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# The geometry of the North Anatolian transform fault in the Sea of Marmara and its temporal evolution: implications for the development of intracontinental transform faults

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## ► To cite this version:

A. M. Celâl Şengör, Céline Grall, Caner Imren, Xavier Le Pichon, Naci Görür, et al.. The geometry of the North Anatolian transform fault in the Sea of Marmara and its temporal evolution: implications for the development of intracontinental transform faults. *Canadian journal of earth sciences*, 2014, 51 (3), pp.222 - 242. 10.1139/cjes-2013-0160 . hal-03455601

**HAL Id: hal-03455601**

**<https://hal.science/hal-03455601>**

Submitted on 1 Dec 2022

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1 THE GEOMETRY OF THE NORTH ANATOLIAN TRANSFORM FAULT IN THE  
2 SEA OF MARMARA AND ITS TEMPORAL EVOLUTION: IMPLICATIONS FOR  
3 THE DEVELOPMENT OF INTRACONTINENTAL TRANSFORM FAULTS

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18 *In pietate memoriam J. Tuzo Wilson*

19 ABSTRACT

20 The North Anatolian Fault is an 1200-km long strike-slip fault system connecting the  
21 East Anatolian convergent area with the Hellenic subduction zone and as such  
22 represents an intracontinental transform fault. It began forming some 13 to 11 Ma ago  
23 within a keirogen, called the North Anatolian Shear Zone, that becomes wider from  
24 east to west. Its width is maximum at the latitude of the Sea of Marmara, where it is 100  
25 km. The Marmara Basin is unique in containing part of an active strike slip fault system  
26 in a submarine environment in which there has been active sedimentation in a  
27 Paratethyan context where stratigraphic resolution is higher than elsewhere in the  
28 Mediterranean. It is also surrounded by a long-civilised rim where historical records  
29 reach well into the second half of the first millenium BCE. In this study we have used  
30 210 multichannel seismic reflexion profiles adding up to 6210 km profile length and  
31 high resolution bathymetry and chirp profiles reported in the literature to map all the  
32 faults that are younger than the Oligocene. Within these faults we have distinguished  
33 those that cut the surface and those which do not. Among the ones that do not cut the  
34 surface, we have further created a time-table of fault generation based on seismic  
35 sequence recognition. The results are surprising in that faults of all orientations contain  
36 subsets that are active and others that are inactive. This suggests that as the shear zone  
37 evolves, faults of all orientations become activated and deactivated in a manner that

38 now seems almost haphazard, but a tendency is noticed to confine the overall  
39 movement to a zone that becomes narrower with time since the inception of the shear  
40 zone, i.e. the whole keirogen, at its full width. In basins, basin margins move outward  
41 with time, whereas highs maintain their faults free of sediment cover making their  
42 dating difficult, but small perched basins on top of them in places make relative dating  
43 possible. These basins in addition permit comparison of geological history of the highs  
44 with those of the neighboring basins. The two westerly deeps within the Sea of  
45 Marmara seem inherited structures from the earlier Rhodope-Pontide  
46 fragment/Sakarya continent collision, but were much accentuated by the rise of the  
47 intervening highs during the shear evolution. When it is assumed that below 10 km  
48 depth the faults that now constitute the Marmara fault family might have widths  
49 approaching 4 km, the resulting picture resembles a large version of an amphibolite-  
50 grade shear zone fabric, an inference in agreement with the scale-independent structure  
51 of shear zones. We think that the North Anatolian Fault at depth has such a fabric not  
52 only on a meso, but also on a macro scale. Detection of such broad, vertical shear zones  
53 in Precambrian terrains may be one way to get a handle on relative plate motion  
54 directions during those remote times.

## 55 INTRODUCTION

56 In 1965, J. Tuzo Wilson defined a new kind of fault that could not be accounted for by  
57 Anderson's theory of faulting in a continuous medium (Anderson, 1951; Wilson's 1965

58 paper was also the founding paper of plate tectonics). Wilson pointed out that the new  
59 class of faults served to connect two plate boundaries (the word 'plate' is Wilson's) with  
60 one another by transforming the motion along them into another kind of motion (Fig.  
61 1). He therefore called them 'transform faults' (Wilson, 1965). Because plate boundaries  
62 are so cleanly delineated in the oceans, there is seldom any doubt as to the nature of  
63 faulting in them, but the scattering of deformation in the continents into areas of  
64 complex strain, in places as wide as 3000 km (as in Asia now), makes it difficult to  
65 decide to what kind of fault class some of the large strike-slip faults moving in such  
66 regions should be assigned.

67 In the case of the North Anatolian Fault (Ketin, 1948, 1969, 1976; Pavoni, 1962;  
68 McKenzie, 1972; Şengör, 1979; Barka, 1992, 1996; Şengör et al., 2005), it is clear that it  
69 connects the East Anatolian zone of convergence (Şengör and Kidd, 1979; Dewey et al.,  
70 1986; Şengör et al., 2003; 2008) with the Hellenic Trench (Dewey & Şengör 1979,  
71 McKenzie and Jackson 1983, Le Pichon et al. 1993; Şengör 2011) and as such qualifies as  
72 a transform fault (Şengör, 1979; Le Pichon and Kreemer, 2010). It consists of a numerous  
73 family of faults going from the well-defined Karlıova Basin in the East (Şengör, 1979;  
74 Şaroğlu, 1985; Şengör et al., 1985) to the more diffuse transtensional region in the  
75 Northern Aegean (Fig. 2). Throughout its length, its various branches have been  
76 mapped in the field (see especially Herece and Akay, 2003), and it is generally agreed

77 that the broad region of deformation, a veritable keirogen<sup>1</sup>, called the North Anatolian  
78 Shear Zone (Fig. 3; Şengör et al., 2005; Le Pichon et al., in press), in which the fault now  
79 moves, came into existence sometime during the later medial Miocene between the  
80 Serravallian and the Tortonian (see especially Şengör et al., 2005; Görür and Elbek,  
81 submitted; between the Sarmatian and the early Pannonian in the Paratethyan scale and  
82 between MN7/8 and MN10 in the European mammal stages; much of the deposits  
83 along the North Anatolian Shear Zone have been dated using these two latter  
84 standards). What so far has been unclear is how the broad keirogen became localised  
85 into the narrow, single-strand (double strand west of Yeniçağa near Bolu: Fig. 2) main  
86 displacement zone(s), constituting the North Anatolian Fault. To study the manner of  
87 that localisation, the Sea of Marmara appears to be an ideal location, because a large  
88 number of both multi-channel seismic and chirp profiles, shallow gravity cores (down  
89 to 30 m from the sea floor, some of which are now dated and enabled assessments to be  
90 made of the sedimentological events recognized in them: Beck et al., 2007; Eriş et al.,  
91 2012 , Çağatay et al., in press), a number of deeper hydrocarbon exploration wells,  
92 recent seismic studies including the water column itself, plus a complete, detailed  
93 bathymetric coverage are available (for references prior to 2005, see Şengör et al., 2005;

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<sup>1</sup> A *keirogen* (from ancient Greek κείρω = shear) is abroad strike-slip shear zone in which more than one major strike-slip fault and associated structures take up the shear and it is the strike-slip counterpart of the shortening *orogens* and extending *taphrogens*. It was first defined in Şengör and Natal'in, 1996, p. 639, endnote 8)

94 Armijo et al., 2005; Carton, 2005; Gazioğlu et al., 2005; Cormier et al., 2006; Seeber et al.,  
95 2006; Carton et al., 2007; Dolu et al., 2007; Géli et al., 2008; Laigle et al., 2008; Bécél, 2009;  
96 Kuşçu, 2009; Bécél et al., 2010; Özeren et al., 2010; Şengör et al., 2011; Gasperini et al.,  
97 2012; Shillington et al., 2012; Sorlien et al., 2012; Zitter et al., 2012; Le Pichon et al., in  
98 press).

99 Moreover, the Marmara Sea and its surroundings are frequently visited by earthquakes  
100 of devastating size with tragic consequences and the long historical records kept since  
101 at least the second half of the first millennium BCE (Fig. 4) give us an unparalleled  
102 knowledge of their repeat times and natures. That yet another imminent large  
103 earthquake is now threatening İstanbul, one of the megalopolises of the world and a  
104 cultural treasure house of mankind (it is the oldest continuously inhabited human  
105 dwelling in the world, since ~250,000 BCE, when the cave dwellers discovered its  
106 advantages as a location: see Howell et al., 2010) has focused the attention of geologists,  
107 geographers and geophysicists on this Sea after the disastrous 1999 earthquakes in İzmit  
108 and Düzce in northwestern Turkey (see Şengör et al., 2005). The fact that the Marmara  
109 area has been a part alternately of the Tethyan and Paratethyan realms since the  
110 beginning of the Miocene has given it a stratigraphy of a resolution higher than the  
111 normal Tethyan regions, because of the complete and abrupt change of biotas during  
112 every transition (so far used only for the youngest sequences and for the frame of the  
113 Sea, except where well information has been released by the petroleum industry). All of

114 these factors make the Marmara region an ideal site to study the evolution of a strike-  
115 slip system below it.

116 We have critically reviewed the maps of the Marmara fault segments (e.g. Parke et al.,  
117 1999, 2002; Le Pichon et al., 2003; Rangin et al., 2004, Grall et al., 2012, Sorlien et al.,  
118 2012) by using that database (Fig. 5A) consisting of 210 multichannel seismic reflexion  
119 profiles adding up to 6210 km profile length, so far unpublished except some segments  
120 of a number of the profiles that had been published by Ateş et al. (2003) and high  
121 resolution bathymetry and chirp profiles reported in the literature to map all the faults  
122 that appear to cut the Miocene to Recent sedimentary rocks in the Sea of Marmara. We  
123 then distinguished those that cut only a portion of the sedimentary sequence and  
124 tentatively established a crude sequence of faulting events, according to the state of  
125 knowledge about the seismic stratigraphy of the Sea of Marmara (the crudeness being a  
126 function of the local paucity of direct litho- and biostratigraphic control). Finally, we  
127 have correlated the faults that cut the present-day sea-floor with the microearthquake  
128 epicentres between the years 2005 and 2011 plus the fault-plane solutions of all the  
129 earthquakes with magnitudes greater than 4 with a view to seeing which faults may  
130 have localised motion, and, where information is available, what kind, since 1960. In the  
131 sections that follow, we describe the faults illustrated in Fig. 5B and give examples of  
132 various faulting events established using seismic sequence stratigraphy.

133 Stratigraphic sequences that are related to 100 ka glacial/interglacial cyclicity over the



134 last 450-540 ka are here depicted following the recent publications by Grall et al. (2012  
135 and in press b) and Sorlien et al., (2012). These authors show repetitions of onlap  
136 geometries (laterally correlated with condensed sections, Grall et al., in press b), and  
137 forset/toplap geometries (Sorlien et al., 2012) and related them to variations of  
138 sedimentation rates during low-stand/high-stand transitions and low-stand deltas  
139 respectively. These authors use the following colour code associated to Marine Isotopic  
140 Stage (MIS) transitions: the Red MIS-5 (140 ka), the Blue MIS-7 (240 ka), the Yellow MIS-  
141 9 (340 ka), the Purple MIS-11(445 ka) and the Green MIS-13 (570 ka) transitions (Grall et  
142 al., in press b). We adopt this colour code, denoting by these colours the sequence on  
143 top of the sequence boundary. In this paper we designate our own figures with a capital  
144 F as Fig. and figures we cite from the literature with a lower case f as fig.

#### 145 THE EASTERN SECTOR

146 Fig. 6 illustrates the faults we have mapped and Fig. 5A shows the locations of the  
147 multichannel seismic profiles that were used to identify them. Fig. 6 shows the eastern  
148 sector of our study area consisting essentially of the Çınarcık Basin (Le Pichon et al.,  
149 2001; Ateş et al., 2003; Carton, 2005; Şengör et al., 2005; Laigle et al., 2008; Uçarkuş, 2010)  
150 and regions to its north (Gökçeoğlu et al., 2009; Özeren et al., 2010; for the neotectonics  
151 of the land area see Şengör, 2011) and south (including the İmralı Basin: Okay et al.,  
152 2000; Sorlien et al., 2012). Çınarcık Basin is a critical area because the exact manner of  
153 the westward prolongation of the faults from it to the Central High remains unclear.

154 In this sector we have chosen profiles 9 (Fig. 7) and 13 (Fig. 8) to illustrate the relative  
155 age sequence of the faults (for locations of the profiles, see Fig. 5A). Fig. 7 shows that  
156 along the profile 9, the main marginal fault cuts to the surface. The seismicity also  
157 shows it to be active localising almost exclusively right-lateral earthquakes (Figs. 6B and  
158 C; Örgülü and Aktar, 2001; Özalaybey et al., 2002, 2003; Pınar et al., 2003). In Fig. 6, the  
159 focal mechanism solution 44 is not readily interpretable in terms of simple shear zone  
160 geometry, but it is still compatible with a tension gash orientation, albeit with a tiny  
161 sinistral component. We note that the term 'tension gash' is here used to denote a  
162 *structure orientation as seen in micro-and meso-scales, where the structure was originally*  
163 *liquid-filled*. The structures we describe are not necessarily 'tensional' but definitely  
164 'extensional' with variable amounts of strike-slip associated with them (from 0% to 50%  
165 in some rare cases). As we have not read the seismograms ourselves we cannot further  
166 comment on that solution. If it is a 'tension gash' orientation normal fault only, then it  
167 is also compatible with pure strike-slip along the northern margin of the Çınarcık Basin.

168 The entire sedimentary fill seen in this basin is of Pleistocene age: In gravity core 27 (for  
169 location, see Fig. 7B), a radiocarbon age of 16.1 to 16.5 ky BP Cal. was obtained at a  
170 depth of 25 metres. In core 25, closer to the main bounding fault to the north, two ages  
171 of 14.3-16 and 15.0-16.5 ky BP Cal. were obtained from depths of 28 and 32 metres,  
172 respectively (Beck et al., 2007). They indicate a mean sedimentation rate of 1.5-2 mm/a  
173 during the Holocene within this basin. These ages make the entire sedimentary section

174 as seen in Fig. 7 Pleistocene in age (Şengör et al., 2005) with the exception of the first 15  
175 to at most 20 metres from the sea-floor, which represent Holocene sedimentation.

176 The fault to the south of the main fault in this profile (Fig. 7) cuts only up into a low  
177 amplitude layer at around 1.5 s twt, but does not penetrate farther up. On the evidence  
178 of the seismic profile that fault seems to have died within the Pleistocene (since 540-450  
179 ka). Neither is there any evidence of its presence in the Chirp profile published by Zitter  
180 et al. (2012). However, there is seismicity correspondent to its location and available  
181 fault-plane solutions associated with that seismicity (solutions 33, 36, 39 in Fig. 6C) all  
182 indicate dextral strike-slip; the fault-plane solution 36 indicates a very slight extensional  
183 component either on a dextral Riedel shear (R) or on a sinistral anti-Riedel (R') shear,  
184 both entirely compatible with the main shear zone (the Y-shear direction) being a right-  
185 lateral one (see Fig. 9 for the terminology we use for various shear zone structures).  
186 These earthquakes indicate a non-negligible strain being taken up on the fault that on  
187 the seismic profile 9 does not seem to reach the surface. The earthquakes' magnitudes  
188 are small enough and the data of the multichannel seismic profiles are coarse enough  
189 that the sea bottom's either not being breached or appearing not to have been breached  
190 is entirely understandable. That the fault does not show up even on the Marnaut 54  
191 Chirp profile (Zitter et al., 2012) may suggest that in levels shallower than the green  
192 sequence the fault may branch out and even merge with the main boundary fault and  
193 reaches the surface via that route.

194 A south dipping fold and thrust along the southern margin of the Çınarcık Basin  
195 separated two small basinal structures (Fig. 7 and Fig. 5B). Farther east there is an Ms  
196 6.4 earthquake that gave a normal fault solution (solution labelled T in Fig. 7C: Taymaz  
197 et al., 1991) where shallow transtensive faults could be mapped on both bathymetry and  
198 chirp sub-bottom profiles, corroborating the view that at least some of the faults in this  
199 set are active (Fig. 5B). The microearthquake activity along these southern normal faults  
200 seems livelier than in the north since 2005, but the southern normal faults do not  
201 generate as large earthquakes as the northern main strike-slip strand, which is what is  
202 expected from their respective rates of motion (see Şengör et al., 2005).

203 Within the Çınarcık Basin, the northern master fault takes up by the largest portion of  
204 the movement which is nearly perfectly parallel with it as seen both from the  
205 earthquake fault plane solutions and from the GPS data (see Şengör et al., 2005 for a  
206 summary of the GPS vectors here) and there is no indication of major oblique-slip,  
207 contrary to what is implied by Seeber et al. (2006), who over-estimated the oblique-slip  
208 component at the crustal level, as they did not take into account the compaction, which  
209 in a mature basin, could result in a seafloor subsidence 2 to 3 times larger than the  
210 subsidence of the basement. We have not taken into account in our study Hergert and  
211 Heidbach's (2010) modelling of the mechanical behaviour of the lithosphere in the  
212 Marmara Region which requires derivation of stresses. Such a derivation is hardly  
213 possible (except in a model), given the uncertainties concerning the properties of the

214 real lithosphere and the stresses acting in it, but if one wished to invoke that model, the  
215 northern margin of the Çınarcık Basin would still require less than 4.5 mm/yr normal  
216 component of motion within its constraints, which we find reassuring.

217 Farther south, west of the Armutlu Peninsula, there is the southerly-to-southeasterly-  
218 tilted İmralı Basin (Okay et al., 2000), also of Pleistocene age. Microearthquake activity  
219 (Fig. 6B and C) is much subdued here, only a few events falling directly on the profile.  
220 This shows that even the ~~inactive~~ faults probably move, albeit rarely, during small  
221 magnitude earthquakes, but contribute essentially nothing to the further straining of the  
222 area. Along its easterly strike continuation, on the Samanlı Mountains (for their detailed  
223 geomorphology, see Bilgin, 1967), there is a livelier earthquake activity. The fault plane  
224 solutions available from there readily fall into two groups: a western group (fault plane  
225 solutions 16, 45, 47, 49, 51 in Fig. 6C) and an eastern group (solutions 53 and 54). It  
226 seems that the western group faults stretch the ground in a NNE-SSW direction and are  
227 related to those that created the İmralı Basin and the ones farther east cause extension in  
228 a NE-SW direction and are associated with the fault family that caused the Ms=6.4  
229 Çınarcık earthquake farther north. Within the İmralı Basin, the main faults strike WNW-  
230 ESE and are related to the faults that cause the western group of earthquakes in the  
231 neighbouring Samanlı Mountains. But they are probably not mainly responsible for the  
232 origin of the İmralı Basin. Otherwise, the neighbouring Samanlı Mountains with similar  
233 and active faults would not have stood as high as they do now.

234 Fig. 8A shows profile 13 which is farther west than the previous one (Fig. 7A) and Fig  
235 8B our interpretation of it. The recognisable sequences indicate a much narrower basin  
236 of a similar age to the one we saw farther east. Although this sub-basin is bounded by  
237 faults that cut to the sea-floor there is one between them that is much older, possibly  
238 pre-Pleistocene. The southerly-tilted basin here appears completely inactive, as none of  
239 the faults that appear related to its formation cut to the surface. The spread of  
240 microseismicity (Figs. 6B and C) corroborates this picture showing only a few scattered  
241 epicentres.

242 Sorlien et al. (2012) have presented a meticulous and exhaustive analysis of the seismic  
243 stratigraphy of the Imrali Basin fill with a view to establishing a chronostratigraphy that  
244 they then used to argue that faulting continued at a constant rate since about half a  
245 million years ago. This is based on the date of what they call the red horizon, which has  
246 been correlated other most of the Sea of Marmara and has been attributed an age of 109  
247 ka based on extrapolation from Marion-Dufresne cores (Sorlien et al., 2012; Grall et al.,  
248 in press b). Lower reflectors have been correlated to low stand deltas and age attributed  
249 with the assumption that the sediment input into the Imrali basin has been of the same  
250 order for each glacial cycle. However, this reasoning also led to the conclusion that the  
251 low-stand delta observed in stratigraphic succession do not systematically correspond  
252 to successive glacial cycles. In the absence of independent age constraints, ambiguities  
253 in the age attributions thus cannot be resolved. In detail, Sorlien et al. (2012) age model

254 taken as face value does suggest some slowing down of the subsidence over the last 130  
255 to 330 ka once compaction is taken into account. In both profiles it is difficult to  
256 ascertain whether the basin fill of the Çınarcık and the İmralı basins is conformable or  
257 unconformable with their stratigraphic basement. Because at least one fault is truncated  
258 by the bottom sequence (Fig. 8B) and some reflectors below the lowest sequence seem to  
259 have a geometry in places not exactly parallel with the basement layering, we assume  
260 an unconformity bounding the basin at the bottom, which is, however, not easy to  
261 defend conclusively on the basis of the available observations.

262 Interestingly, once one goes southward over the Samanlı Mountains on the Armutlu  
263 Peninsula (SA in Fig. 5A), the microseismicity again becomes lively in the Gulf of  
264 Gemlik (GG in Fig. 5A). A major fault strand, which we here term the Gemlik-Bandırma  
265 Fault, goes from there to the Erdek tombolo (ET in Fig. 5) and then turns  
266 southwestward in the county Gönen into the Biga Peninsula (Fig. 5). Young scarps all  
267 along the fault suggest recent activity and at least two estimated  $M=7$  earthquakes (128  
268 and 543 CE) seem to have taken place on it during historical times (Fig. 4). Kuşçu et al.  
269 (2009) mapped a group of active faults in the Gulf of Gemlik and microearthquakes  
270 align along smaller faults on the isthmus separating Lake İznik and the Gulf of Gemlik,  
271 all betraying a lively activity although not accomplishing much strain (no earthquakes  
272 above magnitude 4), except obviously at intervals of a minimum or a millennium-and-  
273 a-half. Although Barka and Kadinsky-Cade (1988) called it the 'middle branch of the

274 North Anatolian Fault', it is much less significant than the two major strands originally  
275 identified by Dewey and Şengör (1979) and Şengör (1979) and is not much different  
276 from the many deactivated faults populating the broad southern shelf of the Sea of  
277 Marmara (Fig. 5A). Therefore, we go along the recent trend in following Dewey and  
278 Şengör (1979) and Şengör (1979) and stick to two major strands, but we nevertheless  
279 point out that the Gemlik-Bandırma fault does maintain an activity, although much  
280 subdued.

## 281 THE CENTRAL SECTOR

282 This sector (Fig. 10), as herein used, corresponds mostly to what is called the Central  
283 High in the literature (it is the high region that completely surrounds the Kumburgaz  
284 Basin, KB; Le Pichon et al., 2001; Şengör et al., 2005) but that also includes the  
285 Kumburgaz Basin and almost the whole of the Central Basin (Le Pichon et al., 2001;  
286 Şengör et al., 2005). The presence of the Kumburgaz Basin atop the Central High  
287 provides a sediment receptacle and thus a welcome means to date the faults cutting the  
288 Central High relative to one another (see especially Sorlien et al., 2012, fig. 6).

289 Profile 23 (Fig. 11A and B) is located in the eastern and central part of the basin and  
290 shows that both its bounding faults there cut to the surface (Fig. 11B). The basin also has  
291 other faults in it, but they seem to have died shortly after the basin initially formed. We  
292 also see that the basin was initially narrower. However, the faults that bound the basin  
293 on this profile are not parts of a single fault family: the northern fault is along the strike



294 of the northern boundary fault of the Çınarcık Basin coming from the east, here forming  
295 a part of the main zone of displacement along the northern branch of the North  
296 Anatolian Fault (Şengör et al., 2005). The southern fault, however, has an WNW strike  
297 and seems to be a somewhat rotated, extensional Riedel shear so identified simply on  
298 account of its orientation. Immediately to its south there is another fault similar to it that  
299 also bounds the basin. Both of these faults cut to the surface and both are active as  
300 shown by the microseismicity they localise. To their east, the southern basin margin  
301 seems unfaulted as can be seen on fig. 6 by Sorlien et al. (2012). Sorlien et al. (2012) carry  
302 what they call the Red-1 reflection into the Kumburgaz Basin where it is clearly cut  
303 twice, whereas in the İmralı Basin it is not, where it records the lower boundary of a  
304 non-faulted floor of a sedimentary infill. Therefore, although a throughgoing main  
305 strike-slip fault appears to have already formed in the Kumburgaz Basin, the auxiliary  
306 faults, such as the two Riedels just mentioned, appear not to have not terminated their  
307 activity. They probably do not localise major shocks (no earthquakes above magnitude  
308 4) and therefore do not take up much displacement, but they nevertheless remain active  
309 and it is hard to tell up to what size earthquakes they might one day localise.

310 Profile 27 (Fig. 12A and B) shows a much wider basin, of almost bovine-head shape  
311 when considered together with the narrower fault trough underlying it, and numerous  
312 faults cut its most recent sequence. Although the major displacement is along the  
313 northern master fault bounding all the sequences and reaching the sea-floor, it seems

314 that during the deposition of the youngest sequence, the basin had already become  
315 much wider. The numerous faults to the south, sub-parallel with the two previously-  
316 mentioned Riedel shears, are all either active or terminated their activity very recently  
317 during the deposition of the top sequence. Many of these faults seem to be short,  
318 therefore insignificant, in terms of the displacement they can accomplish, but they  
319 nevertheless are mostly active.

320 When we look farther south, an active fault that strikes about N45°W, thus in an  
321 extensional orientation in the overall North Anatolian Shear Zone, is seen to be active  
322 with microearthquake epicentres on it. Yet it sits perfectly on strike with, and seems to  
323 be the northwesterly continuation of, a major inactive fault extending down into the  
324 Gulf of Gemlik.

325 Oblique (N120°-N140°) transtensional blind and deep-seated faults, identified in the  
326 western part of the Sea of Marmara, are part of a northeast dipping extensional fault  
327 system named the South Transtensional Zone (STZ) by Grall et al. (2012). Many of  
328 these faults appear still active, although we do not know what portion of the total shear  
329 they accommodate. In particular, the western part of this fault system is suspected to  
330 play a rôle in the landslide activity in the Ganos area (in our western sector: Zitter et al.,  
331 2012). The long fault mentioned in the previous paragraph is the partly active eastward  
332 termination of this transtensional system providing a connection between the main  
333 northern branch of the North Anatolian Fault and the fault system to the south recently

334 named the Southern Marmara Fault by Le Pichon et al. (in press). ~N100° striking,  
335 mainly wrench faults occur also in the eastern part of the Sea. These faults, which may  
336 be blind, are probably a part of the Central Fault System (CFS, Grall et al., 2012). Other,  
337 shorter, faults close to and parallel with the inactive strand are, by contrast, still active.  
338 A small family of short faults with the same orientation to the east of the meandering  
339 submarine canyon seen in Fig. 10C are all active and clearly localise microearthquakes.  
340 By contrast, some abutting against the bounding faults of the spindle-shaped basin  
341 within the Central Basin appear dead, perhaps because they have been recently  
342 decapitated by the main northern branch of the North Anatolian Fault. All of the faults  
343 belonging to Grall et al.'s Central Fault System belong to the tension-gash-orientation  
344 (refer to Fig. 9 to see what that orientation means) normal faults related to the North  
345 Anatolian Shear Zone.

346 Curiously, the longer E-W orientated faults along Y-shear directions (cf. Fig. 9), some  
347 turning into Riedel and P-shear orientations at their ends, seem in some segments active  
348 and in others inactive.

349 Farther west, within the Central Basin, the microearthquake activity shows that the  
350 smaller spindle-shaped basin within the Central Basin is where most of the present  
351 activity is located, but the northeastern border of the larger basin is also active (Figs. 10 B  
352 and C). The active, spindle-shaped basin within the larger Central Basin is in fact where  
353 the Holocene subsidence rate is the highest in the Sea of Marmara (reaching more than 7

354 mm/a, Grall et al., 2012). This centralised subsidence is in stark contrast to the situation  
355 encountered in the Çınarcık and the Kumburgaz basins, where the main displacement  
356 zone is along the northern master faults of the basins.

357 So far, we had encountered no active faults in a shortening orientation. We encounter  
358 such a fault for the first time here, along the main displacement zone of the North  
359 Anatolian Fault in the Sea of Marmara just east of the Central Basin where the main  
360 strand is connected to the small spindle-shaped basinette in the middle of the Central  
361 Basin (Fig. 10B).

362 Within the southern Marmara Shelf, the available seismic profiles show that all the  
363 faults terminated their major movement at the end of the early Pliocene (3.5 Ma), but  
364 there is sparse, broadly spread microseismicity (Le Pichon et al., in press).

## 365 THE WESTERN SECTOR

366 The western sector (Fig. 13) houses what is termed the Western High and the Tekirdağ  
367 (ancient Rodosto) Basin. We selected two profiles to show here, one across the High  
368 (profile 70: Fig. 14A and B) and another, across the Basin (profile 78: fig. 16A and B).

369 Profile 70 (Fig. 14A) shows a much-faulted anticlinal ridge rising in the deepest part of a  
370 trough-shaped basin that is not seriously faulted at either end. Although what faults  
371 there are dip towards the basin, their vertical displacements are insignificant compared  
372 with the marginal faults of the Çınarcık, Kumburgaz and the Central basins. The basin

373 in fact has the appearance of a sag in the midst of which the peculiar anticlinal ridge of  
374 the Western High rises. It seems that the only active faults are those that bound the  
375 Western High, the southern with a higher rate of vertical motion as judged from the  
376 thickness of the green sequence. It is that fault which coincides with the main  
377 displacement zone of the northern branch of the North Anatolian Fault here. Some of  
378 the faults that cut the high are active, others dead and yet both dead and active faults  
379 seem to belong to the same set of steep thrusts. This is another region where we see  
380 active faults in a thrust orientation. In fact, the entire geometry of the Western High,  
381 with its flanking basins, gives the impression of a shortening structure that originally  
382 consisted of a broad synclinal basin, in the middle of which an anticline later rose. This  
383 is also supported by the topography of the sediment/acoustic basement interface,  
384 which reveals a deep trough which extends from the Central Basin towards the east and  
385 to the Tekirdağ Basin towards the West, such as Bayraçkılı et al. (in press) recently  
386 depicted. It seems that the throughgoing faulting chose the anticline to follow and in so  
387 doing sliced it with small and steep thrusts.

388 We have no data as to the nature of the sequences seen in the seismic profiles across the  
389 Western High making up its foundation. It is however very likely that they consist of  
390 the Ganos Flysch wedge that delimits the Thrace Basin to the south and the Mio-  
391 Pliocene gently-dipping sedimentary rocks covering the flysch wedge as one sees on the  
392 Gallipoli Peninsula, where even deeper portions of the accretionary wedge crop out

393 (e.g., Şentürk and Okay, 1984 report blueschists that crop out from under the flysch and  
394 the surrounding younger Cainozoic rocks)

395 The second profile, 78 (Fig. 15), goes across the Tekirdağ Basin. In principle it shows the  
396 same structures that one sees in the Western High, except the central anticline is here  
397 asymmetric and north-vergent, delimited against the northern, deeper half of the basin  
398 by a very steep, essentially vertical fault. Okay et al. (1999) interpreted it as a south-  
399 vergent thrust, which is contrary to what is seen along profile 78 (Fig. 15A). The  
400 southern part of the anticline is not broken by faults (except two very shallow landslide  
401 faults) but has the form of a syncline. The sequences betray a faster subsidence in the  
402 basin centre than along its sides consistent with the syncline interpretation.

403 The steep fault is a continuation westwards of the main displacement zone. The smaller  
404 large fault to the north achieves not much of a vertical displacement and seems a part of  
405 a series of short, *en-échélon* faults that seem an abortive attempt to reach the northern  
406 end of the Ganos (MG in Fig. 13A) constraining bend thrust zone (Dewey and Şengör,  
407 1979; Şengör, 1979; see Fig. 18 herein) by tearing a strike-slip fault from the northern  
408 basin north of the Western High across the northern part of the Tekirdağ Basin. That  
409 attempt seems to have been abandoned, however, while the youngest sequence was  
410 being deposited. There is essentially no microseismic activity along that trend of small  
411 faults.

412 Two other short trends of faults on the basin floor to the east of the middle part of the

413 Ganos constraining bend suggest other attempts at reaching it farther north than its  
414 present southernmost end. Although both attempts seem to have failed, the latter group  
415 of faults are still active as indicated by a lively microseismic activity at their eastern  
416 ends, but this is probably because of the loading of the crust in which they reside by the  
417 Ganos Wedge (see Fig. 16 for the cross-sectional geometry of the Ganos overthrust zone  
418 along the restraining bend).

419 At the latitude of the Western segment, seismic activity on the South Marmara Shelf is  
420 considerably livelier than in the Central Segment. Here we have in fact four earthquakes  
421 with magnitudes larger than 6 (Fig. 17) on the Biga Peninsula, the western continuation  
422 of the southernmost part of the South Marmara Shelf. The microearthquake activity is  
423 also much higher than it is in the Central Sector.

424 In this sector, the Marmara Island (the ancient Proconnesos: the largest island in Fig. 17)  
425 is a divider: to the east nearly all major faults strike W to NW; to its west nearly all  
426 strike NE, indicating that the shear zone takes a fairly sharp bend to the southwest with  
427 accompanying appearance of major thrust earthquakes along the fault segments. The  
428 faults also acquire here a much greater continuity than they have in the east, as one  
429 would expect from thrust faults as opposed to normal faults which are numerous east  
430 of the Marmara Island. Interestingly, west of the Marmara Island all the major strands  
431 are active, presumably because here the 2.5 cm/a motion of the Marmara Block  
432 delimited by the northern and the southern strands of the North Anatolian Fault (Le

433 Pichon et al. 2003) must be compensated in part by shortening and it is more economic  
434 to distribute the shortening and consequent crustal thickening on many faults.  
435 However, here too the strike-slip faults display their characteristic complex strains by  
436 localising earthquakes of all three kinds of fault types (Fig. 17)!

## 437 DISCUSSION

### 438 *Temporal evolution of the North Anatolian Fault in the Sea of Marmara*

439 When we started this study, we were expecting, as expressed in Şengör et al. (2005), an  
440 orderly evolution of a shear belt in which an initially broad shear zone characterised by  
441 such 'secondary' or 'auxiliary structures' as Riedel (R) and Anti-Riedel (R') shears, X,  
442 and P shears and tension gashes (T) with shortening structures (S) gradually evolving  
443 and eventually abandoning the field of deformation to a narrow thorough-going fault  
444 consisting of segments of Y shears, much as in the experiments of Riedel (1929),  
445 Tchalenko (1970), Wilcox et al. (1973), Bartlett et al. (1981), Naylor et al. (1986) and  
446 Ahlgren (2001) and as depicted in our Fig. 9. What we have found instead is a much  
447 more haphazard development. If one separates the fault families expected from the  
448 experiments into R, R', X, P and Y shears and T and S structures according to their  
449 orientations and, wherever available, from earthquake first-motion solutions and  
450 stratigraphic vertical separations and note their sequence of occurrence in the analogue  
451 experiments as R', R, P, X and, if the material being sheared is suitable to form them,  
452 first S and then T structures, Nature shows us that in the Sea of Marmara almost all of



453 these structure types are active now. We have actually seen no X shears that can be  
454 confidently identified and some of the fault segments bending into a P-shear orientation  
455 are all now dead. What is even more peculiar is the observation that some faults may  
456 have both dead and active segments along them (within the time frame of a part of the  
457 Pleistocene, mostly even late Pleistocene-Holocene)! An extreme example has been  
458 recently documented by Le Pichon et al. (in press), who showed that what they called a  
459 South Marmara Fault under the late Pliocene beds in the South Marmara Shelf appears  
460 entirely inactive there at least for the last 2.5 or 2.6 Ma, yet both at its eastern and  
461 western ends it is today active! Kuşçu et al. (2009) showed that the faults in the Gulf of  
462 Gemlik are active and the prolongation of the South Marmara Fault into the Aegean in  
463 the west just localised a Mw 6.2 dextral slip earthquake on 8<sup>th</sup> January 2013 (Le Pichon  
464 et al., in press)! This suggests that some dead fault strands may be connected to active  
465 ones by as yet unrecognised intermediate segments.

466 An extremely interesting observation has been reported recently by Shillington et al.  
467 (2012): they showed the presence of folds forming by downslope creep on steep (3° to  
468 10°) slopes in and around the main trough of the Sea of Marmara. Interestingly, with the  
469 exception of the slope north of the Armutlu Peninsula and to the south of the Central  
470 Basin, all of the slopes on which such creep folds have formed and in places continue to  
471 form are on the slopes of structures that are in a shortening orientation and have been  
472 so since the shear began. The Çınarcık exception is easy to explain in terms of the active

473 normal faulting going on at the foot of the Armutlu Peninsula. The slope south of the  
474 Central Basin may have been accentuated by the faults that there exist along the South  
475 Marmara Shelf (see Le Pichon et al., in press), but the data there are not adequate to  
476 address this question. Shillington et al.'s (2012) work is a very nice possible example of  
477 structures forming concurrently with growing folds on a large scale, perhaps not  
478 dissimilar to the situation as documented by the progressive growth in time (late  
479 Cretaceous to Eocene) of the Hathira (Makhtesh-Hagadol or Kurnub) anticline in  
480 southern Israel (Eyal, 2011).

481 Although most of the elements of the shear zone in the Sea of Marmara are active now,  
482 the strain they accomplish is extremely variable. It seems that almost all the motion of  
483 the Eurasia/Marmara Block (the area delimited by the northern and the southern  
484 strands of the North Anatolian Fault: Le Pichon et al., 2003) is taken up by the Y-shears  
485 making up what Le Pichon et al. (2001) and Şengör et al. (2005) termed the main  
486 displacement zone forming, from east to west, the main northern bounding faults of the  
487 Çınarcık and the Kumburgaz basins, the fault cutting through the Central Basin and  
488 bounding the small spindle-shaped basin in the middle, and the fault delimiting the  
489 Western High and the Tekirdağ Basin to their south. All other active faults seem to cut  
490 to the sea-floor and nucleate microseismic shocks with few exceptions, such as the  
491 normal faults south of the Çınarcık Basin that nucleated an  $M_s=6.4$  earthquake and  
492 possibly another one in 1894 that greatly damaged İstanbul. They contribute very little

493 to the overall strain, yet Le Pichon et al. (2003) documented that the small strain was by  
494 no means negligible and is detectable by GPS observations. However, Sorlien et al.'s  
495 (2012) thesis that the faults have moved with a constant rate during the last half a  
496 million years, which implies in view of the  $\sim 55$  km total offset for the entire Marmara  
497 area (Akbayram and Okay, 2012; Şengör et al., 2005) that the Sea of Marmara has a total  
498 history of less than half a million years, which is certainly incorrect.

499 The difference between the experiments and Nature as it appears to us in the  
500 neotectonics of the Marmara region may be twofold: one is the time scale. What  
501 happens in nature in thousands to hundreds of thousands and even a couple of million  
502 years, happens in experiments in a matter of minutes or at most hours. We may be  
503 seeing a process of slow demise of the auxiliaries in Nature, which to us seem a time of  
504 activity in concert. In reality, the auxiliaries may be slowly passing away as the main  
505 displacement zone gathers the action onto itself. The fact that nearly all the main  
506 displacement today is concentrated on the almost completely throughgoing main  
507 displacement zone may be a result of its gradual climb to prominence, a climb that has  
508 not yet reached the summit.

509 The second difference pertains to the extreme heterogeneity of the natural material  
510 which the shear zone deforms as opposed to the nearly homogeneous materials used in  
511 the experiments. No experiment can hope to reproduce the multifarious natures of the  
512 materials making up the area in and around the Sea of Marmara. Entire domains

513 consisting of very different types of materials dominate the Marmara area and the  
514 developing shear zone no doubt exploited every bimaterial surface to break through on  
515 all sorts of scales. That is mainly why we have been unable to use with confidence the  
516 kind of modelling represented by Hergert and Heidbach's (2011) paper, where rock  
517 properties in the sedimentary carapace are accepted to be distributed homogeneously –  
518 which is manifestly not the case, if for no other reason than the depth differences among  
519 basin basements, even within a single basin—and the great variations in the rock  
520 properties in the basement are not even considered. Şengör et al. (2005) pointed out that  
521 the North Anatolian Fault as a whole seems to have nucleated along the northern 'wall'  
522 of the Tethyside accretionary complexes against the more indurated basement of the  
523 Rhodope-Pontide Fragment (see Fig. 2), very much corroborating the theoretical model  
524 of Ben Zion and Andrews (1998) and Ben Zion (2001). There is little doubt that the  
525 sensitivity of bimaterial surfaces to preferentially produce rupture under stress is also  
526 present at much smaller scales between the tectonic units made up of different materials  
527 under and around the Sea of Marmara and between different rock packages within  
528 those tectonic units.

### 529 *The origin of the Marmara Deeps*

530 Both the Western and the Central Highs have east-northeasterly trends. Fig. 14A  
531 shows the Western High having an overall anticlinal structure albeit sliced by  
532 numerous faults. We had originally hypothesised that the anticline was a function of

533 the shear evolution and asked ourselves that if we reverse the shear along the North  
534 Anatolian Shear Zone, what orientation the high would acquire. Our expectation was  
535 that it would rotate into a more northeasterly direction and we expected a similar result  
536 from the Central High. Fig. 18A shows the present day fault geometry within the Sea of  
537 Marmara and the outlines of the two highs in blue. Fig. 18B shows the situation if a  
538 homogeneous strain is imposed on the system to reverse a 55 km offset. Because the  
539 overall offset along the North Anatolian Shear Zone is 75 to 80 km along its entire length  
540 (Şengör et al., 2005), we subtracted the offset of some 26 km seen along the Pamukova  
541 strand leading to the main southern strand (Şengör et al., 2005) from that to obtain the  
542 55 km left-lateral offset on the entire Sea of Marmara fault family to reverse the present  
543 right lateral shear for the purposes of Fig. 18. This seems consistent with the 52 km  
544 offset reported by Akbayram and Okay (2012) from the northern strand east of the Sea  
545 of Marmara. The timing of the offset reported by Akbayram and Okay (2012) is  
546 problematic, however, because some of it could have formed just after the medial  
547 Eocene collision along the Intra-Pontide suture in a completely different system of  
548 strike-slip faults (Akbayram et al., 2013). It is clear, however, that at least the Marmara  
549 Fault family north of the South Marmara Shelf takes up a displacement that farther east  
550 is concentrated on the northern strand of the North Anatolian Fault and that offset is an  
551 absolute maximum of 57 km. Fig. 18C shows what is seen when we superimpose Figs.  
552 18A and 18B keeping the northern boundary fixed. *The highs do not rotate as we expected,*  
553 *but acquire a lumpier appearance!* In fact, as a result, they present an optical illusion as if

554 they rotated into a more E-W orientation when the displacement is reversed.

555 This to us suggests that the highs, or their much more subdued forerunners, were  
556 probably there before the shear began and were similar to gentle axial culminations  
557 along outer arc highs in present-day forearcs, in this case possibly either accentuated or  
558 even initially caused by the collision. When the shear began they were shortened into S  
559 structures and became elongated in an ENE orientation. The Western High was  
560 probably only a gentle 'suggestion' with a broad basin, a simple 'undulation,' of the  
561 basin floor, which later became accentuated as the shear evolved. The fig. 6 in Sorlien et  
562 al. (2012) very nicely shows the anticlinal structure of the Central High and the smaller  
563 wavelength folds right across it involving even the youngest sediments, although there  
564 is a clear decrease in folding amplitude up section (the reverse of what one sees in the  
565 so-called 'generative folds'; our sorts of folds may thus be called 'degenerative')  
566 indicating a waning of the shortening commensurate with the concentration of the  
567 shear along the northern branch of the North Anatolian Fault.

568 The Çınarcık Basin is clearly not a part of this system as indeed it does not sit in front of  
569 the Thracian fore-arc basin, but in front of the more solid and homogeneous İstanbul  
570 Zone (Şengör 2011, fig. 1) and in a totally different orientation, *viz* WNW. We think that  
571 the interpretation of it by Uçarkuş (2010) as a fault wedge basin *sensu* Crowell (1974) is  
572 so far the best explanation for the origin and the present structure of the basin, so we do  
573 not further elaborate on it here, except to say that it is a completely different sort of

574 basin from the other, more westerly, basins of the Sea of Marmara.

575 *The offset along the northern branch of the North Anatolian Fault*

576 Fig. 18 C further shows that the statement by Le Pichon et al. (2001) that the offset along  
577 the very young northern branch of the North Anatolian Fault, at least west of the  
578 Çınarcık Basin, is only some 4 km, falls out of the distributed shear interpretation. If the  
579 shear has been really as distributed as the Fig. 18 approximates, then the offset along  
580 the level of the Central Basin/Western High comes to at most 5 km. and no more.  
581 Therefore, the fault families we have mapped in the Sea of Marmara must have initially  
582 taken up the entire displacement that falls to the share of the present-day Northern  
583 Branch altogether. That many of them are still active (although probably dying)  
584 corroborates this view. That the South Marmara Shelf fault families are almost all dead  
585 (with the exception of the present microseismic activity), but that more activity is seen in  
586 the northern auxiliary structures, on both seismic and stratigraphic (Sorlien et al., 2012)  
587 evidence, may be an indication of the fact that the displacement is gradually gathering  
588 itself to the northern branch alone.

## 589 CONCLUSIONS

590 The present day fault families younger than about the medial Miocene in the Sea of  
591 Marmara are all parts of the North Anatolian Shear Zone and they all seem to have  
592 shared among themselves a total offset of some 55 km along this zone since the later

593 medial Miocene. Contrary to expectation from analogue claycake and sandbox  
594 experiments, a regular development from distributed structures to a single,  
595 throughgoing strand cannot now be established, but this may be more a function of the  
596 vastly different time-scales between the experiments and Nature herself than a  
597 fundamental difference in the evolutionary sequence. That the Southern Shelf fault  
598 family is now dead and that no major earthquake seems to be nucleated on the  
599 auxiliaries (except in the southern part of the Çınarcık Basin) may indicate a gradual  
600 localisation of the offset along the single strand of the Northern Branch of the North  
601 Anatolian Fault.

602 This study gives support to the older ideas that intracontinental transform faults begin  
603 their lives as broad shear zones and become gradually converted into single- (or at most  
604 a few-) strand structures. What such a broad shear zone at depth looks like may be  
605 displayed by the 15-km wide mylonite zone seen along the giant Irtysh shear zone  
606 (displacement  $\approx 2000$  km) in Central Asia (Şengör and Natal'in, 1996). The numerous  
607 active fault zones we see at the surface no doubt broaden with depth (earthquake  
608 hypocentres are hardly deeper than some 10 km under the Sea of Marmara; below that  
609 the deformation is most likely ductile) and given their density at the surface, they most  
610 likely coalesce into a very broad ductile shear zone. Le Pichon et al.'s (2003) study has  
611 corroborated the presence of such a general, distributed deformation south of the  
612 Northern Branch of the North Anatolian Fault.



613 Fig. 19 shows a situation where the faults depicted on Fig. 18A have been widened to  
614 shear zones of some 4 km width, the presumed with of the San Andreas Fault below 10  
615 km (e.g., McBride and Brown, 1986). In these regions of amphibolite grade  
616 metamorphism what a shear zone looks like is shown in the inset from the Diancan  
617 Shang late Cainozoic metamorphic shear zone Ailao-Shan along the Red River strike-  
618 slip fault in Southeast Asia (copied from Cao et al., 2011). We have flipped Cao et al.'s  
619 figure to make the shear right-lateral and the similarity of the hand sample to the map  
620 pattern displayed by the presumed depth-map of the Sea of Marmara fault family is  
621 staggering: it is one of the best examples of Pumpelly's rule that small structures mimic  
622 big structures and a further demonstration of Tchalenko's (1970) point that shear  
623 structures are independent of scale. . We thus think that at a depth below 10 km, the Sea  
624 of Marmara basement consists of sheared rocks of the kind seen below the Ailao Shan  
625 strike-slip fault, which, by an interesting coincidence, is also nucleated on a suture zone  
626 bearing steep serpentinite screens (Şengör and Natal'in, 1996; Burchfiel and Chen, 2012,  
627 chapret 15, especially fig. 15-1). In deeply-eroded Precambrian terrains, the presence of  
628 such broad, vertical, anastomosing ductile shear zones are most likely expressions of  
629 large strike-slip faults at the surface, many of which were probably transform faults. As  
630 transform faults are dependable indicators of relative plate motion directions, this is one  
631 way to get a handle on such old plate motion directions.

632 ACKNOWLEDGEMENTS

633 This paper was written upon the invitation of Professor Ali Polat to whom we are  
634 grateful for the honour of being included in a volume in memory of Tuzo Wilson.  
635 Professor Namık Çağatay was a great source of knowledge and critique in all matters  
636 stratigraphic. MARSITE funding made substantial parts of this study possible. All  
637 marine work in the Sea of Marmara since 2000 forming the basis of this study was  
638 undertaken with the active and cheerful support of the Turkish Navy. Detailed and  
639 constructive reviews by Professors Kevin Burke and Mark Zoback substantially  
640 improved the original presentation.

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914

#### 915 FIGURE CAPTIONS

916 Fig. 1. Wilson's illustration of the kinds of transform faults (Wilson, 1965, fig. 3). For  
917 'arc' read 'trench' to conform to present common usage: **A**) a. ridge-ridge, b. ridge-arc  
918 facing away from the ridge, c. ridge-arc facing toward the ridge (this is, in principle,  
919 what the North Anatolian Fault is), d. arc-arc with arcs facing away from one another, e.  
920 arc-arc with arcs facing the same way, f. arc-arc with arcs facing toward each other, **B**)  
921 d. transform point at which extension changes to shortening.

922 Fig. 2. The North Anatolian Fault, forming in reality a family of faults that constitute the

923 North Anatolian Keirogen (Şengör et al., 2005). Not all the faults shown on this map are  
924 now active, but all have been active sometime in the last 11 Ma. Most are potential  
925 earthquake generators. Fault traces delineated by heavier lines represent the most active  
926 parts of the keirogen constituting the strands collectively known as the North Anatolian  
927 Fault (NAF). Note that the keirogen is entirely confined to the area underlain by  
928 Tethyside accretionary complexes shown in gray (Slightly modified from Şengör et al.  
929 (2005, Fig. 2)).

930 Fig. 3. The North Anatolian Shear Zone (NASZ: delimited by discontinuous lines) and  
931 the courses of the major rivers traversing it. Key to abbreviations: from east to west  
932 (Black letters): E=Elmalı/Peri (tributary of the Murat before the construction of the  
933 Keban Dam), Ka=Karasu (Elmalı/Peri+Karasu=Fırat {Euphrates} without Murat  
934 {outside this map}, Y=Yeşilirmak, K=Kızılırmak, D=Delice, F=Filyos  
935 (/Yenice/Araç/Soğanlı {formerly Uluçay}/Gerede Suyu), S=Sakarya, Su=Susurluk.  
936 Grey letters in outline show locations of some cities and tectonic features: A=Ankara,  
937 B=Bursa, b=Bolu, E=Erzincan, İ=İstanbul, İ=İzmit (lake) K=Karlıova, OF=Ovacık Fault,  
938 SF=Sungurlu Fault. Notice that significant abrupt deflections of river courses are  
939 confined to the area of the NASZ. Following Şengör et al. (2005), the faults shown to be  
940 parts of the NASZ in the southern part of the Tokat Lobe are here left out of the NASZ  
941 because of their as yet uncertain relationship to the NASZ and to the geometry of the  
942 major river courses (Slightly modified from Şengör et al. (2005, Fig. 6)).

943 Fig. 4. Historical earthquake epicentres in and around the Sea of Marmara compiled by  
944 the ETH Geodesy and Geodynamics Laboratory (GGL), Zurich. Most of these  
945 earthquakes presumably occurred on the faults shown in Fig. 5.

946 Fig. 5. A. The locations of the multichannel seismic profiles used in this study.  
947 CB=Central basin, CiB= Cınarcık Basin, ET=Erdek Tombolo, GG= Gulf of Gemlik,  
948 KB=Kumburgaz Basin, MI=Marmara Island, TB=Tekirdağ Basin. B. The faults mapped  
949 in this study. The black segments are active and the grey segments are **presumably**  
950 inactive.

951 Fig. 6. Maps of the eastern sector of our study area: A. Location map of the profiles  
952 selected to characterise the faults in this sector. B. Distribution of seismicity during the  
953 2005-2011 interval in the eastern sector. C. Distribution of seismicity compared with the  
954 identified faults belonging to the North Anatolian Shear Zone. **Refer to the Fig. 5 for the**  
955 **legend of the faults.**

956 Fig. 7A. Line 9 showing a cross section across the Çınarcık Basin including its broad  
957 southern shelf that contains a subsidiary basin on a southerly-tilted normal fault block.  
958 B. Our interpretation of the structure of the Çınarcık Basin along Line 9 (vertical  
959 exaggeration here is 4 times). In this profile, three onlap surfaces (pinch-outs , indicated  
960 by black arrows) are observed at both basin edges as well as on the 2-km-wide faulted  
961 anticline rising within the basin. These pinch-outs are thus most likely related to  
962 variations in sedimentation rates rather than to changes in fault activity. The onlap

963 surfaces are assigned to the Red, the Blue and the Yellow sequences boundaries (130,  
964 250 and 330 ka respectively). The thickness of 100 ka stratigraphic sequences in this case  
965 ranges between 400 ms twt and 150 ms twt (corresponding roughly to 300 m and 110  
966 m), which is consistent the mean sedimentation rates in this area (Seeber et al., 2006); as  
967 well as the sequences jump correlated by Sorlien et al., (2012, fig. 8). C. Same as Fig. 8B,  
968 but with no vertical exaggeration to show the 'real' geological structure. We have done  
969 this only for this profile to give an idea about the difference between the vertically  
970 exaggerated and non-exaggerated sections for those not used to looking at seismic  
971 profiles.

972 Fig. 8A. Line 13 showing another cross section farther west than line 9 across the  
973 Çınarcık basin showing a subbasin. B. Our interpretation of the structure of the Çınarcık  
974 Basin along Line 13 (vertical exaggeration here is 4 times) the sequences appear less  
975 obvious (exception for the pinch-out on the Red Horizon) but horizons have been  
976 correlated to the line 9, and this correlation looks in agreement with the interpretation  
977 of Sorlien et al. (2012, fig. 8). To the south, across the Imrali Basin, a thick coherent  
978 section of sedimentary horizons (reaching more than 3 s twt) is imaged. As in the  
979 Çınarcık Basin, pinch-outs can be recognized but the most remarkable feature is  
980 probably the toplap geometry (indicated by white arrow) at the base of the Blue (250 ka)  
981 horizon (also drawn on the fig. 4 in Sorlien et al., 2012). We assigned the same sequence  
982 boundary as Sorlien et al. (2012) in this area, and we note that onlap geometry on the

983 sequences boundaries, are also observed in this basin.

984 Fig. 9. Development of a shear zone (from Şengör, in press). Brown areas are structures  
985 associated with extension and/or transtension. In transtension, all extensional structures  
986 become 'flatter' and all shortening structures become 'steeper' with respect to the Y-shear.  
987 However, this is only an instantaneous picture, because as the shear zone evolves, all  
988 structures will rotate and strains will increase as we now see in the Sea of Marmara case. Key  
989 to lettering: F is fold axial trace, P is P-shear, R is a Riedel shear, R' is an anti-Riedel shear, T  
990 is a tension gash (in meso and macro examples, depending on the complexity of the system,  
991 the stress along the 'tension gash orientation' line may not necessarily be purely tensional,  
992 but there must be a major component of extension), X is an X shear.  $\phi$  is the angle of internal  
993 friction of the material being sheared.

994 Fig. 10. Maps of the central sector of our study area: A. Location map of the profiles  
995 selected to characterise the faults in this sector. B. Distribution of seismicity during the  
996 2005-2011 interval in the central sector. C. Distribution of seismicity compared with the  
997 identified faults belonging to the North Anatolian Shear Zone.

998 Fig. 11A. Line 23 showing a cross-section across the eastern moiety of the Kumburgaz  
999 Basin showing the well-localised basin and its temporal enlargement outward. For  
1000 details of the seismic stratigraphy, see the caption of Fig. 10B. B. Our interpretation of  
1001 the geology along line 23.

1002 Fig. 12A. Line 27 showing another cross section farther west than line 23 across the  
1003 western moiety of the Kumburgaz Basin showing a broad, very young, bovine-head  
1004 basin that is nested on the older, narrower, fault-bounded basin. B. Our interpretation  
1005 of the geology along line 27. On both profiles, 23 and 27 pinch-outs of horizons on the  
1006 Red (130 ka) and the Blue (250 ka), as well as on the Violet (430 ka) horizons are  
1007 indicated by black arrows. The depth of these horizons along the Northern Main Fault  
1008 are similar to the one proposed by Sorlien et al., 2012 (figs. 6 and 11), and implies a  
1009 mean sedimentation rate of around 1-1.5 mm/a over 100 ka period. The well-localised  
1010 basin below the blue Horizon, which was widened slowly outward through time, The  
1011 blue sequence (250 ka) time, by contrast, looks like the period when the basin widening  
1012 set on drastically. This change is concomitant to the deactivation/slowdown of the  
1013 faults running across the Kumburgaz Basin. Part of these faults are presumably still  
1014 active as some en-échelon discrete structures can be recognized along them on the  
1015 seafloor (en-échelon structures mentioned in Fig. 11).

1016 Fig. 13. Maps of the western sector of our study area: A. Location map of the profiles  
1017 selected to characterise the faults in this sector. B. Distribution of seismicity during the  
1018 2005-2011 interval in the western sector. C. Distribution of seismicity compared with the  
1019 identified faults belonging to the North Anatolian Shear Zone.

1020 Fig. 14A. Line 70 showing a cross section across the Western High showing its faulted  
1021 anticline structure and the saucer-shaped basin in which the anticline seems to have

1022 risen along its bounding faults. B. Our interpretation of the geology along line 70. This  
1023 basin displays a thick coherent sedimentary section (which can reach 2 s twt of  
1024 thickness), suggesting a long history of sedimentation. Along the basin margins one  
1025 sees the very same seismic facies as one does within the lall perched basins on top of the  
1026 High (to be published in Grall et al., in press, b). In particular one recognises the chaotic  
1027 reflectors intrepreted as slide deposits on the violet horizon (430 ka) and also the bed  
1028 with low amplitude reflectors above the green horizon (which is widespread in the  
1029 entire Sea of Marmara). These observations show that the Western High rose in a basin  
1030 that was simply an eastern continuation of the Tekirdağ Basin! The blue patchy reflector  
1031 (250 ka) and the Red horizon in reverse polarity (130 ka) have been also recognized  
1032 above these characteristic layers. This profile confirms what Grall et al. (in press a)  
1033 pointed out regarding the Mass Transport Deposits on the Violet Horizon which is  
1034 likely related to a regional (at least at the scale of the Western High) process (such as the  
1035 paleo-environmental changes proposed in Grall et al., in press a).

1036 Fig. 15A. Line 78 showing another cross section farther west than line 70 across the  
1037 Tekirdağ Basin showing the central anticline covered with young sedimentary layers. B.  
1038 Our interpretation of the geology along line 78. Note the fault that bounds the anticline  
1039 to the north. It now seems vertical, but the internal structure of the anticline implies  
1040 that it might have originated as a very-steeply south-dipping thrust fault, contrary to  
1041 the interpretation by Okay et at. (1999). 1.5 s of coherent sedimentary section is imaged

1042 across the Tekirdağ Basin, on which onlap sequences (indicated by black arrows) occur  
1043 periodically, as in the Central Basin (Grall et al., 2012), and in the Çınarcık (Fig. 8A). The  
1044 first Red onlap surface (dated at ~130 ka) is at around 300 ms twt (roughly 230 m)  
1045 below the seafloor within the depocentre. This age estimation is thus consistent with the  
1046 mean sedimentation rates (1.5-2mm/a) derived from the last marine/lacustrine  
1047 transition (Mercier de Lepinay, pers. comm). The transparent lens in green (430-540 ka  
1048 interval) in the basin as well as in the bounded anticline, looks very similar to the one  
1049 encountered on the Western High and elsewhere in the Sea of Marmara (Grall et al., in  
1050 press b). The Red and the Blue horizons have been correlated from the basin to the  
1051 southern anticline across the faults. The Red Horizon is tentatively proposed on the  
1052 base of sediments affected by gravitay-gliding, as it has been shown that main slides  
1053 occur preferentially during glacial/interglacial transitions (Grall et al., in press a).  
1054 Nonetheless, we cannot exclude that the prominent reflector on the base of this sliding  
1055 mass, which apparently functions as a *décollement* could have no stratigraphic meaning.  
1056 Besides, it is also probably interesting to note that across this basin as across other deep  
1057 basins, no mass transports deposits have yet been recognized on top of the violet  
1058 horizon, despite the fact that they have been recognized in the ponded basins of the  
1059 Western High (Grall et al., in press b), as well as on the margins of the basin within  
1060 which the Western High seems to rise (Fig. 13B).

1061 Fig. 16. Profile 84 across the Mt. Ganos shelf and slope showing the underthrusting of



1062 the Tekirdağ Basin floor under the rising pillar of Mr. Ganos and the development of a  
1063 small marginal fold-thrust belt. This underthrusting activity is probably the reason of  
1064 the microseismicity seen along the small faults at high angles to the overthrust margin  
1065 breaking up the basin floor (Fig. 12C).

1066 Fig. 17. The locations, magnitudes and the first-motion solutions of the major shocks  
1067 known from the southern part of the western sector studied in this paper. Fault plane  
1068 solutions are from McKenzie (1972), Ekström and England (1989), and Taymaz et al.  
1069 (1991).

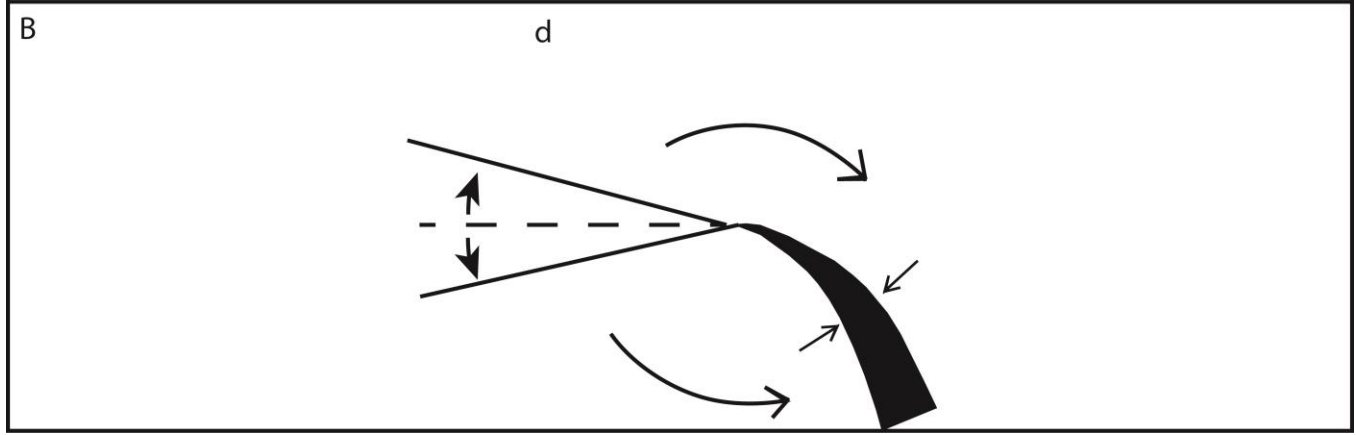
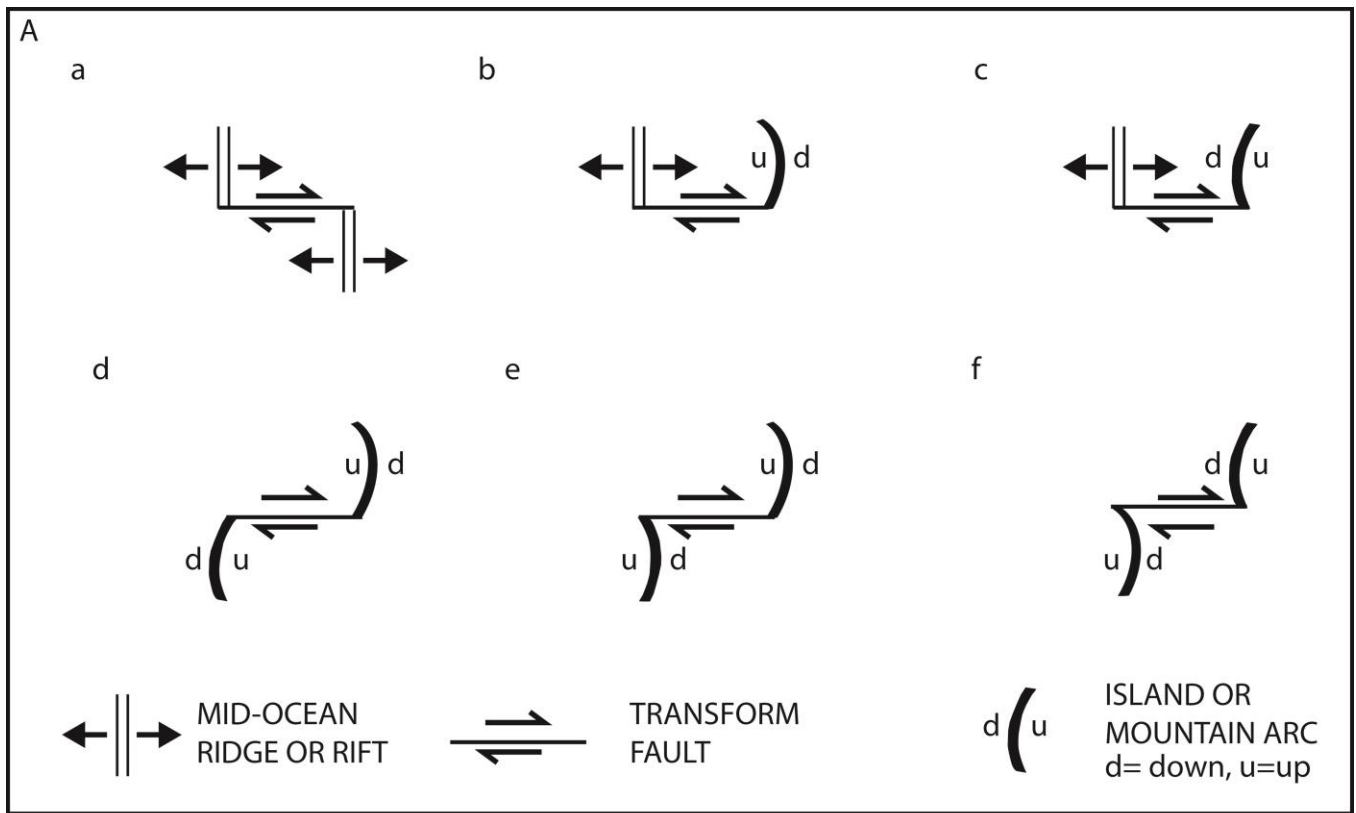
1070 Fig. 18A. A simplified map of the faults reported in this paper. The two light blue  
1071 patches are the Central and the Western Highs. B. Same as A, but with a left-lateral  
1072 shear imposed on all the elements to reverse the right-lateral shear of some 55 km along  
1073 the Sea of Marmara. C. A comparison of the present geometry of the faults belonging to  
1074 the northern part of the north Anatolian Shear Zone in the Sea of Marmara with their  
1075 presumed geometry before a homogeneous right-lateral shear of some 55 km took  
1076 place. Note how little many of the elements rotate. No wonder so many of the R shears  
1077 are so long-lived. Note also that the Central and the Western Highs are compressed into  
1078 a more pronounced NE-SW orientation as the shear evolves. The compression appears  
1079 as a rotation of the entire structure, which, in reality, it is not.

1080 Fig. 19. The presumed geometry of the fault network depicted in Fig. 18A at a depth of  
1081 about 10 km below the floor of the Sea of Marmara. All the faults are assumed to have

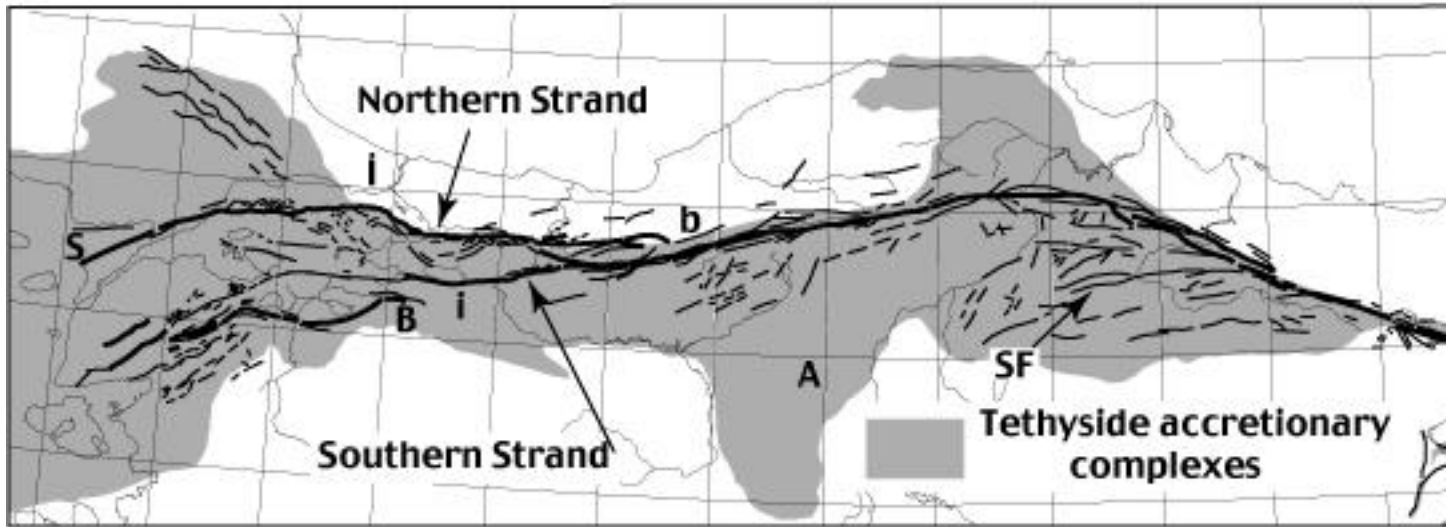
1082 widened into shear zones with a uniform width of 4 km. This is probably wrong, as not  
1083 all faults have the same displacement, but it gives an idea as to what the North  
1084 Anatolian Fault family under the Sea of Marmara may look like close to mid-crustal  
1085 depths. The inset marked 'b' is a hand specimen showing the mesostructure of the Ailao  
1086 shear zone at amphibolite facies metamorphic conditions. Its similarity to our presumed  
1087 North Anatolian fault structure at depth is remarkable.

1088

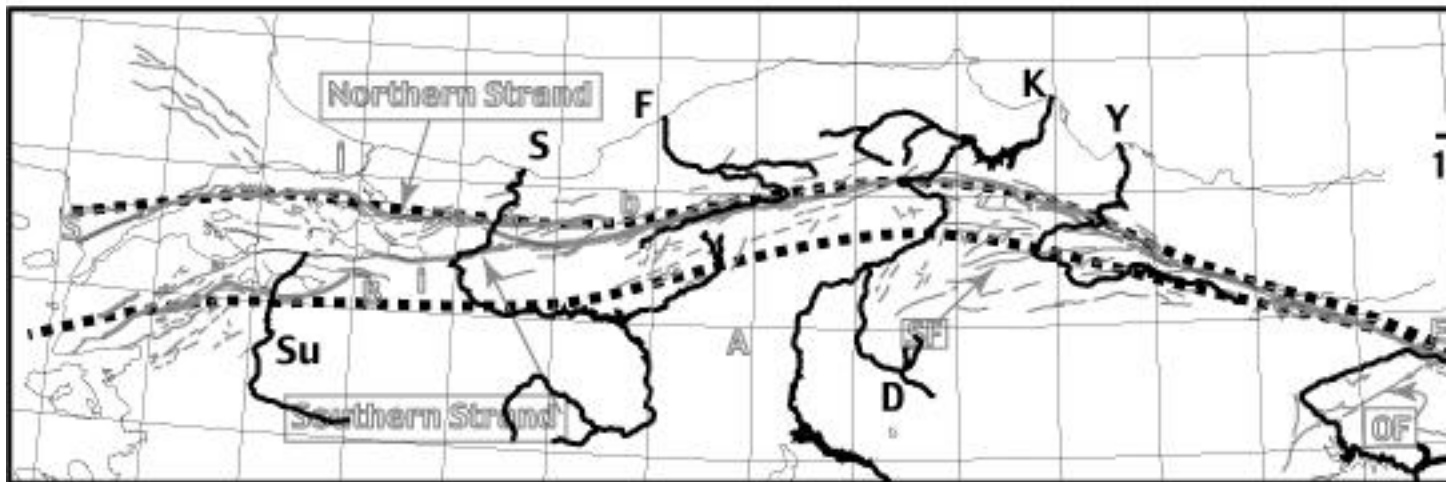
1089 FIGURES

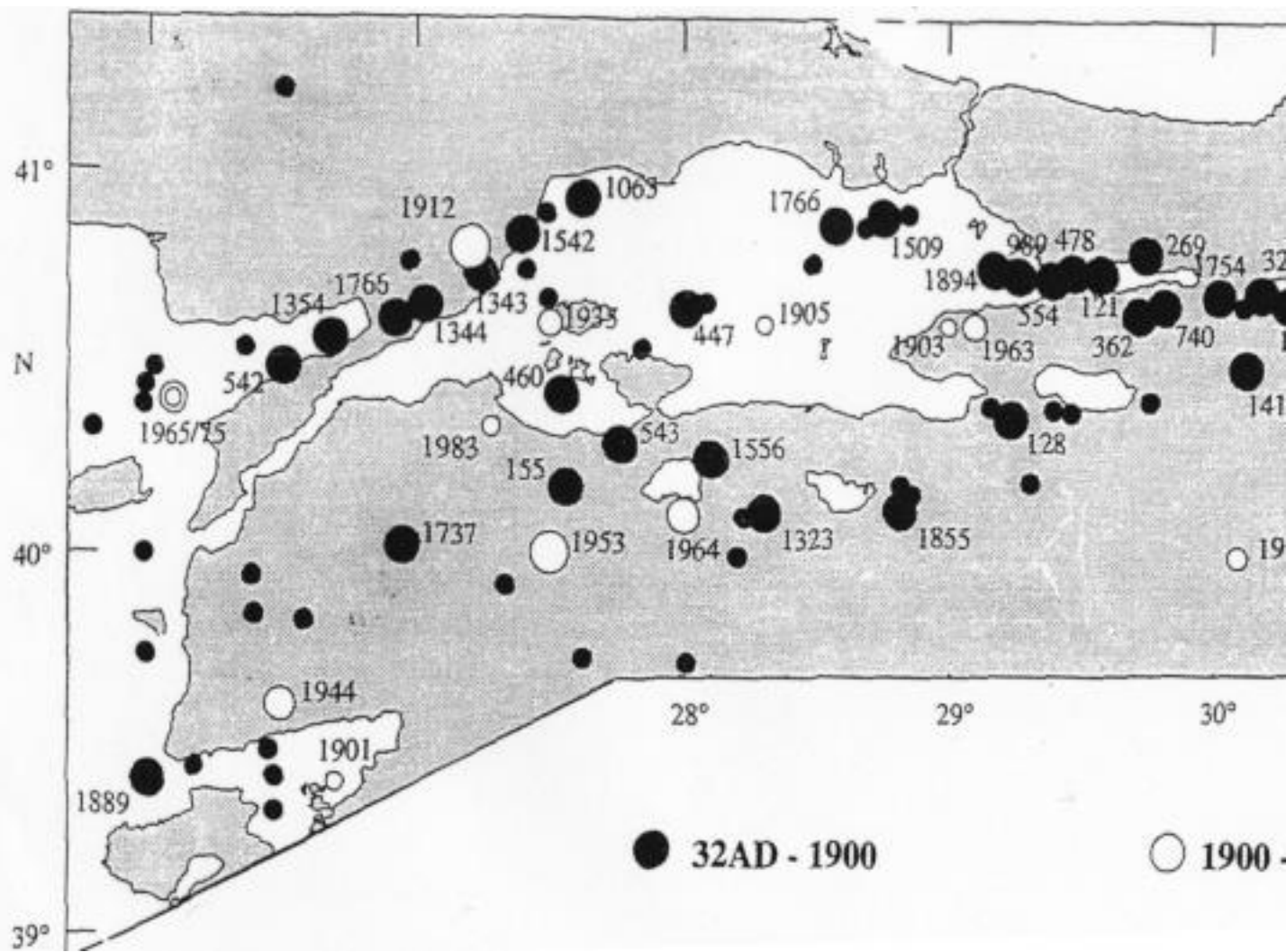


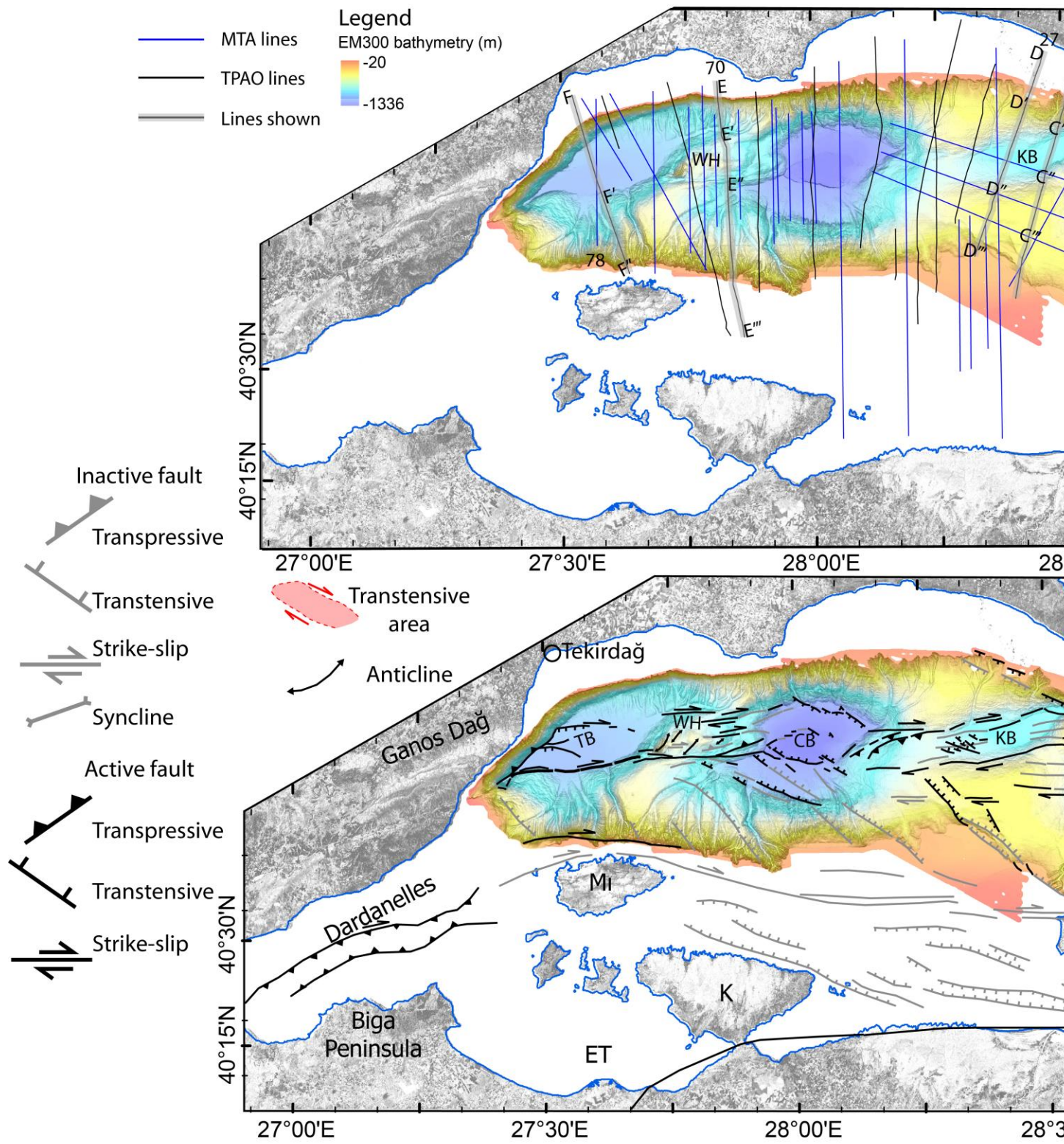
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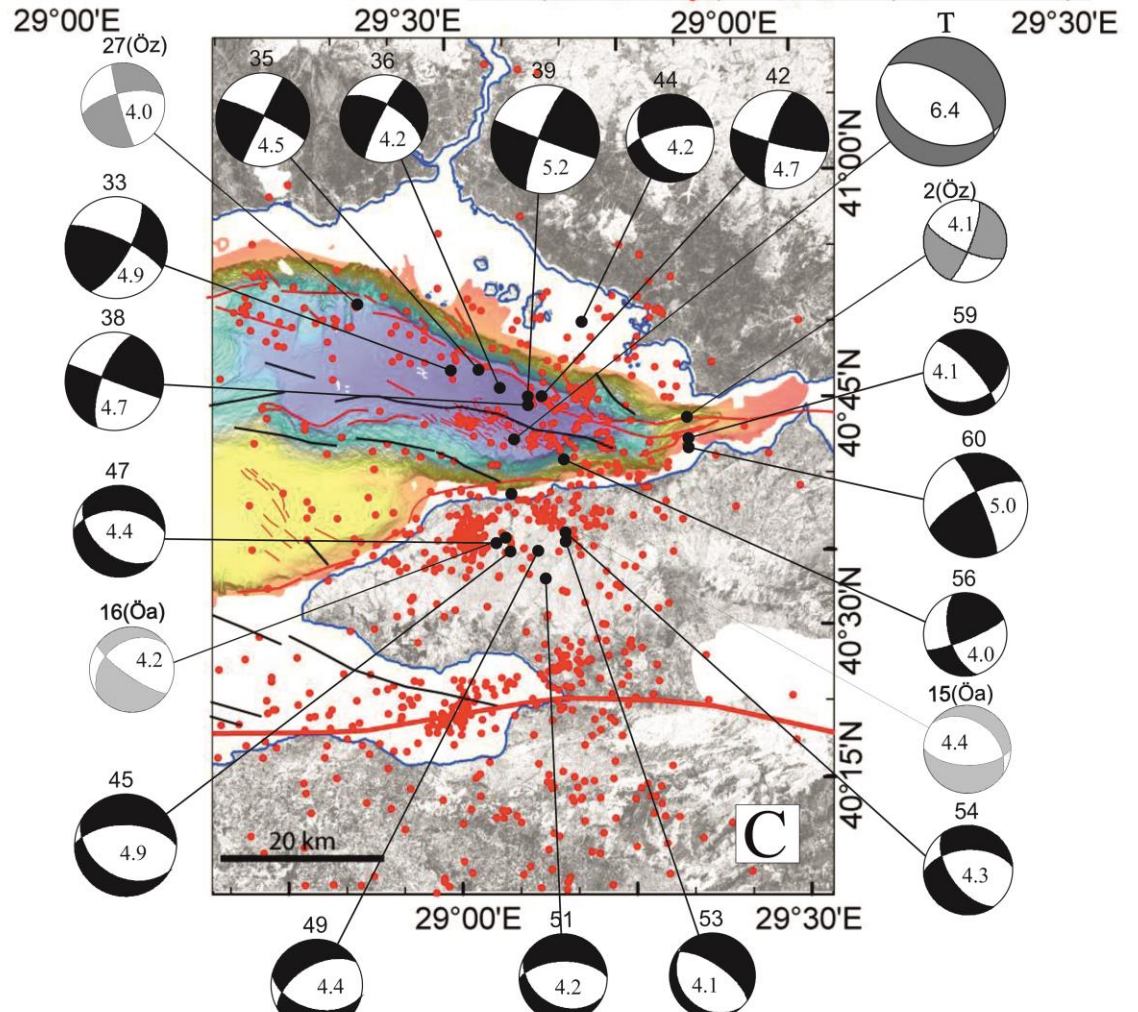
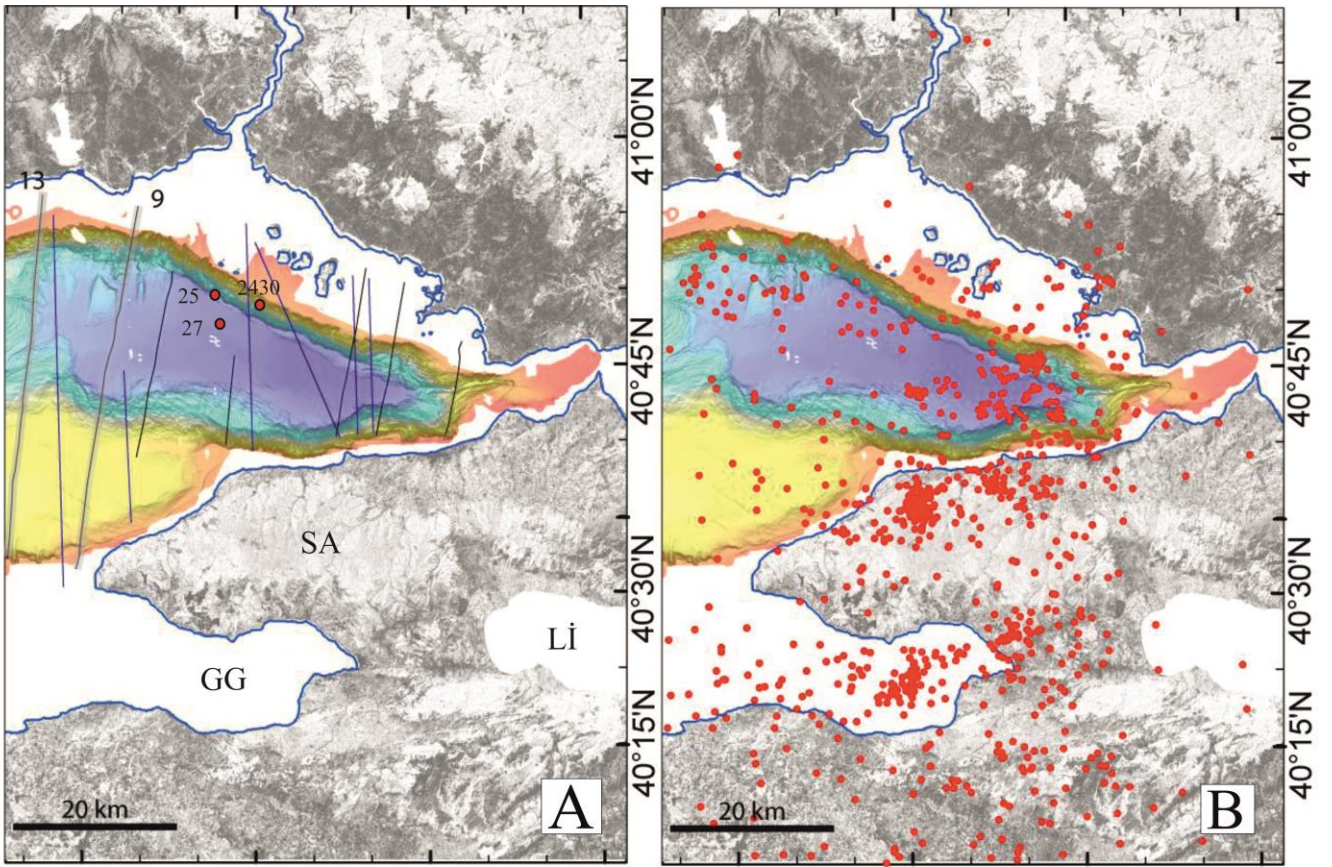


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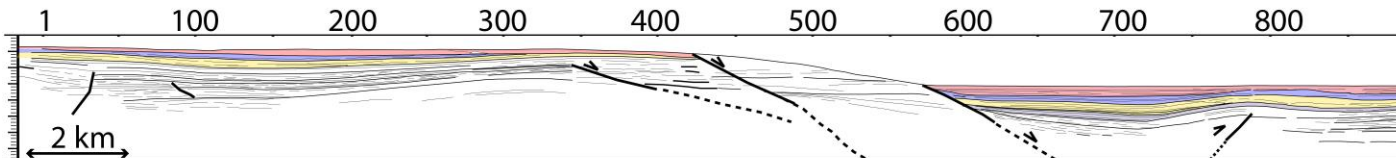
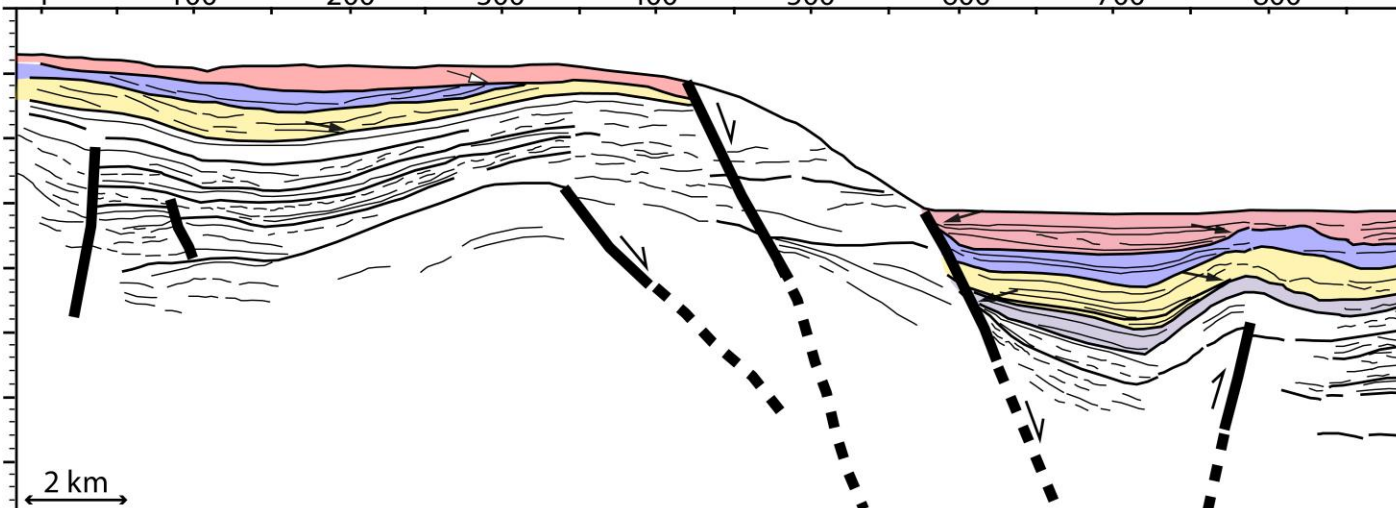
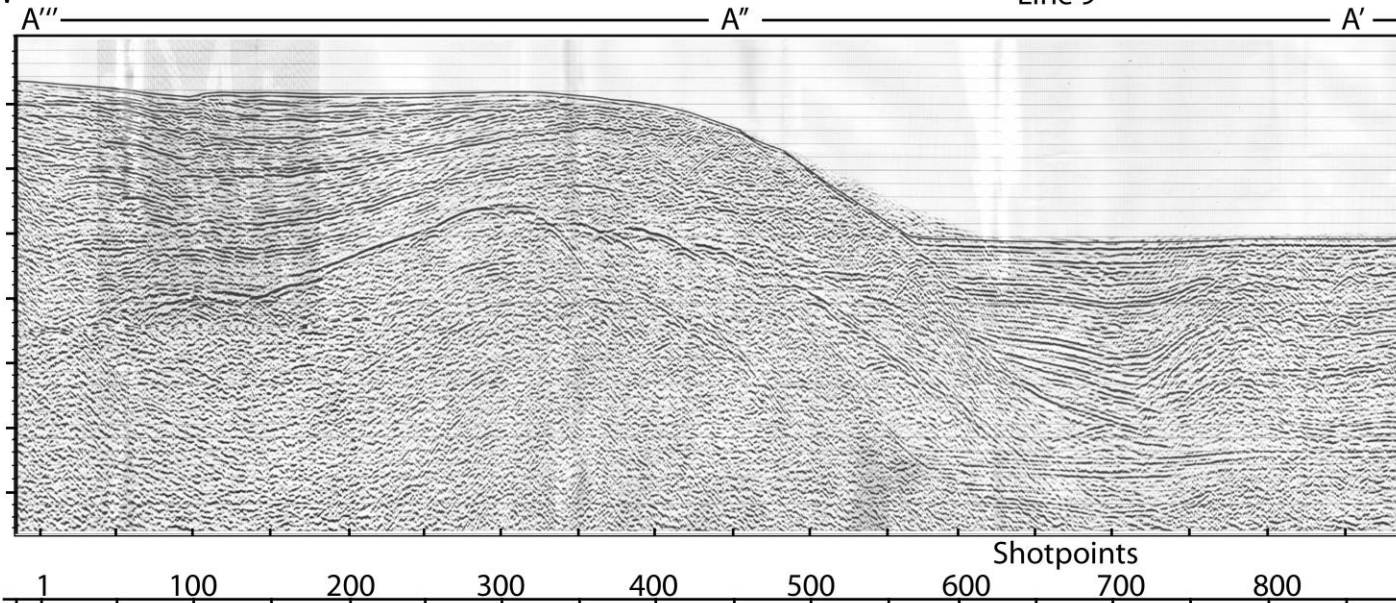






SSW

Line 9

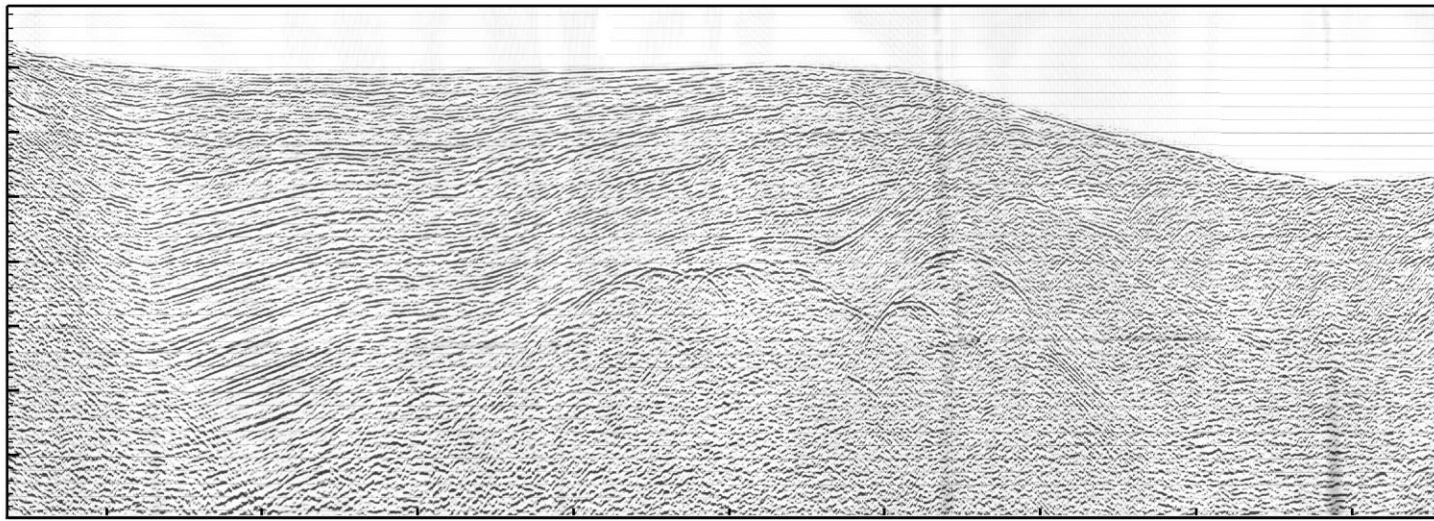




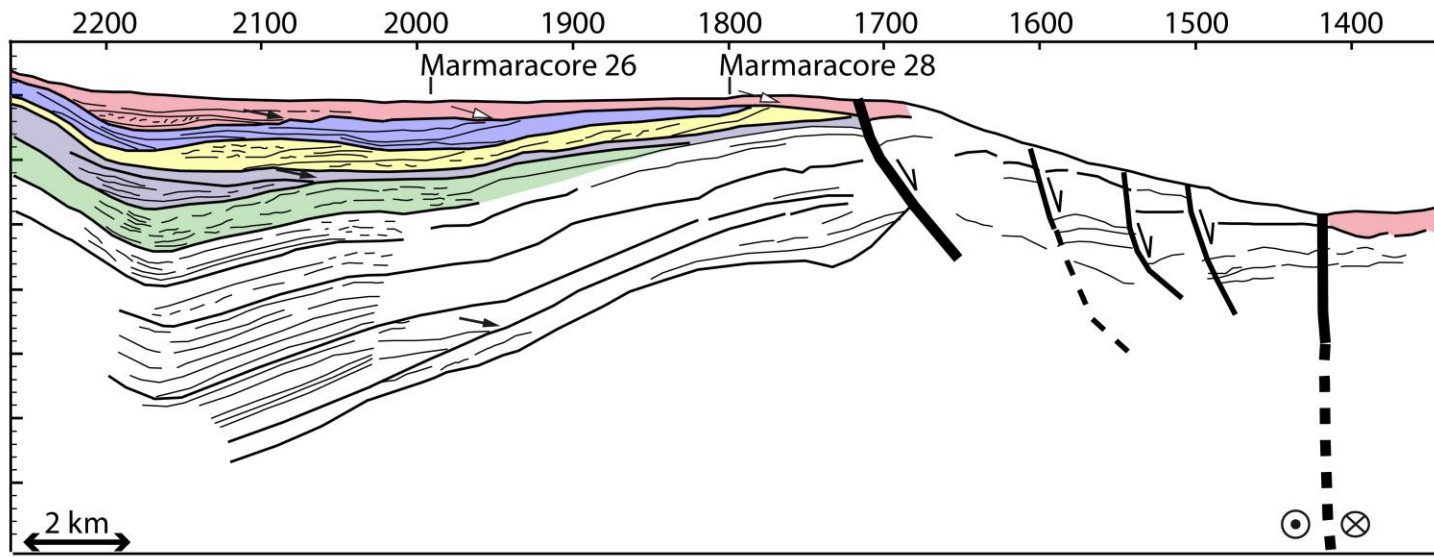
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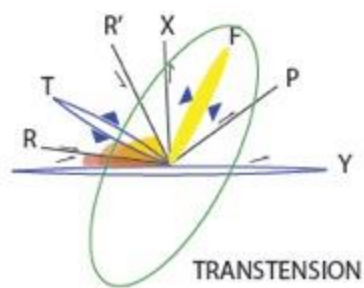
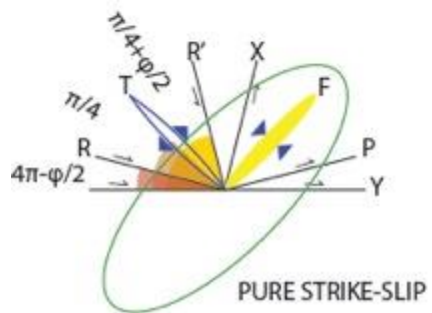
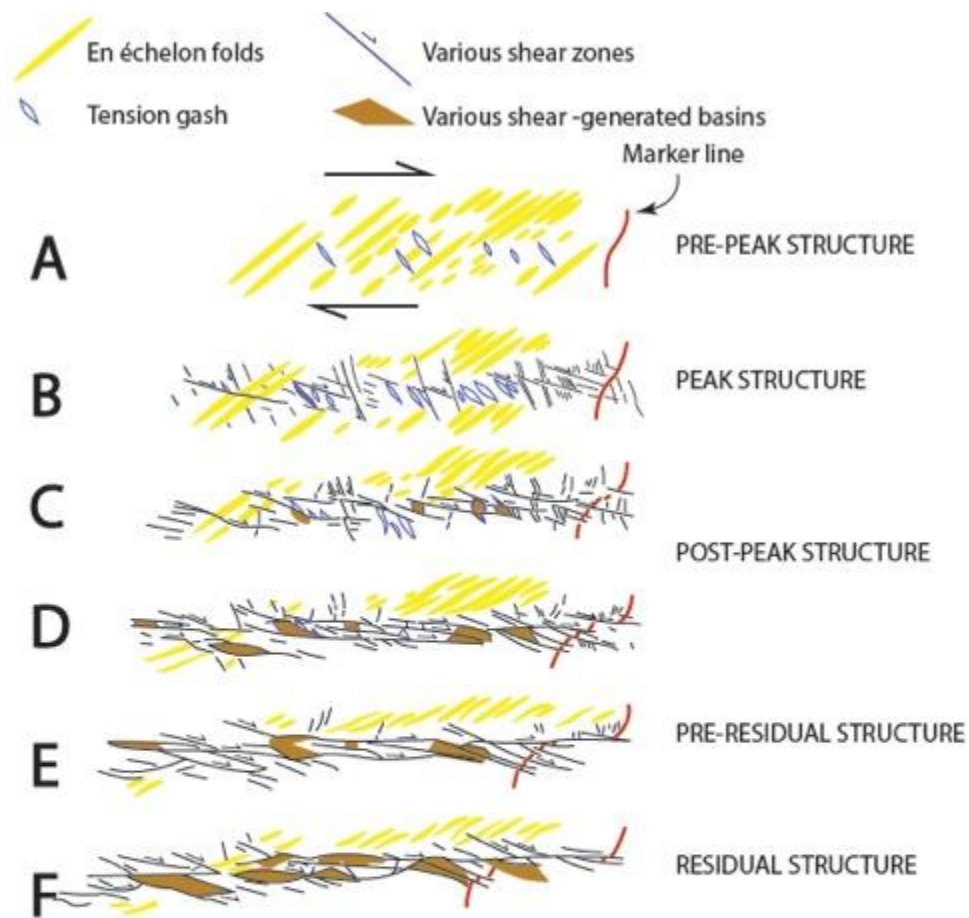
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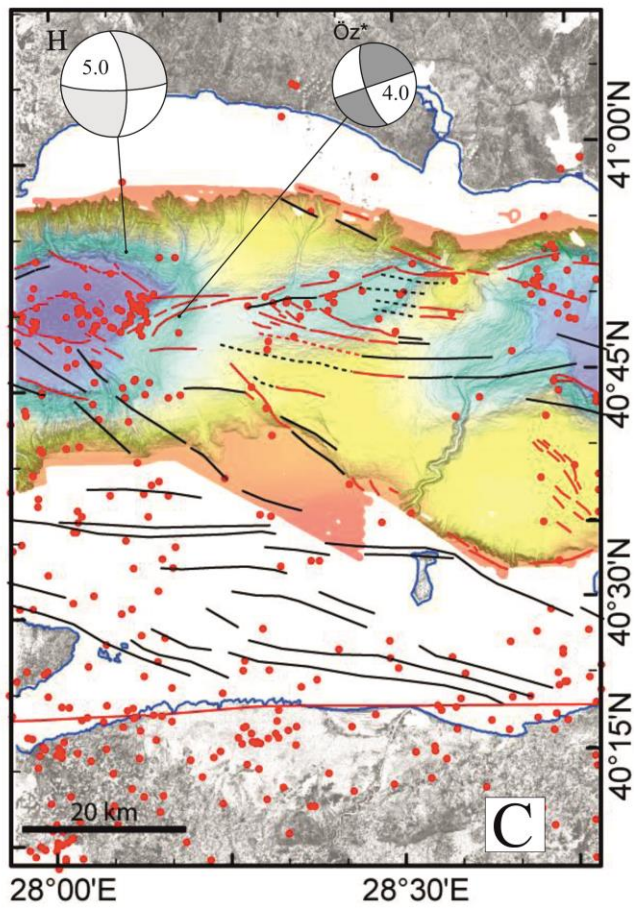
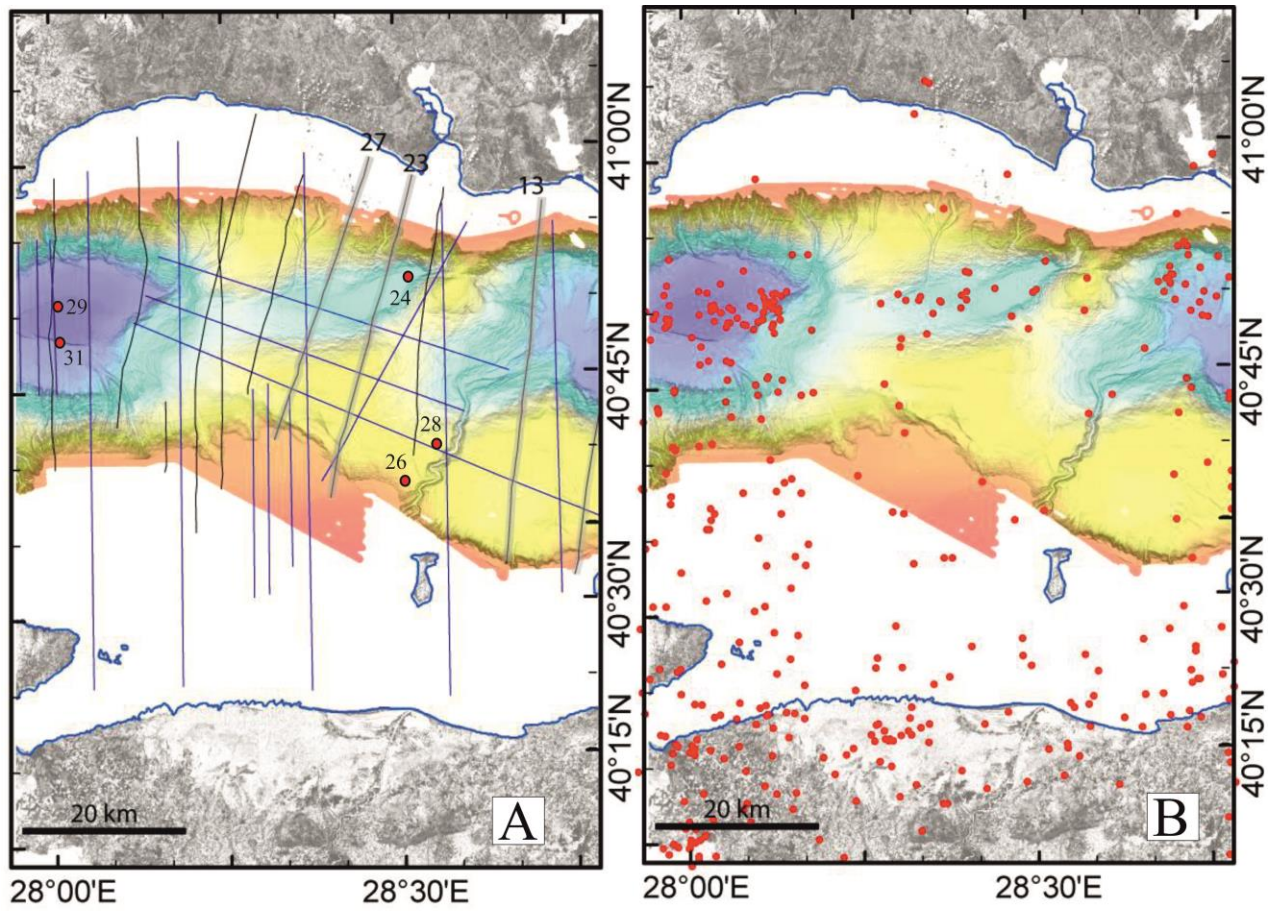
Line 13



Shotpoints





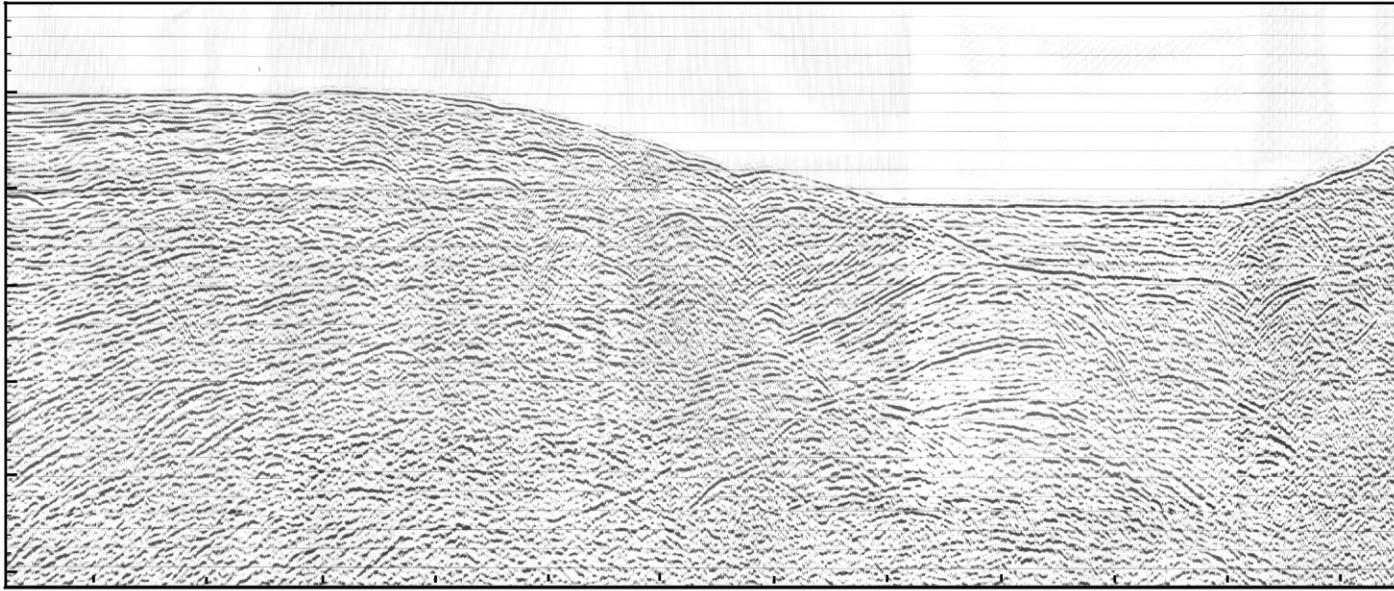


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C'''

Line 23

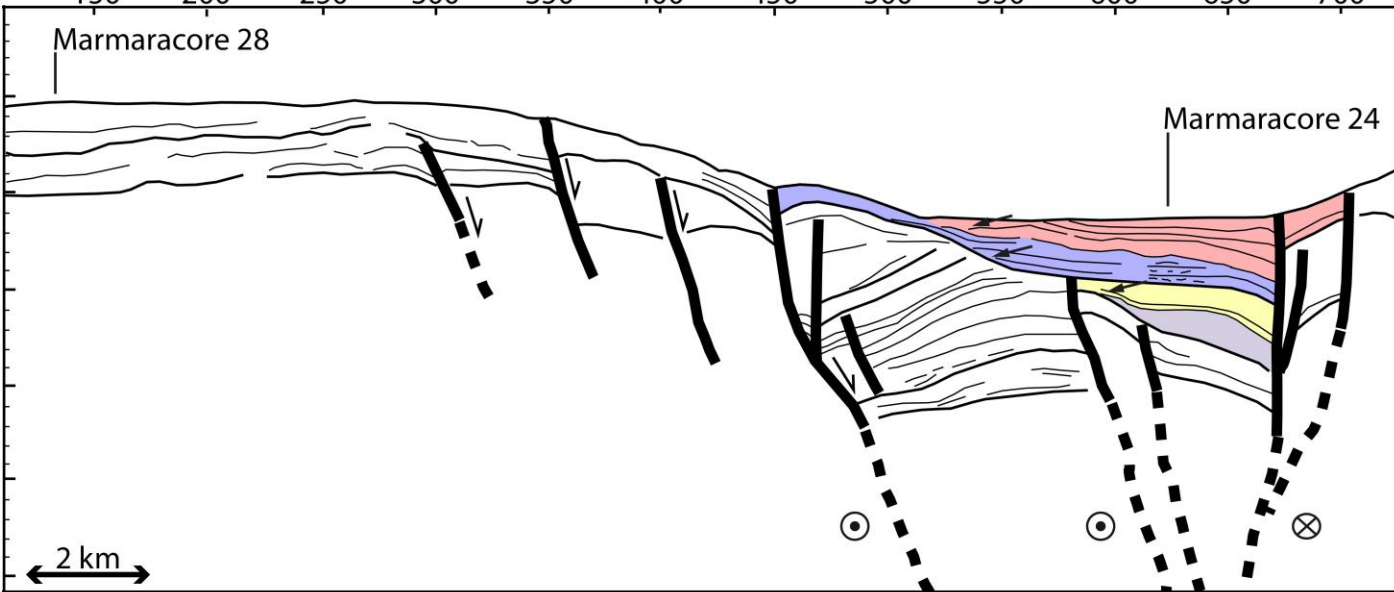
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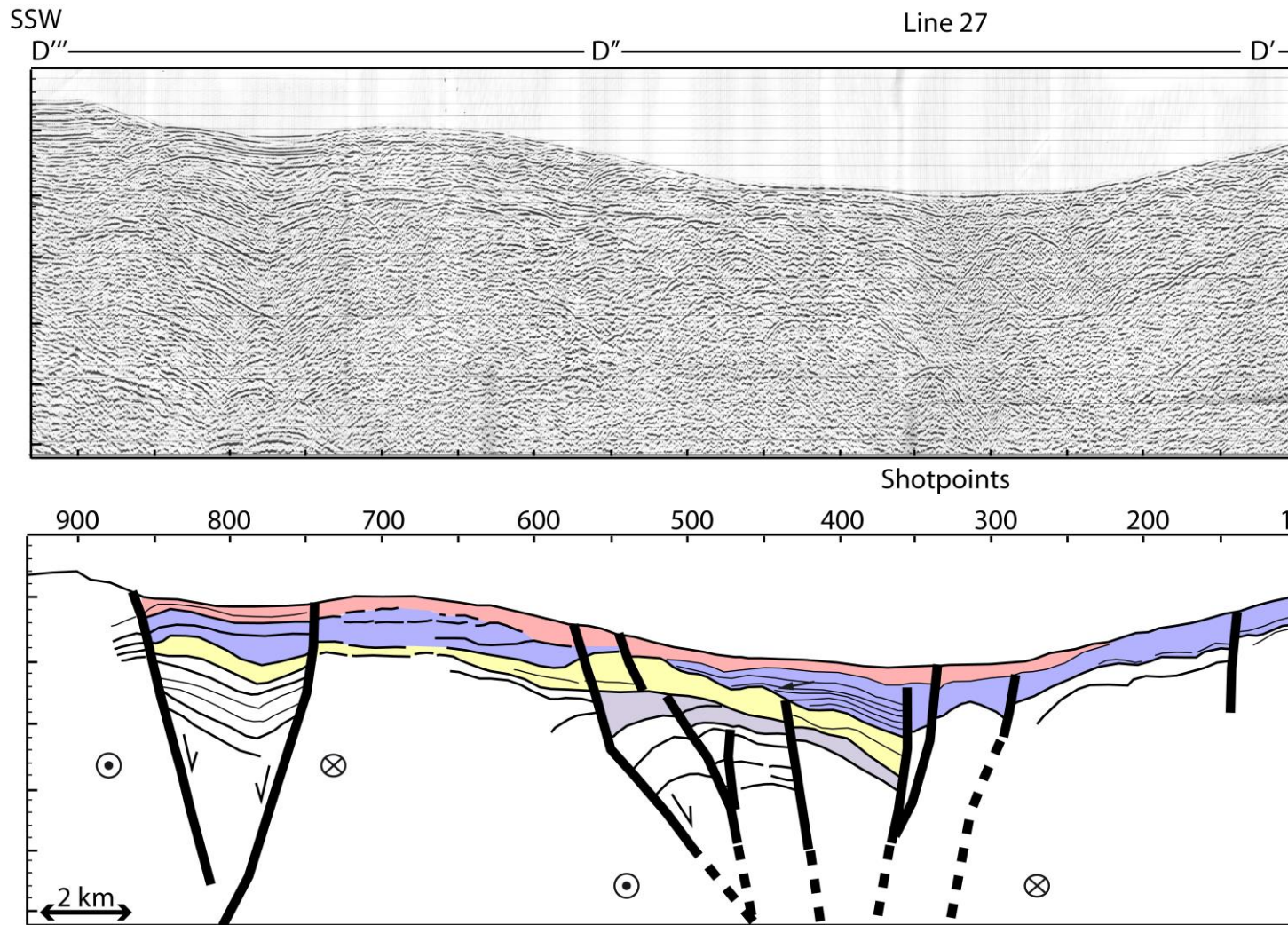
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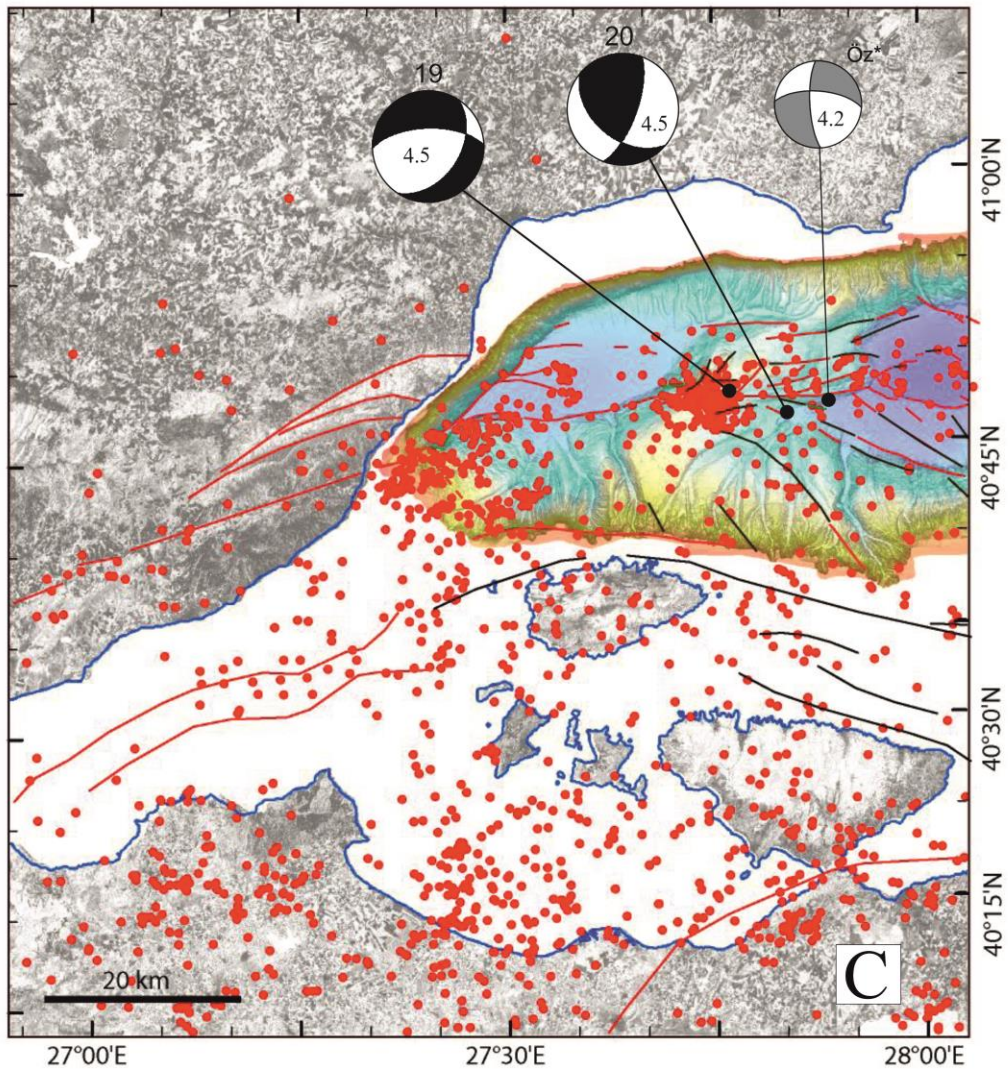
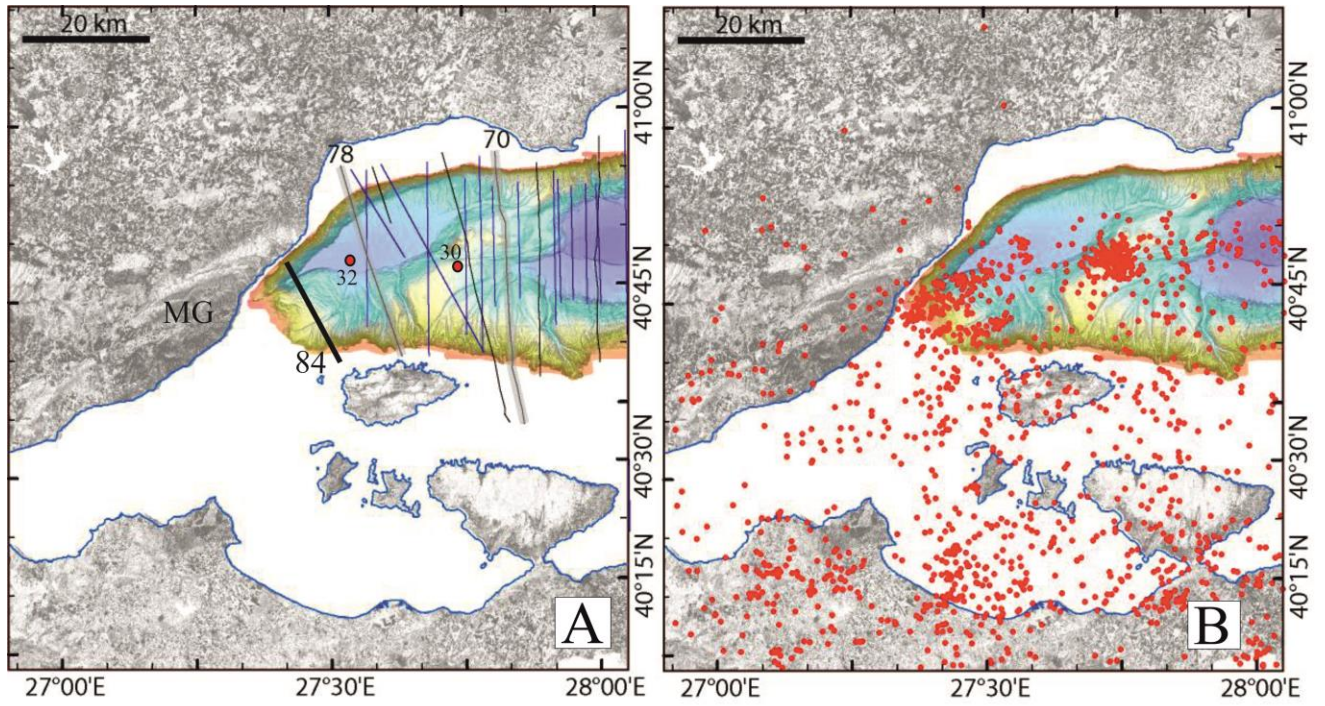
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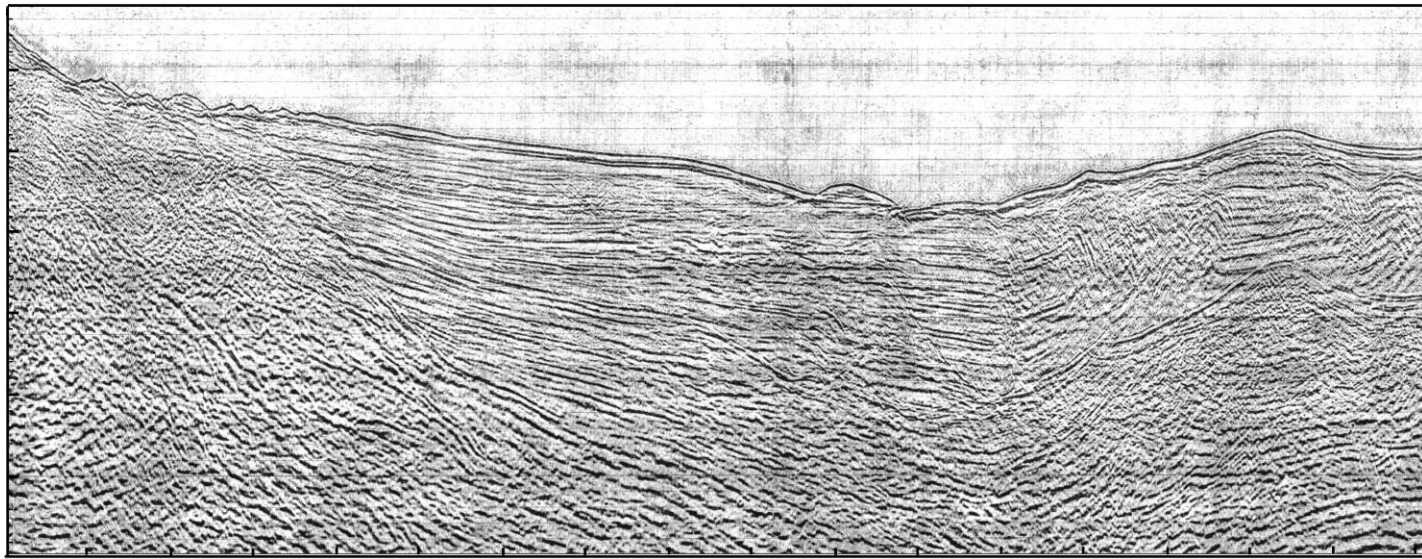


S

Line 70

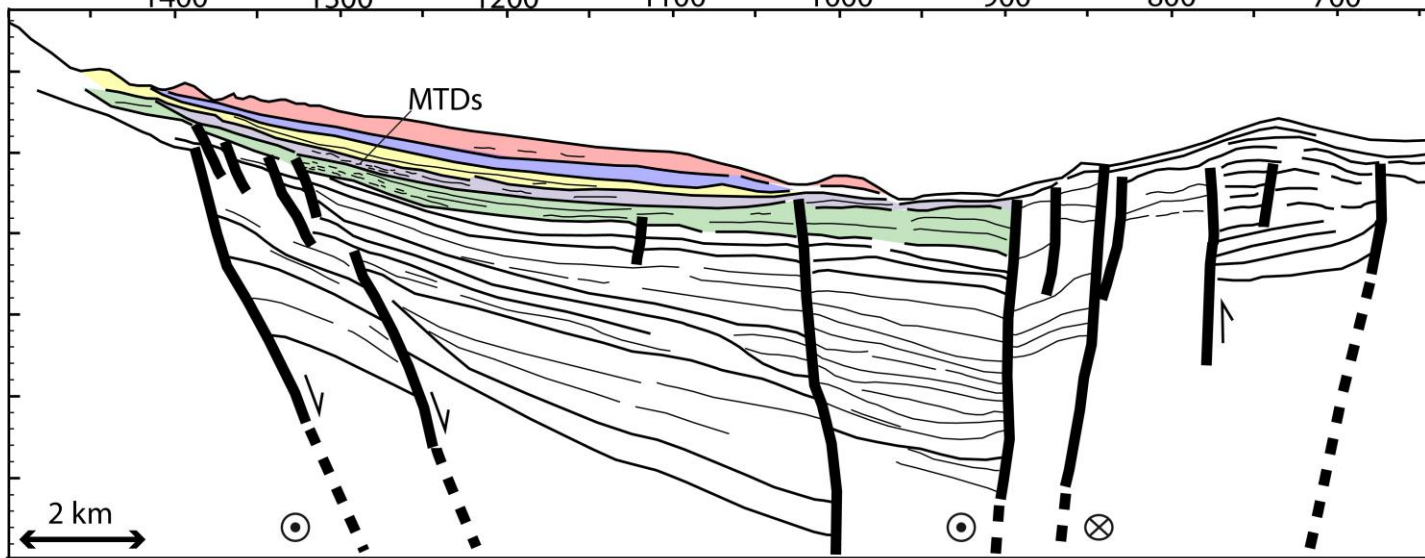
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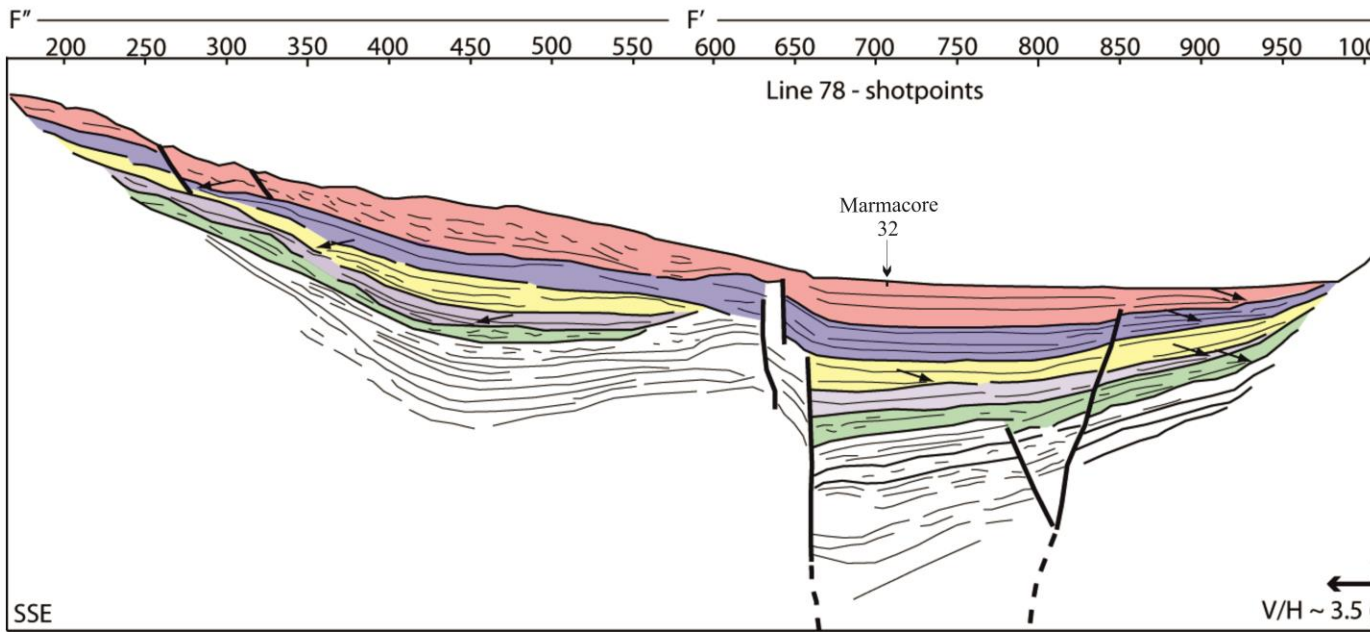
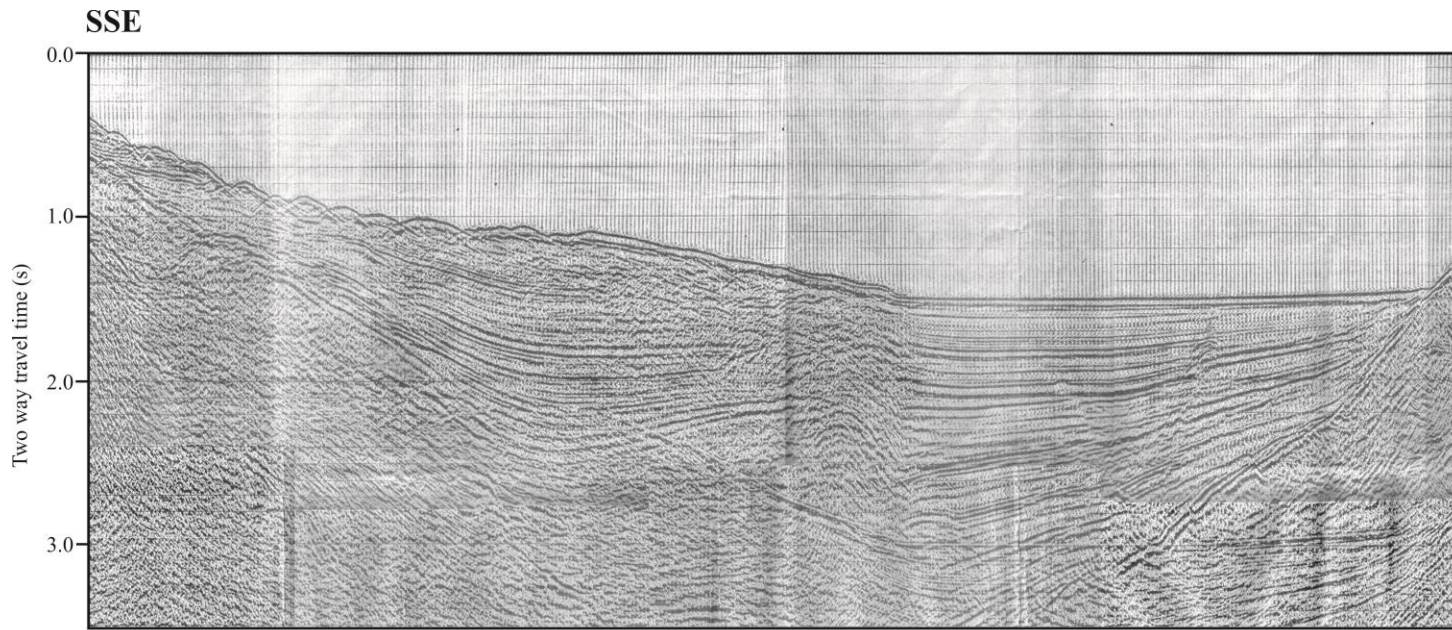
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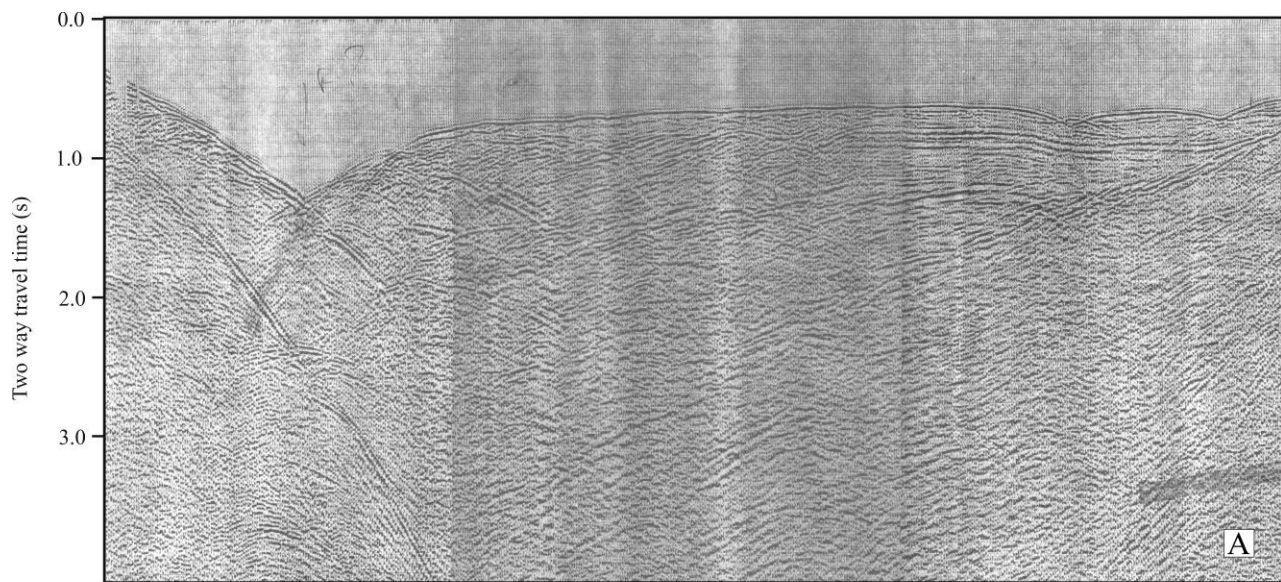
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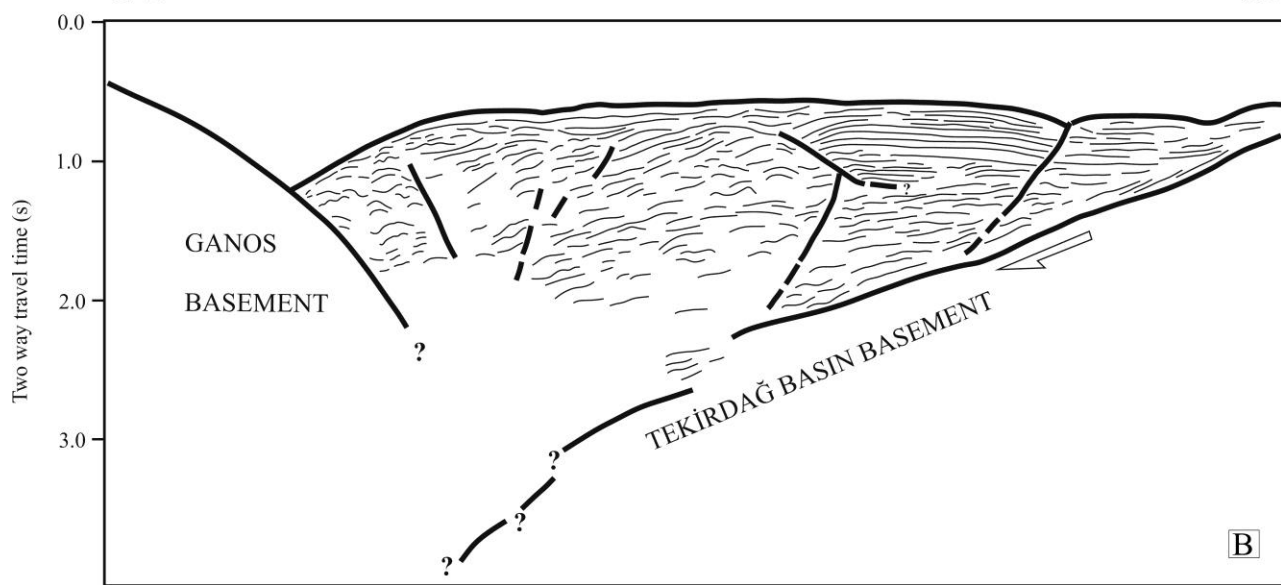


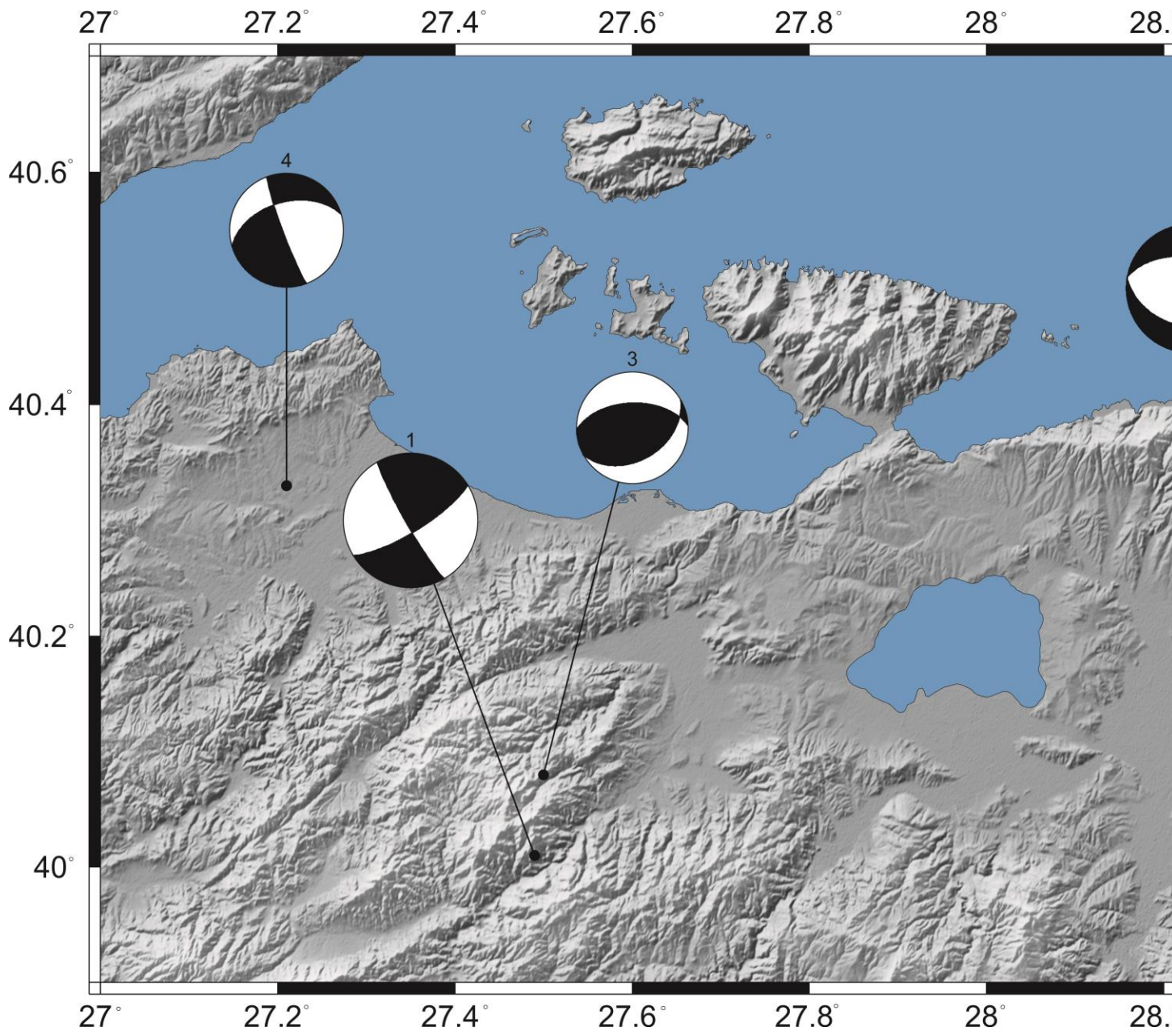




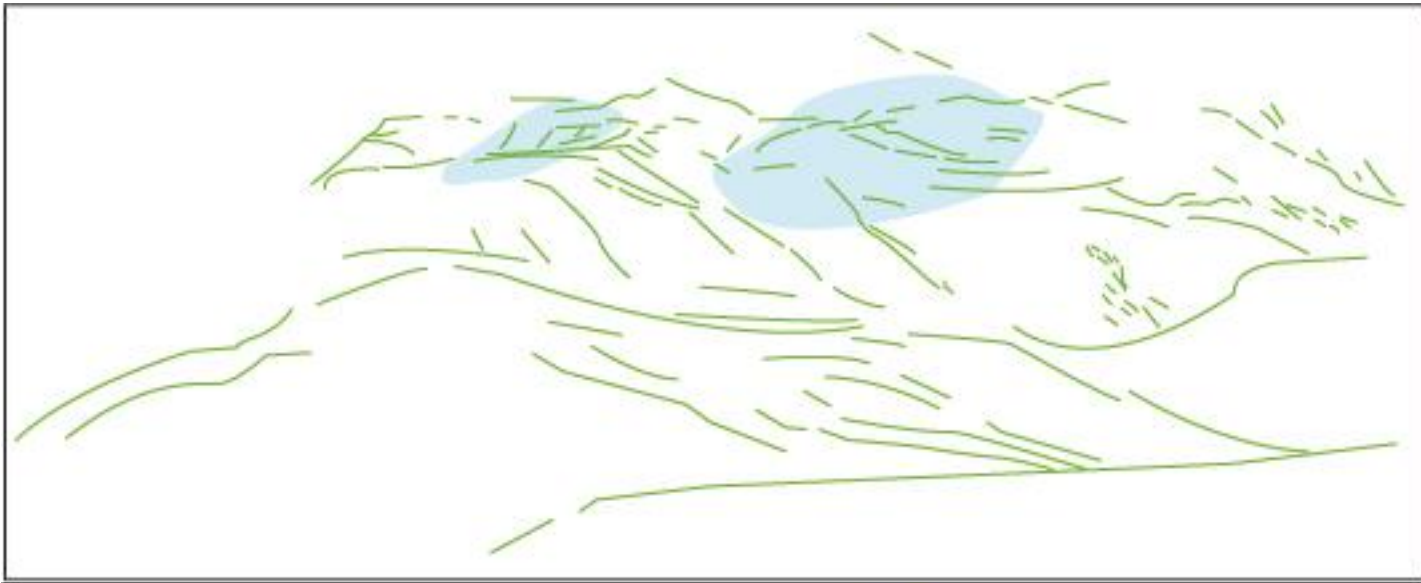
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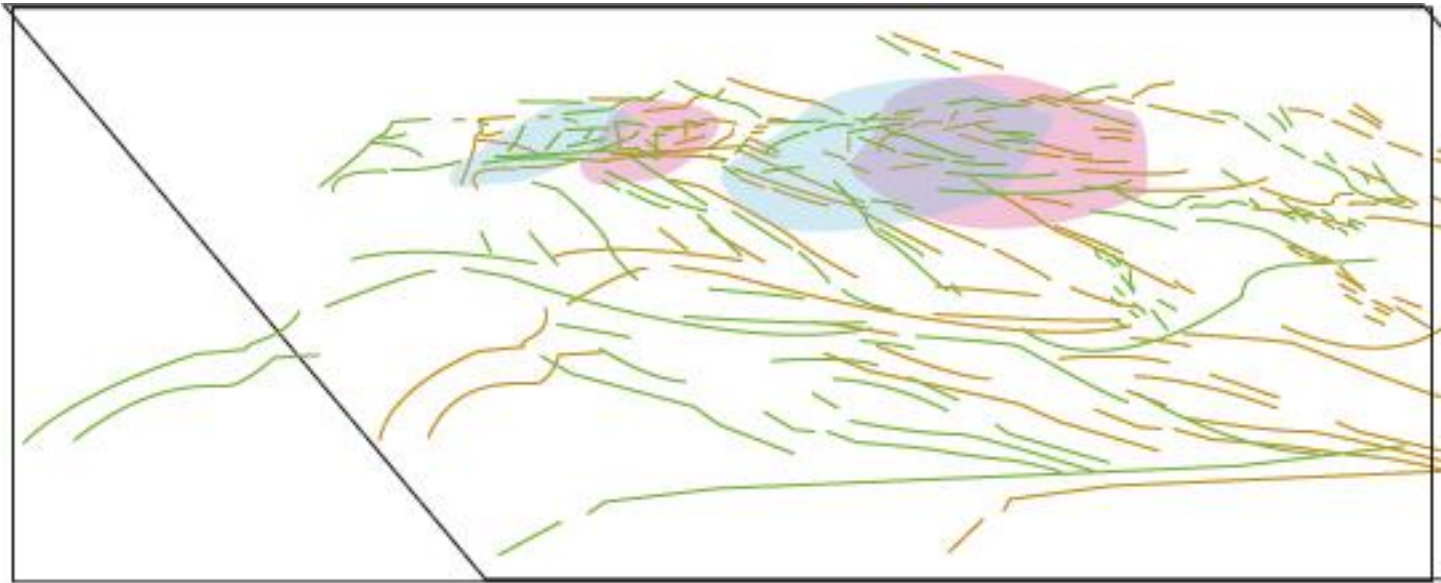
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