# Mass flows, turbidity currents and other hydrodynamic consequences of small and moderate earthquakes in the Sea of Marmara

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**Abstract.** Earthquake-induced submarine slope destabilization is known to cause mass wasting and turbidity currents, but the hydrodynamic processes associated with these events remain poorly understood. Instrumental records are rare and this notably limits our ability to interpret marine paleoseismological sedimentary records. An instrumented frame comprising a pressure recorder and a Doppler recording current meter deployed at the seafloor in the Sea of Marmara Central Basin recorded the consequences of a $M_W$ = 5.8 earthquake occurring Sept 26, 2019 and of a $M_w$ = 4.7 foreshock two days before. The smaller event caused sediment resuspension and weak current (< 4 cm/s) in the water column. The larger event triggered a complex response involving a debris flow and turbidity currents with variable velocities and orientations, which may result from multiple slope failures. A long delay

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of 10 hours is observed between the earthquake and the passing of the strongest turbidity current. The distance travelled by the sediment particles during the event is estimated to several kilometres, which could account for a local deposit on a sediment fan at the outlet of a canyon (where the instrument was located), but probably not for the covering of the whole basin floor. We show that after a moderate earthquake, delayed turbidity current initiation may occur, possibly by ignition of a cloud of resuspended sediment.

## 1. Introduction

Triggering of mass wasting and turbidity currents by earthquakes is a hazard that can damage seafloor infrastructure (Heezen et al., 1954) and may enhance co-seismic tsunami generation (Okal and Synolakis, 2001; Synolakis et al., 2002; Hebert et al., 2005; Ozeren et al., 2010). Earthquake-triggered canyon flushing is also a primary driver of submarine canyon development and material transfer from seismically active continental margins to the deep ocean (Mountjoy et al., 2018). It is often considered that a peak ground acceleration (PGA) of the order of 0.1 g is needed for an earthquake to trigger a submarine slope failure (Dan et al., 2008; Nakajima and Kanai, 2000). A peak ground velocity threshold of 16-25 cm/s for turbidity-current-triggering has been proposed based on observations after 14 November 2016 $M_w$ 7.8 Kaikoura, New Zealand, Earthquake (Howarth et al., 2021). The corresponding peak ground acceleration cannot be accurately determined because the seismic waveform in this study was modelled at long periods (> 2 s). Nevertheless, strong motion records from this earthquake suggest this peak ground velocity threshold does correspond with a peak ground acceleration of the order of 0.1 g (Bradley et al., 2017). On the other hand, a global compilation of cable breaks shows that mass flows have been triggered by individual earthquakes of $M_w$ as low as 3.1 (with esdtimated PGA ≈10$^{-3}$ g) while on other margins where sediment input is relatively low and/or earthquakes frequent, earthquakes >7 $M_w$ failed to trigger cable breaking flows (Pope et al., 2016). In the Mediterranean region, the threshold is reportedly around $M_w$ = 5.

In spite of this high regional variability, turbidite deposits in several seismically active zones have been used as paleoseimological event markers (e.g.: Adams, 1990; Goldfinger et al., 2003, 2012; McHugh et al., 2014; Ikehara et al., 2016; Polonia et al., 2016). For instance, Holocene turbidite records in the Sea of Marmara basins display a recurrence of 200 to 300 years, that roughly corresponds to the recurrence interval

91  of Mw > 6.8 earthquakes (McHugh et al., 2006, 2014; Drab et al., 2012, 2015;
92  Yakupoğlu et al., 2018). Synchronicity of turbidites over a large area is considered as
93  the most robust criterion for recognizing sedimentary events caused by large
94  earthquake ruptures, although this approach has caveats (Talling, 2021; Atwater et al.,
95  2014). Distinguishing seismoturbidites, caused by earthquakes and related mass
96  wasting events, and turbidites resulting from other processes (e.g. floods, storms,
97  sediment loading) from their sedimentological characteristics is particularly challenging
98  (Talling, 2021; Heerema et al., 2022). Seismoturbidites generally comprise a basal silt-
99  sand bearing layer under a layer of apparently homogenous mud (named homogenite
100 or tail) with small or gradual, if any, variations in grain size and chemical composition
101 (Polonia et al., 2017; McHugh et al., 2011; Çağatay et al. 2012; Eriş et al., 2012;
102 Gutierrez-Pastor et al., 2013; Nakajima and Kanai, 2000; Beck et al., 2007). The grain
103 size break between turbidite and homogenite layers is however not specific to
104 seismoturbidites and can result from mud settling processes commonly occurring in
105 turbidity currents (e.g.: Talling et al., 2012). In lakes and closed basins several other
106 characteristics of turbidite-homogenites, such as the alternation of silt/sand and mud
107 laminae within a single turbidite unit and presence of bi-directional cross- or flaser-
108 bedding have been interpreted as indicators of deposition from oscillatory currents
109 associated with seiches or turbidity current reflection (Beck et al., 2007; Çağatay et al.
110 2012; McHugh et al., 2011). Indeed, internal tsunami waves and turbidity current
111 reflection have been recorded after landslides in lakes (Brizuela et al., 2019). However,
112 seismoturbidites on ocean margins have fairly similar characteristics to those in closed
113 basins but their layering has been interpreted differently, as a consequence of
114 confluence (stacked or amalgamated turbidites) or current speed variations (multi-
115 pulsed turbidites) (Gutierrez-Pastor et al., 2013; Nakajima and Kanai, 2000; Goldfinger
116 et al., 2003). There is currently a lack of in situ instrumental records that could
117 substantiate inferred hydrodynamic processes.
118     Monitoring experiments have generated observations of turbidity currents
119 flowing in submarine canyons and initiated by meteorological events, seasonal
120 discharge from rivers and occasionally by landslides (Xu et al., 2004, 2010; Puig et al.,
121 2004; Palanques et al., 2008; Liu et al., 2012; Khripounoff et al., 2012; Hughes Clarke,
122 2016; Gwyn Lintern et al., 2016; Azpiroz-Zabala et al., 2017; Paull et al., 2018; Hage
123 et al., 2019; Normandeau et al., 2020; Heerema et al., 2022). Some turbidity currents
124 originating from sediment remobilization events are driven by a thick dense basal layer,

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a supprimé: Seismoturbidites are generally described as turbidite-homogenites where a basal silt-sand bearing layer is overlaid by a layer of apparently homogenous mud with small or gradual, if any, variations in grain size and chemical composition

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a supprimé: Holocene seismoturbidite records in the Sea of Marmara basins display a recurrence of 200 to 300 years, that roughly corresponds to the recurrence interval of Mw>7 earthquakes (McHugh et al., 2006, 2014; Drab et al., 2012, 2015; Yakupoğlu et al., 2018; Bulut et al., 2019). In lakes and closed basins several characteristics of deposits following earthquake or landslides, such as …t
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a supprimé: Monitoring experiments brought records of representative cases of turbidity currents flowing in submarine canyons and initiated by meteorological events and occasionally by landslides

a supprimé: (Azpiroz-Zabala et al., 2017; Khripounoff et al., 2012; …

a supprimé: . Oscillatory currents resulting from internal waves have been recorded after landslides in lakes (Brizuela et al., 2019).

165 able to displace and bury heavy instruments (Paull et al., 2018). On the other hand, progressive or pulsed build-up of turbidity current energy is considered typical of hyperpycnal flows initiated by river floods (Mulder et al., 2003; Khripounoff et al., 2012). However, the hydrodynamic characteristics of turbidity currents resulting from landslides and floods may not systematically differ, especially when observations are done at a distance from the source (Heerema et al., 2022). Most information on earthquake-triggered events is still indirect based on cable ruptures (e.g.: Gavey et al., 2017; Pope et al., 2017; Hsu et al., 2008), geomorphological and sedimentological observations (Mountjoy et al., 2018; Cattaneo et al., 2012; Piper et al., 1999), and information from displaced instruments (Garfield et al., 1994). In Japan, in situ records of pressure and temperature were obtained from displaced Ocean Bottom Seismometers (OBSs) after the Tohoku 2011 $M_w$ 9.1 earthquake (Arai et., 2013), and from cabled observatories after the Tokachi-Oki 2003 $M_w$ 8.3 earthquake (Mikada et al., 2006) and after a moderate (M 5.4) earthquake off Izu Penninsula (Kasaya et al., 2009). After the large events, strong bottom currents of more than 1 m/s were implied, generally starting 2-3 hours after the earthquake, with no indication of oscillation or pulsing. In the off-Izu case a mudflow was observed with a camera 5 minutes after the earthquake and followed 15 minutes later by a change in current direction and speed.

183 We here present results from an instrumental deployment at the seafloor that accidentally recorded the consequences of earthquakes that occurred 09/24/2019 and 09/26/2019 in the Sea of Marmara with respective $M_w$ 4.7 and 5.8 (Figure 1A). The pressure, temperature and current record from this single instrument demonstrate that both events caused sediment resuspension in turbid clouds, but only the larger event triggered turbidity currents. However, the instrument suffered a rather complex sequence of disturbances and a 10 hour delay is observed between the earthquake and peak current recording. Here, we propose a scenario which could explain the observations and discuss their implications for the understanding of seismoturbidite records.

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**Figure 1.** Context of instrumental deployment. (**A**) bathymetric map of the Sea of
Marmara Central Basin with simplified fault geometry (in red). The hatched zone is a
suspected mass wasting zone (Zitter et al., 2012). Location of instrumented frame
comprising bottom pressure recorder (BPR) and doppler current meter is indicated by
blue square. The blue banana with white dots represents the calculated trajectory of
a sedimentary particle during the waning phase of the turbidity current. Red dots are
CTD profiles 6 and 12 shown in supplementary material S1. Epicenter location of
earthquakes and the focal mechanism of the main shock are indicated. (**B**) Location
of study area. North Anatolian Fault system is shown in red. MMF is the Main

223 Marmara Fault. (**C**) Sediment sounder profile from Marmarascarps cruise (Armijo and
224 Malavieille, 2002). Indicative age of reflector from Beck et al. (2007). The instrument
225 (BPR) was deployed on a depositional fan at the base of slope and canyon outlet that
226 differ in seismic character from the reflector sequence in the basin.

### 2. Context and data collection

228 A series of instrumental deployments was planned to record naturally occurring
229 resonant water column oscillations (seiches) at various locations in the Sea of
230 Marmara with the aim to improve tsunami models (Henry et al., 2021). An instrumented
231 frame was thus deployed at 40.8568° N, 28.1523° E and 1184 m water depth in the
232 Central Basin on May 9, 2019 and recovered six months later (11/19/2019) (Figure 1A).
233 This site is located at the outlet of a complex canyon system with multiple confluence
234 points and tributaries originating from the edge of the continental shelf (Figure 1).
235 Sediment sounder profiles indicate a depositional fan or lobe is present at this location
236 (Figure 1C). Canyons observed on the relatively steep sedimented slope (≈10°) of the
237 Sea of Marmara deep basins are presumably fed by mass flows sourced from the
238 canyon heads and walls (Zitter et al., 2012; Çağatay et al., 2015). In addition, the slope
239 west of the canyons immediately north of the deployment site hosts a landslide
240 covering about 24 km$^2$ and cores taken at the base of the slope contain a sandy debris
241 flow deposit of 35-40 cm thickness buried 2 m below the seafloor (Zitter et al., 2012).
242 The Main Marmara Fault (MMF, Figure 1B), is defined as the part of the northern
243 branch of the North Anatolian Fault system crossing the Sea of Marmara (Le Pichon
244 et al., 2001, 2003). A splay of the MMF runs along the base of this slope (Armijo et al.,
245 2002; Grall et al., 2012; Sengor et al., 2014). The 09/24/2019 and 09/26/2019
246 earthquakes occurred under the canyon system and their epicenters are located 5 km
247 ENE of the instrument, less that 500 m apart (Figure 1). The rupture occurred within
248 the crust at 9-13 km depth on a northward dipping fault located north of the principal
249 displacement zone of the Main Marmara Fault. The focal mechanism indicate right-
250 lateral strike-slip with a reverse component (Karabulut et al., 2021). The rupture did
251 not reach the seafloor, nor caused a tsunami. For instance, tidal gauge records
252 obtained at Marmara Ereglisi do not deviate more than 1 hPa from a fitted tidal model.
253 The instrumentation on the frame comprises (1) an RBR bottom pressure recorder
254 (BPR) with a Paroscientific 0-2000 m Digiquartz pressure and temperature sensor, (2)
255 a Seaguard recording current meter (RCM) equipped with a Z-pulse 4520 Doppler

268 current sensor operating in the 1.9-2 MHz frequency range and other sensors:
269 temperature, pressure (tide sensor Aanderaa 5217), conductivity (Aanderaa 4319),
270 oxygen (Aanderaa optode 4330) (Figure 2). The RBR pressure and temperature
271 recording interval was set to 5s and that of the Seaguard RCM to one hour for all
272 sensors. The Seaguard instrument was fixed on the upper part of the frame and
273 sensors were 1.5 m above the seafloor. The Z-pulse Doppler current sensor is a single-
274 point current sensor, not an acoustic Doppler profiler (ADCP). It emits four narrow (2°)
275 beams paired in opposite directions along two orthogonal axes in a plane (parallel to
276 the seafloor if the frame is standing upright), and measures Doppler backscatter in
277 cells extending 0.5-to-2 m from the instrument (Figure 3). The Doppler current sensor
278 was set in burst mode, averaging 150 pings taken every second at the end of each
279 one-hour recording interval and in forward ping mode, so that only data from sensors
280 measuring a positive Doppler shift, upstream currents moving toward the instrument,
281 are used to calculate current speed. The tide sensor is a piezoresistive sensor with a
282 specified accuracy comparable to that of the Digiquartz sensors (4 kPa for a 0-2000 m
283 sensor vs. 2 kPa for a Digiquartz sensor with the same range) and 0.2 hPa (2 mm
284 ocean depth) resolution and comprises a temperature sensor of 0.2°C accuracy and
285 0.001°C resolution. The tide sensor averages pressure measured at a 2 Hz sampling
286 rate over 300 s at the end of each one-hour time interval. The tide sensor was checked
287 against an atmospheric reference between deployments and found to have a minimal
288 drift, less than 1 hPa. The main purpose of the pressure sensor records was to detect
289 long period variations in water height, related for instance to tides and seiche
290 oscillations but they are also sensitive to pressure variations caused by P-waves. In
291 addition, Digiquartz sensors are intrinsically sensitive to acceleration, but to a small
292 extent, 160 hPa/g for an instrument with 20 MPa range according to the calibration
293 report.
294    As we will show that the 09/24/2019 earthquake caused the instrumented device
295 to lay on its side for several hours and then straighten up, understanding the setup of
296 the seafloor device and its stability is important (Figure 2B). The frame is made of
297 aluminium and has six rigidly bound flotation spheres of 25 kg buoyancy each. The net
298 weight of the instrumented frame in water is -80 kg. The frame is rigidly attached to a
299 12-cm-thick 1.5x1.3 m concrete slab, weighing 300 kg in water. The assembly of the
300 heavy slab and buoyant frame is stable in an upright position in the water and on the
301 seafloor. Moreover, it is estimated that a current of 1 m/s would cause a total horizontal

325 drag of 75 kg (≈750 daN) when the device is in upright position, which is insufficient to

326 destabilize it. If a stronger current, or other external forces, cause the assembly to tilt

327 and lay on one side, the moment of the gravity and buoyancy forces should straighten

328 the device back to upright position when these external forces are removed.

329 Measurement of current speed and direction by a tilted instrument is a related issue

330 that we here consider. The orientation and attitude of the Seaguard RCM is measured

331 with a 2-component accelerometer and a magnetic compass and the recorded data

332 include tilt in X and Y direction and the heading of the X axis. Tilt X and Y components

333 are factory calibrated from -35° to +35° with an accuracy of 1.5°. Tests performed in

334 the laboratory (see supplementary material, Figure S1) showed that tilt information

335 remains consistent outside this range, even when the instrument is upside down. Tilt

336 measurements are accurate within 3° up to 60° but saturate at about 80° (Figure S2).

337 Uncertainty on heading also increases with tilt, especially when the instrument is tilted

338 toward the X-direction. However, measured heading remains ±20° of true heading for

339 a tilting of up to 60° (Figure S3). The current measured in the instrument plane is

340 corrected for tilt assuming current is horizontal. As far as this approximation is valid,

341 the current record should in principle remain fairly accurate when the instrument is

342 tilted beyond the normal range of operation (±35° degree) and at least to 60°. However,

343 the compass was not calibrated for an upside-down configuration. If the top of the

344 instrument would happen to be oriented downward, the measured current direction will

345 be unreliable, even though the absolute speed may still be correctly estimated. Another

346 problem may arise if one of the Doppler sensors is facing down into the sediment so

347 that its measurement cell is below the seafloor. If the sensor pointing upward in the

348 opposite direction is recording a negative Doppler shift, this value will be ignored in the

349 forward ping mode. In this case, the measurement retained to calculate current velocity

350 will correspond to noise from the sensor facing toward the seafloor. In all situations, it

351 remains possible to recalculate the sensor readings retained by the calculator from the

352 current velocity and orientation parameters recorded by the instrument by projecting

353 the velocity vector back in the instrument plane, and thus assess the reliability of data.

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

355 Z-pulse emits in the 1.9-2 MHz band corresponding to a wavelength ($\lambda$) of 750 µm.

356 Doppler backscatter current meters have maximum sensitivity for particles of diameter

357 $D = \lambda/\pi$ and can detect particles down to diameter $D = 0.08 \lambda$, for which backscatter

358 power is less than 1/10 of peak backscatter power (Guerrero et al., 2011, 2012). The

360 seaguard RCM should thus be mostly sensitive to the presence in suspension of larger

361 than 63 µm. This, however, does not imply that the detected particles are all sand

362 grains in the mineralogical sense, as clay flocs of the same size also cause

363 backscattering,

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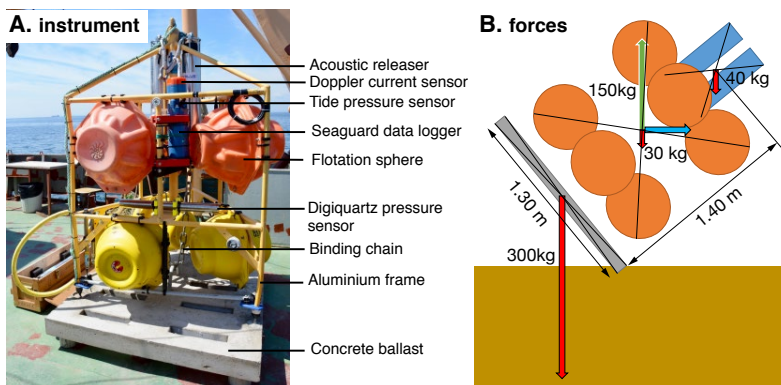
366 **Figure 2.** Instrumented frame. (A) photo of the instrumented frame before

367 deployment. (B) Sketch showing forces applied to the elements of the instrumented

368 frame in water. The red arrows represent the weight in water of the cement ballast, of

369 the instrumented frame and of the acoustic release system on top. The green arrow

370 represents the buoyancy of the flotation spheres. The blue arrow represents the

371 current drag, which depends on current speed and instrument tilt.

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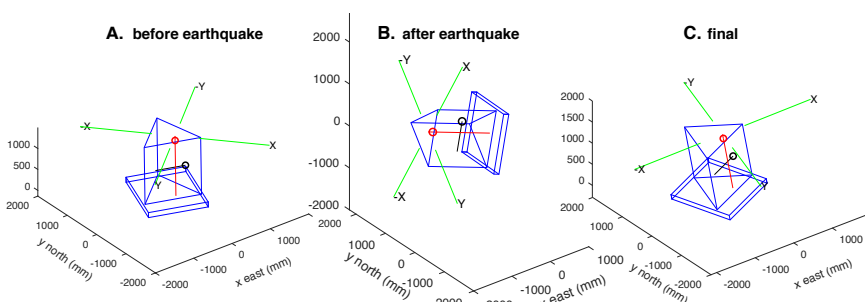
375 **Figure 3.** Reconstruction of frame position based on instrument tilt-meter and

376 compass data: (A) before the earthquake; (B) Tilted, between, 25 minutes and 10.5

377 hours after earthquake; (C) back in nearly upright position 11 hours after earthquake.

380    Position of Digiquartz pressure sensor (black circle), Aanderaa tide sensor (red

381    circle) and Doppler current meter beam cells (green segments)

382

### 3. Results

#### 3.1. Pressure and tilt records

385    Small earthquakes are detected as pressure spikes, while oscillations are

386    recorded after large earthquakes. The $M_w$ 4.7 09/24/2019 caused a short pressure

387    transient of 25 hPa at 08:00:26 followed by small pressure oscillations of less than 3

388    hPa amplitude decaying over a few minutes. The seismic wave train from the $M_w$ 5.8,

389    09/26/1919 earthquake is recorded by the Digiquartz pressure sensor as oscillations,

390    initiated by a pressure drop of 65 hPa between 10:59:22 and 10:59:26 (Figure 4). For

391    the sampling interval of 5s used in this setup, the recorded signal is aliased, which

392    precludes quantitative interpretation in term of velocity or acceleration. However, the

393    initial pressure drop after the 09/26/1919 earthquake may indicate a negative polarity

394    of the first P-wave arrival at the instrument site, located on an ascending ray-path.

395    Twenty-five minutes after the $M_w$ 5.8 earthquake, a new disturbance of the

396    pressure sensor is observed at 11:23:41. The pressure then progressively increases

397    by 30.9 hPa in 15 seconds between 11:24:46 and 11:25:01 before stabilizing. Over the

398    corresponding one-hour-time-interval between successive records, the Seaguard

399    RCM, initially subvertical (tilt less than 2°), acquires a strong tilt (Figure 3). At 11:57:48,

400    measured tilt is -65° along the X-axis and +19° along the Y-axis, with X-axis in a N161°

401    azimuth and these values remain constant ±2° over the next 10 hours, corresponding

402    to an absolute tilt of 68° (Figure 4). The tilting of the instrument causes the Digiquartz

403    and Tide sensors to record different pressure variations because they are located at

404    different positions on the frame (Figure 2). Moreover, the pressure readings by the

405    Digiquartz sensor also depend on its orientation relative to Earth gravity. Pressure at

406    the Tide sensor location increases about 100 kPa, corresponding to a 1 m drop and

407    indicating that the frame was then practically laying on its side. Ten hours later, the

408    device apparently straightens itself in about 5 seconds, between 21:28:29 and

409    21:28:34 as indicated by a rapid pressure variation. After that, the recorded tilt

410    parameters are moderate and stabilize at -11.5° for the X-axis and 5.3° for the Y-axis,

411    with X-axis in a N105.3° azimuth.

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**a déplacé vers le haut [3]:** 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

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421        Baseline changes before and after the earthquake correspond to an increase of
422  23 hPa for the Digiquartz sensor and 20 hPa for the Tide sensor. These concur that
423  the instrumented frame was about 20 cm deeper after returning to upright position.
424  Considering that the slope at the location of the instrument is about 1%, this may
425  correspond to a 20 m lateral downslope displacement. However, in the absence of
426  other information, it is not known whether the pressure baseline change is a
427  consequence of instrument lateral displacement or burial in place.

428        The $M_w$ 4.7 earthquake caused minor disturbances of the attitude of the
429  instrument, with variations of tilt and heading of less than 0.5°. A $M_w$ 3.6 foreshock of
430  the $M_w$ 5.8 occuring 26/09/2019 at 7:32 also caused minor disturbances. These
431  indicate that the seafloor was sensitive to ground shaking caused by these small
432  earthquakes. However, this did not cause the device to sink into the sediment.
433  Changes of the pressure baseline of the digiquartz sensor between before and after
434  these earthquakes are difficult to resolve, and correspond to less than 5 mm vertical
435  displacement for the first event and less than 2 mm for the second one.

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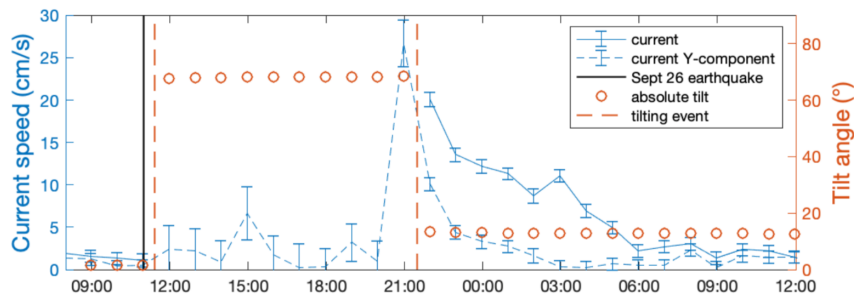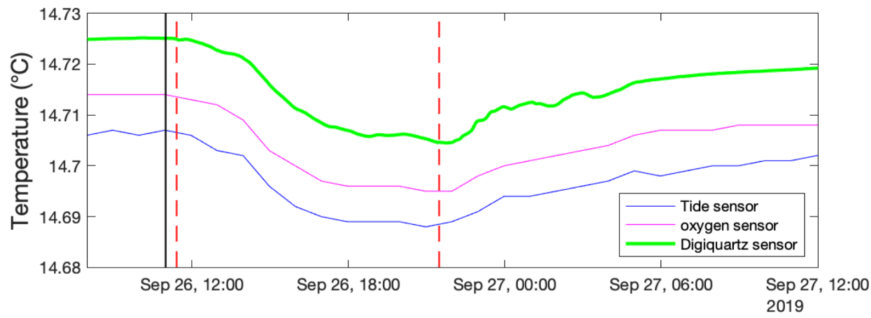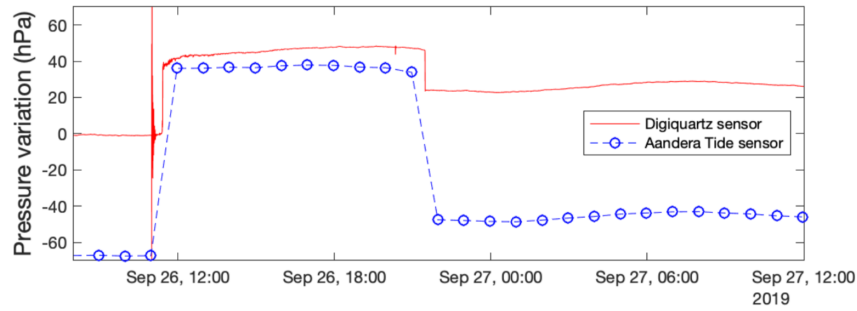
**Figure 4.** Time series around the time of occurrence of a M<sub>w</sub> 5.8 earthquake; (top) pressure variations recorded by two instruments on the instrumented frame; (middle) temperature records from Digiquartz, Tide and oxygen sensors, (bottom) current and tilt data recorded by Seaguard RCM. Between the tilting events only one component of the doppler current meter functioned reliably (Y-component oriented N200) and is here reported.

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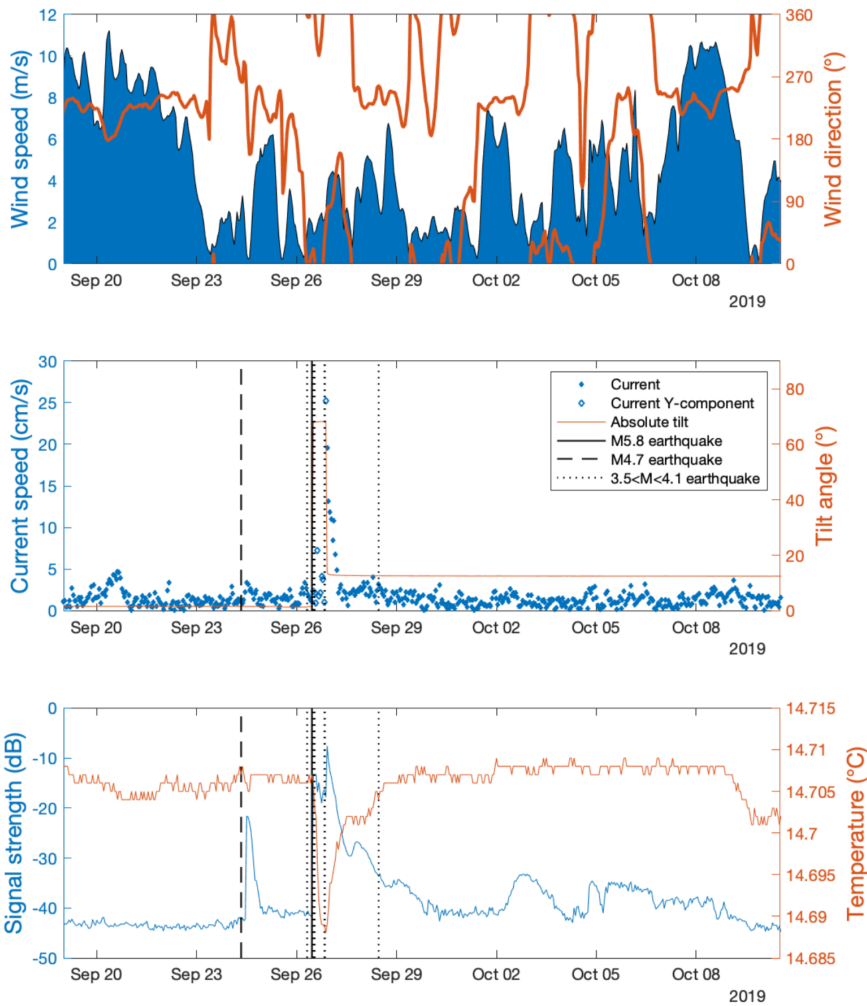
**Figure 5.** Time series acquired with Seaguard RCM during the September 2019
seismicity cluster and ERA5 reanalysed meteorological data (Hersbach et al., 2018);
(top) ERA5 wind data (middle) current speed and tilt; (bottom) backscatter signal
strength and temperature.

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### 3.2. Current records

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, four 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 (with currents up to 3.3 cm/s) just before the $M_w$ 5.8 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 $M_w$ 4.7 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 of 3.4 cm/s (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 the motion of a particle assuming a uniform velocity field (Figure 6). The total drift occurring in the 8 hours following the current inversion is about 500 m. Current direction varies from westward to northward during this time interval.
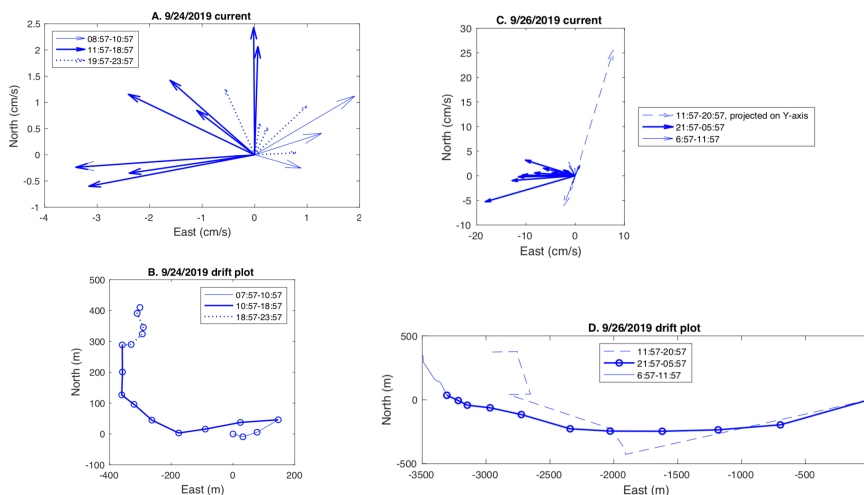
**Figure 6.** Current recorded after $M_w$ 4.7 and $M_w$ 5.8 earthquakes: (A) Current velocity arrows recorded every hour between 08:57 and 23:57 on 09/24/2019; (B) drift plot over the same time interval, the change of current direction and strength between

486    10:57 and 11:57 coincides with increasing backscatter strength (see figure 4),

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

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

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

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

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

492    strongly tilted position.

493

494        After the $M_w$ 5.8 09/26/1919 earthquake, during the 10 hour period when the

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

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

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

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

499    and the measurement cell remained above the bottom (Figure 2B). On the other hand,

500    the X-component may not be reliable as one of the sensors (n°1) is oriented 65°

501    upward in the N160° direction, and the oposite sensor (n°3) is dipping 65° downward

502    in the opposite (N340°) direction. Consequently, measurement cell n°3 lies within the

503    sediment and thus may only record noise. Moreover, because the Doppler current

504    sensor (DCS) is set in forward pinging mode, current speed is calculated with data

505    from sensors measuring positive doppler shifts only. This implies that if the current

506    component toward N160° is positive, sensor n°1 will measure a negative shift and will

507    not be recorded. During the time inteval considered here, the mesured current

508    component in the X-direction (toward N160) is positive, which indicates that data from

509    sensor n°3 was used (Figure 7), and that is probably noise.  It follows that the current

510    component along the Y-direction is the only one reliable. The horizontal current

511    measured along the Y-axis changed sign several times during this time interval, and

512    reached peak values of 6.3 cm/s toward N200 at 14:57:46, about four hours after the

513    earthquake, and of 25 cm/s in the oposite direction at 20:57:46, the last measurement

514    before the instrument straightenned up. Other measurements on both axes remain

515    below 5 cm/s, but the absolute velocity may have been higher because this

516    measurement was only performed in one direction. Yet, these observartions suggest

517    that the stronger current (25 cm/s) recorded 30 minutes before the instrument

518    straightenned up played a role in this event. Once the device got back in an upright

519    position, it recorded a current consistently flowing westward and progressively

decreasing from 20 cm/s to background level (2 cm/s) in nine hours (Figure 4). During

this waning phase, the current drift is about 3.5 km in a westward direction (Figure 6).

The drift estimated during the first 10 hours after the earthquake, while the instrument

was strongly tilted, is in the opposite direction but may not be reliable.
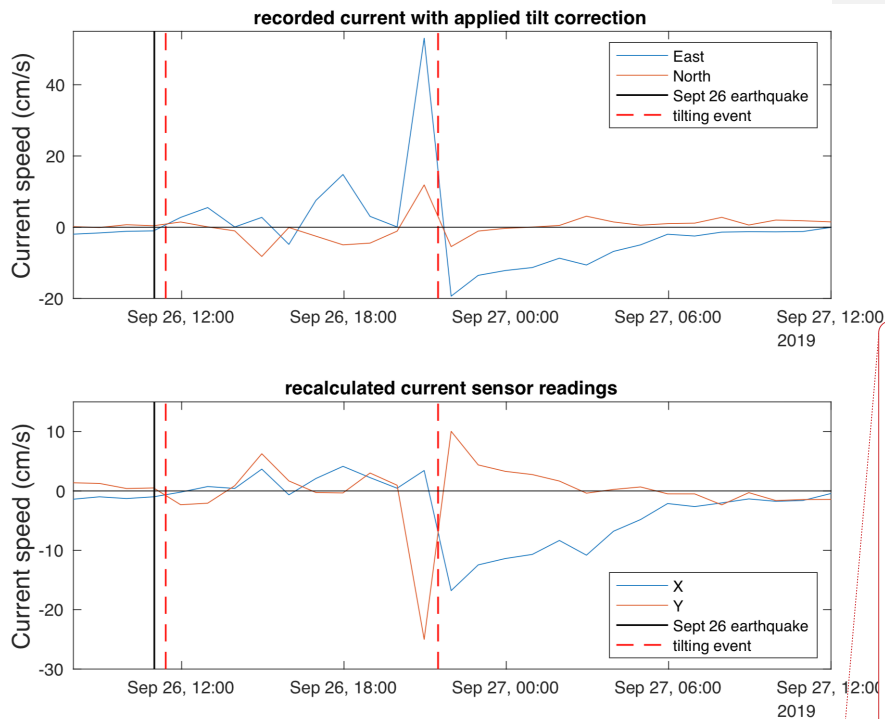
**Figure 7.** Current record acquired around the time of occurrence of a $M_w$ 5.8

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).

### 3.3. Acoustic backscatter signal

The background backscatter amplitude level is -43±1dB before the earthquakes.

Three to four hours after the $M_w$ 4.7 09/24/2019 earthquake, backscatter increases

sharply to -22dB between 11:00 and 12:00, and then decays to -41dB over 12 hours.

The increase of backscatter coincides with a change of current direction and speed,

584 indicating that the turbid cloud was brought to the instrument site by the current.
585 However, the current speed of less than 4 cm/s may have been insufficient to put the
586 particles in suspension. There is no increase of backscatter on Sept 20 when stronger
587 currents coming from the same direction, but not related with an earthquake, were
588 recorded.

589 Backscatter strengh remains -41±1 dB for the 1.5 day interval before the $M_w$ 5.8
590 09/26/1919 and increases to the -20 dB to -13 dB range after the earthquake (Figure
591 5). This implies sand sized particles or flocs were put in suspension soon after the
592 earthquake although the local current speed remained relatively low (about 5 cm/s at
593 most). After the device went back to near vertical position, signal strengh reaches a
594 maximum of -7.6 dB, which correspond to an amplitude ratio of 42 and an intensity
595 ratio of 1800 compared to base level. Similar signal strength levels are typically
596 reached with the Z-Pulse sensor in highly turbid water such as in estuaries. During
597 deep sea deployments signal strength range more typically between -60 and -40 dB.
598 After reaching peak value, backscattered signal strengh progressively decays over 3
599 days to stabilise at about -40 dB (Figure 5). Several turbid events, with signal strength
600 about -35 dB are observed in October and associated with small increases in current
601 velocity (up to 3-4 cm/s). It is unclear whether these passing clouds are residual
602 turbidity from the earthquake. After October 9, backscatter eventually returns to
603 background level while temperature decreases by 0.007 °C over a few hours,
604 indicating replacement of the water mass around the instrument.

605

606 3.4. Temperature record

607 The Sea of Marmara is stratified, with a low salinity (20-22‰) 20-30 m surface
608 layer that displays strong seasonal temperature variability (5-10°C in winter, 20-25°C
609 in summer) overlaying a high salinity (about 38‰) body of seawater at 14-15°C derived
610 from the Aegean Sea (Beşiktepe et al., 1994). Within the high salinity body, the
611 conservative temperature (McDougall et al., 2013) calculated with the Gibbs Seawater
612 oceanographic toolbox of TEOS-10 (McDougall and Barker, 2011) generally
613 decreases with depth. This implies that the adiabatic temperature rise in a turbidity
614 current, flowing downward, should cause a small temperature increase at the location
615 of the instrument. However, the deployment site is prone to seasonal cascading within
616 the deep water body, so that the initial temperature structure may have been disturbed.

632 Examples of CTD profiles recorded in June 2007 (Henry et al., 2007) are shown in

633 Figure 8 and indicate the presence of a slightly warmer water body on the seafloor,

634 only present in the basin along the base of the slope. No CTD profile is available in

635 Sept 2019, but variations in temperature and oxygen concentration associated with

636 mild currents (<5 cm/s) were recorded by the instrument in May-July 2019, and again

637 on Sept 20. It is therefore likely that the temperature at the location of the instrument

638 was slightly higher, by 0.01 to 0.02°C, than at the same depth in the central part of the

639 basin.

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

641 very small, less than ±0.002°C, which confirms that water movements during this event

642 were local. After the $M_w$ 5.8 09/26/1919 earthquake, the recorded temperature

643 decreases progressively by about 0.015°C to reach its minimum value when the

644 strongest current is recorded, around the time when the instrument straightens itself

645 (Figure 5). After that, temperature progressively increases back to reach nearly the

646 same value as before the event. The small variation in temperature indicates that the

647 turbid water originates from within the deep-water body. One remarkable observation

648 is that temperature only starts decreasing very slowly after the tilting of the instrument.

649 Temperature decreases at a higher rate after 14h, which is also when the tilted

650 instrument start measuring significant currents.

651        The slight temperature decrease observed after the earthquake can result from

652 the mixing of the warmer bottom water body originally present around the instrument

653 with the bulk of the deep-water layer in the Central Basin. Moreover, the observation

654 of a temperature drop precludes that the turbid water originates from depths less than

655 600 m, as water present between 600 m and the halocline is at a higher conservative

656 temperature than the deeper water throughout the year (see Beşiktepe et al., 1994,

657 and figure 8). Moreover, an inflow of water from closer to the surface should result in

658 an increase in the $O_2$ concentration in the bottom water, but none is observed in the

659 data.

660

a supprimé: .

a supprimé: did not travel far during this event

a supprimé: ,

a supprimé: , after the first hydrologic disturbance and tilting of the instrument, until the recorded current reaches its maximum value

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676
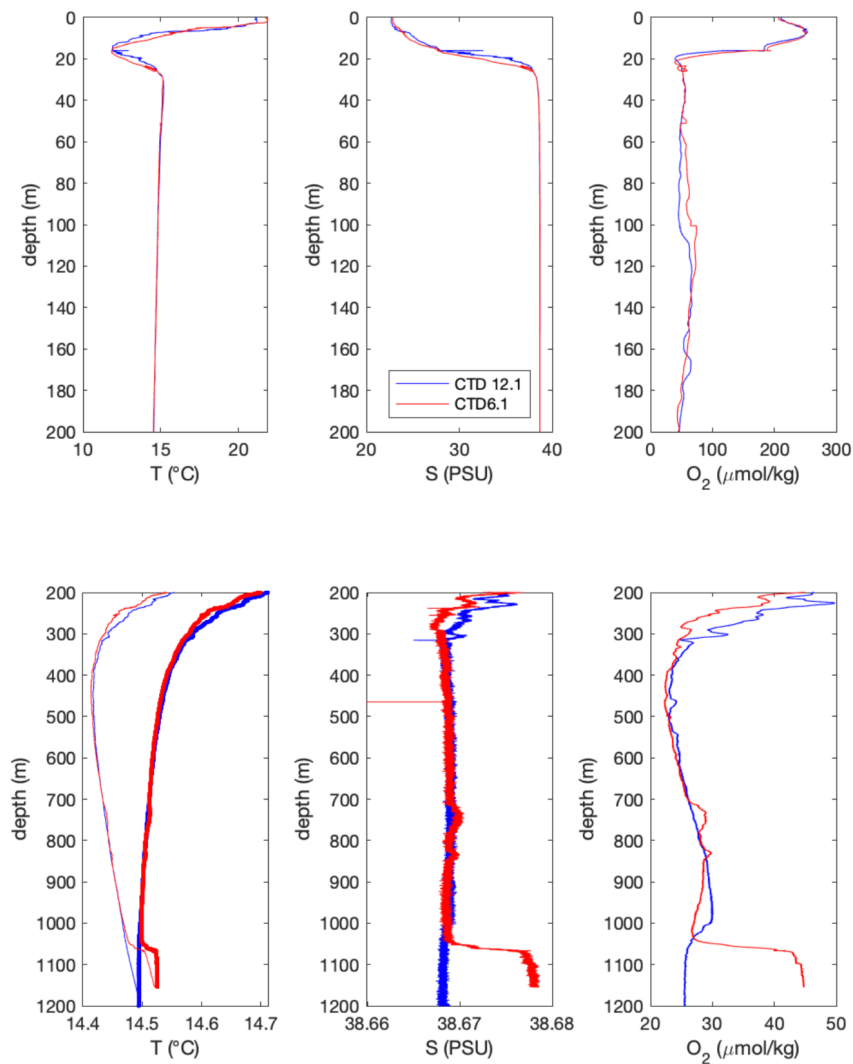
**Figure 8.** Depth plots of Temperature (°C), Salinity (PSU) and oxygen concentration

(μmol/kg) from CTD profiles acquired in the Sea of Marmara in June 2007 during

Marnaut cruise of Ifremer RV L'Atalante (Henry et al., 2007). On the lower

temperature plot, thin lines are measured values and thick lines are conservative

temperatures calculated at 1180 m. Locations are shown on Figure 1

682

> **a supprimé:** potential

## 4. Interpretation and discussion

### 4.1. Sequence of events

Let us first consider the potential influence of meteorology on the events recorded at the seafloor. Reanalysed ERA5 hourly wind and pressure data (Hersbach et al., 2018) interpolated at the location of the instrument indicate relatively low wind (less than 5 m/s) at the time of the earthquakes and during the hydrodynamic disturbances that followed (Figure 5). It is thus unlikely that wind influenced the course of these events. On the other hand, the current event on 20/09/2022 occurs at a time of high wind and follows a change of wind direction. Hypothetically, wind forcing may have caused this event, but probably not through sediment resuspension as acoustic backscatter remained low. A possible influence of wind on the motion of turbid clouds passing over the instrument after October 2 remains open for discussion.

The observations at the seafloor provide some insight on the complex sequence of events that followed the earthquakes and suggest the following scenarios. After the $M_w$ 4.7 09/24/2019 a turbid cloud formed east of the instrument and drifted slowly. Considering the maximum velocity of the current (less than 4 cm/s) and the 4-hour interval between the earthquake and the passing of the turbid cloud over the instrument, the front of turbid water should have formed East North East of the instrument at a maximum distance of about 500 m, and this coincides with the base of the northern slope near the outlet of the canyon. Small scale failures on the steeper slopes on the sides of the canyon and shaking are possible causes of sediment resuspension. The clouds subsequently drifted downslope over a total horizontal distance of at most 1 kilometers before dissipating, adding the 500 m estimate above to the drift calculated after the passing of the front over the instrument (Figure 6).

The $M_w$ 5.8 09/26/1919 caused stronger currents and a small temperature perturbation. Temperature records from turbidity currents invariably display a correlation between current onset and temperature change and this temperature change is nearly always positive (Mikada et al., 2006; Palanques et al., 2008; Kasaya et al., 2009; Xu et al., 2010; Khripounoff et al., 2012; Liu et al., 2012; Hughes Clarke et al., 2006; Johnson et al., 2017; Brizuela et al., 2019; Normandeau et al., 2020; Heerema et al., 2022). Temperature spikes may thus be used to infer turbidity current occurences and provenance (Johnson et al., 2017). That the currents we observe are associated with a temperature decrease rules out that they initiated at the shelf edge

725 as water above 600 m depth is at a higher potential temperature (temperature
726 corrected for the adiabatic gradient) than water at the seafloor (Figure 8). Water may
727 have been mixed locally or flowed down some distance down the slope. For instance,
728 currents may have originated from above the earthquake rupture zone, where the
729 seafloor lies in the 600-to-1200 m depth range.

730        The temperature records also concur with the current record to indicate that
731 currents in the water column remained moderate for several hours after the earhquake
732 and are not the primary cause of instrument tilting. First of all, there is a delay of at
733 least one hour after the earthquake (30 minutes after the tilting event) before
734 temperature starts decreasing significantly. Moreover, an acceleration of the
735 temperature rate of variation correlates with an increase in measured current speed
736 (to about 6 cm/s) between 14h and 15h (about 2 hours later), indicating that the tilted
737 current meter and the termperature sensors are providing concordant information.
738 Even if a short burst of current may have been missed because of the 1 hour interval
739 between current records, this would not explain why the frame remained stable in a
740 tilted position for several hours. Local liquefaction of the sediment beneath the device
741 is also an unlikely cause because the tilting of the instrument occurred 25 minutes after
742 the earthquake. A thin dense flow of remobilized sediment originating from the basin
743 slopes thus appears as a more likely cause. Partial burrial of the device is attested by
744 presence of sandy mud caked on the device in various places: on the frame feet, on
745 the acoustic releases, on the optode connector and, also inside the plastic protection
746 of a flotation sphere from which bindings were broken. On the other hand, the current
747 speed of at least 25 cm/s recorded before the time when the device straightened up is
748 strong enough to cause erosion of mud deposits. It may thus be hypothesized that
749 erosion freed the device from the mud cover. The flotation spheres on the frame and
750 the concrete ballast at its base exert a moment that should keep the assembly stable
751 in an upright position unless the frame is loaded with sediment.

752        Powerful turbidity currents driven by dense basal flows have notably been
753 observed in Monterey Canyon (Paull et al., 2018) and may share some characteristics
754 with the event reported here, although this event is much weaker. These dense flows
755 are relatively thin (< 2 m in the Monterey Canyon case) and have the ability to displace
756 instruments before the development of turbulence in the water column. It appears likely
757 that, after the passing of the seismic wave, failures on slopes adjacent to the
758 deployment site caused a debris flow or dense mud flow that spread on the basin floor

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760 causing the tilting of the instrument and bottom water turbidity while turbulence in the
761 water column remained limited. As the base of the nearest slope is about 400 m north
762 of the instrument, this would imply a minimum velocity of 20 cm/s for the mudflow to
763 reach the device location in 25 minutes.

764 During the following 10 hours, the current record is incomplete but indicates
765 variations in strength and direction. One possible explanation is that widespread slope
766 instabilities triggered by the earthquake have resulted in several turbidity currents
767 recorded as succesive pulses. Other possible explanations include oscillatory currents.
768 However, the role of seiches and surface gravity waves can be ruled out as no tsunami
769 was recorded by near shore tidal gauges around the Sea of Marmara. The relationship
770 between gravity wave amplitude A and bottom current amplitude U in the shallow water
771 linear approximation is given by $U=(g/H)^{1/2}A$, where H is water column height. An
772 oscillatory current of 10 cm/s at 1200 m depth would thus correspond to a free surface
773 oscillation of 1 m (or 100 hPa) for a standing wave (seiche) as well as a progressive
774 wave (tsunami). This should have been easily detected in a sea where tidal amplitude
775 is about 10 cm (Alpar and Yüce, 1998). The influence of baroclinic internal waves on
776 the halocline at 20-30 m depth must also be ruled out as they cannot physically
777 produce currents of more than a few cm/s at 1200 m. Nevertheless, It remains possible
778 that the interface at the top of the turbid cloud is affected by baroclinic waves.

779 The strongest current is recorded after 10 hours, which suggests that a turbidity
780 current initiated further upslope (but deeper than 600 m) may have reached the site
781 after a longer delay but may also have gained more kinetic energy on its downhill path.
782 This event, reaching a speed exceeding 25 cm/s apparently caused enough erosion
783 to free the device from the mud accumulation. The current then stabilizes in a westward
784 direction and decays progressively over the next 9 hours, which suggests the tail of a
785 turbidity current flowing in the canyon E of the deployment site has been recorded. The
786 hours-long delay between the earthquake and the passing of the fastest current over
787 the instrument may hypothetically correspond to the time for the head of the turbidity
788 current to travel from its source to the location of the instrument. Alternatively, a
789 sequence of slope failures may have lasted up to several hours after the earthquake.
790 Longer delays between loading events and turbidity currents, of several days to,
791 possibly, months, have been observed after floods (Carter et al., 2012) or after distant
792 earthquakes (Johnson et al., 2017). Another possibility is delayed ignition, which may
793 occur if the turbidity current develops from the hydrodynamic instability of a dilute turbid

815 cloud, indirectly resulting from slope failures and/or ground shaking (Parker, 1982;
816 Mulder and Cochonnat, 1996; Piper and Normark, 2009; Hage et al., 2019).

817 The distance travelled by the turbidity current on the basin floor cannot be
818 accurately estimated with a single instrumental record. However the drift plot (Figure
819 6) obtained during the waning phase may be roughly indicative of the distance over
820 which particles have been transported beyond the instrument by the turbidity current.
821 The drift distance is 3.5 km, and, when plotted over the bathymetric map, the drift
822 appears to stay within the depositional fan at the outlet of the cayon, the extension of
823 which is known from sediment sounder profiles (Figure 1). These calculations are only
824 a rough estimate of the distance travelled by suspended particles as only the velocity
825 at 1.5 m above the seafloor is known, and at a single point. Nevertheless, considering
826 that the current strength will decrease with distance on the flat seafloor of the basin, it
827 appears unlikely that sediments spread all over the 15x20 km basin floor as this would
828 require velocities of the order of 1m/s, sustained over a wide area for several hours.

829 The decay of the backscatter signal strength over the next three days may
830 reflect the settling of sand size particles, likely clay aggregates, from a dilute
831 suspension. This decay occurs in large part after the 9 hour waning phase of the
832 turbidity current, while current velocity remains lower than 4 cm/s. For a first order
833 assessment, Stokes settling velocity, an upper bound valid in dilute suspensions (e.g.
834 Guazelli and Morris, 2012) may be used. The Stokes settling velocity of 63 µm quartz
835 grains (density 2650 kg/m$^3$) in 13°C seawater is 2.7 mm/s, allowing such grains to drop
836 by 700 m in three days. However, if the particles forming the cloud are mostly
837 composed of clay agregates, which density may be between 1200 and 1700 kg/m$^3$,
838 the settling velocity would be comprised between 0.3 mm and 1 mm/s. In this case the
839 height of the suspended particle cloud could range between 70 and 250 m.

840

841 4.2. Current observations across the earthquake magnitude range
842 In this study a seafloor device located at the oultet of a canyon in the Central
843 Basin in the Sea of Marmara recorded a range of turbid events and currents induced
844 by earthquakes that has been rarely documented. In September 2019, $M_w$ 4.7 and 5.8
845 earthquakes occurred at a 5 km distance from the device as well as a series of smaller
846 foreshocks and aftershocks. In this setting, earthquakes of magnitudes less than 4 did
847 not cause noticeable water column turbidity nor currents. The $M_w$ 4.7 earthquake

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

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

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

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

877 $M_w$ 5.8 earthquake initiated a turbidity current and the data obtained may be compared

878 with more complete records obtained elsewhere with ADCP deployments and/or water

879 column mooring lines. A velocity of several tens of centimeter per second is

880 representative of the slower recorded examples, corresponding to mud rich flows

881 associated with hyperpycnal flows or small landslides (Khripounoff et al. 2012), or to

882 the smaller storm-related events (Normandeau et al., 2019). The event recorded is a

883 very weak event compared to turbidity currents that followed large earthquakes or large

884 slope instabilities. Cable breaks shows that the turbidity current triggered by the Grand

885 Banks 1929 earthquake, $M_s$ 7.2, reached velocities of at least 19 m/s (Piper et al.,

886 1999). Velocity of turbidity currents estimated form cable breaks in the Gaoping

887 Canyon and Manilla Trench system range 5.5-12.7 m/s for the $M_L$ 7.0 Pingtung

888 earthquake in 2006 and 5.9-7.9 m/s for a $M_L$ 6.4 earthquake (Gavey et al., 2017). From

889 instrumental records, velocities of 2-7 m/s were reported for the turbidity current

890 following Tohoku $M_w$ 9.1 earthquake (Arai et al., 2013) and 1.4 m/s in the Tokachi-Oki

891 $M_w$ 8.3 case (Mikada et al., 2006). The downward current after the off-Izu Penninsula

892 earthquake may be constrained with a noisy ADCP record to a maximum of 10-15 cm/s

893 in a 20-30 m layer above the seafloor and lasted about one hour, peaking about 30

894 minutes after the earthquake (Kasaya et al., 2009). This turbidity current thus appears

895 less intense and shorter in duration than the one recorded in the Sea of Marmara, but

896 the triggering earthquake was probably smaller (M 5.4 compared to $M_w$ 5.8) and more

897 distant (10 km). Moreover, the off-Izu event shares an important characteristic with the

898 Sea of Marmara one in that the turbid cloud is observed to form some time before

899 current builds up in the water column. When the maximum velocities reported are

900 plotted against magnitude (Figure 9), they show a tendency for larger earthquakes to

901 trigger stronger currents, which is hardly surprising. It also appears that estimates from

902 cable breaks tend to give higher value than instrumental records. This may perhaps

903 be because instruments give the maximum current speed at a single position while

904 cable breaks yield an integrated estimation of maximum current speed. Moreover, if

905 cable breaks are caused by a dense basal flow, it is yet unclear how its speed relates

906 to that of currents in the water column (Paul et al., 2018). The data set available today

943 remains insufficient to reach general conclusions regarding scaling, and other factors
944 than earthquake magnitude, such as slope, would need to be taken into account.

945

946 **Figure 9.** Maximum measured current velocity as a function of earthquake
947 magnitude for the cases discussed in text.

948

949     4.3.   Implications for sediment transport after earthquakes

950        Several observations from the monitoring records have special relevance for the
951 understanding of sediment resuspension and transport processes during earthquakes.
952 The first one is that earthquakes can induce sediment resuspension in situations where
953 current remains too low to be the primary cause or resuspension. This is apparent went
954 comparing events on Sept 20 (unrelated to earthquake and without turbidity) and Sept
955 24 (after $M_w$ 4.7 earthquake and turbid) that have comparable current speeds.

956 Resuspension may be an immediate effect of ground shaking or results from local
957 slope failures. This process may be important as it opens the possibility of triggering
958 turbidity currents after earthquakes by hydrodynamic instability within the water column.
959 The second one is that a mass flow sufficiently strong and dense to displace a heavy
960 instrument occurred at a time when there was no indication of advection in the water
961 column. Currents in the water column apparently continued to increase in strength after
962 this initial mass flow had stopped. A third observation is that the water displaced with
963 the turbidity currents is deep water, as indicated by the temperature record. Likely, the
964 displaced water originate from where the earthquake triggered sediment mobilization,
965 that is in relatively deep water around the earthquake source area north and west of

the instrument location (Figure 1). Turbidity currents more commonly originate from continental shelf edges or the upper part of continental slopes. This is notably the case when they are related to storms, river discharge or sediment loading. However, triggering by earthquakes may affect any part of the continental slope depending on the location of active faults. The case we reported shows that a moderate earthquake ($M_w$ 5.8) can cause sediment remobilization near the base of the slope rather than at the shelf edge resulting in different flow dynamics than generally assumed for sediment remobilization events.

The geomorphological context of the deployment site also needs to be taken into account. It is located on a depositional fan at the outlet of a canyon and south of a slope identified as unstable from geomorphological criteria (Zitter et al., 2012). We have shown that a debris flow or dense mud flow originating from this unstable slope, followed several hours later by a turbidity current flowing along the canyon could well explain the sequence of event following the $M_w$ 5.8 earthquake. Although this is not the only possible explanation for the observations, we believe it is the most likely one considering the geomorphological context. We estimated that the current during this event was probably too weak to spread a layer of sediment over the entire Central Basin floor. It is also unclear whether this event left on the fan a sedimentary layer that may be identified as a seismoturbidite, as a debris flow or as a layer of homogeneous mud. Differences between the fan and the basin in the number of sedimentary events and of their characteristics could explain why the sequence of seismic reflectors on sediment sounder profiles differs in the basin and in the fan (Figure 1). For all these reasons, the base of slope or canyon outlets are not good sampling locations for obtaining reliable earthquake records. In previous studies in the Sea of Marmara (e.g. McHugh et al., 2014), samples were taken across the basin depocenter for this purpose and events correlated between cores could also be correlated with historical earthquakes of estimated magnitude > 6.8. This approach remains in principle valid.

## 5. Conclusion

Instrumental records obtained in the Sea of Marmara Central Basin near the base of an unstable slope and the outlet of a canyon bring some insight on sediment remobilization by proximal (≈5 km) earthquakes and their hydrodynamic consequences.

-The smaller earthquakes (Mw < 4) are not associated with water column events

-A $M_W$ = 4.7 earthquake caused the formation of a turbid cloud and low currents not exceeding 4 cm/s.

- A $M_w$ = 5.8 earthquake at the same location caused a mass flow strong enough to capsize a heavy instrument. Subsequent movements of the water masses remained local, mixing deep waters at a scale of 5-to-10 km maximum.

This suggests that a continuum of hydrodynamic events of increasing intensity with earthquake magnitude occur above a threshold, corresponding to Mw ≈ 4 at the studied location. Moderate earthquakes can thus generate mass flows and turbidity currents of limited extension that may confuse paleoseismological records in cores taken near the edges of basins. However, the local nature of these events may help distinguish them from the consequences of storms and floods, expected to initiate from near the edge of the continental shelf. Performing new core studies and very high-resolution geophysical surveys in this area would thus have important implications for understanding under which conditions earthquakes leave a distinctive trace in the sediment record.

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