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How does sediment supply influence refugia availability in river widenings?

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ABSTRACT

Habitats that mitigate the effects of a disturbance event (e.g. flood) are referred to as refugia. Their occurrence in heavily impacted river systems is often limited, and their restoration rarely pursued. This paper presents the results of a combined laboratory and numerical modeling study to assess flood refugia availability to mobile aquatic organisms in the context of river restoration and dynamic river widening. We used a calibrated 2D hydrodynamic model based on eight topographies obtained in laboratory experiments to assess refugia availability by analyzing the hydro-morphological conditions under varying sediment supply. Overall, sediment equilibrium sustains more complex hydro-morphological conditions with low bed shear stress zones being maintained during elevated discharges. Furthermore, our results suggest that the floodplain is an important potential refuge that becomes accessible for discharges with a return period of approximately one year. Conversely, sediment deficit results in a homogeneous flow field with steadily increasing hydraulic forces for high flows and impaired lateral connectivity except for very large flood events of a 30- to 100-year return period. Dynamic river widening implemented in a channel with sediment equilibrium conditions as opposed to a sediment deficit is thus more likely to provide flood refugia.

KEYWORDS

river restoration, dynamic river widening, hydrodynamic modeling, habitat availability, flood refugia

1. Introduction

An intricate and episodically changing habitat mosaic is a vital part of a natural river system (Stanford et al. 2005). The habitat mosaic is strongly influenced by the diverse morphologies a river can assume and these are in turn the result of interacting geomorphological, hydrological, and biological processes (Castro and Thorne 2019). These driving processes can be highly dynamic and act on various spatial and temporal scales (Poff et al. 1997; Gurnell et al. 2002; Wohl et al. 2015a). When the characteristics of a biotic or abiotic process, i.e. its intensity or duration, assume exceptional values, the process is referred to as a disturbance (Resh et al. 1988; Lake 2000). Examples of abiotic disturbances include floods (e.g. Lepori and Hjerdt 2006), droughts (e.g. Wood et al. 2010), or ice formation (e.g. Weber et al. 2013). The temporal variation of the disturbing force can be used to classify disturbance events into pulse, press, or ramp events (Lake 2000). Pulses are short, sharply delineated events, press events have an instantaneous start with the pressure subsequently maintained, and a ramp is a steadily increasing disturbing force.

A flood event is a major form of pulse disturbance and induces hydraulic stress as well as sediment and large wood transport in the affected river (e.g. Sedell et al. 1990; Lake 2000; Lepori and Hjerdt 2006; Czuba et al. 2018). Aquatic organisms may be displaced, injured, or killed (Lake 2000). However, the stream communities usually recover fairly rapidly (e.g. Lake 2000; Gjerløv et al. 2003; Lepori and Hjerdt 2006), indicating that a certain number of organisms can resist the disturbing force. Despite the apparent adverse effects of hydrological disturbance events for stream organisms, high magnitude flood events are a critical component of the natural flow regime and are vital for maintaining healthy ecosystem functioning (Poff et al. 1997; Richter et al. 1997).

Disturbance intensity may vary widely between habitats due to their spatial arrangement, substrate characteristics, or the acting hydraulic forces (Weber et al. 2013). Habitats that mitigate the effects of a disturbance on the biota and enable organisms to resist the disturbance are typically referred to as refugia (Lake 2000). They exist at various spatial and temporal scales (e.g. Sedell et al. 1990; Lancaster and Belyea

1997; Lancaster 2000) and their refugia function depends on various factors such as the disturbance type (e.g. Lake 2000), organism characteristics (e.g. Pearsons et al. 1992; Weber et al. 2013; Sueyoshi et al. 2014), substrate composition (e.g. Mathers et al. 2019), and hydro-morphological conditions (e.g. Lancaster and Hildrew 1993; Rice et al. 2007). Examples of local flood refugia are low bed shear stress zones (e.g. Lancaster and Hildrew 1993; Winterbottom et al. 1997; Lancaster 2000), specific turbulent eddies (e.g. Schwartz and Herricks 2005), marginal zones (e.g. Rempel et al. 1999), individual stones (e.g. Townsend et al. 1997; Matthaei et al. 2000), or woody debris (e.g. Palmer et al. 1996). On a larger scale, the floodplain (e.g. Matthaei and Townsend 2000; Schwartz and Herricks 2005), the hyporheic zone (e.g. Dole-Olivier et al. 1997; Stubbington 2012), or tributaries (e.g. Sueyoshi et al. 2014) may serve as refugia. It is commonly assumed that the closer a reach is to its natural state the more heterogeneous its habitats will be, thereby providing ample refugia opportunities under a range of hydrological conditions (Sedell et al. 1990; Pearsons et al. 1992; Gjerløv et al. 2003). In particular, a more morphologically and hydraulically complex reach will likely preserve low bed shear stress zones during periods of elevated discharge (e.g. Pearsons et al. 1992; Lancaster and Hildrew 1993).

However, extensive anthropogenic impact on fluvial environments has destructed or degraded habitats and disrupted ecosystem processes (Dudgeon et al. 2006). Rivers across the globe have become channelized, reducing the available habitat area and homogenizing the remaining habitats (e.g. Hohensinner et al. 2001). Damming, gravel extraction, and protective structures have further disrupted ecosystem processes associated with the flow and sediment regime (Kondolf 1997). At first, the flow regime received most attention (Poff et al. 1997) before it was recognized that other processes, such as sediment and large wood transport, play an equally important role in shaping fluvial ecosystems (Gurnell et al. 2002; Wohl et al. 2015a).

Sediment transport is a highly episodic and heterogeneous process because it depends on the interaction of bed shear stresses inflicted by near-bed hydraulic forces and the mobility of the sediment (Wohl et al. 2015a). In general, the larger the grain the higher the bed shear stress necessary to mobilize it. However, the entrainment process is further complicated by hiding and exposure effects (larger grains are more

exposed to the flow while hiding the smaller grains) that are especially relevant for the poorly sorted sediment of gravel-bed rivers (e.g. Wilcock and Crowe 2003). This leads to a diverse and dynamic mosaic of riverbed patches with active and inactive bed-load transport and different grain sizes (Nelson et al. 2009).

A river’s transport capacity is controlled by channel geometry, discharge, sediment characteristics and supply (Kondolf 1997). Sediment supply that matches the long-term (years to decades) average transport capacity of a river reach is defined as sediment equilibrium. Conversely, the long-term shortage of sediment supply compared to the river’s transport capacity is referred to as sediment deficit (Kondolf 1997). The geomorphic implications of sediment deficit on fluvial systems include channel narrowing and incision into the surrounding floodplain (e.g. Surian and Rinaldi 2003), surface coarsening (e.g. Dietrich et al. 1989), and the expansion of inactive patches, i.e. without frequent changes in river bed elevation and grain size composition (e.g. Nelson et al. 2009). Assuming that habitat units of higher complexity and dynamics are more likely to provide refugia (e.g. Sedell et al. 1990; Gjerløv et al. 2003), sediment deficit may, therefore, impair refugia availability and result in less resistant and resilient stream communities.

In recent decades, great efforts have been made to restore fluvial ecosystems (e.g. Bernhardt et al. 2005). However, lack of awareness for the complex biotic and abiotic interactions often lead to restoration projects that are overly focused on constructing specific structures and may fail to deliver long-term benefits (Beechie et al. 2010; Wohl et al. 2015b). Among others, Beechie et al. (2010) and Wohl et al. (2015b) argue for a restoration practice that emphasizes intact ecosystem processes and functions, such as focusing on the refuge function of certain habitats during disturbance events, as opposed to only residential habitats during mean flow conditions.

An example of process-focused restoration is dynamic river widening, a reach-scale measure to mitigate the effects of channelization (e.g. Rohde et al. 2005; Habersack and Piégay 2007; Williams et al. 2020). On one or both sides of the river channel, bank protection is removed within a limited reach to enable lateral erosion of the adjacent floodplain. During subsequent flood events, the channel is widened by hydrodynamic forces and thus the morphology tends to become more complex (e.g. Garcia Lugo et al.

2015; Williams et al. 2020). The erodible area is usually constricted due to nearby infrastructure. Hidden bank protection measures along the outermost boundary of the widening can limit lateral erosion. 'Dynamic' emphasizes the focus on the reactivation of morphodynamic processes in contrast to widenings that are constructed with a predefined width. Given that dynamic river widening is a reach-scale measure, the restoration effect may be linked to processes acting on a larger scale, for example the sediment supply (Wohl et al. 2015b; Rachely et al. 2018).

As field observations during flood events are very challenging, refugia availability during floods is less well understood than during drought. We therefore used a combined approach of laboratory and numerical modeling (cf. Vanzo et al. 2016) to study flood refugia availability in dynamic river widenings. We were specifically interested in the impact of the sediment supply level, in particular the effects of a sediment deficit versus sediment equilibrium on refugia availability for mobile aquatic organisms such as fish or invertebrates. From the laboratory experiments, we obtained high-resolution topographical data and information on bed-load transport processes. Using numerical simulations, we were able to determine the hydraulic conditions in great detail and extend the study to a wide range of discharges.

We addressed the following research questions:

- (1) Does the sediment supply level influence the inundation of the floodplain and therefore its availability as a potential flood refuge? We hypothesize that the morphological and topographical conditions developed under equilibrium sediment supply enhance the channel-floodplain connectivity (e.g. Hohensinner et al. 2003). In contrast, sediment deficit may lead to incision and the decoupling between main channel and floodplain (e.g. Surian and Rinaldi 2003).
- (2) Does the sediment supply level significantly affect the hydro-morphological conditions across a range of discharges? We hypothesize that a widening subjected to equilibrium sediment supply develops a more diverse morphology in contrast to a reach with a sediment deficit (Rachely et al. 2018). In consequence, the hydraulic variability may be higher and zones of low bed shear stress acting as refugia may persist during high flows (e.g. Lancaster and Hildrew 1993).
- (3) Does the sediment supply level modify the spatial extent of riverbed elevation

change and how does this affect refugia availability? We hypothesize that while equilibrium sediment supply invokes more extensive sediment relocation (Dietrich et al. 1989; Nelson et al. 2009) and thus aggravates the disturbance (e.g. Matthaei and Townsend 2000), the overall impact of equilibrium sediment supply on refugia availability is beneficial.

2. Methods

2.1. Case study Kander River

This study reproduces the general conditions in the Kander River, a subalpine gravel-bed river in Switzerland, as the case study example (Figure 1a). Prior to the channelization of the Kander River in the 20th century, its morphology was wandering to braiding, whereas it is now a single channel with a mostly flat bed. Within the scope of a long-term restoration program, several dynamic river widenings are planned to be implemented, and this study system therefore provides an opportunity to assess the restoration effect of such widenings.

Figure 1. (a) Location of the Kander River in the Canton of Bern, Switzerland, and (b) an example of a dynamic river widening at $N46^{\circ}39'14.755''$ $E7^{\circ}40'48.398''$ (★) in the year 2010. This widening developed in the year 2005 during a flood event with a return period of approximately 50 years (Sources: (a) FOEN (2014); swisstopo (2018), (b) © swisstopo).

An example of an already existing dynamic river widening can be found at $N46^{\circ}39'14.755''$ $E7^{\circ}40'48.398''$, where a flood with a return period of approximately 50 years in 2005 caused lateral erosion over a length of approximately 600 m (Figure 1b). In this river section, the still channelized river has a slope of roughly 0.01, a bed width of 20-25 m, and banks protected by riprap or groins (FOEN 2015). The grain size distribution is that of a poorly sorted gravel-bed river, with a geometric

standard deviation $\sigma_g = (D_{84}/D_{16})^{0.5} = 3.4$, and the characteristic grain diameters $D_{50} = 0.058$ m and $D_{90} = 0.188$ m (Huwiler and Stocker 2019). The flow regime of the Kander River is nivo-glacial (Weingartner and Aschwanden 1992) and the flood season lasts from May to August (FOEN 2017, see Table 1 for characteristic discharges).

Five characteristic discharges of the Kander River displayed as absolute values and multiples of the mean annual discharge Q_m (data series 1981-2017, FOEN 2017)

		Q (m ³ /s)	Q/Q_m (-)	Relevance
Annual discharge	Q_m	20.4	1	Baseflow conditions
Monthly discharge in June	$Q_{m,\text{June}}$	41.3	2	Baseflow conditions during the flood season
1.5 years return period	$HQ_{1.5}$	118.3	5.8	Bedforming discharge; corresponds to steady bedforming discharge (BQ) in the laboratory experiments
30 years return period	HQ_{30}	244	12	Large flood event; corresponds to the peak flow of the flood hydrograph (FQ) in the laboratory experiments
100 years return period	HQ_{100}	313	15.3	Large flood event

2.2. Laboratory experiments

Laboratory experiments investigating the dynamic river widening process were performed at the Laboratory of Hydraulics, Hydrology and Glaciology (VAW) at ETH Zurich. All hydraulic (FOEN 2017), sedimentological (Huwiler and Stocker 2019), and geometric (FOEN 2015) parameters were chosen to correspond to the Kander River. The laboratory setup consisted of a straight, trapezoidal channel with fixed banks and a mobile bed. The channel with a slope of 0.01 is 0.89 m wide and 31.6 m long. One bank of the channel can be partially removed to enable lateral erosion of the adjacent floodplain which consists of the same mobile bed material and longitudinal slope as the channel (Figure 2). Fixed banks delineate the widening area, representing hidden bank protection measures to limit lateral erosion. All data collected in the laboratory model were then upscaled with a Froude scale factor of $\lambda = 30$. In prototype scale, the width of the initial channel is 26.8 m, the widening can assume a maximum width of 107.4 m, i.e. a widening ration of up to 4, the length of the widening is 708 m, i.e. long enough to be hydraulically effective (Hunzinger 1998), and the maximum widening area is 63,800 m².

Figure 2. (a) Plan view and (b) cross-section of the laboratory experiment. The total widening area A_{max} (striped) is composed of the main channel area A_{ch} (striped, dark gray) and the floodplain area A_{fp} (striped, light gray). The straight inlet and outlet sections do not contribute to any spatial analyses in this study.

Two hydraulic conditions were tested: (i) a steady bedforming discharge representing long-term exposure to a morphologically relevant discharge, hereafter referred to as BQ, and (ii) a hydrograph with a peak flow of approximately twice the bedforming discharge representing a single, moderately large flood event, hereafter referred to as FQ (see Table 1 for equivalent discharge values). In addition to the hydraulic conditions, the sediment supply at the inlet was varied to model sediment equilibrium and sediment deficit conditions. Three sediment supply levels (SSL) were applied: (i) 100% of the initial channel’s transport capacity (TC), (ii) 60% of TC, and (iii) 20% of TC (hereafter referred to as SSL 100%, SSL 60%, and SSL 20%). The tests with SSL 100% were run twice, while all other experiments being run once, providing a total of eight topographies (Table 2). For example, the topography FQ_60% was obtained after a run with the FQ hydrograph as inflow and a sediment supply level of 60% TC.

Table 2. Eight topographies obtained from laboratory experiments and the proportion of channel (A_{ch}) and floodplain area (A_{fp}) relative to the maximum widening area of $A_{\text{max}} = 63'800 \text{ m}^2$. $A_{\text{ch}}/A_{\text{max}} = 0.32$ corresponds to no widening of the initial channel. BQ denotes topographies developed under steady bed-forming discharge and FQ under a flood hydrograph. The sediment supply level is indicated with the respective percentage. The sediment equilibrium tests were conducted twice.

Topography	$A_{\text{ch}}/A_{\text{max}}$	$A_{\text{fp}}/A_{\text{max}}$
BQ_100%.1	0.51	0.49
BQ_100%.2	0.39	0.61
BQ_60%	0.34	0.66
BQ_20%	0.36	0.64
FQ_100%.1	0.45	0.55
FQ_100%.2	0.39	0.61
FQ_60%	0.43	0.57
FQ_20%	0.42	0.58

To reproduce initial channel topography, all tests commenced with SSL 100% until an equilibrium between sediment input and output rates was established. In the sediment deficit tests, the SSL was then reduced to 60% TC or 20% TC, again until the sediment output rate equaled the input rate. A weak alternate bar pattern was formed in the straight channel for all SSL. The subsequent removal of the bank on one side enabled the lateral erosion processes to start. This phase was again continued until a transport equilibrium was established, and until the main channel width did not significantly change anymore. These phases were run with the steady bedforming discharge (BQ), before a single hydrograph (FQ) was run at the end of each series.

The flume was drained at 3-hour intervals and the dry bed topography recorded with a terrestrial laser scanner (Leica ScanStation P15) at a measurement accuracy of $\pm 0.06 \text{ m}$ (upscaled value). A Digital Elevation Model (DEM) was subsequently obtained by interpolating the recorded point cloud onto a grid with a spacing of $0.3 \times 0.3 \text{ m}$ (upscaled value). To separate the main channel area A_{ch} from the floodplain

area A_{fp} (Figure 2) for each DEM, a plane of the type $P = aX + bY + c$ was fitted to the floodplain. The plane was offset by $-2D_{90}$ which is larger than or equal to the 99th percentile of the error between floodplain and fitted plane. The topography was then separated into patches below and above this dividing plane. If there were multiple patches below the plane, the largest patch was identified as the main channel with an area of A_{ch} , and the remaining patches were assigned to the floodplain area A_{fp} .

To identify zones of bed elevation change between two subsequent scans, the DEM of Difference (DoD) was obtained by subtracting the respective DEMs. As each DEM was recorded with an error of ± 0.06 m (upscaled value), the propagated error of the resulting DoD is $(0.06^2 + 0.06^2)^{0.5} = 0.085$ m. Probabilistic thresholding with a Student's t-test and a confidence level of 95% was applied to identify areas of true change (Wheaton et al. 2010).

2.3. Numerical model

To collect hydraulic data with a high spatial resolution, numerical simulations were performed with BASEMENT (version 3.0), a freeware simulation tool for hydro- and morphodynamic modeling developed at VAW (Vetsch et al. 2020). We used the 2D model based on the depth-averaged shallow water equations for hydrodynamic simulations. Morphodynamic processes were not modeled, i.e. the bed was fixed for all simulations.

Every topography obtained in the laboratory experiments (Table 2) was transformed into a separate computational mesh with the BASEmesh plugin (BASEmesh 2020) for QGIS (QGIS.org 2020). To remove grain-scale roughness, the DEMs were first smoothed with the SAGA (Conrad et al. 2015) tool 'Mesh denoise' ($T = 0.7$, $n = 20$, $v = 50$) integrated in QGIS. The tool is based on an algorithm by Sun et al. (2007), which removes noise without the loss of sharp topographic features (Stevenson et al. 2010). The filtered DEMs were subsequently interpolated onto an unstructured computational mesh with 238,935 elements. The element sizes are distributed homogeneously across the entire perimeter, with a minimum element size of 0.11 m^2 , a mean element size of 0.32 m^2 , and a maximum element size of 0.5 m^2 .

Uniform flow was assumed at both the upstream and the downstream model

boundary. To robustly distinguish between wet and dry cells, we used a threshold of $h_{\text{dry}} = 0.1$ m (Toro 2001), which lies between the D_{50} and D_{90} of the bed material.

The numerical model was calibrated with longitudinal profiles of the water surface elevation in the laboratory experiment. The profiles were measured with an ultrasonic distance sensor mounted on a movable crossbeam above the flume. Each profile consists of 53 measurement points, 15 m apart (upscaled value). To provide the most accurate fit between measured water surface elevation and recorded topography, the profiles were recorded during the last 8.5 min of each experiment (upscaled value: 47 min). We used the root mean square error (RMSE) to quantify the deviations between measured and calculated values. The only calibration parameter was the roughness, represented by the Strickler coefficient k_{St} ($\text{m}^{1/3}/\text{s}$) and assumed to be equal for bed and banks. The calibration was carried out with the four BQ topographies (Table 2) and resulted in minimal RMSE values of 0.06-0.088 m for Strickler coefficients of 30-33 $\text{m}^{1/3}/\text{s}$. All subsequent simulations were carried out with $k_{St} = 31$ $\text{m}^{1/3}/\text{s}$, which corresponds well to the Strickler coefficient range determined from D_{90} with $k_{St} = (21.1...26)/D_{90}^{(1/6)} = 28...34$ $\text{m}^{1/3}/\text{s}$ (Strickler 1923; Meyer-Peter and Müller 1948).

Steady state simulations were performed for every topography with discharges ranging from 40 to 320 m^3/s in increments of 10 m^3/s , and the five characteristic discharges listed in Table 1 (34 discharges in total). Note that for some simulations, the simulated discharge is higher than the formative discharge of the topography, e.g. when a HQ_{30} discharge is run on any of the BQ topographies. This might lead to some unrealistic flow conditions, as the topography would constantly adjust to the imposed discharge under natural conditions. Nevertheless, the deviations are assumed to be small compared to the general trends in hydraulic variability that are observed when subjecting different topographies to a wide range of discharges (e.g. Reid et al. 2020).

2.4. Data analysis

From the numerical simulations, we extracted the wetted area A_w for each scenario (8 topographies with 29 discharges). Based on each DEM, A_w was divided into the wetted area belonging to the main channel $A_{w,\text{ch}}$ and the floodplain $A_{w,\text{fp}}$ (see Table 2

for channel and floodplain areas). All spatial analyses are restricted to the widening area $A_{\max} = 63,800 \text{ m}^2$ (Figure 2).

From the wetted area, we calculated the shoreline length relative to the channel length (km/km) (Tockner and Stanford 2002). The shoreline length is a measure of complexity of the marginal zones and the lateral connectivity between aquatic, semi-terrestrial, and terrestrial habitats (Ward et al. 1999). It is discharge-dependent and can assume values from 2 km/km for straight channels up to 25 km/km for braided morphologies such as the Tagliamento River in Italy (Tockner and Stanford 2002). We assume that the shoreline length can be used as a proxy for the availability of refugia in marginal zones (e.g. Rempel et al. 1999). To compute the shoreline length, the water depths obtained from the numerical simulations were interpolated onto a $0.3 \times 0.3 \text{ m}$ grid as binary data (1 = wet, 0 = dry). To prevent overestimation of the shoreline length, the identified boundaries with a length < 100 grid elements were removed and the remaining boundaries were smoothed with a filter of window length 21. All boundary lengths were then summed and divided by the streamwise length of the river widening.

A further result of the numerical simulations is the distribution of water depths, flow velocities, and bed shear stress values with high spatial resolution. Water depth and flow velocity are the basic descriptors of a flow field and can be used to assess the hydro-morphological conditions within the model perimeter. Here, we use the dimensionless hydro-morphological index of diversity (HMID, Gostner et al. 2013)

$$HMID = \left(1 + \frac{\sigma_v}{\mu_v}\right)^2 \cdot \left(1 + \frac{\sigma_h}{\mu_h}\right)^2 \quad (1)$$

with the mean flow velocity μ_v (m/s), the standard deviation of the flow velocities σ_v (m/s), the mean water depth μ_h (m), and the standard deviation of the water depths σ_h (m) (Gostner et al. 2013). The higher the HMID the closer the hydro-morphological conditions are to natural conditions, independent of the specific morphology present, i.e. braiding or meandering. Gostner et al. (2013) introduced the classification (i) channelized and morphologically heavily altered for $HMID < 5$, (ii) lightly modified

with limited hydraulic variability for $5 \leq \text{HMID} \leq 9$, and (iii) morphologically pristine for $\text{HMID} > 9$. Note that the HMID was originally defined for mean flow conditions and its value is discharge-dependent.

Bed shear stresses describe the near-bed hydraulic forces acting on any organisms on the riverbed and are therefore relevant for refugia identification. They can also be interpreted as a proxy for potential bed-load transport. By comparing simulated bed shear stresses τ_b to a critical bed shear stress for the entrainment of a characteristic grain size, e.g. $\tau_{c,D50}$, potential bed-load transport zones may be identified. However, if the bed shear stresses are determined from a hydraulic simulation with homogeneous roughness, variations in surface roughness due to grain sorting are not considered. This simplification overlooks relevant morphodynamic processes such as riverbed patchiness, structuring, and armoring. Furthermore, this method does not consider the near-bed turbulent fluctuations of the velocity and pressure field that may locally modify grain mobility. Therefore, we used the aggradation and erosion zones identified from the DoDs obtained in the laboratory experiments to more reliably identify zones of bed elevation change and thus active bed-load transport. The wetted area A_w was divided into the zones: (i) elevation change, (ii) no elevation change on floodplain, and (iii) no elevation change in channel.

3. Results

3.1. Floodplain inundation

Simulated water depths for $HQ_{1.5}$ and all eight topographies obtained in laboratory experiments (listed in Table 2) are shown in Figures 3 and 4. Under sediment equilibrium conditions (Figures 3a,b and 4a,b), topography and hydraulic conditions show a more complex pattern in comparison to sediment deficit conditions (Figures 3c,d and 4c,d). The floodplain is inundated and thus the wetted area A_w is significantly larger for sediment equilibrium cases in contrast to both the SSL 60% and 20% scenarios. Note that this is the case even though the proportion of main channel to floodplain area of BQ_100%_2 and FQ_100%_2 is closer to the sediment deficit scenarios than to BQ_100%_1 and FQ_100%_1 (Table 2).

Figure 3. Simulated water depths for topographies (a) BQ_100%_1, (b) BQ_100%_2, (c) BQ_60%, and (d) BQ_20% and discharge $HQ_{1.5}$. The proportion of (i) wetted area A_w , (ii) wetted main channel area $A_{w,ch}$, and (iii) wetted floodplain area $A_{w,fp}$ to total widening area $A_{\max}$ is determined from the hydrodynamic simulations. The main channel area is defined as the area below the floodplain, and the inlet and outlet channel (white overlay) do not contribute to the reported areas.

Figure 4. Simulated water depths for topographies (a) FQ_100%_1, (b) FQ_100%_2, (c) FQ_60%, and (d) FQ_20% and discharge $HQ_{1.5}$. The proportion of (i) wetted area A_w , (ii) wetted main channel area $A_{w,ch}$, and (iii) wetted floodplain area $A_{w,fp}$ to total widening area A_{max} is determined from the hydrodynamic simulations. The main channel area is defined as the area below the floodplain, and the inlet and outlet channel (white overlay) do not contribute to the reported areas.

To visualize the floodplain inundation process, Figure 5 displays the ratio of wetted area to total area A_w/A_{max} for discharges between 40-320 m^3/s . The sediment equilibrium and deficit scenarios show distinctly different inundation dynamics visible as sharp increase of A_w/A_{max} over a limited discharge range. For topographies developed under sediment equilibrium conditions (SSL 100%), the floodplain is fully inundated

at discharges around 150 m³/s, for SSL 60% and 20% at discharges above 300 m³/s. No significant differences are apparent between BQ (Figure 5a) and FQ topographies (Figure 5b), although the inundation process happens slightly more suddenly for BQ than for FQ topographies, especially for SSL 100%.

Because the water depths on the inundated floodplain are shallow (Figures 3 and 4), A_w is potentially sensitive to the threshold $h_{\text{dry}} = 0.1$ m. However, a sensitivity analysis with $h_{\text{dry}} = 0.05$ m and 0.15 m showed no significant differences in the inundation process compared to $h_{\text{dry}} = 0.1$ m (Supplementary Figure S1).

Figure 5. The ratio of wetted to total area A_w/A_{max} for (a) four BQ and (b) four FQ topographies and a discharge range of 40-320 m³/s (29 discharges in total).

The inundation process is also reflected in the discharge-dependent shoreline length (Figure 6). The shoreline length is elevated and variable for the discharge range of partial floodplain inundation, and close to its minimum level of 2 km/km for lower and higher discharges, i.e. a completely dry or wet floodplain. For example, shoreline length is 3-8 km/km for discharges below 150 m³/s for BQ_100%_1 and BQ_100%_2, whereas the minimum level of 2 km/km is maintained for discharges $< HQ_{30}$ for BQ_60% and BQ_20% (Figure 6a).

Figure 6. Shoreline length for (a) BQ and (b) FQ topographies and a discharge range of 40-320 m³/s. The minimum value of the shoreline length is 2 km/km.

3.2. Hydraulic variability

Figures 3 and 4 show simulated water depths for a discharge of $HQ_{1.5}$ and the separation between main channel and floodplain for the BQ and FQ topographies, respectively. Although the hydraulic forces, i.e. flow velocity and bed shear stress, are presumably more relevant for the refuge function, water depths are suitable to visualize the underlying topography and thus displayed here. For the corresponding visualization of flow velocities and bed shear stresses, see Supplementary Figures S2-S5.

Figures 3 and 4 indicate that the hydraulic variability is higher for sediment equilibrium conditions (SSL 100%) as opposed to sediment deficit conditions (SSL 60% and 20%). Figure 7 quantifies the hydro-morphological diversity as HMID for a discharge range of 40-320 m³/s. Maximum HMID values of 8-17 are attained for discharges $< HQ_{30}$ under sediment equilibrium conditions and for discharge values $> HQ_{30}$ under sediment deficit conditions. Minimum HMID values lie around 5 for sediment equilibrium conditions and $Q > HQ_{30}$, and between 2 and 5 for sediment deficit conditions and $Q < HQ_{30}$. The differences between the sediment supply levels are more pronounced for the BQ topographies than for the FQ topographies.

Figure 7. Hydro-morphological index of diversity (HMID) for the (a) BQ and (b) FQ topographies and a discharge range of 40–320 m³/s with the classification after Gostner et al. (2013). Note that the classification was originally developed for mean flow conditions.

For the five characteristic discharges listed in Table 1, the bed shear stress distributions are shown in Figures 8 and 9. Under sediment equilibrium conditions, the median bed shear stress slowly increases over the entire discharge range (Figures 8a,b and 9a,b). In contrast, the median increases more sharply for the SSL 60% and 20% scenarios, before it drops for HQ_{100} (Figures 8c,d and 9c,d).

The bed shear stress distribution and $A_w/A_{\max}$ indicate that the floodplain is inundated at lower discharges under sediment equilibrium conditions compared to a sediment deficit (cf. Figure 5). The inundation leads to a high proportion of low bed shear stress areas, and a larger wetted area as indicated by the size of the gray area (e.g. Figure 8a at $HQ_{1.5}$). Under sediment deficit conditions, only few low bed shear stress areas are available during flood events smaller than HQ_{100} , and the total wetted area remains limited (Figures 8c,d and 9c,d).

Figure 8. Bed shear stress distribution of the four BQ topographies (a-d) for five characteristic discharges (Q_m to HQ_{100}). The gray area shows the distribution of bed shear stresses, the black bar the interquartile range, and the white circle the median value. The critical bed shear stress values for the entrainment of D_{50} ($\tau_{c, D50}$) and D_{90} ($\tau_{c, D90}$) grains calculated with a dimensionless critical bed shear stress of $\theta_c = 0.047$ are shown for comparison. The ratio of wetted area to total widening area $A_w/A_{\max}$ and the respective proportion of main channel $A_{w, \text{ch}}/A_{\max}$ and floodplain $A_{w, \text{fp}}/A_{\max}$ are reported.

Figure 9. Bed shear stress distribution of the four FQ topographies (a-d) for five characteristic discharges (Q_m to HQ_{100}). The gray area shows the distribution of bed shear stresses, the black bar the interquartile range, and the white circle the median value. The critical bed shear stress values for the entrainment of D_{50} ($\tau_{c,D50}$) and D_{90} ($\tau_{c,D90}$) grains calculated with a dimensionless critical bed shear stress of $\theta_c = 0.047$ are shown for comparison. The ratio of wetted area to total widening area $A_w/A_{\max}$ and the respective proportion of main channel $A_{w,ch}/A_{\max}$ and floodplain $A_{w,fp}/A_{\max}$ are reported.

3.3. Riverbed elevation change

Figure 10 shows the relative frequencies of the bed shear stress distribution of the BQ topographies and their formative discharge $HQ_{1.5}$. The inundated floodplain under

sediment equilibrium conditions is clearly visible as a peak with $\tau_b < 50 \text{ N/m}^2$ in Figures 10a and 10b. Most of the floodplain experiences no elevation change, except for a small part of very low bed shear stress areas (light gray and $\tau_b < 10 \text{ N/m}^2$). These are most likely floodplain zones where a shallow layer of fines was deposited. In contrast, the limited occurrence of low bed shear stresses ($\tau_b < 50 \text{ N/m}^2$) for sediment deficit conditions (Figure 10c,d) indicates that the floodplain was not inundated at this discharge.

For $\tau_b > 50 \text{ N/m}^2$, the bed shear stress distributions are rather similar between sediment equilibrium (Figures 10a,b) and sediment deficit conditions (Figures 10c,d), although the distribution is slightly flatter in Figures 10a and b. However, the relative frequency of the wetted area experiencing bed elevation change is significantly higher for sediment equilibrium and there is, therefore, an extensive zone of active bed-load transport in the main channel. In contrast, most of the wetted area experienced no bed elevation change under sediment deficit conditions (Figures 10c,d). There might still be bed-load transport across such a zone without bed elevation change (no net change). But due to the low sediment supply and the armored bed surface, the active bed-load transport zone is likely very limited (cf. Dietrich et al. 1989).

Figure 10. Relative frequency p of bed shear stress distributions and the partition into areas with and without elevation change for (a) BQ_100%_1, (b) BQ_100%_2, (c) BQ_60%, and (d) BQ_20% and the bedforming discharge $HQ_{1.5}$. The area with elevation change (light gray) was determined from a 3-hour interval in the laboratory experiment.

4. Discussion

This study sought to relate refugia availability for mobile aquatic organisms in a dynamic river widening to the underlying process of sediment transport, which may have contrasting impacts regarding refugia availability. Sediment transport is itself a disturbance causing scour and fill (e.g. Townsend et al. 1997; Effenberger et al. 2006; Czuba et al. 2018), however it is also a key driver of morphodynamics and habitat variability (e.g. Reid et al. 2020) which is considered a prerequisite for the persistence of low-impact habitats during flood events (Sedell et al. 1990).

4.1. Interaction of sediment supply and hydro-morphological complexity

The maximum shoreline lengths observed here are in the range of 5-8 km/km for sediment equilibrium conditions and $Q < HQ_{30}$, and 3-5 km/km for sediment deficit conditions and $Q > HQ_{30}$. These values correspond to the lower end of the range 5-15 km/km reported for a constrained but not channelized reach of the Danube River in Austria (Tockner and Stanford 2002). In contrast, a natural river, such as the Tagliamento River in Italy, reaches shoreline lengths of 25 km/km (Tockner and Stanford 2002). While the shoreline length of the Danube River displays high dependence on discharge similar to this study, the Tagliamento River maintains consistently high shoreline lengths with the exception of very low and very high discharges (van der Nat et al. 2002). Assuming that a higher shoreline length results in more potential refuge areas in the marginal zones (e.g. Rempel et al. 1999), dynamic river widenings developed under sediment equilibrium conditions show a higher refuge potential compared to sediment deficit scenarios. However, due to the limited length and width of river widenings, the attainable shoreline lengths remain limited.

While this general outcome is reasonable, the absolute shoreline length values must be treated with caution. Although the numerical model has a high spatial resolution, it cannot represent the full complexity of marginal zones with crevices, indentations, etc., and absolute values of shoreline length generally depend on the observation scale. Furthermore, the floodplain in the laboratory experiment is unrealistically flat. This causes the shoreline length to drop to its minimum value of 2 km/km once the flood-

plain is completely inundated. Although the shoreline length will likely decrease for elevated discharges, it will not drop to the minimum value in case the floodplain is more structured, for example with patches of vegetation (Tockner and Stanford 2002).

The HMID values follow a similar trend as the shoreline length, with higher values and thus higher hydro-morphological diversity for sediment equilibrium conditions and $Q < HQ_{30}$ and a reversal for $Q > HQ_{30}$, therefore capturing the different floodplain inundation processes. However, at HQ_{30} , equal HMID values of 4-6 result for the BQ topographies under both sediment equilibrium and deficit conditions (Figure 7a), implying that their respective flow fields are equally diverse. However, Figure 3 shows that for sediment equilibrium and $HQ_{1.5}$, the water depth distribution consists of smaller patches and exhibits a more complex configuration than under sediment deficit conditions, and similar differences can be expected for HQ_{30} . The reason for equal HMID despite apparent differences in the respective flow fields is that the HMID is a statistical bulk value to quantify hydro-morphological diversity (Eq. 1). It does not, however, express the spatial arrangement of hydraulic patches (Gostner et al. 2013) which makes it suitable to measure the variability, but not the complexity of a hydraulic habitat mosaic.

For discharges up to HQ_{30} , both the shoreline length and HMID values are higher under sediment equilibrium compared to sediment deficit conditions. For discharges $> HQ_{30}$ the relationship is inverse. Assuming that potentially more refugia are available in morphologically and hydraulically complex reaches (e.g. Pearsons et al. 1992; Gjerløv et al. 2003), both metrics imply higher refugia availability under sediment equilibrium conditions and discharges up to HQ_{30} . This dynamic is closely linked to the lateral channel-floodplain connectivity. Under sediment equilibrium conditions, stronger bed aggradation in the widened main channel led to reduced elevation differences between the main channel and the floodplain and, therefore, the floodplain is inundated at lower discharges (e.g. Hohensinner et al. 2003). Conversely, the incision of the main channel relative to the floodplain under sediment deficit conditions prevents the floodplain inundation except for high discharges (e.g. Surian and Rinaldi 2003).

However, increased lateral connectivity is only one of many factors that likely lead to a higher refugia availability in sediment equilibrium conditions. Low bed shear stress

zones that persist during high discharges also provide potentially valuable refugia (e.g. Lancaster and Hildrew 1993; Winterbottom et al. 1997). Lancaster and Hildrew (1993) distinguish three types of bed shear stress shifts in response to increasing discharge: type I from unimodal skewed to bimodal, type II from unimodal skewed to unimodal bell-shaped, and type III from bimodal to unimodal bell-shaped (see Figure 7 in Lancaster and Hildrew 1993). They concluded that a type I shift in bed shear stresses holds the highest potential for refugia, because low bed shear stress zones persist throughout the discharge range associated with the main channel being most affected by a disturbance event (Lepori and Hjerdt 2006). In this study, the conditions under sediment equilibrium correspond to a type I shift which is mainly associated with the inundation of the floodplain activating an extensive area of shallow water depths, slow flow velocities, and low bed shear stress zones. Conversely, the sediment deficit conditions resemble a type II shift with a narrow distribution of ever increasing bed shear stresses except for HQ_{100} , when the floodplain is inundated and the distribution also resembles type I. The stronger increase of bed shear stresses under sediment deficit conditions can be explained by topographic characteristics. The homogeneous main channel that was preserved under sediment deficient conditions effectively concentrates the flow and leads to higher bed shear stresses with a more narrow distribution in comparison to sediment equilibrium conditions (e.g. Negishi et al. 2002; Czuba et al. 2018). Some benthic organisms may be able to move to the subsurface zone, but the refuge function of the hyporheic zone is highly variable (e.g. Dole-Olivier et al. 1997; Stubbington 2012). It can be concluded that the hydro-morphological conditions developed under sediment deficit conditions are less likely to provide suitable flood refugia.

4.2. The floodplain as a refuge

Under sediment equilibrium conditions, the floodplain is inundated at comparably low discharges ($Q_{m, \text{June}} - HQ_{1.5}$), suggesting that its shallow and slow flowing zones are readily accessible as refuge. But does the floodplain really assume refuge function? Schwartz and Herricks (2005) reported fish moving into the floodplain during high discharges and the same has been observed for macroinvertebrates (e.g. Negishi et al. 2002). However, the organisms must also be able to successfully return to the

main channel once the flood recedes. A field study on macroinvertebrate refugia in a New Zealand stream investigated the refuge function of the floodplain by differentiating between the floodplain acting as a temporary shelter versus a permanent refuge (Matthaei and Townsend 2000). According to their study, the conditions in the main channel during the falling limb of the hydrograph are essential, i.e. when the flow retreats from the floodplain and organisms return back to the main channel. Stranding on the drying floodplain does not seem to be a significant risk to macroinvertebrates (Matthaei and Townsend 2000) or fish (Mannes et al. 2008). However, if hydraulic conditions in the main channel remain unfavorable when the organisms return, i.e. with high bed shear stresses and active bed-load transport, the temporary shelter on the floodplain does not offer any advantage regarding the persistence of organisms.

From the floodplain inundation dynamics, it was determined that the flow retreats from the floodplain below discharges of $Q_{m,June} - HQ_{1.5}$ for sediment equilibrium conditions which corresponds to median bed shear stresses of 23-53 N/m² being present in the main channel. In contrast, the water almost entirely retreats from the floodplain around HQ_{30} under sediment deficit conditions, with median bed shear stresses of 90-140 N/m² in the channel, thus presenting a higher risk for displacement. Note that the bed shear stresses determined from a two-dimensional hydrodynamic model cannot accurately represent the detailed near-bed flow field (Rice et al. 2007), but rather allow comparison of the relative magnitudes of the hydraulic forces.

Another process to consider when evaluating floodplain refugia availability is bed-load transport activity in the main channel upon floodplain drying (Matthaei and Townsend 2000). Under sediment equilibrium conditions, the active bed-load transport zone determined from the laboratory experiments at $HQ_{1.5}$ is more extensive than under sediment deficit conditions (e.g. Dietrich et al. 1989; Nelson et al. 2009). Consequently, organisms returning from the floodplain to the main channel under sediment equilibrium conditions are possibly subjected to destructive bed-load transport (e.g. Effenberger et al. 2006; Matthaei and Townsend 2000). However, even at $HQ_{1.5}$, there is a significant portion of low bed shear stress areas without bed elevation change in the channel (Figures 10a,b). Considering that the floodplain only completely dries up for discharges below $Q_{m,June}$ and bed-load transport activity will have decreased fur-

ther, we can therefore assume that under sediment equilibrium conditions, organisms retreating from the floodplain will find plenty of potentially suitable habitat patches in the main channel.

4.3. Refugia availability versus use

The results of this study illustrate that a dynamic river widening developed under sediment equilibrium conditions has a higher refuge potential compared to sediment deficit conditions. However, the availability of refugia does not necessarily imply positive effects on the biota (cf. Palmer et al. 2010). The use of refugia by aquatic organisms is dependent on a variety of factors that this study does not investigate. Organism characteristics such as their mobility (e.g. Williams and Hynes 1976; Rice et al. 2007) and their body size versus the spatial dimensions of the refuge (e.g. Lancaster 2000; Schwartz and Herricks 2005; Mathers et al. 2019) play an important role. Organisms seek out different refugia associated with varying suitability under differing discharge regimes. Sculpins are for example especially susceptible to scouring events as they hide in the interstitial pore spaces (Swanson et al. 1998). This refuge may therefore be suitable below a critical discharge, above which it is destroyed instantaneously. The higher the flood severity, the more valuable large-scale refugia, such as tributaries and floodplains, become (Sedell et al. 1990).

The spatial configuration and hierarchy of refugia was not considered in this study, but may also have important implications for their use (e.g. Sedell et al. 1990; Lancaster and Belyea 1997). For example, the proximity and accessibility of a refuge to the residential habitat of an organism is crucial regarding its energy expenditure (e.g. Weber et al. 2013; Sueyoshi et al. 2014). In the setting studied here, this means that possibly only the margin areas close to the main channel may be used as a refuge and not the entire floodplain. This effect was observed by Matthaei and Townsend (2000) where floodplain areas close to the main channel were more likely to serve as a refuge compared to a backwater channel further away.

4.4. Implications for management

The results presented here show no substantial deviations between topographies developed during steady bedforming flow (BQ) and during a flood event (FQ). This indicates that refugia availability in dynamic river widenings may be fairly independent of specific hydrological events and far more influenced by the sediment supply conditions. All results exhibit significant differences between sediment equilibrium and sediment deficit conditions, whereas the tested degrees of sediment deficit (60% or 20%) do not seem to be profoundly relevant to the outcome. For the investigated cases, sediment deficit leads to channel incision which disconnects the main channel from the floodplain. The sediment deficit also reduces the tendency for local erosion and deposition of material throughout the reach thus weakening lateral erosion. The result is a distinctly different morphology and degree of lateral connectivity which both negatively impact refugia availability. Our results therefore suggest that an equilibrium sediment supply clearly favors refugia availability in a dynamic river widening. However, the available lab data do not reveal a transition for the degree of sediment deficit at which negative effects start to prevail; this requires further and systematic lab experiments.

This study provides an example of the conflicts that may arise if restoration measures do not target large-scale ecosystem processes but instead focus on habitat structure on a reach scale (e.g. Beechie et al. 2010; Polvi et al. 2020). In order for dynamic river widening to be an effective restoration measure, pressures on a larger scale such as impaired sediment and flow regimes (e.g. Vanzo et al. 2016), insufficient water quality, or land use changes must be addressed and, if possible, restored concurrently (Kondolf et al. 2006). Mitigation measures for sediment supply include various structural (e.g. sediment bypass tunnels), operational (e.g. reservoir drawdown), and replenishment measures (Kondolf et al. 2014). In Switzerland, a concerted effort to mitigate the effects of channel degradation, fish migration barriers, sediment deficit, and hydropeaking is ongoing (Bammatter et al. 2015). Regarding the interruption of sediment continuity, approximately five hundred hydropower dams and weirs as well as other structures must be altered to improve sediment connectivity. This opens opportunities for downstream channel restoration measures, such as dynamic river widenings, to fulfill their ecological potential, for example by providing suitable flood refugia.

5. Conclusions

Our results indicate that dynamic river widenings exhibit higher morphodynamic activity, greater hydro-morphological diversity, and a stronger lateral connectivity under equilibrium sediment supply compared to sediment deficit conditions. Low bed shear stresses are preserved during elevated discharge thus potentially providing more suitable refugia to mobile aquatic organisms. The positive effects of bed-load transport prevail, even though it is itself a disturbing force. In contrast, sediment deficit conditions lead to impoverished hydro-morphological conditions and the floodplain is unavailable, owing to incision, as refuge due to channel-floodplain disconnection. These findings imply that the potential ecological benefits of dynamic river widenings in subalpine gravel-bed rivers are strongly dependent on the sediment supply level. If a river widening is planned in a reach with a sediment deficit, additional measures such as the placement of structuring elements (e.g. groins, engineered log jams) may be necessary to improve refugia availability and increase the resistance and resilience of stream communities. To generalize the relation between sediment supply and refugia availability in the context of river restoration, further studies on different river systems and restoration methods are necessary.

The ecological assessment of restoration measures, such as dynamic river widenings, should not be limited to the investigation of flood refugia, but also consider the impacts of other disturbance events, especially droughts (e.g. Lake 2000; Wood et al. 2010). Due to climate warming, dry periods are becoming more frequent in many regions and the investigation of habitat fragmentation gains in importance (e.g. Vander Vorste et al. 2020). If dynamic river widening leads to large open spaces without shade and with shallow water depths, the restoration measure may not provide suitable conditions for biota in low discharge situations. It is therefore crucial to further investigate the impacts of all relevant types of disturbance events on biota in the context of river restoration.

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