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Abstract

Floods and storms are common phenomena at river mouths with some degree of wave influence. They can have a considerable impact on river-mouth sedimentation and morphological change, although studies elucidating this relationship are relatively rare. The present paper analyzes annual bathymetric changes based on surveys at the mouth of the Sfântu Gheorghe branch of the Danube river delta between 2004 and 2018, and relates bed changes with measured river freshwater and solid discharges, and wave height data. We found a strong inter-annual variability of morphological volume change which ranged from $5.23 \times 10^6 \text{ m}^3$ deposited during the extreme flood of 2006 (maximum discharge of $\sim 4000 \text{ m}^3/\text{s}$), to $-6.88 \times 10^6 \text{ m}^3$ removed during the stormy year of 2012 (wave heights $> 6 \text{ m}$). The sediment budget of the river-mouth area is modulated by the inter-annual variability of storms and floods, and can be estimated by the newly proposed Flood/Storm index based on river water or sediment discharge and wave height proxies ($R^2=0.84$). A selection of discharge and wave thresholds were simulated with Mike 21/3 by DHI (Danish Hydraulic Institute), a coupled hydrodynamic and wave model to characterize circulation and the applied bed shear stresses at this complex river-mouth sedimentary system. The model simulates 3D jet and plume hydrodynamics during floods, and the longshore current and wave dissipation during storms. Finally, field data are integrated with the Mike 21/3 model results to derive a conceptual hydro-morphodynamic model of an asymmetric wave influenced river-mouth bar during the two antagonistic phases of flood-driven

and storm-driven dynamics. During floods, the importance of current circulation and counter-plume currents in trapping sediments in a seaward-thinning depositional wedge is discussed. During storms, the sediment deposited during floods is removed by the high bed shear stresses generated by waves and currents.

Keywords: Wave-river-influenced delta; River mouth; Storms; Floods; Mouth bar; Mike 3; Delta processes

1. INTRODUCTION

Modern deltas are environmental archives of fast evolving coastal configurations, spanning, for many deltas the last 8 ka (Stanley and Warne, 1994; Hori et al., 2004; Tamura et al., 2009; Vespremeanu-Stroe et al., 2017), and involving shifting depocentres associated with river mouths and their interaction with the dispersive oceanographic forces represented by waves, tides and currents (Anthony, 2015a). Delta construction is achieved by the trapping of a large part of sediment supplied by river mouths and longshore currents close to the shoreline in shallow waters. Nevertheless, river mouths worldwide show a wide spectrum of depositional and erosional processes. Large deltas with high sediment supply and comparatively low dispersive agents may trap close to 80% of the sediment delivered by rivers close to their river mouths, forming subaqueous delta fronts (e.g. Milliman et al., 1984; Wright, 1985; Bornhold et al., 1986; Sabatier et al., 2006; Ta et al., 2002; van Maren and Hoekstra, 2005; Correggiari et al., 2005; Maillet et al., 2006; Harris et al., 2008; Xue et al., 2010; Szczuciński et al., 2013; Preoteasa et al., 2016). On the contrary, river mouths encountering basins with efficient sediment-dispersing

agents usually fail to build protruding deltas (Boyd et al., 1992; Walsh and Nittrouer 2009; Ma et al., 2008), although coastal accumulation of Holocene sediment can still be significant. Nevertheless, for both low and high dispersal systems, the initial flood sediment deposition zone near the river mouths often differs spatially from the final depocentre (e.g., Wright and Nittrouer, 1995; Allison et al., 2000; Carlin and Dellapenna, 2014; Marion et al., 2010). The distance to the nearest maximum shelf depocentre (i.e. zone of highest sedimentation over the medium- to long-term) is clearly dependent on significant wave height (H_s) and tidal range. Given sufficient sediment discharge and shelf width, depocentres are found close to river mouths, <10 m depth for H_s <1 m and tidal range <2 m (Walsh and Nittrouer, 2009); whereas for increasing H_s and tidal range, sediment can be deposited far offshore (Walsh and Nittrouer, 2009). Rivers such as the Danube or the Po have built a subaqueous delta front (e.g., Correggiari et al., 2005; Tătui and Vespremeanu-Stroe, 2017) with a sigmoid deposition shape characteristic of relatively low sediment dispersal. Nevertheless, a mid-shelf prodelta clinoform built from fine sediments, offset well alongshore from the river mouths, may be still present (Correggiari et al., 2005; Constantinescu and Giosan, 2017). Rivers draining into narrow shelves associated with active margins and small drainage basins (e.g. Columbia, Eel) are more likely to have their sediment load transported to the deeper shelf, forming mid-shelf mud deposits (Harris et al., 2005), or to the continental slope and down the submarine canyons (Savoye et al., 2009; Warrick 2014). Mechanisms such as wave resuspension and wave- and current-assisted gravity flows become important in carrying sediment offshore and alongshore (Traykovski et al., 2000; Friedrichs and Scully 2007; Harris et al., 2005; Ma et al., 2008). Such ‘wave-supported’ gravity flows, considered as a class of turbidity currents (Wright and Friedrichs, 2006), may be significantly

enhanced by storm activity (Dail et al., 2007). Long-term deposition from these processes may lead to clinoform development, as off the Po (Friedrichs and Scully, 2007).

The Sfântu Gheorghe (Sf. Gheorghe) is one of the world's better-studied deltaic lobes, and an archetype for asymmetric wave-influenced mouths (Bhattacharya and Giosan 2003; Preoteasa et al., 2016). Many authors have described different relevant aspects of the Sf. Gheorghe coastal system: coastal circulation and plume dynamics in the river-mouth area (Bondar et al., 1991; Dinu et al., 2013; Constantin et al., 2017), long-term coastal and shoreface evolution (Vespremeanu, 1983; Preoteasa et al., 2017; Tătui and Vespremeanu-Stroe, 2017), decadal shoreline change linked with storminess (Vespremeanu-Stroe et al., 2007), and longshore transport on the adjacent coasts (Giosan et al., 1999, Vespremeanu-Stroe, 2004; Dan et al., 2009). Clearly, a better understanding of the nature of the sediment dispersal system prevailing at the mouth of this branch and the sediment accumulation and redistribution processes that drive the morphodynamics of the river-mouth area are needed in order to obtain a coherent image of various aspects of the short- to long-term geomorphic evolution of the mouth of the Sf. Gheorghe. The results and their discussion may also be relevant to other wave-influenced delta distributary mouths where significant deposition and resuspension occur close inshore in response to flood and storm-wave alternation.

The objective of this paper is twofold:

- i) establish a relationship between, on the one hand, observed morphological change at the mouth, expressed as volume change (ΔV), and, on the other, water discharge (Q), sediment discharge (Q_s), and significant wave height (H_s);

- 91 ii) characterize the hydrodynamics and wave processes involved in morphological
92 change.

93
94 We address these goals by combining measured data and numerical modelling. To this
95 end, we used data from yearly bathymetric measurements (2004-2018) at the Sf. Gheorghe river-
96 mouth area and applied numerical modelling (Mike 21/3) for wave and hydrodynamic
97 simulations. Measured morphological change was correlated with the results of numerical
98 modelling to derive a conceptual morpho-hydrodynamic model during floods and storms for the
99 river mouth. We further propose a simple, index-based approach in order to explain the
100 variability of yearly sediment budget changes at the river mouth.

102 2. REGIONAL SETTING

103
104 The Danube River basin (~800,000 km²) covers a large part of Central and Southeast
105 Europe and drains a significant part of the Alps and much of the Carpathian Mountains (Fig.
106 1A). It has an average elevation of about 475 m (Stănescu, 1967), a temperate continental
107 climate with a mean annual air temperature of about 10 °C and mean annual precipitation of
108 around 600 mm/y (Malagó et al., 2017). The river has a length of 2860 km, making it the 29th
109 longest in the world and the 2nd in Europe. It discharges into the Black Sea through five main
110 outlets: Oceaakov, Bistroe and Stambul (part of the Chilia main branch, in the north), Sulina
111 (central) and Sf. Gheorghe (south) (Fig. 1).

112 The river's hydrological regime is pluvio-nival with an average discharge (Q) at the delta
113 apex of about 6800 m³/s (based on 2004-2015 INHGA data). About 5-8% of the discharge is lost

to infiltration and evaporation in the delta plain and outflow through minor outlets into the Black Sea (Bondar et al., 2011; Bondar and Iordache 2016). Discharge exhibits a moderate seasonal contrast. Floods usually occur from March to June, last for several weeks to a few months, and are linked to the stationary behaviour of Rossby waves in the Northern Hemisphere associated with strong synoptic depressions that generate persistent precipitation (Blöschl et al., 2013). The maximum Q in April 2006 was about 16,000 m³/s, and was the greatest flood in the last 100 years. Current suspended sediment discharge (Q_s) averages 640 kg/s (2004-2015 INHGA data), a threefold decrease relative to values prior to the construction of the Iron Gates dams (Preoteasa et al., 2016). The monthly suspended sediment discharge values recorded during the last floods (2005, 2006, 2010, 2013), although extreme from a Q perspective, remained lower than the mean annual Q_s of the 1840-1980 period. Although bedload has been considered as currently unaffected (Bondar et al., 2011), bedload values are not known.

The Danube is a Proximal Accumulation Dominated (PAD) dispersal system (following Walsh and Nittrouer 2009), with most of the solid discharge being deposited close to the river mouths. The Sf. Gheorghe displays a prominent river-mouth bar (*RMB*) that is asymmetric in morphology (Bhattacharya and Giosan, 2003; Preoteasa et al., 2016), with a protruding updrift crest with depths of around 1.5 - 2 m and a deeper channel downdrift (south) with depths of around 3 - 3.5 m (Fig. 2A; Suppl. 1). North of the river mouth, two to three longshore linear sand bars exhibit an offshore migration of 20–50 m yr⁻¹ (Tătui et al., 2016). The sand-mud transition during a survey in 2005 occurred at about 5 – 6 m depth in front of the river mouth (Fig. 2B), whereas a few km alongshore, the transition occurred at about 8 m depth. Generally, sorting follows the same distribution of the D50 grain size presented in Fig. 2B, with moderately to well sorted (0.2 to 1 Φ) sands, whereas silts are poorly sorted (around 2 Φ). Samples in the north and

central river mouth-area show a positive skewness (0 to 0.5 Φ), indicative of a higher proportion of fine sediments, whereas southern samples are coarse-skewed (0.3 Φ). A selection of shallow cores taken in 2004 in the river-mouth area (Fig. 2B) demonstrates the interbedded nature of the deposits. A core taken on the RMB crest (Gv1) was composed of mixed silt and sand layers, whereas a core positioned slightly inside the RMB crest (Gv4) was predominantly sandy with a few silt layers. Immediately south of the river mouth, core Gv11 showed a predominantly silty layer down to 20 cm depth and sand afterwards, whereas north of the river mouth, Gv20 was mainly sandy with a few silt layers. Flood-storm interaction is not a dominant occurrence at the river mouth, as severe storms generally occur in winter before the flood season (Fig. 3). The mouth of the Sf. Gheorghe is subject to minor tidal influence with a maximum spring tide range of 0.12 m (Bondar and Panin, 2001). Tides thus have a negligible morphologic imprint. The average significant wave height (H_s) is 0.9 m, and, usually, 20-30 storms exceed the wind speed threshold of 10 m/s and a H_s of 1.5 m each year, and just a few, if any, exceed 20 m/s and a H_s of 4 m (Zăinescu et al., 2017). The most severe are due to the passage of well-developed extratropical cyclones. The dominant northeasterly waves approach the Sf. Gheorghe shoreline at an acute angle, generating a net estimated longshore transport (LST) of sand-sized material of $\sim 1 \times 10^6 \text{ m}^3/\text{yr}$ (Vespremeanu-Stroe, 2004; Dan et al., 2009).

3. METHODS

3.1 Bathymetric data

We used yearly bathymetric measurements from 2004 to 2018, as part of the morphological monitoring programme conducted by the Sf. Gheorghe Marine and Fluvial Research Station (SCMF) of the University of Bucharest for the area > 1 m depth in front of the river mouth. The data were collected along shore-perpendicular profiles, usually down to depths of -20 m. The 2004-2013 data were acquired with a Garmin GPSMAP298C echo sounder (accuracy $\pm 0.1\%$), whereas since 2014 acquisitions have been carried out using a Valeport Midas Surveyor single-beam echo sounder (accuracy ± 0.02 m). Measurements are vertically corrected for water-elevation variability with levels from the Sfântu Gheorghe gauge station.

Based on these data, we constructed Digital elevation models (DEMs) on a 4×4 km maximum grid using Natural Neighbour Interpolation in Golden Software's Surfer with a grid cell dimension of $50 \text{ m} \times 50 \text{ m}$, and with an anisotropy of 0.7 in a North-South direction in order to better represent alongshore features. Some bathymetric surveys such as those of 2004 and 2013 covered a smaller surface. Next, DEMs of Difference (DoDs) were generated by subtracting grids to obtain bathymetric changes, and we ended by calculating the sediment budgets between surveys on overlapping grids for four areas (Fig. 2A): (i) the Danube channel (*Danube_{channel}*); (ii) the river-mouth bar crest region, a dynamic area considered as the zone of strongest morphological change in the immediate vicinity of the river mouth, with maximum depths of around 8 m (*RMB_{crest}*); (iii) the upper shoreface, from the shoreline to depths of -8 to -10 m (*Nearshore*); (iv) the river-mouth offshore area, extending from the upper shoreface down to -16 m depth corresponding to the lower shoreface and prodelta (*Offshore*). The alongshore limits of the last two are about ± 1.5 km N-S from the mouth. Estimating errors from bathymetries is difficult as sources of uncertainties are multiple and not easily quantifiable (Hare et al., 2011). These include errors induced by roll, pitch, heading and sound speed structure. A tentative

estimation of these unaccounted-for errors, for the 8.8 km² common surface retained for volume measurements and for a ± 5 cm error, yielded a total of $\pm 0.44 \times 10^6$ m³. Errors for measurable sources from the instruments used and DEM construction are low, generally near or under $\pm 0.1 \times 10^6$ m³ (Table S1).

3.2 Waves and river discharge

We used wave data from ERA-Interim, a state-of-the-art climate reanalysis made available by the European Centre for Medium-Range Weather Forecasts (ECMWF) (Dee et al., 2011). 3-hourly analysis fields are available at a horizontal resolution of 80 km; more information on the ERA-Interim dataset can be found at <https://www.ecmwf.int/en/forecasts/datasets/reanalysis-datasets/era-interim>.

The following parameters were used: significant height of combined wind waves and swell (H_s), mean wave direction (dir), and mean wave period (T). We retrieved the wave time series from a grid point located SE of the Sf. Gheorghe mouth at -35 m water depth. We assessed the quality of the ERA-Interim data by comparing them with 6-hourly recorded observations from the Gloria oil platform (~45 m depth; Fig. 1) for the period 1 Jan 1998 – 31 Dec 2007. We obtained a good Pearson coefficient of correlation, R , of 0.86 and a root mean square error (RMSE) of 0.4 m. The average H_s of the ERA-Interim data is 0.82, whereas it is 0.92 for the Gloria data, indicating a slight negative bias for higher H_s .

Discharge data comprise daily liquid discharge (Q) and monthly solid discharge (Q_s) reconstructed for the mouth of Sf. Gheorghe. Reliable daily Q and Q_s are measured only at the hydrologic stations upstream of the Danube delta because of the inability to correctly reconstruct

Q based on the water level, which can fluctuate at river mouths due to winds and waves, in addition to river discharge. As such, we derived our daily Q and Q_s data at the Sf. Gheorghe mouth from the Tulcea station (Fig. 1) by applying a correction factor of 0.53, which represents the percentage of Sf. Gheorghe Q at Tulcea Q (53%) (based on INHGA 2004-2015 monthly Q data). This is in agreement with the Delft Hydraulics Report (van Gils et al., 2006), and the ratio remained relatively unchanged between 2004 and 2015. Daily water levels at Tulcea are available for 2004-2018, whereas daily Q only from June 2015 to July 2018. To extrapolate the Q data for the missing period, we applied a rating curve. The Q for Sf. Gheorghe (Q_{sfGhe}) based on level measurements at Tulcea then becomes:

$$Q_{sfGhe} = 0.53 (1E - 7x^4 - 5E - 5x^3 + 0.0142x^2 + 10.523x + 933.92) \quad (1)$$

where,

x is water level at Tulcea.

We obtained an $R^2 = 0.9987$, using a fourth order polynomial with the aim of maximizing representativeness for higher discharges, and a RMSE of 54.5 m³/s. There are still four months in 2015 for which, in the absence of daily Q or water levels, we used average monthly values. For 2016-2018, in the absence of monthly Q_s data, we extrapolated from the $Q - Q_s$ relation in 2004-2015 ($R^2 = 0.83$, RMSE 125 kg/s), based on a power law function, given the fact that sediment concentration increases rapidly with flow:

$$Q_s = 1E - 5Q^{2.0999} \quad (2)$$

Table 1 shows representative percentiles of discharge and wave height, including Q_{75} (the 75th percentile of discharge) and H_{s90} (the 90th percentile of wave height), which were used throughout the paper.

3.3 Numerical modelling

Mike by DHI is a state-of-the-art numerical model used in coastal studies for simulating hydrodynamic, wave, and sediment transport processes. The package used in this study comprises the following modules: Mike 3 FM (Flow module), Mike 21 SW (Spectral waves) Mike 3 MT (Mud transport), with the capability of running in coupled mode. Mike 3 FM is based on the three-dimensional incompressible Reynolds-averaged Navier–Stokes equations based on the assumptions of Boussinesq and hydrostatic pressure. It consists of continuity, momentum, temperature, salinity and density equations with multiple models for turbulence (DHI, 2017a). Mike 21 SW is capable of simulating wave growth due to wind action, propagation in deep and nearshore waters, breaking, dissipation, refraction, diffraction and shoaling, non-linear wave-wave interactions, wave-current interactions (refraction, whitecapping), dissipation due to whitecapping, bottom friction and depth-induced wave breaking (DHI, 2017b). The full spectral formulation is based on the wave action conservation equation as described by Komen et al. (1996), and a instationary time formulation has been employed. In case of breaking waves, energy is extracted from the organized wave motion and is converted into turbulence, where the total production of turbulent energy equals the dissipation energy. The eddy viscosity due to wave breaking is calculated from the transport equation for turbulent kinetic energy (e.g. Deigaard et al., 1986; DHI 2017c). Mike 21/3 Coupled Model FM

enables simulation of the mutual interaction between waves and currents using dynamic coupling between the hydrodynamic module and the spectral wave module by including the radiation stress field yielded by the spectral waves simulation that is driving currents in the hydrodynamic module. Also, water levels and current field variations from the hydrodynamic module are included in the wave simulation. Current velocity has to be taken into consideration in calculating wave propagation speed of the wave action (DHI 2017a) as it generates current-induced refraction. When waves get too steep, the whitecapping dissipation process is accounted for by employing the whitecapping expression of Komen et al. (1994).

Bed shear stress (τ_b) is obtained in the Mike MT module using the Soulsby et al. (1993) maximum current τ_b for flood simulations and maximum combined current and wave τ_b for storm simulations. We use *maximum bed shear stress* because the threshold of motion and entrainment of the sediment are determined by it (Soulsby et al., 1993). Maximum τ_b is usually 2-3 times larger than average τ_b (Soulsby et al., 1993). Numerical aspects and a complete description of the model, with scientific documentation for each module, are available at the Mike website: http://manuals.mikepoweredbydhi.help/2017/MIKE_21.htm#MIKE_21/3_Documentation_

The model simulations and sensitivity tests were calibrated based on field hydrodynamic conditions measured with a Teledyne Workhorse Sentinel Acoustic Doppler Current Profiler (ADCP) in July 2017 and with offshore wave heights from the Gloria Oil Platform. The model was run with calibrated parameters and then with the final flood and storm simulations.

3.3.1 Model calibration and setup

Two simulations were carried out for calibrating the model: (1) hydrodynamic modelling with Mike 3 FM and, (2) wave modelling with Mike 21 SW. A 3D model was chosen due to the importance of stratification in driving hydrodynamics at the river mouth. The first simulation was calibrated for conditions recorded in 28 July 2017 over the Sf. Gheorghe river mouth: $Q = 1100 \text{ m}^3/\text{s}$; wind speed (ws) = 8 m/s, wind direction ($wdir$) = 270° ; and the conditions recorded a few hours before the measurements showed a ws of 6-10 m/s with $wdir \sim 0^\circ$. The model domain used for calibration covers the NW Black Sea region influenced by the Danube delta with 150 km W-E and 175 km N-S dimensions (Fig. 4A). The model employs an unstructured triangular mesh comprising five levels of grid refinement, with $\sim 400,000$ elements distributed over 15 vertical uniform sigma layers. The distance between grid nodes at the lowest level is about 45 m. The source of the bathymetry is the GEBCO 2014 30 arc-second grid <https://www.gebco.net> (Weatherall et al., 2015) for offshore areas with depths > 20 m, and bathymetry measured by the SCMF for shallower waters (2017 for calibration, 2011 for simulations).

To reduce spin-up time for the hydrodynamic simulation, initial conditions of salinity (PSU), temperature ($^\circ\text{C}$) and velocity (m/s) 3D fields were imported from Copernicus Marine, a high-resolution model for the Black Sea <http://marine.copernicus.eu/services-portfolio/access-to-products/> ($1/27^\circ$ to $1/36^\circ$ resolution), 15 days before the ADCP measurements. Daily salinity, temperature, current velocity and sea surface height were added as vertical fields on the E and S boundaries (Flather condition) of the model from the same Copernicus Marine source. Evaporation rates and temperatures were downloaded from Era Interim (<https://apps.ecmwf.int/datasets/>), and precipitation rates, short wave radiation, long wave radiation, and relative humidity from NCEP/NCAR reanalysis (<https://rda.ucar.edu/>) were

prescribed as 2D fields in the model domain as changes in temperature and salinity can have a small influence.

A relatively complex hydrodynamic situation was highlighted during the course of the ADCP survey: a saline intrusion was observed in the river-mouth area. Offshore of the *RMB* crest, a N to S directed buoyant coastal current, the Danubian Current (DC) (Fig. 1B; Fig 5 A,B from 750 to 1500 m offshore) was also detected and was reproduced by the model only when we introduced the influence of the northern Danube river mouths and their buoyant plumes. Buoyant freshwater amplifies the efficiency of wind forcing near the surface (Lentz, 2001) by decoupling the upper buoyant layer from the bottom and thus generating lower shear.

The saline intrusion was calibrated by running sensitivity tests and tuning the Vertical Dispersion parameter, which controlled most of the current velocities and structures, and which is based on a scaled eddy viscosity formulation (DHI, 2017a). The default value of 1 was set to 0.1 close to the river mouth. To avoid stronger stratification than observed for the offshore zone, we used a value of 0.6 offshore of the *RMB* crest. This agrees best with the measurements which show a strong R (Pearson Correlation Coefficient) of 0.82, a low RMSE of 0.09 m/s, and a slight negative bias of -0.03 cm/s, but led to slight underestimation of the depth of the offshore buoyant coastal current (Fig. 5). Also, a lower bed roughness of 0.02 slightly improved correlation with the ADCP profile. Although we do not dispose of salinity measurements, we observed a good agreement between the observed and modelled velocity fields, which implied that the salinity fields were also correctly modelled (Fig. 5 A, B).

The calibration mesh for waves is of lower resolution (excludes first 3 levels near the river mouth), due to the long simulation time (6 months), and because it was not necessary to include detailed nearshore cells. Mike 21 SW was calibrated for waves measured at the Gloria

Oil Platform for a total of 6 months. The Gloria record contained two of the most severe storms in the NW Black Sea (January 1998 and January 2004) over the period of measurement, with local wind data from the Gloria wind station (adjusted at an elevation of 10 m after Rusu et al., 2014). ERA Interim data (H_s , wave period, w_{dir} , directional standard deviation) were prescribed at the E and S boundaries. The model included quadruplet interactions, due to the importance of this process for deep water waves. Default conditions resulted in slightly lower extreme wave heights, and as such, we lowered the Coefficient of Dissipation (C_{dis}) for whitecapping from a default of 4.5 to 4. For the nearshore, we used the Nikuradse roughness value of 0.02 recommended by the Halcrow report for the Romanian coast (Halcrow, 2011). The final run resulted in very good agreement of modelled and measured wave heights $R=0.93$, correctly estimating slope of the relationship and the extreme wave heights (Fig. 5C): the RMSE is 0.33 m (0.29 for the mean normalized RMSE) and the bias is low (0.06), similar to what other studies obtained for the NW Black Sea (e.g. Rusu, 2015).

3.3.1 Flood and storm simulations and setup

Different thresholds of low, medium and high discharge (Q : 1000, 2200, 3200, 4000 m^3/s) and wave height (H_s : 1.5, 2.5, 4, 6 m) corresponding to different percentiles were modelled with the Mike model in order to investigate the hydrodynamics during floods and storms. We used for the flood and storm simulations a high-resolution, 3-level mesh with ~680,000 elements distributed in 20 uniform vertical sigma layers, and a distance of about 30 m between the triangulated grid nodes in the most detailed level and an element surface of ~650 m^2 (Fig. 4B). The river channel was prescribed with a quadrangular grid to improve flow at boundary

conditions due to errors appearing from the use of sigma layers. Higher order solutions for time and spatial integration, a horizontal eddy viscosity Smagorinsky formulation (0.28), and a vertical - k-epsilon model for turbulence were employed, the effect of Coriolis included and Flather conditions prescribed at open boundaries.

We ran simplified flood simulations in Mike 3 FM, neglecting the influence of wind, waves and of the northern Danube river mouths, with increasing Q which reached a maximum of 4000 m³/s. Flood simulations were initialized with averaged salinity fields for the month of April 2005 from Copernicus Marine; temperature influence was excluded, and the model ran for one week for each condition. Hydrodynamic conditions seem to have reached a stable point of diminishing change at the end of the week in the absence of mixing due to winds and waves.

The storm simulations were run in coupled Mike 3 FM and Mike 21 SW modules with typical conditions of Q (1250 m³/s) that appear to be most common during storms in December - February. In the final storm simulation, baroclinic effects were excluded because of the low impact on circulation during storm conditions (Fig. 3). The storm was modelled as a whole event with 12 hour per wave threshold and fixed 45° direction specified at the open boundaries, thus mimicking realistic increase in the peak of the storm which occurs in one or a few days. The currents generated were combined wind and wave currents, and the wind was prescribed with 8, 12, 17, 25 m/s following a relationship between H_s and w_s at the Sulina weather station (Fig. 1). The storm simulation included diffraction, triad and quadruplet wave interactions, a wave breaking approach based on the functional form of Ruessink et al. (2003), bottom friction Nikuradse roughness of 0.02, and a whitecapping C_{dis} coefficient of 0.4 obtained from calibration. It should also be noted that the model is not calibrated for depth-limited breaking. The SW simulation employed a logarithmic discretization of 25 frequencies whereas the wave

action spectrum was discretized in 12 directions on a 0 to 180 degrees directional sector, which resulted in a 15° resolution, sufficient for wind waves. We also reduced the maximum number of levels in transport calculation to 4, necessary in conditions with strong wave breaking.

4 RESULTS

In the following sections, each analyzed time period between successive measurements is named based on the last year of measurements, such that 2004-2005 becomes T_{05} and 2017-2018 becomes T_{18} . We first describe the relationship between bathymetric change and flood and storm forcing.

4.1 Bathymetric change and hydrodynamic forcing

A whole range of bathymetric changes was recorded over the 2004-2018 period (Fig. 6; Suppl. 2). The net difference in volume of the river mouth area between two successive measurements (ΔV) ranged from equilibrium ($\Delta V \sim 0$), to moderate ($\Delta V: \pm 0.5$ to 2×10^6 m³) and extreme $\Delta V: +5.2 \times 10^6$ m³ (T_{06}) to -6.8×10^6 m³ (T_{12}). Considered extreme periods T_{06} (positive) and T_{12} (negative) are around a distance of 2 standard deviations from the mean. Accumulative and erosive periods show opposite morphodynamic patterns. Both are characterized by a zone of high ΔV in the immediate vicinity of the river mouth at 4 to 8 m depth and decreasing ΔV further offshore, alongshore and inside the Danube channel, but with opposite signs (Fig. 6).

The T_{05} ($\Delta V = 3 \times 10^6$ m³) and T_{06} ($\Delta V = 5.2 \times 10^6$ m³) periods are dominated by the occurrence of exceptional floods that peaked at 3500 and 4000 m³/s respectively, during

moderate storminess (Fig. 7A, B, C). Under these conditions, the *RMB* crest migrated offshore by ca. 300 m, as a cumulative effect of the succession of two extraordinary hydrological years, and the depositional layer varied from a maximum of 1 – 3 m on the seaward side of the *RMB* crest to about 0.5 – 1 m offshore to the prodelta. Slight erosion (≤ -0.5 m) occurred in the river-mouth channel. Moderate flood years showed similar patterns (offshore mouth-bar migration, offshore deposition), but of lower magnitude. Nevertheless, during the low flood and low storm periods (T_{07} , T_{08}), 65% of the new deposited sediment (ΔV : $T_{05}+T_{06}$) was removed.

A large sediment volume ($\Delta V = -6.8 \times 10^6 \text{ m}^3$), comparable to that recorded post-major floods, was swept away during T_{12} , a year characterized by multiple episodes of high waves ($H_s > 4$ m) and an extreme storm ($H_s > 6$ m) that lasted from 5 to 10 Feb 2012. The largest erosion in T_{12} occurred up to -8 m depth on the seaward side of the protruding *RMB* crest, with extreme erosion values reaching the -2 to -4 m depth interval. Offshore, the bed-level changes varied from -1 m at the base of the crest to -0.5 m. Accumulation occurred only in the Danube channel. During low to moderate storm years the maximum erosion zone affected constantly the mouth bar crest, linked to low-to-moderate waves breaking on the protruding crest. The *offshore* zone experienced a low but still noteworthy morphological change (-0.1 to -0.5 m). As total erosion increased, the maximum erosion depth moved offshore.

It is also obvious that total ΔV was highly influenced by the strength of the hydrodynamic forcing: the highest positive ΔV occurred during the exceptional flood years, and the highest negative ΔV during the extreme storm year. These results throw light on the high morphological variability that can occur at wave-influenced river mouths, driven by the variability and timing of forcing events.

4.2 Flood/Storm index

To relate morphological changes to the forcing, we propose a Flood/Storm index (FL/ST_i) that can be computed using proxy data: FL/ST_{dur} is the ratio of the duration of floods (Q) and storms (H_s) above a certain threshold (see Table 1 for threshold definition; and the discussion chapter 5.2 for the rationale for using this index):

$$FL/ST_{dur} = \frac{Q_{75}}{H_{s98}} \quad (3)$$

where Q_{75} = duration of discharge above the 75th quantile, H_{s98} = duration of significant wave height above the 98th quantile, and FL/ST_{sed} the ratio of sediment discharge above 125 kg/s and cumulative storm wave power above $H_{s2.5}$ (significant wave height of 2.5 m).

$$FL/ST_{sed} = \frac{Q_s}{Wp} \quad (4)$$

where, Q_s = total sediment discharge > 125 kg/s (50th quantile), and Wp = cumulative storm wave power above the 98th quantile.

Wave power calculations were based on Moritz and Moritz (2006). FL/ST_i is calculated for quasi-yearly periods representing intervals between the analyzed bathymetric measurements. Thresholds were chosen by finding the quantiles that yielded the highest correlation with measured ΔV . Thus, the Q_{75} (the duration of discharge above the 75th quantile or 2200 m³/s for the Sfântu Gheorghe branch) and the H_{s98} (the duration of significant wave height above the 98th quantile or 2.5 m) were found to be most representative. Q_{75} is the highest flood quantile with durations >0 for every measurement interval. It was also necessary to use a threshold for

sediment discharge to obtain the best correlation; this was $Q_s = 125$ kg/s, the Q_s average for the Sf. Gheorghe being ~ 160 kg/s. This requirement can be explained by bedload transport which results in the mobilization of sediment from the river channel into the mouth area and which is insignificant outside medium and high discharge conditions, and by a possible increased flocculation rate during floods, a process hinged on particle collision rate (Winterwerp and van Kesteren, 2004). It is expected that the FL/ST_{dur} , which uses only liquid discharge (Q), is only valid when the Q and Q_s relationship remains unchanged. FL/ST_{dur} and FL/ST_{sed} show good correlation with the measured budget changes, with R^2 of 0.8 and 0.84, respectively, for the whole river-mouth area and R^2 of 0.79 and 0.75 for the RMB_{crest} (Fig. 8).

Even though the overall correlation is good, there may be some significant differences between measured and predicted ΔV which can amount to more than 1×10^6 m³/yr during certain years. In 5.1.1 and 5.1.2, thresholds are discussed in relation with the corresponding hydrodynamics, and may shed some light on the ΔV complexity which is not taken into account by this approach.

4.3 Extreme events

4.3.1 Floods

Simulations with different discharge levels (Q_{15} to $Q_{99.7}$, see table 1 for Q thresholds), and no wind or wave influence, reveal that the circulation in these conditions is dominated by changes in Q , the interaction of flow with bathymetry, and buoyant effects due to the differences in salinity (Fig. 9A, B). The circulation is highly stratified, exhibiting a clear hypopycnal regime. In the

absence of enhanced mixing by winds and waves, there is a sharp horizontal salinity gradient delimiting the buoyant river jet and plume from the Black Sea waters.

The plume liftoff, i.e. the hydraulically forced detachment of buoyant freshwater from the channel bottom or seabed (Poggioli and Horner Devine, 2018) occurs close to the river mouth, and marks the start of the rapidly expanding buoyant plume (Fig. 9). The liftoff point is inside the river channel for low discharges, i.e. Q_{15} (1000 m³/s), which allows for the upriver saline intrusion, and jumps over the *RMB* crest with increasing Q . Diminishing offshore displacement is observed with increasing Q : the liftoff point location at Q_{95} is not much different from $Q_{99.7}$. Plume liftoff location is important in generating an area of lower flow close to the bed and as such, a sharp decrease in τ_b . Further implications of the plume liftoff point, and the secondary circulation on sedimentation will be treated in the discussion.

The large discharge during floods, flowing through a relatively narrow channel outlet (500 m width, 5 m depth), creates strong frictional effects on the *RMB* when depth decreases to 1.5-2.5 m. Frictional effects with the *RMB* bed and especially with its crest cause jet spreading and current deceleration. The main morphological features (bathymetry and river channel orientation) generate preferential steering to the right of the current, complemented further downdrift by Coriolis and Coandă effects.

Simulations for Q_{75} (2200 m³/s), Q_{95} (3200 m³/s) and $Q_{99.7}$ (4000 m³/s) discharges show the Sf. Gheorghe turbulent river jet and plume dynamics during typical and extreme flood conditions. Velocities for Q_{75} rapidly decrease from 1 m/s to 0.5 m/s immediately after the *RMB* crest and to 0.2 m/s offshore, and increase for $Q_{99.7}$ at 1.8 m/s in the river channel, 0.6 m/s after the *RMB* crest and 0.4 m/s offshore. There is a marked decrease in the surficial current velocities

(of 20-30%) at the plume liftoff point, and then a similar increase in velocities after the *RMB* crest due to the convergence with the secondary bottom current directed onshore (Fig. 9A, B).

4.3.2 Storms

The combined fluvial jet current and longshore current (*LSC*) was simulated for different H_s thresholds ($H_{s1.5}$ to H_{s6}), at fixed 45° direction and Q discharge of $1250 \text{ m}^3/\text{s}$ (Fig. 9). Waves start breaking on nearshore bars and on the *RMB* crest to generate the longshore current (*LSC*), which is responsible for longshore sediment transport (*LST*). The *LSC* at $H_{s1.5}$ appears as a weak current, mostly present on the first (proximal) and second bars at the updrift and downdrift coasts with velocities of $\sim 0.5 \text{ m/s}$ (Fig. 10). The northern *LSC* discharges into the river outlet, whereas the *LSC* generated over the *RMB* crest passes over it due to the lower depths $< 2 \text{ m}$ and to the protruding, curved isobaths at the northern part of the *RMB*. Further south (Fig. 10, 4971500) the border between the *LSC* and river jet is not clear, although there is an enhancement in flow at the outer edge of the river jet at the *RMB* crest (Fig. 10A, B). As wave heights increase, wave breaking occurs at progressively greater depths, the dissipation zone widens, and the *LSC* mostly grows by extending its influence offshore. The *LSC* appears to be on the *RMB* crest at $H_{s1.5}$ (Fig. 10B). Waves start breaking at $H_{s2.5}$ on the distal bar on the northern shoreface, and with increasing depths (H_{s4} , H_{s6}) on the upper shoreface slope outside the nearshore bars. As such, maximum velocities are attained at depths of ca. 2 m ($H_{s1.5}$), 4 m ($H_{s2.5}$), 6 m (H_{s4}) and 8 m (H_{s6}) and range from 0.5 m/s ($H_{s1.5}$) to 2 m/s (H_{s6}) (Fig. 10). *LSC* in $H_{s2.5}$ appears discharging partly into the river mouth area, but mostly passing over the *RMB* crest, whereas complete bypassing of

the *RMB* is noticed only with higher H_s . Most of the high waves are dissipated before arriving at the lower depths (Fig 10C) and this explains the offshore increase in current velocities.

The circulation is alongshore-dominated, with a highly downdrift-deflected river jet even at $H_{s1.5}$. The *LSC* interacts with the river jet, although much of the *LSC* over the *RMB* is explained by the morphology of the latter. Nevertheless, the jet appears to have some effect on the river-mouth hydrodynamics, tempering current velocities in the northern part of the river outlet (Fig. 10A). Although not investigated here, higher Q is expected to enhance this effect. Downdrift of the river outlet, the combined jet - *LSC* velocities increase with higher H_s .

5. DISCUSSION

The fate of sediment dispersed from the river into the coastal ocean involves at least four processes: supply via plumes, initial deposition, re-suspension and transport by marine processes, and long-term net accumulation (Wright and Nittrouer, 1995). At present, most of the deposited sediment in the Sf. Gheorghe river-mouth area can be quickly remobilized in the following months and years, regardless of the occurrence or not of high storm waves ($H_s \geq 4\text{m}$). As such, this river-mouth area functions as a temporary depocentre, with most of the captured sediment being then remobilized and transported to the south. The following discussion aims at clarifying the deposition and resuspension processes and at linking measured morphological change with the hydrodynamic data, obtained with the aid of numerical modelling with the Mike 21/3, in order to derive a conceptual hydro-morphodynamic model of an asymmetric wave-influenced river-mouth bar during the two antagonistic phases: (1) flood-dominated; (2) storm-dominated (Fig. 11).

5.1 Processes explaining sediment deposition and erosion patterns

5.1.1 Floods

Floods are characterized by large increases in Q and Q_s at the river mouth which generate changes in hydrology and, ultimately, in the topography of the bed. Interactions between the river effluent with bathymetry and buoyancy effects govern the spreading rate of the river jet, the plume liftoff point, and the bottom plume return current, and ultimately, sedimentation patterns. Sediment settling from the plume produces a seaward-thinning and seaward-fining depositional wedge, which reflects the processes involved in sediment transport and deposition (Fig. 6: T₀₅, T₀₆; Fig. 11). The recorded volumes and bed change in the course of T₀₅ and T₀₆ are remarkably similar to the volumes and bed changes during the exceptional 2004 Rhône flood (Maillet et al., 2006) which peaked at 12,000 m³/s but lasted only a few days.

Stratification is responsible for a reduction in turbulence which can increase the trapping rate by an order of magnitude for silt-size sediment particles (Geyer, 1993). The plume return flow present even during high river discharge contributes in retaining sediment near the shore, as in the case of the Mekong (Eidam et al., 2017). The bottom convergence of flow at the plume liftoff zone is a pervasive and important mechanism for trapping sediments and maintaining the turbidity maximum in estuaries (Schubel and Carter 1984; Dyer 1995; Burchard and Baumet 1998; Geyer et al., 2004). In a modelling study, Thanh et al. (2017) explored suspended sediment dynamics with or without salinity gradients and found that suspended sediments are deposited significantly more rapidly nearshore when salinity gradients are present. This is attributed to two

mechanisms: the onshore bottom circulation (the plume return flow), and the inhibition of sediment spread throughout the vertical water column caused by a reduction in the vertical diffusivity. Even though the plume return flow has low velocities (0.1 - 0.2 m/s) its depth is a few times larger than the buoyant plume above, and this suggests that the former, when present, may greatly favour sequestration. Modelling on the Danube delta coast in realistic conditions showed that increased discharge and increased mixing due to wind push seaward the plume liftoff point to larger depths, thus, reshaping the interaction of the surface plume current and the plume return flow (Bajo et al., 2014), although high stratification may still take place even during extreme flood discharges, as in the case of the Po (Tesi et al., 2011). Notably, the modelled plume liftoff zone is in agreement with the maximum sediment deposition recorded at the Sf. Gheorghe mouth (Fig. 6, T₀₅, T₀₆; Fig. 9A). The bedload and resuspended sands from inside the *RMB* channel due to high τ_b (Fig. 9D) may have contributed significantly to the deposition just outside the *RMB* crest when τ_b decreases abruptly (Anthony 2015a), and fostered the large migration of the *RMB* crest during T₀₅ and T₀₆ (Fig. 6). Additionally, the lower τ_b after the mouth bar crest ensures the preservation of both sands and silts settling over the bed, generating and preserving the *RMB* sand-silt lamination signature. Mapping of grain-size parameters (Fig. 2B) following flood conditions (summer of 2005 and 2006) shows persistent patterns of sand-mud transition in which the contact between the two fractions occurs at -6 m to -8 m in the central sector, and rises to -3 m to -4 m on the lateral northern and southern flanks, in accordance with the position of the liftoff point and the vertical distribution of the current speeds (Fig. 9B, C). It is noteworthy to question the southern limit of the sandy surficial cover which marks the shortest transition from sands to fine silts and clays (100-150 m), and which occurred on relative flat or gently sloping surfaces in the absence of any prominent features (crests,

slopes). Moreover, a 300-400 m wide band of fines occur immediately seaward of the bar zone on the southern (downdrift) coast. This grain-size pattern matches well with the current speed simulations during floods (Fig. 9B) and with the measured deposition (Fig. 6), and seem to be the combined effect of current deceleration and convergence with the plume return flow and of flocculation.

Flocculation processes can significantly restrict transport distances and enhance deposition of silts and clays near river mouths (Dalrymple and Choi, 2007; Hill et al., 2000; Fox et al., 2004a; Milligan et al., 2007, Geyer et al., 2004; Nowacki et al., 2012). The net result is an overall increase in the size and settling velocity of the fine-grained material. Flocculation can increase the falling velocity of particles by several orders of magnitude (Kranck, 1980), and can induce settling velocities in excess of 1 mm/s (Fox et al., 2004a; Milligan et al., 2007). The Po river plume loses most of the suspended load within 1–2 km of the mouth (Fox et al., 2004b). Constantin et al. (2017) found that the spatial extension of the Danube sediment plume is limited to high turbidity close to the river mouths. The interbedded sand-mud deposits that form the river-mouth bar suggest intense flocculation processes even close to the river mouth (Fig. 2B), akin to those of the low-energy shelf setting of the Po (e.g., Fox et al., 2004a; Tesi et al., 2011). Nevertheless, deposition of a few cm can be recorded after floods even at >30 m depth (Frascari et al., 2006) or 10 km offshore of the river mouth (van Maren and Hoekstra, 2005). The offshore edge of our study area (-16 to -20 m), recorded a maximum of 20 to 50 cm bed change, suggesting significant deposition on the prodelta as well.

Sediments from floods are expected to reach greater deposition depths due to increase in sediment residence time caused by increased plume thickness and plume velocity (Milligan et al., 2007; Geyer et al., 2000). Additionally, strong wind-generated currents and wave-induced

turbulence during summer may act to hinder sedimentation and break flocks (Dalrymple and Choi, 2007), and to provide momentum in generating buoyant coastal currents that favour sediment transport. During windy conditions, an episodic buoyant coastal current, connected to the Danube mouths, enhances mixing and circulation alongshore (Fig. 5 A, B; Dinu et al., 2013). Additional complexity in sediment retention during floods is expected from interaction with buoyant coastal currents formed at adjacent river mouths and with episodic wind and wave events.

5.1.2 Storms

During storm conditions, waves control resuspension and sediment transport (Dufois et al., 2014), and usually create shear stresses of an order of magnitude higher than currents (Harris et al., 2008). With decreasing depths, the bed is exposed to increasing circular orbital motion which also becomes more elliptical close to the bed (Knauss, 1997), leading to greater near-bed wave orbital velocities up to the breaking point (Pratson et al., 2007). Wave resuspension and current transport act in concordance, and most of the resuspended sediment travels in dilute suspension downdrift. The different effects of waves, currents, and different sediment grain sizes at the Sf. Gheorghe river mouth can be roughly approximated to occur in two scenarios: (i) combined *LSC*, wave shear stresses and wave breaking acting on a predominantly sandy bed (surf zone); ii) combined *LSC* and wave stresses acting on a predominantly fine sediment bed (lower shoreface, prodelta) (Fig. 10).

Although defining critical shear stresses (τ_{b_crit}) requires caution (Vanoni et al., 1964), a value of 0.9–1.6 τ_{b_crit} seems to be sufficient to explain the resuspension and transport of very

fine sand as suspended load (van Rijn, 1984; Nowacki et al., 2012). Furthermore, studies in natural settings appear to be in relative agreement that a threshold τ_{b_crit} of silt is around 0.2 to 0.4 N/m² (Xu et al., 2014; Sha et al., 2018). When sediment is suspended at τ_b of 0.2 N/m², the quantities involved are limited (Xu et al., 2014), whereas fine sediments exposed to high shear stresses ($\tau_b > 2$ N/m²), may shift the bed into a more erosive state (Amos et al., 1992), indicating that increase in τ_b strongly correlates with increasing sediment remobilization rates. It seems that $H_{s1.5}$ conditions permit fine sand resuspension on the shoreface, and limited fine resuspension from the lower shoreface (Fig. 10D), whereas τ_b for $> H_{s2.5}$ can significantly remobilize fines from deeper waters. Stronger *LSC* velocities and τ_b increase and extension are in agreement with the increased erosion occurring at the *RMB* during storm-dominated periods (Fig. 11; Fig. 7, T₁₂, T₁₇), and explain the changes in the recorded morphology. Preliminary morphological simulations with Mike MT (Mud transport Module) for the Sf. Gheorghe river mouth show that the bed erosion that occurred in 2011-2012 can be explained only by including in the model fine sediments which are dominantly eroded in deeper areas (Zăinescu et al., 2018). Comparatively, “only sand” simulations differed in both patterns and magnitude with the measured data. This is not surprising, as sedimentological data based on shallow cores and surface samples show that cohesive sediments are dominant below -4 m to -8 m depths in front of the river mouth (Fig. 2B), with some seasonal variations. Most of the world’s large deltas are dominated by fine-grained sediment debouching at their mouths, and there is a need to pay greater attention to the morphodynamic significance of these fine sediments at wave-influenced delta fronts.

Cross-shelf transport to greater depths has been found to occur due to gravity-driven flows (wave-assisted) resulting from wave-induced suspension that maintains thin near-bed turbid layers during storms (Cacchione et al., 1999; Ogston and Sternberg, 1999; Wright et al.,

1999; Nittrouer, 1999; Harris et al., 2005; Falcini et al., 2012). This is observed mostly for dispersal-dominated systems (cf. Walsh and Nittrouer, 2009) but has also been recognized for the Po delta, a proximal-accumulated-deposition (PAD) system (Traykovski et al., 2007), similar to the Danube dispersal system. Nevertheless, alongshore sediment transport due to wave resuspension and wind-forced mean currents was an order of magnitude larger than cross-shore flux associated wave-supported gravity flows (Traykovski et al., 2007), which may indicate that most sediment near the Sf. Gheorghe river mouth is transported alongshore by currents in the form of dilute suspension, similar to the Po (Nittrouer et al., 2004; Traykovski et al., 2007), although the model in the present paper still needs more calibration with in-situ measurements. However, downslope transport is not excluded by strong isobath-parallel currents (Ma et al., 2008). Some dispersal systems may thus be more alongshore-dominated (e.g. PAD systems) whereas on other systems, which are more wave-dominated, down-slope wave-supported gravity flows may have a larger contribution. Submarine slides and slumps have been identified in front of the Sf. Gheorghe river mouth using a sidescan sonar (Alexandru Ionescu, Marine Research – pers. comm.). Similar mass failures are present on the Rhône (Maillet et al., 2006) and Po (Bosman et al., 2014) river mouths, albeit with apparently secondary influence on cross-shore transport. Compaction and the action of waves on sediments of different consolidation (Guillén et al., 2006), and biological processes enhancing or limiting erodibility (Xu et al., 2014), should add a further degree of complexity to bed response during storms.

5.2 The importance of extreme events and insight from the Flood/Storm index:

The cycling processes of flood deposition and storm resuspension of sediments are embedded in the very nature of deltas. At the Yangtze river mouth, monthly deposition rates during flood periods are an order of magnitude higher than long-term accumulation rates, which suggests that a major portion of the river-derived sediment is eroded periodically (Liu et al., 2007). The net 2018-2004 bed change at the Sf. Gheorghe river mouth is much lower than would be the combined yearly changes measured at this river mouth (Suppl. 2). A stratigraphic analysis of the asymmetric wave-influenced lobe of the Brazos delta (Gulf of Mexico coast) has shown that flood deposition and storm resuspension are an integral part of the resultant stratigraphy, (Rodriguez et al., 2000), leading these authors to propose that the Brazos delta is a “flood-and-wave” feature, with some degree of control from large-scale climatic teleconnection cycles (Fratelli 2006). Reductions in flood frequency caused significantly less deposition at the Rhône river mouths (Provansal et al., 2015). It seems thus that the interplay between floods and storms, as suppliers and removers of sediment at river mouths, drives most of the morphological changes in many delta river-mouth systems. In turn, changes in flood and storm frequency and magnitude are driven by either natural variability or anthropogenic causes, the effects of which can greatly influence deltaic morphodynamics on many timescales.

Predicting deltaic morphology with simple indexes based on relative ratios of oceanographic vs riverine forcing has always been tempting for coastal researchers, as shown by the diversity of deltaic indexes. First attempts by Wright and Coleman (1973) linked river- and wave-dominated delta morphology to water discharge and wave power, and were followed by Galloway (1975) who proposed the famous tripartite scheme with the inclusion of tidal effects. The Asymmetry Index (A), proposed by Bhattacharya and Giosan (2003), explores the asymmetry controlled by the *LST* and *Q*. Further studies established simple conceptual

representations of the sediment balance at river mouths, which is envisaged in terms of the sediment arriving at the mouth (Q_s) versus the magnitude of the LST . These are the fluvial dominance ratio (R) (Nienhuis et al., 2015), which predicts deltaic plan shape morphology, and the sedimentary index (Si) (Preotasa et al., 2016), which can capture the increase in the influence of wave-driven sediment circulation on river mouth morphology occurring in the past centuries at the Sf. Gheorghe river mouth.

The FL/ST index applied in this study reveals that river mouth ΔV responds according to flood and storm variability. Most importantly, the index shows that for the investigated time period there is a linear response of bathymetric changes in relation to the relative influence of flood and storm events in spite of the bewildering complexity of the river-mouth environment (Anthony, 2015a). This index has the advantage of using simple data of offshore H_s , as opposed to LST , which may be difficult to compute and which ignores fine sediment resuspension. As a rule of thumb, the index suggests that for every 1 hour of $H_s > 2.5$ m, 15 hours of $Q > 2200$ m³/s at the Sf. Gheorghe river mouth are needed to counterbalance the sediment removal by storms. This suggests that the observed bathymetric yearly changes are largely the result of the cumulative effects of storm and flood events. These variations are further accentuated where flood years and high storminess years are negatively correlated, such as in the sf. Gheorghe case. At least for this river mouth where floods and storms occur asynchronously, hydrodynamic interactions at the river mouth are not of first order significance (Fig. 2C; Fig. 3). The timing of these events is largely governed by continental-scale constraints, whereas smaller river basins are more likely to experience both floods and storms at the river mouth from the same atmospheric disturbance. Currently, the effects of storm and flood timing, either synchronous (e.g. Rhône:

Boudet et al., 2017), asynchronous (e.g. Danube, Mekong: Thanh et al., 2017) or mixed (e.g. Red River: van Maren and Hoekstra, 2005) are not clear.

Q_{75} and H_{s98} events can account for most of the variability produced in the river-mouth area ($R^2 = 0.84$). This is unsurprising as floods are responsible for most of the sediment transport in rivers (Marion et al., 2010), and higher waves ($\geq H_{s2.5}$) are necessary to explain the higher erosion of the bed at the lower shoreface, whereas waves as low as $< H_s$ 1.5 m may not be able to apply sufficient τ_b . In addition, since longshore transport formulas describe a power-law relationship between wave heights and LST magnitude, a doubling in H_s causes an order of magnitude increase in longshore sediment transport in most formulas (Mil-Homens et al., 2013; van Rijn 2014; Shore Protection Manual, 1984), without losing view of the fact that further large variability in LST magnitude can be caused by slight variations in other input parameters (Pinto et al., 2006).

RMB variability response to a $H_{s2.5}$ threshold, in comparison with including the Wp for a whole year, may be explained by a balance between frequency and impact of these events. Validity of a similar threshold for other study sites may be possible, especially for wave-influenced river mouths in similar conditions, which are abundant in the Mediterranean basin (Besset et al., 2017). Nevertheless, the relationship embodied in FL/ST_i may be susceptible to change at longer timescales due to morphologic adjustments of the river-mouth area.

6. CONCLUSIONS

The morphological evolution of a wave-influenced river mouth was monitored over the last 15 years, and the observed changes related to external forcing. Simplified conditions for hydrodynamic modelling during relevant flood and storm thresholds revealed the main processes responsible for morphological change. Nevertheless, the complexities and exact contribution of the resulting circulation that can be attributed to each process in sediment transport still remains open for future studies. There is a need to know where the totality of the sediment during floods gets deposited, how this deposition is affected by wind-influenced processes during the flood season, and what the role of stratification is in sediment deposition close to the mouth. For storms, there is still a knowledge-gap on the contribution of fine sediments to delta-front morphodynamics. Moreover, the question of where the sediment gets redistributed after removal from the river mouth area is still a question open to debate, with some studies indicating sand trapping in spits as possible effective mechanisms of deltaic reorganization (Nienhuis et al., 2013; Anthony 2015b; Besset et al., 2017), although this suggestion is not based on a complete assessment of sediment budgets. Nevertheless, aspects such these govern the response of deltas to changes in Q_s , storminess, and sea-level rise, and should spark further research interest.

The conclusions from our investigation are summarized as follows:

(1) A conceptual model was derived that emphasizes the role of floods and storms in driving morphological change at the river mouth.

(2) During floods, a seaward-thinning depositional wedge is formed (from a few m to cm-scale) that is related to plume transport and sediment settling.

(3) During storms, high bed shear stresses (τ_b) generated by combined currents and waves remove the sediment deposited during floods.

(4) A new proposed index (FL/STi) based on storm and flood thresholds explains the sediment budget variability at the Sf. Gheorghe river mouth and reveals a linear response in bathymetric change to the relative influence of storms and floods.

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Figure and table captions

Table 1. Discharge and storm percentiles used throughout the paper and their corresponding thresholds. Q for Sf. Gheorghe is $\sim 1/4$ of the total Danube Q .

1085

1086 Figure 1. The Danube River basin and the Black Sea (A), Danube delta, location of study area,
1087 and the data used throughout the paper (B). The Danubian Current is shown in relation to north
1088 (downwelling, blue) and south (upwelling, red) winds. Wave rose from ERA5data.

1089

1090 Figure 2. Sf. Gheorghe river-mouth bathymetry (A), the areas retained for the sediment volume
1091 calculations (delimited by dashed lines). Grain-size distribution (D50) in the river mouth area in
1092 July 2005 and a selection of shallow cores from 2004 (B). Monthly discharge and wave climate
1093 (C).

1094

1095 Figure 3. Frequency of discharge and wave height (top) in hours/yr. Frequency of wave height
1096 and direction (down). Era Interim data.

1097

1098 Figure 4. Meshes used for Mike modelling.

1099

1100 Figure 5. Comparison of measured data and model results: current velocity (A); current direction
1101 (B), for the ADCP transect shown in fig. 2; wave heights (C).

1102

1103 Figure 6. Spectrum of bathymetric change and selected bathymetric differences illustrating
1104 extreme, intermediate, and low change. T_{05} to T_{18} represent the time intervals between
1105 bathymetric differences, $T_{05} = 2004 : 2005$.

1106

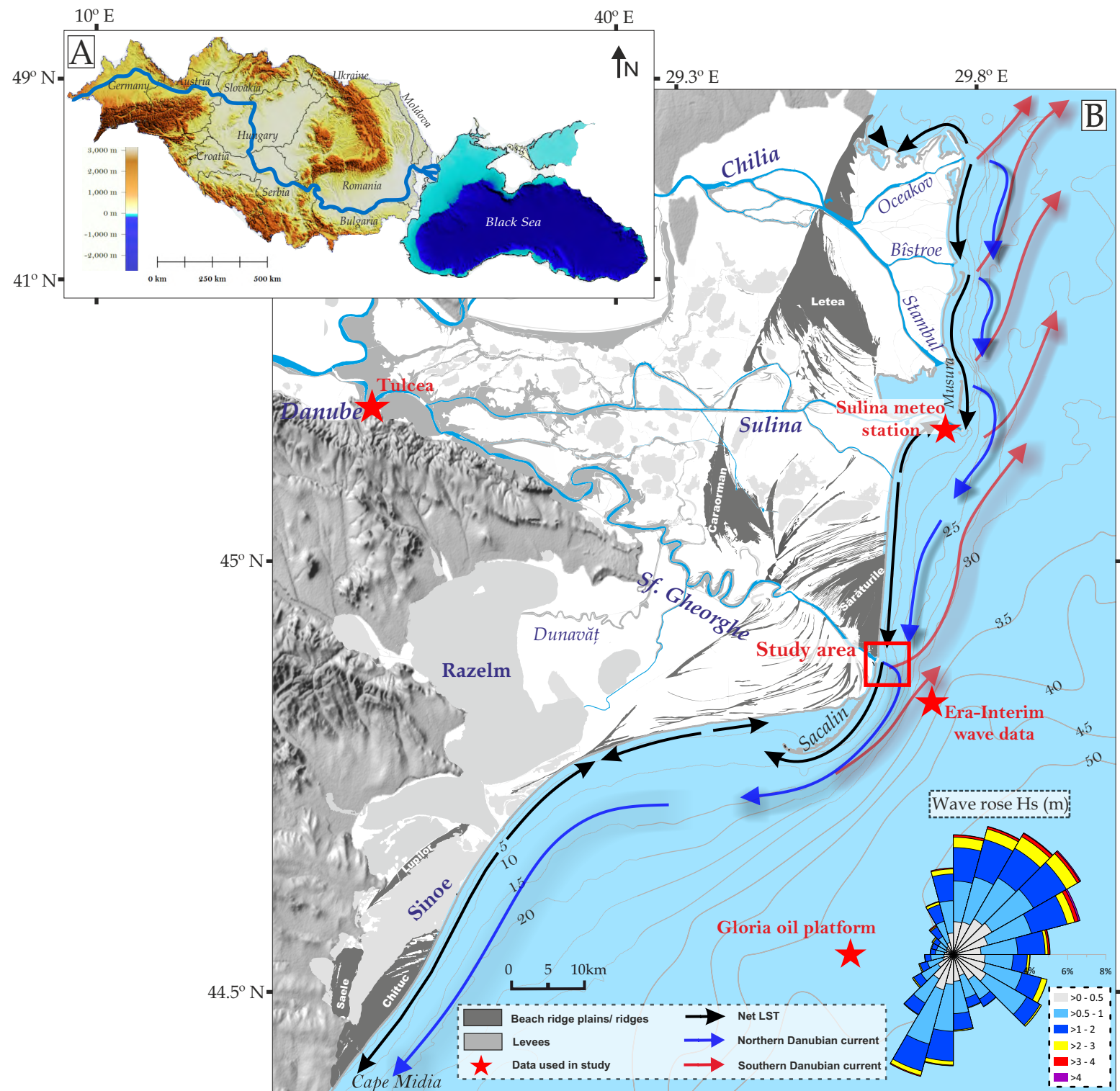
Figure 7. Time series of liquid discharge Q (red line), solid discharge Q_s (red bars) (A); 6-hourly offshore H_s and cumulative storm power for each period (B); duration of selected flood and storm percentiles; note that each percentile has its own scale (C); volume changes for each of the areas shown in Fig. 2. (D); positive, negative and net volume changes (E). T_{05} to T_{18} represent the time intervals between bathymetric differences

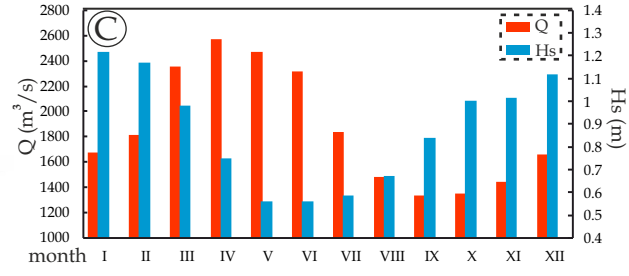
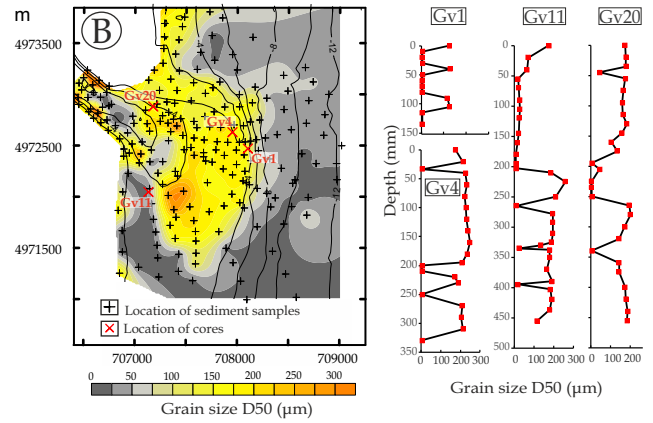
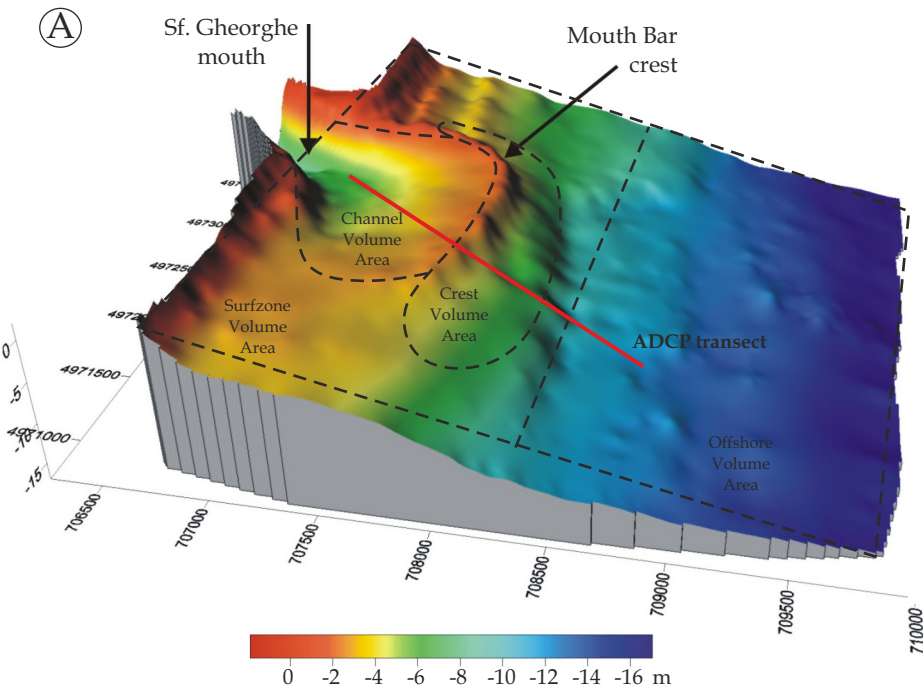
Figure 8. Scatter plots of volume change (ΔV) and FL/ST_i for the whole river mouth area (ΔV Total), left plots; only for the crest area (ΔV Crest), right plots; FL/ST_{dur} based on duration proxies (top); FL/ST_{sed} based on wave power and sediment discharge proxies (bottom). The blue area shows the 95% confidence interval for the linear fit.

Figure 9. Results of flood simulations for different thresholds. Surface velocity fields (A), vertical fields of velocity at 4972000 latitude (B), vertical fields of salinity at 497200 latitude (C), current bed shear stress (D).

Figure 10. Results of storm simulations for different thresholds. Surface velocity fields (A), vertical velocity fields at 4972000 latitude (B), significant wave height (H_s) (C), combined current and wave bed shear stress (D).

Figure 11. Conceptual hydro-morphodynamic model for the river-mouth area during extreme conditions: floods (A), storms (B).





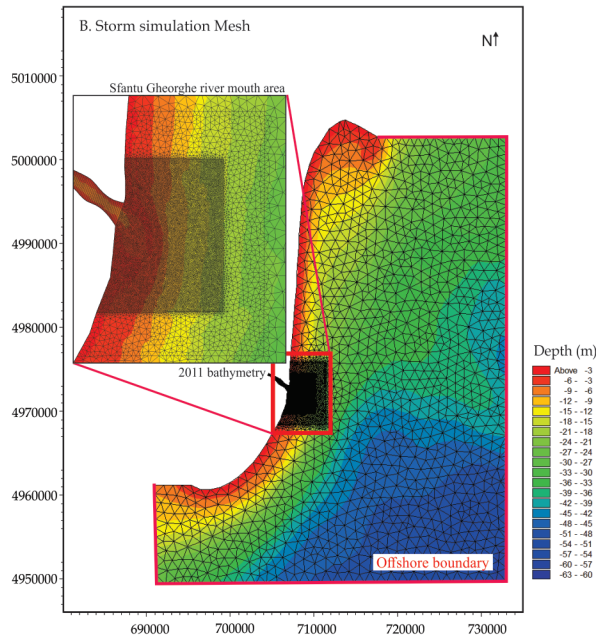
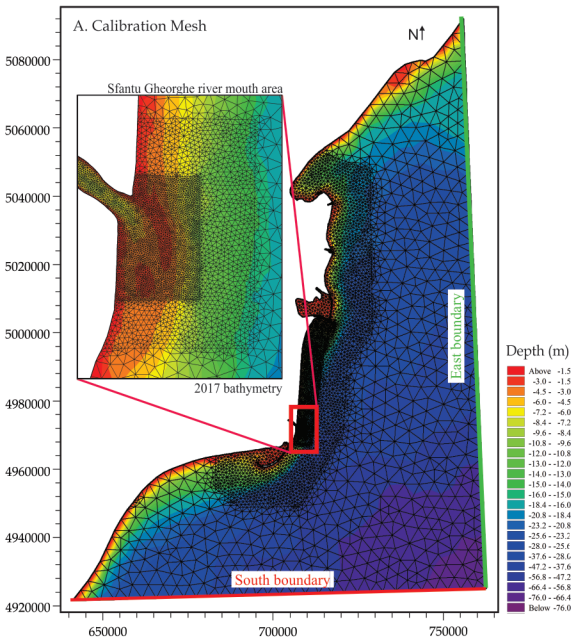
Q (m³/s)													Hours/yr
750	35	381	219	85	42	23	10	4	0.9	2	1		
1250	161	1392	733	282	130	70	37	15	12	2			
1750	109	1026	566	266	109	45	17	11	6	1	0.5	0.5	1
2250	70	602	378	184	81	32	14	5	0.5				
2750	58	413	188	68	16	4	2	0.5					
3250	27	245	121	42	22	12	5	2					
3750	29	204	57	16	5	3	0.5	1					
>4000	13	29	11	7	0.9								

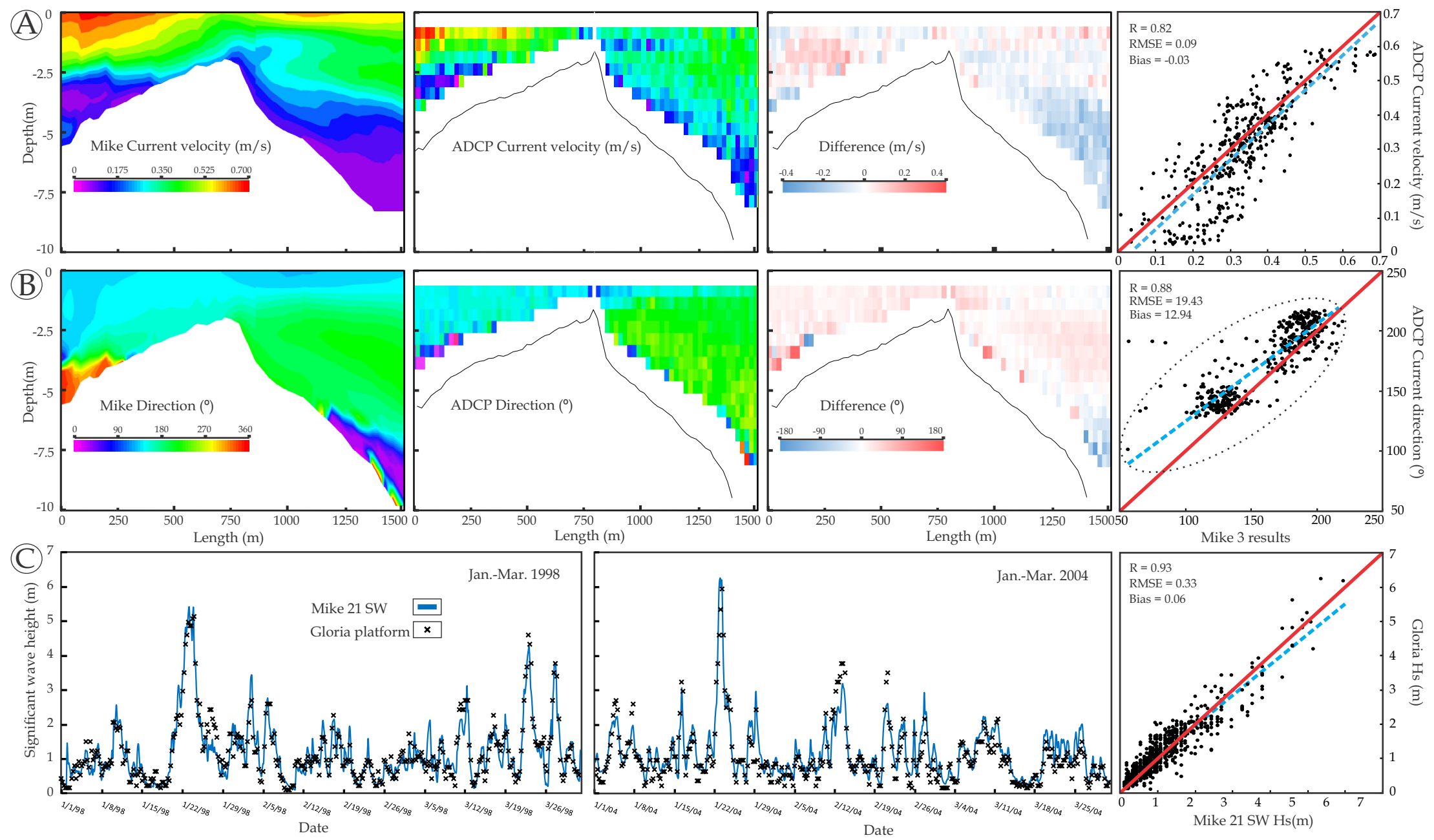
Hs (m) Calm 0.5 1 1.5 2 2.5 3 3.5 4 4.5 5 5.5 ≥6

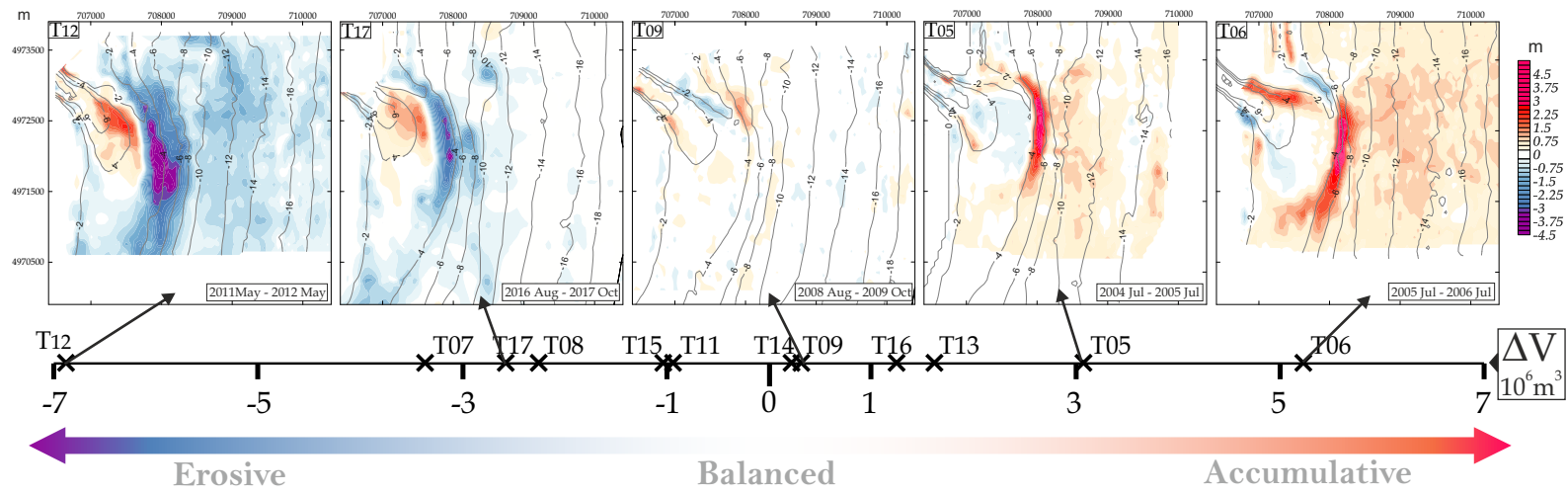
Dir (°)	0													Hours/yr
22.5	43	412	222	116	59	25	16	5	1	2				
45	41	429	228	111	72	41	19	7	5					
67.5	48	390	226	121	54	25	17	12	8	3	1	0.5	1	
90	31	332	153	66	26	9	4	5	2					
112.5	37	362	134	46	22	10	1	1						
135	29	270	97	19	4	3								
157.5	30	192	92	22	7	2								
180	31	250	151	42	11	5	2							
202.5	30	351	202	78	24	12	6	1	0.5					

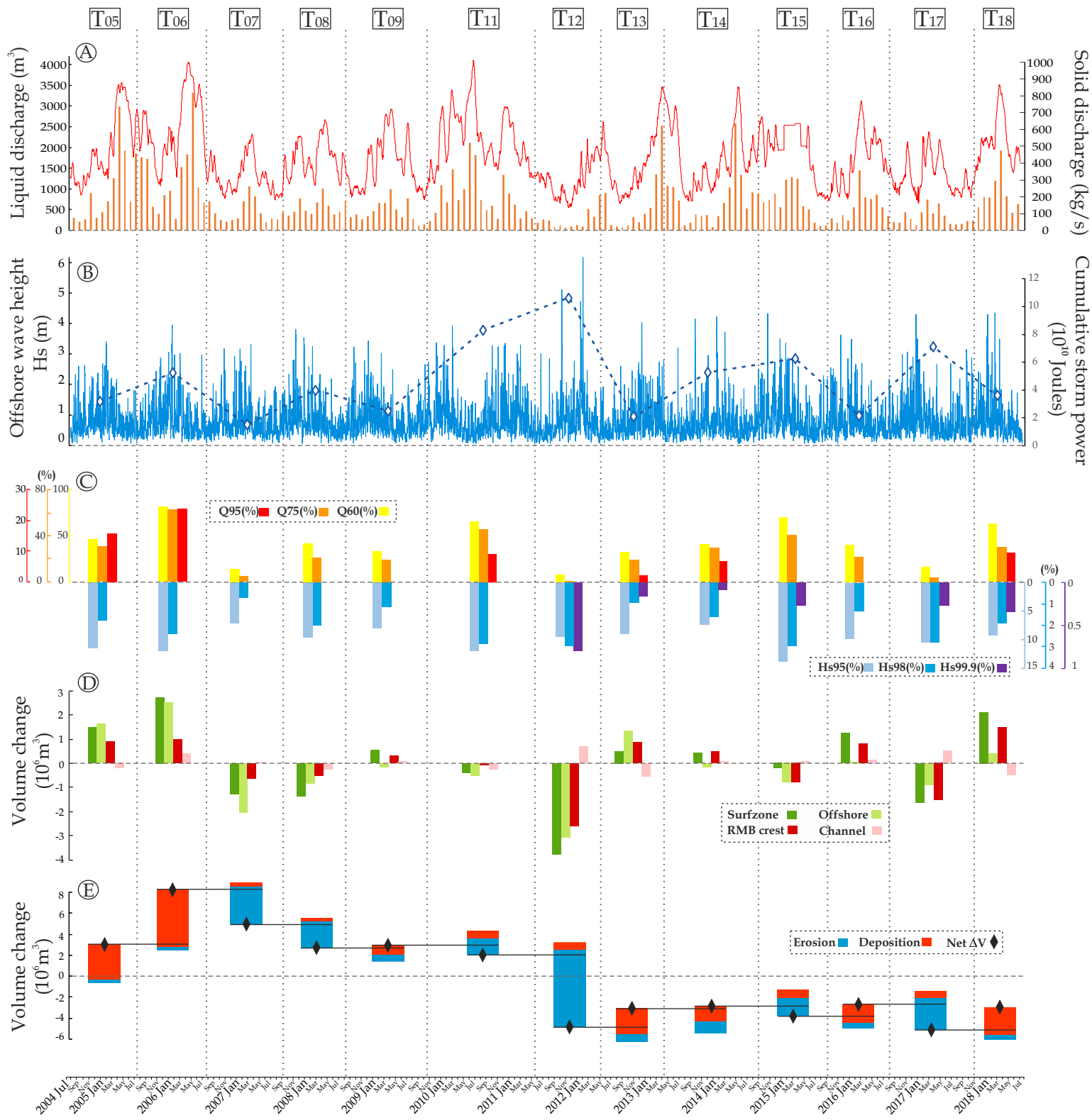
N vs S 1.055 1.067 1.29 2.091 3.373 3.662 5.955 11.8 35

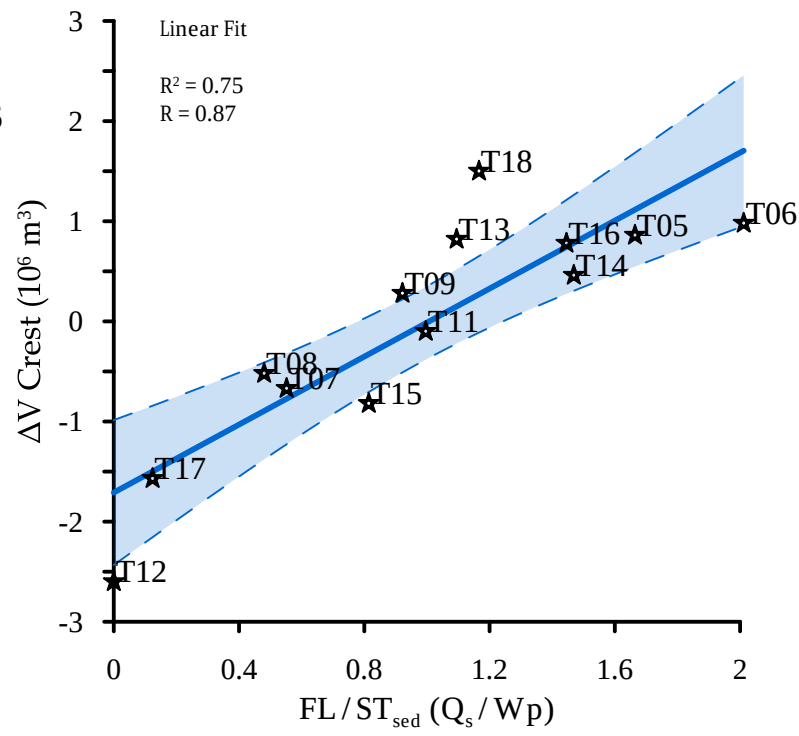
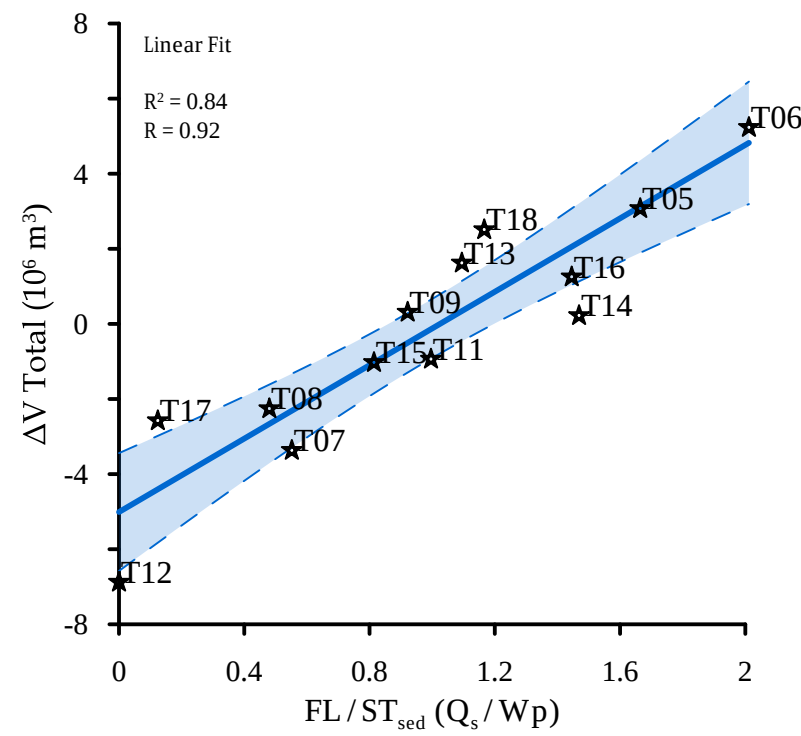
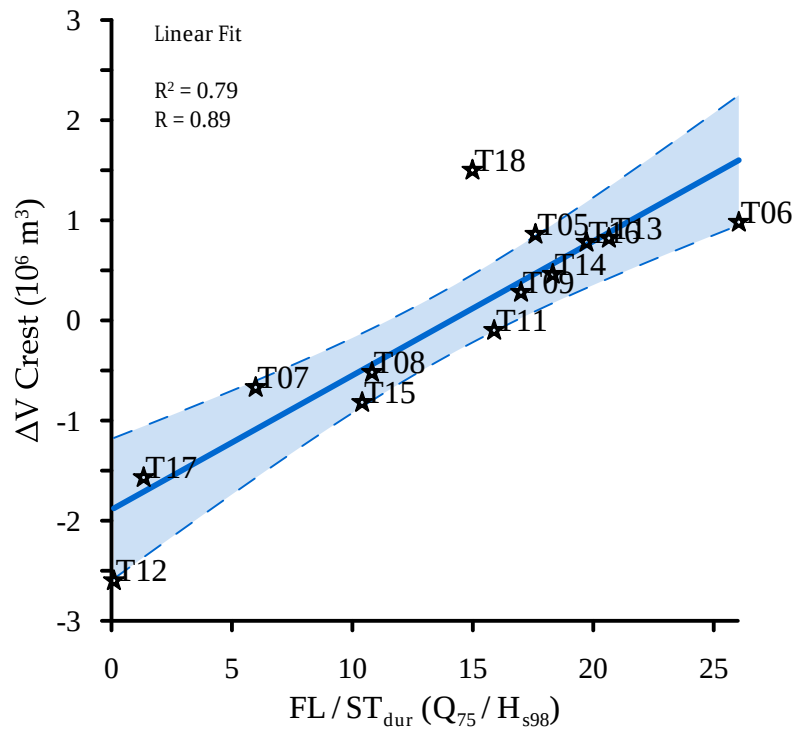
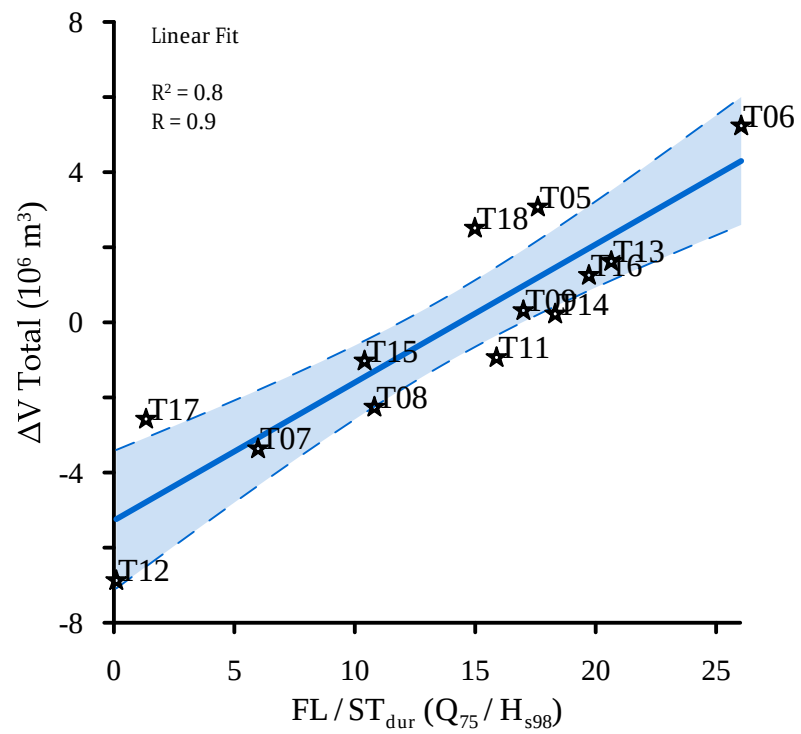
Hs (m) Calm 0.5 1 1.5 2 2.5 3 3.5 4 4.5 5 5.5 ≥6

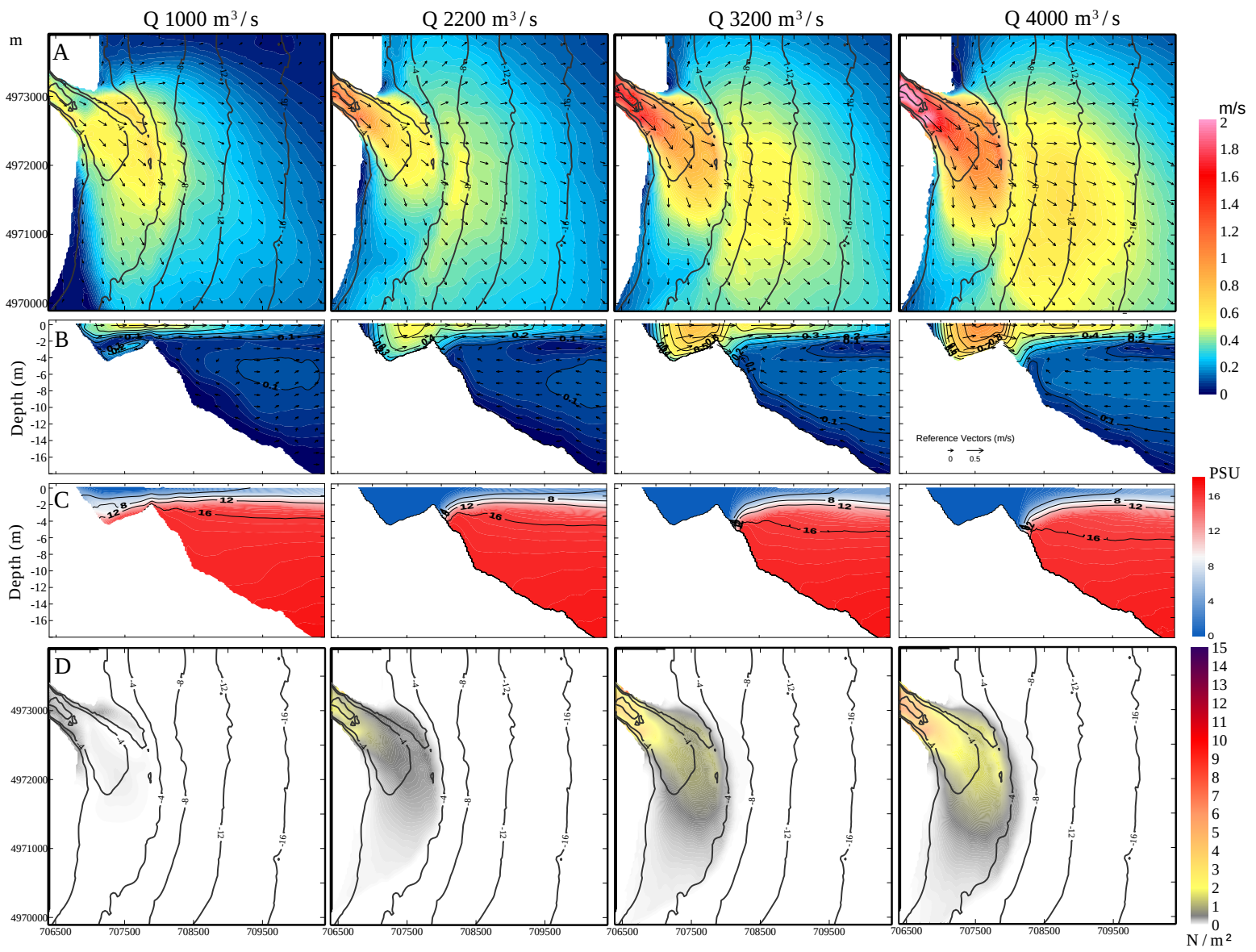


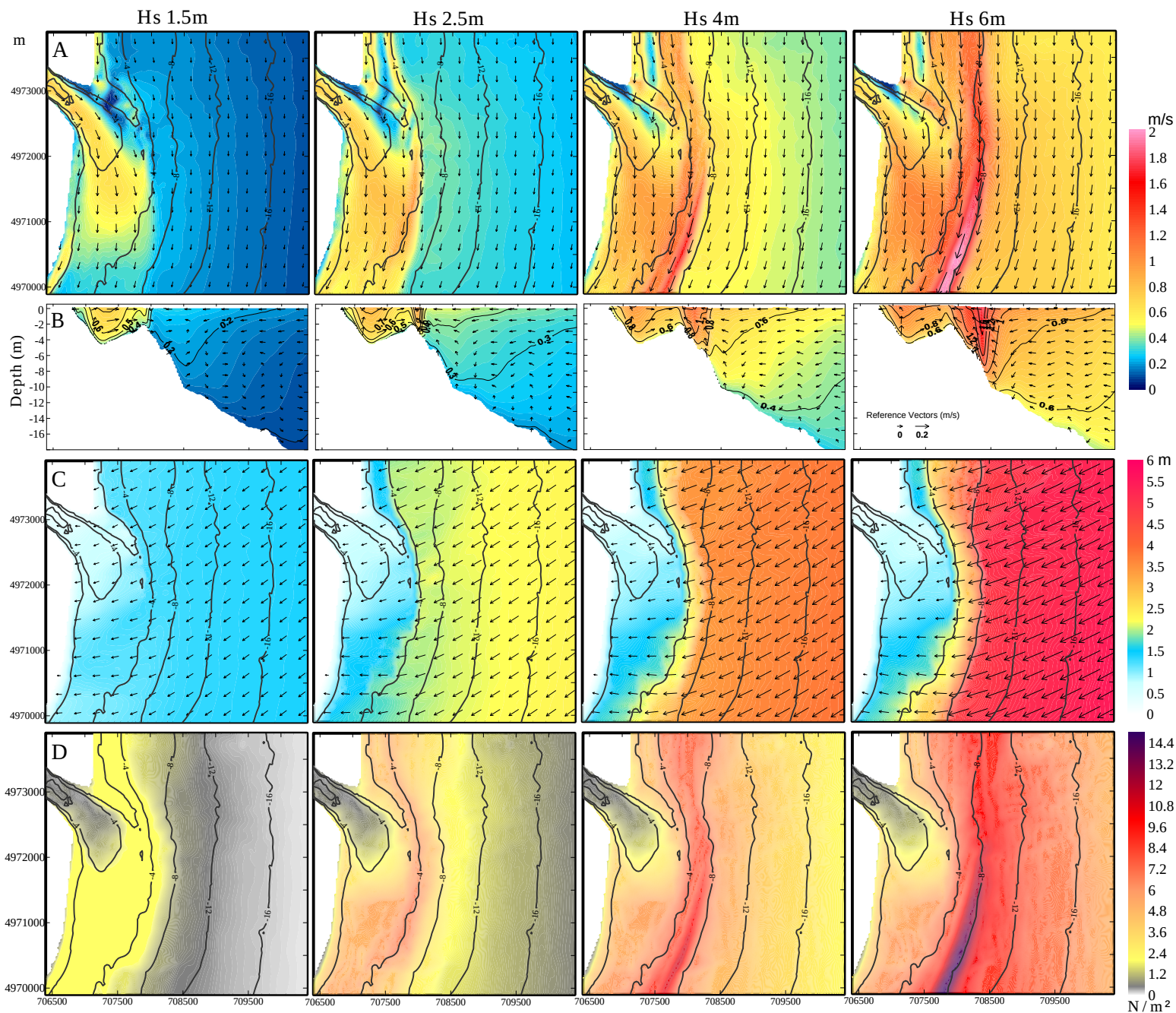




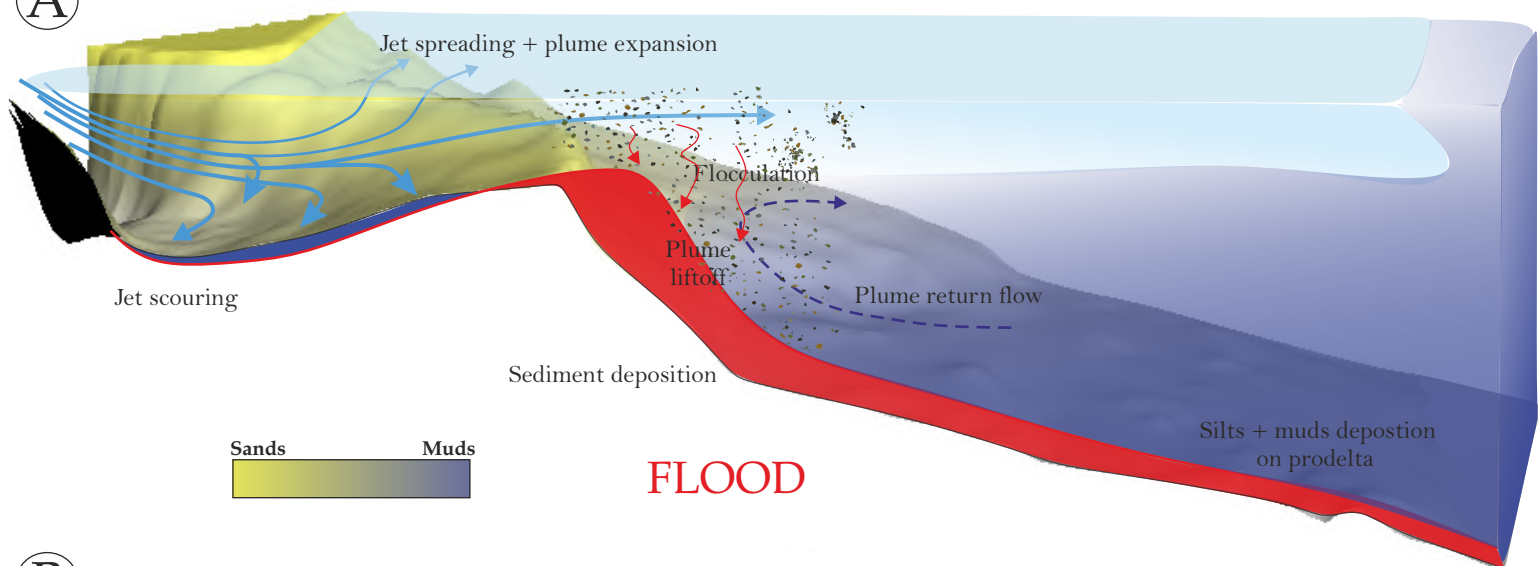




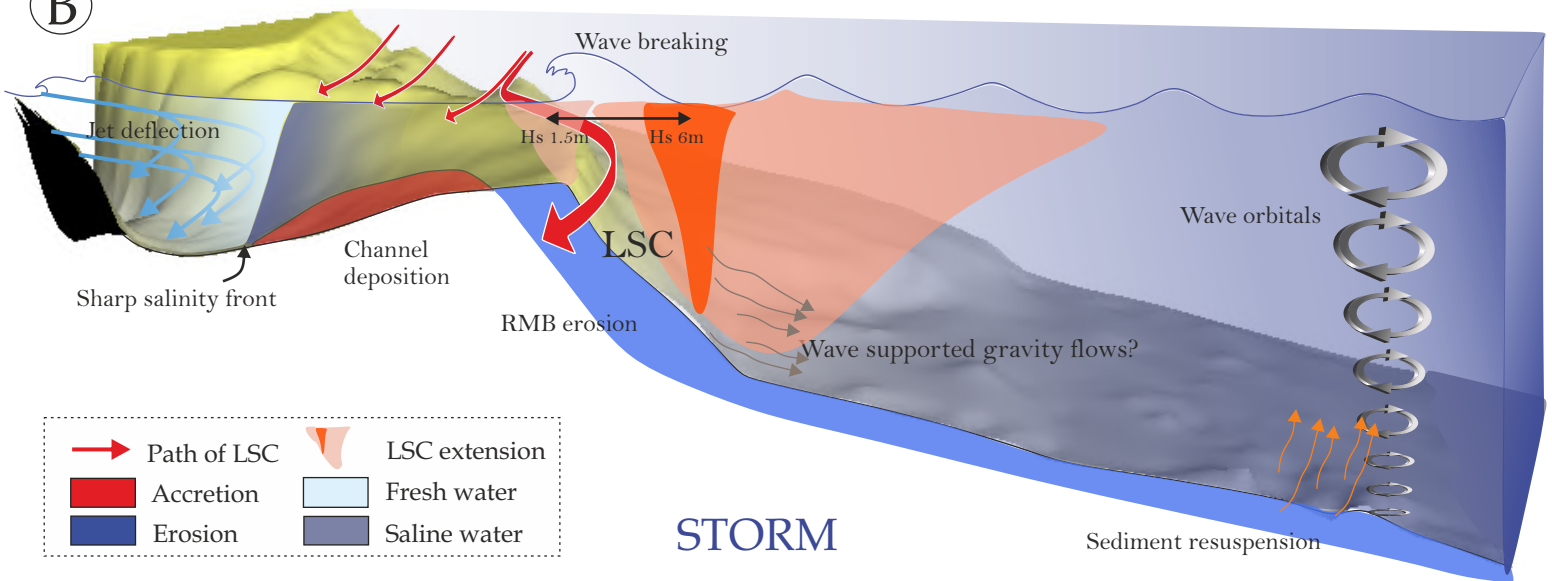




A



B



Discharge thresholds (m ³ /s)		Wave height thresholds (m)	
Q ₁₅	1000	Hs ₉₀	1.5
Q ₇₅	2200	Hs ₉₈	2.5
Q ₉₅	3200	Hs _{99,9}	4
Q _{99,7}	4000	Hs _{99,99}	6

FLOOD

VS

STORM

