



Tectonic-sedimentary architecture of surficial deposits along the continental slope offshore Romania (North of the Viteaz Canyon, Western Black Sea): Impact on sediment instabilities

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1 Tectonic-sedimentary Architecture of Surficial Deposits along the Continental Slope Offshore
2 Romania (North of the Viteaz Canyon, Western Black Sea): Impact on Sediment Instabilities

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Abstract

The upper continental slope offshore Romania is a complex area hosting turbidite deposits, multiple types and ages of deep-seated faults, gas hydrates, gas-escape features, and numerous Mass Transport Deposits (MTDs). Multi-scale seismic data sets (2D-high-resolution and near-bottom very high-resolution) were used to study the interaction between such disparate geological features and determine their impact on slope stability.

At least five main paleo-valleys have been identified in the north of the Viteaz (Danube) canyon/valley. The most recent channelized systems linked to these valleys formed over a basal layer of MTDs. These MTDs are associated with an unconformity corresponding to the Base Neoeuxinian Sequence Boundary formed during the last major sea-level fall. This erosional surface shows scarp alignments that coincide with underlying faults. We argue that gravity-driven fault reactivation, with possible upward gas/fluid migration along these faults, is a determinant factor controlling sedimentary instabilities. Numerous MTDs are also observed during channel-levees building and reveal local sediment instabilities related to localized erosional process in the canyon. Finally, MTDs recorded within the upper draping unit, suggest that sediment instability also occurred during recent sea level highstand. Sediment pulse, seismicity, and gas hydrate dynamics can also play a determinant role in sediment instability throughout the sediment record.

Keywords:

Romanian continental slope; gas/gas hydrates; turbidite systems; sedimentary instabilities; gravitational faults.

1. Introduction

Sediment instabilities and gravity processes strongly affect continental slopes (Masson et al., 2006; Urlaub et al., 2013). Submarine landslides can involve large volumes of sediment (Taillandier et al., 2014) and cause damage to telecommunication cables (Pope et al., 2017 and references therein) and seafloor structures (pipelines). They can also trigger tsunamis (Piper et al., 1999; Løvholt et al. 2020). Their role in sudden gas release and the related potential climate changes is also open to debate (Kennett et al., 2003; Maslin et al., 2004). Numerous features influence the generation or triggering of submarine landslides (Masson et al., 2006). Rapid sediment deposition and high sedimentation rates are well-known instability triggering mechanisms (Lee, 2005). The presence of gas in the sediments (Field and Barber, 1993; Garziglia et al., 2008; Riboulot et al., 2013), melting of gas hydrates (Sultan et al., 2004a, b; Pecher et al., 2005; Kim et al., 2013), and earthquakes can also lead to slope failure where the presence of a weak layer constitutes favourable conditions (Locat et al., 2014). The relative impacts of these factors are often inadequately constrained, mainly due to the difficulty in evaluating the role of each of these factors on the stacking pattern of sedimentary bodies.

Accordingly, this study aims to provide the detailed spatio-temporal tectonic and stratigraphic framework of an area located offshore Romania (NW Black Sea) where numerous sediment instabilities exist within surficial sediment deposits (Fig. 1; Fig. 2). Previous studies of this area have shown that the Mass Transport Deposits (MTDs) are located along the slope where numerous gas seeps and evidence of gas hydrates have been recorded (Popescu et al., 2007; Riboulot et al., 2017; Hillman et al., 2018; Ker et al., 2019). The work presented here is based on a multi-scale seismic database (mainly 2D-high-resolution and near-bottom very high-resolution). The data provide detailed stacking patterns of turbidite systems and MTDs, fault patterns, and their relationship with BSR (Bottom Simulating Reflector) and free gas extents. The main objective is to enhance the identification and definition of the relative importance of factors controlling sediment instability in this particular context of gas/gas-hydrate dynamics, multi-phased tectonics and sea-level fluctuations.

2. General setting and previous work

2.1 Geological setting

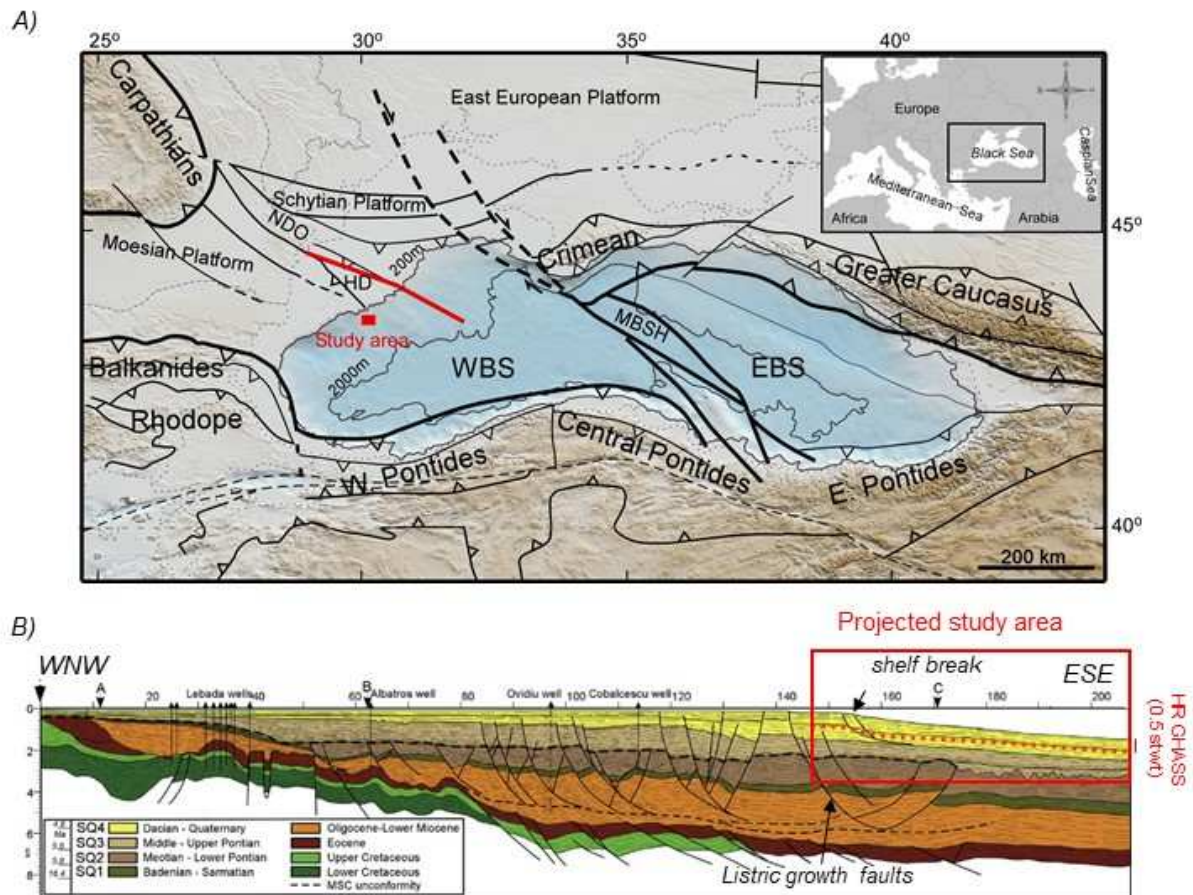


Fig. 1. Geological context of the Black Sea. A) Tectonic map (compilation from Finetti et al., 1988; Yilmaz et al., 1997; Nikishin et al., 2001; Dinu et al., 2005, 2018; Munteanu et al., 2011). Note the North Dobrogea orogenic belt (NDO), the Histria basin/trough well known in literature as Histria depression (HD over NDO) and the Eastern Black Sea (EBS), Western Black Sea (WBS), Mid Black Sea High (MBSH). The red line indicates the cross-section presented below. The red box indicates the location of the study area. B) Cross-section of the Romanian margin (from Matenco et al., 2016) showing the horsts and grabens resulting from the Mesozoic extensional phase and Cenozoic inversion. Note the gravitational

deformation of prograding sediments overlying late Oligocene layers and the associated faults. Note the listric growth faults below the present shelf break. The red box indicates the location of the study area on the upper continental slope.

The Western Black Sea is part of a back-arc domain that has experienced complex evolution with transition from an extensional setting (Cretaceous-early Eocene) to a contractional setting (late Eocene-Middle Miocene) during the northward Neotethyan subduction and collision (Letouzey et al., 1977; Zonenshain and Le Pichon, 1986; Finetti et al., 1988; Görür, 1988; Artyushkov, 1992; Robinson et al., 1996; Nikishin et al., 2001; Georgiev, 2012) (Fig. 1A). The rifting began during the Aptian and continued intermittently until the mid-Turonian (Krezsek et al., 2016). Continental break-up followed in the mid-Turonian caused by the regional uplift and erosion of the basin margin. Late Cretaceous- Paleogene times are marked by rift enlargement and eastward expansion and the development of the deep-sea basin (Munteanu et al., 2012). These two extensional stages led, in the northwestern area, to the formation of complex interplay between isolated blocks organized in horsts and grabens (Fig. 1B and red lines on Fig. 2), such as the Histria Basin (Fig. 1 and Fig. 2) which is located over the North Dobrogea Triassic-Jurassic orogenic belt (Dinu et al., 2005). Repeated periods of inversion from orogenic deformations (Balkans, Pontides, Crimean-Caucasus), are marked by the (re)activation of faults and associated folding during the late Cretaceous and Eocene-Miocene times (Hippolyte, 2002; Munteanu et al., 2011) (Fig. 1A).

Large thicknesses of sediments accumulated during the Pontian, i.e. Miocene-Pliocene transition from 6.04 to around 5 Ma (Dinu et al., 2003, 2005; Konerding et al., 2010; Tambrea, 2007) (Fig. 1B). At that time, which corresponds to the Messinian event, sea-level falls of debated amplitude (100-2300 m) led to large-scale erosion and significant progradation during the subsequent highstand (Hsü and Giovanoli, 1979; Gillet et al., 2007; Munteanu et al., 2012; Krezsek et al., 2016). The Dacian i.e. around 5 to 4 Ma, and the Romanian-Quaternary sediments reached thicknesses of up to 1200 m and 600 m respectively on the continental slope (Konerding et al., 2010) (Fig. 1B).

The high sediment load and high subsidence led to the development of gravitational faults (Konerding et al., 2010) (Fig. 1B) and overpressure of the Oligocene- early Miocene? sediments (Dinu et al., 2003; Bega and Ionescu, 2009), which provided the detachment level for shelf collapse during the Messinian large-scale sea level drop (Tambrea et al., 2002; Dinu et al., 2003; Munteanu et al., 2012; Schleder et al., 2016).

During the Plio-Quaternary, thick successions of MTDs and turbidite deposits took place in front of modern rivers (Munteanu et al., 2012; Matenco et al., 2016). The north-western Black Sea is currently dominated by a rather wide shelf (60-200 km) with a shelf break at 120 to 170 m water depth and canyon systems with large deep-sea fan complexes that mainly developed during sea-level Quaternary lowstands (Winguth et al., 2000; Popescu et al., 2002). The Danube delta system advanced into the Black Sea in the Romanian-Pleistocene in response to the increased sediment supply due to renewed uplift of the Carpathians (Matoshko et al., 2019; Krézsek and Olariu, 2021). The Danube river reached the Black Sea for the first time about 900 ka ago and subsequently built up the Danube fan (Winguth et al., 2000).

Note that several earthquakes were recorded in this area, especially from deeply rooted faults (Fig. 2, Table 1). The north-western Black Sea is also rich in hydrocarbon fields (oil, gas) (Kruglyakova et al., 2004; Georgiev, 2012), gas hydrates (Popescu et al., 2007; Merey and Sinayuc, 2016), and surficial gas systems with abundant gas seeps, pockmarks, and shallow gas fronts (e.g. Polikarpov et al., 1989; Egorov et al., 1998, 2011; Artemov et al., 2007; Naudts et al., 2008; Nikolovska et al., 2008; Greinert et al., 2010; Diaconu et al., 2020; Römer et al., 2020)..

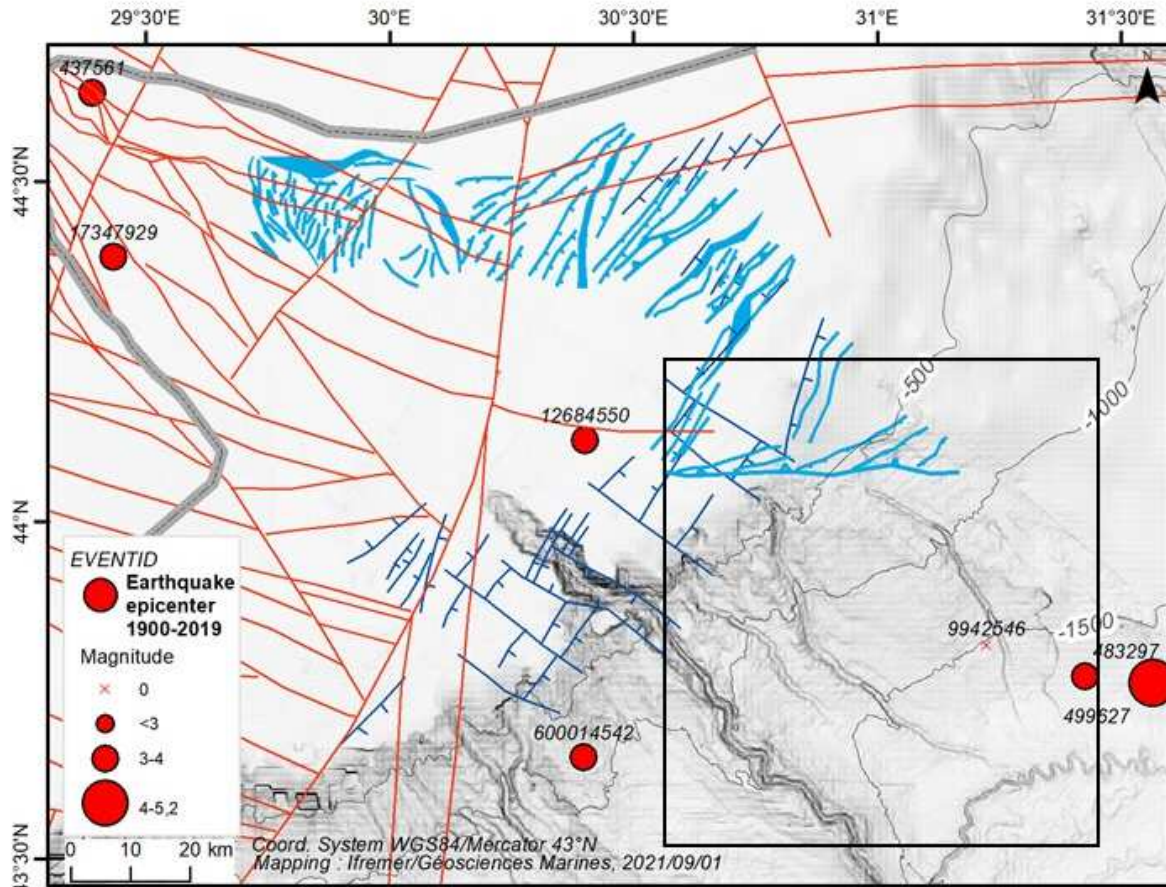


Fig. 2. Morpho-bathymetric map based on EMODNET2016 and GEBCO2014 bathymetric data with superposed structural map, offshore Romania. Note the distribution of canyons and paleo-valleys. The distribution of faults is indicated with: in red, compilation from Diaconu et al., 2020 from the data of Dinu et al., 2005; Morosanu, 2007 and Munteanu et al., 2011; in light blue, gravitational faults from Tambrea et al., 2002 and Dinu et al., 2003; in dark blue gravitational faults from Konerding et al., 2010. Note that the sets of faults of different colors have been mapped from distinct and incomplete grids of seismic profiles thus leading to an incomplete map. In grey dotted line, the border of the Hystria Basin (from Pătruț et al., 1984 and Anton et al., 2019). The earthquake epicenters (time period 1900-2019) are from the ISC catalog (<http://www.isc.ac.uk/iscbulletin/search/catalogue/>). The nearest earthquakes of the study area are presented in Table 1. The study area is indicated by the black box (see Fig. 3).

Event	Lat	Long	Mag	Depth (km)	Origin Time	Station	Distance (° and km)
499627	43.7622	31.5640	4.6	37.5	03/03/1986 07:26:08.21	Sevastopol	1.71° ~189 km
12684550	44.1208	30.4006	3.9	10	20/06/2007 22:34:07.33	Tirgusor	1.47° ~163 km
12701122	43.9301	31.6372	4	3.7	18/02/2009 09:44:10.49	Sevastopol	1.58° ~175 km
600014542	43.6516	30.3969	3.1	12	25/08/2011 08:22:40.42	Mangalia	1.32° ~146 km

165 *Table 1: Earthquake Data extracted from the International Seismological Centre (2020).*

166 *Errors are not reported in the catalog. Distance in km from the closest station was calculated*

167 *with $1^\circ = \sim 111$ km.*

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169 2.2 Previous work on the study area

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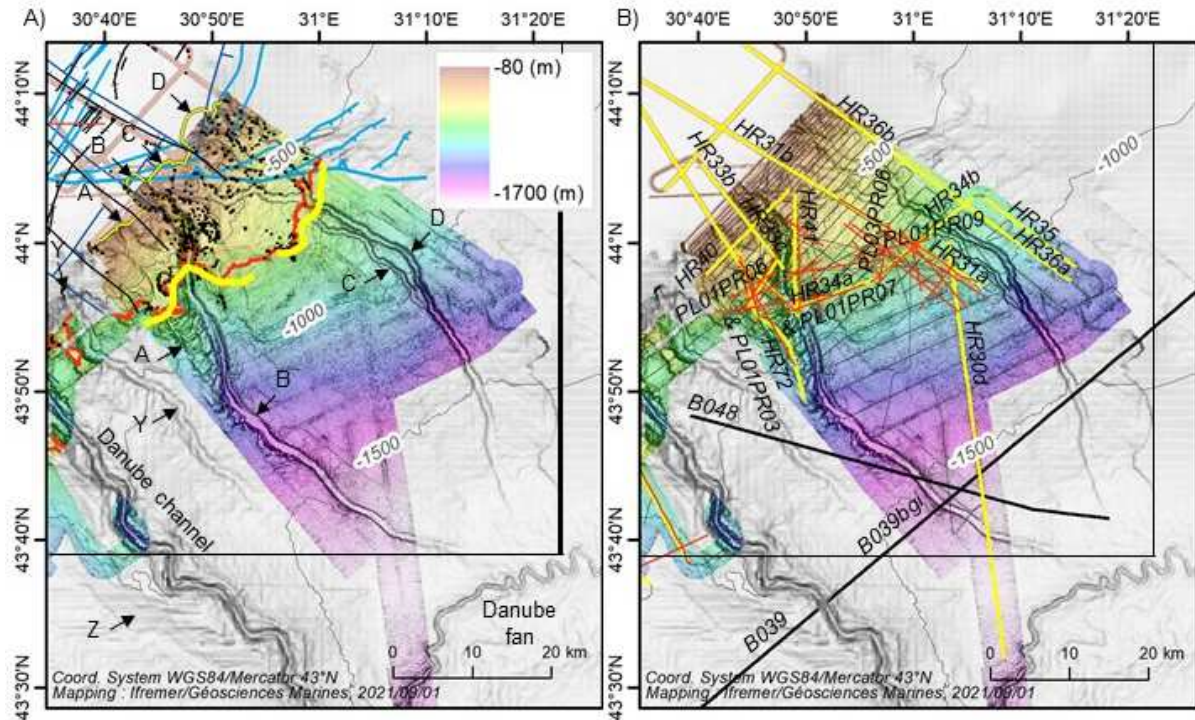


Fig. 3. Presentation of the study area (see location on Fig. 2). A) Geological setting showing the shelf break (dark yellow line) (Riboulot et al., 2017), the canyons/paleo-valleys and associated channel complexes (A, B, C, D, Y and Z), the Danube channel and Fan, the morphology lines i.e. sedimentary ridges (black barbed wires) (Assemblage project), the faulting pattern (see Fig. 2 for references), the upper limit of the model-predicted steady-state GHSZ (red line) (from Ker et al., 2019) and the observed BSR (yellow line) (Colin et al., 2020b) and, the gas flares (black dots, Riboulot et al., 2017). B. Location of seismic data acquired during GHASS and Blason surveys showing high-resolution seismic profiles (yellow lines, 2015 GHASS cruise; bold black lines, 1998 Blason cruise DOI 10.17600/98020030), SYSIF profiles (red lines), sub-bottom profiles (black lines, 2015 GHASS cruise, DOI 10.17600/15000500). Basemap: Morpho-Bathymetric map based on EMODNET2016 and GEBCO2014 bathymetric data (grey map) with superposed bathymetry from GHASS (2015) (colored map). The black box corresponds to the maps presented in Fig. 7, 15, 19, 20.

The study area lies to the northeast of the Danube paleo-delta basinward of the shelf-break from around 180 to 1300 m water depth (Fig. 3A). The continental slope varies around 2°-4°

188 between 200 and 500 m water depth respectively and up to 35° locally (canyon side walls).
 189 The outer shelf and upper slope are incised by several canyons formed by erosion prolonged
 190 on the middle and lower slope to channel-levees (e.g., Popescu et al., 2002; Riboulot et al.,
 191 2017). Around 20% of the seafloor surface is disturbed by scarps linked to instabilities. The
 192 scarps mainly occur in the canyons and paleo-valleys. Scarps also occur on the open upper
 193 slope between 200 and 500 m water depth. MTDs have been observed between 200 and
 194 900 m water depth (Riboulot et al., 2017; Hillman et al., 2018). Seaward of the study area,
 195 the deep-sea thick deposits of mostly clayey sediments were mainly supplied by the Danube
 196 and the Dnepr Rivers (Soulet et al., 2013; Matenco et al., 2016). Faults of interest in our
 197 study area are faults related to the Messinian gravitational collapse and younger
 198 reactivations mainly during middle Pliocene (Dacian, in local stage) (Tambrea, et al., 2002;
 199 Dinu et al., 2003; Munteanu et al., 2012; Schleder et al., 2016). The first generation (Middle
 200 Pontian) gravitational faults are related to the so-called “eastern gravity-driven deformation
 201 system” of Schleder et al. (2016) characterized by an updip extensional domain with SW-NE
 202 oriented normal faults partially accommodated by a downdip contractional domain marked by
 203 a mainly W-E oriented thrust (Tambrea, et al., 2002; Dinu et al., 2003; Munteanu et al., 2012;
 204 Schleder et al., 2016) (in light blue on Fig. 2, Fig. 3A and, Fig. 19B). Second generation
 205 (Dacian) gravitational faults have been observed in a grid of seismic profiles located near the
 206 shelf edge. These faults form mostly NE-SW trending grabens and horsts separated by NW-
 207 SE oriented transfer faults (Bega and Ionescu, 2009; Konerding et al., 2010) (in dark blue on
 208 Fig. 2, Fig. 3A and Fig. 19B).
 209 The presence of gas hydrates is inferred from the identification of the Bottom Simulating
 210 Reflector (BSR), which marks the base of the Gas Hydrate Stability Zone (GHSZ). The
 211 GHSZ corresponds to the zone where gas hydrates (Sloan, 1990, 1998; Kvenvolden, 1993),
 212 i.e. the ice-like solid form of gas, develop under conditions of low temperature, high pressure,
 213 and sufficient gas concentrations (Kvenvolden and McMenanim, 1980; Sloan, 1990, 1998;
 214 Dickens and Quinby-Hunt, 1997). The BSR was first mapped by Ion et al. (2001), then
 215 Popescu et al. (2006). Pockmarks, gas flares, and seismic facies related to the presence of

gas have been also reported. The gas flares were generally acoustically detected in the water column upward the landward limit of the GHSZ at 660 m water depth (Popescu et al., 2004; Riboulot et al., 2017). They mainly consist of gas emissions released to the ocean from the seafloor from underlying sources. Finally, the study area has undergone major changes in environmental conditions relative to sea-level changes and induced connection-disconnection of the Marmara-Mediterranean Sea during last interglacial-glacial cycles (Soulet et al., 2011; Constantinescu et al., 2015; Wegwerth et al., 2015; Ballas et al., 2018). These cycles cause alternating marine and freshwater lake conditions (Badertscher et al., 2011) and connection-disconnection between the Danube River, its canyon, and consequently the deep-sea deposit systems (Lericolais et al., 2009; Constantinescu et al., 2015).

3. Methodology

The results presented in this paper originate from the interpretation of several data sets including (Fig. 3B):

(1) Multi-beam echo-sounder data acquired during the GHASS (2015) campaign (Reson SeaBat 7111 for shallow water 5-500 m, and 7150 for deeper water 200-2000 m) with a bathymetric resolution of 20 m (already published). They have provided the geomorphology of the seafloor from a dense grid of profiles. Water column acoustic data processed on board with SonarScope and GLOBE software have provided the distribution of gas flares. These profiles have not been mapped on Fig. 3B.

(2) 2D-high and very-high-resolution (VHR) seismic reflection profiles acquired during the 1998 BlaSON surveys of Ifremer and GeoEcoMar. High-resolution seismic profiling was implemented using a mini-GI air-gun seismic source (frequency range 150 Hz and vertical resolution of 2.5 m) and a 24-channel streamer. Sub-bottom profiles (Chirp) were acquired simultaneously to investigate surficial sediments.

(3) 2D-high-resolution (HR) multi-channel seismic profiles acquired during the GHASS survey using a single 24/24 mini-GI air-gun seismic source (central frequency at 115 Hz and vertical resolution of 3.2 m) for profiles GAS031, GAS033 to GAS039, two 24/24 for profile GAS030, a single 13/13 for profiles GAS040 and GAS041, and a 96-channel solid streamer (inter-trace of 6.25 m and maximum source-receiver offset of 650 m). These profiles are named HRxxx in Fig. 3B. The seismic dataset was processed using a conventional post-stack sequence with constant velocity (1500m/s) (Ker et al., 2015).

(4) 2D-very high-resolution hull-mounted Sub-Bottom Profiler (SBP) data acquired during the GHASS survey. The SBP operates at the frequency range 1800-5300 Hz, giving a vertical resolution close to 20 cm. The quality control of SBP data was performed using the Ifremer QC Subop software (Dupont, 2015; Ker et al., 2015).

(5) 2D-very high-resolution deep-towed seismic profiles were acquired with the Ifremer SYSIF system (Marsset et al., 2010; Marsset et al., 2014; Marsset et al., 2018) during the GHASS survey. SYSIF is a deep-towed seismic device hosting a Janus-Helmholtz transducer emitting a linear chirp signal in the frequency bandwidth 220-1050 Hz and a streamer, both towed at 2 knots, 100 meters above the seafloor. Processing includes correction of altitude, and seismic signature deconvolution and provides 15-m-lateral resolution at the seafloor and 1-m-vertical resolution.

All seismic data were integrated into IHS Kingdom software for interpretation. Measurements in meters were calculated from Two-Way Traveltime (TWT) values using $v = 1500$ m/s.

Seismic interpretation is based on: (1) analysis of the reflectors' lateral terminations (onlap, down lap, erosional truncation) according to the general principles of seismic stratigraphy (Mitchum et al., 1977); and (2) characterization of seismic facies and correlation of distinctive reflectors between profiles. The cross-correlation between complementary seismic datasets provided mapping of unconformities, sedimentary bodies, faults, and gas systems. SBP profiles are not presented in this paper due to the poor quality of data caused by the presence of shallow gas. Only one Blason profile is presented. SYSIF was mainly used to identify sedimentary discontinuities such as scarps related to MTDs, channel sidewalls, and

faults. It also allowed to identify facies and seismic anomalies related to fluid/gas features and consequently permeable and seal layers. All seismic profiles are presented with overlapped interpretative line drawing in this paper.

4. Results

4.1 Seismic units and facies.

Architecture of the study area consists of the stacking of different types of seismic units classically encountered in similar slope and turbidite environments (e.g. Twichell et al., 1991; Piper et al., 1997; Droz et al., 2003; Madof et al., 2009) and already identified in the Danube Fan (Zander et al., 2017; Hillman et al., 2018; Winguth et al., 2000; Popescu et al., 2004;). The majority of these units are the channel/levee systems and MTDs. For the remaining units, deposits correspond to infilling and draping near seafloor units. A detailed description of these units and their relationship with faults, gas signatures and key surfaces, in terms of sequential stratigraphy, is proposed in sections 4.2 to 4.6.

Mass Transport Deposits

MTDs are of two types and two generations, either 1) irregular masses of wide lateral extent (up to 300 km² in size and up to 80 m in thickness) characterized by a transparent to chaotic high amplitude facies, a very irregular top surface with hyperbolic reflections and an erosive basal surface, truncating underlying reflections. These MTDs are localized over a basal major erosional surface (e.g. MTD12 in Fig. 4); or 2) small masses up to around 10 km² in size and 35 m thickness characterized by hummocky, transparent to chaotic, and irregular seismic facies, with continuous, low to high amplitude reflections and an erosive basal

surface. These MTDs are near the seafloor, localized at the upper slope and channel walls, where they rework superficial deposits (e.g. MTD2 in Fig. 8B).

Channel-levee systems

The channel is generally V-shaped with an erosional base. The levees are wedge-shaped and characterized by low amplitude, high frequency, remarkable continuity of reflectors. The lateral extent and thickness of channel-levees are variable. Some channels form a channel complex (e.g. channel complex B in Fig. 8A).

Filling sedimentation

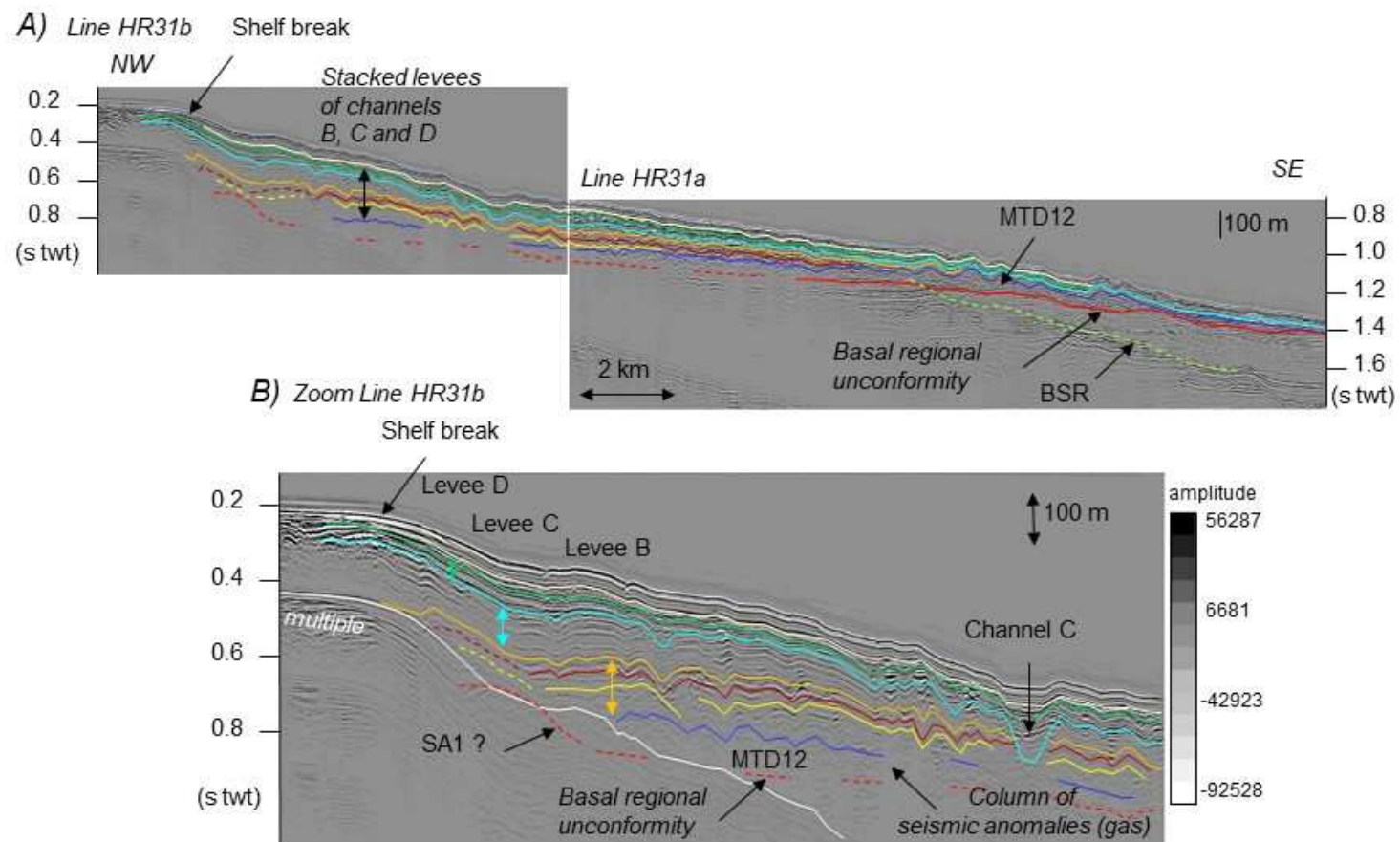
The channel-levees abandonment is marked by channel infilling with transparent facies interpreted as distal overbank deposits from younger channels (e.g. Filling unit in Fig. 8A).

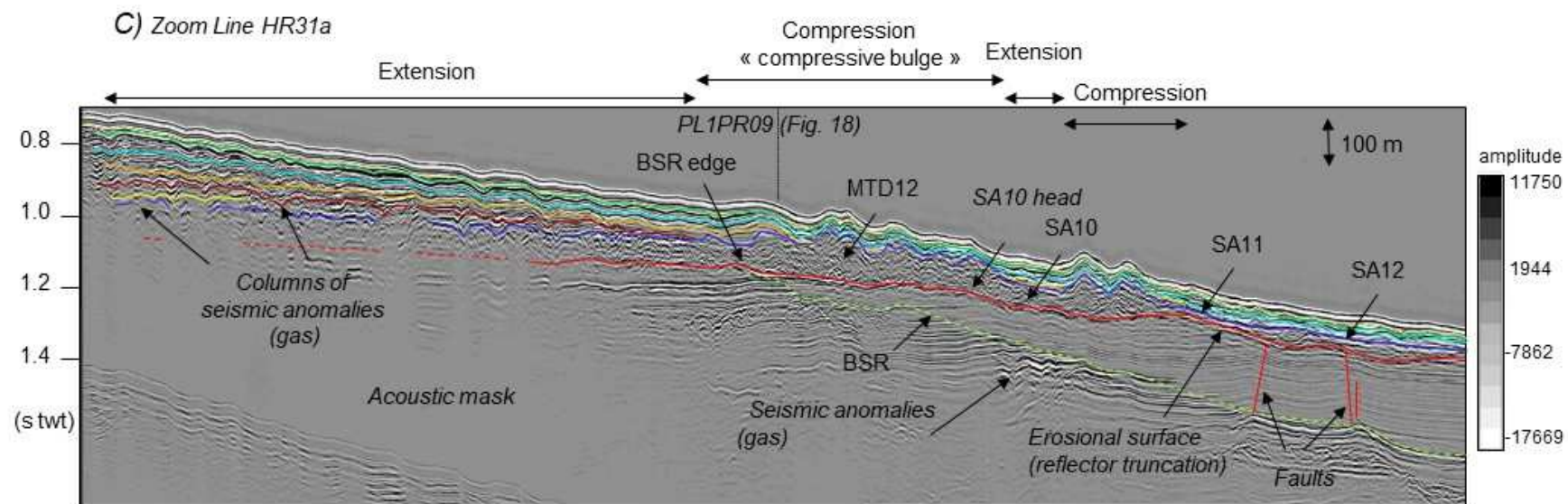
Draping sedimentation

The study area is covered by a draping unit of low amplitude, low frequency and continuous parallel reflectors interpreted as very fine-grained sediments. The thickness of this unit ranges from 20 to 40 m (e.g. Draping unit in Fig. 8A).

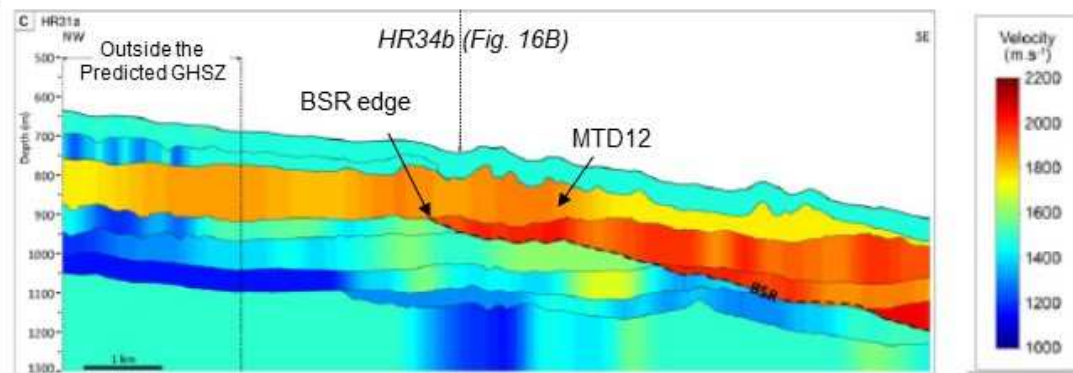
4.2 Basal regional unconformity, slide scarps, and associated MTDs

A basal and regional unconformity, ranging between 300 m below the seafloor (mbsf) upstream to locally 10 mbsf downstream, has been identified despite the scarcity of HR seismic profiles (Fig. 4A). This unconformity corresponds to an erosional surface that truncates reflections of underlying seismic units (Fig. 4C). It is generally well expressed in the middle slope but poorly visible in the upper slope due to the presence of gas (Fig. 4B), except in the western part of the study area (see section 4.3).





D) P-wave velocity analysis of Line HR31a



328

329 *Fig.4. Seismic characteristics of the basal regional unconformity, scarp alignments (SA),*
330 *MTDs, BSR and gas/GH. A) The high-resolution seismic dip Line HR31a-b (for location see*
331 *Fig. 3B) shows the extent of the basal unconformity. B) Zoom of Line HR31b showing the*
332 *shelf break and SA1 (see Fig. 7). The upstream part of the basal unconformity and SA1 are*
333 *not very clear due to the presence of gas. Note the location of MTD12 over the unconformity.*
334 *The upper colored reflectors correspond respectively to the top of the levees of channel*
335 *complex B (yellow, brown, orange), channels C (turquoise blue) and D (green) and upper*
336 *units (light yellow and seafloor). The crossing of the Line with channel C is indicated (Fig.*
337 *3B). C) Zoom of Line HR31a showing the lower part of the basal regional unconformity. Note:*
338 *the truncation of reflectors of underlying layers of this unconformity interpreted as an*
339 *erosional surface; the presence of SA10 to SA12 on the erosional surface (see Fig. 78); the*
340 *extensive part (low deformation) of MTD12 downstream SA10 (Fig. 4C) and the compressive*
341 *part (high deformation) just upstream SA11; the chaotic facies of MTD12 and the hummocky*
342 *top surface of the compressive bulge; the BSR (light green color) and seismic anomalies*
343 *related to gas beneath; the acoustic mask and columns of seismic anomalies upstream the*
344 *BSR edge. D) P-wave velocity analysis of Line HR31a. Note the velocity range (1900-2050*
345 *m/s) which characterize the GH just above the BSR and the velocity range within MTD12*
346 *(1750-1900 m/s). The low velocity values (1200-1500 m/s) indicate the presence of free gas-*
347 *bearing sediments (from Ker at al., 2019).*

348 This erosional surface is characterized by several scarps (Fig. 4C; Fig. 5). The scarps were
349 manually picked and mapped and vary in dimension, i.e. length (L) and height (H). Scarp
350 alignments called SA (SA1 to SA13) have been identified (see for example Fig. 5 and Fig. 6)
351 partly based on their dimensions and/or dip changes of underlying reflectors and their
352 mapping shows a general NE-SW orientation (Fig. 7A). Dip changes in stratifications are
353 characteristic of this area and are discussed in section 5.1. The basal unconformity displays
354 two major erosional troughs up to around 300 m below paleo-valleys A and B (see the
355 detailed description of paleo-valleys in section 4.3) (Fig. 8A). These erosional troughs are

characterized by a notably irregular topography due to numerous scarps along the dip profiles, particularly for paleo-valley A (Fig. 9).

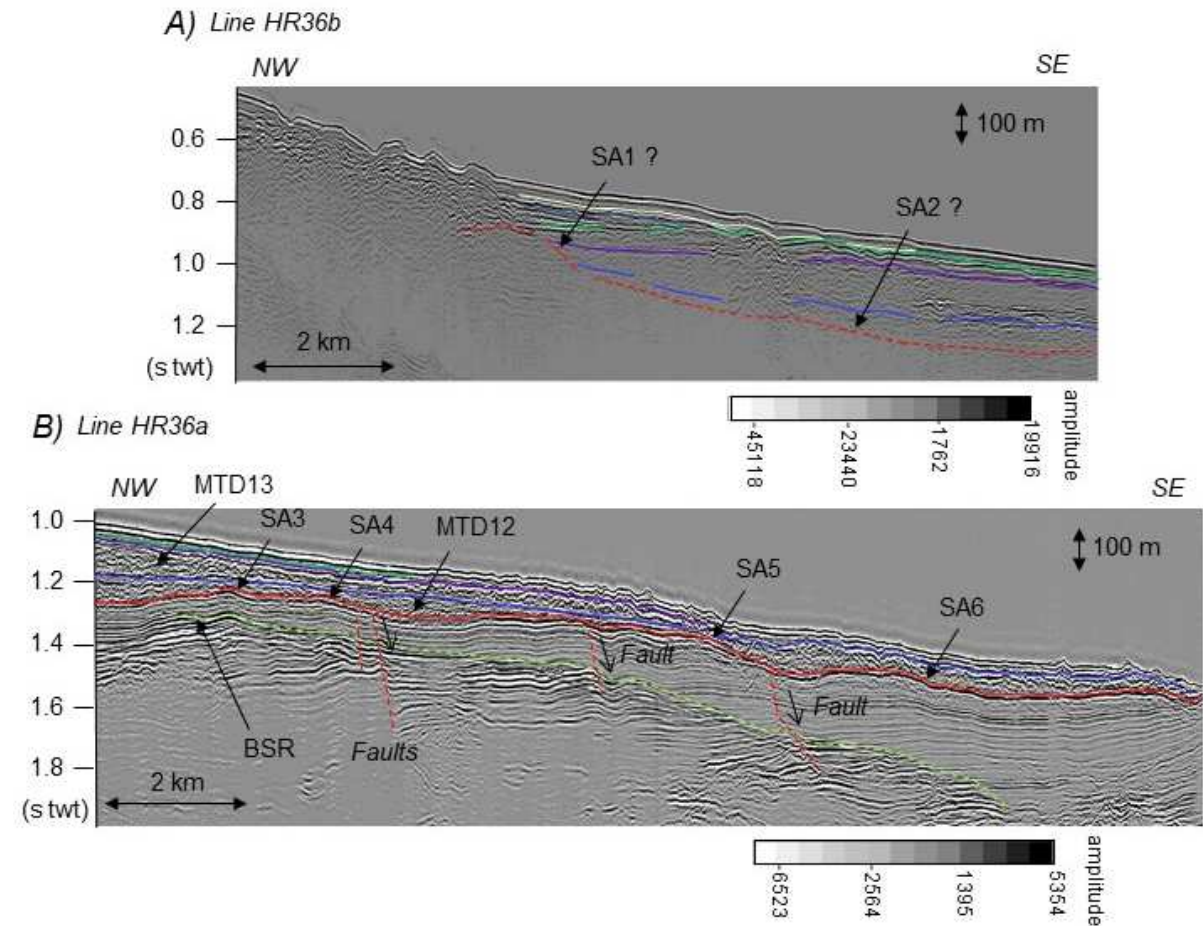


Fig. 5. Example of normal faults below scarp alignments in the eastern part of the study area. A) The high-resolution seismic Line HR36b (for location see Fig. 3B) shows the shelf break, SA1 and SA2 (see Fig. 7) and overlying MTD12 and MTD13. B) Line HR36a showing SA3 to SA6 and normal faults below these scarp alignments within the GHSZ. Note the listric shape below SA5. Arrows to the hanging wall/foot wall define the sense of motion.

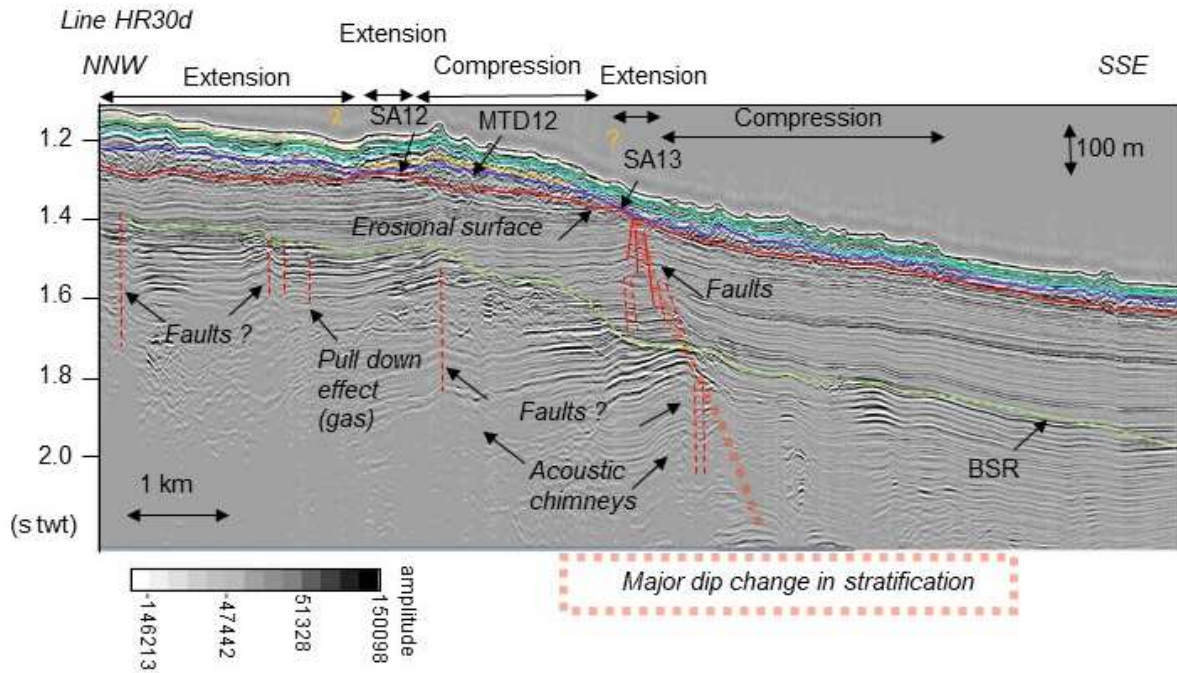
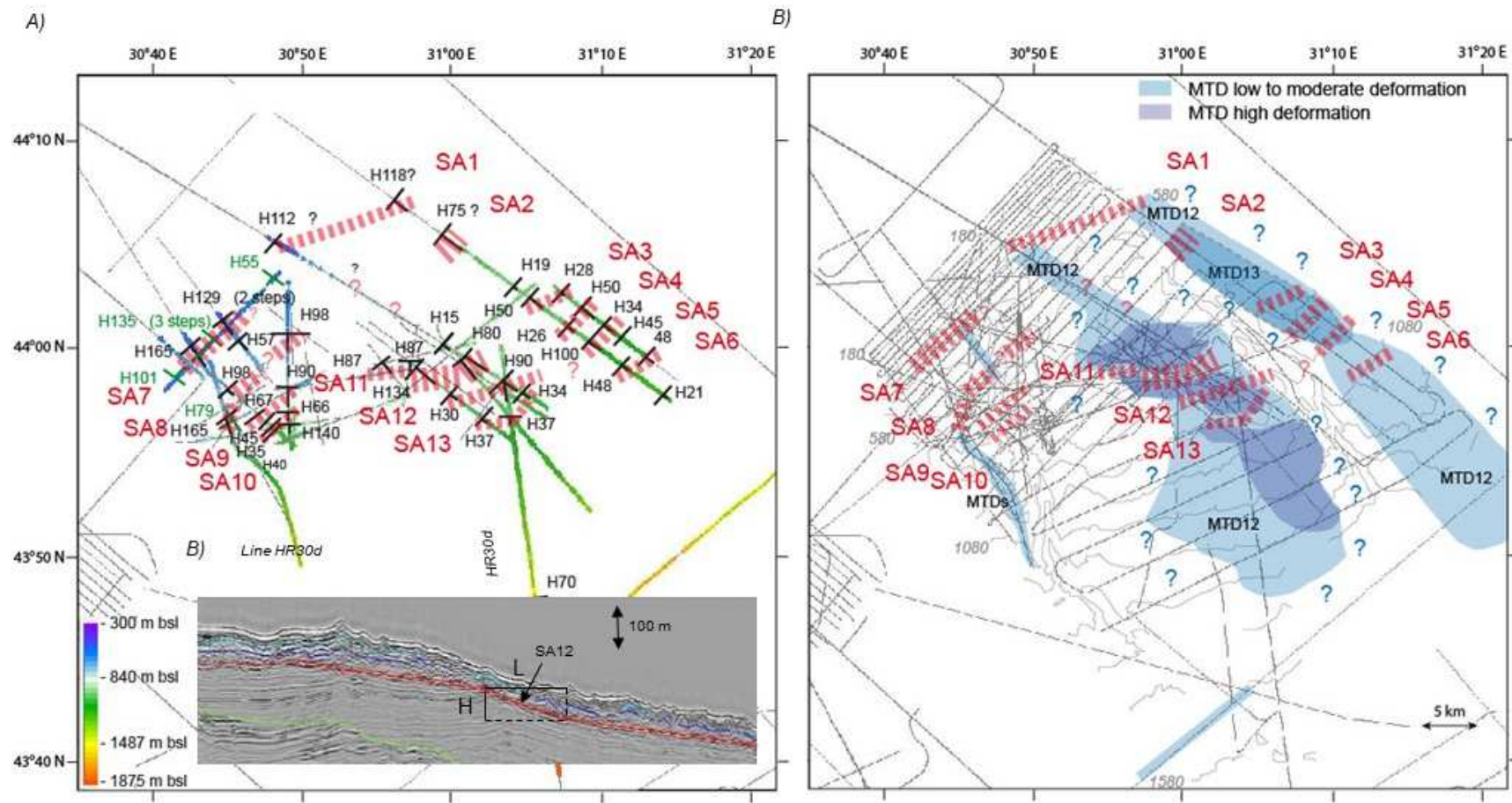


Fig. 6. Relationship between scarp alignments of the erosional surface, MTDs, BSR, and faults. The high-resolution seismic Line HR30d (for location see Fig. 3B) shows the stacked small-offsets of reflectors (faults?) and pull down effects (gas?) below the BSR (in light green). Note the faults and the major dip change in stratification below SA13 at the erosional surface and the accordion-like shape of MTD12 with alternating extensive and compressive parts. Note at SA12, the location of the head of the extensive part which is generally the case on profiles.

375



376

Fig. 7. Map of the basal regional unconformity and associated MTDs (see location on Fig. 3).

A) Map showing the seismic-bathymetry of the basal erosional surface and the scarps from HR and SYSIF seismic profiles. The dash in black crossing the profile corresponds to a scarp, the value in meter (Hxx) indicates the scarp height and the length of the dash in black along the profile indicates the horizontal kilometer length of the scarp. The scarps bounding the paleo-valleys A and B laterally are in green to better distinguish these scarps in this area where scarps are numerous. Scarp Alignments (SA) SA1 to SA13 are indicated by dashed red lines. B) Definition of the height and length of the scarp on the seismic profile. C) Map of MTDs covering the basal regional unconformity. This map is based on all seismic data and presents the distribution of MTDs. Scarp alignments SA1 to SA13 have been reported. Note the correlation between the heads of MTD12 and SA1, the location of MTD13 over MTD12 and, the distribution of low (in light blue) to moderate and high deformation (in dark blue) parts for MTD12 with respect to SA1 to SA13. Basemap: morpho-bathymetric map based on EMODNET2016 and GEBCO2014 bathymetric data.

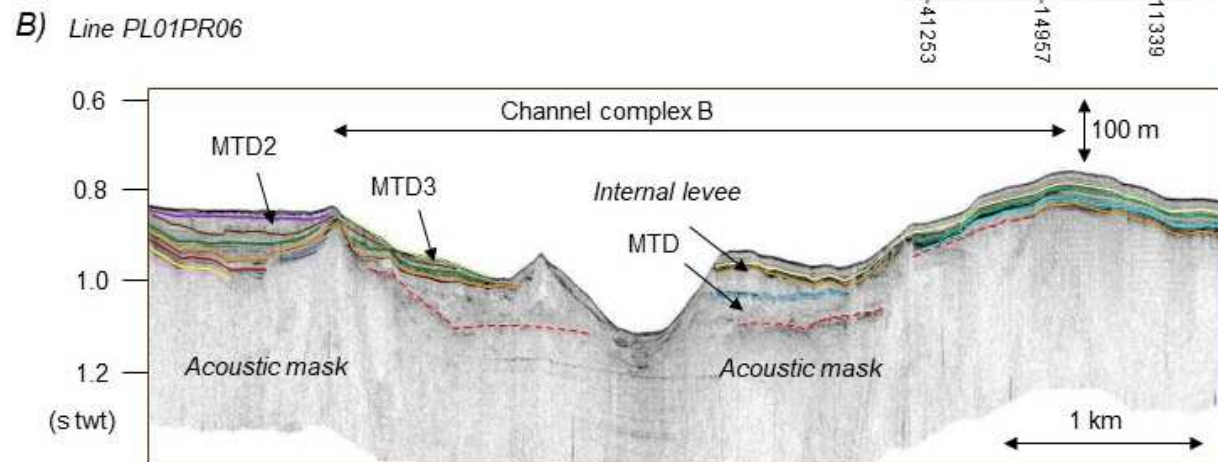
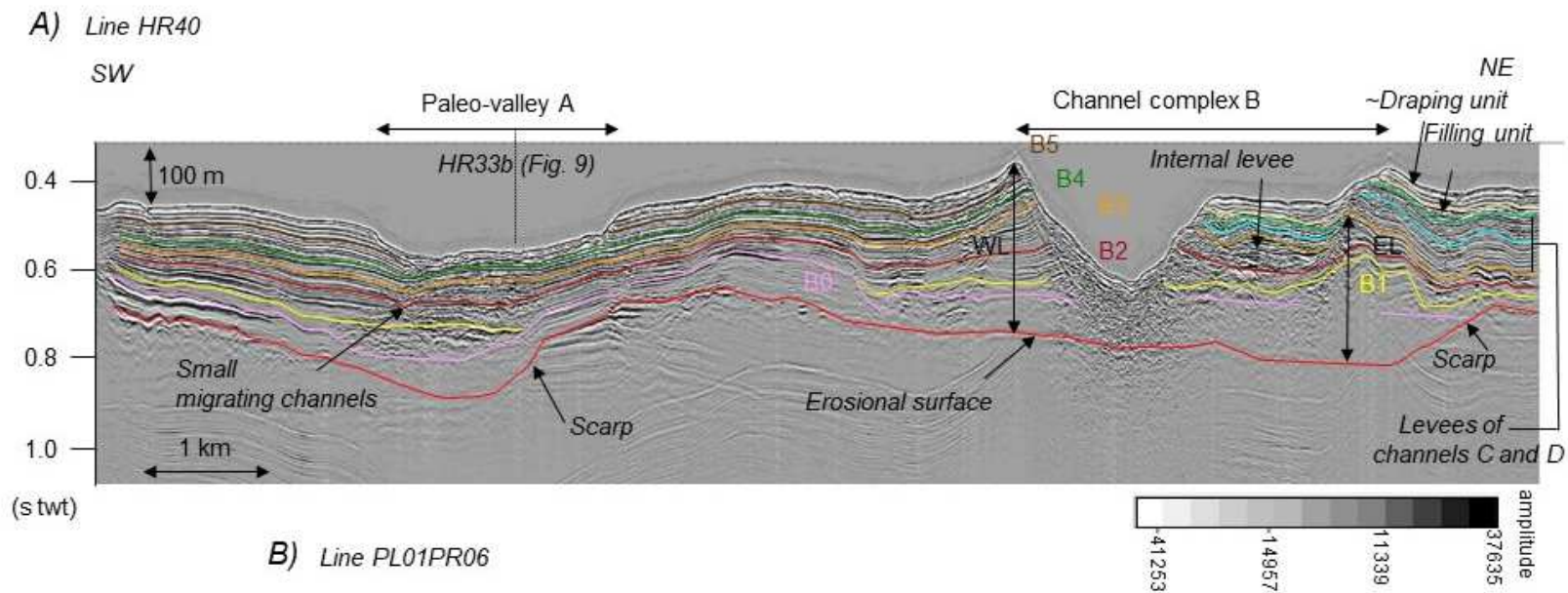


Fig. 8. Cross profile of Paleo-valley A and channel complex B. A) The high-resolution seismic Line HR40 (for location see Fig. 3B) shows the cross-section of Paleo-valley A and channels B0 to B5 and the basal erosional surface below. Note: the erosional trough and the lateral scarps marking the base of paleo-valley A and its filling by the western levees of channels of B0 to B5; the stacked migrating small channels within this filling. The channel complex B is marked by an asymmetric shape with a staircase structure of the eastern side. Note the migration of channels B0 to B2. The double arrows indicate the channels axes of B0 and B1. On the eastern side of channel complex B, the internal levee is considered as coeval to the levees B4 and B5 of the western side. Note the presence of distal levees of channels C and D and the filling and draping upper units. B) The SYSIF PL01PR06 profile (for location see Fig. 3B) shows an example of superficial MTDs i.e. MTD2 and MTD3 within paleo-valley A over the western levee of channel complex B.

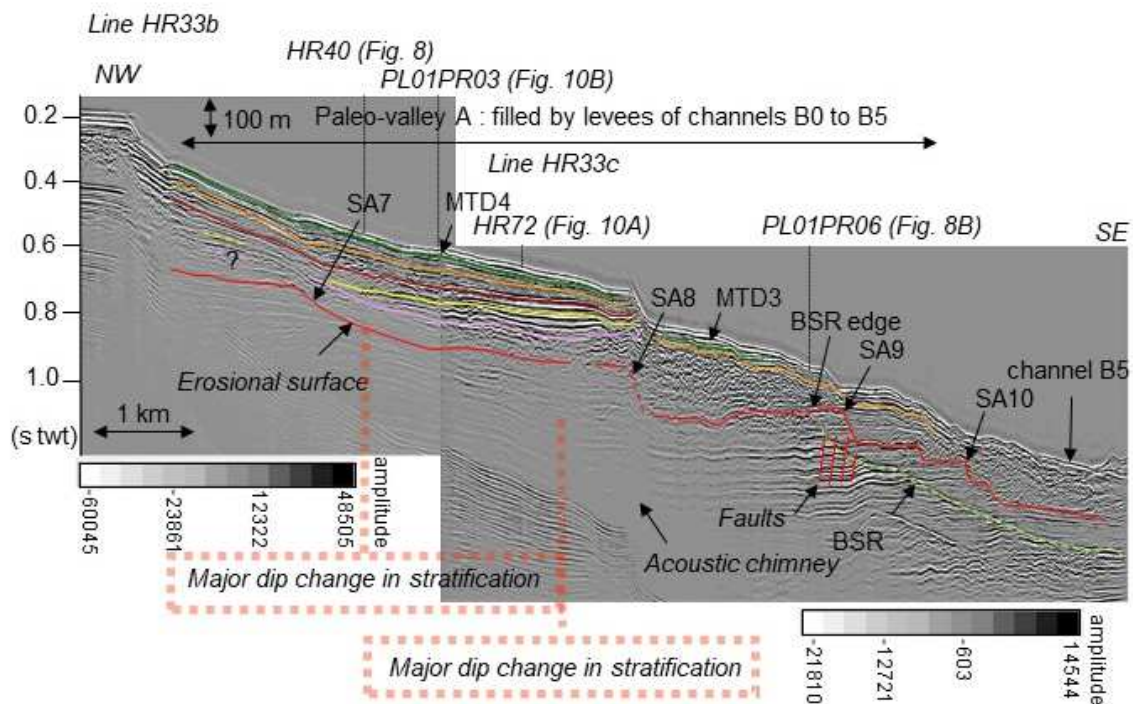


Fig. 9. Dip profile of Paleo-valley A and channel complex B. The high-resolution seismic Lines HR33b and HR33c (for location see Fig. 3B) show the cascading scarp alignments SA7 to SA10. Note the location of MTD4 and MTD3 near the sea floor and over respectively

intervals SA7-SA8 and SA8-SA9; the set of high-angle parallel faults showing small offset at the BSR edge and below SA9.

Wide MTDs lie directly over the erosional surface below the slope (see MTD12 on Fig. 4A and Fig. 4C) and at the bottom of paleo-valleys A and B (see section 4.3, Fig. 10; Fig. 16). The map of MTDs associated with the erosional surface show two MTDs of wide extent i.e. MTD12 and MTD13 (over MTD12) of at least respectively 300 and 75 km² in size and between 10 to 80 m in thickness (total volume > 50 km³) (Fig. 5 and Fig. 7B). The upstream limit of MTD12 correlates with scarp alignments (SAs) SA1 and SA7 (Fig. 7B). MTD12 extends over several SAs basinward. It is characterized by an accordion-like external shape with generally upper extensive parts at the scarp heads and compressive bulges several kilometers from the scarp heads (Fig. 4C; Fig. 6). This implies that the SAs correspond to slide scars and that a polyphase slide event disturbed the slope deposits.

We underline the presence of a deeper (older) erosional surface covered by MTDs (see section 4.3, Fig. 13A, erosional surface 2). This surface is cut by the basal regional surface described above and the erosive axis of paleo-valley B. No further description is possible as only a few seismic lines were available to image such deep structures.

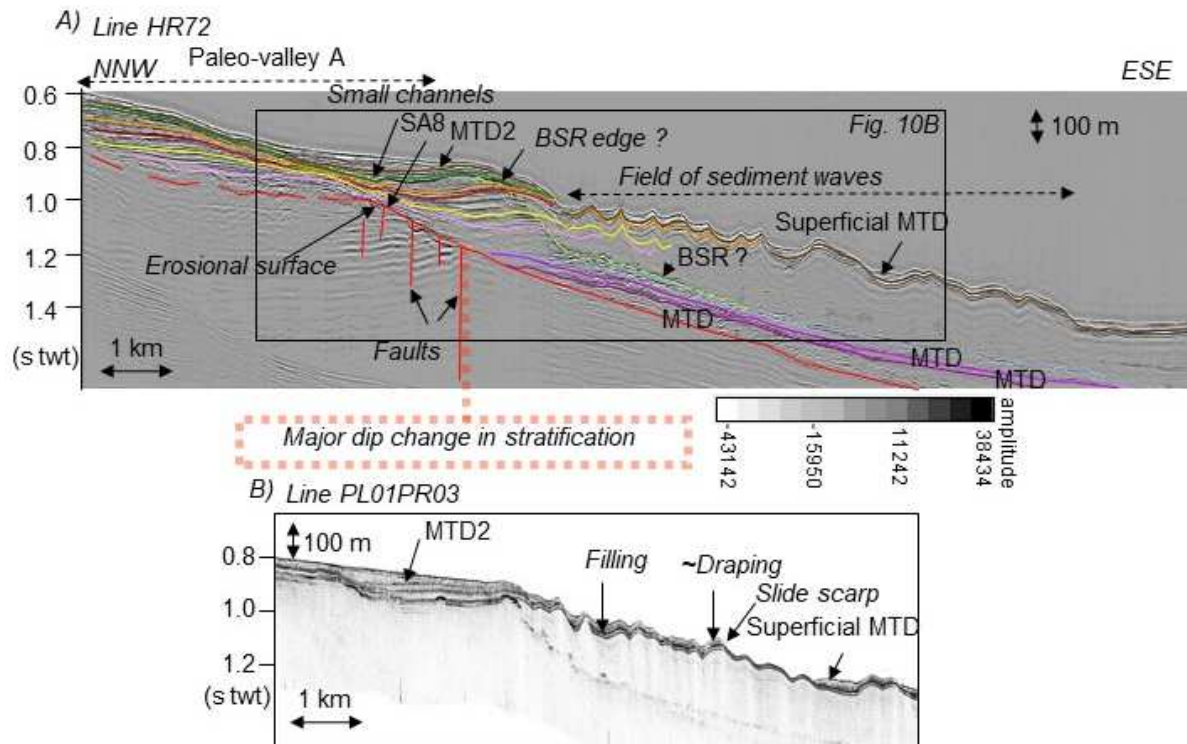


Fig. 10. MTDs and sediment waves field within paleo-valley A. A) The high-resolution seismic Line HR72 (for location see Fig. 3B) shows the MTDs overlying the erosional surface. Note the reworking/deformation of levees deposits of channels B0 to B5. The upstream reworking is characterized by stacked small channels and levees deposits and the downstream reworking corresponds to the sediment waves field. Note also the vertical faults at the BSR edge and below SA8. B) SYSIF profile PL01PR03 showing the sub-surface MTD2 and the superficial MTD associated to the slide scarp within the draping unit at the top of the filling unit.

4.3 Geometry and internal structure of paleo-valleys and channel-levee systems

Five to six main paleo-valleys - A, B, C, D, (Y) and the Danube connected to the Danube (Viteaz) canyon - characterize the region (Fig. 2, Fig. 3). The Danube canyon/valley has already been described by Popescu et al. (2004) as an erosional major trough with steep flanks which incises the shelf at around 200 m water depth. Partial infilling of the canyon

445 followed this initial erosional phase. The flat bottom of this through corresponds to the bottom
446 of the main valley and is incised by a V-shaped thalweg that continues beyond the shelfbreak
447 as an entrenched thalweg with well-developed lateral levees in the youngest Danube
448 channel (Popescu et al., 2004). For consistency, we refer in this study to this description to
449 define the terms of canyon, valley and channel-levees (see section 4.1 for the detailed
450 description of channels and levees).

451 The geomorphology of paleo-valleys B and D has already been described by Riboulot et al.,
452 (2017). Paleo-valley A is not clearly expressed on the seafloor. Paleo-valley C shows a wide
453 thalweg but its head is not clearly identified. The present seismic analysis has provided the
454 geometry and internal structure of paleo-valleys and associated channel-levee systems, as
455 well as their stacking pattern and the relative chronology as follows:

456 -Paleo-valley A appears over a U-shaped erosional trough of 1 to 4 km wide and ~150 m
457 (head) to ~90 m (distal part) deep possibly connected to canyon A. As mentioned previously,
458 the base of the trough is shaped by the erosional surface which is characterized by
459 numerous scarps. More specifically, the lateral scarps define the walls (Fig. 8; Fig. 7 scarps
460 in green) of the trough, and the basal cascading scarps SA7 to SA9 characterize the
461 staircase topography of the bottom along the dip profile (Fig. 9). Three main MTDs lie on the
462 bottom of its downstream part (Fig. 10). It is worth noting that valley A is filled by the stacked
463 western levees of channel complex B i.e. B0 to B5. In the valley axis, the deposits of these
464 levees are locally incised by small migrating channels (Fig. 8; Fig. 10). Downstream, over the
465 basal MTDs, the levee deposits are shaped like a field of sediment waves (Fig. 10). A buried
466 channel-levee system is observed off paleo-valley A and could correspond to the distal part
467 of a channel belonging to channel complex A which seems to have disappeared (Fig. 11,
468 channel A).

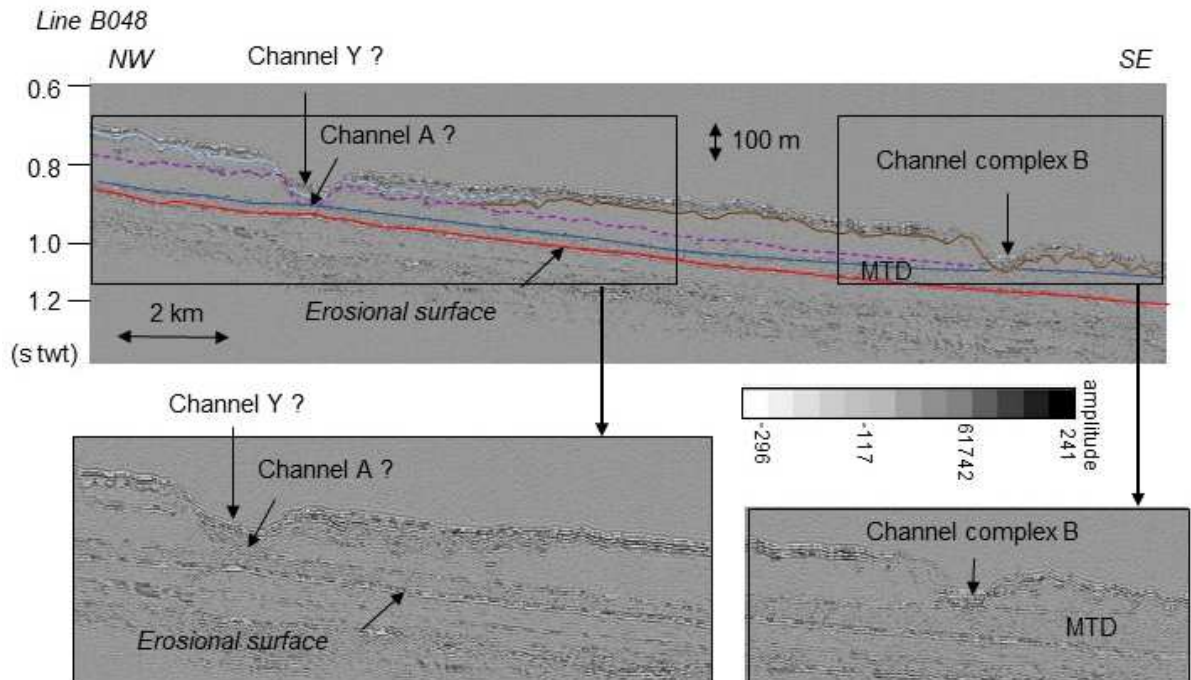
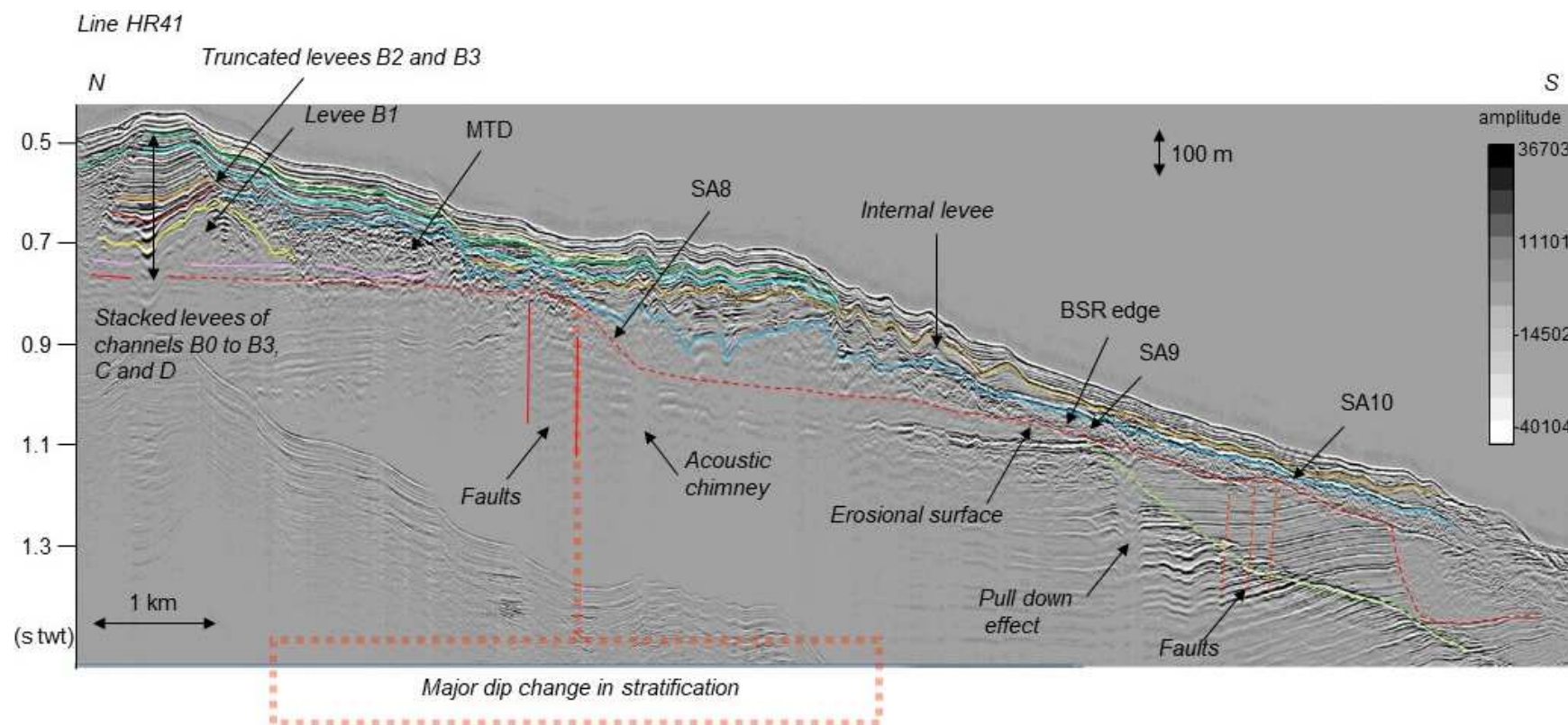


Fig. 11. Possible channels A and Y. The high-resolution seismic Line Blason B048 (for location see Fig. 3B) shows two superposed channels that could correspond from bottom to top to the channels A and Y. Note the stratigraphic location of channel complex B between these two channels.

-Canyon B (2.2 km wide and ~150 m deep) and its associated paleo-valley B are characterized by an axial thalweg (Fig. 3A) which is connected basinward to the youngest channel-levee system i.e. B5 (channel width of 800 m at ~ 1500 m water depth) of channel complex B (Fig. 8). As mentioned previously, an erosional trough, shaped by the basal unconformity and possibly connected to canyon B appears below this channel complex. (Fig. 13A). The singularity of channel complex B is the asymmetry of its walls (Fig. 8 and Fig. 13C). Indeed, the western wall is semi-linear whereas the eastern wall presents a staircase structure with two steps. The first channel B0 is located southwest of the present channel then channel B1 migrated northeastward of the present channel (Fig. 8). Channel B2 migrated to the position of the present channel. Channel B3 has remained in the B2 position; however, a collapse of its eastern levee has led to the formation of an elongated MTD along

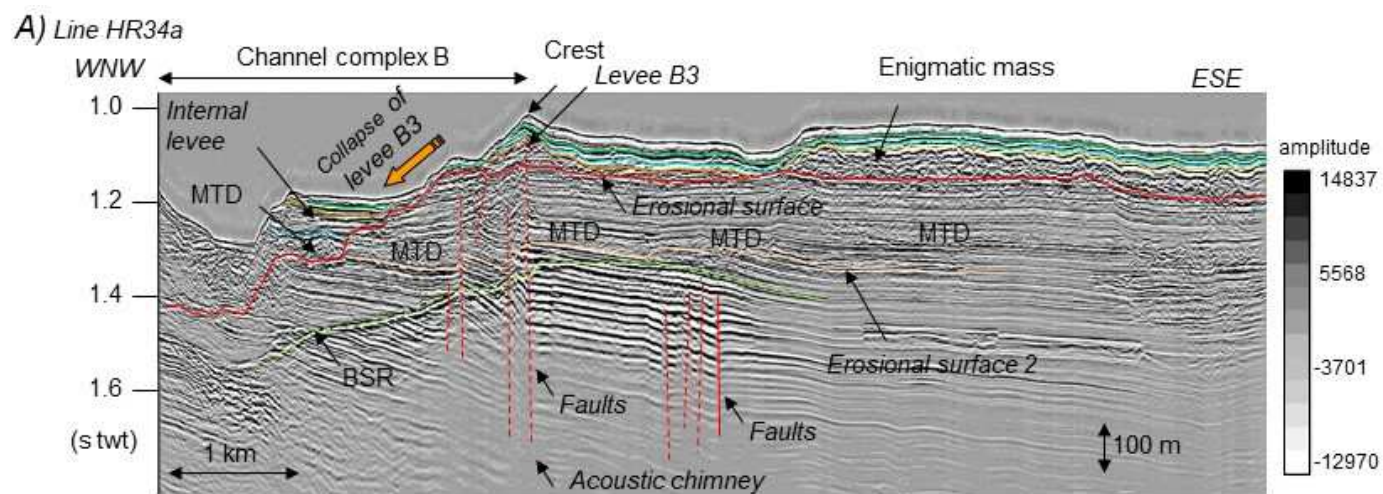
486 the eastern flank of paleo-valley B (Fig. 12). This explains the complex stair shape of the
487 eastern side of channel complex B (Fig. 13A). An enigmatic mass lies eastward of levee B3
488 and at the same stratigraphic level. This enigmatic mass shows chaotic seismic facies
489 suggesting deformed sediments but its external shape is regular (Fig. 13A and C). It has
490 been interpreted as belonging to levee B3. It is worth noting that the morphology of the so-
491 called Crest (Riboulot et al., 2017; Ker et al., 2019) results from the topography of levee B3.
492 Channels B4 and B5 have remained in the B3 position and have built their eastern levee
493 directly on the elongated MTD. This has led to the formation of an internal levee in the
494 eastern side of channel complex B. This internal levee reaches a length of 2.5 km and a
495 thickness of 50 m and shows very clear continuous wedging reflections (Fig. 13B, and Fig.
496 8B).

497

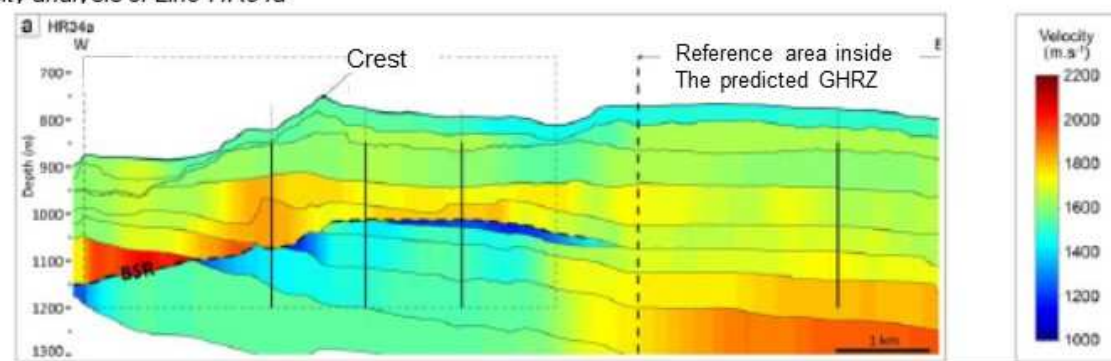


498

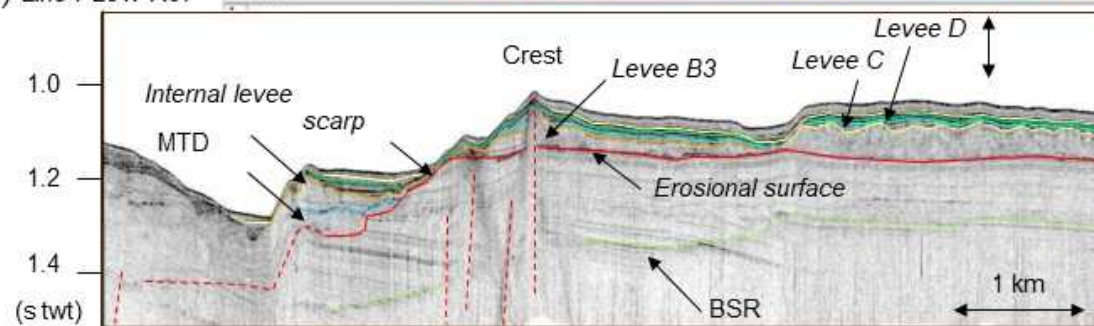
499 *Fig. 12. Dip profile of channel complex B. The high-resolution seismic Line HR41 (for location see Fig. 3B) shows the truncated levees B2 and*
 500 *B3 and the Internal levee coeval to channels B4 and B5 overlying the MTD resulting from the collapse. Basal MTDs on the erosional surface*
 501 *similar to those observed on Line HR72 (see Fig. 10A) are not excluded but the presence of gas does not allow distinguishing these MTDs.*

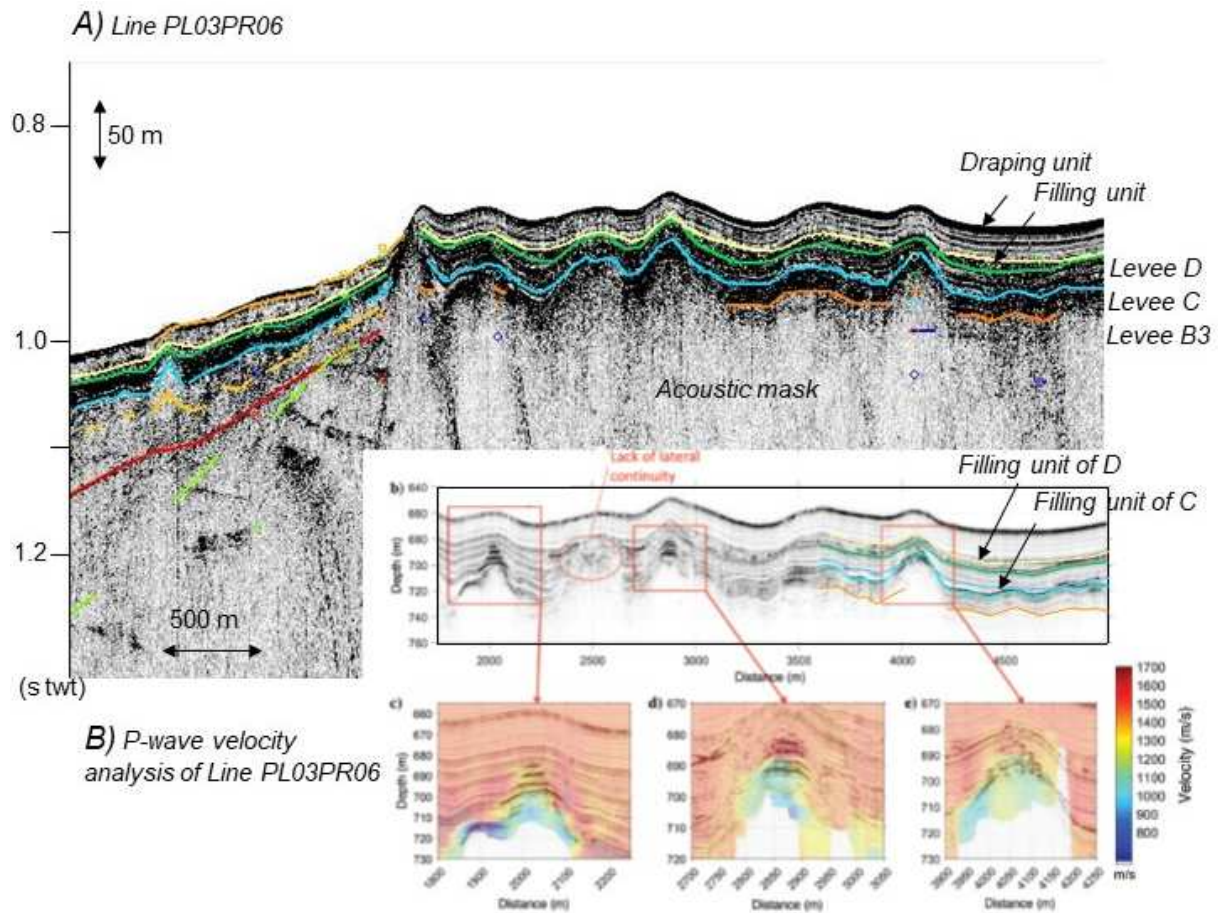


B) P-wave velocity analysis of Line HR34a



C) Line PL01PR07





519

520 *Fig. 14. Presence of gas within levees C and B3. A) The SYSIF PL03PR06 profile (for*
 521 *location see Fig. 3B) shows an acoustic mask at the top of levee B3 and chaotic facies within*
 522 *levee C. B) SYSIF profile migrated and Velocity analysis. The low-acoustic-velocity areas*
 523 *correspond to crests of sedimentary undulations of levee C filled with gas (Colin et al.,*
 524 *2020a).*

525

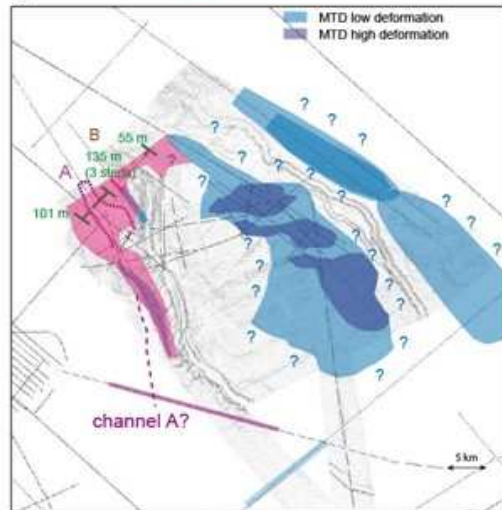
526 The map of the successive channel-levee systems B0 to the youngest shows their extent
 527 within paleo-valley B and on the slope of the study area (Fig. 15A-F). However, B0 to B2
 528 were difficult to map due to the scarcity of HR profiles. The two paleo-valleys C and D have
 529 developed northeast of paleo-valley B and have fed channel complexes (Fig. 3A). Only two
 530 channels of these complexes, called channel C and channel D, were studied. These
 531 channels lie directly over the widespread MTD12 and sometimes over-incise it. (Fig. 15A).

The levees of channels C and D have a wide extent (12 km) and well-stratified deposits and their distal parts overlap the internal levee coeval to channels B4 and B5 (Fig. 8). These levees have been mapped using all seismic datasets (Fig. 15G-H).

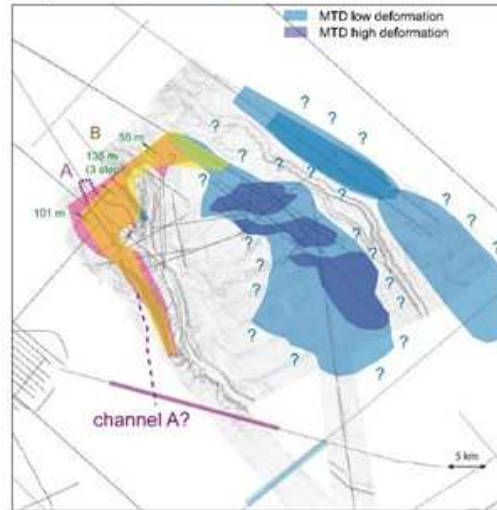
- Paleo-valley Y is located at the seafloor between the Danube canyon and paleo-valley A (Fig. 2; Fig. 3A). Channel Y appears over channel A and over channel complex B (Fig. 11). However, the seismic data are insufficient to map channel Y and to determine its location either between channel complex B and channel C or between channel D and the youngest Danube channel.

- Paleo-valley Danube is located southwest of paleo-valley A. The most recent Danube channel-levee system is observed from the slope to the basin (unit 4 from Popescu et al., 2001) and crosscuts the distal parts of channel complexes B to D (Fig. 2; Fig. 17). This observation implies that the most recent Danube channel is younger than these channel complexes, in agreement with Hillman et al. (2018). Westward, the most recent Danube channel appears over possibly channel A and channel Z. According to Hillman et al. (2018), Z would be below the erosional surface and very old (> 1 Ma) (Fig. 17).

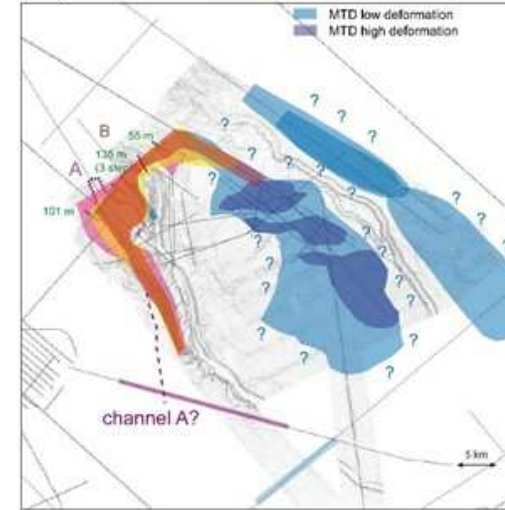
A) channel B0



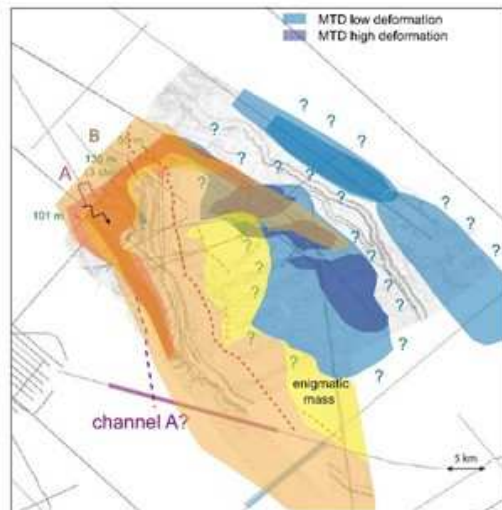
B) channel B1



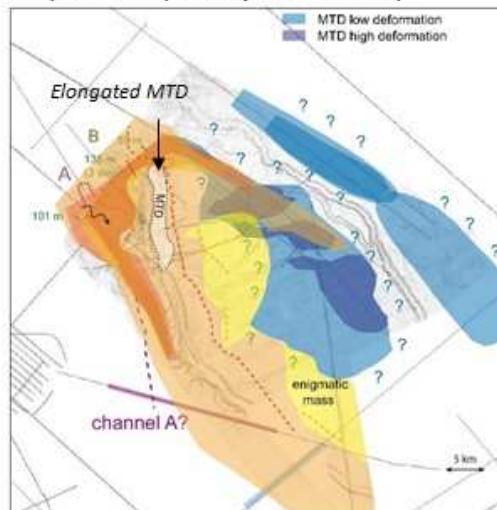
C) channel B2



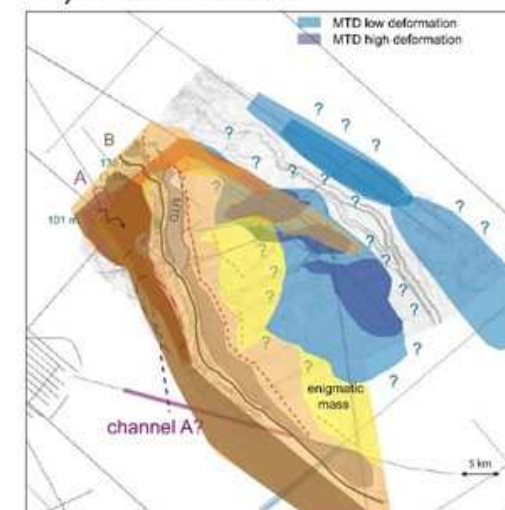
D) channel B3 and Enigmatic mass

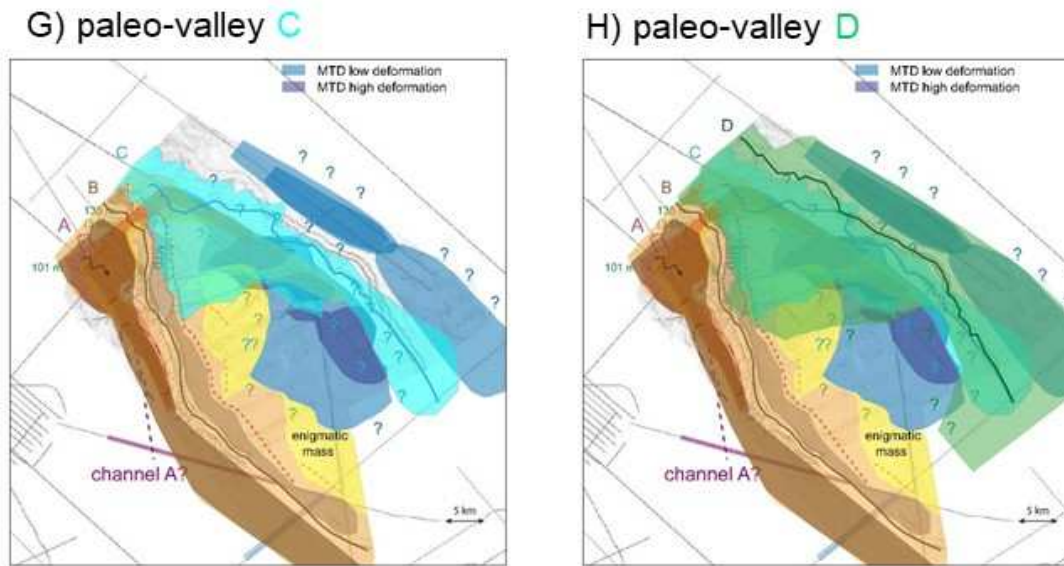


E) MTD (collapse of B3)



F) Internal levee



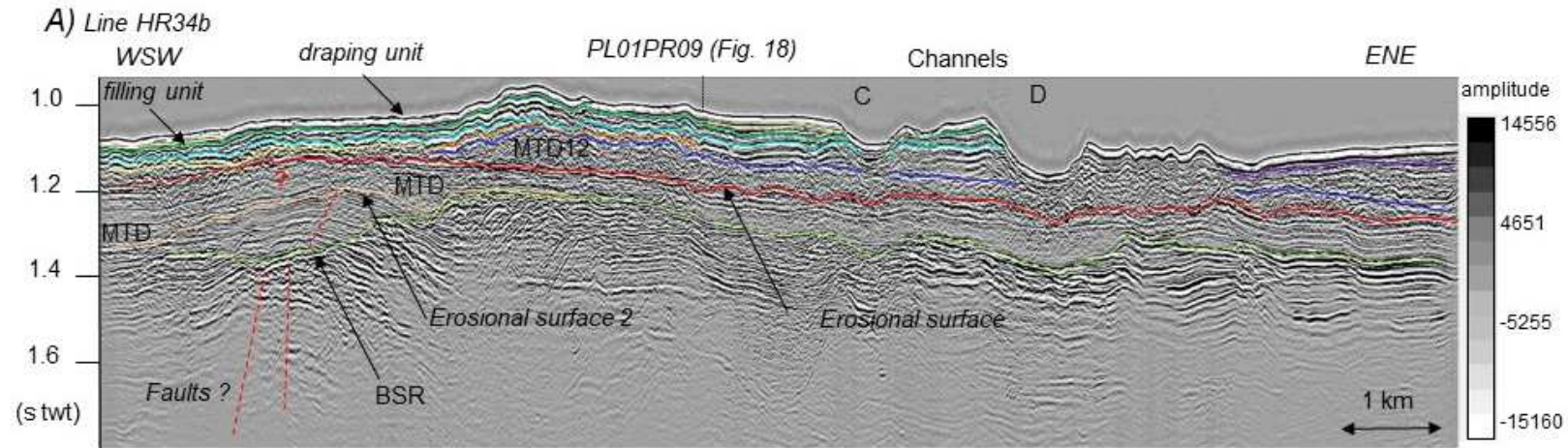


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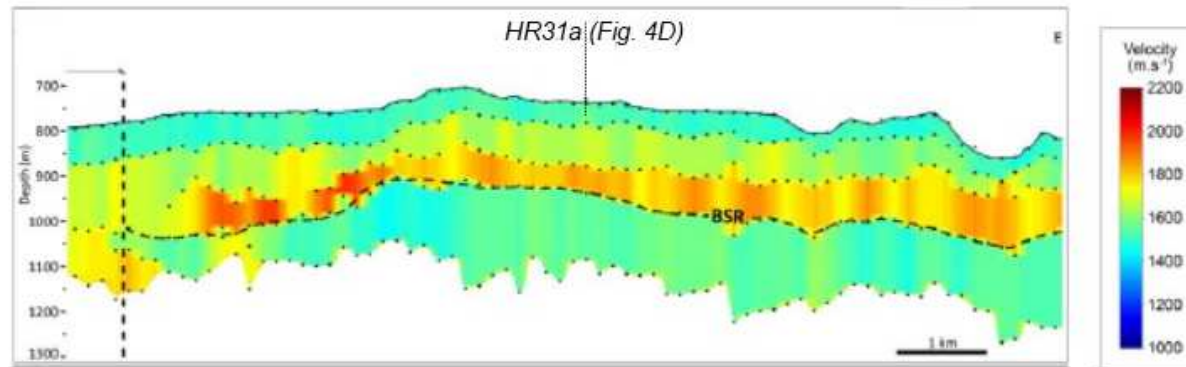
555 Fig. 15. Extent of channel-levees systems of Paleo-valley A, channel complex B, C and D (see location of the map on Fig. 3). MTDs on the
 556 erosional surface (blue polygon) and lateral scarps of erosional troughs A and B have been reported (green color). The axes of channels are
 557 marked by a line of the same color as the color of the levees. The levee extent is indicated by a colored polygon. The channel-levees are
 558 successively stacked on the different maps. A) Valley A and possible channel A (purple color) Channel-levees B0 (pink color); B) Channel-
 559 levees B1 (yellow color) over B0; C) Channel-levees B2 over B0 and B1; D) Channel-levees B3 over B0, B1 and B2 and Enigmatic mass
 560 interpreted as belonging to the levee B3; E) elongated MTD resulting from the collapse of B3 eastern levee; F) Internal levee on the elongated
 561 MTD and channel-levees B0 to B3. G) Channel-levees C (turquoise blue color); H) Channel-levees D (green color). The maps are based on

562 only HR lines for channel-levees B0 to B2 and on all seismic profiles (HR, SYSIF, SDS) for channel-levees B3, C, D, the elongated MTD and
 563 the internal levee.

564

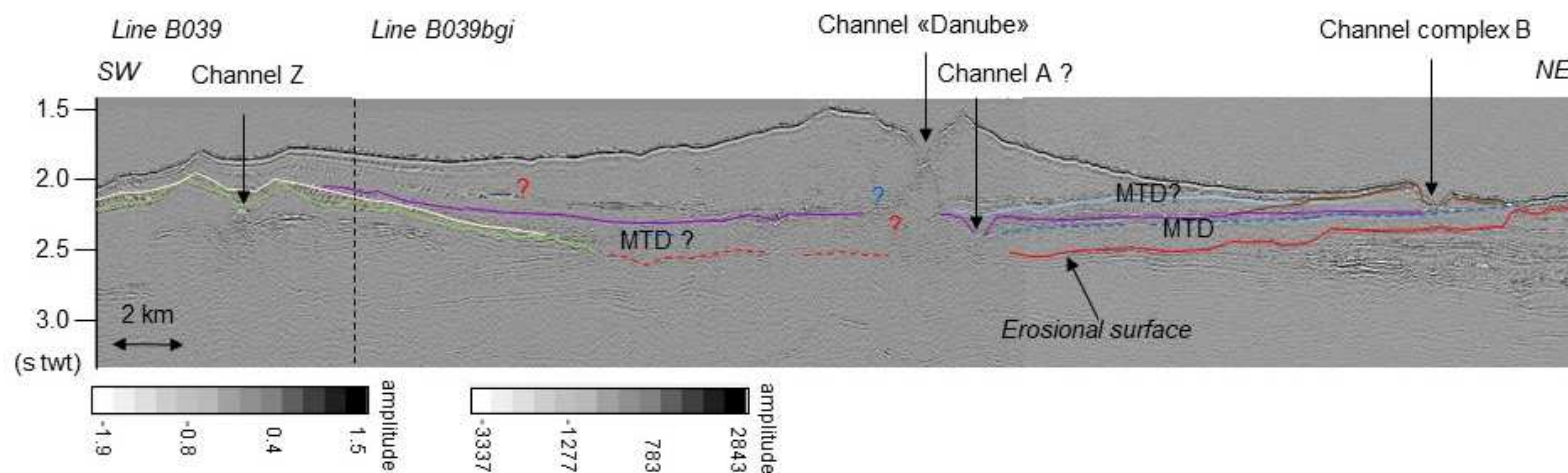


B) P-wave velocity analysis of Line HR34b



565

566 Fig. 16. Cross profile of channel-levees C and D. The high-resolution seismic Line HR34b (for location see Fig. 3B) shows these channel-levees
 567 which lie directly on MTD12.

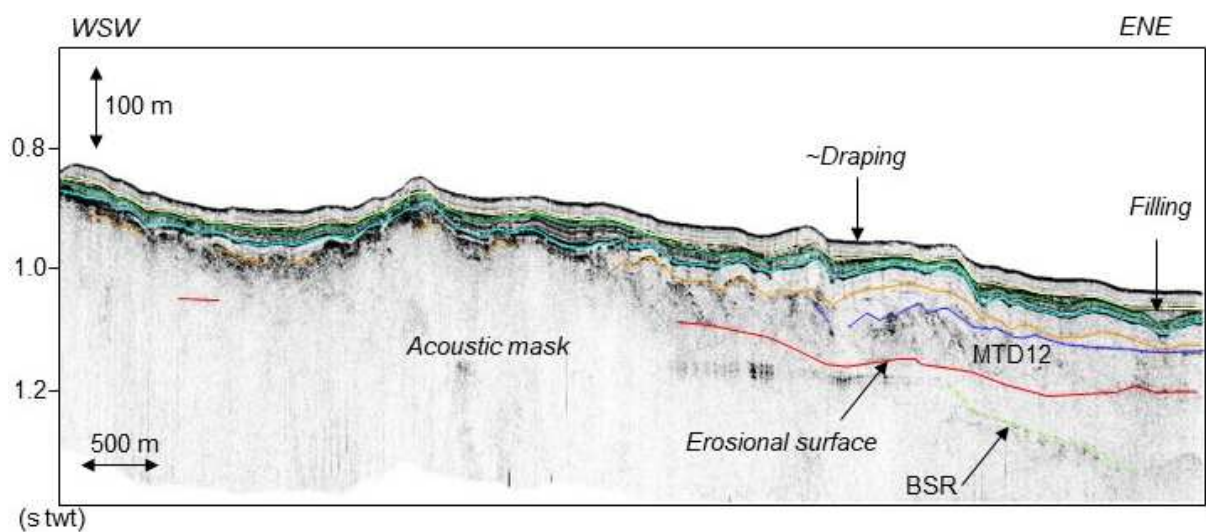


570
 571 Fig. 17. Cross profile of channel “Danube”. A) The high-resolution seismic Line B039bgi (for location see Fig. 3B) shows the youngest channel-
 572 levee “Danube” which lie, respectively from SW to NE, over an infill unit of channel Z, possibly channel A, an MTD (in light blue) and, channel
 573 complex B. Note the erosional surface and overlying MTD (in dark blue).

574

575 The chronology of paleo-valleys and their connected channel complexes is based on the
576 principle of superposition. Channel complex B crosscuts paleo-valley A and is partially
577 covered by the distal part of the levees of channels C and D (Fig. 8). The activity of paleo-
578 valley A is continuous during the construction of channel complex B as revealed by the small
579 channels affecting levee deposits to the seafloor where channel A crosscuts channel
580 complex B (Fig. 3A). Channel D crosscuts channel C (Fig. 15H). The resulting chronology of
581 paleo-valleys is, from the oldest to the youngest, A, B, C, D, and the most recent, the
582 Danube paleo-valley which is connected to the youngest Danube channel.

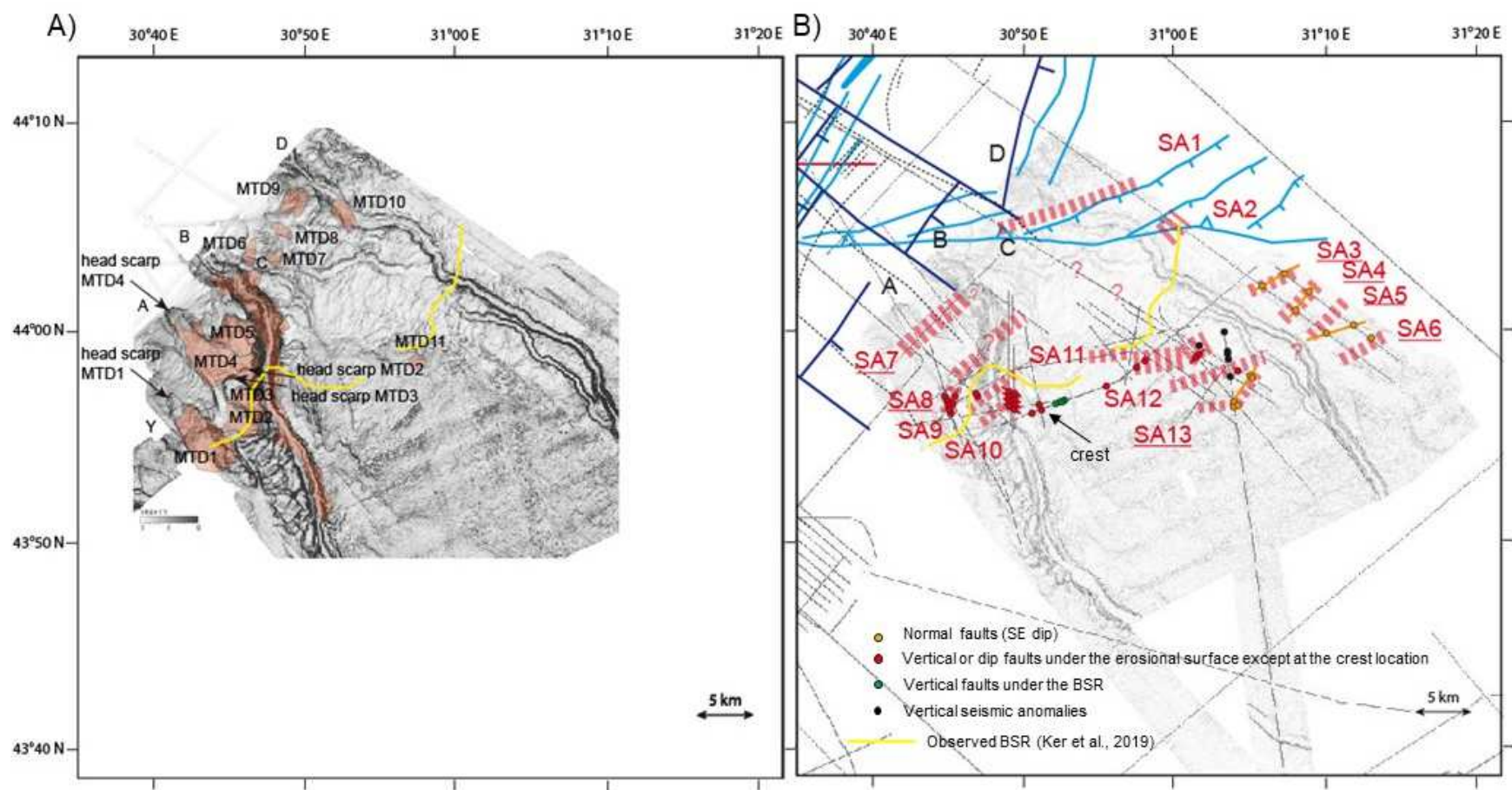
583 The end of activity of paleo-valleys A, B, C, and D is marked by a unit of up to 30 m
584 thickness showing low amplitude infilling seismic facies (Fig. 18). This infilling is attributed to
585 late-stage distal overbank deposits from younger channels, such as possibly channels Y and
586 Danube. The study area is covered by an almost draping unit (Fig. 18). This unit has been
587 described from the core GAS-CS01 (GHASS survey) by Martinez-Lamas et al. (2020) as
588 laminated clayey sequences that resemble the lacustrine varves of Degens et al. (1978).
589 These lacustrine sequences have been interpreted as hyperpycnites i.e. hyperpycnal
590 turbidites produced by discharges of the Danube river at the end of the glacial period (ca.33-
591 17 ka) (Martinez-Lamas et al., 2020). We note that only 1 m of marine sediments has formed
592 since the last connection to the Mediterranean Sea.



593

Fig. 18. Filling and draping upper sedimentary layers. The SYSIF PL01PR09 profile (for location see Fig. 3B) shows the unit (light yellow color) filling the underlying topography of the top of levee D (green color). The overlying draping unit outcrops at the seafloor. These two upper units act as a seal for upward gas migration as gas flares do not occur in this area (see Fig. 3A).

4.4 Surficial sediment instabilities



603

604 Fig. 19. Superficial MTDs and Fault pattern (see location on Fig. 3). A) Map of the superficial MTDs observed on SBP, SYSIF and/or HR
 605 seismic (red polygons). MTD1 appears within paleo-valley Y, MTD2, MTD3 and MTD4 within paleo-valley A with a retrogressive pattern, MTD5
 606 is located between paleo-valleys A and B, MTD6 to MTD10 lie on the upper slope, and MTD11 on the middle slope. Note the MTDs on the walls
 607 of paleo-valley B

608

609 As mentioned previously (section 2.3), the seafloor surface is disturbed by scarps and
610 associated MTDs linked to instabilities. Numerous MTDs disturb the superficial sediments of
611 the headwalls and lateral walls of canyons/paleo-valleys A and B. The walls of channels
612 along their entire length are also disturbed. Eleven MTDs located on the upper slope have
613 been identified on seismic profiles and mapped (Fig. 8B, Fig. 19). In paleo-valley A, three
614 distinct MTDs (MTD2 to MTD4) and their associated head scarps have been identified.
615 MTD2 is at around 10 mbsf, reaches 35 m in thickness, and 2 km² in extent (Fig. 10 and Fig.
616 8B). MTD3 is at around 2 mbsf, reaches 25 m in thickness, and 3 km² in extent (Fig. 8B).
617 MTD4 is at the seafloor and extends over a 9 km² area (Fig. 9). The three distinct MTDs
618 suggest three main episodes of instability in a retrogressive pattern. It is worth noting that
619 MTD5 could have occurred between MTD3 and MTD4 but this has not been identified in the
620 seismic data (Fig. 19). Superficial MTDs can appear at the top of the field of the sediment
621 wave revealing a decollement surface at the base of the draping unit (Fig. 10B).

622

623 4.5 Gas signatures and hydrates in upper sediments

624

625 As mentioned previously (section 2.3), the BSR occurs between ~700 and 1830 m water
626 depth and its edge, i.e. its most landward termination (see Fig. 10 for an example), ranges in
627 the predicted GHSZ (Popescu et al., 2004; Ker and Riboulot, 2015; Riboulot et al. 2017) (Fig.
628 3A). Gas signature is generally absent within the sediments located above the BSR (e.g.,
629 Fig. 5C), as confirmed by the velocity analysis of SYSIF data (Colin et al., 2020a) (Fig. 4D),
630 and the absence of gas flares at the seafloor (Fig. 3A). However, there is one exception
631 along the so-called Crest area where faults cross the BSR and, outcrop at the seafloor and
632 feed gas flares (Ker et al., 2019). Velocity analysis of SYSIF data (Colin et al., 2020a) shows
633 the presence of gas below the BSR under the Crest (Fig. 13B). These observations highlight
634 that gas hydrates play the role of an efficient seal layer in controlling gas-flare distribution.

Under the BSR, rare observable vertical columns of seismic anomalies with a pull-down effect, are interpreted as a localized accumulation of gas implying potential gas breaks through the sediments (Fig. 6 below SA12; Fig. 12). Others seem to show a slight vertical offset of reflectors suggesting potential faults. Acoustic chimneys also occur below the BSR and mainly westwards of channels C and D (Fig. 6).

In the free gas area at the landward termination of the BSR, most of the columns of seismic anomalies generally associated with underlying acoustic chimneys occur in levees associated with channels B to D (Fig. 4B). They do not reach the seafloor but stop at the level of the infilling unit and overlying draping lacustrine and marine sediments (Fig. 18). Velocity analysis revealed the presence of free gas in the crests of sedimentary undulations (Colin et al., 2020a). In our study, these crests have been correlated to levees B and C. (Fig. 14B). This confirms that the draping unit, and possibly the infill unit, bounds the upper limit of the gas front and consequently plays the role of a seal layer. Most of the gas flares are in association with sedimentary discontinuities such as the edge of canyon B and the paleo-valleys and channel sidewalls (B, C and D), and slide scars (Fig. 3A). This is reminiscent of Skarke et al. (2014).

4.6 Fault patterns and distribution

A couple of faults have been identified from seismic profiles (Fig. 19B). They appear as single faults or sets of faults. The single faults are sometimes listric (Fig. 5B), with some of them showing a normal vertical displacement of seismic reflectors (Fig 6, below SA13). The sets of faults include small-offset and high-angle to vertical faults (Fig. 6, below SA1; Fig. 9, below SA9). They are sometimes associated with acoustic chimneys which suggest the presence of gas (Fig. 13A, below the so-called Crest). Some faults coincide with a major dip change in stratification on both sides of the vertical columns of seismic anomalies (Fig. 6, below SA13; Fig. 10, below SA8).

These faults are not randomly distributed in the study area or the sediment stratigraphy (Fig. 19). Almost all these faults affect sediments below the basal regional unconformity. Only the fault system of the Crest seems to have propagated to the seafloor within the surficial sediments (Fig. 13). This fault zone feeds gas flares and seems deep-seated even if its termination was not identified due to lack of penetration of the seismic signal (Ker et al., 2019). The eastern area is characterized by deeply rooted normal faults below the scarps of the basal unconformity. These faults sometimes become listric toward the BSR zone (Fig. 5B; Fig. 6). Sets of vertical faults occur below the BSR and at the BSR edge (Fig. 13A; Fig. 9, below SA9 and Fig. 10, below SA8), but few examples are observed cutting it (Fig. 12, below SA10). We note that this set of vertical faults seems to limit a free-gas zone below a marked BSR from a diffuse-gas zone without a BSR from either side (Fig. 13A). Upstream the BSR edge, the presence of gas facies and acoustic chimneys on seismic profiles does not allow us to identify faults directly (Fig. 4C). However, the abundance of faults reported on the shelf (see Fig. 3A) infers the presence of faults within the free-gas area (Fig. 19B). Precise characterization of fault orientation is not possible in the study area with such a low coverage of HR seismic acquisition (Fig. 3). In the western zone, only a set of small offset faults with a NW-SE orientation can be determined from the correlation of closely spaced and parallel seismic lines along the east side of valley B (Fig. 19). We note that most of the vertical structures are observed from NE-SW seismic lines which could be consistent with transfer structures having a rough NW-SE orientation. In the eastern zone, the high-angle normal faults have NE-SW-oriented alignments.

5. Discussion

On the Dnepr paleo-delta in the NW Black Sea, it has been demonstrated that morphology, underlying stratigraphy, sediment properties, and the presence of gas hydrates control the distribution of gas flares (Naudts et al., 2006). Our results are similar. We show that the

extent of the BSR, faulting patterns, and arrangement of architectural elements such as canyons, channel-levees, slide scars, and MTDs control the distribution of seal layers and sedimentary discontinuities and consequently gas-flare distribution. However, we cannot exclude that seepage may be primarily associated with tectonically driven flows through regional faults (Kruglyakova et al., 2004). Moreover, the release of seismic energy during earthquakes could cause slope failures and subsequent gas escape along slide scarps. Further investigation of the deep fluid plumbing system and sediment properties remains essential to discuss the factors controlling gas/fluid migration pathways and consequently gas-flare distribution.

This work highlighted the evolution of the tectonic-sedimentary architecture of surficial deposits and its impact on sediment instabilities, which is discussed below.

5.1 Origin of the Basal erosional surface and sediment instabilities

The basal erosional surface with the associated layer of MTDs is observable from the upper continental slope down to the deep basin (Popescu et al., 2004; Winguth et al., 2000; Hillmann et al., 2018). This erosional surface corresponds to the Base Neoeuxinian Sequence Boundary of Hillman et al. (2018) attributed to the Würmian with a minimum lake/sea level drop (about -130 m) at around 18000 BP (Panin and Jipa, 2002). The Base Neoeuxinian Sequence Boundary has been recently dated using ¹⁴C dating, at a minimum of ~22 ka (Lericolais et al., 2013) and was considered to mark the onset of activity in the Danube Fan during the last sea-level lowstand in the Black Sea (Popescu et al., 2001). However, the correlation with core Gas-CS01 (base of the draping unit) indicates the age at a minimum of 32 ka (Martinez-Lamas et al., 2020). The presence of several built and abandoned channel-levee systems between this draping unit and the basal erosional surface suggests that this unconformity is significantly older.

It is generally considered that during sea-level falls, migration of the river-mouth to the shelf edge and rapid deltaic sedimentation leads to an increase in sedimentation load,

overstepping and overpressuring in the prodelta and slope sediments, resulting in sediment failure, MTD formation, and canyon development (Weimer, 1990; Posamentier and Kolla, 2003, Madof et al., 2009). A link between the MTDs, the scarp alignments of the basal erosional surface, and the faults has been observed (Fig. 7B). The presence of normal faults under SA3, SA4, and SA6 suggests that these sediment instabilities are related to normal faults underneath, potentially of gravitational origin. Major dip changes in stratification at SA7, SA8, and SA13 are also likely attributed to faults since these scarps are almost parallel to the mapped SW-NE oriented faults from Konerding et al. (2010) (in dark blue on Fig. 19B). We highlight that the observed faults are mainly located below the basal unconformity and are mainly absent in the surficial deposits. This observation suggests that the faulting was predominantly active before or during basal erosional surface formation. We also note that the upper extensive parts of the MTDs lie over the scarp alignments whereas the compressive bulges are downstream (e.g. Fig. 4C). The contribution of fault activity is therefore identified as significant on slope failures and distribution of MTDs.

Sea-level falls may cause an additional effect, such as a reduction in hydrostatic pore pressure in the fluid reservoir that lowers gas solubility and results in gas exsolution. The subsequent accumulation of gas in porous layers interbedded within less permeable layers prevents fluid escape (Magara, 1978; Lafuerza et al., 2009) and facilitates overpressure build-up, which drives fluids along permeable migration pathways (e.g. faults), and may cause a drop in the effective stress of the sediment and favour its failure (Bolton et al., 1998; Dugan et al., 2003; Canals et al., 2004). Reactivation of inherited faults is also possible and the subsequent fluid migration along them may contribute to the initiation and location of surficial sediment instabilities. This potential fault reactivation during sea-level fall does not exclude deep movements of over pressured layers at the regional scale. Indeed, examples of gravity-driven deformations have been demonstrated on the Plio-Pleistocene time scale in several deltaic settings characterized by underlying over pressured layers (dos Reis et al., 2005 for the Gulf of Lion; Knox and Omatsola, 1989 and Rouby et al., 2011 for the Niger delta). We note that regional-scale tectonic influence, such as the flexural uplift of the

Carpathians foreland due to the Wallachian deformations, cannot be excluded from this fault activity along the Romanian margin (Cloetingh et al., 2003).

It is worth noting that the presence of an older erosional surface (erosional surface 2 in Fig. 13A) with associated MTDs, suggests a recurrence of significant instabilities possibly during successive sea-level falls. Glacial and interglacial conditions control the alternating turbidite/hemipelagic deposits and may explain the recurrence of slope failures. This has been demonstrated for the Gulf of Lion where repeated slope failures occurred during glacial periods at specific stratigraphic levels which correspond to weak layers between clay-rich and turbidite sediments (Badhani et al., 2019, 2020).

5.2 The channel-levee systems: Impact of fault systems and instabilities

The data highlight the build-up of channel-levee systems above the basal layer of MTDs and the Neoeuxinian sequence boundary (Fig. 15). This suggests that fluvially derived sediments were transported directly into canyon heads and then as channelized turbidity flows thanks to the connection between fluvial valleys incised in the shelf and canyons. The vertical succession from MTDs to channelized deposits is reminiscent of the “deep-water sequence” of Weimer (1990). The model of Winguth et al. (2000) for the Danube and Dniepr fans can be applied to the study because it is based on the depositional model of Weimer (1990) for deep-sea fan systems, in which sea-level fluctuation controls fan development. We have integrated the tectonics and gas/gas-hydrate dynamics which contribute to sediment instabilities, into the model of Winguth (2000) and propose a diagram of the evolution of the study area since the last sea level fall (Fig. 20). During the sea-level fall, gravity-driven tectonic and/or fault reactivation with subsequent gas expulsion are invoked to trigger major sedimentary instabilities of the slope leading to the formation of MTDs (Fig. 20A). Then during lowstands, the fan system becomes the main site of deposition, and coarse material is confined to the channels, whereas finer material is swept onto the levees and beyond to form overbank deposits (Fig. 20B).

The relative chronology of paleo-valleys (see section 4.3) shows that the turbidite activity migrated northward from paleo-valleys A to D and then southward from paleo-valley D to the Viteaz canyon. We suppose the low sedimentary load on the shelf to be insufficient to induce gravity-deformation and subsequent fluid/gas expulsion.

Sediment instability during canyon activity is related to both erosive action within the canyon (sediment flow dynamics) and the presence of gas in surficial deposits (high sedimentation rate or migration from a fault system). Sediment instabilities were observed during canyon activity and formation of the erosional surface (Fig. 7B) but also during the formation of channel-levee systems (Fig. 8). According to Hillmann et al. (2018), the slope failure located on the western wall of paleo-valley B and labeled slump S2 by these authors shows no evidence for deposits downslope or in the canyon that could be related to this event. This suggests that the canyon was active at the time of slope failure. Our results show the same phenomenon for paleo-valley A, for which the complete removal of sediments within this valley suggests strong turbiditic activity during slope failure. Undercutting at the toe of a slope due to erosive action of turbidity currents in canyons may result in the axial incision of the canyon floor, triggering slope failure. Overpressure generation due to rapid deposition of slope sediments may have also contributed to slope instability (Hillman et al. 2018). It is likely that, in the Viteaz (Danube) turbidite system, the location of gassy facies within the canyon and the upper part of the associated channel favored instability and possibly triggered sediment failures (Popescu, 2002, e.g. Egorov et al., 1998 for the Dnepr Canyon). Gas migration from faults could also contribute to this fluid overpressure and related sediment instability in canyons. The collapse of the eastern levee of channel B3 (Fig. 12) and the sedimentary deformation of the levee, i.e. the enigmatic mass (Fig. 13), are suspected to be attributed to the gas flow observed from the Crest fault system. We, therefore, hypothesize that the combined effect of significant turbidite activity (at least in paleo-valleys A and B) and the presence of gas at the walls of canyons and valleys influences sediment stability and the concurrent development of canyons and/or paleo-valleys A to D (Fig. 20B).

5.3. Surficial sediment deposits and instabilities

The surficial infilling and draping units mark the abandonment of canyon-channel systems in the study area and migration of the turbidite activity to the south, i.e. within canyon Y and/or Danube (Viteaz) canyon/valley. These units record the last evolution of the Danube river activity linked to the Alpine ice-sheet fluctuations during the end of the last glacial 33-17 ka BP (lacustrine deposits, Martinez-Lamas et al., 2020), the last connection to the Marmara-Mediterranean Sea with sapropel and carbonates (Constantinescu et al., 2015) and marine sedimentation. This evolution from channel filling to marine sedimentation suggests a retrogradation of the deltaic system towards the coast during the sea-level rise and is consistent with the conventional sequence stratigraphic models (Mitchum and Vail, 1977; Vail et al., 1977; Van Wagoner et al., 1988) (Fig. 20C).

Some MTDs are recorded within the draping unit along the upper slope, within the free-gas area, and in zones favorable to failure initiation such as the flanks of paleo-canyons and the sediment-wave area (Fig. 19; Fig. 10). According to Lericolais et al. (2009), the Melt Water Pulse 1A at 12.5 ka ^{14}C BP is the source of water level rise between -40 to -20 m below the present-day sea level. Then, during the low water level stage from 11 to 8 ka ^{14}C BP (i.e., younger Dryas), a lacustrine sedimentary wedge system prograded on the Romanian shelf to the Danube canyon head. It is likely that this enhanced sediment supply onto the shelf led to abrupt overloading generating a new and short episode of gravity-deformation and subsequent fluid/gas expulsion underlying slope instabilities.

Sea level has been rising and the Danubian sediment supply decreasing since around 8 kya (Panin and Jipa, 2002). The present-day depocenters in the region are the Danube delta front and prodelta areas on the internal shelf i.e. to 50-60 m depth. According to Winguth et al. (2000), during highstands, the deltaic region is far removed from the shelf edge and is the depocenter for much of the fluvial sediment input. The slope and basin are sediment-starved so that a condensed section is laid down as a result of hemipelagic deposition (Fig. 20D). Sea level is still rising and outpaces subsidence within the coastal area of the Danube delta

(Panin et al., 1983). This suggests that sedimentary load on the shelf is too low to induce present-day gravity-driven deformation. However, rapid sea-level rise is well known to induce bending stress on the continental slope, which can lead to elevated seismicity and potential fault reactivation that can consequentially trigger landslides (Brothers et al., 2013). This could explain the origin of sub-surface MTDs, i.e., those buried under marine sediments (around 1 m thick from GAS-CS01, see section 4.3).

5.4. Potential role of seismicity and gas-hydrate dynamic on sediment instability

Of the mechanisms which can trigger sediment instabilities throughout geological recording are the seismic activity linked to the regional faulting pattern, the presence of gas in the sediment and the dissociation of gas hydrate (see Garziglia et al., 2008 for similar investigation for the Nile delta).

In the study area, earthquakes of magnitude > 7 have occurred along the Romania-Bulgaria coast and could have caused submarine slides and consequential tsunamis. For example, two earthquakes of magnitude = 7.1 and 7.2 were reported offshore NE Bulgaria (Papadopoulos et al., 2011), and near the Sugar canyon (Matova, 2000), respectively (Fig. 2 and Table 1). In the study area, a magnitude 3.9 earthquake occurred in 2007 at the location of the W-E oriented fault (Fig. 2 and Table 1.). Two other earthquakes of magnitudes 4.6 and 4 occurred further offshore the study area in 1986 and 2009 respectively. Despite their low magnitude, these earthquakes could have triggered surficial sediment instabilities, especially along the valley flanks such as for paleo-valleys A and B and also the deformation of levee B3, i.e. the enigmatic mass. Earthquakes of higher magnitude could occur in the study area. The role of paleo-seismic activity in the initiation of MTDs is speculative but cannot be excluded, especially as scarp alignments and faults are spatially linked.

Another mechanism that can trigger sediment instabilities is gas hydrate dissociation and the subsequent free gas release (Bouriak et al., 2000; Bünz et al., 2005; Kwon and Cho, 2012). Several factors lead to gas-hydrate dissociation such as sea level fall (Haq, 1998; Maslin et

al., 1998, 2004; Kennett et al., 2003), penetration in the Black Sea of warmer Mediterranean water during the rapid transgression after the Last Glacial Maximum (Zander et al., 2017; Matenco et al., 2016), or salt diffusion within the sediment (Riboulot et al., 2018). Potentially, the superficial sedimentary instabilities close to the BSR edge such as MTD1 are provoked by gas hydrate dissociation (Fig. 19A). We also note the presence of normal faults which seem to propagate listrically toward the BSR. They are then potentially related to free gas release following the upward migration of gas hydrate since the last glacial maximum (Fig. 5, below SA5) (Dillon et al., 1998). Listric faults can also induce slope instabilities, a process that shaped these basin margins during Miocene-Pliocene times (Fig. 1B, Munteanu et al., 2012). The field of sediment waves could also be interpreted as sediment destabilization by dissociation of gas hydrate (Fig. 10). Indeed, the substratum of sediment waves corresponds to the BSR. Moreover, the sets of vertical faults which coincide with the BSR edge (e.g. Fig. 9, below SA9) are more likely to be fractures linked to free-gas overpressure at the landward termination of the GHSZ than tectonic structures. We add that, conversely, these fractures can increase free-gas migration during BSR evolution and help gas-hydrate dissociation along the landward termination of the GHSZ (Yang and Davies, 2013). In the Dniepr paleo-delta, northward of the study area offshore Crimea, the role of gas hydrate dynamics on slope failures remains unclear (Zandler et al., 2018). Nevertheless, because mass-wasting processes occur on channel-levees upstream the GHSZ, slope failure could have been facilitated by gas-hydrate destabilization (Naudts et al., 2006). As gas hydrate dissociation is still in progress, the slope studied here is gradually weakening.

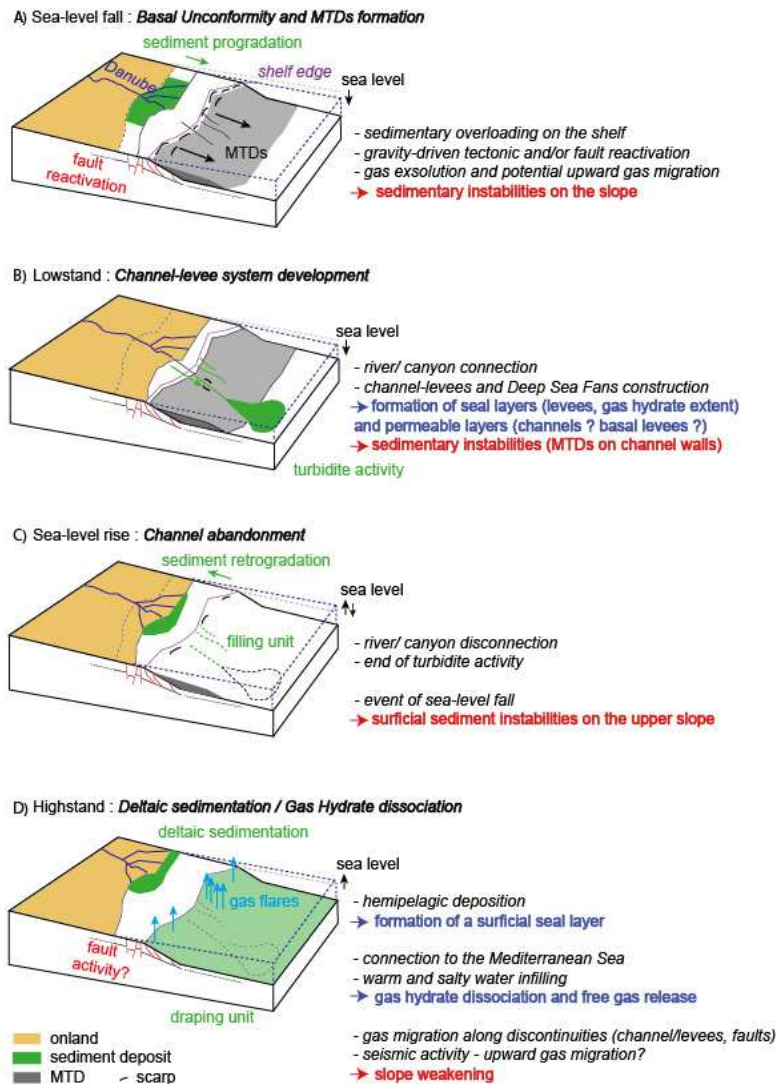


Fig. 20. Evolution of the tectonic-sedimentary architecture of surficial deposits since the last sea level fall. A) Sea level fall, formation of the basal erosional surface boundary and the basal layer of MTDs. B) Lowstand and the development of the canyons and channel-levee systems. Some sediment instabilities are also recorded within the canyon during this stage. C) Sea-level rise and abandonment of the channel. Formation of the infilling and draping unit. Development of the surficial MTDs during a short stage of gravity-driven instability. D) Highstand with deltaic sedimentation and gas hydrate dissociation along the slope. The present-day gas flare location is mainly controlled by the presence of gas hydrates, the location of canyons and slide walls, the fault systems and the sealing draping unit.

Conclusion

This work, based on the detailed analysis of multi-scale seismic data sets, has improved our understanding of the relationships between sedimentary architecture, faulting, slope instability, and fluid systems at the seafloor offshore Romania. The most recent paleo-valleys and their associated channel-levee complex/systems have formed over a basal layer of MTDs associated with an erosional unconformity attributed to the Base Neoeuxinian Sequence Boundary related to the last sea-level fall. The stacking pattern of these channelized systems shows that the turbidite activity migrated northward through at least five main paleo-valleys and then southward to the youngest Danube paleo-valley. The erosional unconformity shows slide scarp alignments, which generally coincide with normal to listric faults possibly linked to gravitational deformation. The geometry of paleo-valleys is strongly influenced by major sedimentary instabilities related to erosional processes and gas accumulation. Surficial (recent) sediment instabilities are interpreted as related to a gravity-driven fault reactivation attributed to the last melt water pulse and potential seismicity along the deep faults as well as gas hydrate dissociation.

The evolution from MTDs to turbidite and then marine deposits is consistent with the conventional sequence stratigraphy models that consider sea-level changes as the main control factor of margin sedimentation. However, gravity-driven deformation with possible seismicity and gas/gas-hydrate dynamics are considered as significant additional parameters impacting slope instabilities offshore Romania.

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Data availability

Data sets used in the current study were acquired during the GHASS expedition (10.17600/15000500) and data set information are available in the GHASS cruise report (<http://archimer.ifremer.fr/doc/00300/41141>).

References

- Anton, E.M., Munteanu, I., Dinu, C., Melinte-Dobrinescu, M.C., 2019. Litho- and biostratigraphy of the Eocene deposits from Istria Basin northern edge (Western Black Sea). *Geo-Eco-Marina*. 25, 55-70.
- Artemov, Y. G., Egorov, V. N., Polikarpov, G. G., Gulin, S. B., 2007. Methane emission to the hydro - and atmosphere by gas bubble streams in the Dnieper paleo-delta, the Black Sea (in Russian). *Rep. Natl. Acad. Sci. Ukraine*. 5, 110-116.
- Artyushkov, E.V., 1992. Role of crustal stretching on subsidence of continental crust. *Tectonophysics*. 215, 187-207. [https://doi.org/10.1016/0040-1951\(92\)90081-G](https://doi.org/10.1016/0040-1951(92)90081-G).
- Badertscher, S., Fleitmann, D., Cheng, H., Edwards, R.L., Göktürk, O.M., Zumbühl, A., Leuenberger, M., Tüysüz, O., 2011. Pleistocene water intrusions from the

944 Mediterranean and Caspian seas into the Black Sea. *Nature Geosci.* 4, 236–239.
 945 <https://doi.org/10.1038/ngeo1106>.

946 Badhani, S., Cattaneo, A., Dennielou, B., Leroux, E., Colin, F., Thomas, Y., Jouet, G.,
 947 Rabineau, M., Droz, L., 2019. Morphology of retrogressive failures in the Eastern
 948 Rhone Interfluve during the Last Glacial Maximum (Gulf of Lions, Western
 949 Mediterranean). *Geomorphology*. 351,.
 950 <https://doi.org/10.1016/j.geomorph.2019.106894>.

951 Badhani, S., Cattaneo, A., Collico, S., Urgeles, R., Dennielou, B., Leroux, E., Colin, F.,
 952 Garziglia, S., Rabineau, M., Droz, L., 2020. Integrated geophysical, sedimentological
 953 and geotechnical investigation of submarine landslides in the Gulf of Lions (Western
 954 Mediterranean). *Geological Society, London, Special Publications*. 500, 359-376.
 955 <https://doi.org/10.1144/SP500-2019-175>.

956 Ballas, G., Garziglia, S., Sultan, N., Pelleter, E., Toucanne, S., Marsset, T., Riboulot, V., Ker,
 957 S., 2018. Influence of early diagenesis on geotechnical properties of clay sediments
 958 (Romania, Black Sea). *Engineering Geology*. 240, 175-188.
 959 <https://doi.org/10.1016/j.enggeo.2018.04.019>Get.

960 Bega, Z., Ionescu, G., 2009. Neogene structural styles of the NW Black Sea region, offshore
 961 Romania. *The Leading Edge*. 28, 1082-1089. <https://doi.org/10.1190/1.3236378>.

962 Bolton, A.J., Maltman, A.J., Clenell, M.B., 1998. The importance of overpressure timing and
 963 permeability evolution in fine-grained sediments undergoing shear. *J. Struct. Geol.* 20,
 964 8, 1013-1022. [https://doi.org/10.1016/S0191-8141\(98\)00030-3](https://doi.org/10.1016/S0191-8141(98)00030-3).

965 Bouriak, S., Vanneste, M., Saoutkine, A., 2000. Inferred gas hydrates and clay diapirs near
 966 the Storegga Slide on the southern edge of the Vøring Plateau, offshore Norway. *Mar.*
 967 *Geol.* 163, 125-148. [https://doi.org/10.1016/S0025-3227\(99\)00115-2](https://doi.org/10.1016/S0025-3227(99)00115-2).

968 Brothers, D.S., Luttrell, K.M., Chaytor, J.D., 2013. Sea-level-induced seismicity and
 969 submarine landslide occurrence. *Geology*, 41, 9, 979-982.
 970 <https://doi.org/10.1130/G34410.1>.

971 Bünz, S., Mienert, J., Bryn, P., Berg, K., 2005. Fluid flow impact on slope failure from 3D
 972 seismic data: a case study in the Storegga Slide. *Basin Res.* 17, 109-122.
 973 <https://doi.org/10.1111/j.1365-2117.2005.00256.x>.

974 Canals, M., Lastras, G., Urgeles, R., Casamor, J.L., Mienert, J., Cattaneo, A., De Batist, M.,
 975 Haflidason, H., Imbo, Y., Laberg, J.S., Locat, J., Long, D., Longva, O., Masson, D.G.,
 976 Sultan, N., Trincardi, F. and Bryn, P., 2004. Slope failure dynamics and impacts from
 977 seafloor and shallow sub-seafloor geophysical data: case studies from the COSTA
 978 Project. *Mar. Geol.* 213, 9-72.

979 Cloetingh, S.A.P.L., Burov, E., Matenco, L., Toussaint, G., Bertotti, G., Andriessen, P.A.M.,
 980 Wortel, M.J.R., Spakman, W., 2003. Thermo-mechanical controls on the mode of
 981 continental collision in the SE Carpathians (Romania). *Earth and Planetary Science*
 982 *Letters*, 218, 57-76. [https://doi.org/10.1016/S0012-821X\(03\)00645-9](https://doi.org/10.1016/S0012-821X(03)00645-9).

983 Colin, F., Ker, S., Marsset, B., 2020a. Fine-scale velocity distribution revealed by datuming of
 984 very-high resolution deep-towed seismic data: Example of a shallow-gas system from
 985 the western Black Sea. *Geophysics*. 85, 5, 161-172. 10.1190/geo2019-0686.1 1.

986 Colin, F., Ker, S., Riboulot, V., Sultan, N., 2020b. Irregular BSR: Evidence of an Ongoing
 987 Reequilibrium of a Gas Hydrate System. *Geophysical Research Letters*. 47, 20.
 988 <https://doi.org/10.1029/2020GL089906>.

989 Constantinescu, A.M., Toucanne, S., Dennielou, B., Jorry, S.J., Mulder, T., Lericolais, G.,
 990 2015. Evolution of the Danube Deep-Sea fan since the last glacial maximum: new
 991 insights into Black Sea water-level fluctuations. *Mar. Geol.* 367, 50-68.
 992 <https://doi.org/10.1016/j.margeo.2015.05.007>.

993 Diaconu, A., Munteanu, I., Rădulescu, V., Moise, G., 2020. Gas escape features on the
 994 Romanian Black Sea shelf - Examples from 2D seismic interpretation. *Geo-Eco-Marina*.
 995 26, 1-9.

996 Dickens, G.R., and Quinby-Hunt, M.S., 1997. Methane hydrate in pore water: a simple
 997 theoretical approach for geophysical applications. *J. Geophys. Res.* 102, 773-783.
 998 <https://doi.org/10.1029/96JB02941>.

999 Dillon, W.P., Danforth, W.W., Hutchinson, D.R., Drury, R.M., Taylor, M.H., Booth, J.S., 1998.
1000 Evidence for faulting related to dissociation of gas hydrate and release of methane off
1001 the southeastern United States. *Journal of the Geological Society, London, Special*
1002 *Publications*. 137, 293-302. <https://doi.org/10.1144/GSL.SP.1998.137.01.23>.

1003 Dinu, C., Tambrea, D., Raileanu, A., 2003. Transition from extensional to compressional
1004 structures in the Upper Miocene Shally deposits, Western Black Sea. *AAPG,*
1005 *Barcelona*.

1006 Dinu, C., Wong, H.K., Tambrea, D. and Matenco, L., 2005. Stratigraphic and structural
1007 characteristics of the Romanian Black Sea shelf, *Tectonophysics*. 410, 417-435,
1008 [doi:10.1016/j.tecto.2005.04.012](https://doi.org/10.1016/j.tecto.2005.04.012).

1009 Dinu, C., Munteanu, I., Tambrea, D., Panin, N., 2018. Gas hydrates, Flares Seeps Offshore
1010 Romania, A Review. *Geo-Eco-Marina*. 24, 5-56.

1011 dos Reis, A.T., Gorini, C., Mauffret, A. 2005. Implications of salt–sediment interactions on the
1012 architecture of the Gulf of Lions deep-water sedimentary systems—western
1013 Mediterranean Sea. *Marine and Petroleum Geology*. 22, 713–746,
1014 <https://doi.org/10.1016/J.MARPETGEO.2005.03.006>.

1015 Droz, L., Marsset, T., Ondreas, H., Lopez, M., Savoye, B., Spy-Anderson, F.-L., 2003.
1016 Architecture of an active mud-rich turbidite system: the Zaire Fan (Congo-Angola
1017 margin southeast Atlantic) results from ZaiAngo 1 and 2 cruises. *AAPG Bull.* 87,
1018 1145e1168. <http://dx.doi.org/10.1306/03070300013>.

1019 Dugan, B., Flemings, P., Olgaard, D.L. and Gooch, M.J., 2003. Consolidation, effective
1020 stress, and fluid pressure of sediments from ODP Site 1073, US mid-Atlantic
1021 continental slope. *Earth Planet. Sc. Lett.* 215, 13-26. [https://doi.org/10.1016/S0012-](https://doi.org/10.1016/S0012-821X(03)00425-4)
1022 [821X\(03\)00425-4](https://doi.org/10.1016/S0012-821X(03)00425-4).

1023 Dupont P., 2015. Compte rendu d'acquisition sismique. Campagne GHASS. N/O Pourquoi
1024 Pas ?, 1 Septembre - 30 Septembre 2015. PDG/REM/GM/EIS-30092015-01.

1025 Egorov, V. N., Artemov, Y. G., Gulin, S.B., Polikarpov G.G., 2011. Methane seeps in the
 1026 Black Sea: discovery, quantification and environmental assessment. J. Black
 1027 Sea/Mediterranean Environment. 17, 171-185.

1028 Egorov, V., Luth, U., Luth, C., Gulin, M.B., 1998. Gas seeps in the submarine Dnieper
 1029 Canyon, Black Sea: acoustic, video and trawl data, in: Luth, U., Luth, C., Thiel, H.
 1030 (Eds.), Methane Gas Seep Explorations in the Black Sea (MEGASEEBS). Project
 1031 Report. Zentrum für Meres-und Klimaforschung, Universität Hamburg, Hamburg, pp.
 1032 11-21.

1033 Field, M.E., Barber, J.H., 1993. A submarine landslide associated with shallow seafloor gas
 1034 and gas hydrates off northern California, in: Schwab, W.C., Lee, H.J., Twichell, D.C.
 1035 (Eds.), Submarine landslides: selected studies in the U.S. Exclusive Economic Zone.
 1036 U.S. Geological Survey.

1037 Finetti, I., Bricchi, G., Del Ben, A., Pipan, M. and Xuan, Z., 1988. Geophysical study of the
 1038 Black Sea. Bollettino di Geofisica Teorica ed Applicata. 30, 117-118, 197-324.

1039 Garziglia, S., Migeon, S., Ducassou, E., Loncke, L., Mascle, J., 2008. Mass-transport
 1040 deposits on the Rosetta province (NW Nile deep-sea turbidite system, Egyptian
 1041 margin): characteristics, distribution, and potential causal processes. Marine Geology.
 1042 250, 180-198. <https://doi.org/10.1016/j.margeo.2008.01.016>.

1043 Georgiev, G., 2012. Geology and hydrocarbon Systems in the Western Black Sea. Turk. J.
 1044 Earth Sci. 21, 723-754.

1045 Gillet, H., Lericolais, G., Réhault, J.-P., 2007. Messinian event in the black sea: Evidence of
 1046 a Messinian erosional surface. Marine Geology. 244, 142-165.
 1047 <https://doi.org/10.1016/j.margeo.2007.06.004>.

1048 Görür, N., 1988. Timing of opening of the Black Sea basin. Tectonophysics. 147, 247- 262.
 1049 [https://doi.org/10.1016/0040-1951\(88\)90189-8](https://doi.org/10.1016/0040-1951(88)90189-8).

1050 Greinert, J., McGinnis, D. F., Naudts, L., Linke, P., De Batist, M., 2010. Atmospheric
 1051 methane flux from bubbling seeps: spatially extrapolated quantification from a Black
 1052 Sea shelf area, J. Geophys. Res.: Oceans. 115. <https://doi.org/10.1029/2009JC005381>.

1053 Haq, B.U., 1998. Natural gas hydrates: Searching the long-term climatic and slope stability
1054 records, in: Henriot, J.P. and Miniert, J., (Eds), Gas hydrates: Relevance to world
1055 margin stability and climatic change: Geological Society (London) Special Publication
1056 137, 303-318. <https://doi.org/10.1144/GSL.SP.1998.137.01.24>.

1057 Hillman, J. I. T., Klauke, I., Bialas, J., Feldman, H., Drexler, T., Awwiller, D., Atgin, O., Çifçi,
1058 G., Badhani, S., 2018. Gas migration pathways and slope failures in the Danube Fan,
1059 Black Sea. *Marine and Petroleum Geology*. 92, 1069-1084.
1060 <https://doi.org/10.1016/j.marpetgeo.2018.03.025>.

1061 Hippolyte, J-C., 2002. Geodynamics of Dobrogea (Romania): new constraints on the
1062 evolution of the Tornquist-Tesseyre Line, the Black Sea and the Carpathians.
1063 *Tectonophysics*. 357, 33-53. [https://doi.org/10.1016/S0040-1951\(02\)00361-X](https://doi.org/10.1016/S0040-1951(02)00361-X).

1064 Hsü, K.J., Giovanoli, F., 1979. Messinian event in the black sea. *Palaeogeography,*
1065 *Palaeoclimatology, Palaeoecology*. 29, 75-93. [https://doi.org/10.1016/0031-](https://doi.org/10.1016/0031-0182(79)90075-0)
1066 [0182\(79\)90075-0](https://doi.org/10.1016/0031-0182(79)90075-0).

1067 International Seismological Centre (2020), On-line Bulletin,
1068 <https://doi.org/10.31905/D808B830>.

1069 Ion, G., Lericolais, G., Nouzé, H., Panin, N., Ion, E. 2001. Seismo-acoustics evidence of
1070 gases in sedimentary edifices of the paleo-Danube realm, in: *Turbidite systems and*
1071 *deep-sea fans of the Mediterranean and the Black seas*. CIESM Workshop Series, 17,
1072 91-95. Bucharest (Romania), 5 - 8 June 2002, 119 p.

1073 Kennett, J.P., Cannariato, K.G., Hendy, I.L., Behl, R.J., 2003. Methane hydrates in
1074 Quaternary climate change: The clathrate gun hypothesis: Washington, D.C., American
1075 Geophysical Union. 213. <https://doi.org/10.1002/9781118665138.ch0>.

1076 Ker, S., Riboulot, V., 2015. GHASS cruise Team (2015). GHASS cruise report.
1077 <https://archimer.ifremer.fr/doc/00300/41141/>.

1078 Ker, S., Thomas, Y., Riboulot, V., Sultan, N., Bernard, C., Scalabrin, C., Ion, G., Marsset, B.,
1079 2019. Anomalously Deep BSR Related to a Transient State of the Gas Hydrate System

1080 in the Western Black Sea. *Geochemistry, Geophysics, Geosystems*. 20, 442-459.
 1081 <https://doi.org/10.1029/2018GC007861>.

1082 Kim, Y-G., Lee, S-M., Jin, Y.K., Baranov, B., Obzhairov, A., Salomatin, A., Shoji, H., 2013.
 1083 The stability of gas hydrate field in the northeastern continental slope of Sakhalin
 1084 Island, Sea of Okhotsk, as inferred from analysis of heat flow data and its implications
 1085 for slope failures. *Marine and Petroleum Geology*. 45, 198-207.
 1086 <https://doi.org/10.1016/j.marpetgeo.2013.05.003>.

1087 Knox, G.J., Omatsola, E.M., 1989. Development of the Cenozoic Niger delta in terms of the
 1088 “escalator regression” model and impact on hydrocarbon distribution. Koninklijk
 1089 Nederlands Geologisch en Mijnbouwkundig Gezelschap Symposium on Coastal
 1090 Lowlands, Geology and Geotechnology Proceeding (1987). Kluwer Academic
 1091 Publishers, Dordrecht, Netherlands, 181-202. doi:10.1007/978-94-017-1064-0_12.

1092 Konerding, C., Dinu, C., Wong, H.K., 2010. Seismic sequence stratigraphy, structure and
 1093 subsidence history of the Romanian Black Sea shelf, in: Sosson, M., Kaymakci, N.,
 1094 Stephenson, R.A., Bergerat, F., Starostenko, V., (Eds.), *Sedimentary Basin Tectonics*
 1095 from the Black Sea and Caucasus to the Arabian Platform. *Journal of the Geological*
 1096 *Society*, London, Special Publications, 340, pp. 159-180.
 1097 <https://doi.org/10.1144/SP340.9>.

1098 Krezsek, C., Schleder, Z., Bega, Z., Ionescu, G., Tari, G., 2016. The Messinian sea-level fall
 1099 in the western Black Sea: small or large? Insights from offshore Romania. *Pet. Geosci.*
 1100 22, 392–399. <https://doi.org/10.1144/petgeo2015-093>.

1101 Krézsek, C., Olariu, C., 2021. Filling of sedimentary basins and the birth of large rivers: The
 1102 lower Danube network in the Dacian Basin, Romania. *Global and Planetary Change*.
 1103 197. doi.org/10.1016/j.gloplacha.2020.103391.

1104 Kruglyakova, R.P., Byakov, Y.A., Kruglyakova, M.V., Chalenko, L.A., Shevtsova, N.T., 2004.
 1105 Natural oil and gas seeps on the Black Sea floor. *Geo Mar. Lett.* 24, 150-162. DOI
 1106 10.1007/s00367-004-0171-4.

1107 Kvenvolden, K.A., 1993. Gas hydrates-Geological perspective and global change. *Rev.*
1108 *Geophys.* 31, 173-187. <https://doi.org/10.1029/93RG00268>.

1109 Kvenvolden, K.A., and McMenamin M.A., 1980. Hydrates of natural gas: A review of their
1110 geologic occurrence. *U.S. Geol. Survey Circ.* C825, 11.

1111 Kwon, T.H., Cho, G.C., 2012. Submarine slope failure primed and triggered by bottom water
1112 warming in oceanic hydrate-bearing deposits. *Energies.* 5, 2849-2873.
1113 <https://doi.org/10.3390/en5082849>.

1114 Lafuerza, S., Sultan, N., Canals, M., Frigola, J., Berné, S., Jouet, G., Galavazi, M., Sierro,
1115 F.J., 2009. Overpressure within upper continental slope sediments from CPTU data,
1116 Gulf of Lion, NW Mediterranean Sea. *International Journal of Earth Sciences.* 98, 4,
1117 751-768. <https://doi.org/10.1007/s00531-008-0376-2>.

1118 Lee, H.J., 2005. Undersea landslides: extent and significance in the Pacific Ocean, an
1119 update. *Nat. Hazards Earth Syst. Sci.* 5, 877-892. [https://doi.org/10.5194/nhess-5-877-](https://doi.org/10.5194/nhess-5-877-2005)
1120 2005.

1121 Lericolais, G., Bourget, J., Popescu, I., Jermannaud, P., Mulder, T., Jorry, S., Panin, N.,
1122 2013. Late Quaternary deep-sea sedimentation in the western Black Sea: new insights
1123 from recent coring and seismic data in the deep basin. *Global Planet. Change.* 103,
1124 232-247. <https://doi.org/10.1016/j.gloplacha.2012.05.002>.

1125 Lericolais, G., Bulois, C., Gillet, H., Guichard, F., 2009. High frequency sea level fluctuations
1126 recorded in the Black Sea since the LGM. *Glob. Planet. Chang.* 66, 65-75.
1127 <https://doi.org/10.1016/j.gloplacha.2008.03.010>.

1128 Letouzey, J., Biju-Duval, B., Dorkel, A., Gonnard, R., Kristchev, K., Montadert, L., Sungurlu,
1129 O., 1977. The Black Sea: a marginal basin: geophysical and geological data, in: Biju-
1130 Duval, B., Montadert, L. (Eds.), *Structural History of the Mediterranean Basins*. Editions
1131 Technip, Paris, pp. 363-376.

1132 Locat, J., Leroueil, S., Locat, A., Lee, H., 2014. Weak Layers: Their definition and
1133 classification from a geotechnical perspective, in: Krastel S., Behrmann, J.H., Völker, D.,
1134 Stipp, M., Berndt, C., Urgele, R., et al. (Eds.), *Submarine Mass Movements and Their*

1135 Consequences. *Advances in Natural and Technological Hazards Research*. 37,
 1136 Springer, Cham, Switzerland, pp. 3-12.

1137 Løvholt, F., Glimsdal, S., Harbitz, C.B., 2020. On the landslide tsunami uncertainty and
 1138 hazard. *Landslide*. 17, 2301-2315.

1139 Madof, A.S., Christie-Blick, N., Anders, M.H., 2009. Stratigraphic controls on a salt-withdrawal intraslope minibasin, north-central Green
 1140 Canyon, Gulf of Mexico: Implications for misinterpreting sea level change. *AAPG Bull.*
 1141 93, 4, 535–561. DOI:10.1306/12220808082.

1142 Magara, K., 1978. Compaction and fluid migration. *Practical petroleum geology.*
 1143 *Developments in Petroleum Science*. 9, 11-46.

1144 Marsset, B., Menut, E., Ker, S., Thomas, Y., Regnault, J.P., Leon, P., Martinossi, H., Artzner,
 1145 L., Chenot, D., Dentrecolas, S., Spychalski, B., Mellier, G., Sultan, N., 2014. Deep-
 1146 towed high resolution multichannel seismic imaging, *Deep Sea Res., Part I: Oceanogr.*
 1147 *Res. Pap.* 93, 83–90. <https://doi.org/10.1016/j.dsr.2014.07.013>.

1148 Marsset, B., Ker, S., Thomas, Y., Colin, F., 2018. Deep-towed high resolution seismic
 1149 imaging II: Determination of P-wave velocity distribution, *Deep Sea Res., Part I-*
 1150 *oceanographic Res. Pap.* 132, 29–36. <https://doi.org/10.1016/j.dsr.2017.12.005>.

1151 Marsset, T., Marsset, B., Ker, S., Thomas, Y., Le Gall, Y., 2010. High and very resolution
 1152 deep-towed seismic system: performance and examples from deep water geohazard
 1153 studies. *Deep Sea Res. I*. 57, 4, 628–637. <https://doi.org/10.1016/j.dsr.2010.01.001>.

1154 Martinez-Lamas, R., Toucanne, S., Debret, M., Riboulot, V., Deloffre, J., Boissier, A.,
 1155 Cheron, S., Pitel, M., Bayon, G., Giosan, L., Soulet, G., 2020. Linking Danube River
 1156 activity to Alpine Ice-Sheet fluctuations during the last glacial (ca. 33–17 ka BP):
 1157 Insights into the continental signature of Heinrich Stadials. *Quaternary Science*
 1158 *Reviews*. 229, 106136. <https://doi.org/10.1016/j.quascirev.2019.106136>.

1159 Maslin, M.A., Mikkelsen, N., Vilela, C., and Haq, B., 1998. Sea-level- and gas-hydrate-
 1160 controlled catastrophic sediment failures of the Amazon Fan: *Geology*, 26, 12, 1107-
 1161 1100. [https://doi.org/10.1130/0091-7613\(1998\)026<1107:SLAGHC>2.3.CO;2](https://doi.org/10.1130/0091-7613(1998)026<1107:SLAGHC>2.3.CO;2).

1162 Maslin, M., Owen, M., Day, S., Long., D., 2004. Linking continental-slope failures and climate
 1163 change: Testing the clathrate gun hypothesis. *Geology*, 32,1, 53-56.
 1164 <https://doi.org/10.1016/j.dsr.2014.07.013>.

1165 Masson, D.G., Harbitz, C.B., Wynn, R.B., Pedersen, G., Løvholt, F., 2006. Submarine
 1166 landslides: processes, triggers, and hazard prediction. *Phil. Trans. R. Soc. A*. 364,
 1167 2009-2039. <https://doi.org/10.1098/rsta.2006.1810>.

1168 Matenco, L., Munteanu, I., Ter Borgh, M., Stanica, A., Tilita, M., Lericolais, G., Dinu C., Oaie
 1169 G., 2016. The interplay between tectonics, sediment dynamics and gateways evolution
 1170 in the Danube system from the Pannonian Basin to the western Black Sea. *Science Of*
 1171 *The Total Environment*. 543, 807-827. Publisher's official version
 1172 <https://doi.org/10.1016/j.scitotenv.2015.10.081>.

1173 Matoshko, A., Matoshko, A., de Leeuw, A., 2019. The Plio-Pleistocene demise of the East
 1174 Carpathians foreland fluvial system and arrival of the Paleo-Danube to the Black Sea.
 1175 *Geol. Carpath.* 70, 2, 91-112. <https://doi.org/10.2478/geoca-2019-0006>.

1176 Matova, M., 2000. Recent geological activity along the northeastern Bulgarian Black Sea
 1177 coast. *Geol. Quaterly*. 44, 355-362.

1178 Merey, S., and Sinayuc, C., 2016. Investigation of gas hydrate potential of the Black Sea and
 1179 modelling of gas production from a hypothetical Class 1 methane hydrate reservoir in
 1180 the Black Sea conditions, *J. Nat. Gas Sci. Eng.* 29, 66-79.
 1181 <https://doi.org/10.1016/j.jngse.2015.12.048>.

1182 Mitchum, R.M., Vail, P.R., 1977. Seismic stratigraphy and global changes in sea-level, part 7:
 1183 stratigraphic interpretation of seismic reflection patterns in depositional sequences, in:
 1184 Payton, C.E. (Ed.), *Seismic Stratigraphy - Applications to Hydrocarbon Exploration*.
 1185 AAPG Memoir. Tulsa, 26, pp. 135–144.

1186 Mitchum, R.M., Vail, P.R., Sangree, J.B., 1977. Seismic stratigraphy and global changes in
 1187 sea-level, part 6: seismic stratigraphic interpretation procedure, in: Payton, C.E. (Ed.),
 1188 *Seismic Stratigraphy - Applications to Hydrocarbon Exploration*. AAPG Memoir. Tulsa,
 1189 26, pp. 117–134.

1190 Morosanu, I., 2007. Romanian continental plateau of the Black Sea, Oscar print, Bucharest,
 1191 176 pp, ISBN 078-973-668-167-7.

1192 Munteanu, I., Matenco, L., Dinu, C., Cloetingh, S., 2011. Kinematics of back-arc inversion of
 1193 the Western Black Sea Basin. *Tectonics*. 30, TC5004. [http://dx.doi.org/10.1029/](http://dx.doi.org/10.1029/2011tc002865)
 1194 2011tc002865.

1195 Munteanu, I., Matenco, L., Dinu, C., Cloetingh, S., 2012. Effects of large sea-level variations
 1196 in connected basins: the Dacian-Black Sea system of the Eastern Paratethys. *Basin*
 1197 *Res.* 24, 583-597. <https://doi.org/10.1111/j.1365-2117.2012.00541.x>.

1198 Naudts, L., Greinert, J., Artemov, Y., Staelens, P., Poort, J., Van Rensbergen, P., De Batist,
 1199 M., 2006. Geological and morphological setting of 2778 methane seeps in the Dnepr
 1200 paleo delta, northwestern Black Sea. *Mar. Geol.* 227, 177-199,
 1201 doi:10.1016/j.margeo.2005.10.005.

1202 Naudts, L., Greinert, J., Artemov, Y., Beaubien, S.E., Borowski, C., De Batist, M., 2008.
 1203 Anomalous sea-floor backscatter patterns in methane venting areas, Dnepr paleo-delta,
 1204 NW Black Sea. *Mar. Geol.* 251, 253-267. <https://doi.org/10.1016/j.margeo.2008.03.002>.

1205 Nikishin, A.M., Ziegler, P.A., Panov, D.I., Nazarevich, B.P., Brunet, M-F., Stephenson, R.A.,
 1206 Bolotov, S.N., Korotaev, M.V. & Tikhomirov, P.L. 2001. Mesozoic and Cainozoic
 1207 evolution of the Scythian Platform-Black Sea-Caucasus domain, in: Ziegler, P.A.,
 1208 Cavazza, W., Robertson, A.H.F., Crasquin-Soleau, S. (Eds.), *Peri-Tethys Memoir*, 6:
 1209 *Peri-Tethyan Rift /Wrench Basins and Passive Margins*. *Memoires du Museum National*
 1210 *d'Histoire Naturelle*, 186, 295-346.

1211 Nikishin, A., Ziegler, P., Bolotov, S.N. and Fokin, P.A., 2011. Late Palaeozoic to Cenozoic
 1212 Evolution of the Black Sea-Southern Eastern Europe Region: A View from the Russian
 1213 Platform. *Turkish Journal of Earth Science*. 20, 571-634.

1214 Nikolovska, A., H. Sahling, and G. Bohrmann, 2008. Hydroacoustic methodology for
 1215 detection, localization, and quantification of gas bubbles rising from the seafloor at gas
 1216 seeps from the eastern Black Sea, *Geochem. Geophys. Geosyst.* 9, Q10010,
 1217 doi:10.1029/2008GC002118.

1218 Panin, N., Jipa, D., 2002. Danube River Sediment Input and its Interaction with the North-
 1219 western Black Sea Estuarine, Coastal and Shelf Science 54, 551-562,
 1220 doi:10.1006/ecss.2000.0664.

1221 Panin, N., Panin, S., Herz, N., Noakes, J. E., 1983. Radiocarbon dating of Danube delta
 1222 deposits. Quaternary Research. 19, 249-255. [https://doi.org/10.1016/0033-](https://doi.org/10.1016/0033-5894(83)90008-X)
 1223 5894(83)90008-X.

1224 Papadopoulos, G. A., Diakogianni, G., Fokaefs, A., Rangelov, B., 2011. Tsunami hazard in
 1225 the Black Sea and the Azov Sea: a new tsunami catalogue. Nat. Hazards Earth Syst.
 1226 Sci. 11, 945–963. <https://doi.org/10.5194/nhess-11-945-2011>.

1227 Pătruț, I., Paraschiv, C., Dăneț, T., Balteș, N., Dăneț, N., Motaș, L., 1984. The Geological
 1228 constitution of the Danube Delta. Anuarul Institutului de Geologie și Geofizică -
 1229 Annuaire de l'Institut de Géologie et de Géophysique. 59, 55-61.

1230 Pecher, I.A., Henrys, S.A., Ellis, S., Chiswell, S.M., Kukowski, N., 2005. Erosion of the
 1231 seafloor at the top of the gas hydrate stability zone on the Hikurangi Margin, New
 1232 Zealand. Geophysical Research Letters. 32, 24.
 1233 <https://doi.org/10.1029/2005GL024687>.

1234 Piper, D. J. W., C. Pirmez, P. L. Manley, D. Long, R. D. Flood, W. R. Normark and W.
 1235 Showers, 1997. Mass-transport deposits of the Amazon Fan, in: Flood, R. D., Piper, D.
 1236 J. W., Klaus, A., Peterson, L. C. (Eds.), Proceedings of the Ocean Drilling Program:
 1237 Scientific Results, 155, 109-146: College Station, Texas (Ocean Drilling Program).

1238 Piper, D.J.W., Cochonat, P., Morrison, M.L., 1999. The sequence of events around the
 1239 epicentre of the 1929 Grand Banks earthquake: initiation of debris flows and turbidity
 1240 current inferred from sidescan sonar. Sedimentology. 46, 79-97.
 1241 <https://doi.org/10.1046/j.1365-3091.1999.00204.x>.

1242 Polikarpov, G.G., Egorov, V.N., Nezhdanov, A.I., Gulin, S.B., Kulev, Y.D., Gulin, M.B.,
 1243 1989. Methane gas seeps in the Black Sea—a new object of molismology, in:
 1244 Polikarpov, G.G. (Ed.), Molismology of the Black Sea. Nauka, Kiev, pp. 10-20.

1245 Pope, E.L., Talling, P.J., Carter, L., 2017. Which earthquakes trigger damaging submarine
1246 mass movements: Insights from a global record of submarine cable breaks? *Mar. Geol.*
1247 384, 131-146. <https://doi.org/10.1016/j.margeo.2016.01.009>.

1248 Popescu, I., 2002. Analyse des processus sédimentaires récents dans l'éventail profond du
1249 Danube (Mer Noire). PhD thesis, Université de Bretagne Occidentale. 282 pp.

1250 Popescu, I., De Batist, M., Lericolais, G., Nouze, H., Poort, J., Panin, N., Versteeg, W., Gillet,
1251 H., 2006. Multiple bottom-simulating reflections in the Black Sea: potential proxies of
1252 past climate conditions. *Mar. Geol.* 227, 163-176. <https://doi.org/10.1016/j.margeo.2005.12.006>.

1254 Popescu, I., Lericolais, G., Panin, N., De Batist, M., Gillet, H., 2007. Seismic expression of
1255 gas and gas hydrates across the western Black Sea. *Geo-Marine Letters*. 27, 173-183.
1256 <https://doi.org/10.1007/s00367-007-0068-0>.

1257 Popescu, I., Lericolais, G., Panin, N., Normand, A., Dinu, C., Le Drezen, E., 2004. The
1258 Danube submarine canyon (Black Sea): morphology and sedimentary processes. *Mar.*
1259 *Geol.* 206, 249-265. <https://doi.org/10.1016/j.margeo.2004.03.003>.

1260 Popescu, I., Lericolais, G., Panin, N., Wong, H.K., Droz, L., 2001. Late Quaternary channel
1261 avulsions on the Danube deep-sea fan. *Black Sea. Mar. Geol.* 179, 25- 37.
1262 [https://doi.org/10.1016/S0025-3227\(01\)00197-9](https://doi.org/10.1016/S0025-3227(01)00197-9).

1263 Posamentier, H. W., Kolla, V., 2003. Seismic geomorphology and stratigraphy of depositional
1264 elements in deep-water settings: *Journal of Sedimentary Research*. 73, 367-388,
1265 doi:10.1306/111302730367.

1266 Riboulot, V., Cattaneo, A., Scalabrin, C., Gaillot, A., Jouet, G., Ballas, G., Marsset, T.,
1267 Garziglia, S., Ker, S., 2017. Control of the geomorphology and gas hydrate extent on
1268 widespread gas emissions offshore Romania. *Bull. Soc. Geol. Fr.* 188 (4), 26.
1269 <https://doi.org/10.1051/bsgf/2017182>.

1270 Riboulot, V., Cattaneo, A., Sultan, N., Garziglia, S., Ker, S., Imbert, P., Voisset, M., 2013.
1271 Sea-level change and free gas occurrence influencing a submarine landslide and

1272 pockmark formation and distribution in deepwater Nigeria. *Earth Planet. Sci. Lett.* 375,
1273 78-91. <https://doi.org/10.1016/j.epsl.2013.05.013>.

1274 Riboulot, V., Ker, S., Sultan, N., Thomas, Y., Marsset, B., Scalabrin, C., Ruffine, L., Boulart,
1275 C., Ion, G., 2018. Freshwater lake to salt-water sea causing widespread hydrate
1276 dissociation in the Black Sea, *Nat. Commun.* 9 (117).
1277 <https://doi.org/10.1016/j.epsl.2013.05.013>.

1278 Robinson, A.G., Rudat, J.H., Banks, C.J., Wiles, R.L.F., 1996. Petroleum geology of the
1279 Black Sea. *Mar. Pet. Geol.* 13, 195-223. [https://doi.org/10.1016/0264-8172\(95\)00042-9](https://doi.org/10.1016/0264-8172(95)00042-9).

1280 Römer, M., Sahling, H., dos Santos Ferreira, C., Bohrmann, G., 2020. Methane gas
1281 emissions of the Black Sea-mapping from the Crimean continental margin to the Kerch
1282 Peninsula slope. *Geo-Marine Letters.* 40, 467-480. [https://doi.org/10.1007/s00367-019-](https://doi.org/10.1007/s00367-019-00611-0)
1283 00611-0.

1284 Rouby, D., Nalpas T., Jermannaud P., Robin C., Guillocheau F., Raillard S., 2011. Gravity
1285 driven deformation controlled by the migration of the delta front: The Plio-Pleistocene of
1286 the Eastern Niger Delta. *Tectonophysics.* 513, 54-67. doi:
1287 <https://doi.org/10.1130/GEOSO1426.1>.

1288 Schleder, Z., Krezsek, C., Lapadat, A., Bega, Z., Ionescu, G., Tari, G., 2016. Structural style
1289 in a Messinian (intra-Pontian) gravity-driven deformation system, western Black Sea,
1290 offshore Romania. *Petroleum Geoscience.* 22(4), 400-410.
1291 <https://doi.org/10.1144/petgeo2015-094>.

1292 Skarke A., Ruppel, C., Kodis, M., Brothers, D., Lobecker, E., 2014. Widespread methane
1293 leakage from the sea floor on the northern US Atlantic margin. *Nature Geoscience.*
1294 DOI :10.1038/NGEO2232.

1295 Sloan, E.D., Jr., *Clathrate Hydrates of Natural Gases.* 641 p. Marcel Dekker, New-York,
1296 1990.

1297 Sloan, E.D., Jr., *Clathrate Hydrates of Natural Gases,* 2 ed., 705 p., Marcel Dekker, New-
1298 York, 1998.

1299 Soulet, G., Ménot, G., Bayon, G., Rostek, F., Ponzevera, E., Toucanne, S., Lericolais, G.,
 1300 Bard, E., 2013. Abrupt drainage cycles of the Fennoscandian Ice Sheet. *PNAS*. 110,
 1301 17, 6682-6687. <https://doi.org/10.1073/pnas.1214676110>.
 1302 Soulet, G., Ménot, G., Garreta, V., Rostek, F., Zaragosi, S., Lericolais, G., Bard, E., 2011.
 1303 Black Sea “Lake” reservoir age evolution since the last glacial — hydrologic and
 1304 climatic implications. *Earth Planet. Sci. Lett.* 308, 245-258.
 1305 <https://doi.org/10.1016/j.epsl.2011.06.002>.
 1306 Sultan, N., Cochonat, P., Canals, M., Cattaneo, A., Dennielou, B., Haflidason, H., Laberg,
 1307 J.S., Long, D., Mienert, J., Trincardi, F., Urgeles, R., Vorren, T.O., Wilson, C., 2004a.
 1308 Triggering mechanisms of slope instability processes and sediment failures on
 1309 continental margins: a geotechnical approach. *Mar.Geol.* 213, 291–321.
 1310 <https://doi.org/10.1016/j.margeo.2004.10.011>.
 1311 Sultan, N., Cochonat, P., Foucher, J.-P., Mienert, J., 2004b. Effect of gas hydrates melting
 1312 on seafloor slope instability. *Mar. Geol.* 213, 379-401.
 1313 <https://doi.org/10.1016/j.margeo.2004.10.015>.
 1314 Tailling, P.J., Clare, M., Urlaub, M., Pope, E., Hunt, J.E., Watt, S.F.L., 2014. Large
 1315 submarine landslides on continental slopes: Geohazards, methane release, and climate
 1316 change. *Oceanography*. 27, 2, 32-45. <http://www.jstor.org/stable/24862153>.
 1317 Țambrea, D., 2007. Subsidence analysis and tectonic-thermal evolution of the Istria
 1318 Depression (Black Sea); Implications for the hydrocarbon generation (in Romanian).
 1319 Ph.D thesis, Bucharest University, 156 p.
 1320 Tambrea, D., Dinu, C., Sampetean, E., 2002. Characteristics of the Tectonics and
 1321 Lithostratigraphy of The Black Sea Shelf, Offshore Romania, in: Dinu, C., Mocanu, V.
 1322 (Eds.), *Geology and Tectonics of the Romanian Black Sea Shelf and its Hydrocarbon*
 1323 *Potential*. Vergiliu, Bucharest, pp. 29-42.
 1324 Twichell, D. C., Kenyon, N. H., Parson, L. M., McGregor, B. A., 1991. Depositional patterns
 1325 of the Mississippi Fan surface: Evidence from GLORIA II and high-resolution seismic
 1326 profiles, in: Weimer, P. and Link, M. H. (Eds.), *Seismic facies and sedimentary*

1327 processes of submarine fans and turbidite systems. *Frontiers in Sedimentary Geology*.
 1328 Springer-Verlag, New York, pp. 349-363. [https://doi.org/10.1007/978-1-4684-8276-](https://doi.org/10.1007/978-1-4684-8276-8_19)
 1329 8_19.

1330 Urlaub, M., Talling, P.J., Masson, D.G., 2013. Timing and frequency of large submarine
 1331 landslides: implications for understanding triggers and future geohazard. *Quaternary*
 1332 *Science Reviews*. 72, 63-82. <https://doi.org/10.1016/j.quascirev.2013.04.020>.

1333 Vail, P.R., Mitchum Jr., R.M., Thompson, S.I.I.I., 1977. Seismic stratigraphy and global
 1334 changes of sea level, part 4, global cycles of relative changes of sea level, in: Payton,
 1335 C.E. (Ed.), *Seismic Stratigraphy - Applications to Hydrocarbon Exploration*. American
 1336 Association of Petroleum Geologists, Memoir 26, 83-97.

1337 Van Wagoner, J.C., Posamentier, H.W., Mitchum, R.M., Vail, P.R., Sarg, J.F., Loutit, T.S.,
 1338 Hardenbol, J., 1988. An overview of the fundamentals of sequence stratigraphy and
 1339 key definitions, in: Wilgus, C.K., Hastings, B.S., Kendall, C.G., Posamentier, H.W.,
 1340 Ross, C.A., Van Wagoner, J.C. (Eds.), *Sea Level Changes: an integrated approach*.
 1341 SEPM, 42, pp. 39-45.

1342 Wegwerth, A., Ganopolski, A., Ménot, G., Kaiser, J., Dellwig, O., Bard, E., Lamy, F., Arz,
 1343 H.W., 2015. Black Sea temperature response to glacial millennial-scale climate
 1344 variability. *Geophys. Res. Lett.* 42, 8147-8154. <https://doi.org/10.1002/2015GL065499>.

1345 Weimer, P., 1990. Sequence stratigraphy, facies geometries, and depositional history of the
 1346 Mississippi fan, Gulf of Mexico. *AAPG Bull.* 74 (4), 425-453.
 1347 <https://doi.org/10.1306/0C9B2321-1710-11D7-8645000102C1865D>.

1348 Winguth, C., Wong, H.K., Panin, N., Dinu, C., Georgescu, P., Ungureanu, G., Krugliakov,
 1349 V.V., Podshuveit, V., 2000. Upper quaternary water level history and sedimentation in
 1350 the northwestern Black Sea. *Mar. Geol.* 167, 127- 146. [https://doi.org/10.1016/S0025-](https://doi.org/10.1016/S0025-3227(00)00024-4)
 1351 3227(00)00024-4.

1352 Yilmaz, Y., Tuysuz, O., Yigitbas, E., Genc, S.C., Sengor, A.M.C., 1997. Geology and
 1353 Tectonic Evolution of the Pontides, in: Robinson A.G. (Ed.), *Regional and Petroleum*
 1354 *Geology of the Black Sea and Surrounding Region*. AAPG Memoir, 68, 183-226.

1355 Yang, J., Davies, R., 2013. Gravity-driven faults: migration pathways for recycling gas after
 1356 the dissociation of marine gas hydrates. *Marine Geology* 336, 1-9.
 1357 <https://doi.org/10.1016/j.margeo.2012.11.013>.
 1358 Zander, T., Haeckel, M., Berndt, C., Chi, W.C., Klaucke, I., Bialas, J., Klaeschen, D., Koch,
 1359 S., Atgin, O., 2017. On the origin of multiple BSRs in the Danube deep-sea fan, Black
 1360 Sea. *Earth Planet. Sci. Lett.* 462, 15-25. <https://doi.org/10.1016/j.epsl.2017.01.006>.
 1361 Zonenshain, L.P., Le Pichon, X., 1986. Deep basins of the Black Sea and Caspian Sea as
 1362 remnants of Mesozoic back-arc basins. *Tectonophysics*. 123, 181-212.
 1363 [https://doi.org/10.1016/0040-1951\(86\)90197-6](https://doi.org/10.1016/0040-1951(86)90197-6).
 1364