1 # Slow build-up of turbidity currents triggered by a moderate

2 # earthquake in the Sea of Marmara

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21 **Abstract.** Earthquake-induced submarine slope destabilization is known to cause

22 debris flows and turbidity currents, but the hydrodynamic processes associated with

23 these events remain poorly understood. Records are scarce and this notably limits

24 our ability to interpret marine paleoseismological sedimentary records. An

25 instrumented frame comprising a pressure recorder and a Doppler recording current

26 meter deployed at the seafloor in the Sea of Marmara Central Basin recorded

27 consequences of a $M_W = 5.8$ earthquake occurring Sept 26, 2019 and of a $M_w = 4.7$

28 foreshock two days before. The smaller event caused sediment resuspension but no

29 strong current. The larger event triggered a complex response involving a mud flow

30 and turbidity currents with variable velocities and orientations, which may result from

31 multiple slope failures. A long delay of 10 hours is observed between the earthquake

32 and the passing of the strongest turbidity current. The distance travelled by the

33 sediment particles during the event is estimated to several kilometres, which could

34   account for a local deposit on a sediment fan at the outlet of a canyon, but not for the

35   covering of the whole basin floor. We show that after a moderate earthquake,

36   delayed turbidity current initiation may occur, possibly by ignition of a cloud of

37   resuspended sediment. Some caution is thus required when tying seismoturbidites

38   with earthquakes of historical importance. However, the horizontal extent of the

39   deposits should remain indicative of the size of the earthquake.

40   1. Introduction

41   Triggering of mass flows and turbidity currents by earthquakes is a hazard that

42   can damage infrastructure at the seafloor (Heezen et al., 1954) and may enhance co-

43   seismic tsunami generation (Okal and Synolakis, 2001; Synolakis et al., 2002; Hebert

44   et al., 2005; Ozeren et al., 2010). It is often considered that a peak ground acceleration

45   (PGA) of the order of 0.1 g is needed for an earthquake to trigger a submarine slope

46   instability (Dan et al., 2008; Nakajima and Kanai, 2000). However, a global compilation

47   of cable breaks shows that, exceptionally, mass flow have been triggered by individual

48   earthquakes of M_w as low as 3.1 (with PGA $\approx 10^{-3}$ g) and that, on the other hand many

49    $M_w > 7$ have failed to break cables, notably in zones (e.g. Japan Trench) where

50   sediment input is relatively low and earthquakes frequent (Pope et al., 2016). In the

51   Mediterranean region, the threshold is reportedly around $M_w = 5$.

52   In spite of this high regional variability, turbidite deposits in several seismically

53   active zones have been used successfully as paleoseimological event markers

54   (Goldfinger et al., 2003, 2012; McHugh et al., 2014; Ikehara et al., 2016). This requires

55   distinguishing between seismoturbidites, caused by earthquakes and related mass

56   wasting events, from those resulting from other processes (e.g. floods, storms,

57   sediment loading). Seismoturbidites are generally described as turbidite-homogenites

58   where a basal silt-sand bearing layer is overlaid by a layer of apparently homogenous

59   mud with small or gradual, if any, variations in grain size and chemical composition

60   (Polonia et al., 2013; McHugh et al., 2011; Çağatay et al. 2012; Eriş et al., 2012;

61   Gutierrez-Pastor et al., 2013; Beck et al., 2007). In lakes and closed basins several

62   characteristics of deposits following earthquake or landslides, such as the sharp

63   boundary between turbidite and homogenite layers, the alternation of silt/sand and

64   mud laminae within a single turbidite-homogenite unit and presence of bi-directional

65   cross- or flaser- bedding have been interpreted as indicators of deposition from

66   oscillatory currents associated with seiches or turbidity current reflection (Beck et al.,

67   2007; Çağatay et al. 2012; McHugh et al., 2011). Seismoturbidites on ocean margins

68   have fairly similar characteristics to those in closed basins but their layering has been

69   interpreted differently, as a consequence of confluence (stacked or amalgamated

70   turbidites) or current speed variations (multi-pulsed turbidites) (Gutierrez-Pastor et al.,

71   2013; Nakajima and Kanai, 2000; Goldfinger et al., 2003). There is currently a lack of

72   in situ records that could substantiate inferred hydrodynamic processes. Monitoring

73   experiments brought records of representative cases of turbidity currents flowing in

74   submarine canyons and initiated by meteorological events and occasionally by

75   landslides (Azpiroz-Zabala et al., 2017; Khripounoff et al., 2012; Xu et al., 2004, 2010;

76   Liu et al., 2012; Hughes Clarke, 2016). Oscillatory currents resulting from internal

77   waves have been recorded after landslides in lakes (Brizuela et al., 2019). On the other

78   hand, most information on earthquake-triggered events is still indirect based on cable

79   ruptures (e.g. Pope, 2017; Hsu et al., 2008), combined with geomorphological and

80   sedimentological observations (Cattaneo et al., 2012; Piper et al., 1999), and

81   information from displaced instruments (Garfield et al., 1994). In Japan, in situ records

82   of pressure and temperature were obtained from displaced OBSs after the Tohoku

83   2011 M_w 9.1 earthquake (Arai et., 2013), and from cabled observatories after the

84   Tokachi-Oki 2003 M_w 8.3 earthquake (Mikada et al., 2006) and after a moderate (M

85   5.4) earthquake off Izu Penninsula (Kasaya et al., 2009). After the large events, strong

86   currents of more than 1 m/s were implied starting 2-3 hours after the earthquake with

87   no indication of oscillation or pulsing. In the off-Izu case a mudflow was observed with

88   a camera 5 minutes after the earthquake and followed 15 minutes later by a change in

89   current direction and speed.

90   We here present results from an instrumental deployment at the seafloor that

91   accidentally recorded the consequences of earthquakes that occurred 09/24/2019 and

92   09/26/2019 in the Sea of Marmara with respective magnitudes 4.7 and 5.8 (Figure 1A).

93   Holocene seismoturbidite records in the Sea of Marmara basins display a recurrence

94   of 200 to 300 years, that roughly corresponds to the recurrence interval of Mw>7

95   earthquakes (McHugh et al., 2006, 2014; Drab et al., 2012, 2015; Yakupoğlu et al.,

96   2018; Bulut et al., 2019). The pressure, temperature and current record from this single

97   instrument demonstrate that this moderate earthquake triggered turbidity currents.

98   However, the instrument suffered a rather complex sequence of disturbances and a

99   10 hours delay is observed between the earthquake and the recording of peak current.

100 We here propose a scenario which could explain the observations and discuss their

101 implications for the understanding of seismoturbidite records.
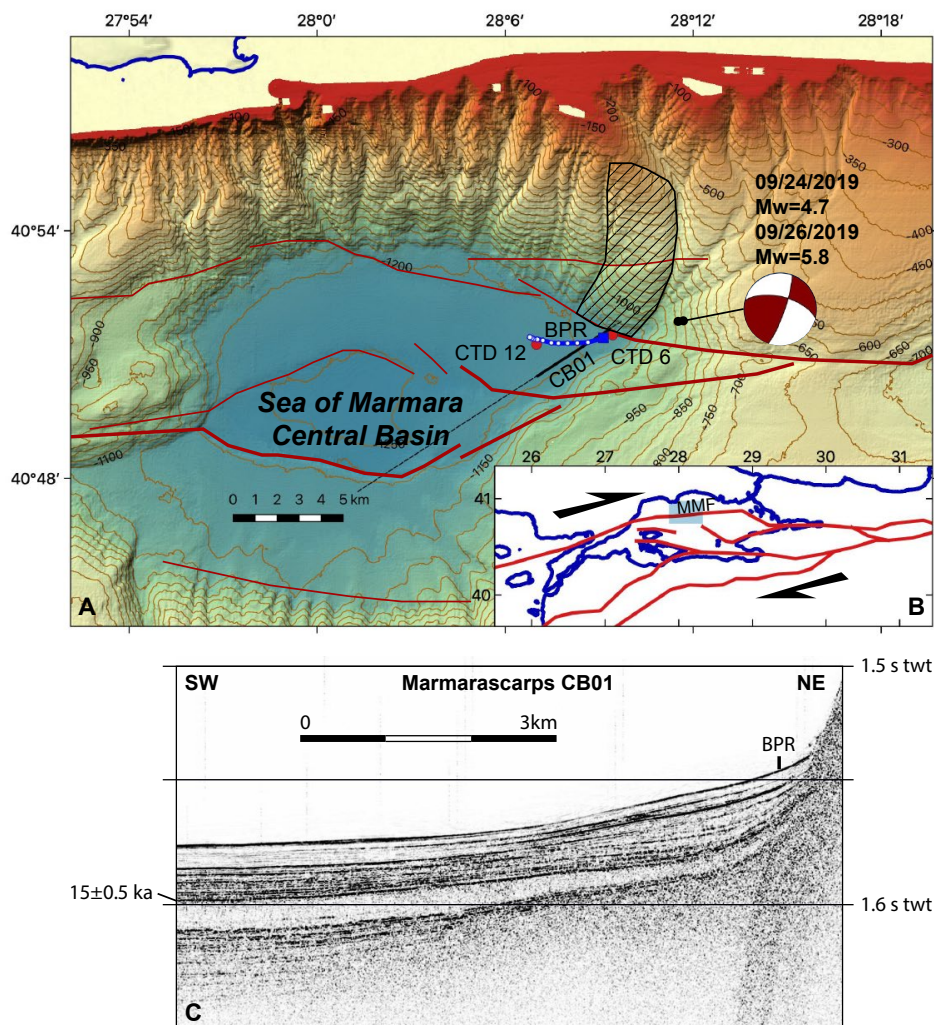
Figure 1. Context of instrumental deployment. (**A**) bathymetric map of the Sea of

104 Marmara Central Basin with simplified fault geometry (in red). The hatched zone is a

105 suspected mass wasting zone (Zitter et al., 2012). Location of instrumented frame

106 comprising bottom pressure recorder (BPR) and doppler current meter is indicated by

107 blue square. The blue banana with white dots represents the calculated trajectory of

108 a sedimentary particle during the waning phase of the turbidity current. Red dots are

109 CTD profiles 6 and 12 shown in supplementary material S1. Epicenter location of

110 earthquakes and the focal mechanism of the main shock are indicated. (**B**) Location

111 of study area. North Anatolian Fault system is shown in red. MMF is the Main

112 Marmara Fault. (**C**) Sediment sounder profile from Marmarascarps cruise (Armijo and

113 Malavieille, 2002). Indicative age of reflector from Beck et al. (2007). The instrument

114 (BPR) was deployed on a depositional fan at the base of slope and canyon outlet that

115 differ in character from the hemipelagite / turbidite-homogenite sequence in the

116 basin.

2. Context and data collection

118 A series of instrumental deployments was planned to record naturally occurring

119 resonant water column oscillations (seiches) at various locations in the Sea of

120 Marmara with the aim to improve tsunami models (Henry et al., 2021). An instrumented

121 frame was thus deployed at 40.8568° N, 28.1523° E and 1184 m water depth in the

122 Central Basin on May 9, 2019 and recovered 6 months later (11/19/2019) (Figure 1A).

123 This site is located at the outlet of a branched canyon system originating from the edge

124 of the continental shelf (Figure 1). Sediment sounder profiles indicate a depositional

125 fan or lobe is present at this location (Figure 1C). The short canyons observed on the

126 relatively steep sedimented slope (≈10°) of the Sea of Marmara deep basins are

127 presumably fed by instabilities of the canyon heads and walls (Zitter et al. 2012;

128 Çağatay et al., 2015). In addition, the slope west of the canyons immediately north of

129 the deployment site hosts a mass wasting feature covering about 24 km^2 (Zitter et al.

130 2012). The Main Marmara Fault (MMF, Figure 1B), is defined as the part of the northern

131 branch of the North Anatolian Fault system crossing the Sea of Marmara (Le Pichon

132 et al., 2001, 2003). A splay of the MMF runs along the base of this slope (Armijo et al.,

133 2002; Grall et al., 2012; Sengor et al., 2014). The 09/24/2019 and 09/26/2019

134 earthquakes occurred beneath the canyon system and their epicenters are located 5

135 km ENE of the instrument, less that 500 m apart (Figure 1). The rupture occurred within

136 the crust at 9-13 km depth on a northward dipping fault located north of principal

137 displacement zone of the Main Marmara Fault. The focal mechanism and aftershock

138 distribution indicate right-lateral strike-slip with a reverse component (Karabulut et al.,

139 2021). The rupture did not reach the seafloor, nor caused a tsunami. For instance, tidal

140 gauge records obtained at Marmara Ereglisi do not deviate more than 1hPa from a

141 fitted tidal model.

142   The instrumentation on the frame comprises (1) an RBR bottom pressure recorder
143   (BPR) with a Paroscientific 0-2000 m Digiquartz sensor, (2) a Seaguard recording
144   current meter (RCM) equipped with a Z-pulse 4520 Doppler current meter operating in
145   the 1.9-2 MHz frequency range and other sensors: temperature, pressure (tide sensor
146   Aanderaa 5217), conductivity (Aanderaa 4319), oxygen (Aanderaa optode 4330)
147   (Figure 2). The RBR pressure recording interval was set to 5s and that of the Seaguard
148   RCM to one hour for all sensors. The Doppler current meter worked in burst mode,
149   averaging 150 pings taken every second at the end of each one-hour recording interval.
150   The SeaGuard instrument was fixed on the upper part of the frame 1.5 m above the
151   seafloor and emit 4 narrow (2°) beams at orthogonal directions in a plane, parallel to
152   the seafloor if the frame is standing upright, and measures Doppler backscatter in cells
153   extending 0.5-to-2 meters from the instrument (Figure 3). The instrument was set in
154   forward ping mode, so that only data from sensors measuring a positive Doppler shift,
155   upstream currents moving toward the instrument, are used to calculate current speed.
156   The tide sensor is a piezoresistive sensor with a specified accuracy comparable to that
157   of the Digiquartz sensors (4kPa for a 0-2000 m sensor vs. 2kPa for a Digiquartz sensor
158   with the same range) and 0.2 hPa (2 mm) resolution and comprises a temperature
159   sensor of 0.2°C accuracy and 0.001°C resolution. The tide sensor averages pressure
160   measured at a 2 Hz sampling rate over 300 s at the end of each one-hour time interval.
161   the tide sensor was checked against an atmospheric reference between deployments
162   and found to have a minimal drift, less than 1 hPa.
163   As we will show that the 09/24/2019 earthquake caused the instrumented device
164   to lay on its side for several hours and then straighten up, understanding the setup of
165   the seafloor device and its stability is important (Figure 2B). The frame is made of
166   aluminium and has 6 rigidly bound flotation spheres of 25 kg buoyancy each. The net
167   weight of the instrumented frame in water is -80 kg. The frame is rigidly attached to a
168   12-cm-thick 1.5x1.3 m concrete slab, weighting 300 kg in water. The assembly of the
169   heavy slab and buoyant frame is stable in upright position in the water and on the
170   seafloor. Moreover, it is estimated that a current of 1 m/s would cause a total horizontal
171   drag of 75 kg when the device is in upright position, which is insufficient to destabilize
172   it. If a stronger current, or other external forces, cause the assembly to tilt and lay on
173   one side, the moment of the gravity and buoyancy forces should straighten the device
174   back to upright position when these external forces are removed.

175   Measurement of current speed and direction by a tilted instrument is a related issue
176   that we here consider. The orientation and attitude of the Seaguard RCM is measured
177   with a 2-component accelerometer and a magnetic compass and the recorded data
178   include tilt in X and Y direction and the heading of the X axis. Tilt X and Y components
179   are factory calibrated from -35° to +35° with an accuracy of 1.5°. Tests performed in
180   the laboratory (see supplementary material, Figure S1) showed that tilt information
181   remains consistent outside this range, even when the instrument is upside down. Tilt
182   measurements are accurate within 3° up to 60° but saturate at about 80° (Figure S2).
183   Uncertainty on heading also increases with tilt, especially when the instrument is tilted
184   toward the X-direction. However, measured heading remains ±20° of true heading for
185   a tilting of up to 60° (Figure S3). The current measured in the instrument plane is
186   corrected for tilt assuming current is horizontal. As far as this approximation is valid,
187   the current record should in principle remain fairly accurate when the instrument is
188   tilted beyond the normal range of operation (±35° degree) and at least to 60°. However,
189   the compass was not calibrated for an upside-down configuration. If the top of the
190   instrument would happen to be oriented downward, the measured current direction will
191   be unreliable, even though the absolute speed may still be correctly estimated. Another
192   problem may arise if one of the Doppler sensors is facing down into the sediment so
193   that its measurement cell is below the seafloor. If the sensor pointing upward in the
194   opposite direction is recording a negative Doppler shift, this value will be ignored in the
195   forward ping mode. In this case, the measurement retained to calculate current velocity
196   will correspond to noise from the sensor facing toward the seafloor. In all situations, it
197   remains possible to recalculate the sensor readings retained by the calculator from the
198   current velocity and orientation parameters recorded by the instrument, and thus
199   assess the reliability of data.
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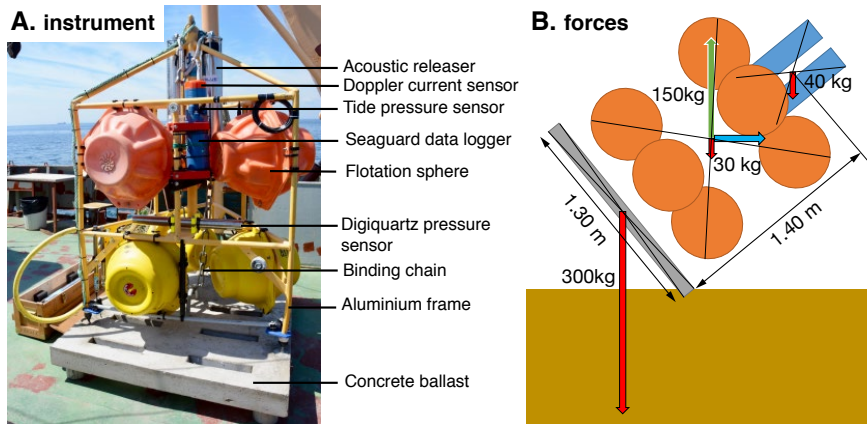
Figure 2. Instrumented frame. (A) photo of the instrumented frame before
deployment. (B) Sketch showing forces applied to the elements of the instrumented
frame in water. The red arrows represent the weight in water of the cement ballast, of
the instrumented frame and of the acoustic release system on top. The green arrow
represents the buoyancy of the flotation spheres. The blue arrow represents the
current drag, which depends on current speed and instrument tilt.
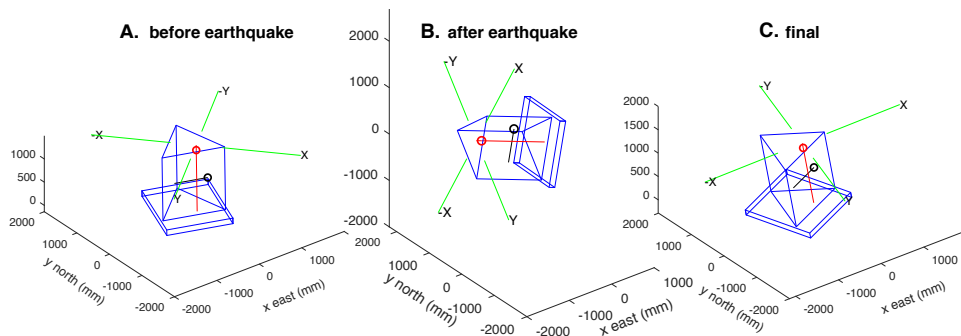
Figure 3. Reconstruction of frame position based on instrument tilt-meter and
compass data: (A) before the earthquake; (B) Tilted, between, 25 minutes and 10.5
hours after earthquake; (C) back in nearly upright position 11 hours after earthquake.
Position of Digiquartz pressure sensor (black circle), Aanderaa tide sensor (red
circle) and Doppler current meter beam cells (green segments)

3. Results and interpretations

3.1. Pressure and tilt records

Pressure sensors are sensitive to pressure variations caused by P-waves and Digiquartz sensors are also intrinsically sensitive to acceleration, but to a small extent, 160 hPa/g for an instrument with 20 MPa range according to the calibration report. Small earthquakes are detected as pressure spikes, while oscillations are recorded after large earthquakes. The M_w 4.7 09/24/2019 caused a short pressure transient of 25 hPa at 08:00:26 followed by small pressure oscillations of less than 3 hPa amplitude decaying over a few minutes. The seismic wave train from the M_w 5.8, 09/26/1919 earthquake is recorded by the Digiquartz pressure sensor as oscillations, initiated by a pressure drop of 65 hPa between 10:59:22 and 10:59:26 (Figure 4). For the sampling interval of 5s used in this setup, the recorded signal is aliased, which precludes quantitative interpretation in term of velocity or acceleration. However, the initial pressure drop after the 09/26/1919 earthquake may indicate a negative polarity of the first P arrival at the instrument site, located on an ascending ray-path.

Twenty-five minutes after the M_w 5.8, earthquake, a new disturbance of the pressure sensor is observed at 11:23:41. The pressure then progressively increases by 30.9 hPa in 15 seconds between 11:24:46 and 11:25:01 before stabilizing. Over the corresponding one-hour-time-interval between successive records, the Seaguard RCM, initially subvertical (tilt less than 2°), acquires a strong tilt (Figure 3). At 11:57:48, measured tilt is -65° along the X-axis and +19° along the Y-axis, with X-axis in a N161° azimuth and these values remain constant ±2° over the next 10 hours, corresponding to an absolute tilt of 68° (Figure 4). The tilting of the instrument causes the Digiquartz and Tide sensors to record different pressure variations because they are located at different positions on the frame (Figure 2). Moreover, the pressure readings by the Digiquartz sensor also depend on its orientation relative to Earth gravity. Pressure at the Tide sensor location increases about 100 kPa, corresponding to a 1 m drop and indicating that the frame was then practically laying on its side. Ten hours later, the device apparently straightens itself in about 5 seconds, between 21:28:29 and 21:28:34 as indicated by a rapid pressure variation. After that, the recorded tilt parameters are moderate and stabilize at -11.5° for the X-axis and 5.3° for the Y-axis, with X-axis in a N105.3° azimuth.
250   The M_w 4.7 earthquake caused minor disturbances of the attitude of the
251   instrument, with variations of tilt and heading of less than 0.5°. A M_w 3.6 foreshock of
252   the M_w 5.8 occuring 26/09/2019 at 7:32 also caused minor disturbances. These
253   indicate that the seafloor was sensitive to ground shaking caused by these small
254   earthquakes, however, this did not cause the device to sink into the sediment. Changes
255   of pressure baseline of the digiquartz sensor between before and after these
256   earthquake are difficult to resolve, and correspond to less than 5 mm vertical
257   displacement for the first event and less than 2 mm for the second one.

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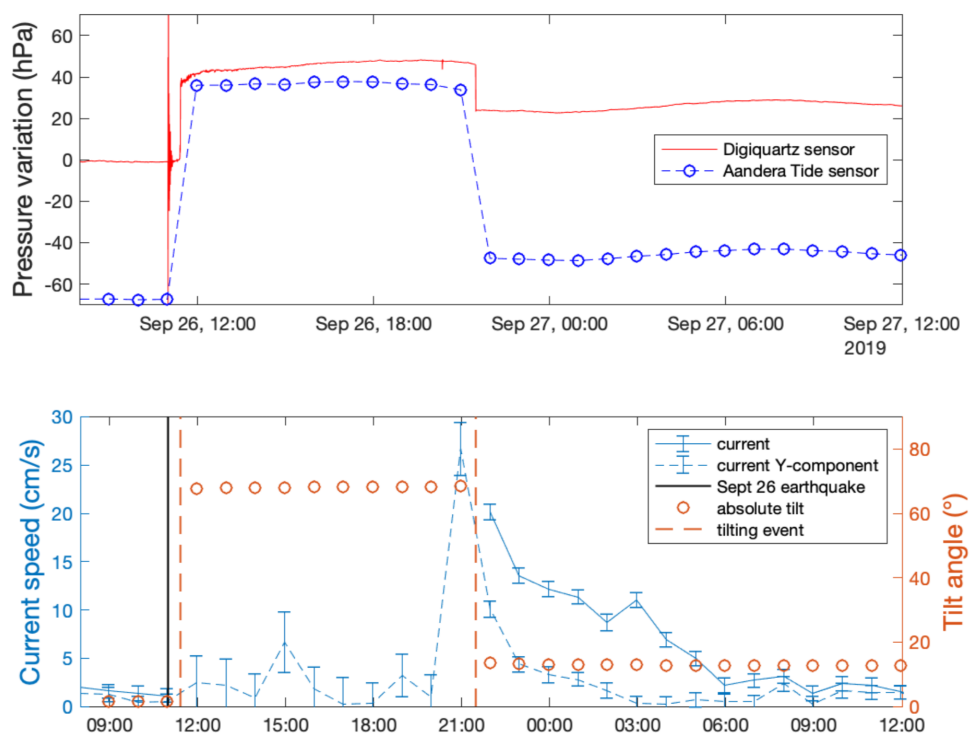
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260   **Figure 4.** Time series around the time of occurrence of a M_w 5.8 earthquake; (top)
261   pressure variations recorded by two instruments on the instrumented frame; (bottom)
262   current and tilt data recorded by Seaguard RCM. Between the tilting events only one
263   component of the doppler current meter functioned reliably (Y-component oriented
264   N200) and is here reported.
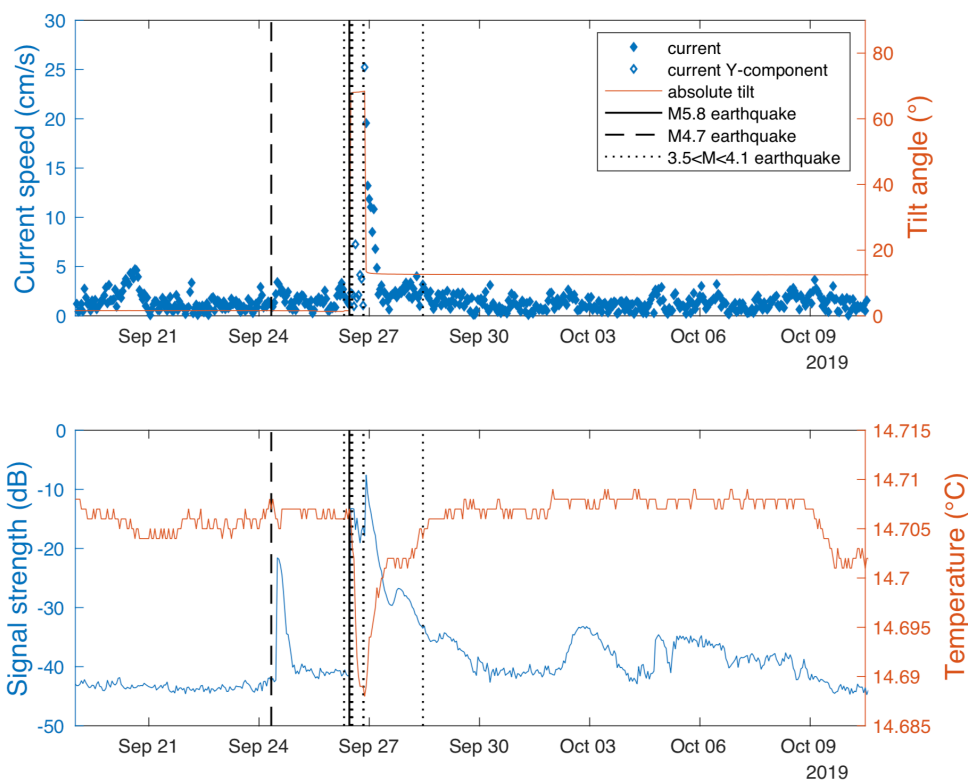
Figure 5. Time series acquired with Seaguard RCM during the September 2019 seismicity cluster; (top) current speed and tilt; (bottom) backscatter signal strength and temperature.

3.2. Current records and possible causes of tilting

The M_w 4.7 09/24/2019 earthquake was followed by a small increase of current strength peaking at 3.4 cm/s at noon, 4 hours after the earthquake (Figure 5). Comparable events in term of duration and strength occurred spontaneously 09/20/2019 (with currents up to 4.7 cm/s) and 09/26/2019 just before the main earthquake. During all three events the dominant current was from the east, thus coming from the direction of the canyon, but there is an important difference between the event that occurred after the earthquake and the two others. During that event a change in current direction occurred from eastward to westward between 10:57 and 11:57 while the current strength increased from 2.2 cm/s to its peak value (Figure 6). During the other events, build-up was more progressive and did not involve a change in direction. A drift plot, calculated by summing velocity vectors over time, reproduces

282   the motion of a particle assuming a uniform velocity field (Figure 6). The total drift is

283   about 500 m and occurs in the 8 hours following the current inversion. Current direction

284   varies from Westward to Northward during this time interval.

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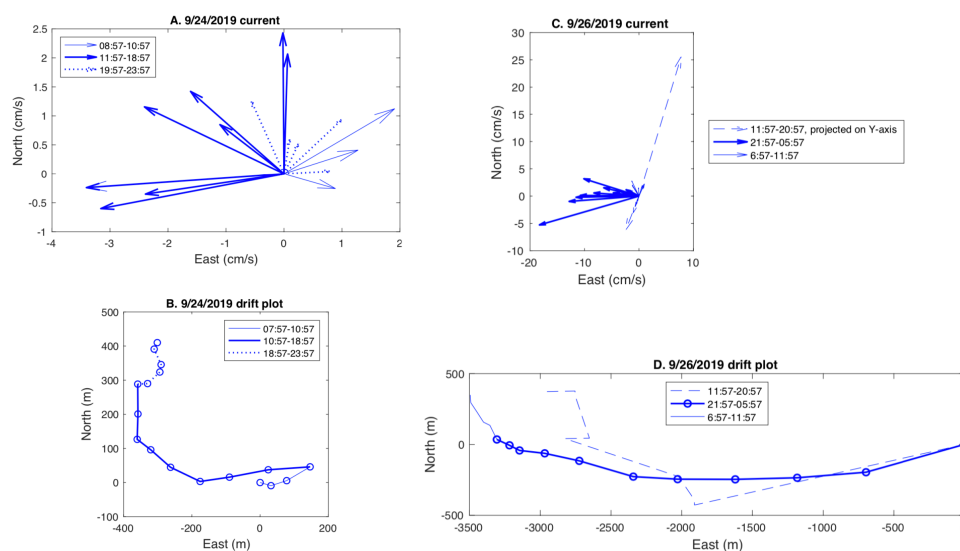
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288   **Figure 6.** Current recorded after M_w 4.7 and M_w 5.8 earthquakes: (A) Current velocity

289   arrows recorded every hour between 08:57 and 23:57 on 09/24/2019; (B) drift plot

290   over the same time interval, the change of current direction and strength between

291   10:57 and 11:57 coincides with increasing backscatter strength (see figure 4),

292   indicative of increased turbidity; (C) Current velocity arrows recorded every hour

293   between 12:00 09/26/2019 and 06:00 09/27/2019. Dashed arrows show

294   measurements acquired in the Y direction when the instrument was strongly tilted

295   (position B in Figure 3), plain arrows when it was back in upright position (C in Figure

296   3); (D) drift plot over the same time interval, the dashed part corresponds to the

297   strongly tilted position.

298

299   After the M_w 5.8 09/26/1919 earthquake, during the 10 hours period when the

300   instrument remained strongly tilted, the instrument recorded currents varying both in

301   speed and orientation, but some precautions are needed when interpreting these data.

302   The current component measured by transducers along the Y-axis of the instrument,

303   oriented N200°, probably remained accurate as the tilt along this axis is less than 20°

304 and the measurement cell remained above the bottom (Figure 2B). On the other hand,
305 the X-component may not be reliable as one of the sensors (n°1) is oriented 65°
306 upward in the N160° direction, and the oposite sensor (n°3) is dipping 65° downward
307 in the opposite (N340°) direction. Consequently, measurement cell n°3 lies within the
308 sediment and thus may only record noise. Moreover, because the Doppler current
309 sensor (DCS) is set in forward pinging mode, current speed is calculated with data
310 from sensors measuring positive doppler shifts only. This implies that if the current
311 component toward N160° is positive, sensor n°1 will measure a negative shift and will
312 not be recorded. During the time inteval here considered, the mesured current
313 component in the X-direction (toward N160) is positive, which indicates that data from
314 sensor n°4 was used (Figure 7), and that is probably noise. It follows that the current
315 component along the Y-direction is the only one reliable. The horizontal current
316 measured along the Y-axis changed sign several times during this time interval, and
317 reached peak values of 6.3 cm/s toward N200 at 14:57:46, about four hours after the
318 earthquake, and of 25 cm/s in the oposite direction at 20:57:46, the last measurement
319 before the instrument straightenned up. Other measurements on both axes remain
320 below 5 cm /s, but the absolute velocity may have been higher because this
321 measurement was only performed in one direction. Yet, these observartions suggest
322 that the strong current recorded before the instrument straightenned up played a role
323 in this event. Once the device got back in an upright position, it recorded a current
324 consistently flowing westward and progressively decreasing from 20 cm/s to
325 background level (2 cm/s) in 9 hours (Figure 4). During this waning phase, the current
326 drift is about 3.5 km in a westward direction (Figure 6). The drift estimated during the
327 first 10 hours after the earthquake, while the instrument was strongly tilted, is in the
328 opposite direction but may not be reliable.
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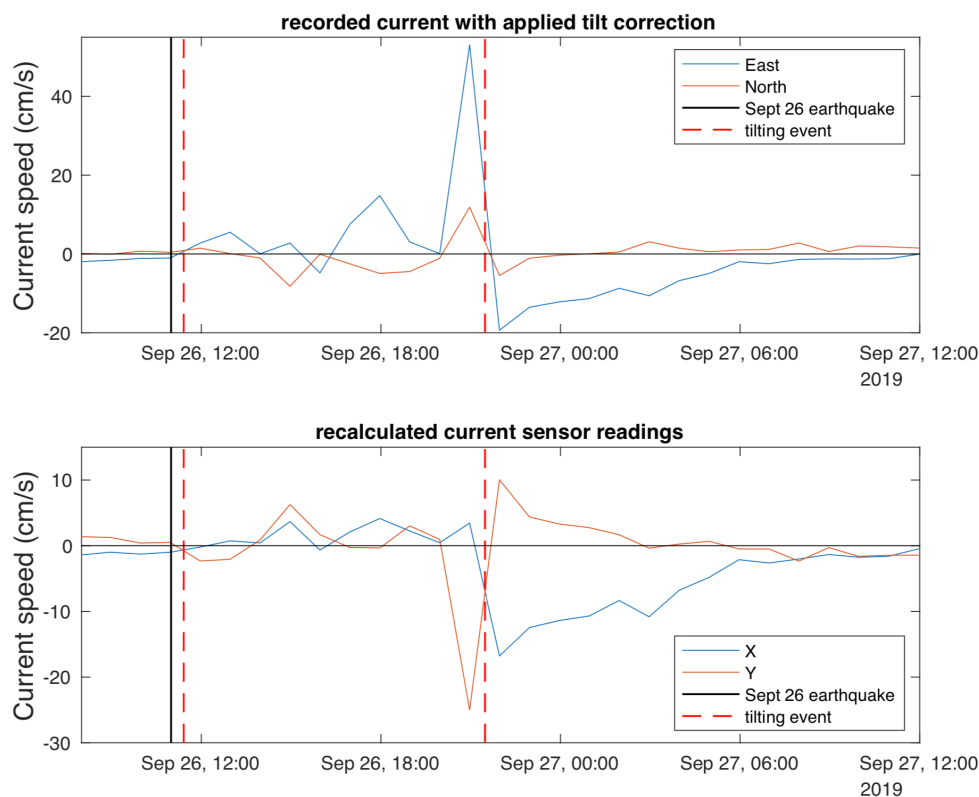


Figure 7. Current record acquired around the time of occurrence of a M_w 5.8 earthquake. A. Instrumental record, automatically corrected for tilt and heading. B. recalculated readings in the X and Y axis of the Doppler sensor (see text for interpretation).

The current speed in the first 2 hours after the main earthquake apparently remained low, at most 5-6 cm/s. It is thus unlikely that the tilt of the device was caused by strong currents. Some short burst of current may have been missed because of the 1 hour sampling interval, but this would not explain why the frame then remained stable in a tilted position for several hours. Local liquefaction of the sediment beneath the device is also an unlikely cause because the tilting of the instrument occurred 25 minutes after the earthquake. A mud flow originating from the basin slopes thus appears as a more likely cause. This hypothesis would also acount for the presence of sandy mud caked on the device in various places: on the frame feet, on the acoustic releasers, on the optode connector and also inside the plastic protection of a flotation

346   sphere, from which bindings were broken and had to be repaired. On the other hand,

347   the current speed in the 20-50 cm/s range recorded before, as well as after, the time

348   when the device straightened up is strong enough to cause erosion of mud or sand

349   deposits. It may thus be hypothesized that erosion freed the device from the mud cover.

350   The flotation spheres on the frame and the concrete ballast at its base exert a moment

351   that should keep the assembly stable in an upright position unless the frame is loaded

352   with sediment.

353

354   3.3. Acoustic backscatter record

355   The strength of the backscattered signal can be used as a proxy for turbidity.

356   The Seaguard RCM emits in the 1.9-2 MHz band corresponding to a wavelength (λ) of

357   750 μm. Doppler backscatter current meters have maximum sensitivity for particles of

358   diameter $D = \lambda/\pi$ and can detect particles down to diameter $D = 0.08\ \lambda$, for which

359   backscatter power is less than 1/10 of peak backscatter power (Guerrero et al., 2011,

360   2012). The seaguard RCM should thus be mostly sensitive to the presence in

361   suspension of sand size particles (larger than 63 μm). The background backscatter

362   amplitude level is -43±1dB before the earthquakes. Three to four hours after the M_w

363   4.7 09/24/2019 earthquake, backscatter increases sharply to -22dB between 11:00

364   and 12:00, and then decays to -41dB in 12 hours. The increase of backscatter

365   coincides with a change of current direction and speed, indicating that the turbid cloud

366   was brought to the instrument site by the current. However, the current speed of less

367   than 4 cm/s may have been insufficient to put the particles in suspension. There is no

368   increase of backscatter on Sept 20 when stronger currents coming from the same

369   direction, but not related with an earthquake, were recorded.

370   Backscatter strengh remain 41±1 dB the 1.5 days interval before the M_w 5.8

371   09/26/1919 and increases to the -20 dB to -13 dB range after the earthquake (Figure

372   5), which implies sand sized sediment was put in suspension soon after the earthquake

373   although the local current speed remained relatively low (about 5 cm/s at most). After

374   the device went back to near vertical position, signal strengh reaches a maximum of -

375   7.6 dB, which correspond to an amplitude ratio of 42 and an intensity ratio of 1800

376   compared to base level. Similar signal strength levels are typically reached with the Z-

377   Pulse sensor in highly turbid water such as in estuaries. During deep sea deployments

378   signal strength range more typically between -60 and -40 dB. After reaching peak value,

379   backscattered signal strength progressively decays to stabilise at about -40 dB 3 days

380   (Figure 5). Several turbid events, with signal strength about -35 dB are observed in

381   October and associated with small increases in current velocity (up to 3-4 cm/s). It is

382   unclear whether these passing clouds are residual turbidity from the earthquake. After

383   October 9, backscatter eventually returns to background level while temperature

384   decreases by 0.007 °C over a few hours, indicating replacement of the water mass

385   around the instrument.

386

387   3.4. Temperature record

388   The Sea of Marmara is stratified, with a low salinity (20-22‰) 20-30 m surface

389   layer that displays strong seasonal temperature variability overlaying a high salinity

390   (about 38‰) body of seawater at 14-15°C derived from the Aegean Sea (Beşiktepe et

391   al., 1994). Within this body, the potential temperature generally decreases with depth,

392   which would in principle imply that a turbidity current, flowing downward, should cause

393   a small temperature increase. However, the deployment site is prone to seasonal

394   cascading, so that the initial temperature structure may have been disturbed. Example

395   of CTD profiles recorded in June 2007 (Henry et al., 2007) are shown in Figure 8. No

396   CTD profile is available in Sept 2019, but variations in temperature and oxygen

397   concentration associated with mild currents (<5 cm/s) were recorded by the instrument

398   in May-July 2019, and again on Sept 20.

399   Temperature variations associated with the M_w 4.7 09/24/2019 earthquakes are

400   very small, less than ±0.002°C, which confirms that water did not travel far during this

401   event. After the M_w 5.8 09/26/1919 earthquake, the recorded temperature decreases

402   progressively by about 0.015°C, after the first hydrologic disturbance and tilting of the

403   instrument, until the recorded current reaches its maximum value (Figure 5).

404   Temperature then progressively increases to reach nearly the same value as before

405   the event. The small variation in temperature recorded indicates that the turbid water

406   originates from the deep-water body. The slight temperature decrease observed after

407   the earthquake can result from the mixing of a warmer bottom water layer with the bulk

408   of the deep-water layer. However, the observation of a temperature drop precludes

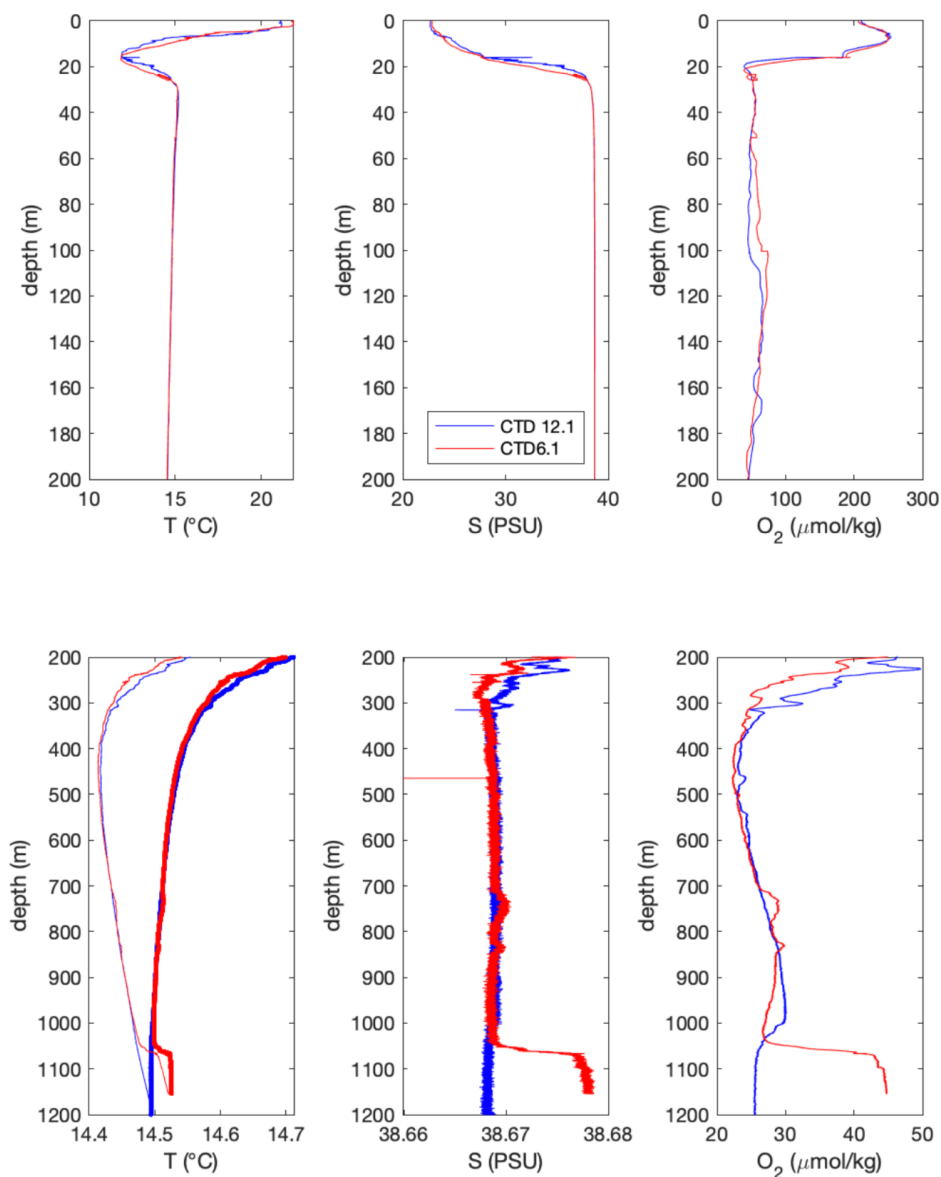
409   that the turbid water originates from depths less than 400 m, as water present between

410   400 m and the halocline is at a higher potential temperature than the deeper water

411   throughout the year (see Beşiktepe et al., 1994, and figure 8). Moreover, an inflow of

412   water from closer to the surface should result in an increase in the O₂ concentration in

413   the bottom water, but none is observed in the data.

414

415

416   **Figure 8.** Depth plots of Temperature (°C), Salinity (PSU) and oxygen concentration

417   (µmol/kg) from CTD profiles acquired in the Sea of Marmara during Marnaut cruise of

418   Ifremer RV L'Atalante (Henry et al., 2007). On the lower temperature plot, thin lines

419   are measured values and thick lines are potential temperatures calculated at 1180 m.
420   Locations are shown on Figure 1
421
422   3.5. Inferred sequence of events
423   These observations provide some insight on the complex sequence of events
424   that followed the earthquake and suggest the following scenario. After the M_w 4.7
425   09/24/2019 a turbid cloud formed east of the instrument and drifted slowly. Considering
426   the maximum velocity of the current (less than 4 cm/s) and the 4-hours interval
427   between the earthquake and the passing of the turbid cloud over the instrument, the
428   front of turbid water should have formed ENE of the instrument at a maximum distance
429   of about 500 m, and this coincides with the base of the northern slope near the outlet
430   of the canyon. It is suspected that instability on the steeper slopes on the sides of the
431   canyon is the primary cause of sediment suspension. The clouds subsequently drifted
432   downslope over a total horizontal distance of at most 1 kilometers before dissipating,
433   adding the 500 m estimate above to the drift calculated after the passing of the front
434   over the instrument (Figure 6).
435   The M_w 5.8 09/26/1919 caused much stronger currents. After the passing of
436   the seismic wave, triggering of instability on slopes adjacent to the deployment site
437   caused mud flows that spread on the basin floor causing the tilting of the instrument
438   25 minutes after the earthquake, and bottom water turbidity. As the base of the nearest
439   slope is about 400 m north of the instrument, this would imply a minimum velocity of
440   20 cm/s for the mudflow to reach the device location in 25 minutes. During the following
441   10 hours, the current record is incomplete but indicates variations in strength and
442   direction. Widespread slope instabilities triggered by the earthquake may have resulted
443   in several turbidity currents recorded as succesive pulses. The role of seiches and
444   surface gravity waves in sediment resuspension can be ruled out as no tsunami was
445   recorded by near shore tidal gauges around the Sea of Marmara. The relationship
446   between gravity wave amplitude A and bottom current amplitude U in the shallow water
447   linear approximation is given by $U=(g/H)^{1/2}A$, where H is water column height. An
448   oscillatory current of 10 cm/s at 1200 m depth would thus correspond to a free surface
449   oscillation of 1 m (or 100 hPa) for a standing wave (seiche) as well as a progressive
450   wave (tsunami). This should have been easily detected in a sea where tidal amplitude
451   is about 10 cm (Alpar and Yüce, 1998). The influence of baroclinic internal waves on

452   the halocline at 20-30 m depth must also be ruled out as they cannot physically
453   produce currents of more than a few cm/s at 1200 m. However, It remains possible
454   that the interface at the top of the turbid cloud is affected by baroclinic waves. The
455   strongest current is recorded after 10 hours, which suggests that a turbidity current
456   initiated further upslope may have reached the site after a longer delay but may also
457   have gained more kinetic energy on its downhill path. This event, reaching a speed
458   exceeding 25 cm/s apparently caused enough erosion to free the device from the mud
459   accumulation. The current then stabilizes in a westward direction and decays
460   progressively over the next 9 hours, which suggests the tail of a turbidity current flowing
461   in the canyon E of the deployment site has been recorded. The hours-long delay
462   between the earthquake and the passing of the fastest current over the instrument may
463   hypothetically correspond to the time for the head of the turbidity current to travel from
464   its source to the location of the instrument. The length of the canyon valley between
465   the device location and the 400 m isobath, inferred to be the minimum depth of the
466   turbid water source, is about 13 km. In this scenario, the average velocity of the head
467   of the turbidity current would be 30-40 cm/s. Alternatively, a sequence of slope failures
468   may have lasted up to several hours after the earthquake. The distance travelled by
469   the turbidity current on the basin floor beyond the instrument may be estimated from
470   the calculated drift during the waing phase, and is found to be about 3.5 km (Figure 6).
471   When plotted over the bathymetric map the drift appears to stay within the depositional
472   fan at the outlet of the cayon, the extension of which is known from sediment sounder
473   profiles (Figure 1). These calculations are only a rough estimate of the distance
474   travelled by suspended particles as only the velocity at 1.5 m above the seafloor is
475   known. However, it appears unlikely that sediments spread all over the 15x20 km basin
476   floor as this would require velocities of the order of 1m/s, sustained over a wide area
477   for several hours.

478   The decay of the backscatter signal strength over the next 3 days may reflect
479   the settling of sand size particles put in suspension in the water column after this
480   sequence of events. For a first order assessment, Stokes settling velocity, an upper
481   bound valid in dilute suspensions (e.g. Guazelli and Morris, 2012) may be used. The
482   Stokes settling velocity of 63 µm quartz grains (density 2650 kg/m^3) in 13°C seawater
483   is 2.7 mm/s, allowing such grains to drop by 700 m in 3 days. However, if the particles
484   forming the cloud are mostly composed of clay agregates, which density may be
485   comprised between 1200 and 1700 kg/m^3, the settling velocity would be comprised

486   between 0.3 mm and 1 mm/s. In this case the height of the suspended particle cloud

487   could range between 70 and 250 m.

488   ## 4. Discussion and conclusion

489   Data obtained with a seafloor device located at the oultet of a canyon in the

490   Central Basin in the Sea of Marmara bring some insight on how earthquakes scale

491   with their hydrodynamic consequences. In September 2019, M_w 4.7 and 5.8

492   earthquakes occurred at a 5 km distance from the device as well as a series of smaller

493   foreshocks and aftershocks In this setting, earthquakes of magnitude less than 4 did

494   not cause noticeable water column turbidity nor currents. The M_w 4.7 earthquake

495   generated a turbid cloud on slopes a few hundred meters from the instrument and the

496   cloud took 3-4 hours to drift down to the instrument location and 10 more hours to

497   dissipate. As the current velocity remained small (less than 4 cm/s), it can be concluded

498   that this cloud did not evolve into a self-sustained turbidity current (Parker, 1982). The

499   M_w 5.8 earthquake initiated a turbidity current and the data obtained may be compared

500   with more complete records of turbidity currents obtained elsewhere with ADCP

501   deployments and/or water column mooring lines. The duration of the event in the Sea

502   of Marmara (about 10 hours) appear fairly typical and comparable with events recorded

503   in other locations regardless of the initiation mechanism, which comprise hyperpycnal

504   flows from river floods (Var and Gaoping canyons), storm waves and dredging (Gulf of

505   St Laurence and Monterey canyon), and slope instabilities triggered by an earthquake

506   (Tokachi-oki) or by other processes (Var canyon) (Normandeau et al., 2019;

507   Khripounoff et al., 2012; Xu et al., 2004; Liu et al., 2012; Mikada et al., 2006). Longer

508   duration events with very different hydrodynamic characteristics have been observed

509   in larger scale systems (e.g. Congo deep sea canyon, Azpiroz-Zabala et al., 2017). On

510   the other hand, events recorded closer to shore on the edge of the continental shelf or

511   on a delta front have much shorter durations (Xu et al., 2010; Hughes Clarke, 2016).

512   In events of comparable scale to the Sea of Marmara one, the velocity of the current

513   generally reaches its maximum several meters above the seafloor, so that the velocity

514   recorded by our instrument at 1.5 m from the seafloor is within the boundary layer, and

515   lower than either the maximum current velocity or the velocity of the head of the

516   turbidity current. A velocity of several tens of centimeter per second is representative

517   of the slower recorded examples, corresponding to mud rich flows associated with

518   hyperpycnal flows, to the smaller landslides (Khripounoff et al. 2012) or to the smaller

519   storm-related events (Normandeau et al., 2019). Turbidites following large
520   earthquakes or large slope instabilities have reached maximum velocities of 20 m/s
521   (Piper et al., 1999). Velocities of 2-7 m/s were reported for the turbidity current following
522   Tohoku earthquake (Arai et al., 2013) and 1.4 m/s in the Tokachi-Oki case (Mikada et
523   al., 2006). The downward current after the off-Izu Penninsula earthquake may be
524   constrained with a noisy ADCP record to a maximum of 10-15 cm/s in a 20-30 m layer
525   above the seafloor and lasted about one hour, peaking about 30 minutes after the
526   earthquake (Kasaya et al., 2009). This turbidity current thus appears less intense and
527   shorter in duration than the one recorded in the Sea of Marmara, but the triggering
528   earthquake was also smaller (M5.4 compared to M5.8) and more distant (10 km).
529   These observations suggest that a general scaling relationship could exist between
530   earthquake magnitudes and the strength and extent of the turbidity currents they
531   induce. Moreover, although the off-Izu event is shorter than the Sea of Marmara one,
532   they share an important characteristic in that the turbid cloud is observed to form when
533   a mud flow hits the observatory site and some time before current builds up in the
534   water column.

535   The 10 hours delay observed in the Sea of Marmara between the triggering
536   event and the peak of the turbidity current is long compared to the 2 hours delay
537   observed after Tohoku and Tokachi-oki earthquakes. We suggested earlier that the
538   long delay may simply result from a slower velocity of the turbidity current or from
539   delayed slope failure. Another possibility is delayed ignition, which may occur if the
540   turbidity current develops indirectly from the hydrodynamic instability of a turbid cloud
541   resulting from slope failures and/or ground shaking rather than by acceleration of a
542   dense mud flow (Parker, 1982; Mulder and Cochonnat, 1996; Piper and Normark,
543   2009).

544   The scenario we propose for the Sept 26, 2019 earthquake involving mud flows
545   from proximal sources, followed by turbidity currents originating at larger distances,
546   and the subsequent settling of sediment in suspension, could relate with the structure
547   of turbidite-homogenites. Progressive or pulsed build-up of turbidity current energy is
548   considered typical of hyperpycnal flows initiated by river floods (Mulder et al., 2003)
549   but reverse grading and pulsing is also observed in seismoturbidites (Gutierrez-Pastor
550   et al., 2013). In the Sea of Marmara, many of the laminated turbidites sampled in
551   Kumburgaz Basin formed from the amalgamation (below the homogenite layer) of at
552   least two flows, the first one being finer and less sorted (Yakupoğlu et al., 2019). The

553   coarsening observed in this context is often associated with an increase of the calcium
554   content indicative of a shallower source, rich in biogenic carbonate material. However,
555   in the case observed in the present study, remobilization of sediment should be limited
556   to the lower slope as the temperature of the displaced water precludes a source
557   shallower than 400 m water depth.

558   The geomorphological context of the deployment site south of a slope identified
559   as unstable from geomorphological criteria (Zitter et al., 2012), and on a depositional
560   fan at the outlet of a canyon is also consistent with the proposed scenario. We
561   estimated by integrating recorded current velocity that the current during this event
562   was not strong enough to spread the sediment over the entire Central Basin floor but
563   that the zone of deposition was probably comparable in size to the fan. It can be
564   inferred that the paleoseismological record from a core taken in the fan should contain
565   more events than one taken at the basin depocenter. A sediment sounder profile
566   (Figure 1) also shows that the character of the seismic reflectors differs in the basin
567   and in the fan and that establishing reliable correlations between them is not simple.
568   However, as hypothesized by previous studies (McHugh et al., 2014), turbidite-
569   homogenite deposits that can be correlated between cores taken at various locations
570   in the basin probably correspond to the larger events, which may be correlated with
571   historical earthquakes. Moreover, it is still unknown whether the Sept 26 event left a
572   trace on the seafloor morphology and in the sediment record. Performing new core
573   sampling and very high-resolution geophysical surveys in this area would thus have
574   important implications for the understanding of seismoturbidite records and for the
575   assessment of geohazards.

576

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Seafloor monitoring data are available through SEANOE (Henry et al., 2021) and
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