Quantifying the combined effects of multiple extreme floods on river channel geometry and on flood hazards

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ABSTRACT

Effects of flood-induced bed elevation and channel geometry changes on flood hazards are largely unexplored, especially in the case of multiple floods from the same site. This study quantified the evolution of river channel and floodplain geometry during a repeated series of hypothetical extreme floods using a 2D full hydro-morphodynamic model (LHMM). These experiments were designed to examine the consequences of channel geometry changes on channel conveyance capacity and subsequent flood dynamics. Our results revealed that extreme floods play an important role in adjusting a river channel to become more efficient for subsequent propagation of floods, and that in-channel scour and sediment re-distribution can greatly improve the conveyance capacity of a channel for subsequent floods. In our hypothetical sequence of floods the response of bed elevation was of net degradation, and sediment transport successively weakened even with floods of the same magnitude. Changes in river channel geometry led to significant impact on flood hydraulics and thereby flood hazards. We found that flood-induced in-channel erosion can disconnect the channel from its floodplain resulting in a reduction of floodwater storage. Thus, the frequency and extent of subsequent overbank flows and floodplain inundation decreased, which reduced downstream flood attenuation and increased downstream flood hazard. In combination and in summary, these results suggest that changes in channel capacity due to extreme floods may drive changes in flood hazard. The assumption of unchanging of river morphology during inundation modelling should therefore be open to question for flood risk management.

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1. INTRODUCTION

Extreme floods are not contained within a river channel, nor are ever composed entirely of water, but rather include considerable sediment transport. Extreme floods exert significant river geomorphological change and these changes can have an extensive and pervasive geological legacy (Alho et al., 2005; Baynes et al., 2015b; Carling, 2013; Carrivick et al., 2010; Guan et al., 2015b). Whilst extreme floods are by definition infrequent and they occur during a very short period of time, field evidence has shown that hydro-geomorphic responses to floods may affect flood hazard and risk due to changes in channel morphology and to subsequent river hydraulics (Borga et al., 2014; Fewtrell et al., 2011; Lane et al., 2007; Marchi et al., 2009).

Types of extreme floods include glacial outburst floods, dam bursts and flash floods due to intense rainfall. The morphological imprint (or adjustment) of river channels to extreme floods can cause short-term and long-term impacts on river hydraulics. Indeed many studies have reported the spatiotemporal morphological response to a single extreme flood and to the effects on flood hydraulics of geomorphological impacts during that flood (Baynes et al., 2015a; Nardi and Rinaldi, 2015; Sloan et al., 2001; Staines and Carrivick, 2015). Considering longer-term impacts, Carling (2013) indicated that a repeated series of floods is likely to result in sediment exhaustion effects and that subsequent high-magnitude floods may disrupt much of the sedimentary evidence of earlier floods. Therefore, sediment dynamics and the resultant channel geometry adjustment are more complex during a series of floods than during a single event.

Increased flooding frequency and/or magnitude are commonly driven by hydrological changes manifest in initial hydrograph properties, such as an increase in peak water discharge. Recently, some studies have reported that geomorphological changes also play a key role in influencing flood hazards (Lane et al., 2007; Neuhold et al., 2009; Slater et al., 2015; Stover and Montgomery, 2001). For instance, bed aggradation decreases channel capacity and bankfull heights, thereby resulting in a wider regional inundation extent. Conversely, channel incision due to in-channel scour during floods lowers bed elevation and increases conveyance capacity of a channel in flood, so leading to smaller overbank flows. Both scenarios imply changes of flood hazard frequency. Therefore, it is reasonable to suggest that effective flood inundation and hydraulic modelling should consider the sensitivity of flood dynamics to the changes in both hydrological and morphological processes. However, recent research on flood modelling has preferred to assume a static river channel geometry both during flood events and between flood events (Guan et al., 2013; Horritt and Bates, 2002; Liang, 2010), and thus ignoring cumulative (long-term) erosion or deposition. Exceptions include the study by (Wong et al., 2014)) who reported that the inclusion of bed elevation changes appeared to alter flood dynamics locally, but that it was not significant for flood inundation, and Slater et al. (2015) who using a large number of field studies and statistical analysis concluded that the changes in channel morphology could lead to significant effects on flood hazard frequency. They also mentioned that morphological effects might be even larger and more widespread than the flow frequency effects.

The significance of channel geometry in flood hazard has also been reported by Lane et al. (2007) which explored the effects of channel aggradation due to upstream sediment delivery on inundation extent. Lane and Thorne (2007) suggested that future flood risk should be conditioned not only by changes in conveyance capacity, but also by morphological adjustment in response to changes in river flows and/or upstream sediment supply. In this regard, Neuhold et al. (2009) has incorporated river morphological changes to flood risk assessment, and emphasised that the influence of bed elevation changes on flood hazard is much higher than the influence of discharge input variations.

Recent research also verified the significant effects of channel adjustments on hydraulics of flood either by field observation (Rickenmann et al., 2015; Wyżga et al., 2015) or by numerical modelling (Guan et al., 2015b; Li et al., 2014; Staines and Carrivick, 2015). In combination and overall, these studies permit a conceptual hypothesis to be proposed that the influence of sediment transport and subsequent channel changes can be a key driver on flood hazard. Testing this hypothesis requires a detailed study on changes in bed elevation during multiple floods and the effects of that morphological change on subsequent flow hydraulics; particularly on flow capacity and flow conveyance.

This study aims to evaluate sediment dynamics within a river channel during a repeated series of extreme floods, and the resultant changes in channel geometry, conveyance capacity of the channel in flood and flood hydraulics. The research questions that this paper poses are:

(1) How is river channel geometry changed during multiple extreme floods?

(2) What effects do river channel geometry changes have on conveyance capacity? and

(3) How do changes in river channel geometry influence flood hazards?

2. STUDY SITE AND TEST CASE

A glacial outburst flood that occurred in 1999 at Sólheimajökull in southern Iceland is used as the study case because of the wealth of morphological (Staines et al., 2014) and hydraulic (Russell et al., 2010; Staines and Carrivick, 2015) information available for this event. The model performance and sensitivities have been presented and assessed by Guan et al. (2015b). Building upon this study, the work here designs and runs a series of experiments in order to elucidate the significance of changes in river morphology and the necessity of including them for flood hazard analyses. It is worth noting that Staines and Carrivick (2015) and Guan et al. (2015b) investigated sediment dynamics within the 1999 FLOOD (individual flood) and their geomorphological impacts, but each using different numerical models. This study differs from both of those by having extended experimental scenarios to address general concerns that exist in flood hazard assessment.

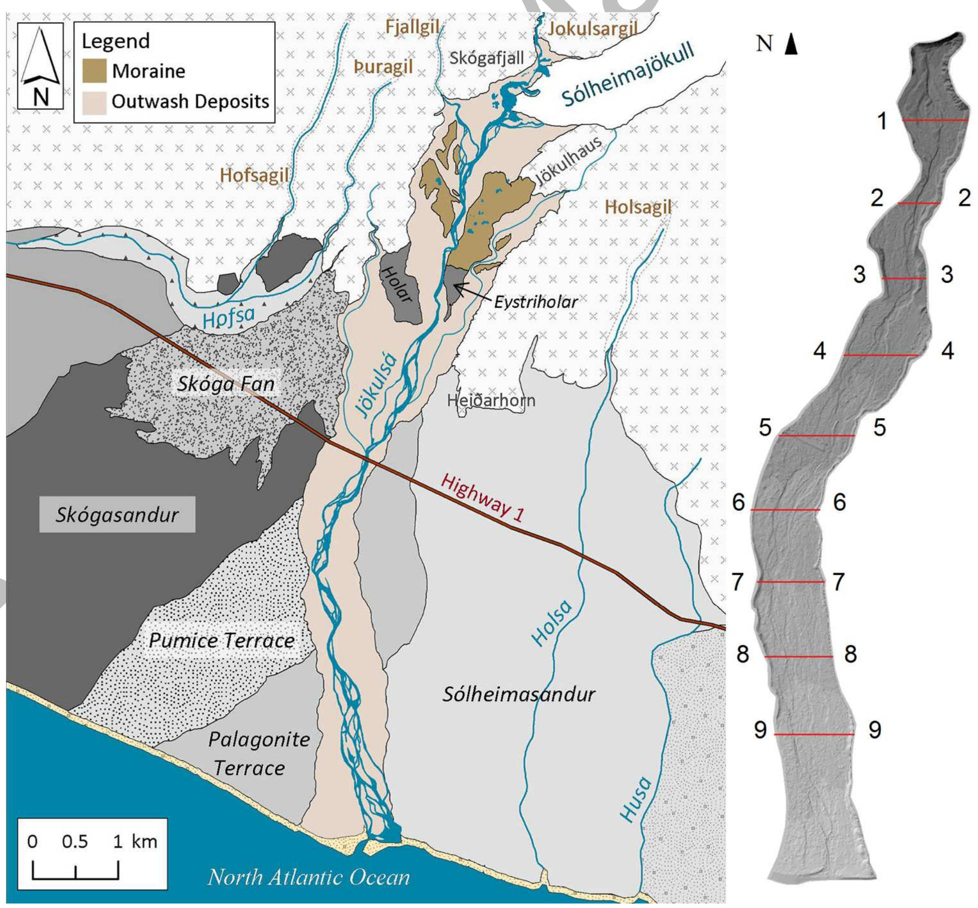


Figure 1. (a) Geomorphology of the study river showing the main Jökulsá channel, and (b) digital elevation model (DEM) of the river channel at 2m × 2m grid cell size resolution

The river channel Jökulsá flows from the 8 km long Sólheimajökull outlet glacier that drains from the Mýrdalsjökull ice cap in the southern volcanic zone of southern Iceland (Figure 1). The river is about 8.7 km in length. Its main flow sources include upstream glacial meltwater and the river Fjallgilsá, flowing into the Jökulsá approximately 2 km downstream of the glacier snout.

The 1999 FLOOD was triggered by subglacial volcanic activity. The flooding process was sudden, lasted ~ 6h and had high discharge. The flood burst initially from the western margin of Sólheimajökull and then flowed into the downstream river channel along with additional meltwater from the glacier. It was reported by Sigurdsson et al. (2000) that the peak water discharge rose rapidly to 1700 m^3s^{-1} as recorded at the bridge of 4 km downstream ~ 1 hour after the flood initiation. Peak water discharge at the glacier terminus has been estimated at 4000 ± 250 m^3s^{-1} from the size of boulders and the velocity required to transport them (Russell et al., 2010). Staines and Carrivick (2015) pointed out that the peak value 4000 m^3s^{-1} was rather high and defined a hydrograph with 40 % of the discharge from the Central Conduit and 60 % from the Western Conduit which was a good-fit to the observations.

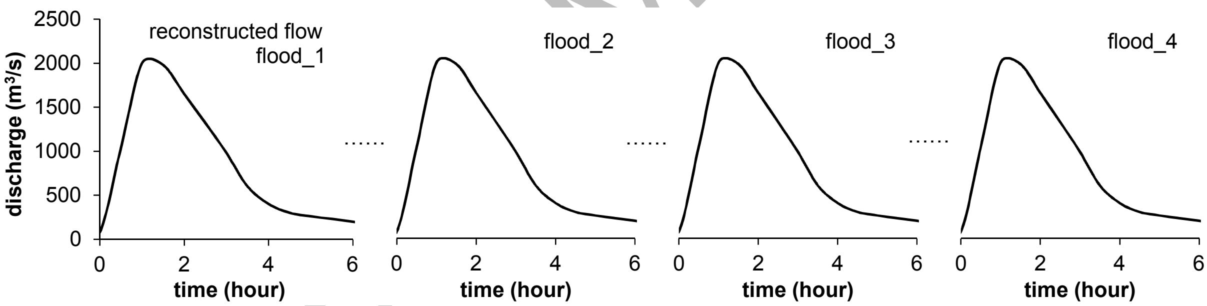


Figure 2. Inflow hydrograph

This study used the input hydrograph provided by Staines and Carrivick (2015) for its first experiment ('flood_1'). Our subsequent scenarios considered that three more extreme floods with the same peak discharge occurred in the river channel (named as flood_2, flood_3 and flood_4). Admittedly, flood sequences in reality differ from each other in terms of hydrograph shape, peak discharge and time. In this study, we choose same hydrograph shape, peak and time for the experimental flood sequence because this study strive to: (1) explore sediment transport activities within a same flood magnitude but a different occurring order, (2) investigate cumulative changes in river morphology during multiple flood events, and (3) elaborate agglomerate effects of multiple channel adjustments on flood hazards. Thus, the experiments can provide multiple spatiotemporal viewpoints on morphological changes and impacts during flood sequences. The inflow hydrograph of the simulated scenarios is plotted in Figure 2.

To be clear, in this paper we are not specifically focusing on the 1999 FLOOD itself, which has already been explored by (Russell et al., 2010; Staines and Carrivick, 2015; Staines et al., 2014), but rather on quantitatively and qualitatively improving understanding of 'flood memory' and 'morphological imprint' during a series of floods. In reality, river morphology may additionally adjust slightly because of human interventions, sediment transport and/or bank erosion caused by perennial flows in channel between each flood. Since the main concern of the study was on the imprint of floods in a riverbed and its flood impact, it was assumed that the slight inter-event changes in river morphology were not significant and neglected. We realise that the exact same scenario will probably not happen repeatedly in the study site. However, widespread and persistent flooding commonly occurs over a short period in reality, such as recent floods across the UK in 2014 and 2015, and this is increasing due to the extreme weather in the context of climate change. Therefore, although this study is running some hypothetical experiments with flood sequences, we consider that the finding from the experiments in this study can profoundly improve the insights into sediment activities within multiple floods and the effects of channel changes on flood hazards, and the finding will be transferable to similar cases suffering from frequent multiple fluvial floods.

Pre-flood aerial photographs were taken in August 1996 and post-flood aerial photographs date from August 2001. Both sets were sourced from Landmaelingar Islands (LMI) and orthorectified in Leica Photogrammetry Suite (LPS) with ground control points (GCPs) generated using a Leica GPS500 dual phase differential Global Positioning System (dGPS) (Staines et al., 2014). Using the photogrammetry, Staines et al., (2014) built the digital elevation models (DEMs) with 2 m resolution before and after the flood. These pre- and post-flood datasets are a very unusual asset for this kind of modelling study. The 1996 DEM was used as the initial input domain for simulation, and the 2001 DEM was used to compare to the simulated bed. DEMs errors and uncertainty were assessed by comparing grid values with the differential Global Positioning System (dGPS) and with a DEM constructed from a 2010 summer LiDAR survey, which is assumed had no errors. The 1999 flood eroded and carried a considerable amount of sediment, causing rapid bed change. However, it is quite challenging to quantify such spatiotemporal river channel adjustment. Therefore, the difference of DEMs (DoD) before (1996 DEM) and after (2001 DEM) the 1999 flood was used to be approximate quantification of ‘real’ bed changes caused by the flood, although there are some extra perturbations over the time scale of five years.

3. METHODOLOGY

3.1 Numerical model - LHMM

The hydro-morphodynamic model (LHMM) that has been presented in previous work (Guan et al., 2014; Guan et al., 2015a; Guan et al., 2015b) was applied in this study and thus is only briefly described here. LHMM solves fully coupled shallow water equations (SWEs) and sediment transport model: both bedload and suspended load. Two-dimensional (2D) depth-averaged SWEs are solved for predicting rapidly varying unsteady flows, and a non-uniform sediment transport model is developed for bed erosion and deposition. The model considers mass and momentum exchange of non-cohesive sediment between the bed and the flow, and updates the hydraulic and sediment quantities per grid cell, per time step. The 2D hydro-morphodynamic model is solved by using a robust Godunov-type finite volume method.

3.1.1 Hydrodynamic model

The depth-averaged SWEs with flow-sediment interactions are written in vector form as follows:

$$\frac{\partial \mathbf{U}}{\partial t} + \nabla \cdot \mathbf{F} = \mathbf{S_o} + \mathbf{S_f} + \mathbf{S_{f-b}} \tag{1}$$

where **U** is the vector of conserved variables, **F** is the flux vector function, $\mathbf{S_o}$, $\mathbf{S_f}$, and $\mathbf{S_{f\text{-}b}}$ are the vector of bed slope term, frictional slope term and flow-bed interaction term, and $\nabla = \vec{\imath}(\partial/\partial x) + \vec{\jmath}(\partial/\partial y)$ is the gradient operator.

$$\mathbf{U} = \begin{pmatrix} \eta \\ hu \\ hv \end{pmatrix},\ \ \mathbf{F} = \begin{pmatrix} h\mathbf{V} \\ hu\mathbf{V} + \frac{1}{2}gh^2\vec{\imath} \\ hv\mathbf{V} + \frac{1}{2}gh^2\vec{\jmath} \end{pmatrix}, \mathbf{S_o} = \begin{pmatrix} 0 \\ -gh\frac{\partial z_b}{\partial x} \\ -gh\frac{\partial z_b}{\partial y} \end{pmatrix}$$

$$\mathbf{S_f} = \begin{pmatrix} 0 \\ -ghS_{fx} \\ -ghS_{fy} \end{pmatrix}, \mathbf{S_{f-b}} = \begin{pmatrix} 0 \\ \frac{\Delta\rho u}{\rho}\frac{\partial z_b}{\partial t}[\alpha(1-p) - C] - \frac{\Delta\rho g h^2}{2\rho}\frac{\partial C}{\partial x} - \mathbf{S_{ad}} \\ \frac{\Delta\rho v}{\rho}\frac{\partial z_b}{\partial t}[\alpha(1-p) - C] - \frac{\Delta\rho g h^2}{2\rho}\frac{\partial C}{\partial y} - \mathbf{S_{ad}} \end{pmatrix} \tag{2}$$

where h = flow depth (m); z_b = bed elevation (m); $\eta=h+z_b$ = water surface (m); u, v = the x and y components of flow velocity (m/s); **V** is the velocity vector defined by $\mathbf{V} = u\vec{\imath} + v\vec{\jmath}$; p = sediment porosity (dimensionless); C = total volumetric sediment concentration including both bedload and suspended load (dimensionless); ρ_s, ρ_w = densities of sediment and water respectively (m^3s^{-1}); $\Delta\rho=\rho_s-\rho_w$; ρ = density of flow-sediment mixture (m^3s^{-1}); S_{fx}, S_{fy} are Manning-based frictional slopes in x and y direction (dimensionless); $\alpha=u_s/u$ = sediment-to-flow velocity ratio (dimensionless) defined by the equation presented by Greimann et al. (2008); $\mathbf{S_{ad}}$ is the additional term vector related to the velocity ratio α defined by

$$\mathbf{S_{ad}} = \frac{\Delta\rho\mathbf{V}}{\rho}(1-\alpha)[C\nabla\cdot(h\mathbf{V}) - (h\mathbf{V})\nabla\cdot\mathbf{C}] \quad (3)$$

where **C** is the sediment concentration vector defined by $\mathbf{C} = C(\vec{\imath} + \vec{\jmath})$.

3.1.2 Sheet flow load

Sheet flow load was defined as bedload dominant transport including portion of suspended load (Pugh and Wilson, 1999; Sumer et al., 1996). Sheet flow has highly concentrated sediment occurs in a layer near the bed with a thickness of several times sediment grain size. The velocity in this layer is usually lower than the water velocity, thus the model considers the sediment-to-flow velocity ratio. As the channel bed was composed of several sediment fractions with different grain sizes, a non-uniform model was preferable. For each size class, the governing equation (mass-balance) of non-equilibrium sheet flow was applied following Guan et al. (2014).

$$\frac{\partial hS_{bi}}{\partial t} + \frac{\alpha\partial huS_{bi}}{\partial x} + \frac{\alpha\partial hvS_{bi}}{\partial y} = -\frac{\alpha(q_{bi} - F_i q_{b*i})}{L_i} \quad (4)$$

where S_{bi}=volumetric bedload concentration of the ith size class; q_{bi}=real sediment transport rate of the ith fraction; q_{b*i} =sediment transport capacity of the ith fraction; L_i= non-equilibrium adaptation length of sediment transport of the ith fraction determined by using the equation in Guan et al. (2014); F_i represents the proportion of i th grain-size fraction in total moving sediment.

As suggested by Guan et al. (2014), this study chose the combination of the modified Meyer-Peter & Müller formula (MPM) (Meyer-Peter and Müller, 1948) and the Smart & Jäggi formula (SJ) (Smart and Jäggi, 1983) based on the bed slopes to calculate transport capacity.

$$q_{b*i} = \varphi\sqrt{g(\rho_s/\rho_w - 1)d_i^3} \tag{5}$$

$$\varphi = \begin{cases} 8\left(\theta_i - \theta_{cr,i}\right)^{1.5} & 0 \le S_o < 0.03 \\ 4\left(\frac{d_{90}}{d_{30}}\right)^{0.2} \frac{h^{1/6}}{n\sqrt{g}} min(S_o, 0.2)\theta_i^{0.5}\left(\theta_i - \theta_{cr,i}\right) & S_o \ge 0.03 \end{cases}$$

where S_o is bed slope; θ_{cri} is critical dimensionless bed shear stress of *i* th fraction; θ_i is the dimensionless bed shear stress of *i* th fraction.

3.1.3 Suspended load transport

Suspended load transport was calculated by solving 2D depth-averaged advection-diffusion equation:

$$\frac{\partial hS_i}{\partial t} + \frac{\partial huS_i}{\partial x} + \frac{\partial hvS_i}{\partial y} = \frac{\partial}{\partial x}\left(\varepsilon_x h \frac{\partial S_i}{\partial x}\right) + \frac{\partial}{\partial y}\left(\varepsilon_y h \frac{\partial S_i}{\partial y}\right) + S_{E,i} - S_{D,i} \tag{6}$$

where S_i = volumetric suspended load concentration of the *i*th size class; ε_x, ε_y = turbulent diffusion coefficients of sediment in the *x* and *y* direction; $S_{E,i}$ = entrainment flux of sediment of the *i*th size class; $S_{D,i}$ = deposition flux of sediment of the *i*th size class; both fluxes is calculated by

$$S_{D,i} = \omega_{f,i}S_{a,i}, \; S_{E,i} = F_i\omega_{f,i}S_{ae,i} \tag{7}$$

where $S_{a,i}$ is the near-bed concentration at the reference level which refers to the depth of the sheet flow layer; $S_{ae,i}$ is the near bed equilibrium concentration at the reference level determined by the empirical equation of van Rijn (1984).

3.1.4 Bed level change

Bed elevation was updated based on simulated bed erosion or deposition at each grid, by

$$\frac{\partial z_b}{\partial t} = \frac{1}{1-p}\sum_{i=1}^{N}\left[\frac{(q_{bi} - F_i q_{b*i})}{L_i} + S_{D,i} - S_{E,i}\right] \tag{8}$$

where N is the number of grain size fractions; the values of the parameters in the right side are calculated according to the equations in previous sections.

3.2 Experimental design

To resolve the research questions, a series of experiments with different scenarios were designed in Table 1. The 1999 FLOOD (R1) was simulated to validate the model performance during real-world events, before then running the experiments. The experiments (R2 – R4) were modelled for understanding morphological records during multiple flood events. The experiments (R5 – R9) were designed in order to quantitatively assess the changes in conveyance capacity of the channel and cumulative effects on flood hazard.

Table 1. Experimental scenarios design

Runs (R)	Inflow	Bed geometry	Explanation	Purpose
R1	the 1999 FLOOD (flood_1)	original bed	with bed change	model validation
R2	flood_2	adjusted bed by flood_1	with bed change	
R3	flood_3	adjusted bed by flood_2	with bed change	morphological imprint
R4	flood_4	adjusted bed by flood_3	with bed change	
R5	the 1999 FLOOD	original bed	fixed bed	
R6	flood_1	adjusted bed by flood_1	fixed bed	
R7	flood_2	adjusted bed by flood_2	fixed bed	quantification of channel capacity, and flood hazard effects
R8	flood_3	adjusted bed by flood_3	fixed bed	
R9	flood_4	adjusted bed by flood_4	fixed bed	

Regarding the model simulations, the river channel was discretised by 1090×288 grids with the grid size being 8×8 m^2. The upstream inflow boundary is the hydrograph shown in Figure 2, and the downstream, left and right boundary are set to be open. For each run, we assumed that Manning's roughness was not affected by channel adjustment. The Manning-Strickler equation $n = 0.038d_{90}^{1/6}$ was used to estimate the value of Manning's roughness. The depth of the eroded bed is considered to be unlimited during the flooding. Sediment material in the channel was composed of various grain-sizes from fine granule to coarse boulder. Three size classes were considered in this study: granules (d_{50} = 2.8 mm), cobbles (d_{50} = 105 mm) and boulders (d_{50} = 400 mm). It was assumed that the outburst flood was initially 'clear water'. In-channel erosion and deposition was the main sediment activity that the flood induced. A variable time step Δt based on a constant Courant–Friedrichs–Lewy (CFL) number, adapted to local flow conditions, was used to maintain the model stability. The model (in)sensitivity to some parameters such as mesh size, manning's roughness, parameterisation of grain-size and choice of sediment transport formulas has been analysed and detailed in Guan et al. (2015b).

4. RESULTS AND DISCUSSION

4.1 Morphological response to the 1999 flood

Although the 1999 flood has been investigated by some researchers (Russell et al., 2000; Sigurdsson et al., 2000; Staines et al., 2014), but an exact record and quantification of flood information has been unattainable. Therefore, in this study we compared our modelled results with palaeocompetence measurements presented by Staines and Carrivick (2015) and with the slope-area reconstructions by Russell et al., (2010). The comparison is given in Table 2. It shows that the differences exist between each other for the hydraulic factors because of the high uncertainty of the data source. Whilst the modelled arrival time of peak flow to the bridge is about 1 h 13 min, which in general agrees with the recorded time by Sigurdsson et al. (2000) and with the reconstructions by Staines (2012).

Table 2. Comparison of reconstructed hydraulics at the glacier terminus

	Velocity (m^3s^{-1})	Flow Depth (m)	Time to peak in the bridge (h)
Point measurements (Russell et al., 2010)	13 (mean)	7.6	~1 h
Slope-area reconstructions (Russell et al., 2010)	~5 - 7	3.3 – 4.8	N/A
Sigurdsson et al. (2000)	N/A	N/A	~1 h
Modelled results at peak	8.5	3.81	1h 13 min

As the DEMs before and after the 1999 flood were reconstructed, the modelled river channel changes were compared to difference of DEMs (DoD) before (1996 DEM) and after (2001 DEM) the flood so as to verify the capability of the model in predicting geomorphological changes. The comparison is demonstrated in Figure 3. It should be clarified that: (1) the time scales of DoD and modelled changes in riverbed are different; the time interval is ~ 5 years for DoD, while it is only 6 hours for modelled result (Staines and Carrivick, 2015); (2) sediment materials from upstream glacial areas were likely brought to downstream river, but this was not quantified; (3) the river channel is complex but its parameterisation for model is in general simplified. Considering a series of uncertainties, we found from Figure 3 that the modelled channel changes are in general agreement with the DoD, which is acceptable particularly without any parameter calibrations. We considered the result is good because both erosion zones and deposition zones were reasonably predicted by the model and the modelled result shows very similar spatial pattern with DoD. For example, in the seven highlighted circle zones, the location and magnitude of bed changes are properly simulated. The uncertainty due to dataset and model parameterisation inevitably leads to some discrepancies. The measurements show that the riverbed changes in a wider area. The mean differences between the two (Figure 3c) are in a range of (-0.78m – 0.92m), which means only two-boulder diameters (diameter of a boulder is 0.4 m). Overall, the present model reasonably reconstructs the full processes of the 1999 flood including inundation and geomorphological changes.

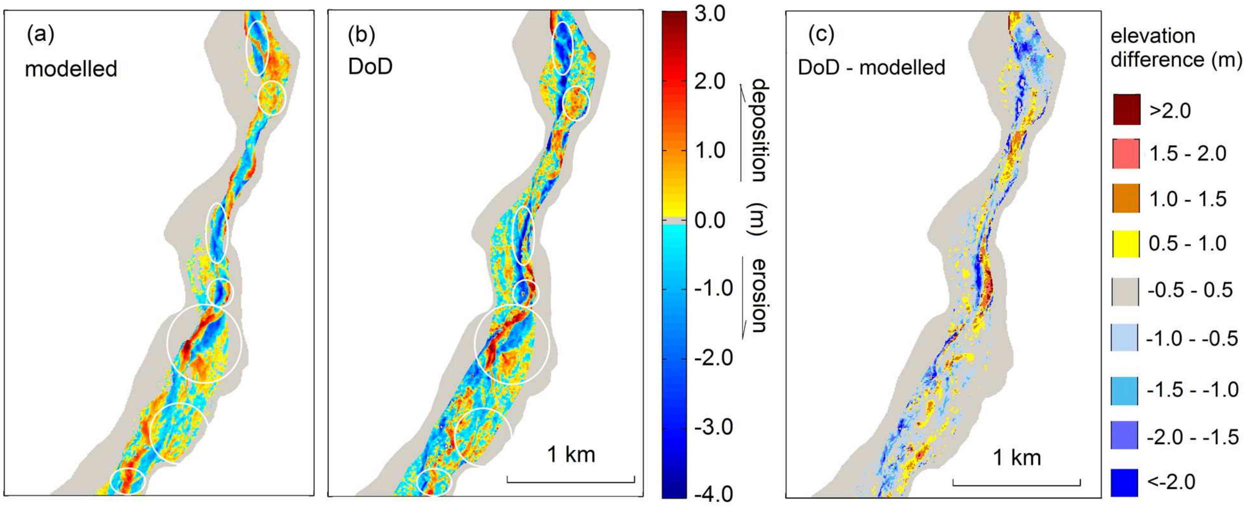


Figure 3. The modelled channel changes, and the DoD of changes in river morphology

4.2 Imprint of a series of floods in river morphology

We have questioned what the morphological records are due to multiple extreme floods of similar magnitude. To answer this question, R1 – R4 (Table 1) was performed. Figure 4 shows the modelled changes in river morphology after flood_1 (R1), flood_2 (R2), flood_3 (R3) and flood_4 (R4), and channel adjustment to these floods in two cross-sections (CS3 and CS7). Intuitively, it is clear that more floods aggravate more riverbed erosion. For example, in the circular highlighted area, the main channel was progressively scoured and more severely with each flood in the sequence. As shown in the representative cross-sections, the channel incision occurred both vertically and longitudinally so that the depth and width of the main channel enlarged due to the series of extreme floods. This channelling has been corroborated in many river cases based on long-term field observation of morphological activities (e.g.(Nardi and Rinaldi, 2015; Wyżga et al., 2015)). The modelling of this study evidences that floods generally lead to some imprints in river morphology and the changes will be exacerbated with more flooding.

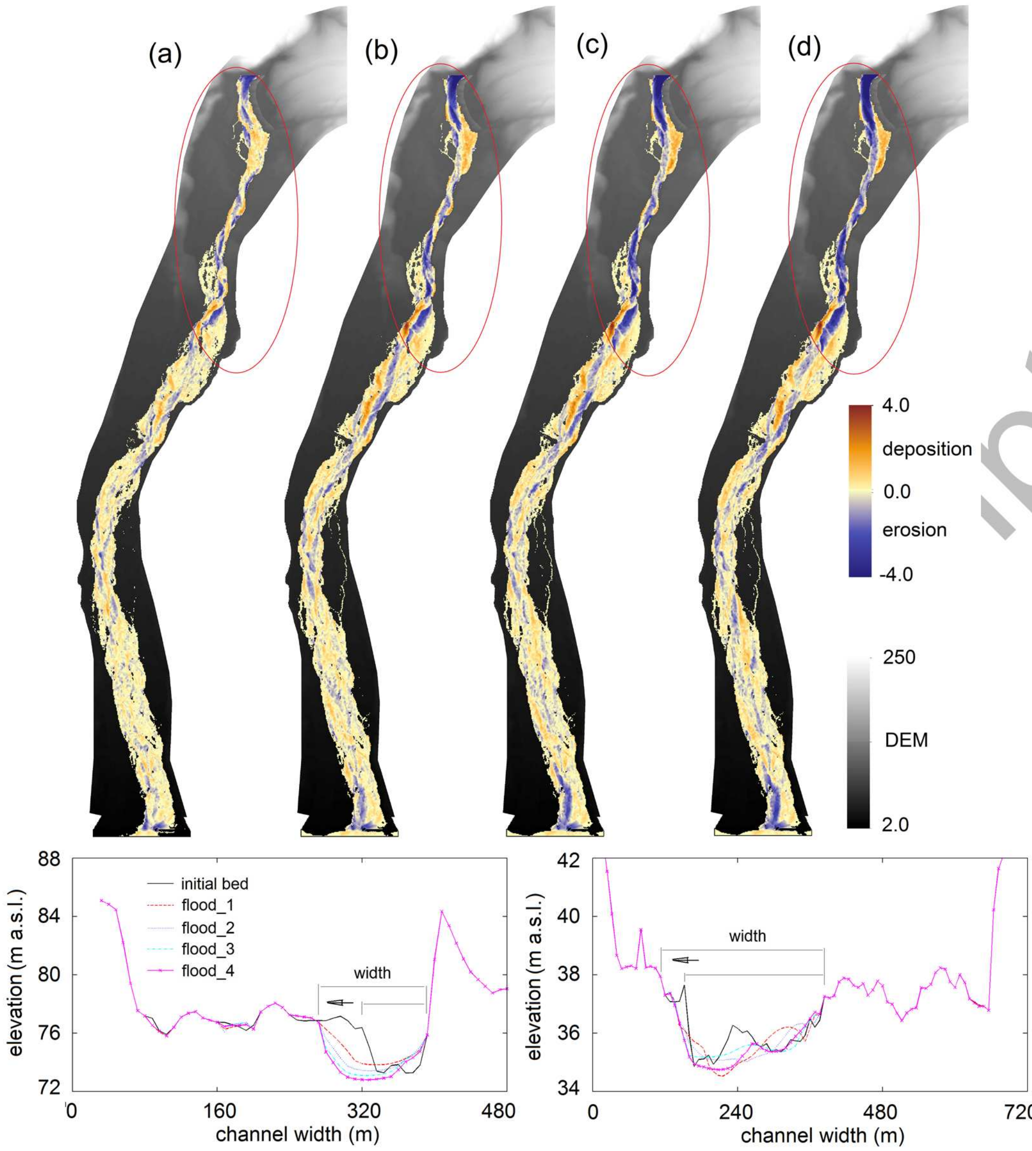


Figure 4. The modelled channel changes after the four successive flood events (a) flood_1, (b) flood_2, (c)) flood_3, and (d)) flood_4, and channel adjustment during the four events in (e) CS3 and (f) CS7. Note: the positive value means deposition, the negative value represents erosion.

From a temporal viewpoint, Figure 5 plots the changes of total net erosion, as well as the erosion and deposition volumes and rates in the whole channel. Surprisingly, we found that even with the same input hydrograph, successive floods led to significantly different changes in sediment volumes and rates for each flood. This raised some interesting points about morphological responses to flooding that previous studies have not detected. The most salient points include that:

(1) Erosion volume increases with subsequent floods (Figure 5a)

(2) Deposition volume decreases slightly with subsequent floods (Figure 5a).

(3) Net elevation changes in the river channel imply that successive floods remove sediment from the bed, i.e. net erosion occurs in the river channel. This commonly occurs in outburst floods: for example Baynes et al. (2015b) reported that erosion during extreme flood events dominates the landscape evolution in Iceland.

(4) Comparing the temporal pattern of erosion and deposition *within* a flood, for successive events, reveals that high erosion occurs during the peak period of the flood. Interestingly, deposition also occurs at peak flow because bedload plays an important role in the flooding. Physically, more bedload is induced into motion from the bed during peak flow and then re-deposit within an equilibrium length. However, both erosion volume and rate are larger than the deposition because of the fact that a majority of suspended load and a portion of bedload were transported within the floodwater.

(5) The majority of channel adjustments (56 % for erosion and 91 % for deposition) take place during the first flood.

(6) Channel changes during subsequent floods become weaker, for bed deposition in particular (Figure 5b). Deposition volume and rate sharply decrease from flood_1 to flood_4, and it appears to be stabilising.

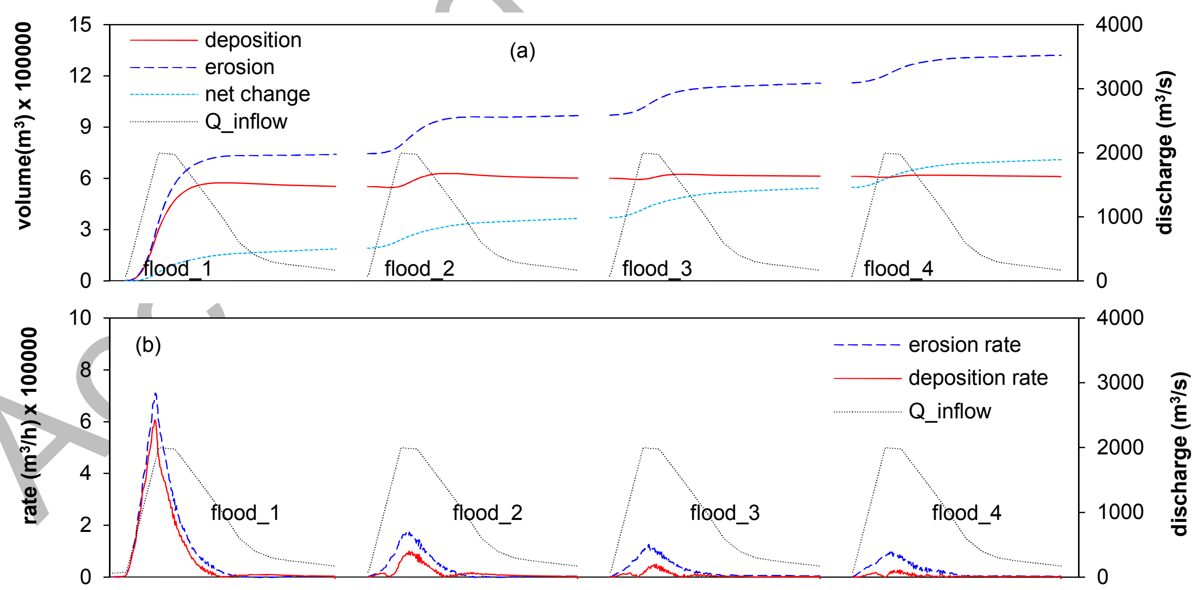


Figure 5. The temporal evolution of: (a) net changes and accumulated erosion and deposition volumes in the river channel, (b) the erosion and deposition rates

In combination these six observations signify that a morphological response to extreme floods is to adjust a river channel towards a more efficient propagation of floodwater. Bed response to flood_1 is the most severe, or significant, because of the riverbed is far from an equilibrium state. Once morphological responses, particularly channel straightening, widening and gradient smoothing proceed, bed response becomes more slight even with the same inflow hydraulics.

4.3 Impact of channel adjustment on conveyance capacity

Flood conveyance represents the ability of a river channel to convey floodwater. The capacity of a channel is typically assumed to be stationary in flood risk analysis and channel design engineering. However, there is evidence that the changes in river morphology and sediment supply are leading to increase and/or decrease of conveyance capacity of a river channel and influence flood frequency (Lane et al., 2007; Lane and Thorne, 2007; Slater et al., 2015; Stover and Montgomery, 2001). The previous section of this study has demonstrated the geomorphological activities within the river channel during multiple floods, such as in-channel scour and sediment re-distribution. Yet it is still unclear what impact of channel adjustments on the flood conveyance is. To answer this question, we simulated five scenarios (R5 – R9 in Table 1).

Using the flood information modelled by the 2D full hydrodynamic model, we calculated the average stage-discharge rating curves at nine cross-sections along the channel and plotted them in Figure 6. It indicates that flood conveyance capacity increases remarkably at some cross-sections such as CS1, CS2, CS3 and CS7, whilst it changes with relative small magnitude at some cross-sections such as CS4, CS5, CS6, CS8 and CS9. It is clear that changes in flood conveyance are attributed to channel adjustments caused by the floods. Overall, we found that the stage appears to be decreasing for a given discharge under the conditions of considering river morphological changes, despite the fact that the changes in magnitude are different at the nine cross-sections. This means that the extreme floods increase the conveyance capacity of the channel to a certain degree in comparison to the original channel. In general, channel aggradation may lead to a reduction of conveyance capacity of a channel, and conversely, bed degradation should results in an increase of flow capacity (Lane et al., 2007; Slater et al., 2015). It is simple to understand the reason from a viewpoint of cross-section. However, river flows are a dynamic process as a whole, thus its conveyance capacity is not only related to the circumstance of one cross-section, but also affected by changes in a reach segment.

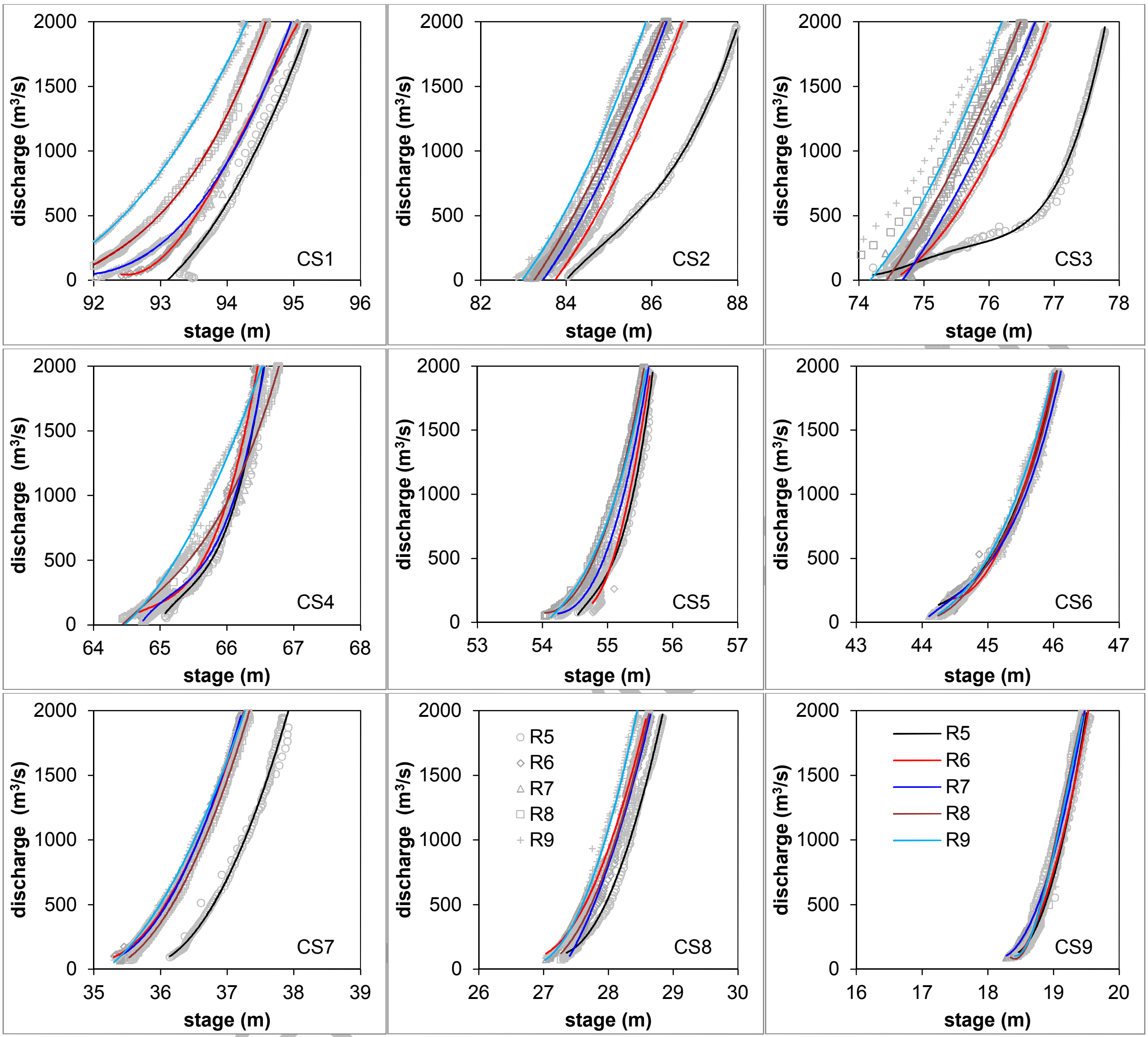


Figure 6. Stage-discharge curves in the nine selected cross-sections during multiple extreme floods

We quantified the areas of changes in bed elevations along with the river channel and plotted the results in Figure 7. It shows that bed erosion dominates in the whole channel (684 in 1090 cross-sections are eroded after flood_1, and after flood_4 it reaches 818 in 1090). In CS2, CS3 and CS7, considerable erosion occurs during flood sequence so that the increase in conveyance capacity is understandable. We can see that in CS1 the water stage for a given discharge is also significantly decreased from the original bed to the adjusted bed after flood_4, however, the net changes after flood_1, flood_2 and flood_3 in CS1 are shown to be aggradation, and then it becomes erosion after flood_4. This finding reveals that bed aggradation in some places does not necessarily cause a decrease of flow capacity of a cross-section; real circumstances of its neighbour areas have equally important impact on flood stages. The same situation happened in CS6 where channel aggradation occurs but the flow capacity in CS6 only has slight changes.

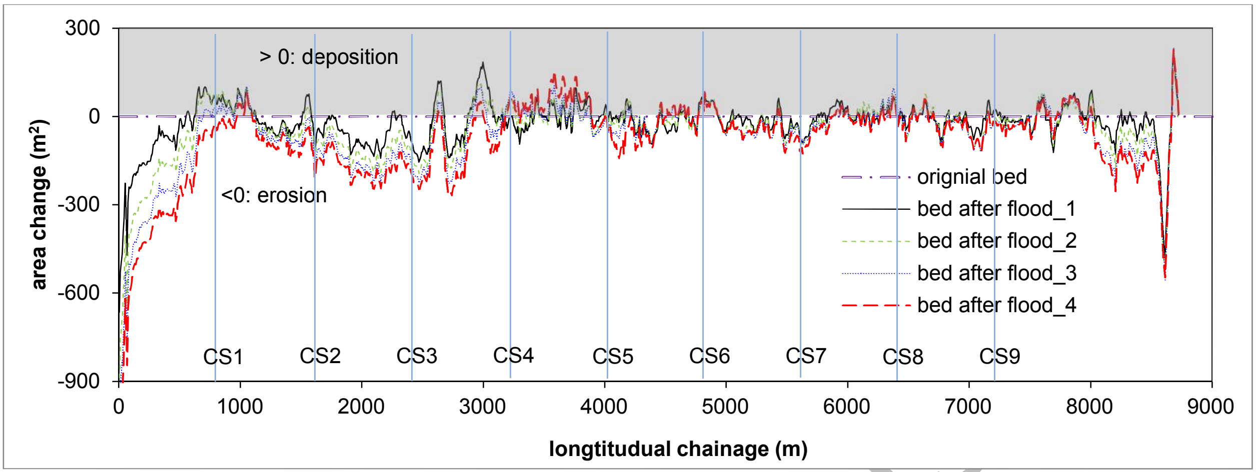


Figure 7. The area of channel changes along with the river channel after each flood

It was found that the overall trend of channel capacity at the nine cross-sections appears to be increasing under the conditions of either degradation or aggradation. As shown in Figure 5a, we may notice that as a whole, the river channel was eroded by multiple extreme floods and the net erosion aggravates along with more flooding. The fact that large account of sediments was washed away from the channel must lead to an increase of conveyance capacity of the whole river segment in flood. Our results suggested that changes in river morphology due to extreme floods is a significant driver of channel conveyance capacity, even though the floods are a short period of time, and that it is a better solution to assess the flow capacity from a reach-scale, not just from a cross-section.

4.4 Effects of changes in riverbed on flood hazard

Changes in riverbed play an important role in affecting flood hazards. Therefore, it is crucial to better understand sediment transport and bed elevation changes during floods, especially multiple floods at a same site. For a single extreme flood, Guan et al. (2015b) has given clear evidence of bed changes and significant effects of sediment transport on flood dynamics such as accelerating flood routing, altering flow velocity field and water stages. We next explored the spatial effects of river morphological changes during multiple floods. Figure 8 demonstrates the spatial distribution of modelled water depth at a reach segment when the peak floods over the original channel and the adjusted channels after each flood. It clearly shows that the water depths differ from each other both

392 in terms of distribution and inundation extent. The thalweg line is modified due to changes in channel
393 morphology after the floods. With more flooding, the main flood pathway appears to be smoother,
394 more straight and wider. All of these adjustments indicate that river channel is scoured and sediment
395 is re-distributed towards more efficient to flood propagation.

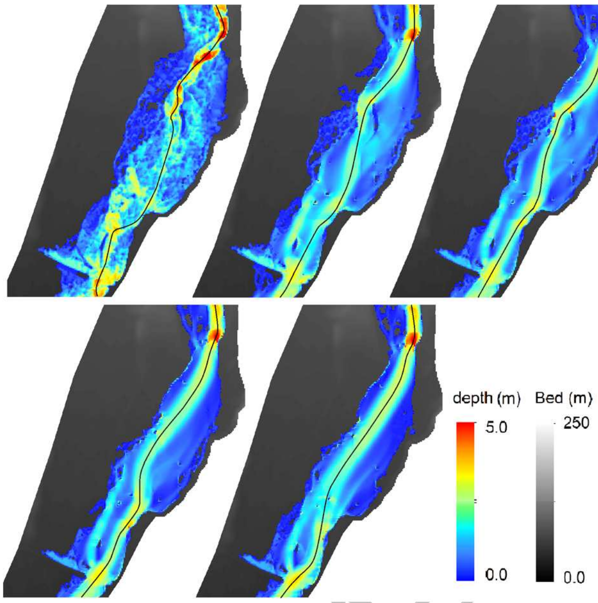
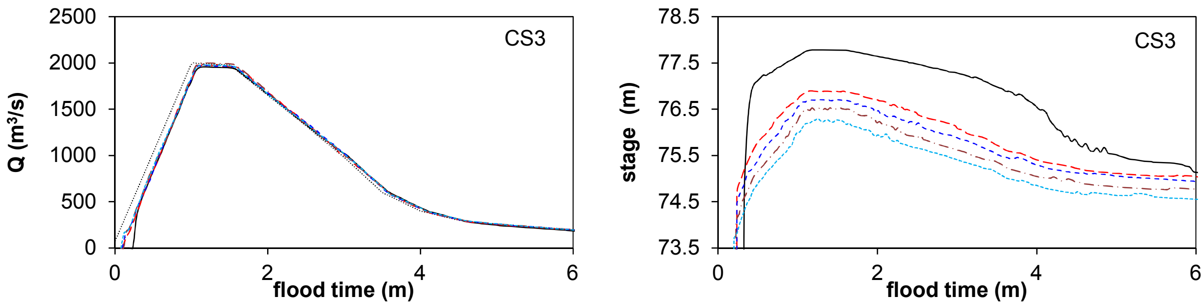


Figure 8. Water depth at the peak discharge for (a) R5, (b) R6, (c) R7, (d) R8 and (e) R9

To further elaborate the impacts of changes in river morphology on flood hazard, we plot the discharges and the average stage during each flood at CS3 and CS7 in Figure 9. Firstly, it was found that floods routing through an altered river channel propagated faster than the flood over an unchanged bed. For example, the difference of the flood arrival time reaches ~12 minutes in CS7. As noted above, this is attributed to the fact that flooding plays a role in scouring the bed to find its pathway. However, the latter floods are not accelerated too much even though more flooding occurs with changes in channel morphology. In alignment with the stage-discharge curves, the averaged stages of the five scenarios (R5 – R9) at CS3 and CS7 differ from each other significantly. For the scenario R5, the stage is over 1 m higher than others at CS3, and it is also over 0.6 m at CS7.



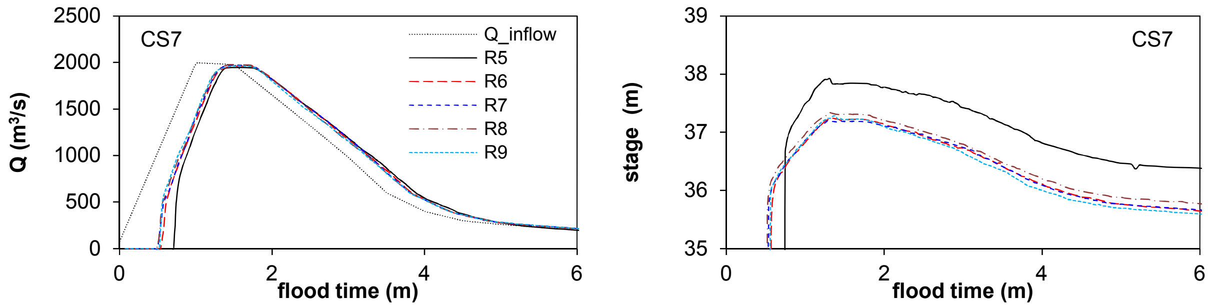


Figure 9. Flow discharge and stage over the flooding (R5 – R9) at CS3 and CS7

In general, flood stage is considered as an important factor for flood hazard assessment because the stage level directly decides the inundation extent and the magnitude of overbank flows during a flood. Figure 10 plots the maximum water stage along with the river channel during the flood over the five scenarios. It shows that the stage is reduced with more net in-channel erosion throughout the river channel, particularly at the upstream reach (Figure 10a). At the mid-downstream reach, the stage decreases in a relatively smaller extent. This coincides with the fact that the bed is scoured more severely in the upstream reach but slightly in distal reaches as shown in Figure 4. The overall reduction of maximum water stage will admittedly result in subsequent changes in local and downstream flood hazard. At CS3 (upstream reach) and CS7 (mid-downstream reach), Figure 11 demonstrates that water stages with a same given discharge (1965 m^3s^{-1}) decrease remarkably because of channel changes during multiple floods. Therefore, this implies that overbank flows to floodplain (where applicable) must be reduced thereby resulting in a decrease in storage of floodwater. A comparison of stage-discharge curves between original bed and adjusted bed after a series of floods (Table 3) indicates that flood-induced in-channel erosion dramatically improved the conveyance capacity of the channel, thereby resulting dramatic impacts on flood hydraulics. With a same stage, the discharge at CS3 increases by 1.79 times, and it increases by 1.39 times at CS7. In other words, flood frequency in the river channel is changed due to changes in river morphology.

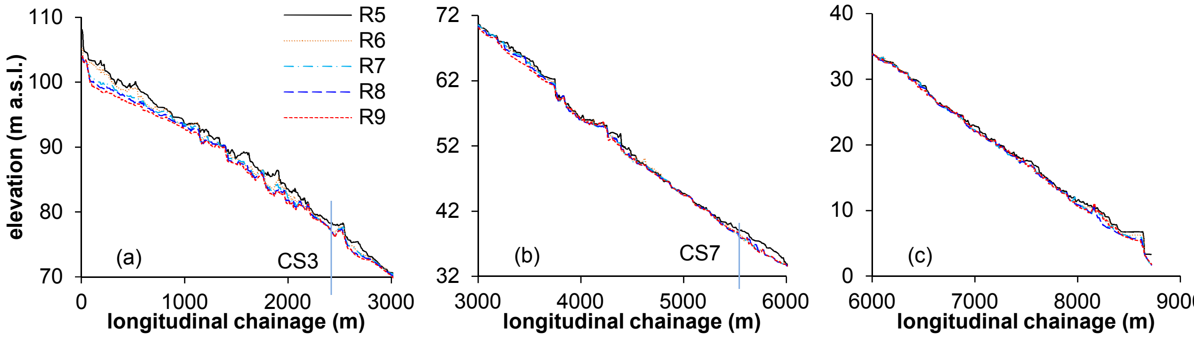


Figure 10. Maximum water level for the flood over the five bed scenarios

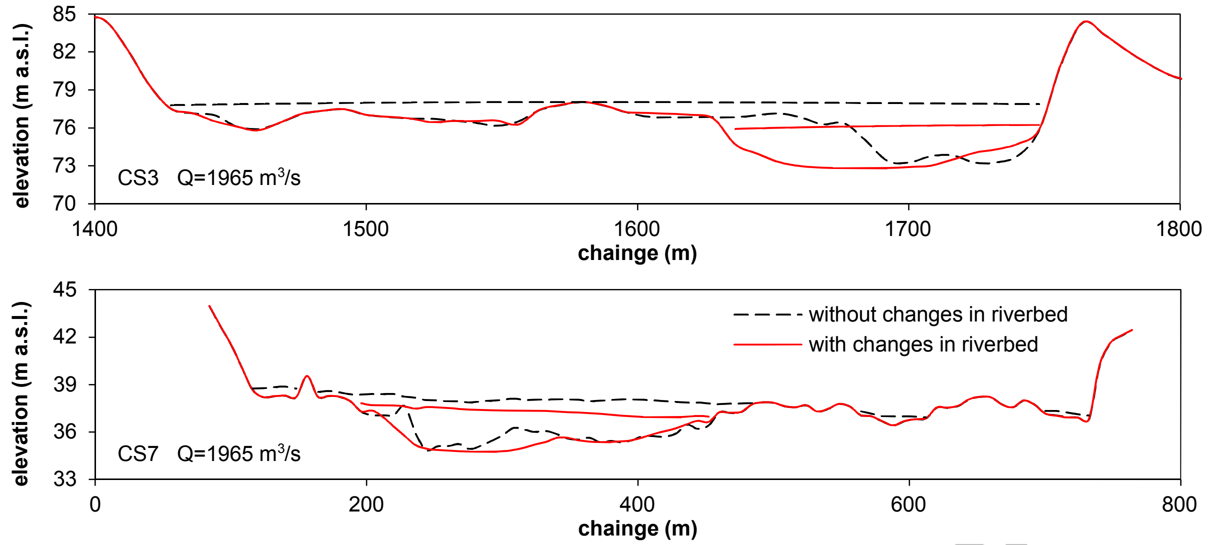


Figure 11. Water stage and bed profile of the original bed and adjusted bed after multiple floods for a given discharge 1965 m^3s^{-1}

Table 3. Changes in discharge with the same stages at CS3 and CS7 before and after in-channel adjustment due to a series of floods

	Original channel		Adjusted channel due to floods	
	Discharge (m^3s^{-1})	Average stage (m)	Discharge with the same stage (m^3s^{-1})	Increase in discharge
CS3	1965	78.0	3402	×1.73
CS7	1965	38.0	2734	×1.39

From a spatial point of view, the overlay of flood areas for three scenarios in Figure 12 clearly shows that the in-channel inundation area decreases with a same magnitude flood with agglomerative bed changes (net in-channel erosion), such as the highlighted zone1 and zone2 in Figure 12. Conversely, the net in-channel aggradation can increase flooding inundation extent as reported by Lane et al., (2007). Apparently, flood hazards were significantly affected by channel morphological changes. The effects can be either positive or negative, which is scale- and case-dependent, such as the location of the eroded channel, and floodplain.

The above results raised a number of key understandings in terms of the potential effects of changes in riverbed on flood hazard, including:

(1) Channel incision due to in-channel scour may lead to disconnect the channel from its floodplain (where applicable) resulting in a reduction of floodwater storage, particularly with further more flooding;

(2) The resultant smaller overbank flows and inundation extent implies less flood hazard in the reach where in-channel erosion severely occurs, because the bankfull water depth considerably increases for a given discharge.

(3) However, at a wider scale, the decrease in overbank flows and floodwater storage must lead to more floodwater propagate to downstream, i.e. increasing flood hazard frequency in downstream reaches or areas.

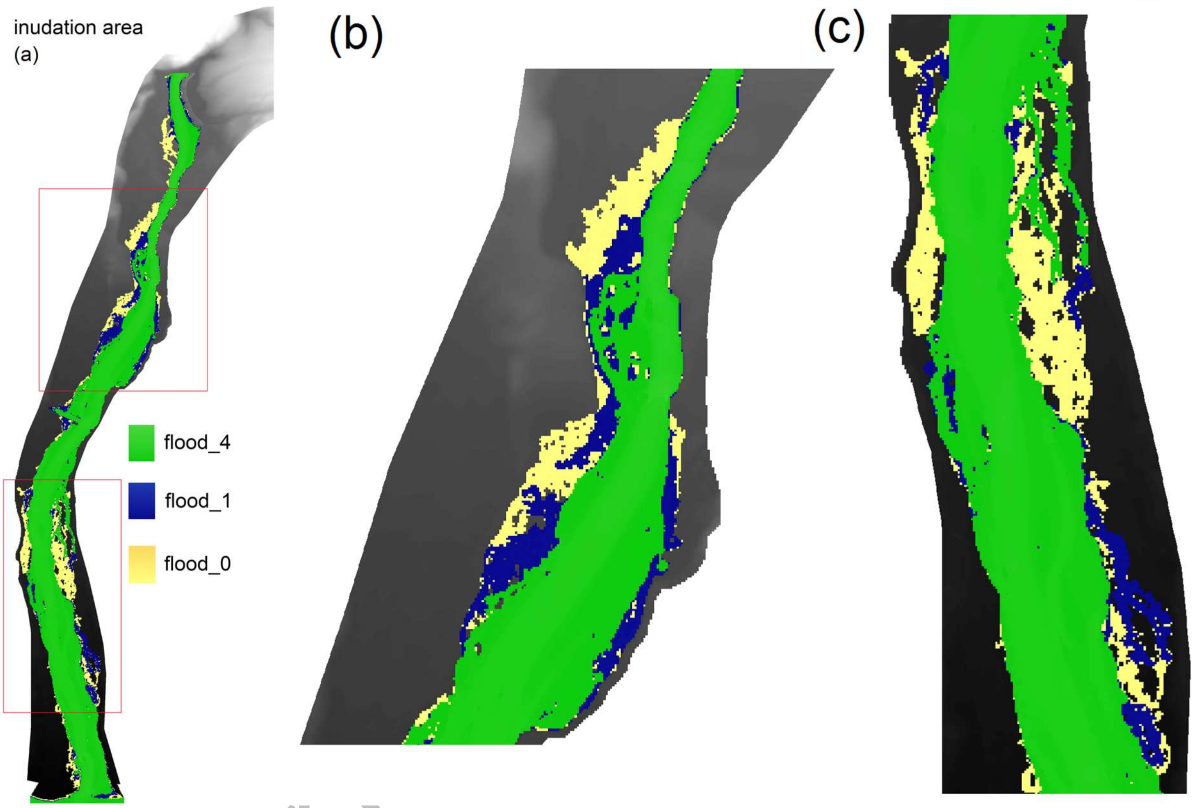


Figure 12. (a) the inundation extent in the river channel; (b) highlighted zone 1; (c) highlighted zone 2

Support for these effects of river channel morphology changes comes from a number of studies that have investigated the effects of long-term channel incision on flood hydraulics (Shields Jr et al., 2010; Wohl, 2004; Wyżga et al., 2015). The case here definitely belongs to channel incision due to in-channel scour caused by extreme floods. Overall, it is emphasised that extreme floods have consistent effects on in-channel scour. The lowering of channel bed and water stage can greatly change local and downstream flood hazard that is influenced not only by flow magnitude, but also by changes in river morphology. This suggests that river channel adjustment caused by flooding can lead to complex impacts on flood inundation by altering the flow pathway in the channel. At a wider viewpoint, it is reported that remarkable channel adjustments have taken place in many rivers throughout the world over the last few decades, not only because of multiple flood events, but also in relation to human interventions (e.g. (Abate et al., 2015; Bollati et al., 2014; Campana et al., 2014; Raven et al., 2009; Scorpio et al., 2015)). There is no doubt that these changes in river morphology will lead to significant effects on flow dynamics and inundation during flooding. Effects of both extreme floods and human interventions on river morphology play crucial roles in subsequent flood inundation and thereby flood hazards. This widely emphasises a point raised in the paper: inundation modelling without considering changes in river morphology, as many studies have done, might be open to question for flood hazard assessment.

4.5 Wider applicability

Although this study conducted a set of hypothetical experiments, the resultant findings have wider implications for understanding geomorphological changes during floods and thus for flood hazard analyses. Firstly, extreme floods are a major cause of geomorphological changes in rivers despite the fact that the time scale of individual floods may be very short. Additionally, the changes can be cumulative with a series of big floods. Secondly, given that multiple floods from single sites is very common, the improved understanding of flood-riverbed interactions in this study is applicable worldwide. Thirdly, the significant effects of river morphological changes on flood dynamics found in this study raise a key point for flood risk management: flood hazard is not just from water, but also from flood-driven sediment and/or debris. Any changes in a floodwater pathway (i.e. river channel) can further lead to different propagation time of floods, inundation extent and water stage both locally and downstream. Such consequences of multiple floods and associated geomorphological changes in river channels cannot be neglected, yet frustratingly presently flood risk management tends to be based on assumptions of ‘clear water’ and a fixed bed. Fourthly, changes in a river channel can typically include localised in-channel scour, aggradation from upland sediment supply, blockage by large debris or large wood, and channel adjustments due to human interventions such as dredging and dumping. Therefore, in future flood hazard assessments should consider both floodwater and its sediment transport and its associated morphological changes. We suggest that the quantified ‘flood memory’ also provides a reach-scale basis for quantifying the effects of sediment/debris from natural flood management, and river restoration might be necessary to effectively manage flood hazard in a river and its floodplain.

5. CONCLUSIONS

This study explored sediment transport processes during a repeated series of hypothetical extreme floods. An emphasis was placed on the cumulative effects of changes in river morphology on conveyance capacity of a channel in flood and thereby on flood hazards. We have shown that during a series of extreme floods, in-channel erosion occurred more severely in the upper river reach than in the distal reaches (Figure 4). The severe in-channel erosion played an important role in adjusting the river channel to becoming more efficient for flood propagation. Net erosion in the river channel implied that the successive floods consistently removed sediment away from the bed. The majority of channel adjustments took place during the first flood, and channel changes during subsequent floods weakened in intensity and coverage. In other words, the response of the river channel geometry to latter floods approached an equilibrium state based on the imprint of former floods.

Our results have shown that both erosion and deposition occurred along with the river channel. This is not surprising but it reinforces the concept that channel capacity increased at some cross-sections and decreased at others. However, overall in-channel scour and sediment re-distribution dramatically increased the conveyance capacity of the river channel in flood in comparison to the original channel. This suggests that it is a better solution to assess the flow capacity from a reach-scale, not just from a cross-section as many studies have done, since flood propagation and transient hydraulics are a fully dynamic process.

We have quantified how changes in river channel geometry due to an extreme flood can have a dramatic impact on subsequent flood hydraulics. Firstly, the propagation of floodwater is speeded up, particularly in the channel adjusted by the first flood. Secondly, the thalweg and longitudinal profile of maximum water depth becomes smoother due to changes in channel morphology. Thirdly, water stage for a given discharge decreases as the conveyance capacity increases. Finally, inundation extent within the river channel also decreases because of the lowering of bed elevation and smoothing of the longitudinal profile.

These changes in flood hydraulics result in significant impacts on flood hazards locally and at a wider scale. The impacts include a decrease in the frequency and extent of overbank flows and floodplain inundation (where applicable) and at a wider scale this means that the resultant increase in channel capacity and the loss of floodwater storage in its floodplain may reduce downstream flood attenuation and increase flood hazard to downstream reaches and areas. This suggests that special efforts should be made to stabilise upper river reaches for retarding severe bed erosion, and to downstream reaches for flood risk assessment.

Overall, this study reinforces the concept that the effects of river channel geometry adjustments on flood hazards are significant and multi-faced. Therefore, properly consideration of changes in river channel geometry during flooding must be made in order to accurately and robustly assess flood risk. We suggest that modelling of floodplain inundation and extreme flooding cannot simply assume river channel consistently unchanged.

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