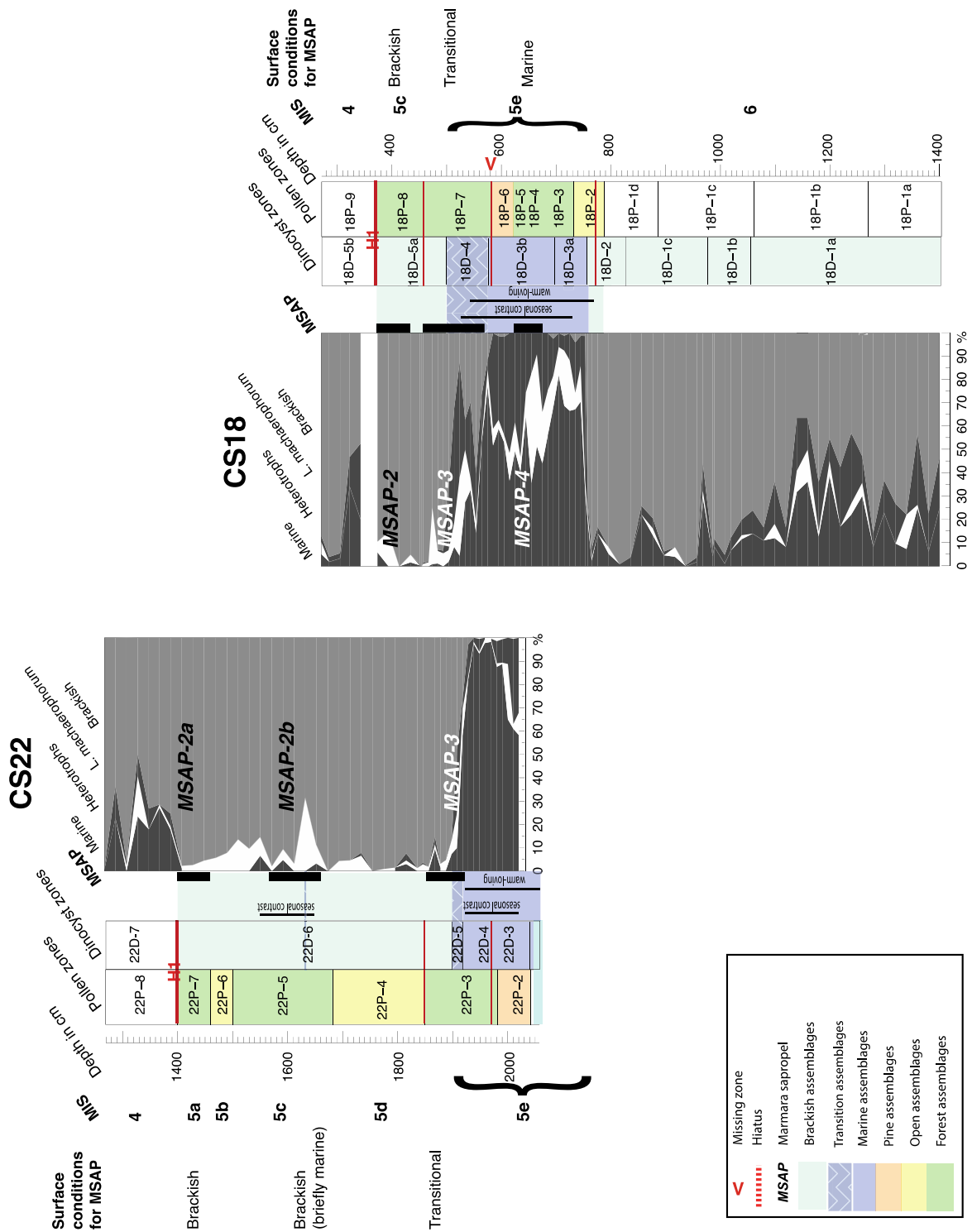
491 *Impact of red-H1 seismic reflector*

492 With the red-H1 reflector (a hiatus), a long-term change in taphonomy is
493 initiated in the SoM. It is seen in the reworking of older marine layers and of soils.
494 Firstly, some foraminifera (linings) are brought in the sediment. Secondly, the non-
495 pollen palynomorphs *Glomus* (a fungal spore) is present, due to inwash of soil
496 erosion products ([Leroy et al., submitted](#)). This reflects a turning point in the aquatic
497 conditions of the SoM.

498 *6.1.2 Sea-surface and bottom conditions*

499 The depths of 2042-1920 cm (dinocyst zones 22D-3 and 4) are fully marine as
500 clearly seen not only in the dinocyst assemblages but also in the alkenone values
501 and the benthic foraminiferal content ([Fig. 3, 5 and 7](#)).

502



503 **Figure 7:** The cumulative diagrams sorting dinocyst groups/taxa by salinity. The four
 504 groups are the marine ones (black) to the left, the brackish Pontocaspian dinocysts
 505 (grey) to the right), and in the middle the heterotrophic cysts (white) and the
 506 *Lingulodinium machaerophorum* forms (black). Dinocyst zones: blue highlight for
 507 marine, light blue for brackish Pontocaspian. The distribution of warm-loving and

508 seasonal contrast indicators as vertical black lines. Pollen zones: green highlight for
509 high arboreal taxa (minus *Pinus*) zones and yellow for low arboreal taxa % and high
510 *Pinus* % zones. Horizontal bold V for pollen and dinocyst zones absent in core
511 CS18.

512

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

519 In zone 22D-4 at 1955 cm, high values of cysts of *P. dalei* and *S. trifida*
520 suggest an increase in temperature seasonality and lower salinity (e.g. Head et al.,
521 2006; Mudie et al., 2017). In contrast, very high concentrations point at surface
522 waters rich in nutrients and possibly bottom waters with low oxygen (Mudie et al.,
523 2017). This may be due to alternating deep-water upwelling - when the heavy saline
524 marine waters penetrate the bottom of the basin from the Aegean Sea - and
525 stratification. This occurs below the horizon with high TOC (sapropel MSAP-3 at
526 1920-1853 cm depth and bottom anoxia, Fig. 2).

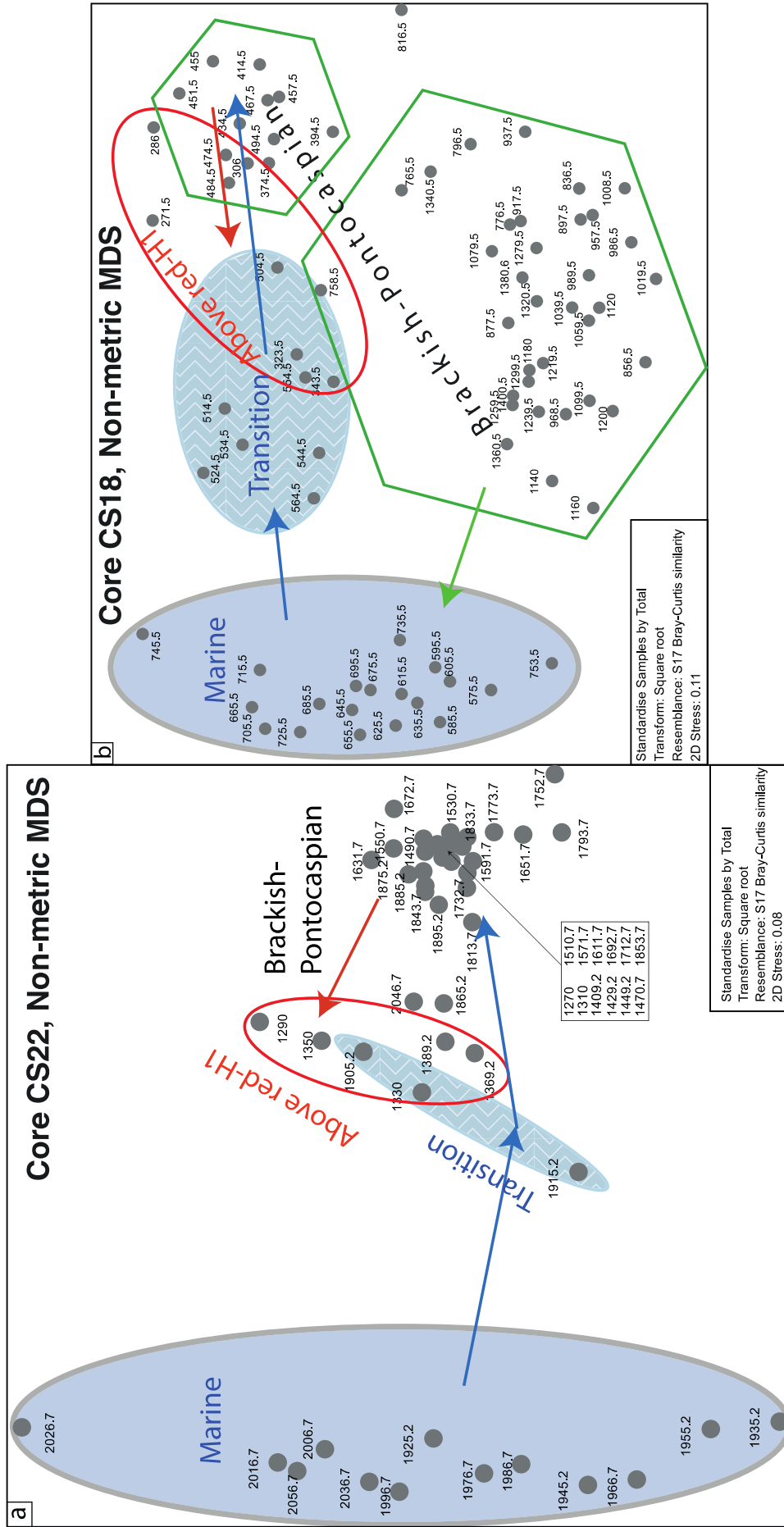
527 To tackle environmental changes at the seafloor, the approach of Cornuault et
528 al. (2016, and references therein) was followed and foraminiferal species were
529 gathered into three functional groups assumed to be representative of contrasted
530 oxygen and trophic conditions at the seafloor: (1) species adapted to high oxygen
531 conditions (HO), (2) low-oxygen indicators (LO), and (3) species adapted to
532 eutrophic conditions (Fig. 5). Ecological assignment is given in Table SI2. The major

533 faunal shift observed at 1995 -1955 cm suggests contrasted oxygen conditions
534 between zones 22D-3 and 22D-4, possibly related to the establishment of eutrophic
535 conditions. In zone 22D-3, benthic foraminiferal assemblages are ecologically
536 diverse and consist of a mix of epibenthic and various endobenthic species. This
537 sequence and the high abundance of HO species suggest a mesotrophic
538 environment where well-oxygenated conditions prevailed at the seafloor. By contrast,
539 HO indicators disappearance in zone 22D-4 together with high abundance of species
540 adapted to high organic fluxes points to reduced oxygen level during this phase.
541 Assemblages from this zone are successively dominated by species adapted to
542 contrasted oxygen conditions, suggesting variable oxygenation level during zone
543 22D-4. In particular, the BFN decrease, high dominance of eutrophic species tolerant
544 to low-oxygenation and slight increase of LO species suggest dysoxic conditions at
545 the beginning of this period at 1935 cm, high dominance of species adapted to
546 moderate oxygen depletion strongly suggest environmental changes recorded at the
547 seafloor directly are related to eutrophication rather than ventilation changes. Overall
548 the foraminiferal assemblages confirm the presence of oxygen at the bottom of the
549 sea in zones 22D-3 and 22D-4.

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

556

557

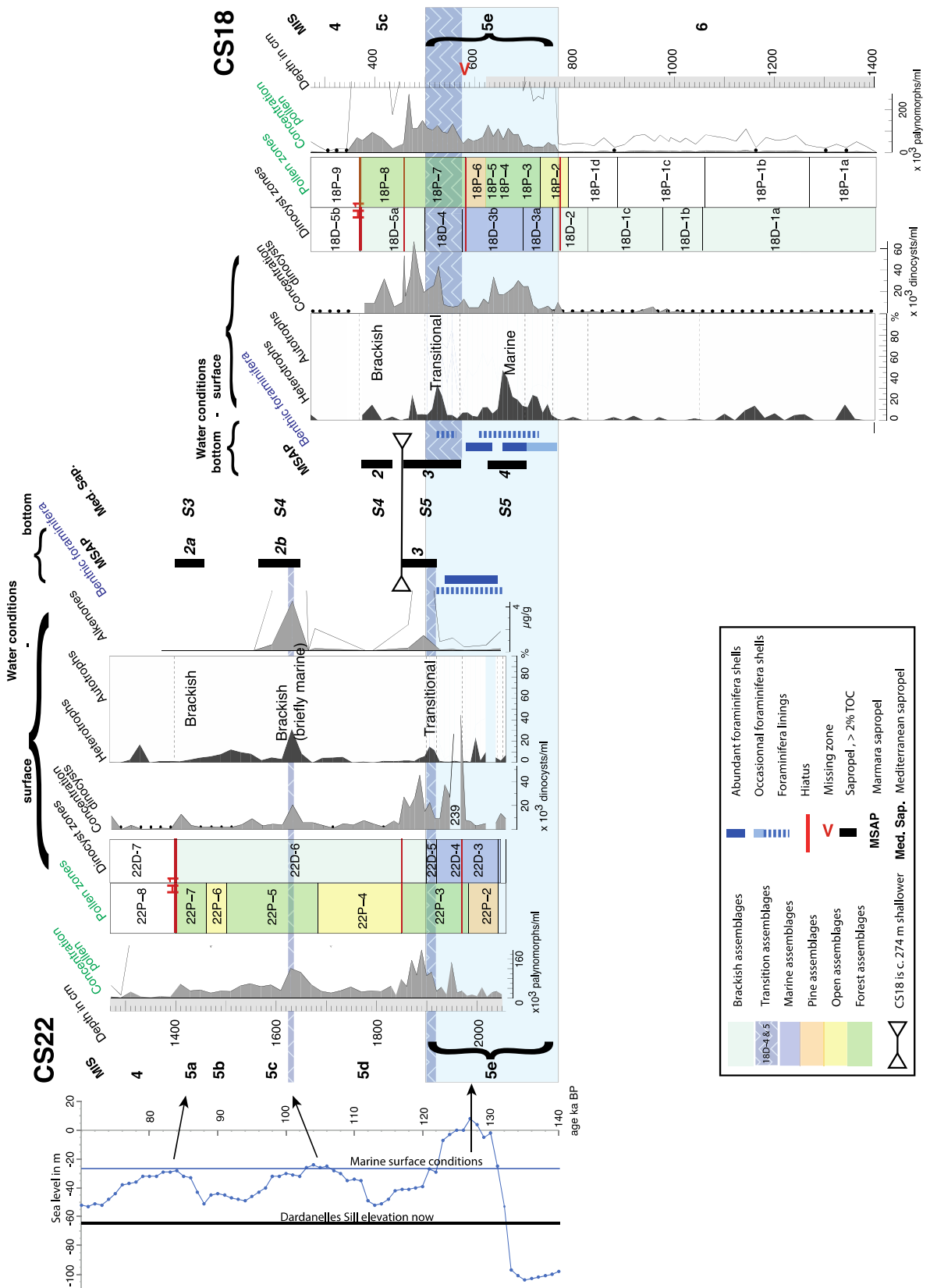


559 **Figure 8:** Results of the NMDS analysis on dinocyst percentages from cores CS22
560 (a) and CS18 (b). Depth in cm.

561

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

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



573

574 **Figure 9:** Water conditions. At the surface, dirocoyst concentration and the ratio
 575 heterotrophic and autotrophic dirocoysts (H/A). In the middle for core CS22, also the

576 alkenones. For the bottom: Marmara sapropels (MSAP, >2% TOC), benthic
577 foraminifera shells and benthic foraminifera organic lining. The top of the marine
578 horizon in core CS18 (hiatus between zones 18P-7 and 8) is c. 274 m deeper than
579 that in core CS22 (hiatus between zones 22P-3 and 4). Horizontal bold V for pollen
580 and dinocyst zones absent in core CS18. For linking with the sister paper on pollen
581 ([Leroy et al., submitted](#)), pollen zones and pollen concentrations are also shown. By
582 combining the two light grey vertical boxes, the most complete record is obtained.
583 Sea-level curve from [Rohling et al. \(2021\)](#) with the position of the Dardanelles Sill
584 elevation now (-65 m) and the sea-level threshold where transition from brackish to
585 marine conditions occurred.

586

587 However this long 22D-6 zone is not homogenous. At 1632 cm, a peak of
588 heterotrophic dinocyst is observed. It seems to correspond to the second peak in
589 alkenones, indicating marine influence, at 1633 cm ([Fig. 9](#)). According to pollen
590 analysis, it is in an interstadial (Saint-Germain I). Thus, a brief marine incursion is
591 probable. It is located at the beginning of sapropel MSAP-2b at 1653-1570 cm.

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

596 *6.1.3 Elements of chronology*

597 It is suggested that the high level of MIS 5e ([Bard et al., 1996](#); [Koeling et al.,](#)
598 [2009](#)) is represented here by zones 22-D3 to 5. In the whole of MIS 5, it is mainly in
599 the maximum sea level of MIS 5e that the Black Sea would have been connected to
600 the Global Ocean and the Turkish Straights would have the two-way exchange of

601 water as today. It was established earlier that stratigraphic succession immediately
602 above the red-H1 reflector belongs to MIS 4 (Kende, 2018; Çağatay et al., 2019),
603 which can now be shown to reflect a brackish-Pontocaspian environment. During
604 MIS 4, the SoM was an oligotrophic lake, with no evidence of water body
605 stratification, and was almost certainly disconnected. During the late phases of MIS
606 5, the inflow of Mediterranean water was restricted and this probably required a
607 Dardanelles Sill depth higher than the present sill depth. We will come back to this
608 point in the discussion and here only consider the phasing of the marine to brackish
609 transition. In the age model proposed in Çağatay et al. (2019), this transition occurs
610 after MIS 5d (in the 110 to 100 ka interval) at a time of sea-level rise, and that may
611 be considered unlikely. This transition more probably occurred concurrently with sea-
612 level drop either later in MIS 5c or during the transition from MIS 5e to 5d. Assigning
613 this transition to later in MIS 5c would keep the sapropel correlation proposed by
614 Çağatay et al. (2019) but imply that MIS 5d is not observed in the pollen record in
615 either core (Leroy et al., submitted).

616 6.2 Core CS18: taphonomy, surface water condition and chronology

617 6.2.1 Taphonomy

618 Before the transition 18D-2 to 3 (756 cm) towards marine assemblages, thus
619 already at the beginning of pollen zone 18P-2 (786.5-730.5 cm), one may see
620 sudden changes. This is firstly a sediment change from homogenous to finely
621 laminated around 772 cm and two round pebbles (of river origin?) are found at 775
622 and 758.5 cm, in between them one finds a 2 mm-thick silt layer (773 cm). Moreover,
623 at 771 cm a sharp increase of the concentration in pollen and spores (from <5000 to
624 > 25,000 palynomorphs/ml) and at 762 cm a sharp and long-lasting reduction of the

625 reworked pollen. The sharp statistical limit at 756 cm in the dinocyst diagram is 17
626 cm above the 2 mm-thick silt layer. These observations are in support of an increase
627 of water level with the influx of marine water from the Aegean Sea (Çağatay et al.,
628 2009) leading to sediment disturbance and taphonomical changes.

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

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

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

642 6.2.2 Sea-surface conditions

643 The transition from brackish Pontocaspian to fully marine conditions (zones
644 18D-2 to 3, main change highlighted by CONISS) seems to occur very abruptly at
645 756 cm depth (Fig. 4 and 7). Then in zone 18D-3, the conditions are only fully
646 marine. Zone 18D-4 is a transitional period, back towards a brackish-Pontocaspian
647 environment. This transition zone is also well marked in the NMDS results (see
648 green arrow in Fig. 8). The surface waters of subzone 18D-5a are clearly brackish-
649 Pontocaspian. Subzone 18D-5b is mostly brackish-Pontocaspian but some

650 influences from the sea are felt, although as in core CS22 a taphonomical impact is
651 very important. Below 756 cm, the assemblages of zones 18D-1 and 2 are brackish-
652 Pontocaspian, with significant reworking in zone 18D-1 both in dinocysts and in
653 pollen.

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

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

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

664 *6.2.3 Elements of chronology*

665 Zones 18D-3 (marine taxa and low P/D) and 18D-4 (transition) reflect the MIS
666 5e high sea levels. The warm-loving taxa are present in the same zone and confirm
667 interglacial conditions as deduced from pollen analysis. However, it is likely that this
668 zone is not complete as a characteristic dinozone in core CS22 (high percentages of
669 cysts of *P. dalei* and cysts of *S. trifida*) is missing from it (Leroy et al., submitted).

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

674 In conclusion, the chronology proposed here represents an in-depth revision
675 of that in [Çağatay et al. \(2019\)](#), with especially the occurrence of a larger part of MIS
676 5e and of MIS 5c and fits the chronology based on geochemistry and pollen analysis
677 proposed in [Leroy et al. \(submitted\)](#).

678 6.3 Comparison of the dinocyst records within the Sea of Marmara and to the 679 Black Sea

680 6.3.1 Diversity

681 The dinocyst assemblages are similar in core CS22 and core CS18, with
682 slightly more marine taxa in CS22 and brackish-Pontocaspian taxa in CS18,
683 reflecting the difference in the extension of glacial versus interglacial periods covered
684 in the two sequences ([Table SI3](#)). The total number of taxa is larger in core CS18;
685 the latter because of the long MIS 6 record. Taxa at play in core 22-GC3 in the SE
686 Black Sea are similar to those in the two Marmara cores, although the Black Sea
687 seems to have more heterotrophic cyst diversity and the SoM more brackish-
688 Pontocaspian taxon diversity ([Table SI3](#)). Modern and Holocene records across the
689 Black Sea show geographic differences in the distribution of heterotrophic taxa
690 ([Mudie et al., 2017](#)). This may be due to variations in water turbidity, limitations to
691 photosynthesis, or possibly competition with other phytoplankton groups.
692 Remarkably, *S. inaequalis* and *S. trifida* are two particularities of the SoM versus the
693 Black Sea.

694 6.3.2 MIS 5e

695 In core 22-GC3 (SE Black Sea), the transition from brackish Pontocaspian to
696 marine conditions during MIS 5e is dated at 128 ka ([Shumilovskikh et al., 2013](#)). The
697 change is progressive contrary to that in core CS18, underlining the possibility of a

698 hiatus in core CS18 at the beginning of the marine phase (see the earlier taphonomy
699 section) linked to flooding by marine waters. The subsequent peak of nutrients in
700 core CS18 is possibly due to shore erosion as sea level rose. The very high
701 percentages of cysts of *P. dalei* and *S. trifida* in zone 22D-4 find no equivalent in
702 core CS18. The transition from marine to brackish (mixed assemblages) is made
703 over 20 cm only within zone 22D-5, whereas it is much longer in core CS18 (zone
704 18D-4, 70.5 cm). These two observations suggest other significant hiatus and deeper
705 waters in the west that were invaded quickly by saline waters.

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

713 The end of the marine phase is not presented in core 22-GC3 due to a
714 turbidite (Wegwerth et al., 2018). The end of the marine phase may however be
715 studied in the two Marmara cores here and seems to have been progressive with a
716 period when the two types of dinoflagellate assemblages co-existed (zones 22D-5
717 and 18D-4), i.e. marine and brackish.

718 It is noteworthy to underline the peaks of *S. trifida* both at the beginning of the
719 marine phase (zone 18D-3a) and at the end of it (zone 22D-4), in both cases
720 attributed to fluctuating salinities between marine and brackish-Pontocaspian. At the
721 end of the marine period in core CS22, in addition, the high values of cysts of *P.*

722 *dalei* may be attributed to terrigenous input, probably related to water level lowering,
723 as proposed for the Holocene (Verleye et al., 2009).

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

734 6.3.3 MIS 6 and 4

735 The NMDS analysis shows that MIS 6 and MIS 4 dinocyst assemblages are
736 clearly different, although both brackish-Pontocaspian (Fig. 8), i.e. more *P. psilata*
737 and clear presence of *S. inaequalis* and *Caspidinium* in MIS 6, contrasted with more
738 *S. cruciformis* in MIS 4. Some common points are noted between the MIS 6 and MIS
739 4 assemblages of the current study and the MIS 3 assemblages of other Marmara
740 cores (Mudie et al., 2004): all show a mix of dominant brackish-Pontocaspian
741 dinocysts with some marine ones and all have low to very low cyst concentrations.
742 The latter may be due to high sedimentation rates diluting the fossil content. Erosion
743 is active in glacial stages, as seen by the percentages of reworked elements that are
744 very high in the MIS 6 of core CS18 and in the MIS 4 of cores CS22 and CS18.
745 Other investigators often do not count them, neither published them, but they are a
746 significant indicator of erosion and inwash. Often reworking also shows proximity to

747 shores due to low water level. The water level must have been relatively low, most
748 likely lower than at any time in MIS 5. The brackish-Pontocaspian periods during MIS
749 5 are clearly different from those of MIS 6 and 4, in that the latter ones have a lot of
750 reworking and contain abundant signs of erosion (Fig. 3, 4 and 8).

751 Core CS18 in the Imralı basin is not far from the North Imralı canyon that the
752 Kocasu carved during lowstands (Çağatay et al., 2015a and b). This relative
753 proximity to a river (if not to the shore) and submarine canyon may also explain the
754 high amounts of reworked elements and the different taphonomy during the glacial
755 times (Pope et al., 2022).

756 It is very likely that, as suggested in Hoyle et al. (2021) based on Sr ratio in
757 the Black Sea, no marine connection occurred during the glacial periods for the SoM.
758 Several indicators (oxygen isotopes of the Sofular Cave along the Turkish coast of
759 the Black Sea (Badertscher et al., 2011) and Sr isotope ratio (Hoyle et al., 2021))
760 suggest an overflow from the Caspian Sea to the Black Sea during MIS 6.
761 Unfortunately, dinocyst assemblages cannot contribute evidence to this debate on
762 water source.

763 For MIS 6, a progressive, even slow, freshening after marine conditions in the
764 previous interglacial (MIS 7) was suggested on the basis of divergent trends in
765 isotopic values, i.e. $\delta^{18}\text{O}$ becomes more negative while $\delta^{13}\text{C}$ becomes more positive
766 (Çağatay et al., 2009). This may explain why here the marine dinocyst group is
767 better represented in subzone 18D-1a than subzones 18D-1b and c, even 18D-2.
768 This also suggests that the base of the core may have just missed the top of MIS 7.
769 As a significant contribution of reworking of marine palynomorphs (as shown here) is
770 revealed, no marine connection would have occurred during MIS 6 (zones 18D-1
771 and 2). For that matter, at the transition from the marine phase to the lake phase at

772 MIS 5 to 4, [Grall et al. \(2014\)](#) and [Aziz et al. \(2019\)](#) suggest slope failure and strong
773 turbidity currents entraining sediment that was stored on the shelf to deeper areas of
774 the SoM, as the sea level fell. Recent investigations (modern discharge and
775 geological data) aiming at disentangling the origin of the sediment in the SoM during
776 low levels have however demonstrated that southern river input was more important
777 than suspended load through one or both straits ([Hiscott et al., 2021](#)). Moreover, the
778 latter authors further estimate that river input was higher than mass-wasting events
779 and shoreface erosion. Nowadays, and most likely in the past, the Kocasu is the
780 largest southern river and it drains a large basin in NW Anatolia ([Kazancı et al.,](#)
781 [2004](#)).

782 The water levels in the Black Sea were most likely low during the cold parts of
783 MIS 6. However, three major meltwater pulses of the Eurasian icesheets occurred;
784 but it is not known if the lake level rise of the Black Sea was sufficient to cause an
785 overspill to the SoM although it is likely ([Wegwerth et al., 2019](#)). So far these pulses
786 cannot be seen in our record. For completeness of the record, it is noteworthy to
787 underline that the glacials (MIS 6 and 4) in the Gulf of Corinth were generally
788 brackish ([Fatourou et al., 2023](#)).

789 6.4 Sapropels

790 *6.4.1 Surface conditions and sapropel formation*

791 The sapropels deposited in SoM during MIS-5 clearly formed under three
792 different surface water conditions ([Fig. 7](#)). MSAP-2 formed under brackish surface
793 conditions (except briefly on a few cm within MSAP 2b), MSAP-4 under clearly
794 marine surface conditions and MSAP-3 in a period of transition from marine to
795 brackish surface conditions.

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

806 The origin of the sapropel formation in the SoM, the decreasing surface
807 salinities and the decrease of the oxic condition at bottom over MIS 5 may be the
808 reason why palaeoproductivity based on the H/A ratio is clearly higher in MSAP-4
809 only.

810 *6.4.2 Bottom conditions and sapropel formation*

811 In core CS22 below MSAP-3, the gradual shift of foraminifera assemblages,
812 starting with genera needing oxic conditions towards suboxic-tolerant genera, is an
813 effect of oxygen loss caused by the formation of water column stratification and
814 organic matter mineralisation ([Fig. 9](#)). It is worth noting that the occurrence of
815 foraminifera linings is only in the marine period. Accordingly in the marine phase of
816 core CS22, the linings occur from the core base and stop at the beginning of the
817 sapropel MSAP-3 at 1920 cm. Their occurrence is similar to that of the foraminifera
818 shells ([Fig. 9](#)).

819 In contrast, in core CS18, foraminifer linings are present at 725.5-605.5 and
820 564.5-524.5 cm depth, thus in sapropels MSAP-4 and 3. Moreover foraminifer shell

821 distribution (Çağatay et al., 2019) (Fig. 9) is like that for the linings, although
822 somewhat shorter. In core CS18, shells are found in the first part of MSAP-4 and
823 between MSAP-4 and 3.

824 This foraminifera distribution in the sapropels is therefore questioning the
825 existence of full anoxia in MSAP-4 (core CS18) and in part of MSAP-3 (mostly in
826 core CS18, no foraminifera in core CS22). For MSAP-2, it is unclear whether the
827 deep water remained sufficiently saline for benthic foraminifera to live.

828 *6.4.3 Sapropel trigger in the Sea of Marmara*

829 Combining the information from bottom and surface conditions (especially
830 dinocysts, foraminifera shells and organic linings), it may be shown that the four
831 Marmara sapropels (4, 3, 2b and 2a) clearly formed over times with decreasing
832 marine influence and increasing stratification. Moreover they are different from each
833 other. Only MSAP-2 may have had fully anoxic conditions during the whole sapropel
834 time.

835 The difference between the two cores may be possibly explained by the
836 shallower position of core CS18 in comparison to core CS22. The water depth
837 difference is 260 m, adding another c. 14 m (at the end of MSAP-3) due to core
838 depth difference, thus a total of c. 274 m difference (Fig. 9). The location of the
839 shallower core CS 18 would make it harder to fall within the depths affected by full
840 anoxia and would justify the occurrence of foraminifera in MSAP-4 and 3 in a dysoxic
841 area. Our results also show that the anoxia did not rise up to 291 m water depth. In
842 the east Mediterranean Sea, sapropels S5 (MIS5e) and S6 (MIS 6e) contained
843 foraminifera. It is interesting to note that during MSAP-1 suboxic conditions extended
844 over the shelf to depths of -75 m (Çağatay et al., 2009). These results (and the
845 difference between cores CS22 and CS18) may be important to further understand

846 anoxia (or euxinia) development during the Eemian sapropel formation in the Black
847 Sea, where it has been shown to be stronger than the Holocene one ([Wegwerth et](#)
848 [al., 2018](#)).

849 The Marmara sapropels occur during warm periods, evidenced in the current
850 joint pollen-dinocyst investigation i.e. Eemian, Saint-Germain I and II, which more
851 likely facilitated algal blooms. Consequently, the proposed correlation between the
852 Marmara sapropels and the Mediterranean sapropels becomes: MSAP-4 – S5,
853 MSAP-3 – S5 too, MSAP 2b – S4 and MSAP-2a – S3 ([Fig. 9](#)), based on ages ([Grant](#)
854 [et al., 2016](#)). The Marmara sapropels do not seem to be preconditioned by
855 astronomical forcing as for the Mediterranean ones ([Rohling et al., 2015](#); [Grant et al.,](#)
856 [2016](#)). Thus further to the astronomical cause of sapropel formation in the
857 Mediterranean Sea, other sapropel formation triggers may be sought for the SoM. A
858 specific trigger mechanism in the SoM is an inflow of marine water causing water
859 column stratification ([Çağatay et al., 2009](#)).

860 6.5 Terrestrial and marine transitions in the Marmara and the Black seas

861 Understanding the facies transitions in the SoM or Lake requires taking into
862 account the hydrodynamic processes and how they control sediment deposition and
863 erosion in the straights. Obviously, the outflow from the Black Sea to the Sea of
864 Marmara is governed by river inputs to the Black Sea and evaporation/precipitation.
865 When the water budget of the Black Sea is positive, it will inevitably overflow into the
866 SoM and then most likely into the Mediterranean Sea as the SoM has a
867 comparatively small surface area for evaporation. On the other hand, the occurrence
868 of a counter current of sea water from the Mediterranean Sea to the SoM when sea
869 level is above the Dardanelles Strait and to the SoM to the Black Sea when sea
870 level is above the Bosphorus Sill cannot be taken for granted. For instance, in the

871 present day with a sill depth of -35 m in the Bosphorus Strait, the seawater bottom
872 current is blocked when the surface current exceeds $30,000 \text{ m}^3 \cdot \text{s}^{-1}$ (Sannino et al.,
873 2017). This situation occurs during transient events in the present conditions, but
874 more generally the hydrodynamics of the straits implies that the influx of seawater
875 is a function of water depth at the sill and of the freshwater outflow, which is in turn
876 governed by river inputs and evaporation/precipitation. Moreover, in steady state, the
877 only way out for seawater entering the Dardanelles Strait is by mixing with the
878 fresher surface water layer in the SoM or in the Black Sea. In the present day, and
879 considering yearly averages, the bottom current flux through the Dardanelles
880 Strait is only about 12 % larger than the bottom current flux through the Bosphorus
881 Strait (Beşiktepe et al., 1994). It follows that the halt of northward flux through the
882 Bosphorus Strait would also greatly reduce the seawater flux into the SoM and
883 thus increase the residence time of the SoM bottom water. The residence time of
884 deep water in the SoM varies from two years in the Tekirdag and the Central Basin
885 to 10-20 years in Çınarcık Basin (Beşiktepe et al., 1994), and this causes an
886 eastward decrease of oxygen concentration from 60 to $8 \mu\text{mol l}^{-1}$ (Henry et al.,
887 2007). Increasing the residence time will further reduce oxygen content, which may
888 in turn favour sapropel deposition.

889 Variations of sill elevation depend on tectonic movements, erosion and
890 sedimentation. The shores of the Dardanelles Strait have been uplifting at a rate
891 estimated in the $0.25\text{-}0.75 \text{ mm yr}^{-1}$ range (Yaltırak et al., 2002). This may suggest
892 that the level of the sill has risen with time. However, the Dardanelles Strait is at
893 least in part structurally controlled (Yaltırak et al., 2000; Gökaşan et al., 2008) and
894 active faults along part of its shore may locally cause subsidence of the channel
895 floor. Two sedimentary units have been deposited on an erosional surface cutting

896 through Miocene and Pliocene sediments and their geometry gives clear indications
897 on the erosion and sedimentation processes ([Gökaşan et al., 2008](#)). The lower unit
898 is thickest (up to 80 m) at the mouths of rivers discharging in the Dardanelles
899 Straight where they formed deltas that were subsequently incised. The upper unit is
900 composed of Holocene sediment drifts that are thickest where the channel is wide
901 and are incised by the active bottom current in the narrow parts. It thus appears that
902 the strength of the bottom current in the Dardanelles Straight, which is in the
903 present-day conditions controlled by the dynamics of the Bosphorus Straight,
904 redistributes sediments and prevents clogging of the channel. All the above
905 observations and reasoning leads to hypothesize that the interruption of the
906 northward flux through the Bosphorus Straight leads to the restriction of the
907 Dardanelles channel by deltas forming at the mouths of local rivers. In the
908 Dardanelles Straight, the current sill depth is defined by the top of Unit I at -65 m
909 while the basement sill is at -85 m and in the Bosphorus Straight, the present-day sill
910 depth is at -35 m while the basement sill is at -70 m ([Çağatay et al., 2009](#)).

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

917 In the SE Black Sea (core 22-GC3), the marine influence however starts only
918 at 128 ka (or 815.5 cm depth) between *Quercus-robur*-type zone (22GC-E1) and the
919 *Juniperus* zone (22GC-E2), i.e. well into the interglacial ([Shumilovskikh et al., 2013](#));
920 whereas here in core CS18 it starts earlier at 756 cm depth, i.e. in the middle of zone

921 18P-2 dominated by *Artemisia* and that is probably still part of the Saalian Glacial.
922 Although a large difference between the SoM and the Black Sea palynological
923 diagrams, this can be explained by the marine inflow that has first to penetrate the
924 SoM from the Aegean Sea and cross two sills, before reaching the Black Sea. For
925 comparison, a delay of 4-6 ka has been observed for the Lateglacial-Holocene, with
926 a marine incursion in the SoM starting at 14.7 ka followed by progressive salinisation
927 with a facies transition defined at 12.6 ka (Vidal et al., 2010; Çağatay et al., 2015)
928 and in the Black Sea at 8.9-8.3 ka (Shumilovskikh et al., 2013; Mudie et al., 2014;
929 Yakupoğlu et al., 2022). This lag has been explained by the difference in sill depth
930 between the Bosphorus and the Dardanelles Straights, the Bosphorus one being
931 close to its present-day level (-35 m) at the time of reconnection and the Dardanelles
932 one being somewhat deeper (-75 ± 5 m) than its present-day level. The situation
933 during the MIS 6 to 5 transition may have been similar with a Bosphorus Strait at
934 about -40 m (Çağatay et al., 2019). The early marine incursion in the SoM at the end
935 of MIS stage 6 indicates that the Dardanelles sill depth was also deeper at that time.

936 A new result from the present study that remains to be explained is the
937 prevalence of brackish Ponto-Caspian conditions in surface water during the later
938 phases of MIS 5 at times when sea level was well above the present-day
939 Dardanelles sill level (Fig. 9). The explanation we propose is that the Bosphorus
940 northward flow was first decreased then interrupted as sea level started decreasing
941 before the end of MIS 5e, and that resulted in a progressive reduction of seawater
942 inflow through the Dardanelles Strait. This decreased oxygen availability in the
943 marine deep waters, as indicated by foraminifer assemblages, eventually enabled
944 the sediment filling of the channel by local rivers, further limiting marine input. Low
945 and/or episodic sea-water inputs during the later part of MIS 5 may have maintained

946 stratified conditions in the SoM with low salinity surface waters and nearly anoxic
947 bottom water of intermediate salinity. Sea level drop during MIS 5d and 5b
948 presumably interrupted the marine inflow, lowering the vertical salinity contrast and
949 allowing better oxygenation of the deep waters. During the following glacial
950 lowstands (MIS 4-3-2), the Marmara lake remained isolated and outflow (either
951 permanent or episodic) may have incised the sedimentary deposits to allow
952 reconnection at \approx -75 m at the end of the glaciation.

953 The $\delta^{18}\text{O}$ of the well-dated Sofular stalagmite record reflects salinity of the
954 Black Sea surface waters by way of evaporation and moisture availability
955 ([Badertscher et al., 2011](#)). These authors propose one clear period of connection
956 between the Black Sea and the Mediterranean Sea at roughly MIS 5e based on a
957 threshold of $-8.5 \pm 1\text{‰}$ $\delta^{18}\text{O}$ values. The next period below that threshold is at MIS 5c
958 and also suggests connection but for a shorter period. The last period at MIS 5a
959 would not have led to sufficiently high water levels to allow flowing of the
960 Mediterranean marine waters into the Black Sea. This fits rather well our results in
961 the SoM, with a full surface water connection at MIS 5e when global sea level
962 reached 8 m above sea level, a marine incursion affecting surface waters during MIS
963 5c when global sea level reached 24 m below sea level and no incursion during MIS
964 5a when global sea level reached only 28 m below sea level ([Rohling et al., 2021](#)).

965 **7 Conclusions**

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

969 From the methodological point of view, as for the Caspian Sea ([Leroy et al.,](#)
970 [2013, 2014](#)) when using the palynological method (pollen and dinocysts), it is
971 essential to study separately and objectively the terrestrial conditions and the aquatic
972 ones. A comparison of the two approaches may then clearly demonstrate if the
973 changes are synchronous or not, and if different forcing factors are at play.

974 As demonstrated for the Iberian margin by pollen and isotopic analyses, the
975 Eemian starts and finishes respectively after the start and after the end of MIS 5e
976 high levels. The same offset can be demonstrated here in this joint pollen - dinocyst
977 study.

978 The combination of the two sequences, CS22 and CS18, following the
979 chronology proposed in [Leroy et al. \(submitted\)](#), allows highlighting one main marine
980 phase corresponding to MIS 5e (marine surface and bottom conditions) and a minor
981 one (mixed surface conditions) in MIS5c. The rest of the record is brackish-
982 Pontocaspian although with some differences across the sequence (e.g. MIS 6
983 versus MIS 4). Based on climatic data (pollen) allowing linking the records, a delay in
984 the marine influence between SoM and SE Black Sea is observed, perhaps due to
985 the closure of the Bosphorus Sill and its late opening or a different sill height.

986 The current water level reconstruction corresponds well to the Global Ocean
987 level curve and Black Sea salinity reconstructions, both indicating progressively
988 lower high stands (i.e. from MIS 5e to 5a) and decreasing connections between the
989 Black Sea and the SoM across MIS 5.

990 The four sapropels (MSAP-4, 3, 2a and 2b) formed under very different
991 surface water conditions ranging from marine to brackish-Pontocaspian, but always
992 under a warm climate. Moreover, the shallower site (CS18) in the east (Imralı basin)
993 does not form as often full anoxia than the western and deeper site (CS22). For

994 example: the top limit of the bottom anoxia in sapropel MSAP-4 would not have
995 reached 291 m water depth (location of core CS18).

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

1000 **Acknowledgements**

1001 We are grateful to Jennifer Bradley and Luke Glascott at the University of
1002 Liverpool (UK) and Dahvya Belkacem at IMBE (Aix-en-Provence, France) for
1003 palynological extraction in the pollen laboratories. Cores were taken during the
1004 MARSITECRUISE of Ifremer/Genavir R.V. Pourquoi Pas?, within the framework of
1005 MARSITE FP7 EU Project (grant agreement no: 308417). Thanks are due to N.
1006 Çağatay, K. Eriş and N. Yakupoğlu from ITU (Turkey) who gave access to core
1007 CS18 for sampling and contributed to the improvement of this publication owing to
1008 intensive discussions over the years. Part of this work has been supported by
1009 Bilateral ANR/TÜBITAK collaborative research project MAREGAMI (ANR-16-CE03-
1010 0010-02) and Tubitak Project (116Y371).

1011 **Data availability**

1012 The dinocyst data are available in PANGEA.

1013 **References**

- 1014 Aydoğdu, A., Pinardi, N., Özsoy, E., Danabasoglu, G., Gürses, O, Karspeck, A.,
1015 2018. Circulation of the Turkish Straits System under interannual atmospheric
1016 forcing. *Ocean Sci.* 14, 999–1019.
- 1017 Aziz, S., McHugh, C.M., Ryan, W.B.F., Henry, P., Kende, J., Delligatti, M.,
1018 Tachikawa, K., 2019. New insights into the rapid transformation of the sea into a
1019 lake, Marmara Sea, Turkey. AGU abstract.
1020 <https://ui.adsabs.harvard.edu/abs/2019AGUFMPP21D1650A/abstract>
- 1021 Badertscher, S., Fleitmann, D., Cheng, H., Edwards, R.L., Göktürk, O.M., Zumbühl,
1022 A., Leuenberger, M., Tüysüz, O., 2011. Pleistocene water intrusions from the
1023 Mediterranean and Caspian seas into the Black Sea. *Nat. Geosci.* 4, 236–239.
- 1024 Balkis, N., Balci, M., Giannakourou, A., Venetsanopoulou, A., Mudie, P., 2016.
1025 Dinoflagellate resting cysts in recent marine sediments from the Gulf of Gemlik
1026 (Marmara Sea, Turkey) and seasonal harmful algal blooms. *Phycologia* 55, 2,
1027 187–209
- 1028 Bard, E., Jouannic, C, Hamelin, B., Pirazzoli, P., Arnold, M., Faure, G.,
1029 Sumosusastro, P., Syaefudin, 1996. Pleistocene sea levels and tectonic uplift
1030 based on dating of corals from Sumba Island, Indonesia. *Geoph. Res. Lett.* 23,
1031 12, 1473-1476.
- 1032 Bennett, K., 2007. Psimpoll and Pscomb Programs for Plotting and Analysis. Version
1033 Psimpoll 4.27. <http://chrono.qub.ac.uk/psimpoll/psimpoll.html> (accessed 27 May
1034 2020).
- 1035 Beşiktepe, S.T., Sur, H.I., Özsoy, E., Latif, M.A., Oğuz, T., Ünlüata, Ü., 1994. The
1036 circulation and hydrography of the Marmara Sea. *Prog. Oceanogr.* 34, 285–334.

- 1037 Bradley, L.R., Marret, F., Mudie, P.J. et al., 2012. Constraining Holocene sea-
1038 surface conditions in the south-western Black Sea using dinoflagellate cysts.
1039 Journal of Quaternary Science 27, 835–843.
- 1040 Bravo, I., Figueroa, R.I., 2014. Towards an ecological understanding of dinoflagellate
1041 cyst functions. Microorganisms, 2, 11–32.
- 1042 Çağatay, M.N., Görür, N., Algan, A., Eastoe, C.J., Tchapalyga, A., Ongan, D., Kuhn,
1043 T, Kuşcu, İ., 2000. Late Glacial-Holocene palaeoceanography of the Sea of
1044 Marmara: timing of connections with the Mediterranean and the Black
1045 Sea. Marine Geology 167, 191-206.
- 1046 Çağatay, M.N., Eriş, K., Ryan, W.B.F., Sancar, Ü., Polonia, A., Akçer, S., Biltekin, D.,
1047 Gasperini, L., Görür, N., Lericolais, G., Bard, E., 2009. Late Pleistocene–
1048 Holocene evolution of the northern shelf of the Sea of Marmara. Marine Geology
1049 265, 87–100.
- 1050 Çağatay, M.N., Wulf, S., Sancar, Ü., Özmaral, A., Vidal, L., Henry, P., Gasperini, L.,
1051 2015a. The tephra record from the Sea of Marmara for the last ca. 70 ka and its
1052 palaeoceanographic implications. Mar. Geol. 361, 96-110.
- 1053 Çağatay, M.N., Uçarkuş, G., Eriş, K.K., Henry, P., Gasperini, L., Polonia, P., 2015b.
1054 Submarine canyons of the Sea of Marmara. In CIESM Monograph 47 [F. Briand
1055 ed.] Submarine canyon dynamics in the Mediterranean and tributary seas- An
1056 integrated geological, oceanographic and biological perspective, pp. 123- 135.
1057 CIESM Publisher, Monaco.
- 1058 Çağatay, M.N., Eriş, K.K., Makaroglu, Ö., Yakuboğlu, N., Henry, P., Leroy, S.,
1059 Uçarkuş, G., Sakınç, M., Yalamaz, B., Bozyiğit, C., Kende, J., 2019. The Sea of
1060 Marmara during Marine Isotope Stages 5 and 6. Quat. Sc. Rev. 220, 124-141.

- 1061 Çağatay, M.N., Eriş, K.K., Erdem, Z., 2022. Morphology and Late Pleistocene-
1062 Holocene Sedimentation of the Strait of İstanbul (Bosphorus): A review. In: Rossi,
1063 V. M., Longhitano, S., Olariu, C. and Chiocci, F. (eds) Straits and Seaways:
1064 Controls, Processes and Implications in Modern and Ancient Systems. Geological
1065 Society, London, Special Publications, 523, [https://doi.org/10.1144/SP523-2021-](https://doi.org/10.1144/SP523-2021-48)
1066 48.
- 1067 Cheddadi, R., Rossignol-Strick, M., 1995. Improved preservation of organic matter
1068 and pollen in eastern Mediterranean sapropels. *Paleoceanography* 10, 2, 301-
1069 309.
- 1070 Chepalyga, A.L., 2007. The late glacial great flood in the Ponto-Caspian basin. in:
1071 Yanko-Hombach, V., Gilbert, A.S., Panin, N., Dolukhanov, P.M. (Eds), *The Black*
1072 *Sea Flood Question*. Springer, pp. 119-148
- 1073 Clarke, K.R., Gorley, R.N., 2015. *PRIMER v7: User Manual/Tutorial*. PRIMER-E,
1074 Plymouth (296 pp.).
- 1075 Cornuault, M., Vidal, L., Tachikawa, K., Licari, L., Rouaud, G., Sonzogni, C., Revel,
1076 M., 2016. Deep water circulation within the eastern Mediterranean Sea over the
1077 last 95 kyr: New insights from stable isotopes and benthic foraminiferal
1078 assemblages. *Palaeogeography, Palaeoclimatology, Palaeoecology* 459, 1–14.
- 1079 de Vernal, A., 2009. Marine palynology. *IOP Conf. Ser.: Earth Environ. Sci.* 5
1080 012002.
- 1081 de Vernal, A., Marret, F., 2007. Organic-Walled Dinoflagellate Cysts: Tracers of Sea-
1082 Surface Conditions. *Development in Marine Geology* 1, 371-408.
- 1083 de Vernal, A., Radi, T., Zaragosi, S., Van Nieuwenhove, N., Rochon, A., Allan, E.,
1084 De Schepper, S., Eynaud, F., Head, M.J., Limoges, A., Londeix, L., Marret, F.,
1085 Matthiessen, J., Penaud, A., Pospelova, V., Price, A., Richerol, T. 2020.

- 1086 Distribution of common modern dinoflagellate cyst taxa in surface sediments of
1087 the Northern Hemisphere in relation to environmental parameters: The new
1088 n=1968 database. *Marine Micropaleontology*, art. no. 101796.
- 1089 Eriş, K.K., Ryan, W.B.F, Çağatay, M.N., Sancar, U., Lericolais, G., Menot, G., Bard,
1090 E., 2007. The timing and evolution of the post-glacial transgression across the
1091 Sea of Marmara shelf south of İstanbul. *Marine Geology* 243, 1-4, 57-76.
- 1092 Eriş, K.K., Çağatay, M.N., Akçer, S., Casperini, L., Mart, Y., 2011. Late glacial to
1093 Holocene sea-level changes in the Sea of Marmara: new evidence from high-
1094 resolution seismics and core studies. *Geo-Marine Letters* 31, 1, 1-18.
- 1095 Fatourou, E., Kaftzidou, A., Marret, F., Panagiotopoulos, K., Kouli, K., 2023. Late
1096 Quaternary Ponto-Caspian dinoflagellate cyst assemblages from the Gulf of
1097 Corinth, Central Greece (eastern Mediterranean Sea). *Marine Micropal.* 179,
1098 102211.
- 1099 Ferguson, S., Warny, S., Escarguel, G., Mudie, P.J., 2018. MIS 5e1 dinoflagellate
1100 cyst analyses and morphometric evaluation of *Galeacysta etrusca* and *Spiniferites*
1101 *cruciformis* in southwestern Black Sea. *Quaternary International* 465, 117-129.
- 1102 Gökaşan, E., Ergin, M., Özyalvaç, M., Sur, H.I., Tur, H., Görüm, T., Ustaömer, T.,
1103 Batuk, F.G., Alp, H., Birkan, H., Türker, A., Gezgin, E., Özturan, M., 2008. Factors
1104 controlling the morphological evolution of the Çanakkale Strait (Dardanelles,
1105 Turkey). *Geo-Mar. Lett.* 28,107–129.
- 1106 Gökaşan E., Tur, H., Ecevitoglu, B., Görüm, T., Türker, A., Tok, B., Çağlak, F.,
1107 Birkan, H., Simşek, M., 2005. Evidence and implications of massive erosion along
1108 the Strait of İstanbul (Bosphorus). *Geo-Mar. Lett.* 25, 324–342.
- 1109 Grall, C., Henry, P., Westbrook, G.K., Çağatay, MN, Thomas, Y., Marsset, B.,
1110 Borschneck, D., Saritas, H., Cifçi, G., Géli, L., 2014. Mass transport deposits

- 1111 periodicity related to glacial cycles and marine-lacustrine transitions on a ponded
1112 basin of the Sea of Marmara (Turkey) over the last 500 ka. Chapter 53. In:
1113 Krastel, S. et al. (eds) Submarine mass movement and their consequences.
1114 Advances in Natural and Technological Hazards Research 37, Springer
1115 International Publishing Switzerland, pp. 595-603.
- 1116 Grant, K.M., Grimm, R., Mikolajewicz, U., Marino, G., Ziegler, M., Rohling, E.J.,
1117 2016. The timing of Mediterranean sapropel deposition relative to insolation, sea-
1118 level and African monsoon changes. Quaternary Science Reviews 140, 125-141.
- 1119 Head, M.J., Lewis, J., de Vernal A., 2006. The cyst of the calcareous dinoflagellate
1120 *Scrippsiella trifida*: resolving the fossil record of its organic wall with that of
1121 *Alexandrium tamarense*. Journal of Paleontology 80, 1, 1-18.
- 1122 Henry, P., Şengör, A.M.C., Çağatay, M.N., 2007. MARNAUT cruise, RV L'Atalante.
1123 <https://doi.org/10.17600/7010070>.
- 1124 Hiscott, R.N., Aksu, A.E., Yaltrak, C., 2021. The uppermost Pleistocene–Holocene
1125 mud drape across the Marmara Sea: Quantification of detrital supply from
1126 southern Marmara rivers. Sedimentary Geology 415, 105851.
- 1127 Hoyle, T., Bista, D., Flecker, R., Krijgsman, W., Sangiorgi, F., 2021. Climate-driven
1128 connectivity changes of the Black Sea since 430 ka: Testing a dual palynological
1129 and geochemical approach. Palaeogeography, Palaeoclimatology, Palaeoecology
1130 561, 110069.
- 1131 Karabanov, E.B, Prokopenko, A.A., Williams, D.F., Khursevich, G.K., 2000. Evidence
1132 for mid-Eemian cooling in continental climatic record from Lake Baikal. Journal of
1133 Paleolimnology 23, 365–371.

- 1134 Kazancı, N., Leroy, S., Ileri, Ö., Emre, O., Kibar, M., Öncel, S., 2004. Late Holocene
1135 erosion in NW Anatolia from sediments of Lake Manyas, Lake Ulubat and the
1136 southern shelf of the Marmara Sea, Turkey. *Catena* 57, 277-308.
- 1137 Kende, J., 2018. Tectonique et hydrologie en mer de Marmara : Histoire de
1138 l'ouverture de la mer de Marmara et reconstitution de la réponse hydrologique aux
1139 variations climatiques depuis le dernier interglaciaire. Thèse de doctorat, Aix-
1140 Marseille Université. 18 pp.
- 1141 Koeling, M., Webster, J.M., Camoin, G., Iryu, Y., Bard, E., Seard, C., 2009. SEALEX
1142 — Internal reef chronology and virtual drill logs from a spreadsheet-based reef
1143 growth model. *Glob. Planet. Ch.* 66, 149-159.
- 1144 Kotthoff, U., Pross, J., Müller, U. C., Peyron, O., Schmiedl, G., Schulz, H., Bordon,
1145 A., 2008. Climate dynamics in the borderlands of the Aegean Sea during
1146 formation of sapropel S1 deduced from a marine pollen record. *Quaternary*
1147 *Science Reviews* 27, 832–845.
- 1148 Leroy, S.A.G., Marret, F., Giralte, S., Bulatov, S. A., 2006. Natural and anthropogenic
1149 rapid changes in the Kara-Bogaz Gol over the last two centuries by palynological
1150 analyses. *Quaternary International* 150, 52-70.
- 1151 Leroy, S.A.G., 2010, Palaeoenvironmental and palaeoclimatic changes in the
1152 Caspian Sea region since the Lateglacial from palynological analyses of marine
1153 sediment cores. *Geography, Environment, Sustainability, Faculty of Geography*
1154 *of Lomonosov Moscow State University and by the Institute of Geography of*
1155 *RAS.* 2, 32-41.
- 1156 Leroy, S.A.G., Tudryn, A., Chalié, F., López-Merino, L., Gasse, F., 2013. From the
1157 Allerød to the mid-Holocene: palynological evidence from the south basin of the
1158 Caspian Sea. *Quaternary Science Reviews* 78, 77-97.

- 1159 Leroy, S.A.G., López-Merino, L., Tudryn, A., Chalié, F., Gasse, F., 2014. Late
1160 Pleistocene and Holocene palaeoenvironments in and around the Middle
1161 Caspian Basin as reconstructed from a deep-sea core. *Quaternary Science*
1162 *Reviews* 101, 91-110.
- 1163 Leroy, S.A.G., Lahijani, H., Crétaux, J.-F., Aladin, N., Plotnikov, I., 2020. Past and
1164 current changes in the largest lake of the world: The Caspian Sea. in: Mischke,
1165 S. (Ed.), *Large Asian lakes in a changing world*. Springer, pp. 65-107.
- 1166 Leroy, S.A.G., et al., submitted. Palynology, palaeoclimate and chronology of stage
1167 MIS 5 from two long cores in the Marmara Sea. *QSR*
- 1168 Lisiecki, L.E., Raymo, M.E., 2005. A Pliocene-Pleistocene stack of 57 globally
1169 distributed benthic $\delta^{18}O$ records. *Paleoceanography*, 20, PA1003
- 1170 Londeix, L., Herreyre, Y., Turon, J.-L., Fletcher, W., 2009. Last Glacial to Holocene
1171 hydrology of the Marmara Sea inferred from a dinoflagellate cyst record.
1172 *Review of Palaeobotany and Palynology* 158, 52–71.
- 1173 Margalef, R., 1958. Information theory in ecology. *Gen. Syst.* 3, 36–71.
- 1174 Marret, F., Zonneveld, K.A.F., 2003. Atlas of modern organic-walled dinoflagellate
1175 cyst distribution. *Review of Palaeobotany and Palynology* 125, 1–200.
- 1176 Marret, F., Leroy, S., Chalié, F., Gasse, F., 2004. New organic-walled dinoflagellate
1177 cysts from recent sediments of central Asian seas. *Rev. Palaeobot. Palynol.*
1178 129, 1–20.
- 1179 Marret, F., Mudie, P., Aksu, A., Hiscott, R.N., 2009. A Holocene dinocyst of a two-
1180 step transformation of the Neoeuxinian brackish water lake into the Black Sea.
1181 *Quaternary International* 197, 72-86.
- 1182 Marret, F., Bradley, L.R., Tarasov, P.E., Ivanova, E.V., Zenina, M.A., Murdmaa, I.O.,
1183 2019. The Holocene history of the NE Black Sea and surrounding areas: An

- 1184 integrated record of marine and terrestrial palaeoenvironmental change. The
1185 Holocene 1–14.
- 1186 Marret, F., Bradley, L., de Vernal, A., Hardy, W., Kim, S.Y., Mudie, P., Penaud, A.,
1187 Pospelova, V., Price, A.M., Radi, T., Rochon, A., 2020. From bi-polar to
1188 regional distribution of modern dinoflagellate cysts, an overview of their
1189 biogeography. *Mar Micropal* 159, 101753.
- 1190 McCarthy, F.M.G., Mudie, P.J., 1998. Oceanic pollen transport and pollen: dinocyst
1191 ratios as markers of late Cenozoic sea level change and sediment transport.
1192 *Palaeogeography, Palaeoclimatology, Palaeoecology* 138, 187-206.
- 1193 Mertens, K. Ribeiro, S., Bouimetarhan, I., Caner, H., Combourieu-Nebout, N., Dale,
1194 B., de Vernal, A., Ellegaard, M., Filipova, M., Godhe, A., Grøsfjeld, K., Leroy, S.
1195 A. G. and co-authors, 2009. Process length variation in cysts of a
1196 dinoflagellate, *Lingulodinium machaerophorum*, in surface sediments:
1197 investigating its potential as salinity proxy. *Marine Micropal* 70, 54-69.
- 1198 Mertens, K.N., Van Nieuwenhove, N., Gurdebeke, P.R., Aydin, H., Bogus, K.,
1199 Bringué, M., Dale, B., De Schepper, S., de Vernal, A., Ellegaard, M., Grothe,
1200 A., Gu, H., Head, M.J., Heikkilä, M., Limoges, A., Londeix, L., Louwye, S.,
1201 Marret, F., Masure, E., Matsuoka, K., Mudie, P.J., Penaud, A., Pospelova, V.,
1202 Price, A.M., Ribeiro, S., Rochon, A., Sangiorgi, F., Schreck, M., Torres, V.,
1203 Uzar, S., Versteegh, G.J.M., Warny, S., Zonneveld, K., 2018. The dinoflagellate
1204 cyst genera *Achomosphaera* Evitt 1963 and *Spiniferites* Mantell 1850 in
1205 Pliocene to modern sediments: a summary of round table discussions.
1206 *Palynology* 42, sup1, 10-44

- 1207 Mudie, P.J., Aksu, A.E., Yasar, D., 2001. Late Quaternary dinocysts from the Black,
1208 Marmara and Aegean Seas: Variations in assemblages, morphology and
1209 paleosalinity. *Mar. Micropal.* 43, 155-178.
- 1210 Mudie, P.J., Rochon A., Aksu, A.E., Gillespie, H., 2002. Dinoflagellate cysts,
1211 freshwater algae and fungal spores as salinity indicators in Late Quaternary
1212 cores from Marmara and Black seas. *Mar. Geol.* 190, 203-231.
- 1213 Mudie, P.J., Rochon, A., Aksu, A.E., Gillespie, H., 2004. Late glacial, Holocene and
1214 modern dinoflagellate cyst assemblages in the Aegean–Marmara–Black Sea
1215 corridor: statistical analysis and re-interpretation of the early Holocene Noah’s
1216 Flood hypothesis. *Rev. Palaeobot. Palynol.* 128, 1-2, 143-167.
- 1217 Mudie, P., Yanko-Hombach, V., Kadurin, S., 2014. The Black Sea Dating Game and
1218 Holocene Marine Transgression. *Open Journal of Marine Science* 4, 1-7.
- 1219 Mudie, P.J., Marret, F., Mertens, K., Shumilovskikh, L., Leroy, SAG, 2017. Atlas of
1220 modern dinoflagellate cyst distributions in the Black Sea Corridor, including
1221 Caspian and Aral Seas. *Mar Mic* 134, 1-152.
- 1222 Mudie, P.J., Marret, F., Gurdebeke, P.R., Hartman, J.D., Reid, P.C., 2021. Marine
1223 dinocysts, acritarchs and less well-known NPP: tintinnids, ostracod and
1224 foraminiferal linings, copepod and worm remains. Geological Society, London,
1225 Special Publications 511, SP511-2020-2055.
- 1226 Paillès, C., Blanc-Valleron, M.M, Poulin, M., Crémière, A., Boudouma, O., Pierre, C.,
1227 2014. A new fossil diatom, *Entomoneis calixasini*, from the Turkish Marmara
1228 Sea sediments. *Diatom Research* 29, 4, 411-422.
- 1229 Pope, E.L., et al., 2022. First source-to-sink monitoring shows dense head controls
1230 sediment flux and runout in turbidity currents. *Sci. Adv.* 8, eabj3220.

- 1231 Rohling, E.J., Marino, G., Grant, K.M., 2015. Mediterranean climate and
1232 oceanography, and the periodic development of anoxic events (sapropels).
1233 Earth Sci. Rev. 143, 62-97.
- 1234 Rohling, E.J., Yu, J., Heslop, D., Foster, G.L., Opdyke, B., Roberts A.P., 2021. Sea
1235 level and deep-sea temperature reconstructions suggest quasi-stable states
1236 and critical transitions over the past 40 million years. Sci. Adv. 2021; 7.
- 1237 Sala-Pérez, M., Lattuada, M., Flecker, R., Anesio, A., Leroy, S.A.G.,
1238 2020. Dinoflagellate cyst assemblages as indicators of environmental
1239 conditions and shipping activities in coastal areas of the Black and Caspian
1240 Seas. Regional Studies in Marine Science 38, 101472.
- 1241 Sannino, G., Sözer, A., Özsoy, E., 2017. A high-resolution modelling study of the
1242 Turkish Straits System. Ocean Dynamics, 67, 397–432,
1243 <https://doi.org/10.1007/s10236-017-1039-2>.
- 1244 Schmiedl, G., Mitschele, A., Beck, S., Emeis, K.-C., Hemleben, C., Schulz, H.,
1245 Sperling, M., Weldeab, S., 2003. Benthic foraminiferal record of ecosystem
1246 variability in the eastern Mediterranean Sea during times of sapropel S5 and S6
1247 deposition. Palaeogeography, Palaeoclimatology, Palaeoecology 190, 139-164.
- 1248 Shackleton, N.J., Sanchez-Goñi, M. F., Pailler, D., Lancelot, Y., 2003. Marine
1249 Isotope Substage 5e and the Eemian Interglacial. Global and Planetary
1250 Change 36, 151-155.
- 1251 Shumilovskikh, L.S., Marret, F., Fleitmann, D., Arz, H.W., Nowaczyk, N., Behling, H.,
1252 2013. Eemian and Holocene sea-surface conditions in the southern Black Sea:
1253 Organic-walled dinoflagellate cyst record from core 22-GC3. Mar. Micropal.
1254 101, 146-160.

- 1255 Sonzogni, C., Bard, E., Rostek, F., Dollfus, D., Rosell-Melé, A., Eglinton, G., 1997.
1256 Temperature and Salinity Effects on Alkenone Ratios Measured in Surface
1257 Sediments from the Indian Ocean. *Quaternary Research* 47, 3, 344–355.
1258 <http://doi.org/10.1006/qres.1997.1885>
- 1259 Sorlien, C.C., Akhun, S.D., Seeber, L., Steckler, M.S., Shillington, D.J., Kurt, H.,
1260 Imren, C., 2012. Uniform basin growth over the last 500 ka, north Anatolian
1261 fault, Marmara Sea, Turkey. *Tectonophysics* 518, 1-16.
- 1262 Sorrel, P., Popescu, S.-M., Head, M. J., Suc, J.-P., Klotz, S., Oberhänsli, H., 2006.
1263 Hydrographic development of the Aral Sea during the last 2000 years based on
1264 a quantitative analysis of dinoflagellate cysts, *Palaeogeography,*
1265 *Palaeoclimatology, Palaeoecology* 234, 304–327.
- 1266 Tolun, T., Çağatay, M.N., Carrigan, W.J., 2002. Organic geochemistry and origin of
1267 Holocene sapropelic layer and associated sediments in Marmara Sea. *Mar.*
1268 *Geol.* 190, 47–60.
- 1269 Triantaphyllou, M.V., Antonarakou, A., Kouli, K., Dimiza, M., Kontakiotis, G.,
1270 Papanikolaou, M. D., Ziveri, P., Mortyn, P. G., Lianou, V., Lykousis, V.,
1271 Dermitzakis, M. D., 2009. Late Glacial–Holocene ecostratigraphy of the south-
1272 eastern Aegean Sea, based on plankton and pollen assemblages. *Geo-Mar.*
1273 *Lett.* 29, 249–267.
- 1274 Tudryn, A., Leroy, S.A.G., Toucanne, S., Gibert-Brunet, E., Tucholka, P., Lavrushin,
1275 Y.A., Dufaure, O., Miska, S., Bayon, G., 2016. The Ponto-Caspian basin as a final
1276 trap for southeastern Scandinavian ice-sheet meltwater. *Quat Sci Rev* 148, 29-43.
- 1277 Vidal, L., Ménot, G., Joly, C., Bruneton, H., Rostek, F., Çağatay, M.N., Major, C.,
1278 Bard, E., 2010. Hydrology in the Sea of Marmara during the last 23 ka:

- 1279 Implications for timing of Black Sea connections and sapropel deposition.
1280 *Paleoceanography* 25, PA1205.
- 1281 Wegwerth, A., Eckert, S., Dellwig, O., Schnetger, B., Severmann, S., Weyer, S.,
1282 Brüske, A., Kaiser, J., Köster, J., Arz, H.W., Brumsack, H.-J., 2018. Redox
1283 evolution during Eemian and Holocene sapropel formation in the Black Sea.
1284 *Palaeogeography, Palaeoclimatology, Palaeoecology* 489, 249–260.
- 1285 Wegwerth, A., Dellwig, O., Wulf, S., Plessen, B., Kleinhanns, I., Nowaczyk, N.T.,
1286 Jiabo, L., Arz, H.W., 2019. Major hydrological shifts in the Black Sea “Lake” in
1287 response to ice sheet collapses during MIS 6 (130-184 ka BP). *Quaternary*
1288 *Science Reviews* 219, 126-144.
- 1289 Yakupoğlu, N., Henry, P., Uçarkus, G., Eris, K.K., Demory, F., Crouzet, C., Çağatay,
1290 M.N., 2022. Factors affecting thickness and frequency of turbidites triggered by
1291 earthquakes in Kumburgaz Basin, Sea of Marmara. *Marine Geology* 452, 106900.
- 1292 Yaltrak, C., Alpar, B., Sakınc, M., Yüce, H., 2000. Origin of the Strait of Çanakkale
1293 (Dardanelles): regional tectonics and the Mediterranean-Marmara incursion. *Mar.*
1294 *Geol.* 164, 139-156 (with erratum 167, 189-190).
- 1295 Yaltrak, C., 2002. Tectonic evolution of the Marmara Sea and its surroundings.
1296 *Marine Geology* 190, 493-529.
- 1297 Zwiep, K.L., Hennekamp, R., Donders, T.H., van Helmond, N.A.G.M., de Lange,
1298 G.J., Sangiorgi, F., 2018. Marine productivity, water column processes and
1299 seafloor anoxia in relation to Nile discharge during sapropels S1 and S3.
1300 *Quaternary Science Reviews* 200, 178-190.
- 1301

22/03/2023

1302 **Supplementary information**

1303 **Table SI1:** Benthic foraminifera census data in nine samples from core CS22

Core CS22											
benthic foraminifera census data		depth (cm)	1631	1632	1915	1935	1955	1975	1995	2015	2035
shallow-water species	<i>Ammonia parkinsoniana</i>				2						
	<i>Ammonia</i> sp						1				
	<i>Discorbinella bertheloti</i>								1	3	3
	<i>Elphidium</i> sp				1						
	<i>Haynesina depressula</i>				5						
	<i>Rosalina</i> sp							1		1	2
	<i>Porosonion granosum</i>				1						
hyaline bathyal species	<i>Amphicoryna scalaris</i>								1		
	<i>Bolivinita quadrilatera</i>							1			
	<i>Bolivina striatula</i>										2
	<i>Bolivina alata</i>					64	371	101			
	<i>Bolivina dilatata</i>					16	4	5			1
	<i>Bolivina spathulata</i>					1	2	78	9	7	2
	<i>Bulimina aculeata</i>							5	6	8	11
	<i>Bulimina inflata</i>							1	4	151	235
	<i>Bulimina elongata</i>						7	6	5		1
	<i>Bulimina marginata</i>						98	5	1	3	4
	<i>Bulimina</i> sp									7	2
	<i>Cassidulina carinata</i>						15	2	8	32	9
	<i>Cassidulina obtusa</i>										
	<i>Chilostomella oolina</i>									2	3
	<i>Chilostomella ovoidea</i>							11	10	3	
	<i>Chilostomella</i> sp							3			
	<i>Cibicidoides pachyderma</i>							1	6	35	15
	<i>Cibicidoides pseudoungerianus</i>									7	
	<i>Cibicidoides</i> sp								1	8	
	<i>Dentalina</i> sp										1
	<i>Fissurina staphyllearia</i>									1	
	<i>Globobulimina</i> spp								1		3
	<i>Globocassidulina oblonga</i>									1	2
	<i>Gyrogoninoida lamarckiana</i>						1		2	17	10
	<i>Gyrogoninoida soldanii</i>										1
	<i>Hyalinea balthica</i>					1	105	9	27	108	8
	<i>Lagena</i> spp										2
	<i>Lenticulina calcar</i>										1
	<i>Lenticulina</i> sp						1				
	<i>Melonis affinis</i>						8	1	23	44	3
	<i>Melonis pompilioides</i>						8		9	15	9
	<i>Neolenticulina variabilis</i>										
	<i>Nonionella turgida</i>										1
	<i>Planulina ariminensis</i>								20	45	13
	<i>Pullenia quadriloba</i>										
	<i>Robertina</i> sp									1	
	<i>Robertina translucens</i>										1
<i>Saracenaria</i> sp											
<i>Siphogenerina columellaris</i>											
<i>Siphonina</i> sp											
<i>Sphaeroidina bulloides</i>									2		
<i>Stainforthia concava</i>						3					
<i>Uvigerina mediterranea</i>									12		
<i>Uvigerina peregrina</i>								4	3	6	
<i>Valvulineria bradyana</i>								13	2	11	
porcelaneous species	<i>Adelosina</i> sp								1		1
	<i>Articulina tubulosa</i>									1	2
	<i>Biloculinella</i> sp										11
	<i>Comuloculina foliacea</i>									2	
	<i>Pyrgo depressa</i>								1	2	2
	<i>Pyrgo elongata</i>								2		4
	<i>Pyrgoella irregularis</i>							1	2		
	<i>Quinqueloculina padana</i>							2	2	2	11
	<i>Quinqueloculina stelligera</i>							1	2		3
	<i>Quinqueloculina</i> sp									2	8
	<i>Sigmoilopsis schlumbergeri</i>						1		1	1	3
	<i>Siphonaperta</i> sp									1	
	<i>Spiroloculina excavata</i>										2
	<i>Triloculina tricarinata</i>									2	
	<i>Triloculina</i> sp										1
<i>Textularia</i> spp											
unidentified species				4	4	1			2	1	2
total number of counted foraminifera	0	0	14	332	423	331	535	377	310		
total number of shallow water individuals	0	0	9	0	1	1	1	4	5		
total number of bathyal individuals	0	0	5	332	422	330	534	373	305		
BFN (bathyal species only, ind.g ⁻¹)	0	0	0	386	30	209	277	459	194		
S (number of bathyal taxa)	0	0	1	13	13	24	34	30	40		

1305 **Table SI2:** Ecological assignment of main foraminiferal species in core CS22
 1306 Fontanier, F., Garnier, E., Brandily, C., Dennielou, B., Bichon, S., Gayet, N., Eugene,
 1307 T., Rovere, M., Grémare, A., Deflandre, B. 2016. Living (stained) benthic
 1308 foraminifera from the Mozambique Channel (eastern Africa): Exploring ecology of
 1309 deep-sea unicellular meiofauna. *Deep-Sea Research I* 115, 159–174.
 1310

Species adapted to high oxygen conditions (HO)			
<i>Cibicidoides</i> spp			
<i>Gyroidina lamarckiana</i>			
Miliolids excluding <i>Articulina tubulosa</i>			
Low-oxygen indicators (LO)			
<i>Chilostomella</i> spp			
<i>Globobulimina</i> spp			
<i>Stainforthia concava</i>			
Species adapted to high organic fluxes (eutrophic indicators)			
<u>adapted to low oxygenation</u>			
<i>Bolivina alata</i>			
<i>Bolivina dilatata</i>			
<i>Bolivina spathulata</i>			
<u>not tolerant to low oxygenation</u>			
<i>Bulimina marginata</i>			
<i>Hyalinea balthica</i>			
(after Cornuault et al. 2016 and references therein; Fontanier et al. 2016)			

1311

1312

1313

1314

1315 **Table SI3:** Comparative list of taxa between the Marmara Sea (cores CS22 and
 1316 CS18) and the SE Black Sea (core 22-GC3; [Shumilovskikh et al., 2013](#)).

	CS22	CS18	22-GC3	
	Pyxidinoopsis psilata, Spiniferites cruciformis/Galeacysta etrusca,	Pyxidinoopsis psilata, Spiniferites cruciformis/Galeacysta etrusca, Galeacysta etrusca,	Pyxidinoopsis psilata,	VV
	S. cruciformis A, S. cruciformis B, S. cruciformis C,	S. cruciformis A, S. cruciformis B, S. cruciformis C,	S. cruciformis, S. cruciformis, S. cruciformis,	0
	S. Inaequalis,	Pterocysta cruciformis, S. Inaequalis,		0
		Romanodinium areolatum, Caspidinium rugosum, C. rugosum rugosum, Impagidinium sp. A, I. caspiense,	Caspidinium rugosum,	VV
				0
6/13				only in MIS6
				only in MIS6
				0
	Lingulodinium machaerophorum ss,	Lingulodinium machaerophorum ss, L. machaerophorum B, L. mach. bulbous,	Lingulodinium machaerophorum,	0
1/3				0
		Dubridinium caperatum, Lejeuncysta marieae,	Dubridinium caperatum,	0
			Lejeuncysta sabrina? Quinquecuspis sp. Quinquecuspis concreta	0
	Votadinium calvum, Brown baggy cyst,	Votadinium calvum, Brown baggy cyst,	Votadinium calvum, Brigantodinium sp. B. cariaceense B. simplex C. Gymnodinium microret./nolleri	VV
				V
				V
	Xandarodinium xanthum, C. Protoperidinium nudum, C. Polykrikos hartmanii,	Xandarodinium xanthum, C. Protoperidinium nudum, C. Protoperidinium stellatum, C. Polykrikos hartmanii,	Xandarodinium xanthum, C. Protoperidinium cf nudum C. Protoperidinium stellatum, C. Polykrikos schwartzii Echinidinium sp. E. zonneveldiae	0
5/8				0
	S. belerius, S. bentorii,	S. belerius, S. bentorii,	S. belerius, S. bentorii, S. bulloideus	0
	S. elongatus, S. halnanensis,		S. elongatus, S. hyperacanthus S. membranaceus	V
				VV
	S. ramosus, S. septentrionalis, Spiniferites sp., S. mirabilis, S. pachydermus, Polysphaeridium zoharyi, Tuberculodinium vancampoae, Operculodinium israelianum, O. centrocarpum, Ataxodinium choane, Bitectatodinium tepikiense, Tectatodinium pellitum, C. Pentapharsodinium dalei, C. Scripsiella trifida, Pyxidinoopsis reticulata,	S. ramosus, S. septentrionalis, Spiniferites sp., S. mirabilis, Polysphaeridium zoharyi, Tuberculodinium vancampoae, Operculodinium israelianum, O. centrocarpum, Ataxodinium choane, Bitectatodinium tepikiense, Tectatodinium pellitum, C. Pentapharsodinium dalei, C. Alexandrium sp., C. Scripsiella trifida,	S. ramosus, S. septentrionalis, Spiniferites sp., S. mirabilis, S. pachydermus, Tuberculodinium vancampoae, O. centrocarpum, Ataxodinium choane, Bitectatodinium tepikiense, Tectatodinium pellitum, C. Pentapharsodinium dalei, Achomosphaera cf andalousiense	VV
				0
				VV
				0
21/17	Nematosphaeropsis labyrinthus,	Nematosphaeropsis labyrinthus,	Nematosphaeropsis labyrinthus, Achomosphaera cf andalousiense	VV
	33	41		
				0 = insignificant
				V = different
				VV = very different

1317

1318

22 March 2023

We have no competing interests

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

Leroy S.A.G. ^{1-2-3-*}, Henry P. ¹, Marret F³, Pailles C. ¹, Licari L. ¹, Kende
J. ¹, Rostek F. ¹, Bard E. ¹