



HAL
open science

Mass Transport Deposits Periodicity Related to Glacial Cycles and Marine-Lacustrine Transitions on a Pounded Basin of the Sea of Marmara (Turkey) Over the Last 500 ka

Céline Grall, Pierre Henry, Graham K. Westbrook, M Namik Çağatay, Yannick Thomas, Bruno Marsset, Daniel Borschneck, Hakan Saritaş, Günay Çifçi, Louis Géli

► To cite this version:

Céline Grall, Pierre Henry, Graham K. Westbrook, M Namik Çağatay, Yannick Thomas, et al.. Mass Transport Deposits Periodicity Related to Glacial Cycles and Marine-Lacustrine Transitions on a Pounded Basin of the Sea of Marmara (Turkey) Over the Last 500 ka. Sebastian Krastel; Jan-Hinrich Behrmann; David Völker; Michael Stipp; Christian Berndt; Roger Urgeles; Jason Chaytor; Katrin Huhn; Michael Strasser; Carl Bonnevie Harbitz. Submarine Mass Movements and Their Consequences. 6th International Symposium, 37, Springer International Publishing, pp.595-603, 2014, Advances in Natural and Technological Hazards Research, 978-3-319-00971-1. 10.1007/978-3-319-00972-8_53 . hal-03455606

HAL Id: hal-03455606

<https://hal.science/hal-03455606v1>

Submitted on 1 Dec 2022

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.



Distributed under a Creative Commons Attribution 4.0 International License

1 **Mass Transport Deposits periodicity related to**
2 **glacial cycles and marine-lacustrine transitions**
3 **on a ponded basin of the Sea of Marmara**
4 **(Turkey) over the last 500 ka**

5 **C. Grall^{1,2*}, P. Henry¹, G.K. Westbrook², M.N. Çağatay³, Y. Thomas², B.**
6 **Marsset², D. Borschneck¹, H. Saritas⁴, G. Cifçi⁴, L. Géli²**

7 ¹ CEREGE, CNRS, Aix Marseille University, Marseille, France

8 ² IFREMER - French Research Institute for Exploration of the Sea, Marine Geosciences,
9 Plouzané, France

10 ³ Istanbul Technical University, EMCOL and Department of Geological Engineering,
11 Istanbul, Turkey

12 ⁴ Institute of Marine Sciences and Technology, Dokuz Eylul University, Izmir, Turkey

13 *grall@cerege.fr

14 **Abstract**

15 The Sea of Marmara (SoM) is affected by large earthquakes occurring on the
16 North Anatolian Fault. Numerous submarine mass movements have occurred and
17 the most recent turbidites in the basins of the SoM have been related to historical
18 earthquakes. Within the SoM, the occurrence of submarine mass movements and
19 their size appears modulated by eustatic changes that can be accompanied by transi-
20 tions between a salty marine environment and a brackish lake environment. De-
21 tailed analysis, using a 3D high-resolution seismic dataset, of stratigraphy over the
22 last 500 ka, within a ponded basin of the Western High, shows that intervals of
23 draped sedimentary reflectors alternate with onlap sequences that followed episodes
24 of rapid sea-level rise, with a periodicity of approximately 100,000 years (corre-
25 sponding to glacial cycles). Mass Transport Deposits (MTDs) occur within the
26 onlapping sequences. Detail analysis of the youngest large slide, which probably
27 followed the lacustrine transition during Marine Isotopic Stage 4 is presented; and
28 the possible triggering processes are discussed. The potential triggers of MTDs dur-
29 ing this transition, in the context of the SoM are: (i) gas hydrate dissociation by
30 pressure drop; (ii) changes in sediments supply and transport dynamics; (iii) varia-
31 tions in pressure and/or ionic strength in pores. The latter case appears the most
32 suitable hypothesis, as salt diffuses out of the pores of the marine clay-rich sediment
33 dominated by smectite at the beginning of low stand/lacustrine stages. The pore
34 water freshening induces clay swelling, which can potentially drive sediment slope
35 failure.

36

37 **Keywords:** High Resolution 3D seismic data, Mass Transport Deposits,
38 lacustrine/marine transitions, Sea of Marmara

39 1. Introduction

40 A great variety of factors can cause and/or promote submarine mass movements,
41 including earthquakes (Masson et al. 2006), environmental changes such as sea-
42 level changes and associated variations in sediment supply (Rothwell et al. 1998),
43 and gas hydrate dissociation (e.g. Maslin et al. 2004). The Sea of Marmara (SoM)
44 is an active seismotectonic setting, located along the western part of the North An-
45 atolian major continental transform fault (Wong et al. 1995; Le Pichon et al. 2001).
46 Submarine gas hydrates occur in the sediment at fluid emission sites (Bourry et al.
47 2009). It is likely that hydrates were more widely distributed in the past and were
48 dissociated massively during the deglaciation (Menot and Bard, 2010). The SoM
49 has also been subjected to a history of connections and disconnections with Black
50 and Aegean Seas via the Bosphorous and the Çannakale Straits respectively (Figure
51 1A, e.g. Cagatay et al. 2009). These oscillations are related to global eustatic
52 changes, and induce transitions between lacustrine and marine conditions. In such
53 an environment, submarine mass-movements are very common, and 75% of the
54 sedimentation in basins is composed of gravity-induced deposits (e.g. Beck et al.
55 2007) and many of which were triggered by earthquakes (McHugh et al. 2006).
56 Mass-wasting occurrence and size increased during the end of the glaciation period
57 which may be attributed to thermal destabilization of gas hydrate or variations in
58 sediment supply (Zitter et al. 2012).

59 During the last glacial period, the SoM was a mildly brackish lake (e.g. Londeix
60 et al. 2009), as the global sea-level dropped below the level of the Çannakale sill
61 (e.g. Çağatay et al. 2009) at the beginning of Marine Isotopic Stage 4 at ~70 ka
62 (thereafter MIS-4). Sedimentation rates were at least two times higher during these
63 glacial times than during modern marine environment (Cagatay et al. 2003; Vidal
64 et al. 2010). Pore fluid composition profiles indicate diffusion of salts, between the
65 lacustrine sediments and the seawater above, affecting a layer of 12-40 m thickness
66 below the seafloor, that does not strongly depending on the sedimentation rates in
67 the area considered (Zitter et al. 2008; Çağatay et al. 2004).

68 Little is known about older paleo-environmental changes induced by sea-level
69 variations. Considering global sea-level variations (e.g. Lisieck and Raymo, 2005),
70 marine to lacustrine transitions may not have occurred during all the glacial
71 100,000-year cycles, as the timing of connection and disconnection depends not
72 only on global sea-level but also on the depth of the Çannakale sill, which probably
73 varied with time (e.g. Çağatay et al. 2009; Badertscher et al. 2011).

74 In this paper, the relationship between mass wasting and sea-level changes over
75 the last 400-500 ka is evaluated using high resolution seismic records. The Mass-

76 Transport Deposits (MTDs) related to sea-level oscillations and marine/lacustrine
 77 transitions are here characterized, and the possible MTDs trigger processes dis-
 78 cussed.

79 **2. Main observations and interpretations**

80 ***2.1. Overview of stratigraphic sequences and Mass-Transport*** 81 ***Deposits position***

82 The 3D seismic survey covers an area of 33 km² on the western part of the West-
 83 ern High, along the main strike-slip fault branch of the Main Marmara Fault (MMF,
 84 Le Pichon et al. 2001, Figure 1B). We focus here on the Eastern Fault-Zone Basin,
 85 an asymmetric basin bounded on its southern side by the MMF (Figure 1B). The
 86 seismic imaging (migration, bin size 6 m, Thomas et al. 2012) has a maximum sub-
 87 seabed seismic penetration of about 600 m in this basin. H0, the shallowest seismic
 88 horizon (Figure 2) encountered over most of the 3D HR survey, is at around 15 m
 89 depth on core MD01-2430 location, at the same depth as an ash layer(Lericolais and
 90 Henry, 2004) Mean sedimentation rates over this time-scale have been derived
 91 from the H0 mapping. The sedimentation rates range between 0.3 and 1.1 mm/a, in
 92 the Eastern Fault-Zone Basin. Considering the seismic penetration and the sedimen-
 93 tation rates, the dataset gives access to the history of sedimentation over a time scale
 94 of about 500 000 years.

95 Sedimentary sections consist of the repetition of four sequences, each sequence
 96 has between 0 and around 80 meters of thickness and contains two seismic units
 97 (Figure 2A and B) : (1) Draped units with moderate lateral variations of thickness,
 98 either acoustically transparent or bearing laterally continuous internal sub-parallel
 99 reflectors (2) Basin filling units displaying strong lateral variations of thicknesses
 100 and laterally correlative with either condensed sections (Figure 2B) or hiatuses on
 101 the slopes and topographic highs. Sedimentary bodies that contain irregular lenses
 102 and chaotic reflections are observed at the base and/or within the filling sequences
 103 (Figure 2A and B), and are interpreted MTDs (e.g. Alves et al. 2010).

104 Three remarkable MTDs thus occur within the basin filling
 105 units (Figure 2A). Prominent reflective horizons (H1, H4 and H6; Figure 2B) which
 106 present significant disruption, mark the base of the MTDs. The top horizons of the
 107 MTDs (H1⁰, H3 and H5, Figure 2B), pinch out against a continuous basal horizons
 108 and display small-scale topography. Horizon H5 displays erosive channel geometry
 109 that may have developed concurrently with the deposition of the basin filling unit
 110 (Figure 2A and B). Erosion affects the underlying MTDs and therefore must at least
 111 in part post-date the MTDs events. Horizon H3 is relatively smooth but small ero-
 112 sive channels affect H4 where H3 is absent, suggesting the MTD strata between H3
 113 and H4 is correlative with an erosion surface (Figure 2A and B).

114 The high-resolution age model from ^{14}C calibrated ages obtained from core
115 MD01-2430 (Figure 1B) indicates a sharp reduction of sedimentation rate from 0.6
116 to 0.2 mm/a following the reconnection at 14.7 ka (Vidal et al. 2010). In the Holo-
117 cene, sedimentation rates subsequently increased back to 0.4 mm/a, although still
118 not reaching the glacial sedimentation rate again (Vidal et al. 2010). We surmise
119 this has been the case during previous glacial cycles, at least in cases when the SoM
120 had been disconnected during a global sea-level low-stand, and possibly during all
121 episodes of rapid sea-level rise. In this case, the repetition of filling unit/draped
122 section corresponds to the sea level low and high stands oscillation at 100 ka time-
123 scale. In particular, the onlap sequence on H1' (Figure 2C and D), first low-
124 stand/filling units below the seafloor, probably results from a decrease of sedimen-
125 tation rate during the marine high stand at MIS-5 (~130 ka). The MTDs lying on
126 H1 thus occurred during the sea level fall at MIS-4 (~70 ka).

127 Variations of clay composition could explain differences of slope instabilities
128 frequency between marine and lacustrine rich-clay sediments. Powder of clay frac-
129 tion of 38 samples (of both marine and lacustrine muds) has been characterized by
130 X-ray diffraction. The relative amount of smectite, illite, kaolinite and chlorite were
131 determined by comparing the major XRD reflection peaks surfaces obtained on un-
132 treated, ethylene-glycolated, hydrazine monohydrate and heated smear slides. La-
133 custrine and marine clay compositions do not display significant differences. The
134 clays fraction is dominated by swelling clay as the clay fraction is composed of, on
135 average, 75% smectite, 13% illite, and around 5-7% kaolinite and chlorite.

137 ***2.2 Characterization of MTDs on H1***

138
139 The MTDs above H1 form a thick unstratified lens with a rough upper surface.
140 Within this lens, we observed locally some highly reflective patches interpreted as
141 stratigraphic reflectors within coherent blocks floating in a matrix of flowed mud
142 that has lost coherence, at least at the scale of the HR data (Figure 3D). The top of
143 the MTDs is a highly reflective and rough reflector of reverse polarity to that of the
144 seafloor, suggesting that the deposit has probably a higher content of fluid than the
145 upper sediments. The top of the MTD is undulated, and we observed small-scale
146 ridges of higher reflectivity (in absolute value) with a NE-SW strike (Figure 3B and
147 C). The geometry of the structures suggests the mud flow occurred perpendicular to
148 the western ridges (Figure 3A) along a NW-SE sliding direction. The MTD covers
149 an area of about 0.47 km² and has a total volume of 5.5 hm³. Along the NW edge
150 of the basin, the MTD thickness is around 7-8 m. It increases toward the southeast,
151 reaching a maximum thickness of 23 m in the central part of the basin (Figure 3C).
152 H1 displays variable reflectivity below the MTDs. On the whole western edge of
153 the MTDs but also its northeastern edge, H1 has negative polarity and high reflec-
154 tivity (Figure 3A). Below the central part of the MTDs, H1 appears composite with
155 several sub-reflectors resulting in a concentric reflectivity pattern (Figure 3A). On
156 the slope west of the MTD, H1 is absent, or displays low reflectivity, over a 0.28
157 km² area (labeled eroded area in Figure 3A). This area corresponds to a place where

158 H1 but also others horizons above have been eroded and thus appears as the proba-
159 ble main source of the material forming the slide masses. The thickness of sediment
160 removed during the event is estimated to 11 ± 5 m comparing the thickness of sedi-
161 ment on either side of the slide with the thickness of sediment in the scar (Figure
162 3D, section 1-1'). It also seems that most of the sediment deposited between H1 and
163 H0 has been removed. The volume of sediment removed is thus $3\pm 1.4 \times 10^6$ m³,
164 which represents only about 1/2 to 1/3 of the MTD volume. This paradox may be
165 resolved if sediments deposited above H1 in the basin have been displaced and in
166 part incorporated into the MTD. The eastern edge of the MTD is not sharp (Figure
167 3D, section 2-2') but appears as a gradational transition between disturbed sedi-
168 ments and an undisturbed layer of about 7-8 m thickness. Also, the base of the MTD
169 is not erosive in the deepest parts of the basin but appears to follow closely H1,
170 which may represent the basal sliding surface or layer immediately beneath it. In
171 fact the volume and geometry of the MTD may be accounted for, if a 7.4 m average
172 thickness layer of sediment covering both the area of sediment removal and the area
173 now covered by the MTD was redistributed by the sliding event to the deeper part
174 of the basin. This can suggest liquefaction occurred as the slide progressed, a com-
175 mon feature in slides and slumps, which can evolve into debris flow deposits
176 (Mulder and Cochonat, 1996).

177 **3. Discussions about possible triggering processes and** 178 **conclusions:**

179 During the last glacial cycle, the draped sequence in the basin corresponds to
180 low-stand lacustrine deposits with a relatively high sedimentation rate. Parts of the
181 underlying deposits, presumably marine sediments from the last interglacial (MIS-
182 5), have been redistributed by gravitationally driven mass transport, resulting in lat-
183 eral variations in sediment thickness. Sediment remobilization likely occurred after
184 sea-level dropped at MIS-4. This pattern of alternating draping sedimentation and
185 sediment redistribution is found to repeat cyclically in the stratigraphic record. This
186 suggests that comparable sequences of events occurred during earlier glacial cycles.
187 This is also supported by the thickness of these sequences, which would correspond
188 to about 100,000 years of sedimentation extrapolating the mean sedimentation rates
189 measured over the last ~29 ka.

190 Sediment fluxes into the deep basins of the Sea of Marmara from continental
191 shelves and slopes are higher during low stand than during high stand, and slope
192 instabilities clustered at the end of the glaciation (Zitter et al. 2012). However, the
193 Western High is relatively isolated from this influence, as canyons originating from
194 the shelf edges bypass the Western High and supply sediments to the deep Central
195 and Tekirdag basins (Figure 1A). Nevertheless, we still observe cyclic local redis-

196 tribution of sediments between slopes and ponded basins but probably with a differ-
197 ent timing as the MTD in this study is inferred to correlate with the beginning of a
198 low stand.

199 At the beginning of MIS-4 (~70 ka), sea-level fell from -25 m to -85 m below
200 the modern sea level resulting in the disconnection of the Sea of Marmara (Çağatay
201 et al. 2009). This also induced a pressure drop of 0.6 MPa and thus possible disso-
202 ciation of gas hydrate (Liu & Flemings, 2009). Nonetheless, a temperature decrease
203 of 1°C at the seafloor of the Western High associated to this pressure drop should
204 be sufficient to keep hydrates in their stability field (e.g, Henry et al. 1999). It is
205 thus unlikely that widespread gas hydrate dissociation occurred at this time. On the
206 other hand, pore fluid freshening by salt diffusion must have occurred. Since the
207 post-glacial reconnection, water column salinization (that has been progressive over
208 2000 years, Vidal et al. 2010) lead to the diffusion of salts downward into the pore
209 fluids of the lacustrine sediments (Zitter et al. 2008) and the reverse process had
210 occurred during the last glaciation after the disconnection. The diffusion length be-
211 ing proportional to the square root of time, the salt diffusion gradient below the
212 seafloor affects a sediment layer of thickness 10 to 100 m for times ranging from
213 1000 to 100.000 years. A layer of 11 ± 5 m of marine sediments has been remobi-
214 lized, which appears consistent with the thickness of leached sediments after only a
215 few thousand years. As the clay-fraction is dominated by smectite, clay swelling
216 induced by osmotic stress during the desalinization of the pore water is expected
217 (Charpentier and Bourrié, 1997), reducing the cohesion of sediment and promoting
218 submarine slope failures (Masson et al. 2006). Consequently, systematic destabili-
219 zation of marine sediments on slopes at the beginning of lacustrine periods should
220 occur.

221 We remarked that the cyclic occurrence of MTDs observed in a ponded basin of
222 the Western High in the SoM could correlate with glacial-interglacial sequences.
223 Despite that slope instabilities are presumably triggered by earthquakes, their size
224 and/or frequency appears to be modulated by glacio-eustatic oscillations and the
225 lacustrine/marine transitions associated to them. The swelling of clays in marine
226 sediments under low salinity brackish water can promote slope instability after ma-
227 rine to lacustrine transitions, and may be one factor contributing to long-term cy-
228 clicity of slope instability occurrence in the Sea of Marmara.
229

230 **Figure captions**

231 Figure 1 : A. Tectonic map of the Sea of Marmara region with the Main Marmara Fault (MMF, Le
 232 Pichon et al. 2001) and secondary fault system (Grall et al. 2012). The red box locates the 3D
 233 seismic survey data. B. Seafloor mapping from the 3D-seismic data, and the location of the Eastern
 234 Fault-Zone basin (annotated EFZ-basin). Locations of the seismic lines present on Figure 2,
 235 MD01-2430 core, and sites where gas hydrates have been sampled are shown. The black box
 236 indicates the area shown in Figure 3A.

237 Figure 2 : A. Seismic section across the northern edge of the Eastern Fault-Zone Basin. MTDs are
 238 highlighted in yellow, and numbers 1, 2, 3 refer to the three most remarkable MTDs discussed in
 239 the text. B. Line drawing of the seismic section above. Horizon numbers mentioned increase from
 240 top to bottom. Stratigraphic sequences contain one basin filling unit (in grey) in which MTDs
 241 occurred (indicate by black stars), and one draped section (in white). The four main stratigraphic
 242 sequences in this characteristic seismic section have a thickness of around 65-75 ms twt, which
 243 represent around 50-63 m (P-wave velocity ranges between 1515 (above H1) and 1660 m/s
 244 (between H6 and H4), within the sediments, Thomas et al, 2012). C: Seismic section across the
 245 Eastern Fault-Zone Basin, showing details of the first stratigraphic sequence below the seafloor.
 246 D. Line drawing of the seismic section above. The filling unit contains a condensed section (H1'
 247 H1) on which MTDs occurred (between H1 and H1⁰). The condensed section extension
 248 corresponds to the basin depocentre (represented in Figure 3A) when the slides occurred. See
 249 Figure 1C for the location of the seismic lines.

250 Figure 3 : Mapping and characteristics of the first remarkable MTDs below the seafloor, which
 251 probably occurred after the sea-level fall at MIS-4 (~70 ka). A. Amplitude map of the basal surface
 252 of the MTDs (horizon H1). This map is located in Figure 1B. Eroded area and area of deposits
 253 (which is detailed in Figure 2B and C) correspond to the main scars of the slides and the zone of
 254 slide depositions. The depocentre is derived from the mapping of the condensed section below the
 255 MTDs (Figure 2C and D). B. Amplitude map of the top horizon of the MTDs. Amplitude anomalies
 256 alignment is indicated by white dash lines. C. Topography of the top horizon of the MTDs. The
 257 white dash lines and the black arrows indicate the orientation of small undulations and the flow
 258 lines respectively. D. Seismic sections along the scars of the slides (1-1') and across the MTDs (2-
 259 2').

260

261 **Acknowledgements**

262 The Marmara-DM cruise was supported by ESONET NoE, Network of Excel-
 263 lence, coordinated by IFREMER and co funded by EU as part of FP6 (2009). This
 264 work is also co-funded by CNRS. We would like to thank Cedric Amiel, for his
 265 participation to the XRD laboratory manipulation. We thank Cecilia McHugh and
 266 Lies Loncke for their constructive comments that improved this manuscript.

267 **4. References:**

268 Alves TM (2010), 3D Seismic examples of differential compaction in mass-transport deposits and
 269 their effect on post-failure strata, *Marine Geology*, 271: 212-224.

- 270 Badertscher S, Fleitmann D, Cheng H et al (2011), Pleistocene water intrusions from the Mediter-
271 ranean and Caspian seas into the Black Sea, *Nature Geosci*, 4: 236-239.
- 272 Beck C, Mercier de Lepinay B, Schneider JL et al (2007), Late Quaternary co-seismic sedimenta-
273 tion in the Sea of Marmara's deep basins, *Sedimentary Geology*, 199: 65-89.
- 274 Bourry C, Chazallon B, Charlou JL et al (2009), Free gas and gas hydrates from the Sea of Mar-
275 marama, Turkey: Chemical and structural characterization, *Chemical Geology*, 264: 197-206.
- 276 Cagatay MN, Ozcan M, Gungor E et al (2004), Pore-water and sediment geochemistry in the Mar-
277 marama Sea (Turkey): early diagenesis and diffusive fluxes, *Geochemistry-Exploration Environ-*
278 *ment Analysis*, 4: 213-225.
- 279 Cagatay MN, Eris K, Ryan WBF et al (2009), Late Pleistocene-Holocene evolution of the northern
280 shelf of the Sea of Marmara, *Marine Geology*, 265: 87-100.
- 281 Charpentier, S. Bourrié, G. (1997), Deformation of saturated clays under mechanical and osmotic
282 stress and its relation with the arrangement of the clays, *European Journal of Soil Science*, 48:
283 49-57.
- 284 Grall C, Henry P, Tezcan D et al (2012), Heat flow in the Sea of Marmara Central Basin: Possible
285 implications for the tectonic evolution of the North Anatolian fault, *Geology*, 40: 3-6.
- 286 Henry P, Thomas M, Ben Clennell, M (1999), Formation of natural gas hydrates in marine sedi-
287 ments 2. Thermodynamic calculations of stability conditions in porous sediments, *J. Geophys.*
288 *Res.*, 104 : 23005-23022.
- 289 Lericolais, G., Henry, P., 2004. Cruise report of Marmara VT / Marmacore 2. <http://www.cdf.u-3mrs.fr/~henry/marmara/marmaraVT.html>2004.
- 290 Le Pichon X, Sengor AMC, Demribag E, et al (2001), The active Main Marmara Fault, *Earth and*
291 *Planetary Science Letters*, 192: 595-616.
- 292 Lisiecki LE., Raymo ME (2005), A Pliocene-Pleistocene stack of 57 globally distributed benthic
293 $\delta^{18}O$ records, *Paleoceanography*, 20: PA1003.
- 294 Londeix, L., Herreyre Y, Turon JL, et al (2009), Last Glacial to Holocene hydrology of the Mar-
295 marama Sea inferred from a dinoflagellate cyst record, *Review of Palaeobotany and Palynology*,
296 158: 52-71.
- 297 Maslin M, Owen M, day S, et al (2004), Linking continental-slope failures and climate change:
298 Testing the clathrate gun hypothesis, *Geology*, 32: 53-56.
- 299 Masson DG, Harbitz CB, Wynn RB, et al (2006), Submarine landslides: processes, triggers and
300 hazard prediction, *Philosophical Transactions of the Royal Society A: Mathematical, Physical*
301 *and Engineering Sciences*, 364 : 2009-2039.
- 302 McHugh CMG, Seeber L, Cormier MH et al (2006), Submarine earthquake geology along the
303 North Anatolia Fault in the Marmara Sea, Turkey: A model for transform basin sedimentation,
304 *Earth and Planetary Science Letters*, 248 : 661-684.
- 305 Menot, G., and E. Bard (2010), Geochemical evidence for a large methane release during the last
306 deglaciation from Marmara Sea sediments, *Geochimica Et Cosmochimica Acta*, 74 : 1537-
307 1550.
- 308 Mulder T, Cochonat P, Classification of offshore mass movements, *Journal of Sedimentary Re-*
309 *search*, 66: 43-57.
- 310 Rothwell RG, Thomson J, Kahler G. (1998), Low-sea-level emplacement of a very large late Pleis-
311 tocene "megaturbidite" in the Western Mediterranean Sea, *Nature*, 392 : 377-380.
- 312 Seeber L, Cormier MH, McHugh C, et al (2006), Rapid subsidence and sedimentation from oblique
313 slip near a bend on the North Anatolian transform fault in the Marmara Sea, Turkey, *Geology*,
314 34 : 933-936.
- 315 Thomas Y; Marsset B; Westbrook GK.;et al (2012), Contribution of high-resolution 3D seismic
316 near-seafloor imaging to reservoir-scale studies: application to the active North Anatolian
317 Fault, Sea of Marmara, *Near Surf. Geophys.*, 10 : 291-301.
- 318 Vidal, L., Menot G., Joly C. et al (2010), Hydrology in the Sea of Marmara during the last 23 ka:
319 Implications for timing of Black Sea connections and sapropel deposition, *Paleoceanography*,
320 25: PA1205.
- 321 Wong, H. K., Ludmann T., Ulug, A. et al (1995), The Sea of Marmara - a Plate Boundary Sea in
322 an Escape Tectonic Regime, *Tectonophysics*, 244 : 231-250.
- 323

324 Zitter TAC; Henry P; Aloisi G; et al (2008), Cold seeps along the main Marmara Fault in the Sea
 325 of Marmara (Turkey), Deep-Sea Research Part I-Oceanographic Research Papers, 55: 552-570.
 326 Zitter T AC., Grall C; Henry, P et al (2012), Distribution, morphology and triggers of submarine
 327 mass wasting in the Sea of Marmara, Marine Geology, 329: 58-74.
 328

329 **5. Figures:**

330
 331





