Topographical change caused by moderate and small floods in a gravel bed ephemeral river - depth-averaged morphodynamic simulation approach

Elisaa Loutsari, Mikel Calle, Gerardo Benito, Antero Kuikko, Harri Kaartinen, Juha Huypa and Petteri Alho

1 Department of Geographical and Historical Studies, University of Eastern Finland, Joensuu, Yliopistonkatu 2, P.O. Box 111, 80101 Joensuu, Finland
2 National Museum of Natural Sciences, Spanish Research Council (CSIC), Madrid, Calle de Serrano 117, 28006 Madrid, Spain
3 Department of Remote Sensing and Photogrammetry, Finnish Geospatial Research Institute, National Land Survey of Finland, Kaisanieminki, Geoetensinenv 2, 60430 Mariehamn, Finland
4 Department of Built Environment, Aalto University, Espoo, Vesi ja Vesi 8, P.O. Box 15800, 00076 Aalto, Finland
5 Department of Geography and Geology, University of Turku, Turku, 20014 Turun yliopisto, Finland

Correspondence to: Elisaa Loutsari (elisaa.loutsari@uef.fi)

Abstract. In ephemeral rivers, channel morphology represents a snapshot at the end of a succession of geomorphic changes caused by floods. In most cases, the channel shape and bedform migration during different phases of a flood hydrograph are not recognized from field evidence. This paper analyzes the timing of river bed erosion and deposition of a gravel bed ephemeral river channel (Rama de la Vieda, Spain) during consecutive, moderate (March 2013) and low-magnitude (May 2013) discharge events, by applying a morphodynamic model (Delt 3D) calibrated with pre- and post-event surveys by RTK-GPS points and mobile laser scanning. The study reach is mainly depositional and all bedload sediment supplied from adjacent areas is trapped 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 each flood event. 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 platform. The model results showed that erosion and deposition were in total greater during the long-lasting recession phase than during the rising phase of the flood hydrograph. In the case of moderate-magnitude discharge event, deposition and erosion peaks were predicted to occur at the beginning of the hydrograph, whereas deposition dominated throughout the event. On the contrary, the low-magnitude discharge event only experienced the peak of channel changes after the discharge peak. Thus, both type of discharge events highlight the importance of modeling 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 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, which can be divided into two types: the force of the flow and related channel changes throughout the discharge events. For emergency costs and enhancing preventive flood mitigation measures, understanding the forces of the flow and related channel changes throughout the discharge events is very 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.

Most geomorphological studies have been mostly concentrated on these large discharge events, due to their impacts on the river channel changes related to river environments and human infrastructure (Greenbaum and Bergman, 2006; Goodek et al., 2012; Nardi and Rinelli, 2015; Hooke, 2016). However, moderate and low flows have also been shown to cause great morphological changes in gravel bed river channels (Calle et al., 2015; Hooke, 2016a), as a small discharge over long time spans can substantially rework the sediment and bedload of the river channels produced by greater floods (Greenbaum and Bergman, 2006).

According to Hooke and Maryn (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 peak flow phases in perennial rivers (e.g. Gendron et al., 2013), but similar knowledge for ephemeral river channels is still lacking. Therefore it is important to understand the capacity of ephemeral river for sediment deposition and flooding due to the combined effects of water flow and sediment transport during flood situations. Ephemeral river channel changes are mainly due to the recession phase in which the receding phase should not be ignored when planning flood mitigation measures.

Conversely, 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. Eiler et al., 2017). New insight on morphodynamic changes and their driving parameters during flash floods can be gained by applying simulation methods. Simulations may provide information about the channel dynamics from the times when it has not been possible to perform measurements (Loutsari et al., 2014a) and thus increase our understanding of sediment dynamics during flood events (Hooke et al., 2005). However, morphodynamic modelling of gravel migration have so far been more common in perennial braided gravel bed rivers (e.g. Williams et al.,
according to the measurements (see section 3.3 below), which had been performed in the areas of topographical changes. These have also been published in Calle et al. (2015). The movement of these gravels caused the development of the bar fronts (Calle et al., 2015).

Figure 1: The study site of the Rambles 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 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 listed in Fig. 5). Also the sediment sample locations are shown (see also Table 2 and Fig. 2).

Calle et al. (2015) described the main morphological 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 (Delft3D) 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 D40 of 32–45 mm) and absence of a well-developed armoured layer (Calle et al., 2015). Thresholds for sediment transport were proven to fit the Hjulström graph (1935) in this ephemeral environment. However, simulations did not include topographical change and its influence on the hydraulic parameters during the floods, and could not answer how 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

Figure 2: The conceptual graphic presented by Calle et al. (2015: 5.10, www.schweizerbart.de/journals/fp): the pre-stage of 2012 (a), high flood stage during March 2013 event (b), low flood 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, H: high 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 at the downstream boundary condition for the hydro- and morphodynamic model (2D module of Delft3D-FLOW). As no direct measurements were possible during the study reach during the discharge events, such as commonly done for modelling 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 rain gauges. The hydrograph shape of the study site can be assumed similar to Vall d'Alba gauge station, due to 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 HWM 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). 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 2D implementation of Deli3D morphodynamic model required channel geometry in grid format. Orthogonalized curvilinear grids of two different resolutions were created from both measured topographies, one with "course" 1.51–5.31 m cells and one with "fine" 0.76–3.03 m cells (e. g. circle half of the courser grid cell size). 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 than the "fine" resolution (i.e. 0.76–0.08 m) did not enhance the results, and those only increased the computational time. Thus, curvilinear grids of two resolutions, "course" 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 simulations was defined by adding the MLS 2012 measurement points to the curvilinear grids, both to the coarse and fine grids, and averaging the point values for the grid cells. There was also available a digital elevation model (DEM) from 2009 (from airborne laser scanning), which had 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 at 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 validation topography after the May 2013 flood, was defined by adding 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 areas), was used in the 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 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). Different methods did not recover the differences between upper layer and sublayer sediment distribution and what kind of particles the 2013 discharge events had moved. Therefore, the gravel moved by the 2013 spring events were measured using a US standard gravimeter (US SAR-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 sampling of sediments (c. 10–60 cm in 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 Alt (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, WBT and 9 were measured in March 2013, and SUB1–6 and UP 1–6 were measured in June 2014. W= Wollen, C= channel, B=Bar, SUB=sublayer, UP=upper layer. Most of the samples were within the calibration area, and their locations can be seen in Fig 1 and also in Fig 11 of Calle et al. (2015). Due to the spatial scarcity of the 2013 measurements (only two samples) were not applied as the surface grain sizes for the model. The armour ratio was calculated following Lide and Madej (1992), i.e. ratio of the surface-to-subsurface D50.

<table>
<thead>
<tr>
<th>Sample</th>
<th>D50 (mm)</th>
<th>D94 (mm)</th>
<th>D10 (mm)</th>
<th>Armour ratio</th>
</tr>
</thead>
<tbody>
<tr>
<td>WCM</td>
<td>18</td>
<td>22</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>WCU</td>
<td>30</td>
<td>101</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>WB1</td>
<td>22</td>
<td>44</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>WB2</td>
<td>17</td>
<td>33</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>WB3</td>
<td>17</td>
<td>34</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>WB5</td>
<td>12</td>
<td>21</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>WB6</td>
<td>17</td>
<td>37</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>WB7</td>
<td>14</td>
<td>22</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>WR1</td>
<td>18</td>
<td>35</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>WR2</td>
<td>20</td>
<td>31</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>UP1 and SUB1</td>
<td>22.5</td>
<td>39</td>
<td>18.5</td>
<td>1.126</td>
</tr>
<tr>
<td>UP2 and SUB2</td>
<td>35.5</td>
<td>79</td>
<td>21.3</td>
<td>1.667</td>
</tr>
<tr>
<td>UP3 and SUB3</td>
<td>18.8</td>
<td>31</td>
<td>14.3</td>
<td>1.314</td>
</tr>
<tr>
<td>UP4 and SUB4</td>
<td>18.0</td>
<td>30.5</td>
<td>14.4</td>
<td>1.230</td>
</tr>
<tr>
<td>UP5 and SUB5</td>
<td>31.2</td>
<td>55.4</td>
<td>18.6</td>
<td>1.677</td>
</tr>
<tr>
<td>UP6 and SUB6</td>
<td>41.2</td>
<td>80</td>
<td>16.7</td>
<td>2.467</td>
</tr>
<tr>
<td>Average UP1-6</td>
<td>26.3</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>Average SUB1-6</td>
<td>17.1</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>Average UP and SUB1-6</td>
<td>21.1</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
</tbody>
</table>

(5) The D50 and D90 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 D50 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
carrying input sand sediment fractions. This was similar to earlier studies done in gravel bed river (Williams et al., 2018a), and was based on earlier experiences of the river (Calie 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. 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 utilized. The transverse slope affects transport towards the down-slope direction (Defuera, 2011). The Bagrold (1966) equation was applied for the longitudinal slope and Ikeda (1982), as presented by Van Rijn (1993), was applied for the transverse slope.

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

<table>
<thead>
<tr>
<th>Simulation</th>
<th>Transport</th>
<th>Elevation</th>
<th>Discharge</th>
<th>Transverse</th>
<th>Grid size</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>EHS</td>
<td>1</td>
<td>1.5</td>
<td>coarse</td>
<td>varying, 2013</td>
</tr>
<tr>
<td>2</td>
<td>EHS</td>
<td>1</td>
<td>1.5</td>
<td>coarse</td>
<td>varying, upper, 2014</td>
</tr>
<tr>
<td>3</td>
<td>EHS</td>
<td>1</td>
<td>1.5</td>
<td>coarse</td>
<td>varying, upper, 2014</td>
</tr>
<tr>
<td>4</td>
<td>EHS</td>
<td>1</td>
<td>1.5</td>
<td>coarse</td>
<td>varying, upper, 2014</td>
</tr>
<tr>
<td>5</td>
<td>EHS</td>
<td>1</td>
<td>1.5</td>
<td>coarse</td>
<td>varying, upper, 2014</td>
</tr>
<tr>
<td>6</td>
<td>EHS</td>
<td>1</td>
<td>1.5</td>
<td>coarse</td>
<td>varying, upper, 2014</td>
</tr>
<tr>
<td>7</td>
<td>EHS</td>
<td>1</td>
<td>1.5</td>
<td>coarse</td>
<td>constant, average upper, 2014</td>
</tr>
<tr>
<td>8</td>
<td>EHS</td>
<td>1</td>
<td>1.5</td>
<td>coarse</td>
<td>constant, average upper, 2014</td>
</tr>
<tr>
<td>9</td>
<td>EHS</td>
<td>2</td>
<td>1.5</td>
<td>coarse</td>
<td>varying, upper, 2014</td>
</tr>
<tr>
<td>10</td>
<td>EHS</td>
<td>2</td>
<td>1.5</td>
<td>coarse</td>
<td>varying, upper, 2014</td>
</tr>
<tr>
<td>11</td>
<td>MPM</td>
<td>2</td>
<td>1.5</td>
<td>coarse</td>
<td>varying, upper, 2014</td>
</tr>
</tbody>
</table>

Altogether, 61 morphodynamic simulations were needed for calibrating the morphodynamic model. These 61 simulations included simulations with Q1, Q2, and Q3 discharge hydrographs. During these simulations the match of the simulated water levels to HWMs was also checked. The roughness values were still valid. Out of these 61 morphodynamic simulations, 11 simulations done during calibration were selected to be presented in the supplementary material of this paper (see also Table 4). Reasons for their selection was that these 11 simulations showed the effects of grid size (before [2012] and after [2014] floods, and grid 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 [EHS], Meyer-Peter and Muller [MPM]) on model performance (Table 4, and supplementary material).

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. Kavi 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 Q2 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).

Figure 4: The comparison between the observed and simulated (simulations 9 and 10) elevations. The negative values (red) mean that the observed (A)/simulated (B) March 1/May 2 2013 topographies were higher (i.e. deposition had occurred) than observed in 2012. The negative values also mean that simulated elevations of March/May 2013 were higher than was observed in March/May 2013 with simulations 9 (C) and 10 (D). 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 FC).

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 underestimated 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).
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 6th and 7th 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 display of the gravel bars, i.e. alternating bars and pools, the river channel has also a broader meandering platform. The simulated area can be considered as one bend. The downstream lobe area located 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 (Lotzari et al., 2014b). Thus, during the higher flow stage 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 flow 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 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 1st hour and between 16th and 19th 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 5th March and 14:10 on 30th 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.
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 (9th hour) and the peak flow situation (24th 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 the 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, can be improved with the quality and temporal density of the available calibration topography, i.e. pre- and post-flood bedforms geometries. We had two high accuracy MLS 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 3D morphodynamic model uncertainties have related to the channel topographies (Suryal, 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 (Verheugen et al. 2008; Lotani 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 measurements, which are measured during the events for calibrating models of perennial rivers, are always difficult and dangerous to perform in this particular ephemeral river. 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 Suryal (2017) the sediment transport is always inherently approximate in nature, and sediment load added to the model causes uncertainty to the results, despite detailed sediment load measurements have been used as model input.

Despite this lack of the data during the flood events, we had a good control on sediment volume and gravel particle-size moving downstream as the flood event progressed 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. As already mentioned, the channel changes at the outflow downstream boundary of 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
5.2 Moderate- and low-magnitude flow events as channel modifiers

Our study has shown that geomorphic responses to the two analysed discharge events differed. Previously this uniqueness of geomorphic response has been shown for perennial rivers (Pettick, 1993), where also topographical and sedimentary data has been applied as initial conditions. Hooker (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 (Hooker, 2016a).

Moreover, Hooker et al. (2005) noted the importance of simulating and analysing the feedback effects of consecutive events. At Rambøll 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 simulation results of the Rambøll 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 fluvial bar lobe, particularly on the right bank side of the channel. Thus, the continuous channel changes were similar to those for braided perennial rivers (Lottarsi et al., 2014b). However, the channel changes differed from a recent study of Gerdaaske et al. (2013), who studied the gravel perennial riverbed changes during moderate- (65 m³/s) and high-flow (159 m³/s) 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 (Gerdaaske 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 1990 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 Rambøll 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 our study area, there was a situation resembling to chute cutoff, as the channel was cut more on the right bank side than on the bar. The modelling was capable in producing this observed chute cutoff from the right bank side during the moderate discharge events (March 2013). The high values of bed shear stresses related to this initial cut off (8 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 chocking is the main process for braiding development in their studied perennial river. However, we were not possible to observe chocking 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 resemble more the ones in a meandering river bend. The greatest erosion and bed shear stresses occurred on the right/main 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; Lottarsi et al., 2014b). However, during lower flow stage, such as at the initial rise of the flood 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 at 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 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 deposition 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-organic than ephemeral rivers (Reid et al., 1996; Hassen et al., 2009). Although, their armouring can decrease as bed load concentrations increase (Müller and Pettick, 2013). Hysteresis of both kinds have also been shown in ephemeral 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.
Lotuari 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 Rambles 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, Kaaritonen, and Alho. The laser scanning data has been processed by Kukko and Kaaritonen. Hyyppa J. and Hyyppä H. and have contributed by providing funding for the study and they have also worked on the development of the laser scanning approaches. Benito, Calle and Lotuari 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.

Acknowledgements
We would like to thank the Instituto Cartografico Valencia, Servicio Automatico de Informacion Hidrologicas of the Confederacion Hidrograficas del Jucar for supporting data. Financial support is provided by the Spanish “Ministerio de Economia y Competitividad” (projects CGL2011-29176 and CGL-2014-58127-C3-1-R), the Academy of Finland (Extreme and annual fluvial processes in river dynamics – ExRIVER [grant number 267345], the Centre of Excellence in Laser Scanning Research – CoELaSR [grant number 272195], and the Strategic Research Council project “COMBAT” [grant number 293389]) and Maj and Tor Nessling Foundation (Aärmännien ja vuononisten fluviaali-prosessien vaikutukset jokidynamikkaan [grant number 2013067]).

References


