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(Article begins on next page)

Hydrodynamics at a Microtidal Inlet: Analysis of Propagation of the Main Wave Components

Lorenzo Melito^a, Matteo Postacchini^a, Alex Sheremet^b, Joseph Calantoni^c,
Gianluca Zitti^a, Giovanna Darvini^a, Pierluigi Penna^d, Maurizio Brocchini^a

^a*Department of Ingegneria Civile, Edile e Architettura, Università Politecnica delle Marche,
Ancona, Italy*

^b*University of Florida, Gainesville, FL, USA*

^c*Marine Geosciences Division, U.S. Naval Research Laboratory, Stennis Space Center, MS,
USA*

^d*National Research Council CNR IRBIM, Ancona, Italy*

Abstract

The evolution of different wave components as they propagate within a microtidal inlet during a storm occurring from 24–26 January 2014 is analyzed, in order to improve knowledge on how microtidal river mouths typical of the Adriatic Sea behave. For the first time, the “low-pass filter” mechanism previously ascertained at several macrotidal oceanic inlets around the world has been observed in the field with remarkably specific hydrodynamic conditions, i.e. low tide excursion, permanent connection with the sea and generally milder wave climate than in the ocean. Sea/swell (SS) waves were strongly dissipated before entering the river mouth, through the combined action of wave breaking due to reducing depths and opposing river currents enhanced by rainfall. Infra-gravity (IG) waves propagated upstream and significant IG wave heights of up to 0.4 m, about 13% of the local water depth, have been observed 400 m upriver (about 10 times the local SS peak wavelength) during storm climax. The IG wave energy here represented over 4% of the maximum offshore storm energy. IG wave components travelled upriver at estimated velocities between 3.6 m/s and 5.5 m/s (comparable with speeds of nonlinear long waves) during intense storm stages up to 600 m into the river channel (about 15 times the local SS

Email address: l.melito@pm.univpm.it (Lorenzo Melito)

peak wavelength), and are enhanced by tide-induced increase in water depths. It is estimated that tide-induced excursion accounted for about 80% of the total mean water elevation at storm peak at about 400 m into the river. Finally, tidal oscillations are detected up to 1.5 km upstream (about 40 times the local SS peak wavelength). This study highlights the dominance of astronomical tide over both wave setup and storm surge in controlling the upriver propagation of IG waves, even in a microtidal environment.

Keywords: river mouth, estuary, microtidal estuary, wave-current interaction, infragravity waves, storms, calm-storm transition

1. Introduction

River mouths are strongly dynamic environments, where a great number of hydrological and morphological processes take place simultaneously and influence each other. The complex nonlinear interactions among bathymetry, waves and currents play a key role in defining the hydro-morphodynamic behaviour of coastal settings like estuaries, deltas and inlets [1, 2, 3, 4]. However, the relevant features of these nonlinear interactions are still ambiguous and need more research to be fully understood.

Much attention over the last decades has been devoted to a better comprehension of infragravity (IG) waves [5]. IG waves are defined as low-frequency ocean surface oscillations with period ranging from 25 seconds to five minutes [6, 7], generated either as group bound long waves [8, 9] or by a temporal variation of the breakpoint [10]. Further, IG waves may also generate from the nonlinear interaction between swell components, then propagate in the surf zone and be refractively trapped on a mildly sloping beach [11]. Such low-frequency oscillations have an important influence on the surf zone hydrodynamics (e.g., see [12, 13, 14, 15, 16, 17, 18, 19] and cited literature therein) and on swash properties, particularly on dissipative beaches [12, 19, 20, 21, 22]. The influence of IG motions in sediment transport processes in the nearshore has been also widely investigated (e.g., see [23, 24, 25, 26, 27] and references therein).

21 Notwithstanding such interest, the study of IG motions at microtidal estuaries
22 has received comparatively less attention than for other coastal features and
23 structures such as tide-dominated inlets [28, 29], wave-dominated inlets [30],
24 reefs [16, 31], coastal dunes [32] and harbours [33, 34, 35, 36].

25 Measurements within the small estuary of the Ría de Santiuste (Spain) [37]
26 have revealed a consistent upriver propagation of waves in the IG band (4.3–4.8
27 minutes) and current velocity amplitudes of up to 0.1 m/s despite the strong
28 discharging flow. Such waves and related currents were amplified as they pro-
29 gressed toward the inner parts of the river, probably due to the effect of edge
30 waves entering the estuary and producing resonance. Field measurement cam-
31 paigns conducted at the shallow, wave-dominated Albufeira Lagoon inlet (Por-
32 tugal) [2, 30] have also shown the presence of long period oscillations in water
33 levels in the back-barrier lagoon, due to IG waves developing in the region of the
34 ebb delta. IG wave propagation appears to be tide-modulated; IG waves are the
35 most intense during the flood phase and are blocked during the ebb phase. An
36 extensive analysis of field measurements at the shallow inlet of the Pescadero
37 river, North California [28] have similarly revealed large IG-related oscillations
38 in flow velocity, often of the same order of magnitude of tide-induced currents,
39 as well as upstream-propagating low frequency waves within the estuary. IG
40 velocities appear to have maximum amplitude with high flood velocities, while
41 they are damped during ebb.

42 Here we present observations of the hydrodynamic phenomena occurring at
43 the microtidal estuary of the Misa River (Senigallia, Italy) in two successive
44 storm events during an experimental campaign performed in January 2014 [38].
45 A combined analysis of offshore and riverine forcing actions is performed to clar-
46 ify the processes that either generate or evolve before, during and after storm
47 events, e.g. the evolution, dissipation and relevance of the main wave modes
48 (sea/swell (SS) waves, IG waves, tides) motion and dissipation within the river,
49 and the interactions between sea and river forcings. Additional information
50 from a recently installed hydrometer complement the study by giving insight
51 into the penetration of tidal forcing into the river. The regional setting, the



Figure 1: Study site (images adapted from Google Earth). **a)** Map of Italy and locations of Ancona and Senigallia. **b)** Locations of deployed instrumentation during the field sampling campaign in January 2014. **c)** Locations of hydrometers measuring in February-March 2018.

experimental setup deployed during the original field campaign and the analytical tools adopted in the present study are briefly described in Section 2. Section 3 presents observations on the evolution of wave components and energetic content from offshore to into the river mouth, along with considerations on the interaction between sea and river forcings in different space and time scales. The main findings are discussed in Section 4. Concluding remarks are presented in Section 5.

2. Materials and Methods

2.1. Regional setting and instrumentation

Two field campaigns have been carried out in September 2013, during summertime conditions, and January 2014, in a wintertime regime, in the area around the microtidal estuary (mean tidal range less than 0.6 m) of the Misa River (MR hereinafter), a natural stream located in central Italy and flowing

65 into the Adriatic Sea. The final reach of the river flows within a fixed engineered
66 channel through the town of Senigallia (Marche Region), one of the most im-
67 portant touristic towns of the Italian Middle Adriatic coast (Figure 1a). At the
68 time of the measurements (2013–2014) the river was regularly dredged.

69 The MR has a total watershed of about 383 km^2 and a river discharge of
70 about $400 \text{ m}^3/\text{s}$ for a return period of 100 years. Although showing a low flow
71 regime during summer [39], in wintertime the MR features a strongly increased
72 surface flow that dominates over the action of waves and tides.

73 The MR estuary is classified as a salt-wedge estuary, where freshwater flows
74 out to sea in the upper layer of the water column, while seawater intrusion oc-
75 curs in the lower layer [40]. Salt water intrusion is confirmed by salinity values
76 larger than 10 psu 700 m upstream of the river mouth, and smaller, although
77 non-zero values up to 1.5 km upstream of the mouth during summer [39]. Small
78 landward velocities in the lower part of the water column (up to 1–1.5 m above
79 the bed) during salt wedge intrusion have been also detected 500 m upstream of
80 the estuary during winter (see the top and center panels of Figure 3 in [38]). As
81 typical for rivers originating in the Apennine Mountains, the MR is also char-
82 acterized by large sediment transport rates in spite of its moderate discharge.
83 The intense sediment outflow has been postulated to have an influence on the
84 evolution of morphological features of the unprotected beaches southward of the
85 estuary [41].

86 The two field campaigns have been performed in the context of the EsCoSed
87 project, which aimed at characterizing estuarine morphodynamics in summer
88 and winter conditions as well as their differences [38]. To this purpose, specific
89 equipment was deployed to measure hydrodynamics features in the lower part
90 of the MR and the close nearshore area. The present work largely focuses
91 on the observations from equipment deployed over the wintertime campaign of
92 January 2014. Two quadpods were deployed in the sea at locations QS2 and
93 QS1, at water depths of around 6.5 m and 5.3 m, respectively (Figure 1b). Other
94 quadpods were placed at various locations inside the final reach of the MR at
95 locations QR1, QR2, and QR3 (at 530 m, 400 m and 290 m upstream from the

mouth, and located at mean water depths of 2.7, 3.0 and 2.9 m, respectively). A complete description of the instrumental setup can be found in [38]. Two additional tide gauges, TGdown and TGup, were also located in the river at about 280 m and 580 m from the MR mouth, respectively. The locations of the equipment involved in the present study are illustrated in Figure 1b.

To observe the propagation of the sea forcing actions within the MR, additional gauge measurements have been included in this study. Data collected by the river hydrometers installed at “Ponte Garibaldi” (~ 1.5 km from the MR estuary, RG1 hereafter) and “Bettollele” (~ 10 km from the MR estuary, RG2 hereafter; Figure 1c) have been analyzed. The “Ponte Garibaldi” river gauge has been installed in 2016 by the Civil Protection of the Marche Region at a distance of around 1.5 km from the mouth, just upstream of Ponte Garibaldi, a 40 m-long bridge located in the city center. It has been placed in an area along the ending reach of the river where sea intrusion is still detectable [39]. Data from RG1 and RG2 are used in Section 3.6 to discuss some significant flood events in the first few months of 2018, occurred with contextual wave and wind states similar to those observed in January 2014.

The river gauges have a resolution of one datum every 30 min, much lower than that of the majority of the EsCoSed instrumentation (2 Hz). Although SS waves cannot be captured by river gauges, the tidal influence can be easily observed at these locations. Comparisons are shown in Section 3.6 between RG signals and the signal of the tide gauge of the Ancona harbour, located about 30 km south of Senigallia (Figure 1a). The Ancona harbour tide gauge records water level data at a frequency of 0.1 Hz. Preliminary observations on tidal propagation into the MR estuary have been previously presented in [42].

2.2. Methods

In order to evaluate the wave state energetic content in the offshore and into the river estuary, a spectral analysis has been performed. Starting points for the analysis are the water elevation time signals detected by the respective pressure sensors: quadpod QS2 for the analysis of offshore sea state (Section 3.1), and

126 river gauge QR2 for the analysis of wave propagation into the estuary (Sec-
 127 tion 3.2). For each recording hour, the corresponding wave spectral energy den-
 128 sity \mathcal{E} has been determined through a discrete Fourier transform of the hourly
 129 de-tided water elevation signals over a frequency domain 0–1.0 Hz. All spectra
 130 have been obtained with a number of DOF equal to 41. Once \mathcal{E} is obtained, the
 131 energy flux density \mathcal{F} at that hour is $\mathcal{F} = \mathcal{E}c_g$, where c_g is the group velocity.
 132 An evaluation of the spectral centroidal frequency f_c of each hourly spectrum
 133 is moreover performed as the weighted mean of the frequency range, with the
 134 flux density as the weights.

135 To study the temporal evolution of energy fluxes \mathcal{F} at a specific location,
 136 the entire frequency range is divided into three bands representing domains
 137 characterized by different physics: the IG band (0.0016–0.05 Hz), the SS band
 138 (0.05–0.3 Hz) and the wind waves band (0.3–1 Hz). The band-specific flux Φ_{band}
 139 is defined by integrating the flux density \mathcal{F} over each band:

$$\Phi_{\text{band}} = \int_{f_{\text{start}}}^{f_{\text{end}}} \mathcal{F}(f) \, df, \quad (1)$$

140 where f_{start} and f_{end} are the start and end frequencies of each band. Ad-
 141 ditionally, band-specific significant wave heights H_{sig} have been calculated as
 142 $H_{\text{sig}} = 4\sqrt{m_0}$, where m_0 is the 0-th moment of the spectral energy density \mathcal{E} in
 143 the specific band.

144 Section 3.5 gives an analysis of wave energy propagation from the offshore
 145 to within the river mouth. To this purpose the wave energy (total, of SS waves
 146 or of IG waves) is evaluated as a function of the significant wave height at a
 147 specific sensor through the classical relation valid for random gravity waves:

$$E = \frac{\rho g H_{\text{sig}}^2}{16}. \quad (2)$$

148 Wave energy is also used to evaluate the cross-shore radiation stress, main

149 driver of the wave-induced setup [9, 43]:

$$S_{xx} = E \left(2n - \frac{1}{2} \right), \quad (3)$$

150 where $n = c_g/c$ is the ratio of group velocity to wave celerity. In the assumption
 151 of approaching shallow waters c tends to \sqrt{gh} , where h is the water depth. For
 152 a given hour, group celerity c_g is evaluated as the mean of the group celerities
 153 $c_g(f)$, with $f \in (0-1.0]$ Hz.

154 Also in Section 3.5 an evaluation of upriver propagation velocities of IG wave
 155 components is attempted. In order to remove the sea and swell wave components
 156 from signals, a low-pass filtering (with passband frequency of 0.05 Hz) of the
 157 de-tided signals of river sensors has been performed. A cross-correlation of
 158 low-frequency signals between consecutive gauges was performed to furnish an
 159 estimate for the time t_{lag} it takes for IG wave components to travel from a
 160 gauge to another, and therefore give a proxy for the velocity of propagation of
 161 IG oscillations:

$$c_{\text{IG}} \approx \frac{\Delta x}{t_{\text{lag}}} \quad (4)$$

162 where Δx is the distance between two consecutive gauges inside the river. We
 163 here stress the fact that, by using Equation (4), we assume that only IG progres-
 164 sive waves are taken into consideration and possible reflected and/or standing
 165 waves developing into the channel are therefore not considered.

166 An IG wave attenuation rate R_{IG} is also used in Section 3.5 to better appraise
 167 the decay of low frequency components across the lower reach of the MR:

$$R_{\text{IG}} = \frac{H_{\text{sig,IG,TGdown}} - H_{\text{sig,IG,TGup}}}{H_{\text{sig,IG,TGdown}}} \cdot 100, \quad (5)$$

168 where $H_{\text{sig,IG,TGdown}}$ and $H_{\text{sig,IG,TGup}}$ are the significant heights of IG compo-
 169 nents at sensors TGdown and TGup, respectively.

170 Finally, in Section 3.6 comparisons between signals from river hydrometers
 171 RG1 and RG2 and the Ancona harbour tide gauge TG are proposed to evaluate

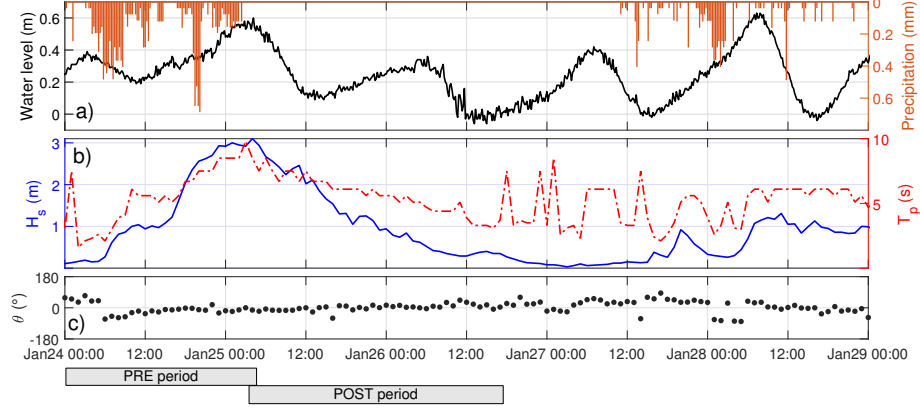


Figure 2: Wave climate in January 24–29. **a)** Water elevation as measured at the Ancona harbour (black line) and precipitation (orange bars). The datum is the mean sea level. **b)** Estimated significant wave height (blue line) and peak period (dashed red line) offshore of sensor QS2. **c)** Predominant direction of incoming waves with respect to the shore normal. The ranges of PRE and POST periods (Sections 3.1 and 3.2) are also labelled with the grey boxes.

the influence of tidal forcing over river discharge and water levels. The difference between signals at river hydrometers is made:

$$\eta_{\text{diff}} = \eta_{\text{RG1}} - \eta_{\text{RG2}}, \quad (6)$$

and discrepancies between the tide gauge signal η_{TG} and η_{diff} are evaluated by means of a cumulative parameter defined as follows:

$$TGRG_{\text{cum}} = \frac{1}{T_{\text{rec}}} \sum (\eta_{\text{TG}} - \eta_{\text{diff}}) \delta t, \quad (7)$$

where T_{rec} is the recording period over which the calculation is made and δt is the time discretization interval.

3. Results

3.1. Analysis of offshore sea state (sensor QS2)

The characterization of the sea state in front of the MR estuary is made by elaborating wave data measured seaward of and at sensor QS2, located approximately 650 m offshore of the river mouth. Figure 2 shows the offshore wave

183 climate in front of the MR mouth as measured by an ADCP deployed seaward
184 of QS2 between January 24 and January 29, 2014, when the wintertime section
185 of the EsCoSed project took place [38]. The elevation datum for the water level
186 in Figure 2 is the mean sea level. Precipitation data are collected from the
187 database of the Civil Protection.¹

188 The event occurring from January 24 to January 26, 2014 was a severe storm,
189 characterized by offshore significant wave heights of up to 3 m during the first
190 hours of January 25 (Figure 2a). The storm was sustained by strong northerly
191 (Bora) winds, whose relatively short fetch in the Adriatic sea contributed to the
192 generation of short and steep swells propagating almost perpendicularly to the
193 shoreline (Figure 2c), also thanks to the relatively deep and engineered river
194 mouth. Such swells were capable of penetrating the river mouth at least up
195 to sensor QR2, where consistent wave oscillations were observed (Section 3.2).
196 The storm was also accompanied by intense precipitation in terms of both rain
197 rate and total daily rain [38].

198 The storm duration has been divided into two segments with a 1-hour over-
199 lapping. The PRE period, encompassing the storm growing up to its climax,
200 spans from January 24, 00:00 to January 25, 05:00 (29 hours). The POST pe-
201 riod, that includes the storm decay, runs from January 25, 04:00 to January 26,
202 18:00 (38 hours).

203 Figure 3 presents energy flux densities and evolution of band fluxes at off-
204 shore sensor QS2. The left panels show data from the PRE period, whereas
205 right panels show data from the POST period.

206 In the PRE period, the offshore wave energy increased by nearly four or-
207 ders of magnitude in the range 0–0.3 Hz throughout the investigated period
208 (Figure 3a), the peak frequency being in the range of wind wave frequencies,
209 0.1–0.15 Hz (wave periods of 6.5–10 s) for both the initial calm state (January
210 24, 00:00–07:00; black to light grey spectra in Figure 3a) and the fully-developed
211 storm condition (January 25, 00:00–05:00; orange and red spectra in Figure 3a).

¹<http://www.regione.marche.it/Regione-Utile/Protezione-Civile>

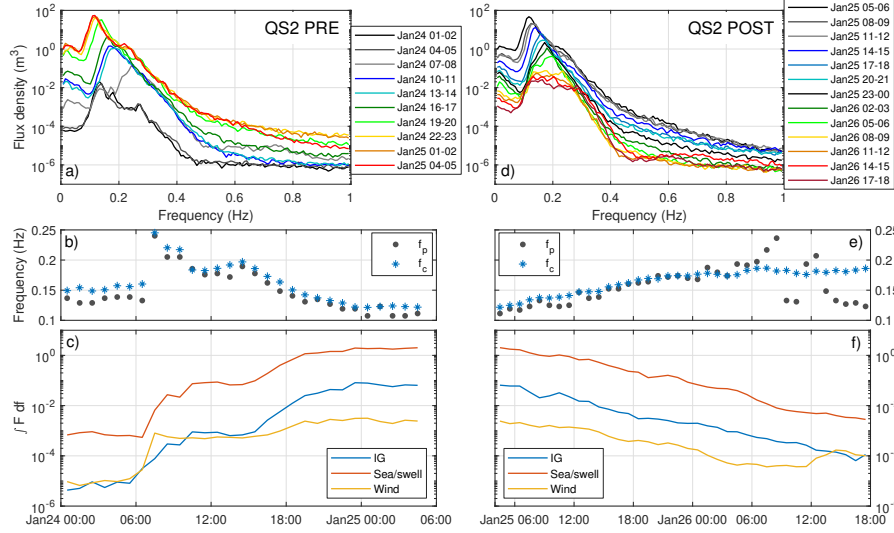


Figure 3: Evolution of energetic content at QS2 at selected hours during the PRE period (left panels) and the POST period (right panels) of the January 24–26 storm. **a) d)** Flux density spectra. **b) e)** Temporal evolution of the peak frequency f_p (grey dots) and the spectral centroidal frequency f_c (blue stars). **c) f)** Temporal evolution of integral fluxes for IG waves (blue line), SS waves (red line), and wind waves (yellow line).

At the very start of storm growth, on January 24, 07:00–08:00, the peak frequency showed an abrupt shift towards higher frequencies after January 24, 06:00 (see the light grey spectrum in Figure 3a and the peak in both f_p and centroidal frequency f_c in Figure 3b), due to short swell waves produced by sustained strong northerly winds that reached speeds up to 20 m/s (recorded by a weather station located within the harbor area) on relatively short fetches [38]. This is also confirmed by the coincident increase in integral flux for swell and wind wave frequencies (red and yellow lines in Figure 3c). As the storm increased its intensity, the bulk of storm energy transferred to lower frequencies and the peak frequency migrated back to the regular lower swell wave frequency range (0.1–0.15 Hz, see Figure 3b). The integral fluxes of IG and sea/swell waves seems to be proportional (compare blue and red lines in Figure 3c). In more detail, during the hours of storm climax, i.e. when the highest wave energy levels are attained, the IG wave energy E_{IG} increases with the sea-swell energy E_{ss} following a nonlinear trend, i.e. $E_{IG} \propto E_{ss}^{1.5}$ (Figure 4), in agreement with

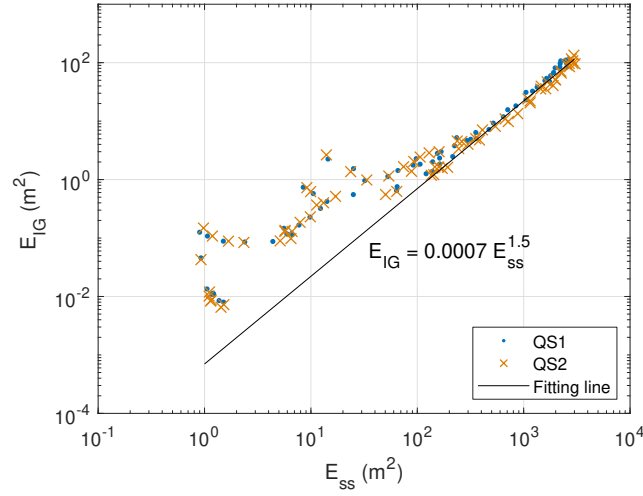


Figure 4: IG band energy E_{IG} versus sea/swell energy E_{ss} measured at QS1 (blue dots) and QS2 (orange crosses) across the investigated period. The black line represents a fitting line for the data with highest energy level.

previous literature findings [44].

Wind wave fluxes (yellow line in Figure 3c), on the other hand, quickly reached saturation after the steep increase in January 24, 06:00 and showed very little evolution afterwards, in spite of the storm growing steadily until January 25. Both IG and SS waves experienced more or less the same increase in absolute energy flux across storm growth up to the climax.

Offshore flux density spectra in the POST period are shown in Figure 3d. Also here, the peak frequency moved from the low swell frequency range (on January 25, 04:00–12:00; black to light grey spectra in Figure 3d) to higher frequencies (see Figure 3e), and the spectrum flattened until its typical peak shape is lost altogether and a clearly dominant frequency could not be singled out, although there still seems to be a dominance of low swell frequencies (on January 26, 11:00–18:00; yellow to red spectra in Figure 3d). Throughout storm decay, though, the frequency range pertaining to swell waves (0.05–0.3 Hz) lost up to four orders of energy content (Figure 3d). The same loss is retrieved in Figure 3f, where a steady decrease in energy flux is noticeable at all frequency bands and the link between IG and SS waves is again apparent.

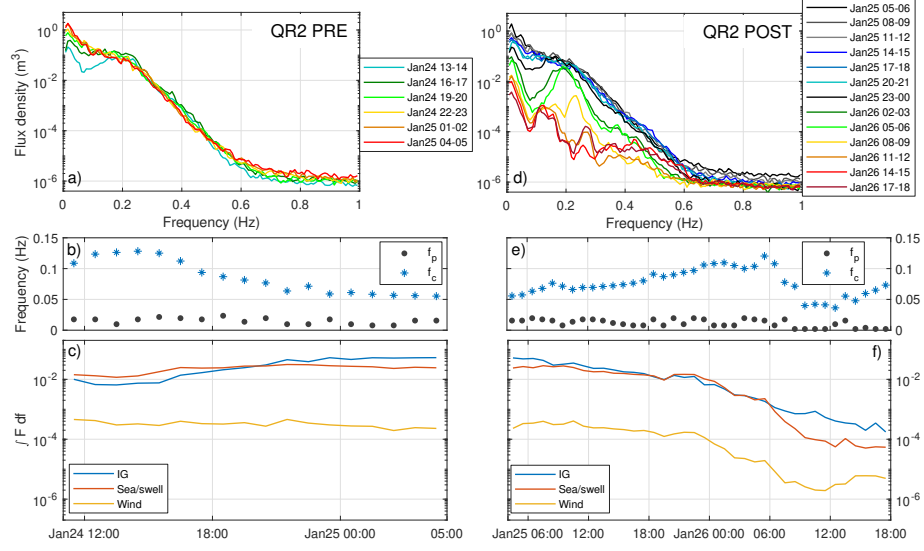


Figure 5: Evolution of energetic content at QR2 at selected hours during the PRE period (left panels) and the POST period (right panels) of the January 24–26 storm. **a) d)** Flux density spectra. **b) e)** Temporal evolution of the peak frequency f_p (grey dots) and the spectral centroidal frequency f_c (blue stars). **c) f)** Temporal evolution of integral fluxes for IG waves (blue line), SS waves (red line), and wind waves (yellow line).

3.2. Analysis of wave energy into the Misa River (sensor QR2)

The same analysis described in Section 3.1 has been also performed for sensor QR2, located within the final reach of the MR at about 400 m upstream of the mouth. As already evidenced in [38], the January 24–26 storm was energetic enough to drive waves up the channel in spite of the strong breaking outside the mouth, so a spectral analysis at this location is appropriate and gives a sense of how the wave energy content has evolved from the offshore to the river channel. The total measuring period is also here split into a PRE period and a POST period. Sensor QR2 was not in operation during the initial calm state, in January 25, 00:00–11:00.

Figure 5 presents selected flux density spectra and temporal evolution of band-specific fluxes at sensor QR2 during the PRE period (left panels) and the POST period (right panels). None of the spectral forms in the investigated period show peaks within the SS frequency range. Only the IG band (0–0.1 Hz) increased its energy flux by roughly one order of magnitude as the storm ap-

259 proached its climax; on the other hand, energy flux at wave frequencies greater
 260 than 0.15 Hz was largely unaltered (Figure 5a). The frequency peak was con-
 261 sistently inside the IG band (Figure 5b). Figure 5c also shows that the IG band
 262 had the largest increase in energy flux and eventually became the most ener-
 263 getic band after January 24, 18:00 (blue line in Figure 5c). Integral fluxes of
 264 swell and wind waves were largely unchanged, with the wind waves flux being
 265 particularly low. However, the centroidal frequency (blue stars in Figure 5b)
 266 falls in the sea-swell band at the earlier stages of the storm (this recalling the
 267 sea-swell dominance also in terms of integral fluxes, see Figure 5c) and towards
 268 the end of the storm (Figure 5e). This demonstrates the larger importance of
 269 the IG band during the core of the storm, while the sea-swell components gain
 270 more (relative) relevance into the river during the rise and fall stages of the
 271 storm, when both wave breaking and IG generation are reduced.

272 In the POST period the storm attenuation led to a general reduction of
 273 energy flux at all frequencies, although the energy loss was stronger for swell
 274 and low-frequency wind waves (0.15–0.4 Hz; see Figure 5d and yellow and red
 275 lines in Figure 5f). This is particularly evident in the spectra related to the end
 276 of the period, when a calm state characterized by a dominance of IG energy was
 277 achieved (orange to red spectra in Figure 5d). Also in the POST period the
 278 peak frequency f_p was located into the lower IG band at all times (0.01–0.03 Hz;
 279 see Figure 5e).

280 3.3. Wave setup

281 An estimate of wave-breaking-induced radiation stress in the area surround-
 282 ing the MR estuary during the January 24–26 storm as been performed. Figure 6
 283 gives space and time maps for the cross-shore radiation stress S_{xx} (left panel)
 284 and its cross-shore gradient dS_{xx}/dx (right panel) along an ideal transect from
 285 the offshore into the river channel. The maps have been obtained by estimating
 286 and interpolating hourly values of radiation stress from both sea sensors (QS1
 287 and QS2) and all river sensors (TGdown, QR1, QR2, QR3, and TGup).

288 The highest values of both radiation stress ($S_{xx} \approx 2500 \text{ J/m}^2$) and its gra-

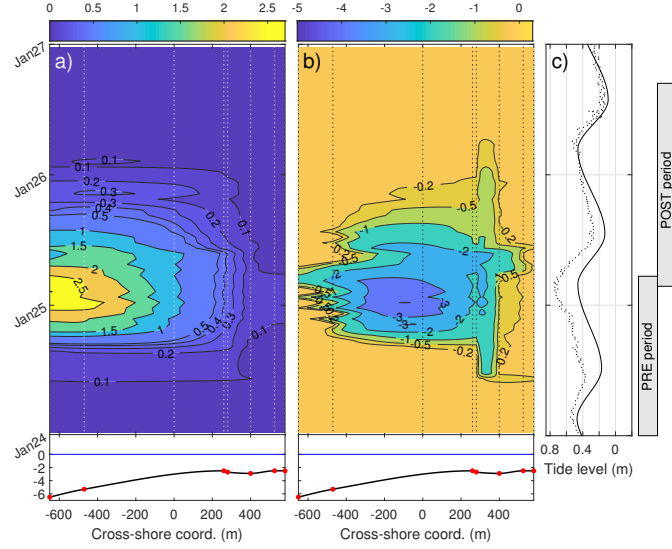


Figure 6: **a)** Cross-shore radiation stress S_{xx} , **b)** cross-shore gradient of S_{xx} , **c)** astronomical tide at Ancona (black line) and tidal level as measured at Ancona harbour (dotted line). Bottom panels represent the bottom profile. Dotted white lines in panels a) and b) and red dots in the bottom panels identify the position of sensors and the river mouth ($x = 0$ m). Positive x -coordinates are directed into the river. The ranges of PRE and POST periods are labelled with grey boxes.

289 dient ($dS_{xx}/dx \approx -5 \text{ J/m}^3$) have been achieved at the hours of storm peak
 290 (January 25, 00:00–06:00). Strong alongshore gradients of S_{xx} , corresponding to
 291 wave-breaking-induced setups, occurred landward of sea sensor QS1 (470 m off-
 292 shore of the mouth), where SS waves broke as an effect of reducing water depths.
 293 River current seems not to be strong enough to sufficiently enhance wave steep-
 294 ening and force waves to break. Specifically, during the Bora storm, the river
 295 current speed at the surface $U_{\text{river,surf}}$ never exceeds 1.0 m/s, and the maximum
 296 wave-induced horizontal velocity $U_{\text{wave,max}}$, estimated using linear wave theory,
 297 is comparable to such value. This leads to a ratio $U_{\text{river,surf}}/U_{\text{wave,max}} \leq 1.2$.
 298 On the other hand, based on simulations [3], much larger ratios are required
 299 ($U_{\text{river,surf}}/U_{\text{wave,max}} > 2.2$) to reach the condition of river-current-induced wave
 300 breaking. The absence of river current-induced wave breaking was also con-
 301 firmed by visual observation.

302 The highest potential for wave-induced setup has been estimated right in

front of the river mouth, up to 200 m offshore of it (second panel of Figure 6). Relevant setup gradients is also identified well into the river channel, at least up to sensor QR2, 400 m upriver from the mouth. Notably, a persistence of moderate radiation stress is observed in the region outside the mouth also in January 25, 12:00–18:00, simultaneously with low tide. Other than this, the patterns of radiation stress and its gradients seem not to be tide-modulated. An estimation of wave setup across a distance of $\mathcal{O}(100\text{ m})$, a typical length scale for local wave evolution, gives that maximum wave setups of around 0.05 m are expected to happen. Such values are small if compared to the local tide excursion (a few tens of centimeters) and the storm surge elevation, thus the relative importance of wave setup in raising mean water levels outside and into the estuary is expected to be low.

3.4. Interplay of sea and river currents

Figure 7 shows a comparison of depth-averaged current velocities and directions at river sensors QR1, QR2 and QR3 and sea sensor QS2 in the wintertime campaign (January 24–29), along with data from tides, precipitation and wave climate in the same period. See Figure 1 for the location of sensors.

At the most inland sensor QR1 (cyan diamonds in Figure 7c and d) seaward velocities reached up to 0.2 m/s until January 24, 10:00. River currents in this phase were strengthened by precipitation prior to the analyzed period. A slight upriver inversion of the current at QR1 is noticeable at the beginning of January 24, in association with a small flood tide and weak sea forcing.

A strong dominance of river flow was observed at QR2 (red squares in Figure 7c and d) with current velocities of more than 0.4 m/s throughout both storm build-up and decay. The peak velocity occurred around January 25, 06:00, about 11–12 hours after the strongest rainfall in January 24, 18:00–20:00. **Despite a reduced influence of the wind stress on the river current (max $\sim 20 - 25\%$ of the total speed), the river outflow was strong to the point of generating a plume that propagated up to 1.3 km off the mouth [38], while halving the intensity and altering the prevalent direction of offshore currents measured at QS2 (pur-**

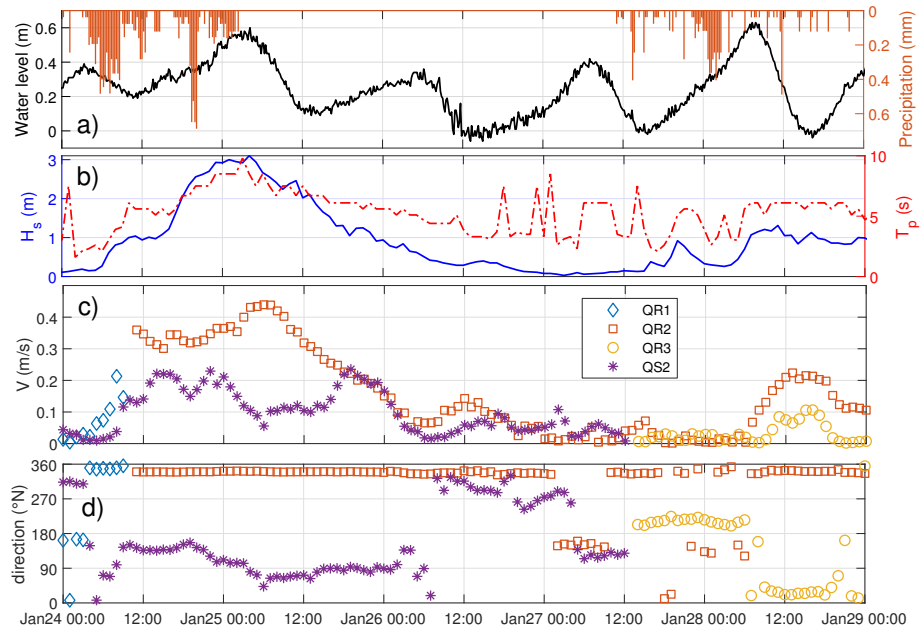


Figure 7: Interplay of river and sea forcings in the MR estuary. **a)** Tide at the Ancona harbour (black line) and precipitation (orange bars). The datum is the mean sea level. **b)** Offshore wave significant heights (blue line) and peak period (red dash-dotted line). **c)** Depth-averaged current velocities at river sensors QR1, QR2 and QR3 (cyan diamonds, red squares and yellow circles, respectively), and sea sensor QS2 (purple asterisks). **d)** Depth-averaged current directions at the same sensors.

333 ple stars in Figure 7c and d) during and immediately after storm peak. During
 334 storm decay sea and river currents were comparable, but the outflow still im-
 335 parted a westward (seaward) orientation to the offshore current detected at QS2.
 336 The mean intensity and orientation of the river current at QR2 appeared not
 337 to be influenced by the alternating tide, although flow reversal was detected in
 338 January 27, 02:00–12:00, again in coincidence with flood tide and absent storm
 339 action. The river current, at least landward of sensor QR2, was strong enough
 340 to not be altered by tide- or wave-induced oscillations entering the channel, as
 341 already postulated in [38].

342 Only measurements from river sensors QR2 (red squares) and QR3 (yellow
 343 circles) are available in the period of the second, minor storm (January 27,
 344 12:00 and afterwards). Prior to the development of the second storm the small
 345 currents detected at sensors QR2 and QR3 were oppositely oriented; mainly
 346 northerly oriented at QR2, and southerly oriented at QR3. This may suggest
 347 the presence of a convergence region between QR2 and QR3 in case of low river
 348 outflow, as evidenced also during the summertime campaign [38]. Afterwards,
 349 river currents at both sensors were directed seaward with maximum values in
 350 average flow velocity (0.2 m/s at QR2 and 0.1 m/s at QR3). No tide-induced
 351 oscillations in flow velocity or orientation were visible.

352 3.5. Propagation of low frequency waves in the lower reach of the Misa River

353 The temporal evolution for the total significant wave height H_{sig} , the IG
 354 contribution $H_{\text{sig,IG}}$ and SS contribution $H_{\text{sig,ss}}$ at sensors QS2 (offshore) and
 355 QR2 (into the river) for the January 24–26 storm is illustrated in Figure 8.
 356 While the most of the significant wave height at the offshore sensor QS2 was to
 357 be ascribed to the action of sea and swell waves, the contribution from the IG
 358 band $H_{\text{sig,IG}}$ was up to 0.4 m at the peak of the storm and made for about 10–
 359 20% of the total height (top panel in Figure 8). Similar values of $H_{\text{sig,IG}}$ about
 360 0.4 m (corresponding to about 13% of the local water depth) have been observed
 361 at sensor QR2, 400 m from the mouth. This distance corresponds to about 10
 362 times the local SS peak wavelength. In addition, the relative magnitude of

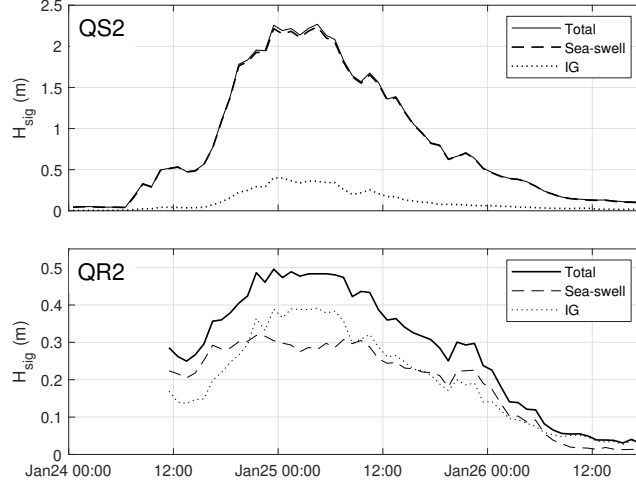


Figure 8: Band significant wave heights during the January 24–26 storm event at sensors QS2 (offshore; top panel), and QR2 (400 m into the river; bottom panel). The total height H_{sig} , SS contribution $H_{\text{sig,ss}}$, and IG contribution $H_{\text{sig,IG}}$ are represented with solid lines, dashed lines, and dotted lines respectively.

the IG contribution considerably increased in the river channel, to the point of reaching about 80% of the total wave height and overcoming the contribution of SS components in the hours of storm climax (bottom panel in Figure 8).

Presence of IG components up to the most inland river gauge is also apparent in Figure 9, in which maps of wave energy variation in time (the period of the first storm) and space (from the offshore to within the river mouth) are shown. For the maps in Figure 9 information from both sea sensors (QS1 and QS2) and all river sensors (TGdown, QR1, QR2, QR3, and TGup) have been elaborated and interpolated to obtain hourly estimates of wave energy from the offshore to into the river mouth. SS-band and IG-band wave energy (Figure 9b and c) are expressed as a percentage of the maximum total wave energy at the most offshore sensor, attained on January 25, 05:00. The SS-band-related energetic content experienced the largest decay as the river mouth is approached and entered, falling down to 10% of the total offshore magnitude (corresponding to $H_{\text{sig,ss}} \approx 0.65\text{--}0.7$ m, about 21%–23% of the local water depth) at around 200 m past the mouth (about 10 times the offshore SS wavelength) and 2% at around

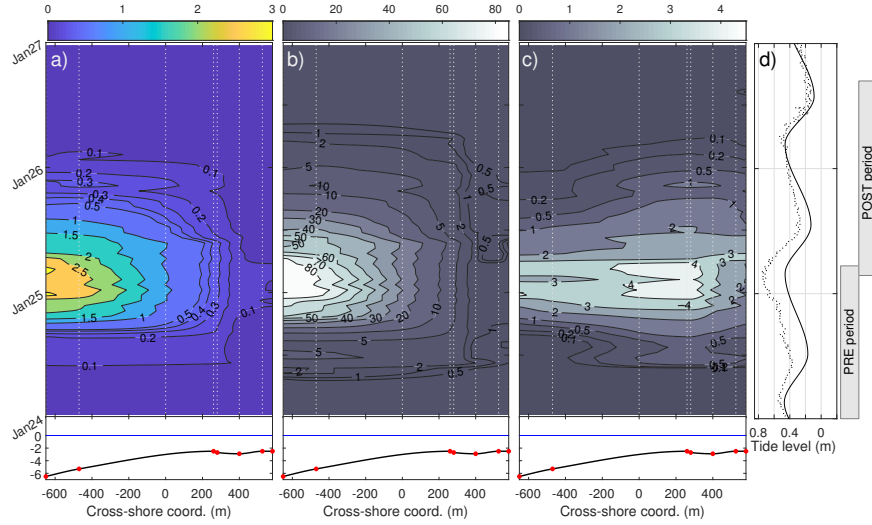


Figure 9: **a)** Total wave energy, **b)** percentage of SS wave energy (with respect to maximum offshore total wave energy), **c)** percentage of IG wave energy (with respect to maximum offshore total wave energy), **d)** astronomical tide at Ancona (black line) and tidal level as measured at Ancona harbour (dotted line). Bottom panels represent the bottom profile. Dotted white lines in panels a) to c) and red dots in the bottom panels identify the position of sensors and the river mouth ($x = 0$ m). Positive x -coordinates are directed into the river. The ranges of PRE and POST periods are labelled with grey boxes.

379 400 m ($H_{\text{sig,ss}} \approx 0.25$ m, about 8% of the local water depth) even at the hours of
 380 strongest storm energy (Figure 9b). Conversely, the IG contribution underwent
 381 a modest enhancement in the river channel, where it held up to 4–5% of the
 382 maximum offshore wave energy ($H_{\text{sig,IG}} \approx 0.45$ –0.5 m, about 15–17% of local
 383 water depth) at storm peak, in a region within 300 m (15 times the offshore
 384 peak SS wavelength) upstream of the river inlet.

385 Figure 10a gives estimates of IG significant wave heights at river sensors
 386 TGdown and TGup. The evaluation of the IG waves attenuation rate R_{IG}
 387 (Equation (5)) between the two sensors is also performed. Values of R_{IG} are
 388 shown in Figure 10b. $H_{\text{sig,IG}}$ of about 0.5 m (17% of local water depth) have
 389 been detected at TGdown in the hours of storm climax (January 25, 00:00–05:00;
 390 solid line in Figure 10a); in the same hours the IG wave height contribution at
 391 TGup, 300 m upriver of TGdown, was more than 0.3 m (10% of local water
 392 depth), resulting in R_{IG} of less than 40% as the wave field travels from TGdown

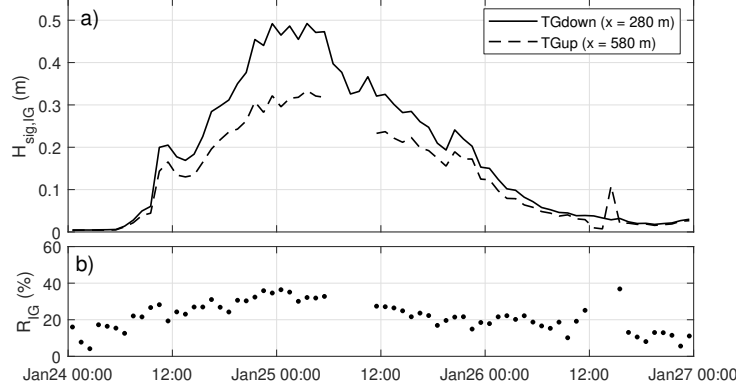


Figure 10: Decay of IG significant wave height throughout the river estuary. a) IG significant wave height $H_{\text{sig,IG}}$ at sensors TGdown (280 m from river mouth; solid line) and TGup (580 m from river mouth; dashed line) during the January 24–26 storm event; b) reduction rates of $H_{\text{sig,IG}}$.

to TGup (Figure 10b).

Figure 11 illustrates excerpts from signals at gauges TGdown (280 m upstream of the mouth) and TGup (580 m upstream of the mouth; Figure 1) on January 25, 00:00–00:20 are presented, along with the respective low-frequency components obtained through low-pass filtering. Patterns of low-frequency oscillations with zero-crossing periods ranging between 30 seconds and 3 minutes and amplitudes 0.1–0.4 m are clearly detectable beneath the SS wave train and can be retrieved at upward gauges at later times as they propagate (see the time shift of low-frequency components t_{lag} in Figure 11).

Table 1 provides estimates for IG components time lags between consecutive gauges at selected hours during the January 24–26 storm event, as well as for their propagation velocity. The test hours have been chosen as representative of the main stages of storm evolution. For each test hour, the first number is the estimate for the time shift t_{lag} of the IG wave component between two consecutive gauges, in seconds. Cross-correlation between IG signals at consecutive gauges always gave good results with correlation coefficients greater than 0.92 and p -values very close to 0 in all cases. The second number represents an estimate for the IG waves propagation velocity c_{IG} in m/s, as calculated with

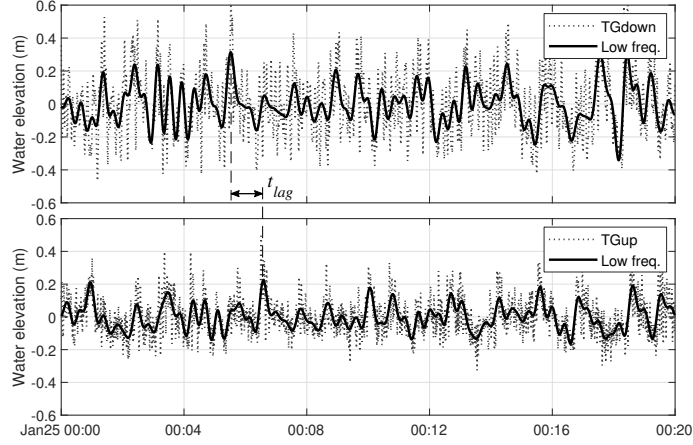


Figure 11: Time series of de-tided water elevation at sensors TGdown (280 m from river mouth; top panel) and TGup (580 m from river mouth; bottom panel) on January 25, 00:00–00:20. The dotted lines represent the de-tided signals; the low-frequency components obtained through low-pass filtering are shown with solid lines.

Equation (4). The estimates have been all computed between gauges TGdown and QR2 ($\Delta x \approx 120$ m; first row) and gauges QR2 and TGup ($\Delta x \approx 180$ m; second row), with the exception of the first test hour (January 24, 00:01; first column in Table 1). For this hour, only a correlation between gauges TGdown and TGup ($\Delta x \approx 300$ m) could have been established since QR2 was not operative at that time.

Table 2 shows estimates for **phase speeds** at sensor QR2 during the same

Table 1: Estimates of time lags and velocity propagation of low-frequency wave components into the Misa River estuary during the January 24–26 storm event. The tags under the hours indicate the storm stage.

Gauge	Hour									
	Jan 24, 00-01		Jan 24, 12-13		Jan 25, 00-01		Jan 25, 18-19		Jan 26, 08-09	
	[calm]		[storm rise]		[climax]		[storm fall]		[calm]	
	t_{lag}	c_{IG}	t_{lag}	c_{IG}	t_{lag}	c_{IG}	t_{lag}	c_{IG}	t_{lag}	c_{IG}
	[s]	[m/s]	[s]	[m/s]	[s]	[m/s]	[s]	[m/s]	[s]	[m/s]
TGdown			24.5	4.9	22	5.5	23.5	5.1	23.5	5.1
QR2	67.5	4.4	50	3.6	45.5	4.0	46.5	3.9	46.5	3.9
TGup										

418 storm. Reported are also the depth-averaged current velocities V at QR2 on the
 419 same hours (also shown in Figure 7). Assuming a mean undisturbed water depth
 420 $h = 3.1$ m at QR2, the linear theory shallow water speed $c = \sqrt{gh}$ is equal to
 421 5.5 m/s. The observed perturbation to the mean (undisturbed) water depth, $\langle\eta\rangle$
 422 (made of proper tide elevation and contributions from storm-related wave setup
 423 and storm surge) ranged from -0.2 to $+0.25$ m/s. This leads to a modified phase
 424 speed $c^* = \sqrt{g(h + \langle\eta\rangle)}$ between a minimum of 5.33 m/s in January 24, 12:00–
 425 13:00, when storm rising was paired with low tide (Figure 2), and a maximum
 426 of 5.73 m/s in January 25, 00:00–01:00, when the storm climax was paired with
 427 high tide (Figure 2) and the propagation velocity of the low-frequency signal
 428 reached its maximum value both for the TGdown–QR2 (5.5 m/s) and the QR2–
 429 TGup (4.0 m/s) intervals (Table 1).

430 When the storm was rising and the effect of storm-related setup and surge
 431 should thus increase (January 24, 12:00–13:00), low tide dominated by giving a
 432 total negative mean water elevation ($\langle\eta\rangle = -0.2$ m; Table 2). Also during storm
 433 climax in high tide (January 25, 00:00–01:00), when all agents in water level
 434 variation (tide, surge, and setup) are at their maximum potential, surge and
 435 setup are accountable for only a minor part of the mean water level elevation
 436 $\langle\eta\rangle$ achieved at that hour. A tidal positive excursion is expected of the same
 437 magnitude of the negative excursion, 0.2 m. Since the total observed excursion is
 438 $\langle\eta\rangle = +0.25$ m, tide alone is accountable for around 80% of the total excursion,
 439 with the total effect of setup and surge being roughly 20%.

440 A fair approximation of estimated c_{IG} between TGdown and QR2 can be
 441 done by subtracting mean river flow velocities V from the theoretical speeds
 442 c^* . This gives slightly underestimated values in storm climax with high tide
 443 ($c^* - V = 5.73 - 0.37 = 5.36$ m/s, comparable with $c_{IG} = 5.5$ m/s), and slightly
 444 overestimated values during both storm rise ($c^* - V = 4.98$ m/s, comparable
 445 with $c_{IG} = 4.9$ m/s) and storm fall ($c^* - V = 5.11$ – 5.19 m/s, comparable with
 446 $c_{IG} = 5.1$ m/s).

447 Tables 3 and 4 collect the same estimates evaluated for selected hours during
 448 the second major storm event, occurred on January 28–29. The second storm

Table 2: Evaluation of **modified phase speeds** at sensor QR2 ($h = 3.1$ m) during the January 24–26 storm event. The tags under each hour indicate the storm stage and the tide stage. High tide and low tide are marked in bold. The elevation of mean water level $\langle\eta\rangle$ is evaluated as excursion from the undisturbed water depth h . $c^* = \sqrt{g(h + \langle\eta\rangle)}$ is the **modified phase speed**.

	Hour							
	January 24		January 25				January 26	
	12–13	19–20	00–01	08–09	12–13	18–19	00–01	08–09
	[<i>rise</i>] [low]	[<i>rise</i>] [<i>flood</i>]	[<i>climax</i>] [high]	[<i>fall</i>] [<i>ebb</i>]	[<i>fall</i>] [low]	[<i>fall</i>] [<i>flood</i>]	[<i>fall</i>] [<i>flood</i>]	[<i>calm</i>] [high]
$\langle\eta\rangle$ [m]	-0.2	0	+0.25	0	-0.1	-0.1	-0.05	0
c_{NL} [m/s]	5.33	5.51	5.73	5.51	5.42	5.42	5.47	5.51
V [m/s]	0.35	0.32	0.37	0.4	0.31	0.23	0.15	0.07

449 was less energetic than the previous one due to weaker winds characterized
 450 by an almost constant direction for a shorter period [38]. This led to **water**
 451 **surface levels oscillating between -0.25 m and 0.25 m**, while a complete tide
 452 cycle occurred. **Specifically, almost constant wave heights ($H_s \sim 0.5$ m) were**
 453 **recorded between the two high tides, with no apparent influence from the tide**
 454 **phase, as shown in Figure 12. On the other hand,** the tidal level played a major
 455 role over the propagation velocity of low frequency components in spite of the
 456 microtidal nature of the estuary, mainly by increasing water heights into the
 457 river and thus letting waves propagate with a larger celerity (see Table 3). High
 458 tide on January 28, 07:00–08:00 and January 29, 08:00–09:00 promoted a quicker
 459 propagation of low frequency waves thanks to higher water columns within the
 460 channel. The effect of high tide was felt up to TGup, the most landward tide
 461 gauge inside the river, giving slightly higher c_{IG} also between QR2 and TGup
 462 (see the second and sixth columns in Table 3). The lowest c_{IG} , conversely, were
 463 attained on January 28, 15:00–16:00 (fourth column in Table 3), when low tide
 464 led to a strongly negative $\langle\eta\rangle$ (-0.55 m; Table 4) in spite of storm climax. The
 465 major influence of tide is also highlighted by the fact that positive mean water
 466 elevation $\langle\eta\rangle$ is achieved even for a falling storm (January 29, 08:00–09:00). Also
 467 for the second storm, $c^* - V$ gives slightly overestimated values of c_{IG} for both
 468 storm rise in high tide ($c^* - V = 5.58$ m/s if compared to 5.4 m/s) and storm
 469 climax in low tide ($c^* - V = 4.79$ m/s if compared to 4.6 m/s).

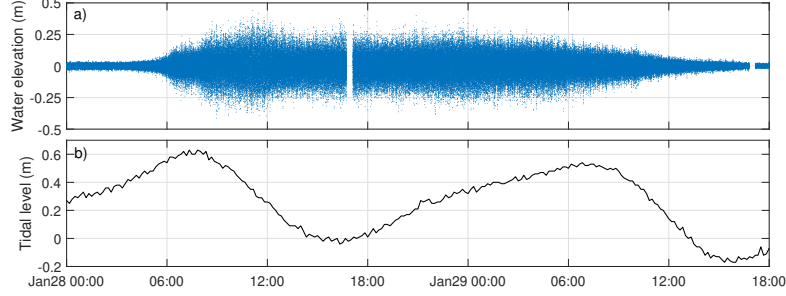


Figure 12: **a)** De-tided and de-trended water elevation time series for the January 28–29 storm event at TGdown. **b)** Storm tide level as measured in Ancona harbour during the same period.

Table 3: Estimates of time lags and velocity propagation of low-frequency wave components into the Misa River estuary during the January 28–29 storm event. The tags under the hours indicate the storm stage.

Gauge	Hour											
	Jan 28, 01-02		Jan 28, 07-08 <i>[high tide]</i>		Jan 28, 11-12		Jan 28, 15-16 <i>[low tide]</i>		Jan 29, 01-02		Jan 29, 08-09 <i>[high tide]</i>	
	t_{lag} [s]	c_{IG} [m/s]	t_{lag} [s]	c_{IG} [m/s]	t_{lag} [s]	c_{IG} [m/s]	t_{lag} [s]	c_{IG} [m/s]	t_{lag} [s]	c_{IG} [m/s]	t_{lag} [s]	c_{IG} [m/s]
TGdown	22	5.4	22	5.4	23.5	5.1	26	4.6	24	5.0	23	5.2
QR2	46	3.9	43.5	4.1	47.5	3.8	51.5	3.5	47	3.8	46	3.9
TGup												

Table 4: Evaluation of **modified phase speeds** at sensor QR2 ($h = 3.1$ m) during the January 28–29 storm event. The tags under each hour indicate the storm stage and the tide stage. High tide and low tide are marked in bold. The elevation of mean water level $\langle \eta \rangle$ is evaluated as excursion from the undisturbed water depth h . $c^* = \sqrt{g(h + \langle \eta \rangle)}$ is the **modified phase speed**.

	Hour						
	January 28				January 29		
	01-02 <i>[calm]</i> <i>[flood]</i>	07-08 <i>[rise]</i> [high]	11-12 <i>[climax]</i> <i>[ebb]</i>	15-16 <i>[climax]</i> [low]	01-02 <i>[climax]</i> <i>[flood]</i>	08-09 <i>[fall]</i> [high]	
$\langle \eta \rangle$ [m]	-0.15	+0.15	-0.2	-0.55	-0.1	+0.2	
c_{NL} [m/s]	5.38	5.65	5.33	5	5.42	5.69	
V [m/s]	0	0.07	0.19	0.21	–	–	

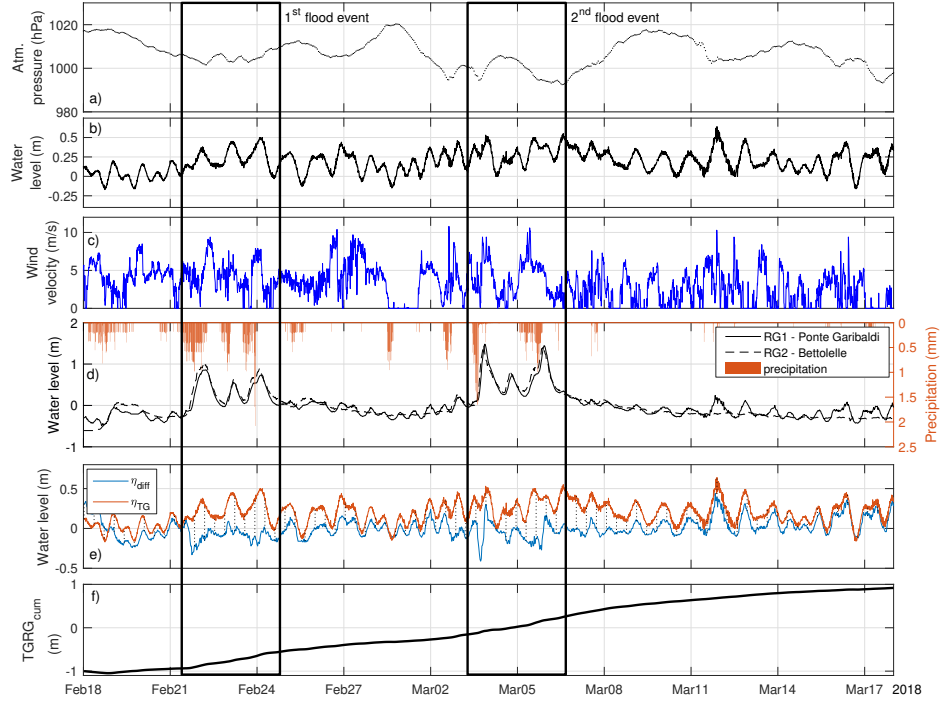


Figure 13: Measurements collected by the tide gauge in Ancona harbour in February–March 2018 and comparison with river gauge measurements. **a)** Atmospheric pressure, **b)** water level (the ordnance datum is the mean sea level), and **c)** wind velocity at the Ancona harbour. **d)** Comparison between water level observations at river gauges RG1 (1.5 km from estuary mouth; solid black line) and RG2 (10 km from estuary mouth; dashed black line), and precipitation levels (orange bars). **e)** Difference between hydrometer measurements (η_{diff} , blue line) compared with storm tide (η_{TG} , orange line). **f)** Cumulative difference between η_{TG} and η_{diff} . The two major flood events are evidenced in black boxes.

3.6. Interaction between tide and river forcings in the Misa River

In view of the ascertained role of tidal action over wave propagation dynamics in the MR estuary, the sea surface oscillations recorded by the tide gauge in the harbour of Ancona (TG hereafter) is here compared to data from the hydrometers at “Ponte Garibaldi” and “Bettolle” (respectively RG1 and RG2 in Figure 1) to better understand the penetration of tidal forcing into the MR estuary. Figure 13a–c illustrate data of atmospheric pressure, water level and wind speed recorded at TG in February–March 2018, while Figure 13d–f show measurements at TG, RG1 and RG2, as well as precipitation data.

The February–March 2018 time period has been chosen because the recorded

480 conditions are comparable to those observed during the Bora storm occurred
 481 on 24–26 January 2014 (see [38] for comparison), when measurements at RG1
 482 and RG2 were not available. Specific high-flow conditions have been identified
 483 in two distinct time ranges, i.e. February 21–24 and March 3–6 (see black boxes
 484 in Figure 13). Although they seem to show periods of about 24 hours, the three
 485 contiguous peaks in water level at RG1 and RG2 during these high-flow events
 486 are due to both the time interval between contiguous precipitations and the
 487 hydrologic routing of precipitation through the MR watershed t_c .

488 Five rainfall gauges have also been selected² to evaluate the mean precip-
 489 itation on the MR watershed (orange bars in Figure 13d). Three significant
 490 precipitation events (1–2.5 mm) occurred in February, just before the peaks in
 491 the first flood event, and two significant events (1–2 mm) occurred in March,
 492 just before two flood peaks in the second flood event.

493 Oscillations periods of 12 to 24 hours are observed at the downstream gauge
 494 RG1 (see also [42]) in phase with tidal cycles at TG. This shows that tide
 495 reached relatively large distances from the river mouth, especially during low-
 496 to moderate-flow conditions, although the MR environment is classified as mi-
 497 crotidal. To better analyze this aspect, the TG recordings (η_{TG}) have been
 498 compared to the difference between RG1 and RG2 measurements (η_{diff} ; see
 499 Equation (6)), with the aim to remove the river forcing effects, similar at RG1
 500 and RG2, and highlight the tide influence. Figure 13e shows that the two time
 501 series are in phase during low-flow conditions (outside the black boxes) and
 502 suggests that the tide only affected RG1, with negligible effects on RG2. On
 503 the other hand, η_{diff} and η_{TG} are not in phase during flood events (inside the
 504 black boxes); this suggests that, in high-flow conditions, the river forcing largely
 505 prevailed over tide action and prevented the tide from reaching the downstream
 506 station.

²Regional database of the “Centro Funzionale Multirischi per la Meteorologia, l’Idrologia e la Sismologia”, Marche Region

507 4. Discussion

508 4.1. Evolution of wave components

509 The January 24–26 storm featured an intense sea state with a sustained
510 energy injection at high frequencies and a homogeneous redistribution of energy
511 among all frequencies as the storm reached its climax (left panels of Figure 3). A
512 link between swell waves and low-frequency waves is apparent across the whole
513 storm. This can be explained by IG wave frequencies receiving energy from
514 shorter wave frequencies, typically as a byproduct of several mechanisms like
515 nonlinear interactions, shoaling and bore merging in the surf zone [5, 12, 14].

516 Although nonlinear interactions between sea/swell components may occur in
517 the shoaling zone and generate IG waves, the bound wave mechanism [5] seems
518 to be the dominant IG generation process in the coastal zone of the MR estuary.
519 Such mechanism is typical of mild slope regimes characterized by a normalized
520 slope parameter $\beta_b < 0.3$ [45]:

$$\beta_b = \frac{h_x}{\omega} \sqrt{\frac{g}{h}}, \quad (8)$$

521 where h_x is the bed slope, ω the wave angular frequency, and h a characteristic
522 depth value. For the studied area, $h_x \sim 0.005$, while the dominant offshore
523 wave frequency during storm climax is $f = 0.1$ Hz (Figure 3b), which provides
524 $\omega = 0.63$ rad/s. Since off the mouth the typical depth is $h = (3\text{--}6)$ m, β_b is
525 around 0.010–0.014, suggesting a dominance of the bound wave mechanism.

526 Storm decay, conversely, showed an homogeneous reduction of energy flux at
527 all frequencies, with the SS frequency range still being energetically dominant
528 (right panels of Figure 3).

529 As waves moved into the channel, IG motions gained a more relevant role
530 over swell and wind waves. As the storm intensified offshore, the IG band
531 experienced the largest increment in flux energy density into the river while SS
532 energy was unaltered (left panels of Figure 5). This was favoured by intense
533 wave breaking and energy dissipation affecting mainly SS components outside
534 the river inlet (Figure 6).

535 The contribution of IG waves to wave height in the river closely followed the
 536 offshore significant wave height [28, 46] and reached values of 0.5 m, 0.4 m and
 537 0.3 m respectively 280 m, 400 m and 600 m from the river mouth (Figures 8
 538 and 10). This corresponds to a decay from about 25% of the local depth at
 539 QR3, to about 10% of the local depth at TGup. The contribution of sea and
 540 swell, on the other hand, came to saturation (bottom panel of Figure 8) and is
 541 reflected by a sharp drop in wave energy (Figure 9).

542 Moreover, due to a river discharge evaluated as $Q = (45 - 50) \text{ m}^3/\text{s}$ during
 543 the storm peak and to the plume extending up to 1.3 km off the mouth [38],
 544 the increased average velocity of the river flow evidenced after intense rainfall in
 545 January 24 (up to 0.45 m/s at 400 m from the mouth) was strong to the point
 546 of damping and deviating storm-driven sea currents as much as 650 m offshore
 547 of the river mouth (Figure 7), but not enough to block low frequency waves.
 548 This is in agreement with observations at the Albufeira Lagoon inlet [30], where
 549 gravity waves are damped as long as the ebb-tidal delta is crossed, and IG waves
 550 propagate beyond the inlet provided the opposing current (given by ebb tidal
 551 flows) is not strong enough.

552 Low-frequency components manage to hold moderate, tide-modulated prop-
 553 agation speeds that are always directed upriver (in the range 3.6–5.5 m/s; Ta-
 554 bles 1 and 3). Tide is indeed dominant over storm surge and wave setup in
 555 controlling the upriver propagation velocity of IG components against the river
 556 current.

557 Overall, we can say that the MR estuary and nearshore zone acted as a
 558 low-pass filter capable of removing higher-frequency waves (dissipation outside
 559 the mouth) while retaining low-frequency energy. A similar behaviour has been
 560 so far observed and documented only for oceanic inlets. IG waves propagating
 561 upstream as bores have been identified at the Pescadero inlet [28]. Similarly,
 562 fluctuations in the IG range, responsible for up to 50% of flow velocity oscil-
 563 lations, have been observed in the Albufeira Lagoon [2, 29], and IG velocity
 564 amplitudes of 0.1 m/s are detected in the Ría de Santiuste estuary [37]. These
 565 inlets are subjected to higher tide excursions, periodic cut-off from the open

566 sea and much more severe wave regimes than the MR estuary (tidal excursions
567 between 1.4 m and 3.6 m **and** significant wave height between 1.9 m and 4 m).
568 No studies of IG energy propagation have been performed for inlets with low
569 tide excursion, to the best of authors' knowledge. Therefore, our work marks
570 the first observations of such "filtering effect" in a microtidal setting and com-
571 paratively milder wave climates common in enclosed basins like the Adriatic
572 Sea.

573 We expect that the highly engineered final reach of the MR, the effects
574 of which are not investigated in this work, might play a role on the observed
575 phenomena. Uncles et al [37] suggested that the dominant frequencies of IG
576 waves penetrating into the Ría de Santiuste are close to resonance periodicities
577 typical of that estuary, and their effects are thus possibly amplified by resonance.
578 The concrete jetties that bound the flow of the MR in its final stretch may have
579 a similar effect and furnish an explanation to the focussing and amplification
580 of IG waves energy observed inside the channel, up to 300 m landward of the
581 mouth (Figure 9c). Our data sets, however, did not allow us further analyses
582 on this regard.

583 *4.2. Influence of tide in the estuary*

584 Several studies [1, 28, 30, 47, 48] have ascertained that tide and related
585 processes are crucial factors in determining how much of the incident wave
586 energy is able to cross an inlet mouth or estuary, by inducing depth-limited wave
587 breaking over shoals and bars [48] and enhancing/reducing residual wave heights
588 in backbarrier lagoons and estuaries when associated to ebb/flood currents [1, 2,
589 30, 47]. These observations, though, are often obtained for coastal environments
590 with significant tidal excursions.

591 Our study proposes novel findings that tide exerts some effect also in mi-
592 crotidal estuaries. In particular, similarly to what observed in macrotidal en-
593 vironments (see Figures 7 and 8 in [30] as an example), it is shown that the
594 tide has a fundamental role in controlling the upriver propagation of IG waves.
595 Further, IG waves propagation velocities are enhanced during flood tide and

high tide (Table 3), mainly due to an increase of water levels into the estuary. The majority of the water levels increment has to be ascribed to tide also in the phases of greatest storm intensity, when the combined effect of storm-induced setup and surge account for no more than 20% of the total mean water elevation. Moreover, albeit the tidal range at the MR estuary is low (less than 0.6 m between high tide and low tide in the investigated period), tidal oscillations are visible at all river sensors up to 1.5 km from the mouth (Figure 13). Tidal currents in the MR estuary, conversely, are low and rarely affect the mean velocity and orientation of the river current (Figure 7). The MR current is directed up-river only when flood/high tide occurs in coincidence with low-flow conditions in both river and sea. Other than this, river forcing is generally dominant over tide in wintertime conditions, and the effect of tide is obliterated during river floods, as observed in [38].

Wave setup, forced by wave breaking and energy dissipation in front of the estuary mouth, was relatively small and not critically modulated by the tide. Nonetheless, the tide is shown to promote some persistence of wave-induced setup in the region immediately off the estuary, where moderate gradients of radiation stress appear to be maintained by low tide [2] during storm decay (Figure 6).

5. Conclusions

The present study inspected the hydrodynamic behaviour of the microtidal, wave-dominated estuary of the Misa River (Italy). Microtidal inlets with wave dominance are common along the Adriatic Sea. Analysis of field measurements during a storm event in January 2014 and signals from hydrometers in February–March 2018 allowed us to observe evolutionary trends for waves of different frequencies (SS waves, IG waves, and tide) both offshore and into the river mouth.

For the first time, the mechanism of “low-pass filtering” of wave energy exerted by several tide-dominated inlets around the world [2, 28, 30] has been

625 assessed also for a microtidal environment. In tide-dominated inlets, wave prop-
626 agation, enhancement and dissipation are often altered by strong ebb/flood cur-
627 rents [2, 30]. Inlets and back-barrier lagoons may get “cut off” from offshore
628 wave action during low tides [28]. In coasts characterized by relevant sediment
629 transport dynamics, inlets may also be filled by the concurrent action of tide
630 and waves [29]. All the aforementioned processes may modify the penetration
631 of wave energy in a significant way. This study demonstrates that also in a
632 microtidal estuary, with near-zero tidal currents and always in connection with
633 the sea, a natural filtering of wave frequencies occurs.

634 High-frequency storm waves from wind and swell strongly dissipate and de-
635 cay outside of the MR mouth in the most seaward portion of the river, their
636 energy contribution being negligible after 400 m from the inlet. Low frequency
637 (IG) waves, on the other hand, slightly increased their relative energy and even-
638 tually overpowered high-frequency oscillations in the channel. Short wave dissi-
639 pation is promoted by shoaling and breaking processes in shallow depths, as well
640 as by strong opposing river currents after heavy rainfall. The enhanced river
641 flow, though, is not able to dampen low frequency waves as much as strong ebb
642 currents do in tidal inlets [2, 30].

643 In spite of the microtidal nature of the MR estuary, tidal action is felt
644 across a large distance. In the final reach of the river, high tide promotes a
645 faster IG waves upriver propagation by means of increased water depths. Tidal
646 oscillations are also detected across all river sensors except the most landward
647 one, particularly during low- to moderate-flow conditions. Due to the modest
648 tidal excursions, however, tidally-generated currents are not able to alter river
649 flows significantly, if not during extremely low river flow regimes.

650 A more detailed analysis would surely benefit from other comprehensive,
651 simultaneous measurements of waves, tides, and river discharge. Information
652 from a video-monitoring station recently installed near the estuary, moreover,
653 would complement the study with a reconstruction of morphodynamic features
654 in the nearshore. Finally, although we believe it may have a determinant role,
655 the effects of river engineering on the processes studied in this work and the

656 potential for phenomena of wave resonance have not been investigated. All of
657 the abovementioned aspects will have to be analyzed in future research.

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