

# Topographical changes caused by moderate and small floods in a gravel ~~bed~~<sup>ly</sup> ephemeral river - ~~2D~~<sup>-depth-averaged</sup> morphodynamic simulation approach

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**Abstract.** In ephemeral rivers, channel morphology represents a snapshot at the end of a succession of geomorphic changes ~~performed~~<sup>caused</sup> by a flood. In most cases, the channel shape and bedform ~~evolution~~<sup>migration</sup> during different phases of a flood hydrograph are not recognized from field evidence. ~~Understanding of these processes would be needed particularly for flood mitigation purposes.~~ This paper analyzes the timing ~~onf~~<sup>of</sup> bed-river bed erosion and deposition ~~changes~~<sup>of</sup> a gravel bed ephemeral river channel (Rambla de la Viuda, Spain) during consecutive, moderate- (March 2013) and low-magnitude (May 2013); ~~discharge events, by applying a morphodynamic modelling (Delft 3D) approach~~<sup>calibrated with pre- and -post-events survey by RTK-GPS points and mobile laser scanning</sup>. Based on the simulations, we analyze 1) when and to what extent the channel changes occur during the phases of moderate- and low-magnitude discharge events, and 2) which flow characteristics explain these changes throughout the discharge events. ~~Based on the simulations, we analyze 1) when and to what extent the channel changes occur during the phases of moderate- and low magnitude discharge events, and 2) which flow characteristics explain these changes throughout the discharge events.~~ This paper analyzes the capabilities of morphodynamic modelling (Delft 2D) for resolving the evolution of a gravelly ephemeral river channel during consecutive, moderate- and low-magnitude discharge events. We pursue for schematic concepts for simulations in ephemeral gravelly rivers that provide an outcome with the closest similarity to the post-flood reality. Based on the simulations, we analyze the morphodynamic evolution of Rambla de la Viuda (Spain) to find out when and to what extent the channel changes occur during the phases of moderate- (March 2013) and low-magnitude (May 2013) discharge events, and what influence the discharge has on the rate of morphological changes. The model performance is examined with different parameterizations.

The calibrated model worked well, however, spatially varying grain size data and transport equations were the most important factors, in addition to the quality of recorded discharge, for the simulation results of the channel evolution model performance. The total load equation worked better, compared to the deterministic equation. The studied system reach was mainly depositional based on the field evidence and all bedload sediment supplied from upstream adjacent areas is trap in the study segment forming gravel lobes. Therefore, estimates of total bedload sediment mass balance can be obtained from pre- and - post field survey for each flood event. The spatially varying grain size data and transport equations were the most important factors for model calibration, in addition to the flow discharge for the model performance. The channel acted as braided channel during the lower flows of the two discharge events, but when bars were submerged in higher discharges, the high fluid forces followed a meandering river planform. The model results also showed that erosion and deposition can be were in total greater during the hydrograph long-lasting receding phase than during the rising phase of the flood hydrographs. In the case of moderate-magnitude discharge event, The deposition and erosion peaks are were predicted to occur at the beginning of the hydrograph-moderate-magnitude discharge event, whereas deposition dominates throughout the event. On the contrary, the low-magnitude discharge event only experienced the peak of channel changes after the discharge peak, suggesting a rupture of upper layer before sediment is moved. Thus, both type of discharge events highlight the importance of receding phase for this type of gravel bed ephemeral river channel.

## 1 Introduction

The hydrology of ephemeral rivers is dominated by occasional large flash floods that cause morphological fluvial changes (Tooth, 2000; Benito et al., 2011). The costs of these floods due to their catastrophic nature include major economic, social and environmental aspects (Petersen, 2001). These are caused by both hydro- and morphodynamics during the discharge events. For reducing the emergency costs and enhancing the preventive flood mitigation measures, understanding of the forces of the flow and related channel changes throughout the discharge events would be important. For being able to allocate the measures temporally most efficiently, the understanding of the timing of the morphodynamics and their effects to the adjacent river environment throughout the discharge events are needed.

~~Consequently, m~~Most geomorphological studies have been mainly concentrated on these greatest-large flow-discharge events, due to their impacts on the river channel changes, related river environments and human infrastructure (Greenbaum and Bergman, 2006; Grodek et al., 2012; Nardi and Rinaldi, 2015; Hooke, 2016b). However, moderate and low flows have also been shown to cause great morphological changes in gravelly-gravel bed river channels (Calle et al., 2015; Hooke, 2016a), as a small discharge over long time spans can substantially rework the sediment and fluvial bedforms produced by the greater floods (Greenbaum and Bergman, 2006).

According to Hooke and Mant (2000), the pattern and magnitudes of fluvial morphological changes show the best relationship with the magnitude of peak discharge. However, during flash floods, it is difficult to perform sediment transport or bedform migration measurements to detect the timing of topographical changes, i.e. whether the greatest changes occur for example due to the peak discharge, the slope of the rising limb, or the length of the receding limb. Case studies have reported most erosion during the rising and the peak flow phases in perennial rivers (e.g. Gendaszek et al. 2013), but similar knowledge for ephemeral river channels is still low.- It would be important to understand the capacities of ephemeral rivers for sediment deposition and flooding due to the combined effects of water flow and sediment transport during flood situations. If the erosion and deposition would be in total greater during the receding phase than in rising phase, the receding phase should not be ignored while planning flood mitigation measures.

~~Thus, the timing of the morphological changes during the different phases of flow discharge events hydrographs is still unresolved, and how they differ from the gravelly bed perennial rivers, with e Case studies have reportinged most erosion during the rising and the peak flow phases of in perennial rivers (e.g. Gendaszek et al. 2013). It would be important to understand the capacities of ephemeral rivers for sediment deposition and flooding due to the combined effects of water flow and sediment transport during the flood situations. If the erosion and deposition would be in total greater during the receding phase than in rising phase, the receding phase should not be ignored while planning flood mitigation measures. The erosion~~

and deposition have been shown to be greater during the rising limb of flood hydrographs in perennial rivers (Gendaszek et al., 2013), but knowledge of the ephemeral river channels is still low.

Conventionally, ~~these~~ ephemeral river channel changes associated with specific flood events are interpreted on the basis of the post-flood bedform and grain size distribution (e.g. Euler et al., 2017). New insight on morphodynamic changes and their driving parameters during flash floods can be gained by~~The only way to gain information and to show any concepts about topographical changes during the flash floods is to~~ applying simulation methods. Simulations may provide information about the channel dynamics from the times when it has not been possible to perform measurements (Lotsari et al., 2014a) and thus increase our understanding of sediment dynamics during flood events (Hooke et al. 2005). However, morphodynamic ~~simulations-modelling~~ of gravel migration have so far been more common in perennial braided gravel bed rivers (e.g. Williams et al., 2016b3) or laboratory flumes (Kaitna et al., 2011) than in ephemeral rivers. ~~There are~~ examples of simulating sediment transfer in ephemeral channels (Graf, 1996), although most of the few recent simulations of ephemeral rivers have been performed in sandy reaches (Billi, 2011; Lucía et al., 2013) and on alluvial fans (Pelletier et al., 2005). Ephemeral rivers with gravel-sized bedload particles have seldom-rarely been simulated. One of the few existing examples was carried by Hooke et al. (2005), simulating morphological changes during flash floods with a cellular automata model. Their model worked well with simulations using moderately large discharges during clear water conditions, but discharge events with sediment loads had some tendency for excess deposition (Hooke et al., 2005). Further examination of different discharge events with modelling approaches is thus required.

It is possible to calibrate the morphodynamic model based on the ~~between-pre- and post-after~~ flood topographies of different consecutive flood situations. This causes uncertainties to the modelling, as calibration and validation data is measurable only during dry channel conditions. However, ~~but~~ it is the best available way to get understanding on the temporal evolution of the ephemeral river channels during flooding. In particular, the differences and similarities in model performance between different flood magnitudes can be detected (Lotsari et al., 2014a; Williams et al., 2016a). Recently, the measurement techniques for deriving this calibration data for morphodynamic modelling have increased. One of them is accurate laser scanning (mobile and terrestrial), which enables to capture the channel topography before and after flooding in detail (Milan et al., 2007; Vaaja et al., 2011; Calle et al., 2015; Kasvi et al., 2015; Kukko et al., 2015). The laser scanning enables rapid measurements of gravel bed rivers at the sub-grain level resolution (Milan et al., 2007). In ephemeral rivers the quality of topographical data can be very good and the uncertainties are less than in perennial rivers, either with laser scanning or traditional RTK-GPS measurements, because these rivers can be surveyed when the river bed is dry. The high uncertainties in topographical measurements of sub-water areas in gravel bed perennial river have been related particularly to the high bed load velocities and temporal variability of bed load (Williams et al., 2015). Mobile laser scanning, for example from earvehicle, or backpack and boat, provides in an ephemeral river a practical scanning angle to survey channel banks, bar lobes and vertical surfaces (Vaaja et al., 2011; Kasvi et al., 2015; Kukko et al., 2015). ~~and a~~ Also data collection over large channel bars has been

enhanced with these methods. (~~Vaaja et al., 2011; Kasvi et al., 2015; Kukko et al., 2015~~). Digital elevation models produced from LiDAR-airborne laser scanning (ALS) data have been applied for detecting the geomorphic effects of different discharge events (Hauer and Habersack, 2009; Croke et al., 2013; Thompson and Croke, 2013; Nardi and Rinaldi, 2015) and for recording and calibrating sediment transport models (Rodriguez-Lloveras et al., 2015). These reveal the reorganization of the channel morphology due to flood events even over large areas with great detail (Thompson and Croke, 2013). These accurate topography data, together with RTK-GPS measurements, enable more detailed analysis-calibration of the morphodynamic model than before. ~~Despite these advances in measurement techniques and simulation approaches, there is a need for testing the model performance in comparison to the detailed measurements related to flow events of different magnitudes, and not only to high/large floods. In addition, morphodynamic models are suited with a range of sediment transport equations to further test for simulating moderate and small floods of ephemeral channels.~~

This paper analyzes ~~the capabilities of morphodynamic modelling for resolving~~ the evolution of a gravelly gravel bed ephemeral river channel (Rambla de la Viuda, Spain) during consecutive, moderate- (March 2013) and low-magnitude (May 2013), discharge events, by applying a morphodynamic modelling (Delft 3D) approach ~~moderate- and low-magnitude, discharge events~~. Based on the simulations, we analyze 1) the timing of river bed erosion and deposition in relation to the flow hydrograph phases during moderate- and low-magnitude discharge events, 2) the hydraulic characteristics (e.g. shear stress) explaining these channel and bedform morphodynamics, and 3) the prevailing fluvial processes, and related sediment transport routing, during these different magnitude discharge events in a gravel bed ephemeral stream. Based on the simulations, we analyze 1) the timing on bed river erosion and deposition modes leading hydrodynamic changes in relation to the flow hydrograph phases during moderate- and low-magnitude discharge events, and 2) the hydraulic characteristics (velocity, depth, shear stress) explaining channel and bedform morphodynamics, and 3) the implications of this modelling to infer river flow processes (e.g. sediment rates of bedload transport) in gravel bed ephemeral streams. Understanding of these processes would be needed particularly for river management works and flood mitigation purposes. ~~Based on the simulations, we analyze 1) when and to what extent the channel changes occur during the phases of moderate- and low-magnitude discharge events, and 2) which flow characteristics explain these changes throughout the discharge events.~~ We pursue for schematic concepts for simulations in ephemeral gravelly rivers that provide an outcome with the closest similarity to the post flood reality. This study is based on both accurate topographical measurements with RTK-GPS survey and mobile laser scanning, before and after each discharge event, and morphodynamic simulation (2D implementation of Delft-23D). Thus, the model is calibrated with data from moderate magnitude event, and then validated based on the consecutive low-magnitude event. ~~two flash flood events.~~ Based on the simulations, we analyze the morphodynamic evolution to find out when and to what extent the channel changes occur during the different phases of discharge events, and what influence the discharge has on the rate of morphodynamics. With this simulation method it is possible to improve and deepen the analyses earlier made based on only the survey work (e.g. Calle et al., 2015).

## 2 The study area

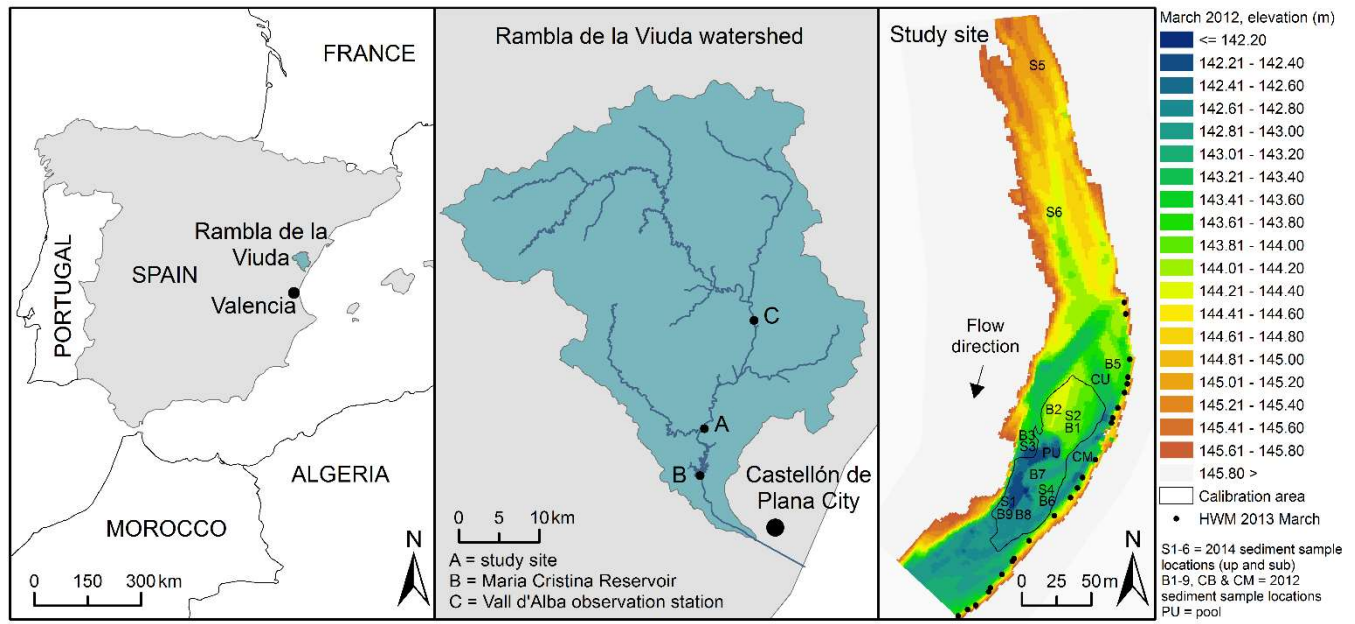
Rambla de la Viuda is an ephemeral stream with a catchment area of 1523 km<sup>2</sup>, located in eastern Spain (Fig. 1). The river has braided pattern associated ~~to-with~~ a high sediment supply (Calle et al., 2015). The river is prone to flash floods, and the stream flow occurs on an average of 20 days per year. Due to the high infiltration ~~and sporadic nature of rain events~~, only intensive rains, i.e. a run-off threshold of 65 mm, cause flow in the river corridor (Segura-Beltrán and Sanchis-Ibor, 2013). This run-off threshold may vary between seasons, due to soil moisture and the frequency of rains. The rain producing floods in this catchment are caused by mesoscale convective complexes (autumn heavy rains), and those may be up to 100 km in diameter. The rains producing these floods occur in autumn and spring, and are related to Atlantic fronts, which do not give much spatial variability.

The study site is located 8 km upstream of the dam of the María Cristina reservoir. The simulation area is circa 440 m long (length along the channel) and includes an advancing gravel lobe bar front, which is the lowest 200 m of the simulation area. ~~This lowest 200 m long reach is where this applied as the calibration area -study will be focused on-~~ (Fig. 1). This area was selected essentially because it was possible to ~~assess-estimate~~ the ~~bedload sediment input gravel movement~~ with certainty (Calle et al., 2015). This is due to the gravel extraction ~~/as part of river~~ mining ~~activities~~ in recent decades, which has produced a flat valley bottom on which gravel bars were deposited. Recently, new bars have been deposited by floods, and gravel lobes with a sharp front have been formed in the downstream edge of the area. The gravel bars and lobes of the simulation area are non- or weakly-armoured (see also Table 2) and move freely during low-magnitude flows.

The two events under study occurred in spring 2013. These were caused by two rain events, which were recorded with 5 minute interval at the precipitation and gauging station provided by the Automatic System of Hydrologic Information (SAIH-Jucar) at Vall d'Alba (Fig. 1 and Calle et al., 2015). These intense spring rains caused the two flood events. The first rain event (March 2013) started on 4<sup>th</sup> of March and lastinged for three days with a total .All together 70 mm-precipited. The second rain event (April-May 2013) started on 27<sup>th</sup> April and lasted four days with .All togethera total of 72 mm-precipited during those days. These rain eventsEvent caused flows that lasted 13 days (started at 12:00 on 5<sup>th</sup> March) and and 9 days (started 14:10 on 30<sup>th</sup> April), respectively. -Even though-The peak discharges of 23 (at 11:05 on 6<sup>th</sup> March) and 12.5 m<sup>3</sup> s<sup>-1</sup> (at 9:05 on 1<sup>st</sup> of May) were registered in the gauging station of Vall D'Alba, respectively. The hydrograph and precipitation data has already been presented at Calle et al. (2015). Due to the continuous widely spread rain events, the peak discharges on the study site were estimated to be greater at the study site than at Vall D'Alba station that due to the rain occurring between the study site and the observation stations (Calle et al. 2015). The hydrograph of the study site can be assumed to be the same shape as measured at Vall D'Alba, due to the widespread continuous character of the rain events. The hydrograph and precipitation data has already been presented at Calle et al. (2015). Thus, no sporadic events occurred. Event flows lasted 13 and 9 days, respectively. These two discharge events transported 12–41 mm sized gravel (D<sub>50</sub> values) according to the measurements (see



section 3.3 below), which had been performed in the areas of topographical changes. These have been also published at Calle et al. (2015). The movement of these gravels caused the development of the bar fronts (Calle et al., 2015).



5 **Figure 1: The study site of the Rambla de la Viuda, Eastern Spain. The elevation on March 2012 is derived mainly from mobile laser scanning data, but the edges are from the 2009 national digital elevation model. The analyzed-area calibration area was defined, based on the data coverage of March 2013. The high water mark (HWM) measurement locations of 2013 March are shown (their values are plotted in Fig. 3). Also the sediment sample locations are shown (see also Table 2 and Fig. 2).**

Calle et al. (2015) described the main morphological and topographic changes in the same river reach (i.e. around the calibration area of Fig. 1) caused by these moderate (March 2013) and small (May 2013) discharge events based on multi-temporal mobile laser scanning (MLS) and RTK-GPS surveys before and after the floods (Fig. 2). They also related the observed morphological and sediment textural changes with hydraulic parameters (flow velocity, depth and discharge) estimated by a two-dimensional implementation of hydrodynamic model (Delft-2D) and investigated whether the combination of the applied techniques is a reliable method to study the morphodynamics of a flood event. It was shown that MLS surveys and additional RTK-GPS surveys are suitable for a dryland river environment. In addition, a two-dimensional hydrodynamic simulation was able to estimate the hydraulic characteristics associated with the discharges. Change detection and spatial grain size distribution analyses after flood showed a high availability of material (up to  $D_{84}$  of 32–45 mm) and absence of a well-developed armoured layer (Calle et al., 2015). Thresholds for sediment transport were proven to fit with the Hjulström graph (1935) in this ephemeral environment. However, simulations did not include topographical change and its influence on the water level and the hydraulic parameters during the floods, and could not answer how sediment transport and topographic changes evolved during rising limb, peak stage and receding limb of the hydrographs. In this paper, we investigate the modelling further for answers to those questions and deepening the analyses based on only with topographical measurements.

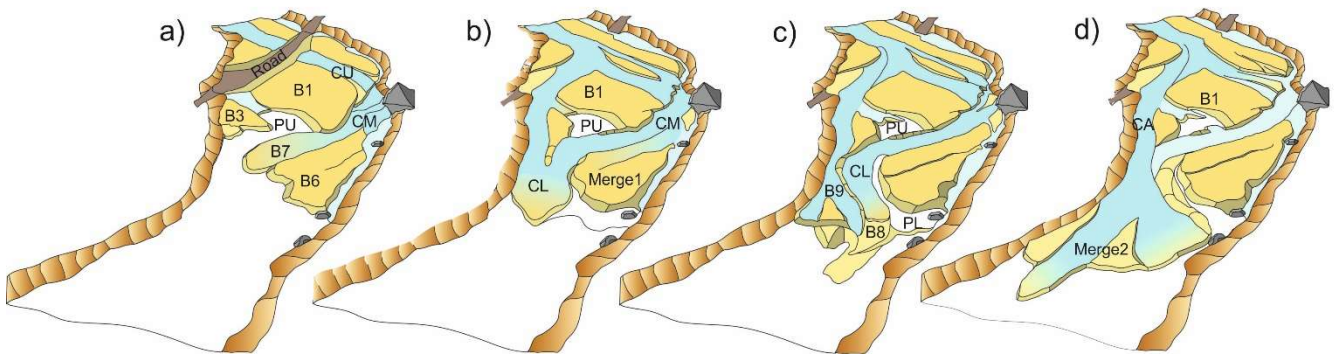


Figure 2: The conceptual graphic presented by Calle et al. (2015: Figure 10, [www.schweizerbart.de/journals/zfg](http://www.schweizerbart.de/journals/zfg)): the pre-stage of 2012 (a), high flood stage during March 2013 event (b), low floor stage during May 2013 event (c) and final stage after May 2013 (d). CU: Upper Channel, CM: Middle Channel, CL: Lower Channel, PU: Upper Pool, PL: Lower Pool, B1-9: gravel Bars (cf. Calle et al., 2015).

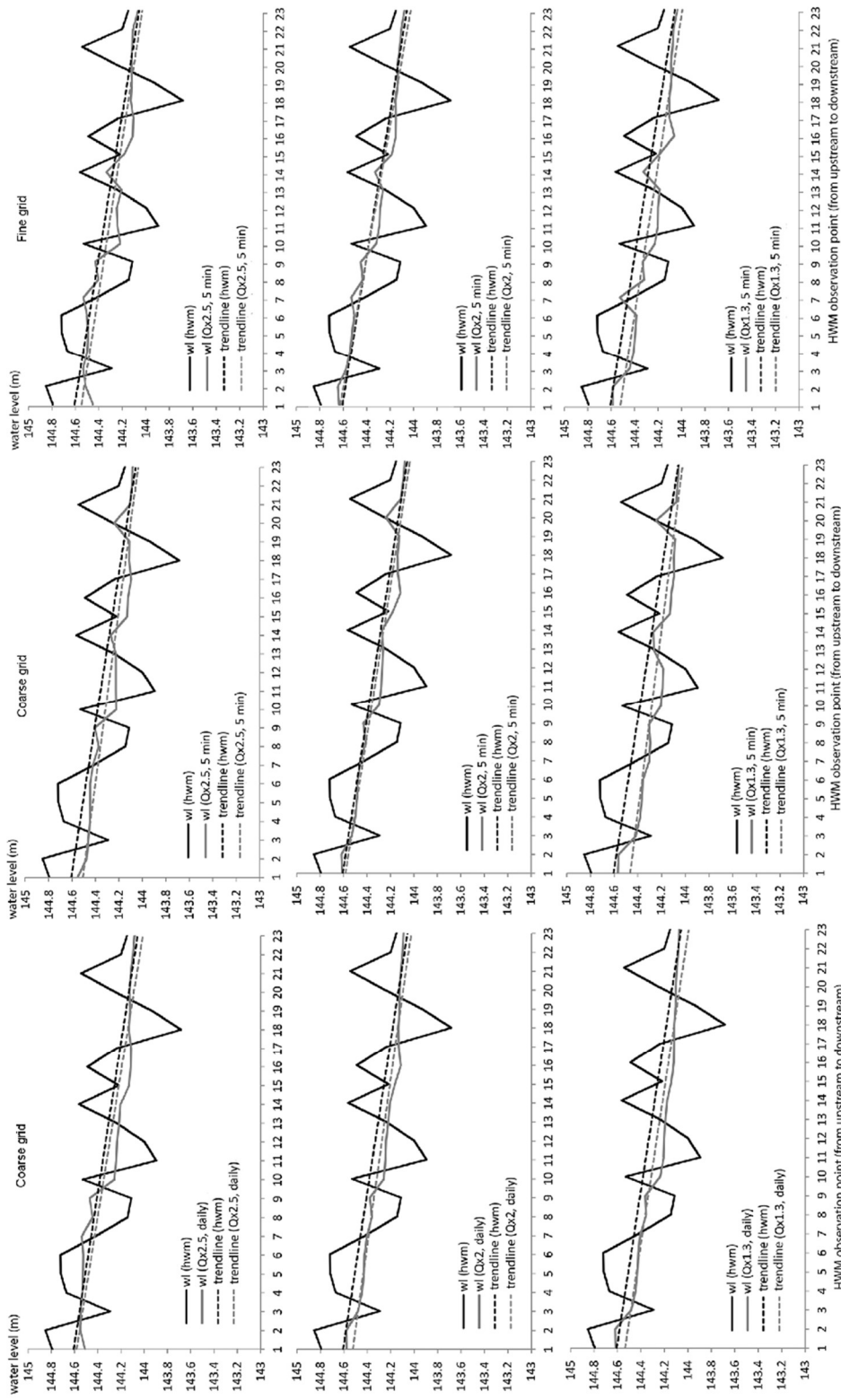
### 3 Morphodynamic simulation approach

#### 3.1 Boundary conditions: discharges and water levels

Two flow hydrographs recorded in 2013 were defined as the upstream boundary condition and the water levels as the downstream boundary condition for the hydro- and morphodynamic model (2D module of Delft3D-flow). As no direct measurements were possible to do at the study reach during the discharge events, such as normally has been done in the case of perennial rivers (Williams et al., 2013), the input data and calibration procedures differ from the traditional ones done for simulating perennial rivers. At the study reach, the hydrograph peak discharge was estimated from continuous lines of flotsam (i.e. high-water marks, HWM) emplaced by the floodwater. The hydrograph shape of the study site can be assumed similar to Vall d'Alba gauge station, due to the ~~widespread the~~ widespread continuous character of the rain events. The hydrograph shape was verified to be similar to Vall d'Alba observation station also by recent installation of water level sensors (in late 2014) in the study site (see supplementary material). The high-water marks HWM (HWMs) left by the March 2013 discharge event provided evidence that the peak discharge was greater at the study site than that measured at Vall d'Alba (Figs. 1 and 3). Due to the absence of direct discharge measurements at the study site, discharges recorded every 5 minutes at the gauge station of Vall d'Alba (Located 18 km upstream from the study site; cf. Fig. 1) were used. Also daily discharges were calculated from these 5 minutes measurements. The hydrographs were re-scaled by using different multipliers to match the peak discharge calculated during the calibration procedure of the model (see Sect. 3.4 below, and from Calle et al., 2015). ~~The high-water marks (HWMs) left by the March 2013 discharge event provided evidence that the peak discharge was greater at the study site than that measured at Vall d'Alba. Thus, the discharges for the study site were re-scaled by using different multipliers to match the peak discharge calculated during the calibration procedure of the model (see Sect. 3.4 below, and from Calle et al., 2015).~~







**Figure 3: The hydrodynamic calibration results based on the coarse grid (1.51–5.31 m cells) and fine grid (0.76–3.03 m cells). In this figure, the high-water marks (HWM) and simulated water levels are presented. In addition, the trendlines fitted along these measurements and simulation results are shown. These calibration results of the fine grid were previously presented in Calle et al. (2015) by the same authors. The Qx1.3, Qx2 and Qx2.5 refer to the discharges scaled to match the peak discharge at the study site. Daily and 5 min refer to the observation interval of discharges at Vall d’Alba station. wl = water level based on HWMs and simulations, trendline = the trendline fitted in the measured HWMs and the simulation results. The location of these HWM measurement points have been shown in the Fig. 1.**

Figure 3: The hydrodynamic calibration results based on the coarse grid (1.51–5.31 m cells) and fine grid (0.76–3.03 m cells). These calibration results of the fine grid were previously presented in Calle et al. (2015) by the same authors.

Water level (hereafter WL) was defined, based on the actual ~~HWM high water marks (HWM)~~ measurements along the longitudinal profile (Calle et al., 2015). The HWM measurements varied slightly along the longitudinal profile and the trendline, which was fitted between these ~~field~~ measurements, showed ~~anthe- accurate water surfaceWL slope-of the water surface-better(WL)~~. ~~Note that this ese HWM water surface elevation is taken~~ were measured at the valley side and it mightay  
5 ~~not to-have reflected the actual-water levelstage across the whole channel as bed changes mayight have locally influenced water level. Therefore, the trendline constituted a good indicative of water slope and level.~~ The downstream-most value of this WL trendline was applied as the peak WL value of March 2013 discharge event (Fig. 3, and see also Sect. 3.4). The changes in WL over time were calculated ~~for both March and May 2013 events~~ based on the relative change (%) of the discharge between each discharge measurement time interval. ~~The WL curves for both the March and May 2013 discharges were defined~~  
10 ~~and used as the downstream boundary conditions~~

### 3.2 Channel geometry

The channel topography was measured, using mobile laser scanning (MLS) and RTK-GPS (Calle et al. 2015). ~~These three~~ topography data sets were applied. ~~These were as~~ 1) initial topography (MLS, March 2012), 2) calibration topography between the floods (RTK-GPS, March 2013), 3) final ~~calibration-validation~~ topography after the floods (MLS, June 2013). The MLS was performed in March 2012 (pre-flood situation) and in June 2013 (after the two discharge events of spring 2013) (Calle et al., 2015 and Table 1). The MLS was performed, both with a backpack and on a platform mounted on a 4x4 vehicle. A detailed description of the MLS data set can be found in Calle et al. (2015) and Table 1. According to Calle et al. (2015) and Kukko et al. (2015) the RMSE in elevation was 18.2 mm, and it was 36 mm for the 3D position of the targets. In addition, the edges of the bar lobes, which had been moved by the March 2013 discharge event, were measured with RTK-GPS just after the discharge event in March 2013 (Trimble 4700, 1 Hz, accuracies specified by the manufacturer: XY 0.02 m, Z 0.05 m, absolute systematic measurement error  $\leq 0.054$  m). ~~Thus, these error estimates of the topographic data had been performed in Calle et al. (2015) and therefore are not anymore presented in this paper.~~ This RTK-GPS data ~~thus~~ represented the geometry after the March 2013 event and before the May 2013 event, and ~~were as therefore therefore also~~ applied as a calibration data set. This enabled better estimation of the model performance than only using one ~~calibration~~ data set after the two consecutive flood events.

Table 1. The properties of the MLS measurements of 2012 and 2013 (adapted from Calle et al. 2015).

	2012 March	2013 June
Laser sensor / GPS sensor	FARO Photon 120	FARO Focus 3D 120 S
Receiver	NovAtel DL4+	NovAtel Flexpak6
Inertial measurement unit	Honeywell HG1700 AG11	Northrop-Grumman UIMU-LCI
IMU frequency	100 Hz	200 Hz
Referencing system	GPS (1 Hz)	GNSS (1 Hz)
Wavelength	785 nm	905 nm
Scan frequency	61 Hz	95 Hz
Point acquisition	488 Hz	488 Hz
Angular resolution	0.045° (0.8 mrad)	0.07° (1.2 mrad)
3D RMSE	0.034m	0.019 m

The ~~2D implementation of~~ Delft-~~2D~~ morphodynamic model requireds channel geometry in grid format. Orthogonalized curvilinear grids of two different resolutions were created from both measured topographies, one with “coarse” 1.51–5.31 m cells and one with “fine” 0.76–3.03 m cells (i.e. circa half of the coarser grid cell sizes). These cell sizes were selected for testing the impacts of cell sizes on simulation results, but also due to their computational effectiveness. Cell ~~sizes smaller~~ ~~finer~~ than ~~the “fine” resolution (i.e. 0.76–3.03 m)~~ did not ~~make more difference to enhance~~ the results, ~~thus no further finer cell sizes were included in this paper. Furthermore, and cells finer than 0.76–3.03 m those would have also only highly~~ increased the computational time. Thus, curvilinear grids of two resolutions, “coarse” 1.51–5.31 m cells and “fine” 0.76–3.03 m cells, were created from the topography measurement times.

The initial input channel topography for all of the ~~Delft 2D~~ simulations was defined by adding the MSLS 2012 measurement points to the curvilinear grids, both to the coarse and fine grids (~~see above~~), and averaging the point values for the grid cells. There was also available a digital elevation model (DEM) from 2009 (from airborne ~~LiDAR~~ laser scanning), which ~~hasd~~ a 1 m resolution. After adding the MLS data, this 2009 DEM was added to the grid cells, which located outside the laser scanning perimeter, in order to cover the higher banks. The coarser resolution of the 2009 DEM did not affect the simulation results of the channel, because the high-water levels of the 2013 spring events barely reached these higher bank elevations.

For calibrating and validating the modelling results, the topographies were also created from the March 2013 and June 2013 measurements so that those geometries had similar resolution as the initial input channel geometry. The model outputs after each of the discharge events were then compared to these topography grids created from the March and June 2013 measurements. The grid-form geometry of March 2013, i.e. calibration topography between the floods, was created by adding only the GPS measurements to the grid cells, and the rest of the channel was excluded from the area to be analyzed during calibration (Fig. 1). The grid-form geometry of June 2013, i.e. the final ~~calibration-validation~~ topography after the ~~two-May 2013 flood~~ floods, was defined by first adding ~~first~~ the laser scanning data set, which had been processed to include both backpack and car MLS data, to the grid cells. Then the 2009 DEM was added to the grid to cover the higher bank areas. However, only the same area, as applied in the case of March 2013 (i.e. the gravel bar lobes area), was used in the ~~final analyses of morphological change detection and comparison~~ validation of the model performance. These observed elevation and volumetric changes between the events were compared to the simulated changes. For calculating the volumetric changes the curvilinear grid topographies were needed to convert into regular grids. To minimize the errors, a 0.5 m regular grid's cell size was selected, as the original cells were mostly divisible by that value.

~~These calibration curvilinear grids of “between floods” and “after floods” situations were used for analysing the capabilities of the model to result correct channel bed elevations. These simulated and measured bed elevations were compared. For volumetric calculations, these curvilinear grids and the model outputs from the same time steps were also converted into regular grids. To minimize the errors, a 0.5 m regular grid's cell size was selected, as the original cells were mostly divisible by that value.~~

### 3.3 Grain sizes

Spatially varying grain sizes were measured from the gravel lobes area prior to the first flood and between the floods, using the Wolman (1954) sampling method (Table 2). However, this method did not recover the differences between upper layer and sublayer sediment distribution and what kind of particles the 2013 discharge events had ~~in reality~~ moved. ~~We were able to measure these only after the discharge events had occurred.~~ Therefore, the gravel moved by the 2013 spring events were measured using a US standard gravelometer (US SAH-97 handheld particle analyzer) in summer 2014 (Table 2). This was



possible, as no discharge had occurred between May 2013 and summer 2014. These measurements represented different active forms, from bars to the channel bed, which had evolved during the 2013 spring floods. The bulk sieving measurements (c. 80 kg from 1.1x1.1 m area, 0–10 cm from the surface) were performed at six different locations, and from both the upper layer (UP1–6) and sublayer (SUB1–6, c. more than 10 cm below the surface) to evaluate armouring (Fig. 1 and Table 2). The upper layer-sublayer contact was established following the criteria described by Bunte and Abt (2001) considering the size of the largest particles embedded depth. The difference in the average D50 values of upper layer samples between 2012 (18.5 mm) and 2014 (21.1 mm) measurements was only 1.6 mm (Table 2).

Table 2. The grain sizes measured in 2012–2014 (see their locations from Fig. 1). WCM, WCU and WB1–7 were measured in 2012, WB8 and 9 were measured in March 2013, and SUB1–6 and UP 1–6 were measured in June 2014. W= Wolman, C= channel, B=Bar, SUB=sublayer, UP=upper layer. Most of the samples were within the analyzed area calibration area, and their locations can be seen in Fig. 1 and also in Figure 11 of Calle et al. (2015). Only the samples UP/SUB 5 and 6 were further in the upstream section of the simulation area. Due to the spatial scarcity of the 2013 measurements (only two samples) were not applied as the surface grain sizes for the model, but were used for Limerinos (1970) calculations. The armour ratio was calculated following Lisle and Madej (1992), i.e. ratio of the surface-to-subsurface D<sub>50</sub>.

<b>Sample</b>	<i>upper layer</i>		<i>sub layer</i>	
	<i>D50 (mm)</i>	<i>D84 (mm)</i>	<i>D50 (mm)</i>	<i>armour ratio</i>
WCM	18	29	-	-
WCU	30	101	-	-
WB1	22	41	-	-
WB2	17	33	-	-
WB3	17	34	-	-
WB5	12	21	-	-
WB6	17	37	-	-
WB7	14	22	-	-
WB8	18	35	-	-
WB9	20	31	-	-
UP1 and SUB1	22.5	-39	18.5	1.216
UP2 and SUB2	35.5	-79	21.3	1.667
UP3 and SUB3	18.8	-31	14.3	1.314
UP4 and SUB4	18.0	30.5-	14.4	1.250
UP5 and SUB5	31.2	55.4-	18.6	1.677
UP6 and SUB6	41.2	80-	16.7	2.467
Average UP1–6	26.3	-	-	-
Average SUB1–6	17.1	-	-	-
Average UP and SUB1–6	21.1	-	-	-

The D<sub>50</sub> and D<sub>84</sub> grain sizes of 2012 and March 2013 were first used for the calibration and initial testing of the hydrodynamic model (see Sect. 3.4). For the calibration of the morphodynamic model, the spatial distribution of the D<sub>50</sub> grain sizes of both 2012 and 2014 were then applied. The grain size values were assigned to equivalent morphological elements defined in the geomorphological map. These grain sizes were transferred to each cell of the curvilinear morphodynamic model's grid.

Different input grain size distributions were applied in the model tests: 1) spatially varying upper layer grain sizes, 2) spatially varying sublayer grain sizes and 3) constant average grain sizes (average of upper layer, sublayer or both upper layer and sublayer). The values were applied to the whole active layer of the river bed. The D50 grain size values were used in the model.

### 3.4 Hydrodynamic simulations

- 5 The same calibrated hydrodynamic model (2D ~~implementation~~version of the Delft3D-FLOW) was applied as in Calle et al. (2015). The flow was simulated as depth-averaged flow by using the well-known Navier-Stokes and shallow-water equations. Thus, the modelled fluid was considered vertically homogeneous. These well-known equations ~~can be were~~ found in the user manual of the model (Deltares, 2011). ~~The model applies the alternating direction implicit (ADI) time integration method for the shallow water equations (Leendertse, 1967).~~

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Table 3. The Manning's n-values and discharges that gave the best simulation results when compared to the measured HWMs and their trend line during the calibration. Qx1.3, Qx2 and Qx2.5 stand for Vall D'Alba's discharges multiplied by 1.3, 2 and 2.5, respectively. This same table is presented in Calle et al. (2015).

Element type	<u>n (Qx1.3)</u>	<u>n (Qx2)</u>	<u>n (Qx2.5)</u>
Channel/active bar	<u>0.063–0.065</u>	0.05–0.053	<u>0.03–0.04</u>
Extract area	<u>0.07</u>	0.06	<u>0.03</u>
Block	<u>0.068</u>	0.058	<u>0.05</u>
Exposed area	<u>0.07</u>	0.06	<u>0.033–0.04</u>
Scarce vegetation bar	<u>0.065</u>	0.055	<u>0.035–0.04</u>
Vegetated bar	<u>0.07</u>	0.06	<u>0.04</u>
Pleistocene terrace	<u>0.07</u>	0.06	<u>0.05</u>
Rocky bank	<u>0.07</u>	0.06	<u>0.035</u>
Vegetated bank	<u>0.07</u>	0.06	<u>0.05</u>
Opencast mine	<u>0.07</u>	0.06	<u>0.04</u>

- 15 The hydrodynamic model was calibrated by adjusting the Manning's n-values so that the simulated water levels matched with the measured high-water level marks, and their WL trendline, of the 6<sup>th</sup> of March 2013~~6.3.2013~~-discharge peak situation (Fig. 3 and Calle et al., 2015). The hydrodynamic model was calibrated by applying first daily discharge values as the upstream boundary condition. The n-values were defined based on Limerinos (1970) equations for D<sub>50</sub> and D<sub>84</sub> grain sizes (Table 3). It is noteworthy that vegetation was almost non-existent in the channel. There was only scarce vegetation in the upstream part,
- 20 and higher up in the banks, i.e. out of reach of the 2013 discharge events. Therefore, the roughness values were possible to define by the particle size on the active river channel. ~~It is noteworthy that there was no vegetation in the river channel area affecting the two (moderate and small) discharge events, and therefore the roughness values were possible to define by the particle size on the active river channel.~~ The Limerinos equation was applied for the whole range of water levels (i.e. hydraulic radiuses) for the March 2012 and March 2013 grain sizes of D<sub>50</sub> and D<sub>84</sub>. These represented the best of the preceding conditions
- 25 of the March 2013 and May 2013 flows. The Limerinos calculations were performed for a cross-section located at the

downstream side of the simulation area. This cross-section was defined from 2012 geometry, i.e. pre-flood geometry, which remained unchanged during the 2013 discharge events. At the 144.151 m water level, which was the measured high-water mark elevation at this location, the wetted perimeter of the cross-section was 52.09 m and the hydraulic radius was 0.95 m (max flow depth 1.28 m). Each of the geomorphological elements, and thus also grid cells, received their own Manning's n-value during the calibration procedure. Note that the assignment of the n values was static throughout the simulations, i.e. the spatial distribution of the roughness values was not possible to change between simulation time steps.

The calibration was first done for the coarser grid of 1.51–5.31 m cell sizes. Because the discharge estimated at the study site was greater than the one recorded at the Vall d'Alba gauge station, three different daily discharges ("Qx1.3", "Qx2" and "Qx2.5") were tested during the calibration of the hydrodynamic model. The "Qx1.3" was defined based on the increase in watershed area (30%) between the study area and the Vall d'Alba observation station. This Qx1.3 discharge simulation matched the high-water-level HWMs-marks using the highest n-values calculated with the Limerinos 1970 equation, i.e. for shallow flow (Table 3). In this Qx1.3 discharge, each of the observations was multiplied by 1.3. The "Qx2" (observations multiplied by 2) water surface elevation matched the HWMs using an average n-values calculated from Limerinos (1970) equations. Also, the effects of bedform (from +0.01 to +0.015) and bank roughness (+0.02) had been added to these average n-values (Chow, 1959; Acrement and Schneider, 1989). The simulated water levels also matched with HWMs when the "Qx2.5" discharge and n-values representing the high-flow stage values (i.e. lower n-values on the value range) calculated from Limerinos (1970) equations were selected. Similar n-value ranges were gained for different water level situations, when either the equation for D<sub>84</sub> or D<sub>50</sub> grain sizes was applied. Based on the Limerinos equation calculations, the Qx2 was expected to work the best due to its average nature and inclusion of bedform and bank roughness effects.

When the calibration against the HWMs and their trendline was successful with these daily discharges and coarse grid, the same Manning's n-values were applied for simulations with a hydrograph with a 5-minute measurement interval. The "Qx1.3", "Qx2" and "Qx2.5" versions of these 5-minute hydrographs were tested. These simulations also reproduced the water levels, and the best fit was obtained with "Qx2" (Table 3) sufficiently well, and, thus, no more modifications-adjustments were performed-done for the Manning's n-values. All in all, the best fit was obtained with "Qx2" (Table 3). Finally, the 5 min hydrograph simulations were performed simulations were also performed with the more detailed grid, i.e. the fine grid with (0.76–3.03 m cells, (Fig. 3). These simulation hydrodynamic simulation results corresponded the observations with the fine grid cells were better than with-if the coarse grid cell sizes were used (as also shown in Calle et al., (2015).

In addition Finally, both March and May 2013 discharge events were tested simulated in the final hydrodynamic simulations. Similar multipliers (Qx1.3, Qx2, Qx2.5) and Manning values were applied for both events. Both these discharge hydrographs events of 2013 were simulated as an unsteady simulation-flow and by applying the bathymetry derived from 2012 topographical measurements. The Qx2 discharge was proven to best produce the observed hydrodynamics (Fig. 3; see also Calle et al., 2015).

All in all, 21 hydrodynamic simulations were run during the calibration of the hydrodynamic model.

### 3.5 Morphodynamic simulations

Calibration of the morphodynamic model consisted on the adjustment of the model parameterization so that the model output geometries matched the first-order measured bedform geometries as well as possible. The model applied an “upwind” bed update scheme, where the elevation of the bed was dynamically updated at each computational time step. In all simulations, the initial boundary conditions of the input and output sediment transport load amount were defined as 0. This is because the flood events started from 0, and no significant transport, i.e. evolution of bedforms was observed upstream and downstream of the simulation area. The model then calculated the transport based on the selected transport equations and using equilibrium concentration for carrying input sand sediment fractions. This was similar to earlier studies done in gravel bed river (Williams et al., 2016b), and was selected based on earlier experiences of the river (Calle et al., 2015), and because no suspended load measurements were available and the sand sediment fraction was almost non-existent in the river. Based on the experience in the study site, the load should be in equilibrium particularly in erosional areas of the study site. The simulation area began further upstream than the actual analyzed area, so that sediment transport input in the study reach was steady. Because the bed level gradient affects the bedload transport, the slope in the initial direction of the transport (referred to as the longitudinal bed slope) and the slope in the direction perpendicular to that (referred to as the transverse bed slope) were utilised. The transverse slope affects transport towards the downslope direction (Deltares, 2011). The Bagnold (1966) equation was applied for the longitudinal slope and Ikeda (1982), as presented by Van Rijn (1993), was applied for the transverse slope.

~~Due to the better performance with Qx2, all these selected morphodynamic simulations have these input data. All simulations with Qx1.3 and Qx2.5 discharges were discarded as those did not reproduce the correct topographical changes. The 11 selected simulations showed the effects of grain size (before [2012] and after [2014] floods, and grain sizes from different layers [2014]), grid size (coarse: 1.51–5.31 m, fine: 0.76–3.03 m), transverse slope (user defined coefficients in the bed load transport equations: default 1.5 and increased to 3) and transportation equations (Engelund-Hansen [EH], Meyer-Peter and Müller [MP]) on model performance (Tables 4 and 5). These parameter tests were selected for the calibration procedure, as these have earlier been found important for Delft 2D model simulations, albeit in perennial rivers (e.g. Kasvi et al., 2014). The morphodynamic simulation results were compared to the measured topographies (Figures 4–6), in particularly the volumetric change of river bed and displacement of the lobe front (Table 6).~~

Table 4. The morphodynamic simulations and applied parameters. The fine grid size is 0.76–3.03 and the coarser size is 1.51–5.31 m. EH=Engelund-Hansen, MPM=Meyer-Peter and Müller. Events: 1=only the March 2013 event was simulated, 2=both the March and May 2013 events have been simulated. The simulations that were selected for the hourly channel change analyses are **bolded in bold**.

Simulation	Transport equation	Discharge events	transverse slope	grid size	grain size
1	EH	1	1.5	coarse	varying, 2012

2	EH	1	1.5	fine	varying, 2012
3	EH	1	1.5	coarse	varying, upper, 2014
4	EH	1	1.5	coarse	varying, sub, 2014
5	EH	1	1.5	coarse	varying, average upper+sub, 2014
6	EH	1	1.5	coarse	constant, average upper, 2014
7	EH	1	1.5	coarse	constant, average sub, 2014
8	EH	1	1.5	coarse	constant, average upper+sub, 2014
9	<b>EH</b>	<b>2</b>	<b>3</b>	<b>fine</b>	<b>varying, upper, 2014</b>
10	<b>EH</b>	<b>2</b>	<b>1.5</b>	<b>fine</b>	<b>varying, upper, 2014</b>
11	MPM	2	1.5	fine	varying, upper, 2014

Altogether, 61 morphodynamic simulations were needed for calibrating the morphodynamic model. These 61 simulations included simulations with Qx1.3, Qx2 and Qx2.5 discharges hydrographs. During these simulations the match of the simulated water levels to HWMs was also checked. The roughness values were still valid. While performing the calibration, the sensitivity of the model to the different parameter combination was analyzed.

Out of these 61 morphodynamic simulations, 11 simulations done during calibration, which produced the best and most interesting results, were selected to be presented in the detail in this supplementary material of this paper for sensitivity analyses (see also Tables 4 and 5). Reason for their selection was that these 11 selected simulations showed the effects of grain size (before [2012] and after [2014] floods, and grain sizes from different layers [2014]), grid size (coarse: 1.51–5.31 m, fine: 0.76–3.03 m), transverse slope (user defined coefficients in the bed load transport equations: default 1.5 and increased to 3) and transportation equations (Engelund-Hansen [EH], Meyer-Peter and Müller [MPM]) on model performance (Table 4, and supplementary material Tables 4 and 5). These parameter tests were selected for the calibration procedure, as these had earlier been found important for Delft3D (2D implementation) model simulations, albeit in perennial rivers (e.g. Kasvi et al., 2014). The morphodynamic simulation results were compared to the measured topographies within the calibration area, in particular to the volumetric change of river bed and displacement of the lobe front. Due to the better correspondence of simulations with Qx2 discharges to the observed channel evolution, all these selected morphodynamic simulations had these input data. The longitudinal slope did not affect the results, and the default value 1.0 was selected to be used. The model was first calibrated with March 2013 discharge event (simulations 1–11), and then validated with May 2013 discharge event (simulations 9–11).

best Reid et al. (1996) showed that the Meyer Peter and Müller (MPM, 1948) equation, which was originally developed for steady flow conditions and used by Reid et al. (1996) in non-armoured gravel bars, performs well and better than the Bagnold (1980) and Parker (1990) equations in ephemeral gravel river channels.

Because the bed level gradient affects the bedload transport, the slope in the initial direction of the transport (referred to as the longitudinal bed slope) and the slope in the direction perpendicular to that (referred to as the transverse bed slope) were utilised. The transverse slope affects transport towards the downslope direction (Deltares, 2011). The Bagnold (1966) equation was applied for the longitudinal slope and Ikeda (1982), as presented by Van Rijn (1993), was applied for the transverse slope. The

longitudinal slope was tested with both the default value (1.0) and double the default value (2.0). However, the longitudinal slope did not affect the results, and it was decided that the default value 1.0 would be used in the selected simulations, which are shown in Tables 4–6. Due to the great effects caused by the transverse slope on the morphodynamic simulations of perennial rivers (e.g. Kasvi et al., 2014), we were particularly interested in the effects of this transverse transport component on the ephemeral river simulation, and thus tested the model with the component being 3.0, which is double the default value (1.5). These values were defined based on earlier publications (e.g. Kasvi et al., 2014).

Table 5. The simulations used for comparing the effects of different parameters on model performance. x = parameter tested in the simulation.

Simulation	Tested parameters					
	grid-size effect	grain-size effect: before (2012) & after flood (2014)	grain-size effect: different layers	grain-size effect: constant vs. varying	transverse slope effect	transportation equation effect
1	x	x				
2	x	x				
3	x	x	x			
4			x			
5			x	x		
6			x			
7			x			
8			x	x		
9					x	
10	x	x			x	x
11						x

The bedload transport was tested initially with the Van Rijn 1993 equation (VR), which is widely used, despite the fact that it was developed with smaller grain sizes than those of the study area. Also, small flow depths may affect the performance of the VR equation (Van Rijn, 1984). The depths of the flow were relatively shallow at the Rambla de la Viuda, in comparison to the channel width. As expected, the channel bed did not evolve at all with VR equation. Thus the VR equation was discarded at once (not included in Tables 4 and 5). Instead, the widely applied equations of “Engelund Hansen 1967” (EH) and “Meyer-Peter and Müller 1948” (MPM) were selected from the Delft 2D model’s assortment of equations (cf. Deltares, 2011), because they had been developed by using initially non-armoured bed conditions and were the most appropriate for the study area, according to the  $D_{50}$  grain sizes. Also, MPM has previously been proven to perform well, i.e. better than, for example, Bagnold 1980, in flash flood simulations (Reid et al., 1996; Cao et al., 2010). The MPM was expected to perform the best, as it has originally been developed for particles of 0.4–29 mm overall diameters. Earlier simulations done for ephemeral rivers, e.g. Hooke et al (2005), have applied the Bagnold (1966) total load equation. However, it was not included in the set of equations provided by the standard Delft 2D model, and thus it was not applied in this study. Instead, the EH was selected as the total load equation, due to its proven performance in a variety of environments (e.g. Kasvi et al. 2014), even though it has originally been tested on up to 0.93 mm median particle sizes (Engelund and Hansen, 1967).



Two simulations, which show after calibration the best correspondence with the observed riverbed changes and bed formations, were selected for the final channel evolution analysis. Based on these two simulations, the temporal evolution of the riverbed was analyzed during the two hydrographs of moderate and low magnitude discharge events.

#### 4 Morphodynamic simulation results

##### 4.1 The morphodynamic model's sensitivity to the parameterization

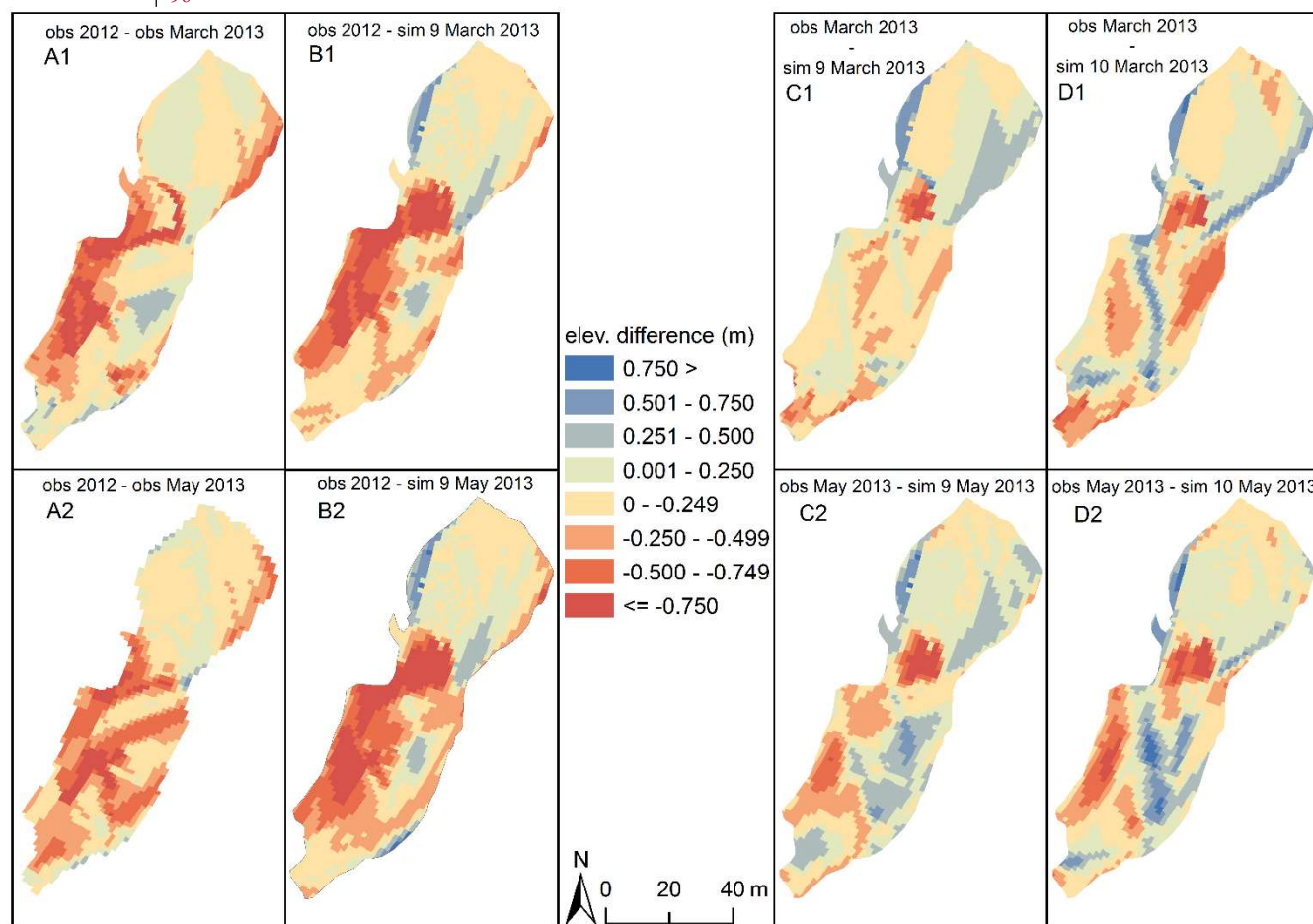
###### 4.1.1 Effects of applied grain size on simulation results

The grain size and its spatial variation affected the model results greatly and needed the largest number of tests during the calibration of the model (Tables 4–6, Figures 4–6). This was first analyzed by comparing the measured and simulated topographies and volumetric changes produced by the March 2013 flow. When the spatially varying upper layer sediments (representing the sediments actually moved by the two discharge events, i.e. the ones measured after floods in 2014) were used in the simulation (number 10), the volumetric changes were less and fitted the observations better than if spatially varying “before floods” grain sizes of 2012 were applied (simulation 2). The effects of spatially varying sublayer grain sizes (simulation 4) and spatially varying upper layer grain sizes (simulation 3) were compared: the deposition increased by 352 m<sup>3</sup> when applying spatially varying sublayer grain sizes. The bedforms were also represented better with upper layer sediments. When the spatially varying average grain sizes (average of the upper and sublayer grain sizes) were applied, the simulation (number 5) resulted in a slightly greater deposition than with spatially varying upper layer grain sizes (simulation 3, Table 6). When the constant grain sizes were applied (simulation 8), the deposition was greater and resembled the observations better than a corresponding simulation with spatially varying grain sizes (simulation 5), but the erosion was also greater and thalweg was excavated too deep with constant grain sizes (simulations 6–8). Altogether, the spatially varying upper layer grain sizes (measured in 2014), which were known to have been moved by the discharge events, resulted in the most realistic results.

Table 6. The evaluation of modelled volumetric changes and lobe movement, in comparison to observations. Based on the performance analyses of the model during March 2013 event's simulations, the May 2013 discharge event was only simulated with a fine grid cell size, and using upper layer sediments measured in 2014 (cf. Table 4). The simulations that selected for the final hourly channel change analyses are bolded. Gravel storage was calculated by subtracting the erosion from deposition.

simulation	grid size	erosion (total m <sup>3</sup> ; difference compared to observed %)	deposition (total m <sup>3</sup> ; difference compared to observed %)		Gravel storage (total m <sup>3</sup> ; difference compared to observed %)	lobe movement (m)
<b>observed</b>	<b>coarse</b>	89.116; 0	1252.066; 0		1162.95; 0	58.5

observed	fine	154.564; 0	984.721; 0	830.157; 0	58.5
1	coarse	267.558; +200	1071.393; -14	803.835; -31	40.6
2	fine	286.240; +85	1332.001; +35	1045.761; +26	69.4
3	coarse	204.457; +129	878.286; -30	673.829; -42	47.9
4	coarse	155.742; +75	1230.629; -2	1074.887; -8	47.9
5	coarse	204.460; +129	992.506; -21	788.046; -32	47.9
6	coarse	210.668; +136	992.720; -21	782.052; -33	47.9
7	coarse	210.277; +136	1245.116; -1	1034.839; -11	42.5
8	coarse	185.326; +108	1106.058; -12	920.732; -21	40.23
9	fine	164.688; +7	1119.282; +14	954.594; +15	59.9
10	fine	178.686; +16	1190.846; +21	1012.16; +22	52.1
11	fine	6.498; -96	30.382; -97	23.884; -97	3.7



**Figure 46: The comparison between the observed and simulated (simulations 9 and 10) elevations after the March 2013 and May 2013 discharge events. The negative values (red) mean that the observed (A)/simulated (B: simulation 9) March (1)/May (2) 2013 topographies were higher (i.e. deposition had occurred) than was observed in 2012. The negative values also mean that simulated elevations of March/May 2013 were higher in those locations than was observed in March/May 2013 with simulations 9 (C) and 10 (D). The fine grid cell sizes were applied in these simulations. Simulation 9 corresponded slightly better the observations than simulation 10, and thus its results are shown in more detail. The pool area, in the middle of the area, was the only area where simulation results clearly showed more deposition than observed (see also Fig. 2 PU).**

Here we present the summary of calibration results, which can be found in more detail from the supplementary material of this paper. The best simulation results in relation to the surveyed volumetric changes were achieved with the fine grid simulations.

The coarser grid size overestimated incision and was therefore discarded. The grain size and its spatial variation affected the model results greatly and needed the largest number of tests (supplementary material). When the spatially varying upper layer sediments of 2014 (representing the sediments moved by the two discharge events) were used (simulation 10), the volumetric changes were less and fitted the observations better than if spatially varying 2012 “before floods” grain sizes were applied (simulation 2). The bedforms were also best represented when upper layer sediments were applied (simulations 9 and 10).

The deposition and erosion amounts resembled the observed elevations, and the lobe movement distances were the best when the transverse slope was increased (simulation 9). Despite, the best correspondence of the simulated and measured spatial bedform pattern was obtained with simulations number 9 and 10, the default (1.5) transverse slope of simulation 10 resulted in a more excavated thalweg, when compared to the simulation 9 with an increased transverse slope (Fig. 4).

Earlier simulations done for ephemeral rivers, e.g. Hooke et al (2005) and Graf (1996), had applied the Bagnold (1966) total load equation. However, Meyer-Peter and Müller (MPM, 1948) equation had previously been proven to perform well, i.e. better than, for example, Bagnold 1980 and Parker (1990) equations in flash flood simulations and in ephemeral gravel river channels (Reid et al., 1996; Cao et al., 2010). The widely applied equations of “Engelund-Hansen 1967” (EH) and “Meyer-Peter and Müller 1948” (MPM) were selected from the standard Delft3D model’s assortment of equations (cf. Deltares, 2011), because they had been developed by using initially non-armoured bed conditions and were the most appropriate for the study area, according to the  $D_{50}$  grain sizes. The MPM was expected to perform the best, as it had originally been developed for particles of 0.4–29 mm overall diameters. The EH was selected as the total load equation, due to its proven performance in a variety of environments (e.g. Kasvi et al. 2014), even though it had originally been tested on up to 0.93 mm median particle sizes (Engelund and Hansen, 1967). The transport equation had a crucial role in the simulation results. The EH (simulations 9 and 10) was superior in reproducing the channel morphology. MPM resulted in much smaller and incorrect transport values (simulation 11) during both March and May 2013 discharge events.

Thus, simulations 9 and 10 produced the best correspondence with the observed riverbed changes and bed formations (Fig. 4).

These were used for analysing the temporal riverbed evolution during the moderate- and low-magnitude discharge events.

#### 4.1.2 Effects of grid size on simulation results

The grid size effects on the simulation results were compared between simulations 1 and 2, and simulations 3 and 10 (Tables 4–6, Figures 4–6). The fine grid size tends to overestimate both erosion and deposition, whereas a coarser grid size overestimates incision. When compared to the observed topographic river bed model, the coarse grid simulations show also an underestimation in the depositional amounts. The coarse grid also caused a moderate magnitude flood to flow mostly on the left bank side, which was opposite to the fine grid simulations and reality. The best simulation results (from the simulations 1, 2, 3 and 10) in relation to the surveyed volumetric changes were achieved with the fine grid simulation (number 10). It was decided to discard the coarse grid from the simulations of the May 2013 discharge event, and from further analyses.

#### 4.1.3 Effects of transverse slope on simulation results

The increased transverse slope affected positively the simulation results (simulation 9), when they were compared to the corresponding simulations with the default 1.5 transverse slope (simulation 10) (Table 6, Figures 5–6). The deposition and erosion amounts resembled the observations more, when the transverse slope was increased (Table 6). The lobe movement distance was also the best with an increased transverse slope (simulation 9). Noticeable from the elevation difference analyses is that the simulations with an increased transverse slope resulted in the greatest correspondence to the observed elevations (Fig. 6). However, particularly one location in these simulations had experienced more deposition than observed: the pool area downstream of the large lateral bar (PU: Figure 2 and Calle et al., 2015).

When compared to the geomorphological evolution described in Calle et al. (2015) and the measured topographies, the spatial bedform pattern was visually compared for both the discharge events. Of all simulations, the best results were obtained with these simulations number 9 and 10. However, the default 1.5 transverse slope resulted in a more excavated thalweg (simulation 10), when compared to the simulation with an increased transverse slope (simulation 9) (Figures 5–6). However, simulation 10 also showed satisfying bar accretion from the left bank side of the channel.

#### 4.1.4 Effects of transport equation on simulation results

Based on the comparisons of observed and simulated topographies and volumetric changes (Table 6 and Figure 4–6), the transport equation had a crucial role in the simulation results. The EH equation (simulation 10) was superior in reproducing the channel morphology. MPM resulted in much smaller transport values (simulation 11) during both March and May 2013 discharge events (simulation 10). Basically, no movement occurred when the MPM equation was used. Thus, the MPM equation was proven not to be able to produce the correct movement during the moderate or low magnitude discharge events of this ephemeral gravel river. The EH equation (simulation 10) was superior in reproducing the channel morphology. MPM resulted in much smaller transport values (simulation 11) during both March and May 2013 discharge events (simulation 10). Basically, no movement occurred when the MPM equation was used. Thus, the MPM equation was proven not to be able to

produce the correct movement during the moderate or low magnitude discharge events of this ephemeral gravel river

## 4.2 The temporal evolution of Rambla de la Viuda

### 4.2.1 The evolution during moderate-magnitude flood hydrograph discharge event

All in all, lobe movement of c. 60 m was observed during the March 2013 discharge event, and additional c. 30 m during the May 2013 discharge event. The bedload sediment mass balance was possible to estimate. Deposition was greater than erosion based on measurements (i.e. in the calibration area) (Table 5). Hydrographs and erosion and deposition rates were also plotted against time for the simulations 9 and 10 (Fig. 6), which had reproduced the best the observed morphodynamics within the calibration area (see Sect. 3.6). Sediment erosion and deposition were calculated from the total channel bed change amount of each hour from this calibration area. In addition, the bed elevation changes between the key time steps, i.e. times of clear changes in discharge or bed evolution, were also defined from the whole simulation area (Fig. 6: simulation 9 as an example).

Table 5. The volumetric changes and lobe movement of observations and the two selected simulations (see also supplementary material). Gravel storage was calculated by subtracting the erosion from deposition.

1. event (March 2013)								
sim	grid	erosion		deposition		Gravel storage		
		total m <sup>3</sup>	difference compared to observed %	total m <sup>3</sup>	difference compared to observed %	total m <sup>3</sup>	difference compared to observed %	lobe movement (m)
	fine	154.56	0	984.72	0	830.16	0	58.5
9	fine	164.69	+7	1119.28	+14	954.59	+15	59.9
10	fine	178.69	+16	1190.85	+21	1012.16	+22	52.1
2. event (May 2013)								
obs	fine	63.67	0	1087.59	0	1023.93	0	87.1
9	fine	182.53	+187	1222.32	+12	1039.79	+2	61.3
10	fine	160.09	+151	1289.38	+19	1129.29	+10	79.5

Hydrographs and erosion and deposition rates were plotted against time for the simulations 9 and 10 (Fig. 7), which had reproduced the observed morphodynamics the best (see Sect. 4.1). Sediment erosion and deposition were calculated from the total channel bed change amount of each hour. The bed elevations from the key time steps, i.e. times of clear changes in discharge or bed evolution, are also shown (Fig. 8: simulation 9 as an example). Sediment erosion and deposition were calculated from the total channel bed change amount of each hour. During this moderate flow event, the channel evolved and bar lobes advanced on both sides of the large lateral bar, due to the diverted flows (B1 location of Fig. 2, and Figs. 4–6 and 8). However, the main flow and bar lobe movement occurred on the right bank side of the channel and diagonal bar aggradation took place downstream part of the study site.

Both simulations experienced more deposition than erosion during March 2013 event (Table 5 and Fig. 5). The greatest total erosion and deposition (m<sup>3</sup>) occurred at the beginning of this discharge event (Fig. 75). However, the initial changes were

slightly smaller when the default transverse slope was applied (simulation 10). The peak of channel bed changes occurred at the beginning of the moderate-magnitude discharge event, i.e. before the discharge peak. This implies to possible positive hysteresis phenomenon. Significant changes were observed in small peaks at 9h and 71h, which give an idea of the high system response to steep changes in the discharge, rather than a progressive increase (as seen in the absolute peak of 23h). The deposition in simulations 9 and 10 started very slowly decreasing after 13 hours, but the decrease in erosion was slightly faster. However, all in all, the deposition and erosion remained high for a long time and only started clearly declining after 71 hours, i.e. the secondary discharge peak. The erosion and deposition volumes followed the changes in receding discharge more faithfully than during the rising phase, occurring simultaneously to the changes in discharge. Thus, the morphodynamics of both simulations 9 and 10 followed the changes in discharge better during the falling stages of the discharge peak, i.e. when the discharge was  $30 \text{ m}^3 \text{ s}^{-1}$  or less (Fig. 5). During the rising phase, there were temporal differences between changes in discharge and changes in deposition and erosion. When the default transverse slope was applied (simulation 10), the erosion and deposition continued to be high longer in the receding phase than when the transverse slope was increased to 3 (simulation 9). The great deposition amount during the receding phase was mainly due to the propagation of the bar lobe front.

When detecting the channel changes spatially during this March 2013 moderate flow event from the whole simulation area (based on simulation 9, as an example), the channel evolved and bar lobes advanced on both sides of the large lateral bar, due to the diverted flows (B1 location of Figs. 1 and -2, and B in Fig. -Figs. 4-6 and 86). However, the main flow and bar lobe movement occurred on the right bank side of the channel (B in Fig. 6) and diagonal bar aggradation took place downstream part of the study site. Both topographical observations and model simulations showed the development of the diagonal bar alongside with the cutoff of the bar (B). The model results showed that the initial cutoff and simultaneous initiation of the diagonal bar took place during the rising limb, more precisely the couple of first hours of the flood event (Fig. 6). However, during these first hours of the flood event, the changes in the river bed were local, and greatest elevation changes occurred especially in the bar lobe area, and right upstream of the large lateral bar B1 (A location in Fig. 6). Noteworthy is that the diagonal bar would not have developed to its full extent without the long receding phase of the flood hydrograph. Spatially, the most changes occurred throughout the whole simulation area during the 24 hours following the discharge peak, i.e. between 6<sup>th</sup> and 7<sup>th</sup> of March. The diagonal bar formation was slightly greater in the model outcomes than based on observations. Despite this, the model showed potential in producing the channel development following the established theories of gravel bed evolution. Similarly to the diagonal bar movement in the downstream part of the study site (B, Fig. 6), also further upstream another lateral bar (A, Fig. 6) experienced excavation on the right bank side, and the propagation of lobe movement downstream particularly after the discharge peak. Thus, throughout the study area the initial hours of the channel changes caused the selection of the flow and sediment transport routes where the most erosion would take place later during the moderate-magnitude discharge event.

In addition to the braiding structured display of the channel gravel bars, i.e. alternating bars and pools, the river channel has also

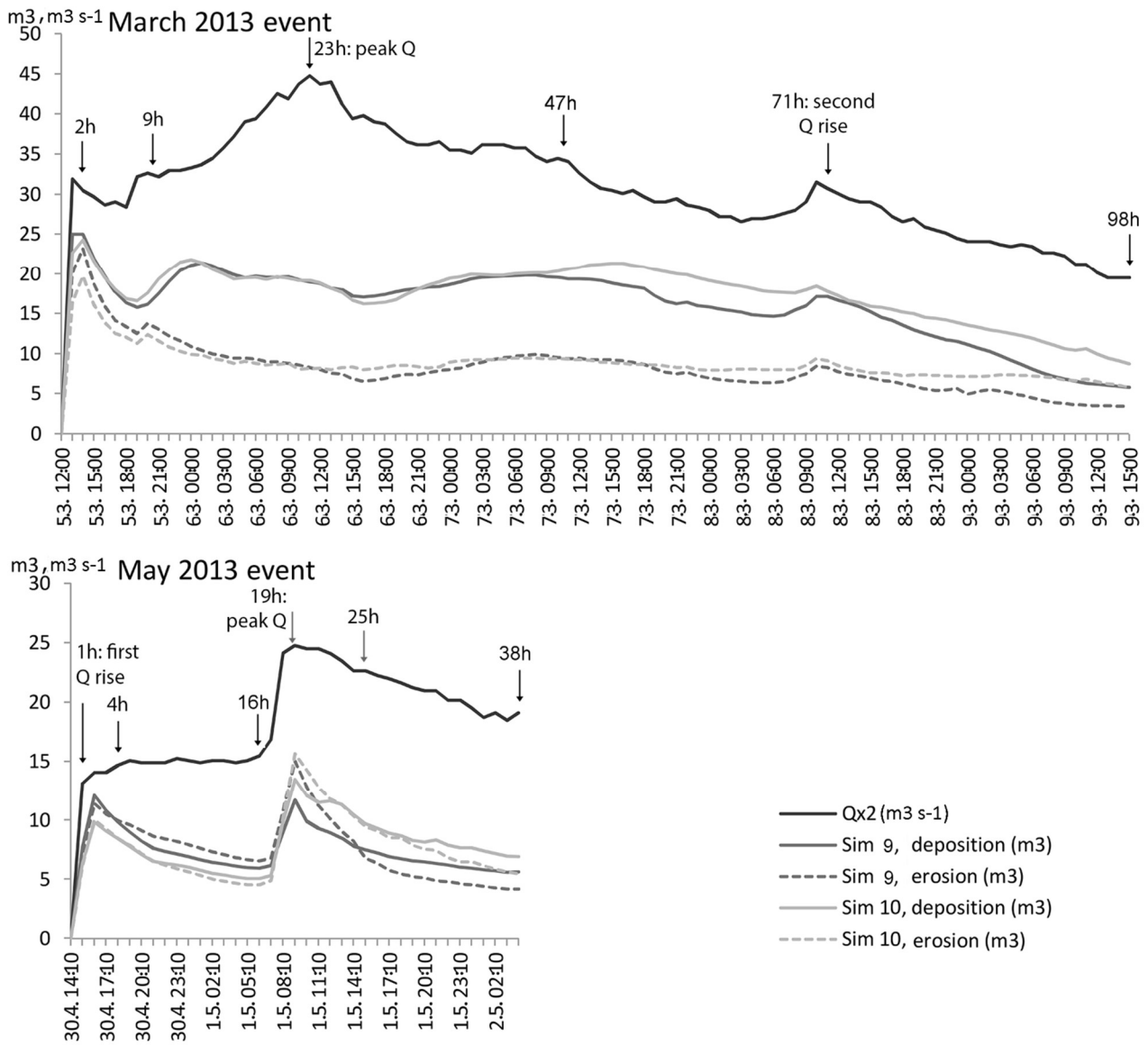


a broader meandering planform. The simulated area can be considered as one bend. The downstream lobe area locates downstream of the bend's apex, and the greatest erosion occurred during the lowering flood phase on the right bank side, which is the inner bank of the bend at the inlet area to the bend (A). This followed also the results of other meandering river studies about the erosion locations (Lotsari et al., 2014b). Thus, during the higher flow stages the flow routing and channel started acting more like in a meandering river, but during lower flow stage, such as during the initial rising flood stages, when not all of the bars were covered with water, the fluvial processes and related channel changes resembled more to braided channel development.

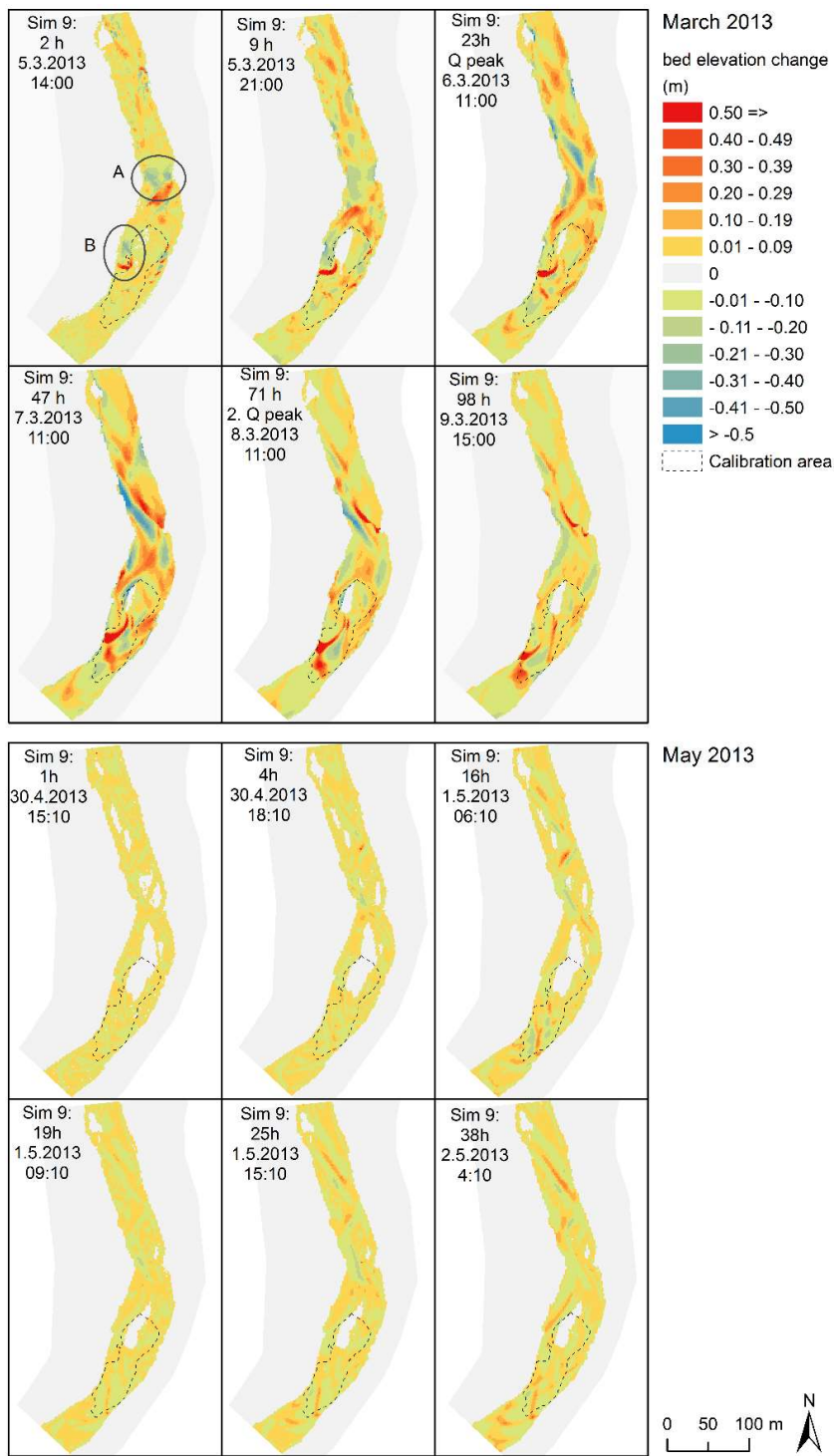
#### **4.2 The evolution during low-magnitude discharge event**

During the May 2013 discharge event, the hourly changes in erosion and deposition within the calibration area followed the discharge evolution more than during the March discharge event (Fig. 5). The discharge event of the May 2013 had two peaks, of which the latter discharge peak was greater. The erosion and deposition peak occurred approximately an hour after this greatest peak discharge had been reached. This implies to possible negative hysteresis phenomenon. Thus, the deposition and erosion peaks did not occur immediately at the beginning of the discharge event, as in the case of the March 2013 event. The erosion became greater than the deposition four hours after the beginning of the discharge event (simulation 9). Six hours after the sediment transport peak (at 25 h), the deposition dominated again. During the March discharge event, the erosion was never greater than the deposition. Note also that the discharge remained constant after both peaks, whereas a decrease in the deposition and erosion rates was observed (Fig. 5).

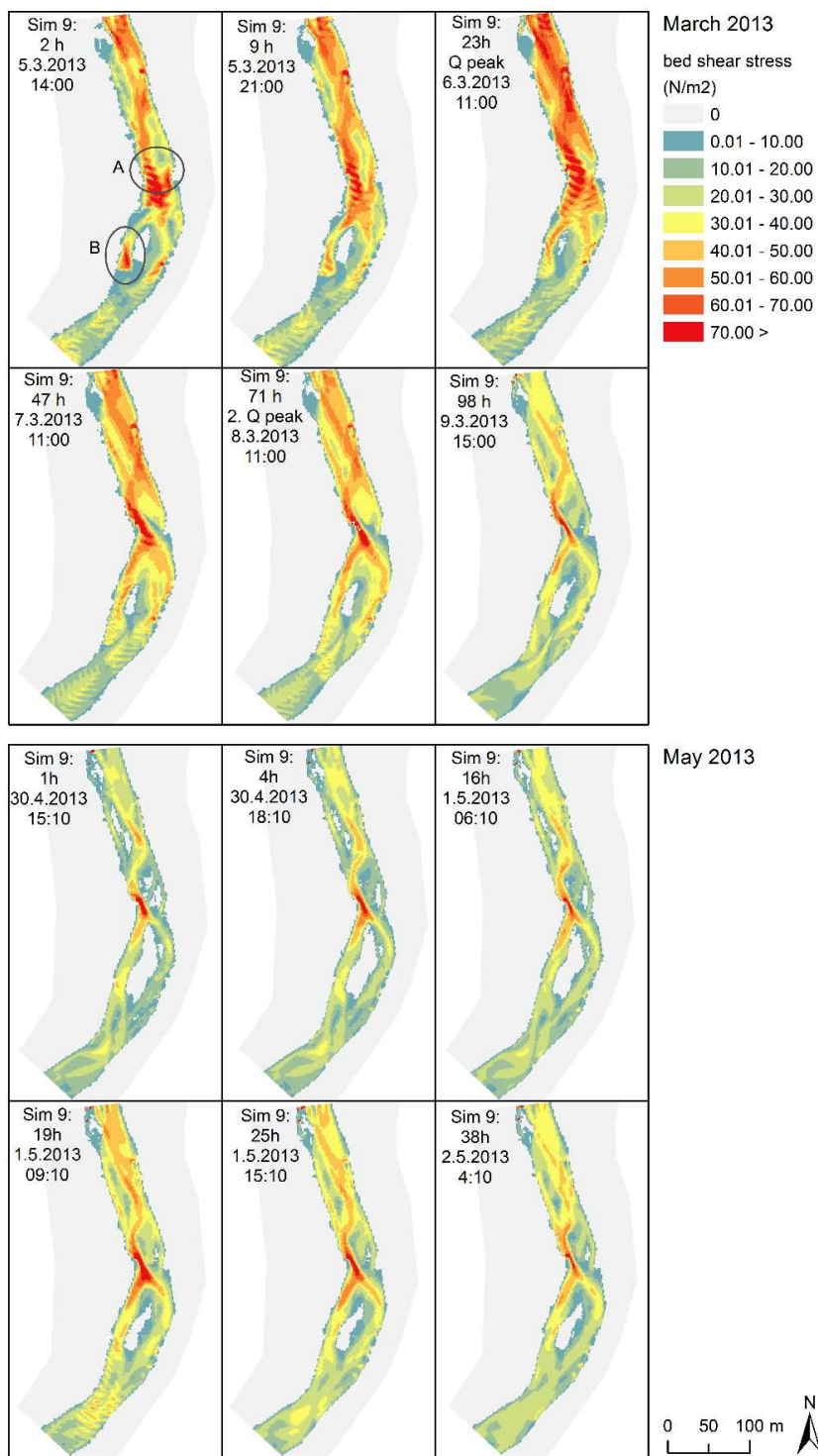
When detecting the whole simulation area (based on simulation 9, as an example), the main flow path was again on the right bank side of the channel, where the greatest changes also occurred during this low-magnitude discharge event (B in Fig. 6). The quick rise in discharge during the 1<sup>st</sup> hour and between 16<sup>th</sup> and 19<sup>th</sup> hours of this discharge event did not cause spatially great changes to the river bed, as most of the changes were within +/- 10 cm throughout the simulation area. Thus, the spatial morphodynamics of this low-magnitude event differed from the moderate-magnitude discharge event. The channel acted more like a braided river during this May 2013 low-magnitude event, as there was continuous small bed elevation changes throughout the event. Thus, there was not such clear high erosion or deposition periods, but continuous steady changes. The bar lobes progressed downstream, as the sediment was transported from the proximal side of the bars to their distal side.



**Figure 5. The hourly volumetric changes during the moderate- and low-magnitude flow events of March and May 2013. The results are based on simulations 9 and 10. The graphs show the erosion and deposition from the beginning of the flow events (12:00 on 5<sup>th</sup> March and 14:10 on 30<sup>th</sup> April) until the time when the erosion and deposition had declined and levelled out during the receding phases. The key time steps are pointed out as hours from the beginning of the flow events.**



**Fig. 6. The bed elevation changes between the key time steps of March and May 2013 events (cf. Fig. 5). The changes of the 5<sup>th</sup> March 2017 at 14:00 are from the beginning of the discharge event.**



**Figure 7. The bed shear stress of simulations 9 during moderate and small discharge events. The key time steps are presented (cf. Figure 5).**

### 4.3 The flow characteristics during the moderate and small discharge events

The bed shear stress (based on simulation 9, as an example) during the first hours of the March 2013 moderate-magnitude discharge event revealed that the bend apex (A in Fig. 7) and right bank side of the downstream lateral bar (B in Fig. 7) experienced the most fluid forces. These explained the locations of the greatest changes, and the initial cutoff of the lateral bar (B). Noteworthy is that these locations were the initial high erosion and bed load transport locations, but the high bed shear stresses occurred spatially more widely during the rising stage (9<sup>th</sup> hour) and the peak flow situation (23<sup>rd</sup> hour). Even though the changes in the river channel had been great between the 24 hours following the discharge peak (Fig. 6), the fluid forces (bed shear stress) had already started to concentrate on the thalwegs of the channel (Fig. 7). Noteworthy is that throughout the receding phase of the moderate-magnitude March 2013 discharge event the greatest shear stresses occurred mainly in thalwegs. Thus, the initial forces of the flood were greater throughout the channel area, but later concentrated on the channel routings formed by the initial stages of the flood discharges, particularly at right bank side at the apex (A) during the receding phase of the flood.

The spatial distribution of the bed shear stresses of the low-magnitude discharge event (May 2013) differed from that of the moderate-magnitude event (Fig. 7). Throughout the discharge event, the greatest fluid forces concentrated in the thalwegs, which had been formed by the preceding moderate-magnitude discharge event of March 2013. Particularly, it could be seen that the bed shear stresses were the greatest at the inner bank side at the apex (A), and then the main flow route was at the right bank side (B) in the lobes area. The greatest bed shear stresses occurred during the second rise of the low-magnitude discharge event, i.e. between 16 and 19 hours from the beginning of the event. Thus, similarly as revealed in Fig. 5, the transport followed the changes in discharge more during this low-magnitude May 2013 event than during the March 2013 event (Figs. 6 and 7).

## **5 Discussion**

### **5.1 Uncertainties related to the simulations**

The morphodynamic simulation of the ephemeral river provides a good quantification of the sequence of channel changes described by Calle et al. (2015). It also deepens the analyses done based on only the topographical data, and is the only way to gain concepts about topographical changes during the flash floods. The reliability of the model, which is calibrated against the events under interest, results improves- can be improved with the quality and temporal density of the available calibration topography, i.e. between-pre- and after-post-flood bedform geometries. We had two high accuracy MLS topographical calibration data sets and one RTK-GPS topographical data sets, which is more than in many other studies. In recent studies done in perennial rivers, where topographical measurements have been sparse, the greatest 2D morphodynamic model uncertainties have related to the channel topographies (Sanyal, 2017). The high uncertainties in topographical measurements

of sub-water areas in gravel bed perennial river have been related particularly to the high bed load velocities and temporal variability of bed load (Williams et al., 2015). However, despite the high quality data from two events at Rambla de la Viuda, we think that further research with multiple yet-to-come events needs to be run to assess the repeatability and validation of the model even better. For example, at Rambla de la Viuda, large floods have not yet occurred since the beginning of the MLS measurement approaches. As also earlier has been stated (Verhaar et al. 2008; Lotsari et al., 2015), the roughness conditions defined for small discharge events, might not be suitable for simulating extreme events. Therefore, the work and refinement of the model will continue, and the applicability of the model for larger floods will be tested, when validation data will be available.

Transport rates (suspended and bed load) and flow velocity measurements, which are measured during the events for calibrating models of perennial rivers, would be suitable to calibrate models during the event, but they are always difficult and dangerous to perform in this particular ephemeral river environment. The uncertainties of the present model approach thus relate to the lack of sediment transport, flow and topographical data during the events. However, the selected initial boundary conditions seem to be congruent with the flooding mechanisms of Rambla de la Viuda and also with other published works, i.e. Williams et al (2016b), for loosely consolidated sand and gravel. So, the modelled flow carried each sand sediment fractions (suspended) adapted to the local flow conditions at inflow boundary, and the model assumed that very little accretion or erosion was experienced near the model boundaries. Based on measurements, the presence of sand size particles and their concentration in the study site were also almost non-existent, and no channel changes occurred at downstream boundary of the simulation area. Thus, this equilibrium load condition was considered valid, and it was also the only option for the present modelling approach, as no input suspended load measurements were available. Similarly as Williams et al. (2016b) state, the model results, i.e. the modelled deposition and erosion, when compared to observations, could have been possibly enhanced if the input suspended sediment load observations would have been available. However, according to Sanyal (2017) the sediment transport is always inherently approximate in nature, and sediment load added to the model causes uncertainties to the results, despite detailed sediment load measurements have been used as model input.

Despite this lack of the data during the flood events, as the forefront lobe prograded over a flat valley bottom (as gravel bed had been mined), we had a good control on sediment volume and gravel particle-size moving downstream as the forefront lobe prograded over a flat valley bottom (as gravel bed had been mined). Total volume input and total transport rates observed in earlier study of Rambla de la Viuda by Calle et al. (2015), had already proved the high availability of sorted gravel particles, and were the basis of the decisions made while building up the model. In fact, as already mentioned, the channel changes bedload sediment output at the output downstream boundary from of the the studied reach was zero. In addition, the simulation result supports the hypothesis of Calle et al. (2015) that moderate- and low-magnitude events reworked sediment locally within the reach. This means that these flows were not able to establish a sediment connection upstream, i.e. between



larger reaches, and the transported sediment originated from the erosion of adjacent areas.

In addition, we were able to use a discharge and precipitation stations further upstream of the study area, and thus also a hydrograph with a known shape (cf. supplementary material). Sometimes only the high water marks are known, and for example, Cao et al. (2010) analyzed the bedload transport by using a symmetrical hydrograph. This could be why their simulations showed less transport during the receding phase than ours. An et al (2017a, 2017b) and Viparelli et al. (2011) had raised the issue related to the cyclic hydrographs and hydrograph boundary layer (HBL), which defines the spatial region within which riverbed topography and grain size respond to the hydrograph. They state that the HBL causes problems, if the material is poorly sorted. The calibration area of the present study was in the downstream part of the simulation area, i.e. outside of the influence of the possible HBL. The gravel of Rambla de la Viuda was well sorted and also therefore we can state that the HBL was not an issue for the calibration of our model.

As the curvilinear grids of fine 0.76–3.03 m and coarse 1.51–5.01 m cells were tested for morphodynamic simulations, the finer grid was proven to better predict the channel evolution, including both flow and bar lobe movement paths. Fine grids of circa 3 m in length have been also found by Williams et al. (2016b) suitable for Deflt3D model applications. Furthermore, the grain sizes and their spatial distribution affected the morphodynamic simulation results greatly. Hooke et al. (2005) have shown that grain sizes have a major effect on the morphological impacts of floods. Simulation results may be enhanced when using techniques that are able to obtain spatially more complete and detailed grain size data (such as e.g. Wang et al., 2013) from the surface layers, instead of lower resolution site sampling, such as Wolman method. Earlier studies have Parker (1990) stated that a calculation of the bedload transport rate of mixtures should be based on the availability of each size range in the surface layer (Parker, 1990), and that during the calibration of Deflt3D model uniform sediment cannot be an acceptable assumption to study the long-term response of natural gravel-bed braided rivers (Singh et al., 2017). The best simulation results of Rambla de la Viuda were indeed gained when spatially varying upper surface layer sediment sizes were applied. However, further detailed analyses of spatial and vertical grain size distribution effects on ephemeral gravel-bed river morphodynamic simulations are suggested. Next step should be to apply the surface grain size variation derived from the laser scanning data sets (Casas et al., 2010a and b; Kukko et al., 2015).

According to Kasvi et al. (2014), The patterns of channels, bars and braiding properties are highly sensitive to the bed slope parameterization, and particularly transverse slope the transverse bed slope parameterization affects the sediment transport in the downslope direction particularly in long-term simulations (Schuurman et al., 2013; Kasvi et al., 2014; Singh et al., 2017). However, a Also during these rather short-term flood event simulations at gravelly gravel bed Rambla de la Viuda, the effects

of the increasing transverse slope on channel evolution was evident, and the simulation results resembled more the observed topographies. Kasvi et al. (2014) had noticed that ~~the model run with the higher the high~~ transverse bed slope effect (i.e. value 3) ~~was, the predicted~~ more filling of ~~the pools occurred~~ in a meandering river, ~~when compared with the model run with no transverse bed slope effect~~. Similarly to Kasvi et al. (2014), the pool area of Rambla de la Viuda ~~had~~ experienced more deposition with the increased transverse slope value (Fig. 4). This pool area was the only clearly wrongly simulated location ~~with this higher transverse slope value, but otherwise the lobe movement and elevation resembled the best to the observations,~~ when this high transverse slope value was applied, ~~but otherwise the lobe movement and elevation resembled the best to the observations~~. Thus, in a gravel bed river with short lasting flow events, adjustments are clearly needed in this parameter for being able to replicate the observed changes while simulating with the currently available sediment transport equations. Thus, similarly to the perennial river simulations (e.g. Kasvi et al., 2014), the transverse slope was important for the simulated changes.

~~Earlier, Hooke et al. (2005) and Graf (1996) have applied the Bagnold equation to an ephemeral channel, but we wanted to use the equations that have often been used together with the Delft 2D model. Reid et al. (1996) showed that the Meyer-Peter and Müller (MPM, 1948) equation, which was originally developed for steady flow conditions and used by Reid et al. (1996) in non-armoured gravel bars, performs well and better than the Bagnold (1980) and Parker (1990) equations in ephemeral gravel river channels. Despite the~~ This justified the application of the MPM equation ~~had earlier been successfully applied in the ephemeral gravelly gravel bed Rambla de la Viuda rivers. Reid et al. (1996) stated that the reason for this is partly due to the fact that sediment was not supply limited by channel bed armour development (Reid et al., 1996), which is the case in many perennial rivers, the~~ However, the movement was minimal ~~with the MPM equation in Rambla de la Viuda~~ and occurred only in the beginning of the moderate event. ~~The second discharge event could not cause much movement with the MPM equation, even though movement had occurred based on the field measurements. The transport rate is imposed as bed-load transport due to currents in both EH and MPM equations, which are used in Delft3D-FLOW. The MPM is more advanced than EH as the MPM includes a critical shear stress for transport (Deltares, 2011). However, this could be one reason for less movement, if critical stresses for transport required in MPM equation did not exceed during the lower discharges of the hydrographs. According to Barry et al. (2004) the formulae containing a transport threshold typically exhibit poor performance when compared to the observed gravel bed river's bed load data. These equations include MPM (1948) and Bagnold (1960) equations, which have been applied in ephemeral channels. In addition, Singh et al. (2017) state that during variable discharge conditions, the MPM formula can not produce the braided river pattern because MPM is not suitable if the value of the Shields parameter falls outside its applicability range. These above mentioned could explain, why The MPM equation is a deterministic equation. According to Van Rijn (1993), in a strongly decelerating flow, the deterministic formulae do not give realistic results because the mean bed shear stress approaches zero in the case of flow separation and reattachment, and thus sediment transport also approaches zero. This may have happened in the Rambla de la Viuda, as the MPM equation did not move particles anymore during the lower discharges and during the receding phase of the hydrographs of Rambla de la Viuda, where flow is~~

temporally highly variable. The application of the ~~deterministic~~ MPM equation could also partly explain why the receding phase's transport was less ~~according to the study of~~ Cao et al. (2010), than based on the present study. Despite being unable to produce the correct morphodynamics caused by the moderate- or low-discharge events of Rambla de la Viuda, the MPM equation could have potential in the simulation of greater floods of ephemeral gravel bed rivers, similarly as Reid et al. (1996) and Cao et al. (2010) have shown. However, further research on this matter should be done also with higher discharges at the river reach.

Engelund and Hansen (1967) developed their equation, so that the effects of dunes, e.g. transport from stream-side to lee-side and form drag, were included. This could be the reason why the EH worked the best at the Rambla de la Viuda, where clear sharp lobe edges occur, and the local slope varies a lot. In addition, the EH is a total load equation (Engelund and Hansen, 1967), whereas, for example, the MPM equation calculates bedload. Meyer-Peter and Müller (1948) equation applies an energy slope, which, for example, Van Rijn (1984) has stated not to be an appropriate parameter for morphological mathematical modelling. The disadvantages of the use of the energy gradient may be according to Van Rijn (1984) the variations due to non-equilibrium phenomena of rising and falling stages. This could partly explain why the EH worked better than the MPM in this ephemeral river.

There are no universal sediment transport formulas, which can cover all coarse sediment transport conditions from low discharges to peak discharges (Li et al. 2016; Singh et al., 2017). The grain sizes, which had been moved by the 2013 discharge events, were on the upper limit of the available sediment transport equations. Also a matter of further research is to detect, whether the transport equations would able to move the sediments at higher discharges, if in reality the higher discharges would possible transport even larger sediment particles.

~~According to Barry et al. (2004) the formulae containing a transport threshold typically exhibit poor performance when compared to the observed gravel bed river's bed load data. These equations include MPM (1948) and Bagnold (1960) equations, which have been applied in ephemeral channels. Barry et al. (2004) noted that simulation results done with different sediment transport equations show substantial differences in performance, but no consistency has been found between equation performance and the complexity of the equation. Despite being unable to produce the correct morphodynamics caused by the moderate or low discharge events of Rambla de la Viuda, the MPM equation could have potential in the simulation of greater floods of ephemeral gravelly rivers, similarly as Reid et al. (1996) and Cao et al. (2010) have shown.~~

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1967), whereas, for example, the MPM equation calculates bedload. Meyer-Peter and Müller (1948) equation applies an energy slope, which, for example, Van Rijn (1984) has stated not to be an appropriate parameter for morphological mathematical modelling. The disadvantages of the use of the energy gradient may be according to Van Rijn (1984) the variations due to non-equilibrium phenomena of rising and falling stages. This could partly explain why the EH worked better than the MPM in this ephemeral river.

However, despite the challenges of applying morphodynamic simulation approaches to ephemeral river channel change analyses, the results are promising, and show the importance of both rising and receding phases. The observations of sediment transport during flow hydrographs would still be important, in addition to the continuous time series of pre- and post-flood topography, but at the moment, their measurement is still a complex and risky issue. By continuing the long-term observations of the ephemeral channel changes with the high-precision equipment, more reliable calibration and validation of the models may be gained.

## 5.2 Moderate- and low-magnitude flow events as channel modifiers

Our study has shown that geomorphic responses ~~in ephemeral rivers, such as Rambla de la Viuda, to the two~~ are unique to each analysed discharge events differed, and the model needs to be calibrated based on consecutive discharge events, and not just one event. Previously this uniqueness ~~to of the geomorphic~~ response has been shown for perennial rivers (Pitlick, 1993), ~~where also topographical and sedimentary data has been applied as initial conditions. The methods and conceptual results of this present study can be applicable in ephemeral rivers worldwide (e.g. USA, House and Pearthree, 1995).~~ Hooke (2016a) stated that the flow events of similar magnitude can have differing effects, depending on the state of the system, as the long-term evolution of the ephemeral river channel and its material greatly influence the response to the stream flow. Some events are more erosional and some are more depositional (Hooke, 2016a). Moreover, Hooke et al. (2005) noted the importance of simulating and analysing the feedback effects of consecutive events. At Rambla de la Viuda, the riverbed morphology formed by the March 2013 influenced the later channel changes during the May 2013 flow. For example, the flow was diverted to the right bank side during the first flood (of March 2013), which also therefore acted as the main channel for the May 2013 flow, which was lower in magnitude.

~~The model was calibrated based on the observations before/between and after the discharge events, because it is knowingly impossible to measure during the risky flash flood events, albeit moderate and low in their magnitude. Therefore, the following discussion is purely based on the model outputs.~~

The simulation results of the Rambla de la Viuda showed that the differences between rising, peak and receding phases of a

moderate-magnitude discharge event are very important in an ephemeral river environment. Higher total amount of channel changes occurred during the receding phase than at the early stages of the discharge events. Deposition dominated due to the progradation of the frontal bar lobe, particularly on the right bank side of the channel. Thus, the continuous channel changes were similar to those for braided perennial rivers (Lotsari et al., 2014a). However, the channel changes differ from a recent study of Gendaszek et al. (2013), who studied the gravel perennial riverbed changes during moderate- (65 m<sup>3</sup> s<sup>-1</sup>) and high-flow (159 m<sup>3</sup> s<sup>-1</sup>) events. They found that most erosion occurred during the rising and the peak flow phases, but did not mention great changes during the receding phase. They found only some scour during sustained high flows following the flood peak (Gendaszek et al., 2013). Noteworthy is that they applied one sensor per reach, and thus the site selection could have greatly affected on their results.

Ferguson stated already in 1993 the potential in numerical modelling of the coupling between geometry, flow, and bedload transport, if it can be applied successfully for the braided channels. The results were promising at Rambla de la Viuda, which also has a braided pattern. According to Wheaton et al. (2013) the chute cutoff mechanism, already described by Ferguson (1993), is the most common braiding mechanism, but that the cutoff is not only an erosional process, but more the result of deposition during the construction of diagonal bars. In the-our study area-of-ours, there was a situation resembling to chute cutoff, as the channel was cut more on the right bank side of the bar than before. The modelling was capable in producing this observed chute cutoff from the right bank side during the moderate discharge event (March 2013). The high values of bed shear stresses related to this initial chute cut off (B location in Figs. 6 and 7). Both erosion and deposition related to these changes, as both topographical observations and model simulations showed the development of the diagonal bar alongside with the cutoff of the bar. However, the diagonal bar formation was slightly greater in the model outcomes than based on observations. Despite this, the model showed potential in producing the channel development following the established theories of gravel bed evolution. In addition, the model showed that the initial cutoff and simultaneous initiation of the diagonal bar took place during the rising limb, but the diagonal bar would not have developed to its full extent without the long receding phase of the flood hydrograph. Williams et al. (2015) had found that the choking is the main process for braiding development in their studied perennial river. However, we were not possible to observe choking processes within our study reach. Further analysis of this braiding process are needed to perform from longer river reaches of ephemeral rivers.

In addition to these braiding processes, the fluvial processes at higher discharges resembled more the ones in a meandering river bend. The greatest erosion and bed shear stresses occurred on the right/inner bank side at the inlet area of the bend (A in Figs. 6 and 7) during the peak and receding flood phases of the moderate-magnitude (March 2013) event. This followed also the results of the erosion and high velocity core locations of perennial meandering rivers (Dietrich and Smith, 1983; Lotsari et al., 2014b). However, during lower flow stage, such as during the initial rise of the floods and the whole May 2013 discharge event, the spatial distribution of channel changes resembled more to braided channel development. Therefore, the

morphodynamics of the low-magnitude May 2013 event differed from the preceding moderate-magnitude discharge event. During May 2013 low-magnitude event, there was continuous steady small bed elevation changes throughout the event. Noteworthy is, that throughout the receding phase of the moderate-magnitude March 2013 discharge event and the whole May 2013 event, the greatest shear stresses occurred mainly at the channel routings formed by the initial stages of the moderate-magnitude flood discharges.

Noteworthy is that during the moderate-discharge event the erosion and deposition peak occurred much earlier than the discharge peak. However, the low-magnitude discharge event experienced the greatest channel changes an hour after the discharge peak. Our model results (e.g. Figs. 7 and 8) suggest the possible existence of hysteresis in the rate of bedform changes, being ~~clockwise or~~ positive in the case of moderate-magnitude flow (bedform change peak occurs before flood peak), and negative in the case of low-magnitude flow (bedform change peak occurs during/after flood peak). The hysteresis phenomenon has been described well in sediment transport studies, and their effect is due, among other factors, to sediment depletion or surface gravel consolidation in the channel (Reid et al., 1985) or a long-lasting portion of the baseflow during the recession limb (Walling, 1974). Cao et al. (2010) have shown, based on their 1D simulations, that bedload transport in an ephemeral river can have similarities to a perennial river. However, even though perennial rivers may have sharp rising phases in their discharge hydrographs (e.g. Long, 2009), they more likely have a greater initial threshold for particle movement by bed-armouring than ephemeral rivers (Reid et al., 1996; Hassan et al., 2009). Although, their armouring can ~~though~~ decrease as bed load concentrations increase (Müller and Pitlick, 2013). Hysteresis of both kinds have also been shown in perennial rivers regarding turbidity, but their flashy storm hydrographs have more often caused anti-clockwise (i.e. negative) hysteresis phenomenon (Lloyd et al., 2016). This would indicate that ephemeral rivers act more similarly to perennial rivers during their low magnitude flow events. Even though further research is needed, the results indicate that the greater the discharge event's magnitude in an ephemeral river is, the more different the channel evolution and its timing are compared to perennial braided gravel bed rivers.

~~These hysteresis phenomena show a time dependence on the fluvial system outputs, depending on a previous sediment transport size and gravel bar morphology, which affects the internal state and the future morphodynamic response.~~ In our study reach,  $30 \text{ m}^3 \text{ s}^{-1}$  was the threshold discharge for the channel changes to become gradual and to follow the discharge changes during the receding phase. According to Hooke (2016a), the threshold values of hydraulic conditions for erosion and deposition in ephemeral channels vary between sites, mostly due to the size and nature of the bed material. Threshold discharge has been observed to be for the deposition and erosion of 15 mm particles  $35 \text{ m}^3 \text{ s}^{-1}$  and  $1 \text{ m}^3 \text{ s}^{-1}$ , respectively, and of 50 mm particles  $14 \text{ m}^3 \text{ s}^{-1}$  and  $40 \text{ m}^3 \text{ s}^{-1}$ , respectively (Hooke, 2016a). The average upper and sublayer  $D_{50}$  grain sizes at the Rambla de la Viuda were 26.3 mm and 17.1 mm, respectively, and the maximum  $D_{50}$  grain size was 40.2 mm. The discharges of the Rambla de la Viuda were overall within the ranges of Hooke's thresholds for the movement of these-sized particles, and the simulations clearly showed their transport. In addition, the threshold discharge of  $30 \text{ m}^3 \text{ s}^{-1}$  for starting the channel changes of the Rambla

de la Viuda to follow the discharge changes during the receding phase coincide well with Hooke's (2016a) threshold analysis of deposition. However, the threshold could be also site dependent, such as for Spanish Mediterranean region, where both Hooke et al (2016a) and the present study have been performed.

## 5 6 Conclusions

A 2D morphodynamic model was successfully implemented for inferring the ~~Due to the proven~~ morphological changes caused by ~~low and moderate-~~ and low magnitude discharge events. The morphological starting and final conditions were derived from high-resolution laser scanning topography, bedform mapping and sediment texture analysis of both surface and subsurface layers on the main morphological bars and channel units. The flow hydrograph was based on a continuous (5 min) record from an upstream gauge station, which was re-scaled to the peak flow estimated at the study reach based on flotsam evidences. The model parameters were calibrated to best fit the pre-and post morpho-topographic field evidences. ~~the preventive measures against flash flood-induced channel changes should take into account moderate and low magnitude flows. This has received support from earlier studies (e.g. Hooke, 2016a and b). As the receding phase can cause substantial channel changes, it plays a major role in inducing damages/modifications to the river environments, and it needs to be acknowledged while planning flood mitigation measures. We have demonstrated the feasibility/goodness of 2D implementation of modelling for analysing ephemeral river channel changes.~~ The following ~~schematic concepts can be concluded~~ conclusions can be made about the morpho- and hydro dynamics ~~and their simulations in~~ an ephemeral river during moderate- and low-magnitude discharge events:

1) The 2D implementation of morphodynamic ~~2D~~ model was proven to work during moderate- and small-magnitude flash flood events in an ephemeral river ~~affected by flash floods. Instead of applying calibration data from only one discharge event, the model results can be enhanced by applying topographical pre and after flood data from consecutive flood events. The spatially varying grain size data, the applied transverse slope parameter value, and sediment transport equations and the streamflow hydrograph were~~ are the most important factors affecting the , apart from the quality of recorded discharge, for the simulation results of the bedform and channel evolution. The selection of sediment transport equations was also critical to match the sediment mass-balance, concluding that the total load equation (Engelund-Hansen, EH), worked the best. When modelling events of moderate and low magnitude in ephemeral rivers of 50 to 100 meters wide, a resolution of at least 1m is recommended (i.e. one cell is 1–2 % of the total width). ~~At Rambla de la Viuda, the curvilinear grid with mesh edge lengths of 0.76–3.03 resulted topographically more realistic outcomes than with coarser grid of i.e. grid size 1.51–5.31 m. A coarser resolution results in incorrect flow and sediment transport routing, as the edges of the bar lobes are smoothed, which affects simulation of water level, hydraulics and bed form evolution, because it is . The finer grid was able to show the steep lobes forefront scarp and showed that laser scanned data is the way to produce the detailed scale topography for simulating the bar lobe movement. Otherwise, procedures would need to include manual modification of the topography to include the lobe~~



edges.

2) ~~M2D morphodynamic models can be a good approach for shedding light on the temporal evolution of channel evolution~~ channel evolution during flash flood events, ~~as measurements during flash floods are risky to perform. With simulations it is possible to improve and deepen the analyses earlier made based on only the survey work or hydrodynamic~~ modelling, as it is possible to analyze whether the greatest changes occur for example due to the peak discharge, the slope of the rising limb, or the length of the receding limb of the flood hydrograph.

3) ~~The spatially varying grain size data, the applied transverse slope parameter value and sediment transport equations are the most important factors, apart from the quality of recorded discharge, for the simulation results of the channel evolution. The total load equation worked better, compared to the deterministic equation.~~

4) Both rising and receding phases of discharge events were predicted to be important for bar movement and channel evolution and thus should not be ignored while planning flood mitigation measures. The erosion and deposition can be greater during the long-lasting receding phase than at the rising phase of moderate- and low-magnitude discharge events, despite the typical hydrograph shape of a flash flood. The receding phase contributed also greatly on the shaping of the bed forms and channel pattern.

5) The deposition and erosion peak rates ~~we~~are predicted to occur at the beginning of the moderate-magnitude discharge events (e.g. March 2013 ~~event~~), whereas deposition dominates ~~s~~d throughout the event, i.e., even during the rising phase. On the contrary, the low-magnitude discharge events (e.g. May 2013 ~~event~~) only experienced ~~d~~ the greatest channel changes after the discharge peak.

46) These different predicted erosion/deposition patterns suggest ~~that the timing of the channel change peak, compared to the timing of the discharge peak, differs between different magnitude discharge events. a hysteresis effect on the morphodynamic changes, These closely related to sediment transport rates, and~~ stress the importance of previous flood history (timing, succession and magnitude) in understanding the geomorphic response of ~~gravelly~~gravel bed ephemeral rivers. ~~The peak of channel changes during the moderate flood occurred before the discharge peak. On the contrary, the low-magnitude discharge event experienced the peak of channel changes after the discharge peak, suggesting a rupture of upper layer before sediment was moved. These differences between the events were at least evident with these two events and the applied parameterization. However, further studies of this possible hysteresis effect are needed from multiple discharge events.~~

5) ~~The results showed that the fluid forces of the initial stages of the moderate-magnitude discharge event caused the initiation of chute cutoff and diagonal bar formation, and defined the flow routing of the rest of the moderate magnitude discharge event. The flow during the following low-magnitude discharge event followed these same thalwegs caused by the preceding moderate-magnitude event. ¶~~

67) The clearest difference in the predicted riverbed changes between the rising and receding phases ~~wis~~ere that erosion and deposition characteristics followed ~~e~~d the temporal discharge changes more during the receding phase than during the rising phase. The threshold discharge, below which the channel-bed changes started following the discharge changes temporally, was around  $30 \text{ m}^3 \text{ s}^{-1}$ . ~~Noteworthy is that this ephemeral channel acted like the braided river channel during these lower flow~~

conditions, but when the bars were submerged in higher discharges the high fluid forces followed the meandering river planform of Rambla de la Viuda. Thus, there were different fluvial processes working at different flow stages, particularly during the moderate-magnitude discharge event.

- 67) ~~P~~ the preventive measures against flash flood-induced channel changes should take into account moderate- and low-magnitude flows. This has received support from earlier studies (e.g. Hooke, 2016a and b). ~~A~~ as the receding phase can cause substantial channel changes, it plays a major role in inducing damages/modifications to the river environments, and it needs to be acknowledged while planning flood mitigation measures.

~~This threshold should be further tested in other ephemeral channels in Mediterranean, and elsewhere.~~

### Data availability

- 10 The data has been stored in the universities of the authors. ~~We have provided the initial geometry, calibration area delineation, sedimentary data (2014 samples) and final topography from the calibration area as supplementary material. Due to the great size of the data, Because some of the data was gained from third parties, we are not able to provide the whole model as open access data. Data those~~ can be also accessed by request from the authors. Each request will be processed separately.

### Author contribution

- 15 Lotsari has done most of the writing and all of the model simulations. Calle and Benito have been also greatly contributed in the writing of the paper and have enabled the study with their projects, as they have initiated the studies of Rambla de la Viuda. All the other authors have contributed to the writing process by commenting the manuscript and its content. All authors have contributed in collecting the field data. Laser scanning has been done by Kukko, Kaartinen, and Alho. The laser scanning data has been processed by Kukko and Kaartinen. Hyypä J. and Hyypä H. and have contributed by providing funding for the
- 20 study and they have also worked on the development of the laser scanning approaches. Benito, Calle and Lotsari have measured the sedimentological data. GPS measurements of topography and high water marks have been done by Benito and Calle.

### Competing interests

The authors declare that they have no conflict of interest.

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