Dear Sébastien Castelltort,

We submit a revised version of our manuscript entitled '*Alluvial channel response to environmental perturbations: Fill-terrace formation and sediment-signal disruption*' for consideration for publication in *Earth Surface Dynamics*. First, we would like to apologize for our delay in submitting the revised version. We appreciated the reviews and the efforts made to help improve our manuscript and we heavily modified the manuscript even beyond the suggestions of the reviewers. Therefore, before replying to the reviewer's comments in detail below, we would like to summarize the main modifications to the manuscript.

Reviewer #1 suggested to better emphasize the main focus of the manuscript, as the differentiation between novel observations and confirmation of earlier ideas was not clear. Initially, the manuscript had a strong focus on fluvial terraces only. However, we think that the major strength of our experimental setup is the opportunity to track the evolution of two different records of landscape evolution simultaneously – namely (1) fluvial-fill terraces in the transfer zone and (2) sediment discharge ($Q_{s,out}$) to the deposition zone. We have thus reoriented the scope of the manuscript in this direction. The main modifications involve:

- A strong reorientation of the introduction. We included information on the general behavior of alluvial channels as well as background knowledge on variability in $Q_{s,out}$. To balance this, we have removed the extensive background section on fluvial-fill terraces (former chapter 2) and included the most important information in the introduction.
- For clarification and a better visualization we included two new figures: figure 1 and 9. Figure 1 is a conceptual summary of sediment-routing systems (modified after Castelltort and van den Driessche, 2003), the important parameters that shape the transfer zone as well as the two landscape-evolution records that we investigate in our experiments. Figure 9 is a conceptual summary of our observations.
- With the modified focus of the paper, we also strongly modified the structure and (partly) the content of the discussion section. Within the new structure, we discuss (1) fill-terrace records; (2) $Q_{s,out}$ records; (3) what can be learned about the coupling of the two; (4) the limitations of our experimental approach, especially when compared to natural settings (following the main concerns of reviewer #2); and (5) implications of our observations for future field studies.

In addition to the changes listed above or stated below within the detailed responses to the reviewers, we performed slight modifications to the figures, captions or to the wording of the main text. These modifications were either only stylistic for increased precision and clarity, or are discussed in the detailed responses below.

Yours sincerely,

Steffi Tofelde and co-authors

Response to review by Luca Malatesta:

First, we would like to thank Luca Malatesta for his detailed and constructive review of our submission. We believe that addressing his thoughtful comments has increased the quality of the manuscript. Luca Malatesta's two main comments were related to (1) the usage of terminology and (2) the structure of the manuscript. In addition, he provided several line-by-line comments related to science and bibliography. While the original review comments are shown in *italics*, our responses are given in regular, blue font. Our line numbers refer to the newly submitted version of the manuscript.

1. Fill vs. Cut-in-fill terraces:

The authors introduce the object "fill-terrace" on page 2 and thereafter it is inferred that all terraces recorded in their flume experiments are such. I would object to this use of the term. A fill terrace, as described on page 2, is a morphologic datum recording the culmination of sediment aggradation immediately preceding a phase of incision and thus abandonment (Howard, 1959; Bull, 1991; Pazzaglia, 2013). In several experimental runs, it seems that the entire active floodplain is being eroded before it narrows its width and starts entrenching, thus abandoning terraces. In that situation, these terraces are not "fill-terraces" but cut-in-fill as they record a moment during the incisional phase and not the culmination of alluvial aggradation. The title of the article needs to be accordingly modified. Then, the difficulty resides in reliably identifying if a given "top" terrace (top as in being the highest from the last incision episode) is indeed a fill terrace. To me it is very interesting that the authors identify cases where barely any fill terraces are abandoned. And that instead two large cut-in-fill terraces replace the fill terraces one would commonly expect. It appears to capture the moment when vertical incision is promoted over lateral erosion leading to fast autogenic entrenchment of the channel (Malatesta et al., 2017; Bufe et al., 2018) but the two experiments with a drop in Qs suggest that this inflexion point does not always occur at a similar moment. Finally on that point, the rationale behind picking the terraces TA and TB should be fleshed out because at least in the case of the DQsin run, they capture cut-in-fill terraces. More about that with the comment on p. 12 l. 13.

Indeed, the terrace terminology in the literature is rather inconsistent. Often, terraces are subdivided into two main categories: strath and fill (e.g. Howard 1959, Pazzaglia 2013). Fill terraces have been further subdivided into the 'highest' terrace that preserves the original deposited surface and 'lower' terraces with surfaces below the original deposited surface that have been eroded laterally into the fill. While the first type is referred to as 'filltop' (Howard 1959) or just 'fill terrace' (e.g., Bull 1990, Merritts et al., 1994), the second type has been described as 'fill-strath' (Howard 1959), 'cut-terrace' (e.g. Merritts et al., 1994), 'fill-cut terrace' (Mizutani 1998, Bull 1990, Pazzaglia 2013, Malatesta et al. 2017) or 'cut-fill terrace' (e.g., Norton et al., 2015). As such, for simplification, we only referred to fill terraces in general with the aim to include both subtypes. Especially since a distinction between the two subtypes in the field is often not possible without detailed stratigraphic or geochronological analysis.

However, we agree that a distinction of the two fill terraces sub-categories would be helpful to clarify the terrace description, and also the discussion on lag-times between perturbation and terrace cutting. A distinction of the two subtypes within the experiments can easily be made, as we covered the surface with a thin layer of red sand prior to each perturbation. The preservation of the red sand is a clear indication for no further overwash after the perturbation and, as such, identifies the first subtype of fill terraces (fill-top). Any later formed terraces will consequently be fill-cut terraces. Therefore, we added a formal definition of the two subtypes to the introduction (l. 72-73) and the methodological distinction to the methods section (l. 162-164). For a better visualization, we modified Fig. 3 (former Fig. 2) and included photos of each terrace section with the according labels of terrace type and lag-times. We chose, however, to keep the original title as the general term ‘fill terraces’ covers all types. For later analysis (e.g. terrace surface slope) we always chose the most extensive terrace surfaces on each side of the river. With this approach we aim to mimic common field approaches. We clarified this in the results section (l. 235-236).

We also agree that the definition of what constitutes a paired or unpaired terrace is not clear (see reviewer comments p.9 l. 21-22 and p. 12 l. 5). Often, paired or unpaired terraces are distinguished based on height similarities or differences. However, as far as we are aware, there is no common rule where to draw the threshold. Instead, we followed the reviewer’s suggestion and instead referred to the ages/ lag-times of the terrace surfaces and describe successive abandonment instead of referring to ‘unpaired’ terraces. All former statements of paired/unpaired terraces have been removed.

Also, we agree that the cut-terraces capture the moment when vertical incision outcompetes lateral erosion. However, we disagree that this process should necessarily be referred to as ‘autogenic entrenchment’. In the literature, the term ‘autogenic’ has been used inconsistently and no one definition of the term seems to exist. In the new version of the manuscript, we have decided to reduce this topic to the following statement: “In some cases, internal dynamics of the system, sometimes referred to as “autogenic processes”, lead to terrace formation that cannot be directly linked to external forcing (e.g., Erkens et al., 2009; Limaye and Lamb, 2016; Malatesta et al., 2017; Patton and Schumm, 1981; Womack and Schumm, 1977) .”

2. Structure of the manuscript

I think that a weakness of the current manuscript structure is that it is difficult to understand what the novel advances are and what the narrative of the work is. That is especially true for readers who are familiar with the existing, extensive, body work on alluvial geometry dating starting with Gilbert and Murphy (1914). The results are presented as if they almost provided a first-time observation of such alluvial dynamics. However, most of the observations from the flume experiments have already been observed, predicted, or discussed in previous bodies of work. What is novel is the documentation of the transient response itself. The manuscript could be somewhat modified to make this clearer and better highlight the contribution of the authors to this larger body of work. In that spirit, I would suggest to move elements of the discussion to the review section “2 Formation of fluvial fill terraces” so as to clearly establish what is acquired knowledge and to underline the gap that the authors want to fill here. In particular, section 2.1 could be augmented with large parts of sections “5.1 Channel response to perturbations and conditions of terrace formation” and “5.3 Differences in terrace surface slope”. By explicitly introducing the theoretical framework used to describe the relationships between alluvial slope and fluxes of sediment and water (Qs and Qw), the authors would build a better launchpad for their study in my opinion. The Meyer-Peter Müller (MPM) equation revised byWong and Parker (2006) or more recent derivations of slope as a function of Qs and Qw (e.g. by Malatesta and Lamb, 2017 GSAB, or Wickert and Schildgen, 2019) can help establish clearly what is known so far, and what is not. The latter being a good understanding of the transient behaviour from one equilibrium configuration to the next. I believe that this modification to the structure of the manuscript would help the reader better navigate the coexistence of the review and experimental aspects of the paper.

We apologize for giving the impression that all our observations were novel. This was not our intention. We followed the suggestion of the reviewer to better elaborate the main focus of the paper. We think that the strength of the paper is the tracking of the simultaneous evolution of two different records of landscape evolution – fluvial fill terraces in the transfer zone and sediment deposition rate (~$O_{s,out}$) in the deposition zone. The majority of the changes of the manuscript relate to better structure the manuscript around this point. Following adjustments were made:

- We significantly rewrote the introduction to better lead to the point of comparing the two records.

- Within the new introduction, we follow the reviewer's suggestion and extend the background on channel geometry and sediment transport. To do so, we have moved parts of the former discussion (section 5.3) to the introduction, and we have also included an explanation of the relationship between the geometrical adjustment, bedload transport and transport capacity after Meyer-Peter Müller (1948), as was both suggested by reviewer #1. In line with these modifications, we have moved the most important information on terrace formation (former section 2) to the introduction and removed section 2 instead.
- We also added a new overview figure 1 that summarized the source-to-sink framework as well as the important parameters for our study.
- Within the results, we slightly rearranged the order by moving the description of terrace abandonment to the beginning and by distinguishing between fill-top and fill-cut terraces (l. 182-201).
- With the more defined focus of the paper, we basically discuss for each of the two records (fill-terraces and $Q_{s,out}$/sedimentation rate) (1) how the information stored in them can be modified and (2) how different forcings can lead to a similar stratigraphy (ambiguity) (new discussion sections 4.1 and 4.2). Therefore, we decided to remove the discussion part on channel width evolution, as this point is interesting, but does not directly contribute to our main focus anymore. Instead, we are currently working on a manuscript that addresses this point separately. We enhanced the discussion on why we observed different lag-times between perturbation and terrace surface abandonment for the different forcing mechanism (l. 290-325).
- We also expanded the discussion on $Q_{s,out}$ by including a paragraph about the times of Qs signal modification (new section 4.2.1), as well as an entire section on the combination of the two records (new section 4.3), and included a new figure 9 to summarize these points.
- Last, we adjusted the conclusion by focusing more on the coupling of the two records.
- Also, we carefully rephrased the sentences that implied our observations are novel despite being a confirmation of earlier observations or ideas (e.g. p. 13 l. 11-12, p. 16 l. 18-19), if they have not been removed from the new version.

3. Science and bibliography comments

p. 1 l. 9-10: This is a pretty strong statement. I would argue that published work provide a pretty good understanding of the impacts of such forcing on terrace formation and sediment dynamics. What is lacking and provided by the authors here is rigorous observations of the transient response.

We agree that the original statement was rather vague and as such could be understood in several ways. Therefore we adjusted the sentence to: "However, we currently lack a systematic and quantitative understanding of the transient evolution of fluvial systems and their associated sediment storage and release in response to changes in base level, water input, and sediment input."

p. 2 l. 27-30: Malatesta, Prancevic and Avouac (2017, JGR) explicitly target lateral feedbacks with a numerical model.

We did include the reference.

p. 2 l. 31: Limaye and Lamb (2016, JGR) could also be mentioned here as an example of an excellent bedrock model.

We agree that the work of Limaye and Lamb (2016) is an important paper. The sentence the reviewer refers to has been removed from the modified introduction. Please note though, that we cite this paper in the section on autogenic terrace formation, as it particularly focuses on the formation of autogenic terraces.

p. 3 l. 8-10: I strongly encourage the authors to have a look at the 2003 Geology paper by Bonnet and Crave. Therein the authors investigate the impact of climatic (Qw) vs. tectonic forcing (base level) on an experimental landscape. While not targeting terraces in particular, it is one of the most insightful papers I've read on the subject. I strongly encourage the authors to read through it and incorporate some thoughts in their work.

We thank the reviewer for this suggestions. We have implemented the paper in the introduction (l. 79, 82 and 84) as well as in the discussion (section 4.2.2, l. 367, 376 and 389).

p. 3 l. 20: "upstream" [and along stream] (to take into account extra Qs from local incision)

This sentence has been removed.

p. 4 l. 3: If incision supplies sediment to Qsin along stream, then Qsin is not the input sediment flux. It might be useful to separate Qsin, Qsc (sediment transport capacity at any point along stream), and Qsout.

We agree that this point was confusing. Thus, we have now separated sediment flux into: $Q_{s,in}$ (=entrance to the transfer zone), Q_s (=sediment flux at any point within the transfer zone) and $Q_{s,out}$ (=sediment discharge at the outlet of the transfer zone). We have added the three parameters to our new overview figure 1. Also, we have adjusted the text accordingly.

p. 5 l. 10-13: I understand and appreciate the distinction here, and it is quite useful to separate the two. But is it a new refined definition? It seemed to me that fill-cut terraces are commonly considered both "complex response" and "autogenic" at the same time (Schumm's work and Pazzaglia's review paper). If you indeed propose this new,useful, distinction here, I would encourage you to take ownership of it.

The section on autogenic terraces has been reduced to the following statement (l. 68-70) "In some cases, internal dynamics of the system, sometimes referred to as "autogenic processes", lead to terrace formation that cannot be directly linked to external forcing (e.g., Erkens et al., 2009; Limaye and Lamb, 2016; Malatesta et al., 2017; Patton and Schumm, 1981; Womack and Schumm, 1977)."

p. 5 l. 28: There is a new paper by Johnson and Finnegan that is in revision at Geology on "Tributary Channel Transience Triggered by Bedrock River Meander Cutoffs." I don't know when it will come out. But regardless, it might interest you for the future.

We thank the reviewer for this suggestion.

p. 6 l. 5: As the reference codes of the experiments are going to be used thereafter, I would suggest to make a reference to Table 1 here.

We included it.

p. 6 l. 16: what is the vertical resolution?

We included the information on the vertical resolution (1 mm).

p. 6 l. 29: It could be helpful to mention that water is tainted blue in the photos.

We added a sentence within the methods section explaining that the water was dyed blue (l. 141-142).

p. 6 l. 32: why can it be considered unaffected?

We agree that this statement was too strong as we cannot 'prove' the upstream part to be unaffected. Instead, we changed the statement to "we considered this sector of the channel to be least unaffected by the fixed location of the outlet." The second and more important reason to analyze the upstream part is because the terraces were preferentially formed in this part.

p. 7 l. 30: I would argue that change in channel width is not required to form fill terraces. What needs to be reduced is the breadth of the active floodplain (in which the channel, of potentially fixed width, migrates left and right).

We agree with the reviewer. We have changed the sentence to 'Fill-terrace formation requires changes in the channel-bed elevation and width of the active floodplain'.

p. 9 l. 4: The nature of terraces TA and TB could be mentioned here to simplify the reading of the paragraph.
We moved the description of the terraces to the beginning of the results section (in accordance with the new details on all terraces in the updated Fig. 3) and included an explaining sentence regarding T_L (former TA) and T_R (former TB) at its very beginning (l. 174-175).

p.9 l. 21-22: Is there a threshold for what constitutes a pair? Is there a way to define that objectively, or at least in a consistently arbitrary way?
We removed the terms 'paired' and 'unpaired' as no clear definition exists. Instead, we refer to the terraces based on their lag-times. Please also see our comment to 1.

p. 10 l. 6-7: Not sure I understand the rationale behind the ratio of vertical and horizontal erosion. A terrace of width W is preserved for a time T with a river lateral erosion Eh such that T=W/Eh. Preservation is independent from the vertical incision rate. However, deep incision will result in higher walls that are costlier to erode.
What we meant is that vertical incision needs to outcompete lateral erosion to even form terraces. We agree that the term 'preservation' used in the text was misleading. To avoid further confusion, we changed the sentence to "The cutting of fluvial-fill terraces requires vertical incision and a simultaneous reduction of the active floodplain width."

p. 10 l. 15-19: Field studies such as Tofelde et al. (2018), Malatesta et al. (2017, Basin Research), or, and especially, Dzurisin (1975). More on the latter below.
In this paragraph, we discuss the evolution of the longitudinal channel profiles and how our results relate to other physical or numerical studies that applied the same or similar forcings. The difference between physical and numerical model studies compared to field studies is that in physical and numerical modelling studies the input forcing parameters are known, such that the resulting profiles can be directly related to the forcing mechanism. For field studies, however, the longitudinal profiles can be reconstructed from terraces, but the main driver can only be inferred, but not known. Therefore, we prefer not to compare our results to field studies at this point.

p. 11 l. 4-5: a comment only valid if the theoretical framework for alluvial rivers is not beefed up above: I suggest to state that +Qs leads to +S in order to preserve eq. 1 under constant Qw, just as to explain the rationale between Qs and S which is not directly derived from Eq. 1 and 2.
The theoretical framework is now included in in the introduction and this part of the discussion has been removed.

p. 11 l. 8: This dynamic is described and discussed by Malatesta et al. (2017, JGR). It is also worth noting two earlier flume experiments by Schumm et al. [1987, chapter 6] and Meyer et al. [1995] describe the evolution of a channel profile after it reaches a new equilibrium post-incision (see description of that work in section 5.1 in Malatesta et al. 2017, JGR).
With the new scope of the paper, we have decided to remove the detailed discussion on channel width changes as it does not contribute to the adjusted focus of the paper.

p. 12 l. 1: What exactly is the degree of reworking of terrace material? The amount of vertical incision?
We consider the degree of reworking rather as the time the river still reworks the active layer before the terrace surfaces get abandoned and the sediment 'trapped'. For clarification, we changed the sentence to: "The lag time between an external perturbation and the onset of terrace cutting determines how much time the fluvial system has to modify the terrace sediments before their abandonment."

p. 12 l. 5: I am a little hung up on paired/unpaired and the threshold it implies. Wouldn't it be more informative to simply write that the terraces are abandoned successively?
We have removed the terms 'paired' and 'unpaired' entirely. Please see comment to 1 for details.

p. 12 l. 13: Runs DQsin and IQsin_DQsin both lead to entrenchment when sediment flux drops. So, why does the same forcing cause very different terrace creation, or at least be considered as two different systems? To me, it seems that the different terrace record of the two runs could be explained as reflecting the inherent variability in the abandonment of cut-in-fill terraces. See point about fill terraces written at the beginning of the review. It should be however noted that, in the experiment DQsin, there are two slivers of what was probably the original floodplain datum. As such, these slivers should be TA and TB for comparison with IQsin_DQsin.
We clarified in the manuscript that the long profiles and lag-times shown in Fig. 7 and 6, respectively, were extracted from the most extensive terraces surface to resemble common field approaches. We expanded the discussion on incision rates, by relating incision rates to excess transport capacity following Wickert & Schildgen (2019). We included a potential explanation why one of the experiments, in which we reduce $Q_{s,in}$ behaves differently than the other (l.313 -316): "However, while one of the two experiments with a reduction in $Q_{s,in}$ ($DQ_{s,in}$) is consistent with this theory (Fig. 6D), in the other one ($IQ_{s,in}$ _ $DQ_{s,in}$), we observed relatively short lag times (Fig. 6E). These unexpectedly short lag times might be related to how the incision phase was preceded by an aggradation phase (due to an increase in $Q_{s,in}$). Possibly, the system rapidly settled back to the initial conditions because it had not completely adjusted to the preceding increase in $Q_{s,in}$."

p. 12 l. 17: this feedback has also been extensively discussed and explored by Malatesta et al. (2017, JGR).
We implemented this reference (l. 302).

p. 12 l. 19-21: yes, but the two effects mitigate each other. If the incision rate is slow, the later terrace will also not have been lowered that much such that the geometrical difference remains about the same.
The sentence refers to the time when the switch from dominantly lateral erosion to dominantly vertical incision happens. The earlier the switch, the better the preservation of the initial profile. When the channel continues to planate laterally, it lowers the entire bed surface and when rapid incision initiates, the cut-fill terrace has a lower slope than the channel at the onset of the perturbation. Given the observations we make (Fig. 7), lateral erosion and incision do not seem to completely trade-off so as to keep the geometry constant as suggested by the comment. Instead, we see a good preservation of profiles in cases of instant incision (very low lag-times), compared to lower channel profiles in cases with longer lag-times.

p. 13 l. 11-12: The formulation used here suggests that the authors have observed and established ("we found that") this relationship for the first time, along the 2018 Wickert & Schildgen paper. Yet, the fact that terraces have a steeper gradient than the stream's for Qs or Qw forcing is not a new observation or theoretical construct, it is built-in in theory since early fluvial geomorphology work (Mackin, 1948; Meyer-Peter & Müller, 1948; Léopold & Maddock, 1957; Hooke, 1968; Schumm, 1973; Leopold and Bull, 1979; Wells and Harvey, 1987; Harvey et al., 1999; DeLong et al., 2008; Rohais et al., 2012). Recently Malatesta & Lamb (2018) used a derivation of MPM to constrain alluvial slope as an explicit function of Qs and Qw. This passage is one that inspires my earlier suggestion to provide a more complete overview of current knowledge, in particular in terms of theories of transport and geometry.
We did not intend to pretend that we observed those relationships for the first time. We have removed the expression "we found that" entirely from the manuscript. And we expanded the discussion by comparing our observations to predictions from theory (l. 352-355), as well as with the numerical model results (Lane, 1955; Mackin, 1948; Malatesta and Lamb, 2018; Meyer-Peter and Müller, 1948; Wickert and Schildgen, 2019; Wobus et al., 2010). In addition, we added another field example (Pepin et al., 2013) to the one that was already included (Poisson and Avouac, 2004). Taken all those adjustments together, we are convinced that we do not create an impression of novelty about this point anymore.

p. 13 l. 14: I would also point to the absolutely remarkable site of the Gower Gulch alluvial fan in Death Valley. There, a man-made diversion instantaneously changed the hydrology of the catchment leading to sudden incision of the alluvial channel. Details are found in the work of - Troxel, B.W. (1974, ManÂ made diversion of Furnace Creek Wash, Zabriskie Point, Death Valley, California: California Geology, v. 27, p. 219– 223), - Dzurisin (1975,Channel responses to artificial stream capture, Death Valley, California: Geology, v. 3, p. 309–312, doi:10.1130/0091Â 7613(1975)3<309 :CRTASC> 2.0.CO;2.), - Snyder & Kammer (2009), - Malatesta & Lamb (2017). [you will find the two 70's papers on Gower Gulch attached hereby]

We thank the reviewer for this suggestions. However, instead of artificial river capture, we decided to implement another natural example of terraces with reduced slopes due to climatic changes by Pepin et al. (2013) from the southern Central Andes as well as the numerical model exercise by Wobus et al. (2010). Both of their observations are in agreement with our experimental results as well.

p. 13 l.30 - p. 14 l. 9: I am not sure that I follow the argument here. When terrace treads are used to quantify tectonic deformation, the gradient of the terrace does not matter as it is always detrended to retrieve local deformation (e.g. from an anticline, Lavé Avouac, 2000). As long as the tread is straight, tectonic deformation can be well-constrained.

We agree that this paragraph was a little confusing. We have clarified that the slope changes after upstream perturbations (Q_w, $Q_{s,in}$) mainly affect the approach, in which incision rates (and thus uplift rates) are inferred from terrace height-age plots. As incision is higher at the upstream end after upstream perturbation, the terrace height varies along the profile. We have adjusted the text and references accordingly (l. 494-500).

p. 14 l. 12-15: this context could be introduced much earlier in the manuscript to better motivate the study.

This part has been moved to the introduction.

p. 15 l. 7: It can be noted that this illustrates predictions of laws like MPM whereby no geometric change at the downstream end of the reach demands that the sediment flux transport capacity does not change either.

Unfortunately, we do not follow the comment of the reviewer. The sentence refers to changes in upstream sediment supply and the according adjustment of the channel reach. Although the base level at the downstream end is fixed, changed in upstream sediment supply do result in changes of channel geometry, i.e. slope and width of the channel reach.

p. 16 l. 6-7: Wouldn't chemical signals be best transferred during phases of bypass? Or is recycling more important in such phase than during aggradation?

We would expect that recycling due to lateral movement plays a greater role during bypass that during an aggradation event. Bypass, in the sense of no net deposition or erosion because the channel is in equilibrium, does not exclude the mixing of older and younger material during lateral movement. However, we have added another sentence stating that the degree of signal modification is a function of the mixing- ratio of fresh and remobilized material (l. 516-518).

p. 16 l. 18-19: I understand that these are observations from the runs, but I think it would be advisable to add that these "findings" validate existing theories. Though grammatically correct, the word suggests an unwarranted degree of novelty to my ears (non-native english hearing ears, mind you). That is well known and demonstrated already. The same comment is also valid for point 5 of the conclusion.

This part of the conclusion has been rewritten and the sentence, the reviewer refers to, was deleted.

Response to Reviewer #2:

First, we would like to thank reviewer #2 for the feedback on our manuscript. Reviewer #2 provided general as well as specific comments which we will address below. While the original review comments are shown in italics, our responses are given in regular font. Our line numbers refer to the newly submitted version of the manuscript.

General Comments:

This well written paper describes a set of seven flume experiments in a sand box in order to mimic conditions and controls of fill-terrace formation. The main controls explored are changes in water Qw and sediment Qs discharge and changes in base level. The paper gives a nice and consistent description of current terrace formation theories, models and controls. It gives a clear description of the experiments and relates them in a transparent way to current model insights on fluvial dynamics. The derived conclusions are supported by the sand box experimental evidence but the translation to field evidence is not equally well considered and not always supported by evidence (there a quite some constraints related to the physical experiments). The main limitation of this investigation is that all results and relationships found are only valid for a flume sand box system which cannot be linearly scaled up to real world system without some critical considerations and reflections.

First of all is the sand box experiment dealing with a relatively short and steep fluvial system with Qw,in = Qw,out. The setup resembles, in a qualitative way, more an alluvial fan system than a large mature fluvial system that are usually studied in the cited terrace studies.

First, we thank the reviewer for the positive feedback on our work.

We further acknowledge that upscaling is common problem when transferring experimental results to natural settings. Therefore, we included a new section within the discussion that addresses the different limitations of our experimental setup (new section 4.4). In this section we discuss the following limitations:

(1) We investigate the transfer zone separated from the erosion zone, such that any natural coupling between hillslope processes and channel activity is not included.
(2) We treat $Q_{s,in}$ and Q_w as two independent parameters, although they are known to be coupled in natural systems.
(3) We only investigate a single, braided channel and can therefore make no statements about tributary – main stem interactions or terraces forming in meandering rivers.
(4) We have geomorphically effective flow 100% of the time. As natural rivers have variable flow conditions, including times of no geomorphic activity, a direct comparison of lag times or response times is complicated.
(5) Discussion on number of experiments and reproducibility.

We agree that the channel system is relatively short and steep, but includes the fundamental feedbacks: water and sediment inputs, base level, and a channel that responds to these forcings. And we do believe that our setup differs from an alluvial fan setting because it has a narrow, defined outlet, which ensures that the river stays within a confined valley. Typical processes observed on alluvial fans during aggradation are gradual channel migration and avulsion (sudden changes in channel position), which results in an overall widening of the actively reworked alluvial fan area in downstream direction. This can be seen for example in experimental setups from Whipple et al. (1998), Kim et al. (2006a,b) and Martin et al. (2009). In our experimental setup, however, the confined outlet forces the river to stay 'in place' and limits avulsions. In addition, alluvial fans are often characterized by superelevation, i.e. elevation in the central part can be higher compared to the fan margins. As our main purpose is to study process-behavior of a system, we assume that the preservation of processes (e.g., dominance of lateral migration over avulsion) is more important than the absolute scaling of the slope. The slope of the river is a function of the sediment supply and water discharge. As such, we could have chosen a gentler slope of the river. The reason for the stepper references slopes was to produce pronounced differences in channel geometry within all the different settings.

Also, the results of the experiment are qualitatively similar to those of the numerical alluvial channel simulations of Wickert & Schildgen (2019).

Secondly, is the 'fluvial system' studied a braided system only, while many studied and cited terrace systems are thought to be initiated when the fluvial system switched from a braided to (more) meandering state (and back).

We indeed cite field studies of terraces that were formed within braided as well as within meandering channel systems. The purpose of the introduction is to give an overall overview of the different processes of terrace formation.

However, a large number of the studies cited refer to terraces that were formed in braided channels system only. These studies include for example: Scherler et al. (2015), Schildgen et al. (2016), Tofelde et al. (2017), Norton et al. (2015), Faulkner et al. (2016), Fuller et al. (1998), Malatesta et al. (2018), Malatesta and Avouac (2018), Bookhagen et al. (2016), McPhillips et al. (2014), Dey et al. (2016), Steffen et al. (2009), Steffen et al. (2010) and Litty et al. (2016).

Nevertheless, we agree that our experimental approach is restricted to terrace formation in braided systems only. For clarification, we added an extra paragraph to the discussion stating that our setup restricts us to only investigate terraces that form parallel to the main stem and that we cannot investigate any terrace formation at the junctions between the main stem and tributaries or terraces that form due to meander cut-off (l. 449 - 458). We also clarified in the abstract and introduction, that our transfer zone is represented by "a single braided channel in non-cohesive sediment".

Finally has the used methodology the issue of reproducibility. If we would repeat the same experiments in the same sand box would we get the same terraces (properties) and results? This is crucial to know because the laser scanning allows us to measure very small changes (with known uncertainties) but if there is significant other uncertainty ('noise') in the sand box data of a higher magnitude we might be over interpreting the data. As long as we do not know the 'noise' in the experiments we should be reluctant to draw too many conclusions from relative minor changes in elevation. I recommend to address these potential limitations in the discussion in a separate section.

This is a good point and we agree that reproducibility is crucial. In the set of experiments contained within this manuscript we only repeated the control experiment (*Ctrl_1 and Ctrl_2*). The purpose of the control experiments was to investigate 'noise' within the system. We only interpret changes in morphology that are beyond the variability within the control experiments as externally driven adjustments.

Although we did not repeat the experiments that included external perturbations with exactly the same settings, we consider the last phase of the two experiments during which we performed two changes ($DQ_w\_IQ_w$ and $IQ_{s,in}\_DQ_{s,in}$) as repetition of the experiments with only one perturbation (IQ_w and $DQ_{s,in}$), although with different absolute values of Q_w and $Q_{s,in}$. The comparison of those experiments with each other reveals that the trajectories of channel evolution (longitudinal profiles, slope, width (Fig. 5 and 6)) is robust. In addition, the final Q_s and Q_w settings of the experiments with two changes ($DQ_w\_IQ_w$ and $IQ_{s,in}\_DQ_{s,in}$) were equal to the reference settings (*Ctrl_1*, *Ctrl_2* and 'spin-up' time setting of all experiments but *BLF*). When comparing the slope values to which all those sub-experiments evolve, the final slope values are very similar (around 0.07). Although not being exact repetitions of the same experiments, the evolution to the same equilibrium conditions is an indicator that the results are reproducible.

However, we agree that despite the apparent repeatability based on two different experiments, our number of repeat experiments is very limited. We therefore included a new paragraph to the discussion elaborating on these points (l. 464-473).

Having raised these concerns I do believe the experiments generate an interesting set of criteria and hypotheses that could and should be more rigorously tested on real world systems and be evaluated in numerical models. I will certainly test some of the proposed relationships on existing terrace field evidence and with numerical modelling. I therefore recommend to publish this publication after revisions.

Thank you.

Specific comments:

The validity of the results and relationships observed are certainly more valid for fluvial fan type settings where also transport distances are relatively short and gradients are steep and we only observe braided behavior. In such real world systems we actually do observe differences in gradients between different fill type terraces. The large and longer fluvial systems are often characterized by almost parallel gradients of preserved terraces. Often terrace formation and preservation is linked to tributaries causing reach specific changes in Qs and Qw, something that has not been evaluated in the experiments.

As already discussed above, we think that our setting is not entirely representative of an alluvial fan system due to the confined outlet, the absence of superelevation and the dominance of lateral migration over avulsion. In real world systems, terraces along the main stem can be parallel to the active channel (e.g., Hanson et al., 2006; Faulkner et al., 2016), but they can also vary in gradient (e.g., Tofelde et al., 2017; Poisson and Avouac, 2004; Baker and Gosse, 2009; Burgette et al., 2017). As terrace sequences along the main stem can be up to tens of kilometers in extent, changes in slope might not be so obvious locally and can only bet determined by detailed surface elevation surveys of those terraces.

We agree that many terraces are preserved at confluences of tributary channels and the main stem. Within this set of experiments, we only focus on terraces that form along the main stem to keep the setting as simple as possible and investigate the direct effects of changes in Q_s, Q_w or base-level on changes in bed elevation and terrace formation. Adding a tributary channel adds another level of complexity due to possible internal feedback mechanism between the main stem and the tributary. We also have performed experiments in which we focus on the interaction of a tributary and the main stem. This work is currently in preparation. We think that including another set of experiments, with a detailed focus on tributary-main stem interactions, would overload this manuscript and also draw the focus in a different direction. However, for clarification, we will add an explaining sentence that this study only investigates terrace formation along the main stem.

The link between landscape dynamics and Qs,in is another scaling challenge. Landscapes often display a delay between environmental changes and sediment flux responses. These response lags can be even an order magnitudes larger than the lag-times within the fluvial system itself. This is related to coupling and decoupling of hillslope dynamics to the fluvial system.

We agree that changes in sediment supply from the hillslopes to the channels can lag behind any changes in environmental conditions that might cause an adjustment of the supply rate. In this study, we only investigate the response of the fluvial part (=transfer zone) to variations in input conditions, and we do not have the ability to address lag times between environmental forcing and hillslope responses (erosion zone). Following pioneers like Stanley Schumm and Philipp Allen, we consider a sedimentary source-to-sink system as systems that can be subdivided into three zones – the erosion zone, the transfer zone and the deposition zone. Each of those zones has its own responses and response timescales to external perturbations. We only investigate the transfer sub-system of a source-to-sink sediment transport system. The transfer sub-system connects the erosion zone (hillslopes) with the final deposition zone (e.g. a terrestrial or marine basin). As such, we only investigate response or lag-times of the transfer sub-system and do not investigate delays between sediment supply from hillslopes to river channels. Although we have stated this in the original manuscript (p. 2 l. 2, p.2 l. 17-19, p. 6 l. 4 of original version), we clarify this in the introduction by introducing the new figure 1, as well as by stating that we only investigate Qs-modifications within the transfer zone (e.g., l. 110-114, 529-530 and many more).

In additions, we include a paragraph in the discussion stating that we cannot investigate the potential coupling between the hillslopes (erosion zone) and channel (transfer zone) with our setup (l. 436-441).

The autogenic dynamics analysis requires more thought. We can only discard them if they do not occur after longer repeated runs under 'stable' conditions. It seems there is more autogenic dynamics related in the transient response of channel width, an aspect in the model results that are not as detailed analyzed as the terrace profiles, surface slopes and signal propagation.

We apologize, but this seems to be a misunderstanding. We do not discard autogenic terrace formation. On p. 11 l. 3-10 (original version) we state that we did not observe any autogenic terrace formation after the 'spin-up' time, but that the absence of such terraces does not mean that autogenic terraces do not exist. We also state that most mechanisms of autogenic terrace formation, could not be tested with our experimental setup. This part of the discussion has been moved to the section 4.4 (Limitation of experiments) and states:

"…the lack of terrace formation in the two control experiments after the 'spin-up' time does not imply that autogenic terraces do not exist in natural systems, because several potential mechanisms of autogenic or complex-response terrace formation like meander-bend cut-off (Erkens et al., 2009; Gonzalez, 2001; Limaye and Lamb, 2016; Womack and Schumm, 1977) or internal feedbacks between the main-stem and tributaries (Schumm, 1979, 1973, Gardener 1983, Schumm and Parker 1973, Slingerland and Snow 1988) could not be tested with our experimental set-up.

I like the prediction that net deposition along the channel leads to the majority of the grains at the outlet being freshly delivered from hillslopes (assuming hillslope coupling). While during incision older material is reworked in the outlet material, potentially yielding older ages (with cosmogenics).

Thank you.

In terms of the boundary conditions of the physical experiments I have the following remarks/questions: How realistic is a constant Qs,in input? In reality sediments are released as sediment waves into the fluvial system.

We agree that sediment supply from hillslopes to the channel (erosion zone to transfer zone) can be highly variable. The further downstream transport of the sediment in the river, however, is then limited by the availability of water. As alluvial rivers are limited by their transport capacity and not by the availability of sediment, we consider the $Q_{s,in}$ for a given channel reach within the transfer zone as less variable compared to sediment supply from hillslopes to the channel itself. For clarification, we have adjusted the text such that our experiments only investigate the geomorphic response of the transfer zone of a source-to-sink system (see comment above). The constant water discharge prescribed in the experiments is also a difference to natural channels that are dominated by variable discharges. In a way, we are 'compressing time' and assume that the experiments integrate over a number of large floods in natural channels; therefore, the timescales cannot be scaled directly (see new paragraph in section 4.4., l. 459-463).

How important are the initial conditions? (referring to initial channel and 'spin-up' phase).

We assume that the initial conditions play a minor role as we only look at changes in the system once the system is close to equilibrium. If the initial conditions were different, we expect the time to reach steady conditions to be longer or shorter (depending on the initial conditions). The two experiments during which we performed two changes ($IQ_s\_DQ_s$ and $DQ_w\_IQ_w$), both result at the initial slope value after the conditions have been changed back to reference conditions (Fig. 6C, E). As such, we expect the initial conditions mainly affect the 'spin-up' time required to reach stable conditions.

What is the effect of stopping the experiment for the laser scanning? Doesn't this 'disturb' the experiment? A comparison between two equal runs with and without stopping could answer this issue? If this has been investigated before, please cite the relevant literature on this.

Unfortunately, it is not possible to scan the surface without stopping the experiment for two reasons: (1) The laser scanner is mounted directly above the setting and it scans the surface in five lines parallel to the flow direction. Those five lines largely overlap and are merged after finishing the scans. The scanning of all five lines requires about 5 min. A continuation of the experiment would alter the surface morphology during the scanning time, such that the overlapping parts could not be merged anymore. (2) The water supplied to the experiments is dyed blue (Fig. 2). The reason is to enable the automatic detection of wet and dry pixels from the overhead photos. For the automatic detection, significant color differences between the water and the surrounding sand is necessary. The laser scanner, however, cannot penetrate the dyed water. As such, the experiments have to be stopped to be able to scan the surface topography. For the two reasons listed above, a comparison as suggested by the reviewer is unfortunately not possible.

However, the experiments have also been stopped overnight. In those cases, a laser scan was performed after stopping the experiment in the evening and before starting it again in the morning. The DEM of difference (DOD) between those scans reveal no major changes in topography for example through drying of the surface and collapse of channel banks. Finally, the time to drain the system took only a few minutes, and therefore does not leave a lot of time for significant reworking of the surface. Our approach is common for these type of experiment, and so far, there is no indication in the literature that it causes significant problems.

You give temporal lags in measured time. How would you scale this up to reality? (see fig 5)

As already mentioned above (see new paragraph in section 4.4., l. 459-463), our experiments are simple in a way that sediment supply and water discharge are constant through time, such that we assume that the experiments integrate over a number of large floods in natural channels. This makes an absolute scaling of channel response time and lag-times between perturbation and terrace abandonment complicated. Rather, we see the advantage of our approach that we can observe the form of the response (e.g. decrease in slope follows and exponential pattern and not a linear one). As such, we can differentiate whether a terrace was abandoned instantly after the onset of perturbation or rather later during the transient channel response phase.

A difference between the Qw and Qs experiments compared to the base level change scenarios is the there is far less accommodation space in the upper part for terrace preservation (a narrow steep incision) compared to the downstream section and its response to base level change. Shouldn't this not be included in the impact analysis of perturbations?

This is an interesting point. We agree that if the channel widens downstream (as channels tend to do in real systems), there is indeed more "space" to accommodate terraces downstream than upstream. That might be reflected in the width of terraces formed upstream and downstream. Because of the fixed outlet, we have a limited capacity of the system to widen downstream, and therefore are not sure we can make a strong statement about downstream changes in accommodation space with our setup.

I fully agree with the statement that simulating long-profile evolution requires an improved understanding of the transient response of channel width. I presume that the Wickert and Schildgen, 2018 relationship between S, Qs ,in and Qw are also only valid for braided sand box systems under transport limited conditions?

Wickert and Schildgen (2019) derive a general set of equations for gravel-bed river long-profile evolution -- meaning that flows are bedload-dominated and lack bedforms. They also note that transient width response is a needed direction of future research, and limit their approach to the assumption that such channels will tend to have a near-equilibrium width (e.g., Parker, 1978), which is appropriate for gradual changes discharge or other drivers of width change. This equilibrium width is set such that the Shields stress against the bank is equal to the critical Shields stress for initiation of motion, which is also appropriate for experiments such as ours, in which the banks are not held together by cohesive forces. For further questions on this study, we refer the reviewer to the final (2019) published paper.

Also, the detailed discussion of channel width evolution has been removed from the revised version of the paper.

This also implies uniform 'bedrock' lithology. In reality (all cited real world examples) tectonic stability doesn't exist, nor do uniform lithologies or transport limited conditions. I am not suggesting to exclude the comparison but be more sensitive of the differences.

In our experimental setup, we only study alluvial rivers. Therefore, uniform bedrock lithologies are not of major concern compared to studies of bedrock channels, in which a lowering in slopes requires the erosion of bedrock, which indeed is influenced by lithology. In our case, the material that needs to be moved is sediment, and we consider its lithology of minor importance.

As already noted above, we make the assumption that the system is always in 'transport-limited' conditions. The same conditions characterize some of the cited real world examples. Several of the cited field studies refer to braided, alluvial rivers in mountain basins that are characterized by massive alluvial fills (Tofelde et al. (2017), Schildgen et al. (2016), Dey et al. (2016), Malatesta et al. (2018), Malatesta and Avouac (2018), Scherler et al. (2015), Huntington (1907), Litty et al. (2016), Steffen et al. (2009, 2010)). In those settings, the amount of sediment that is transported out of the basin is restricted by the transport capacity of the river. As such, we consider those sites to be in transport-limited conditions. Concerning tectonic stability, we agree that it is unlikely to be maintained over very long time periods, but even over the millennial timescales that many alluvial features are formed, it is not uncommon to find areas where there is no substantial change in tectonic forcing.

The view of terraces/floodplains as temporal storage space is a realistic one. The percentage of Qs,in is in temporary storage during experiment in total in time, in Fig 5 could be used to quantify this effect and the possible effect on cosmogenic age.

Thanks for the suggestion. Indeed, as the absolute $Q_{s,in}$ values are known, we can quantify the percentage of sediment discharge ($Q_{s,out}$) that has been supplied from upstream ($Q_{s,in}$) and that has been remobilized from within the channel. In figure 6 (bottom panel) we plot $Q_{s,in}$ and $Q_{s,out}$, both normalized to the reference value of 1.29 ml/s (as stated in the methods). As such, the numbers read from the y-axis multiplied by 1.29 give the absolute volumes of $Q_{s,in}$ and $Q_{s,out}$ for each point in time. As the values are all normalized by the same value (1.29), the ratio of $Q_{s,out}$ and $Q_{s,in}$ tells us how much sediment has been remobilized within the channel compared to $Q_{s,in}$. For example in the IQ_w experiment, the $Q_{s,out}$ increases to about 20 right after the doubling in discharge, while the $Q_{s,in}$ stays at 1. Consequently, 20 times as much sediment has been remobilized from the transfer zone compared to the upstream supply.

Response to Reviewer #3:

First, we would like to thank reviewer #3 for taking the time to review our submission. Reviewer #3 provided general, as well as line-by-line comments, which we will address below. While the original review comments are shown in *italics*, our responses are given in regular blue font. Our line numbers refer to the newly submitted version of the manuscript.

First, I enjoyed reading this well written manuscript. I appreciate that the authors crafted an accessible background literature review (from the perspective of a nonexperimentalist). In their manuscript, Tofelde et al., develop interesting and timely scientific questions and knowledge gaps–what are the responses of alluvial fill terraces to modulation of base level, and changes in upstream water discharge and sediment supply (Qw, Qs respectively)–which they then address using seven experiments. I echo the sentiment of Reviewer 2 that this paper has the ring of a review paper, yet that is not a problem for me, and I actually appreciated the good explanations of current knowledge (theoretical, field, and experimental). I thought the amount of review in the introduction was appropriate to bring a non experimentalist/expert up to speed on the current thinking of how terrace incision-aggradation functions with respect to changes in upstream or downstream (base level) boundary conditions. I thought the figures are well made and that the captions are effective as well.

Thanks for this kind assessment.

The results of the seven experiments performed by the authors show there are distinct responses in the slope of pre-perturbation and post-perturbation alluvial surface elevations that are dependent upon the type of forcing mechanism, and the authors document interesting transient behavior of fill terrace, channel elevations/width, and Qs out of the experimental system with time. In experiments with increased Qw or Qs, gradients in the new equilibrium channels decrease significantly compared to the pre upstream perturbation channel gradients. This is a somewhat intuitive, yet interesting result, and one that presumably has the potential to be tested in the sedimentary/geomorphic record. I thought that the rationale for the experiments and the results are thought provoking to those interested in not only morphologic response of alluvial fill terraces to external forcing, but also the implications of their response to external forcing in terms of chemical signatures preserved (or not) in sediment/sedimentary systems (end of Section 5).

The experimental design did not include simulations of increased Qw + Qs, or decreased Qw + Qs, as conceivably might occur/be expected in a natural sedimentary system undergoing upstream changes in boundary conditions. Thus its possible the C2 results of these experiments (pure perturbations in Qw or Qs) may be difficult to invert from sedimentary records or be more pronounced in experiments than nature. I don't consider this a shortcoming of the manuscript, it's just an observation, and perhaps the authors could include a statement about this in the discussion?

We agree that changes in environmental conditions (e.g. tectonics, climate) that have the potential to affect either Q_s or Q_w are likely to affect both in reality. For example, a change to wetter conditions (increase in Q_w) might also trigger a pulse of sediment release from the hillslopes to the channels (e.g. Steffen et. al (2009, 2010)). Thus, considering the entire sediment routing system, Q_s and Q_w are often coupled. With our experimental set-up, however, we only investigate the response of the transfer sub-system to changes in surrounding conditions, and we de-couple Q_s and Q_w to investigate the potential effect that each of those two parameters can have on the evolution of channel morphology. Also, although both parameters are thought to vary simultaneously, thick fluvial fills and fill terrace formation in the field are often related to either significant changes in either Q_s or Q_w (hillslope-driven and discharge-driven models as described in Scherler et al. (2015); see p.3 l.27 to p.4 l.4 of the original version). As such, we investigate the two end-members of those models. Many variations in-between those endmembers are possible though. For clarification, we included the above mentioned points within the discussion section on 'Limitations of experiments' (l. 442-448).

Other reviewers have suggested ideas to help improve the communication of what results are novel by the restructuring of the literature review and parts of the discussion (e.g. Malatesta's comment #2). I concur that the authors should consider improving the way in which they communicate how to interpret these experimental results in the context of existing theoretical and experimental knowledge.
We agree with both reviewers that the introduction on the theoretical background should be extended. Please see our reply to Malatesta's comments for details on how we intend to adjust the section on background knowledge.

Recommendation: I recommend that this manuscript ultimately be accepted for publication after the authors implement minor revisions.

Line-by-line comments:
P1 L9 suggest "…tectonic histories" rather than "…tectonic conditions"?
We appreciate this suggestion, however, we prefer to maintain the term 'tectonic conditions' for the following reason: Terraces form under certain environmental conditions. As such, the terraces can be used to reconstruct those certain conditions that persisted at a certain point in time. They are not a continuous archive (as for example a varved lake core would be). Therefore, fill terraces cannot be used to infer entire climatic or tectonic histories.

P2 L20-21 You may want to specify that (at least for Schaller et al 2004) the methods used to interpret paleo discharge were in part based on cosmogenic nuclide concentrations, not simply the age of terrace formation. Interpretations from those concentrations are in turn subject to assumptions of the systematics of cosmogenic nuclides and sedimentary dynamics.
The main point about this sentence was to state that fill-terrace deposits have been used in various ways, including for example the reconstruction of paleo-discharge or paleo-denudation rates. Schaller et al. (2004) did not reconstruct discharge, but paleo-denudations rates. If we explained the Schaller work in detail, we would also need to explain the other applied approaches, which would not benefit the purpose of the sentence. As such, we prefer to not extend the explanation. Please also note that this sentence has been moved and slightly rearranged within the new structure of the introduction.

P3 L10 The following sentence needs to be rewritten: "To our knowledge, there are no experimental studies that systematically compare how fill terraces formed through various mechanisms may differ from one another, or investigate the impacts of terrace formation on downstream sediment discharge."
With the new structure of the introduction, the sentence has been changed to "To our knowledge, there are no experimental studies that consider the combined evolution of two records of landscape evolution – fill terraces in the transfer zone and sediment discharge to the deposition zone – in response to environmental perturbations."

P5 L33 The end of the second Section (2 Formation of fluvial fill terraces) seems abrupt; would it help to provide one or two statements that help summarize and transition into Section 3 here?
Section 2 has been removed from the new version of the manuscript.

P8 L2 Suggest "channel incision" rather than "river incision"?
OK, was changed.

P9 L23-26 "When comparing terrace slopes to the active channel slopes (blue lines) at the end of each run, terrace slopes are steeper in all experiments in which upstream conditions (Qw, Qs,in) were changed 25 (Fig. 6 A-D). In contrast, the slopes of the terraces and the active channel in the BLF experiment are similar to each other (Fig. 6E)." This is a really interesting relationship, and one I would not have expected (though I don't often think about these kinds of experiments), but that does seem intuitive. Is this pre-perturbation terrace slope and upstream-downstream boundary condition relationship something that is seen in other experimental studies? In nature? I see your discussion includes some mention of this explicitly, and introduces the active tectonic aspect that unfortunately complicates interpretations and adds non uniqueness to potential interpretations of terrace slope history. Can you predict/offer guidelines for which kind of natural systems your experimental results would be best applied?
Variability in terrace slopes has been reported from field studies (e.g., Tofelde at el., 2017; Baker and Gosse, 2009; Burgette et al., 2017; Poisson and Avouac, 2004), while others have observed parallel or semi-parallel terrace surface slopes (e.g., Faulkner et al., 2016; Hanson et al., 2006). The slope-comparison is one of the parameters that should be investigated to identify the main terrace driving mechanism. However, it cannot stand alone, as both changes in Q_w or $Q_{s,in}$ can results in a reduction in slope. As such, this characteristic should be seen in combination with other observations, as we have summarized in the new figure 9.

P10 L22 add a space after "…Fig 5)."
OK.

P15 L30-31 Perhaps cite the figure # again for clarity, for which grey vs. yellow circles relate to this sentence.
Done.

P16 L6-8 The last sentence of Section 5 suggests chemical signals may be propagated more efficiently through systems during phases of aggradation, rather than phases of incision when mixing of older stored sediment might overprint the chemical signature of "fresh" hillslope derived sediment. This is interesting…Your statement makes sense, however would it also be fair to say that the chemical signature would be a function of the ratio of the "fresh" to recycled sediment (and obviously the erosion rate upstream)? And that those ratios could vary greatly given different system scales (I'm thinking about the ratio of upstream derived Qs vs excavated volume)? Perhaps this is a tangential idea more suitable for its own paper?!
Following the thoughts of the reviewer, we have added a sentence stating that the degree of signal modification is a function of the mixing- ratio of fresh and remobilized material (l. 516-518).

Alluvial channel response to environmental perturbations: Fill-terrace formation and sediment-signal disruption

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Abstract. The sensitivity of fluvial systems ~~fill terraces~~ to tectonic and climatic boundary conditions allows us to use the geomorphic and stratigraphic records as quantitative ~~make them potentially useful~~ archives of past climatic and tectonic conditions. Thus, fluvial terraces that form on alluvial fans and floodplains as well as the rate of sediment export to oceanic and continental basins are commonly used to reconstruct paleo-environments. However, we currently lack a systematic and quantitative understanding of the transient evolution of fluvial systems and their ~~impacts of base-level, water discharge, and sediment discharge changes on terrace formation and~~ associated sediment storage and release in response to changes in base level, water input, and sediment input. Such~~. This~~ knowledge is necessary to quantify ~~gap precludes a quantitative inversion of~~ past environmental change~~changes~~ from terrace records or sedimentary deposits, and to disentangle the multiple possible causes for terrace formation and sediment deposition.~~terraces.~~ Here, we use a set of seven physical experiments to explore terrace formation and sediment export from a single, braided channel ~~system~~ that is perturbed by changes in upstream water discharge or~~and~~ sediment supply, or through downstream base-level fall. Each perturbation differently affects (1) the geometry of terraces and channels, (2) the timing of terrace cutting~~formation~~, and (3) the transient response of sediment export from the basin.~~discharge.~~ In general, an increase in water discharge leads to near-instantaneous channel incision across the entire fluvial system and consequent local terrace cutting, thus preserving ~~preservation of~~ the initial channel slope~~profile~~ on terrace surfaces, and it also produces a transient increase in sediment export from the system~~. that eventually returns to its pre-perturbation rate.~~ In contrast, a decreased ~~changes in the~~ upstream sediment~~-~~ supply rate may result in longer lag ~~-~~times before terrace cutting, leading to terrace slopes that differ from the initial~~a less well-preserved pre-perturbation~~ channel slope~~profile~~, and ~~may~~ also lagged responses~~produce a gradual change~~ in sediment export~~output towards a new steady-state value~~. Finally, downstream base-level fall triggers the upstream propagation~~migration~~ of a diffuse knickzone, forming terraces with upstream-decreasing ages. The slope~~gradient~~ of terraces triggered by base-level fall mimics~~mimicks~~ that of the newly-adjusted active channel, whereas slopes~~gradients~~ of terraces triggered by a decrease~~variability~~ in upstream sediment discharge or an increase in upstream~~or~~ water discharge are steeper compared to the new equilibrium channel. By combining fill-terrace records with

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constraints on sediment export, we can distinguish among environmental~~Our findings provide guidelines for distinguishing between different types of~~ perturbations that would otherwise remain unresolved when using just one of these records~~interpreting fill terraces and sediment export from fluvial systems~~.

1 Introduction

Sediment-routing systems are commonly subdivided into three zones: a sediment-production zone, typically a mountainous region; a transfer zone of alluvial and fluvial systems that transport and/or temporarily store sediment; and a sedimentation (or deposition) zone, comprising continental or oceanic basins (Fig. 1; Allen, 2017; Castelltort and Van Den Driessche, 2003). Because climate and tectonics can affect sediment production rates, any changes in those conditions may lead to the formation of fluvial terraces in the transfer zone or changes in sedimentation rates in the deposition zone (Alloway et al., 2007; Bull, 1990; Scherler et al., 2015; Zhang et al., 2001). Many past studies have used such records to reconstruct paleoenvironmental conditions (fluvial terraces: Litty et al., 2016; Poisson and Avouac, 2004; Schaller et al., 2004; sedimentation rates: Hay et al., 1988; Zhang et al., 2001). Quantitative interpretations of either record, however, require a clear understanding of how terraces are formed or how sedimentary signals are altered in the transfer zone (Romans et al., 2016 and references therein). In addition, both records suffer from ambiguity, because variability in different environmental parameters can produce similar sedimentary responses. For example, changes in either sediment or water inputs can create fill terraces (Scherler et al., 2015) and affect sediment deposition rates (e.g., Armitage et al., 2011; Simpson and Castelltort, 2012).

Alluvial rivers adjust their slope and width with respect to the local base-level such that in a graded (steady) state, the incoming water discharge (Q_w) can transport the incoming sediment supply ($Q_{s,in}$) downstream (Buffington, 2012; Gilbert, 1877; Lane, 1955; Mackin, 1948). When graded, the slope (S) scales nearly linearly with the ratio of $Q_{s,in}$ and Q_w (e.g., Blom et al., 2017; Malatesta and Lamb, 2018; Parker, 1979; Wickert and Schildgen, 2019):

$$S \propto \left(\frac{Q_{s,in}}{Q_w}\right) \qquad (1)$$

Changes in boundary conditions (Q_w, $Q_{s,in}$, and base level) therefore cause alluvial rivers to adjust their geometries through sediment deposition (aggradation) or incision, until a new graded profile is reached. Incision or aggradation result from the dependence of bedload-transport capacity on slope and water discharge (Meyer-Peter and Müller, 1948). For example, if Q_w increases while $Q_{s,in}$ is held constant, the transport capacity exceeds $Q_{s,in}$, which leads to the entrainment of additional sediment from the channel bed, thus incision. As incision proceeds, the channel slope decreases until the transport capacity drops to match $Q_{s,in}$. Conversely, if $Q_{s,in}$ exceeds the transport capacity of the channel, sediment will be deposited to steepen the channel, thus increasing the transport capacity until it matches $Q_{s,in}$. These adjustments can be recorded through (1) fill-terrace formation in the transfer zone (e.g., Bridgland and Westaway, 2008; Bull, 1990; Merritts et al., 1994) and (2) changes in sediment export to basins (e.g., Allen, 2008; Castelltort and Van Den Driessche, 2003; Romans et al., 2016).

Fluvial fill terraces form when rivers incise their formerly deposited sediments (Bull, 1990; Howard, 1959), preserving former channel floodplains as terrace surfaces in a process we call “terrace cutting”. Such changes in channel-bed elevation can be triggered by changes at the upstream end of the river, namely the sediment to water discharge ratio, $Q_{s,in}/Q_w$ (eq. 1; e.g., Dey et al., 2016; Scherler et al., 2015; Schildgen et al., 2016; Tofelde et al., 2017), or by base-level changes at the downstream end (e.g., Fisk, 1944; Merritts et al., 1994; Shen et al., 2012). Drivers for terrace formation through the first mechanism include climatically driven variability in Q_w (Hanson et al., 2006; Penck and Brückner, 1909; Scherler et al., 2015; Schildgen et al., 2016; Tofelde et al., 2017) and variability in $Q_{s,in}$, due to, for example, changes in regolith-production rates on hillslopes (Bull, 1991; Norton et al., 2015; Savi et al., 2015), changes in vegetation cover (Fuller et al., 1998; Garcin et al., 2017; Huntington, 1907), exposure of additional regolith following glacier retreat (Malatesta et al., 2018; Malatesta and Avouac, 2018; Savi et al., 2014; Schildgen et al., 2002) or changes in landslide activity (e.g., Bookhagen et al., 2006; McPhillips et al., 2014; Scherler et al., 2016; Schildgen et al., 2016). River incision and terrace cutting through an upstream-migrating knickzone have been related to changes in glacio-eustatic sea-level (Fisk, 1944; Merritts et al., 1994; Shen et al., 2012) or lake-level (Farabaugh and Rigsby, 2005). In some cases, internal dynamics of the system, sometimes referred to as “autogenic processes”, lead to terrace formation that cannot be directly linked to external forcing (e.g., Erkens et al., 2009; Limaye and Lamb, 2016; Malatesta et al., 2017; Patton and Schumm, 1981; Womack and Schumm, 1977) .

When studying terraces in the field, it can be difficult to distinguish between terraces that mark a sudden switch from aggradation or stable conditions to incision (“fill-top” terraces of Howard, 1959) and those that preserve surfaces that were cut by a river moving laterally during a period of overall incision (“fill-cut” terraces of Bull, 1990 and Pazzaglia, 2013). In the latter case, there can be a substantial lag between the onset of the environmental perturbation and the abandonment of the terrace surface (e.g., Steffen et al., 2010, 2009). Consequently, from fill terraces alone, both the formation mechanism (change in Q_w, $Q_{s,in}$ or base level) and the timing of the perturbation can be ambiguous.

Numerical and experimental work has demonstrated that the geometrical adjustment of alluvial rivers to external perturbations not only creates fluvial terraces, but also affects sediment discharge at the outlet ($Q_{s,out}$; Allen and Densmore, 2000; Armitage et al., 2013, 2011; Bonnet and Crave, 2003; Simpson and Castelltort, 2012; Tucker and Slingerland, 1997; van den Berg van Saparoea and Postma, 2008; Wickert and Schildgen, 2019), which may be recorded by changes in sedimentation rates in the deposition zone. For example, increases in either Q_w or $Q_{s,in}$ increase $Q_{s,out}$ (Allen and Densmore, 2000; Armitage et al., 2013, 2011; Bonnet and Crave, 2003; Simpson and Castelltort, 2012), but each has a characteristic signature. Whereas a change in $Q_{s,in}$ triggers a permanent change in $Q_{s,out}$, a change in Q_w leads to a transient change in $Q_{s,out}$ (Armitage et al., 2011; Bonnet and Crave, 2003; Wickert and Schildgen, 2019). However, because environmental forcings can be cyclic rather than step changes, it can be difficult to relate variability in sedimentation rates to a distinct forcing. Moreover, changes in $Q_{s,out}$ in response to changes in $Q_{s,in}$ or Q_w may be buffered, amplified, or directly transmitted through sediment routing systems (Armitage et al., 2013; Godard et al., 2013; Romans et al., 2016 and references therein; Simpson and Castelltort, 2012). We propose that to correctly interpret changes in sedimentation rates, the modifications of $Q_{s,in}$ within the transfer zone (then referred to as Q_s) due to sediment deposition (channel aggradation) and remobilization (channel incision) must be understood.

The long temporal and broad spatial scales of fill terrace-formation and sediment deposition preclude direct observations of their potential links in nature. Numerical models provide an inroad to understand the evolution of river long profiles and/or $Q_{s,out}$ after perturbations (Blom et al., 2017, 2016; Malatesta et al., 2017~~Sediment is moved across the Earth's surface from the production zone (mountainous regions), through the transfer zone (fluvial channels and floodplains), to the final depositional zone (continental and oceanic sedimentary basins) (Allen, 2017; Castelltort and Van Den Driessche, 2003). Because sediment production in mountainous regions is thought to vary with climatic and tectonic conditions, any changes in those conditions may be reflected in the sedimentary deposits in the transfer or depositional zones (Alloway et al., 2007; Zhang et al., 2001). However, reliable reconstructions of past conditions from sedimentary deposits require a detailed understanding of sediment transport along the sediment-routing (or source-to-sink) system, including any potential alteration of signals through the transfer zone, as well as the preservation of the sedimentary deposits and its signals over time (Romans et al., 2016 and references therein).~~

~~Fluvial fill terraces represent transient sediment storage along river channels, and therefore they are an important component of the sediment-routing system (e.g., Allen, 2008). They are generated by variations in river-bed elevations due to sediment deposition followed by river incision into the formerly deposited sediments (Bull, 1990). As a result of incision, remnants of the former floodplain can be abandoned by the active channel and preserved as terraces, a process we refer to as "terrace cutting". Fill terraces, as such, are an indicator of unsteadiness in the parameters that control fluvial-channel geometry. Aggradation and incision can be triggered by changing conditions at the upstream end of the river, namely the sediment to water discharge ratio,~~ ~~$Q_{s,in}/Q_w$~~ ~~(e.g., Buffington, 2012; Gilbert, 1877; Lane, 1955; Mackin, 1948), or by base-level changes at the downstream end (e.g., Fisk, 1944; Merritts et al., 1994; Shen et al., 2012). In some cases, internal dynamics of the system, sometimes referred to as "autogenic processes", may lead to terrace formation which cannot be directly linked to any external forcing at the upstream or downstream end of the channel (e.g., Erkens et al., 2009; Limaye and Lamb, 2016; Malatesta et al., 2017; Patton and Schumm, 1981; Womack and Schumm, 1977). The cutting of terraces can either coincide with or lag behind the onset of the perturbation that drives terrace formation. The formation of fill terraces in response to external perturbations has two major implications: (1) fill terraces potentially provide a record of past environmental conditions (e.g., Bridgland and Westaway, 2008; Bull, 1990; Merritts et al., 1994); and (2) the deposition and erosion of fill terraces can alter downstream sediment signals, complicating signal propagation from catchment headwaters to long-term depositional sinks (e.g,. Allen, 2008; Castelltort and Van Den Driessche, 2003; Romans et al., 2016).~~

~~Fill-terrace deposits have been used to infer past variability in discharge (Litty et al., 2016; Poisson and Avouac, 2004) or sediment supply (Bookhagen et al., 2006; Schaller et al., 2004). For a reliable reconstruction of such parameters, however, it is essential to understand how closely terrace formation tracks environmental perturbations. Because most studied fill terraces are thousands to millions of years old and form over the course of years to thousands of years (e.g., Bookhagen et al., 2006; Schaller et al., 2004; Schildgen et al., 2002, 2016; Tofelde et al., 2017), fill-terrace formation can rarely be observed~~

~~directly in nature. Consequently, we need alternative ways to investigate the formation of fill terraces and their impacts on downstream sediment discharge.~~

~~Numerical models provide an opportunity to predict the evolution of alluvial river-bed elevation over time (Blom et al., 2017, 2016~~; Simpson and Castelltort, 2012; Slingerland and Snow, 1988; Wickert and Schildgen, 2019), but most simulate river-profile ~~2018). However, those predictions commonly are limited to the~~ evolution without taking~~of the longitudinal profile and do not take~~ into account modifications of ~~the~~ channel width or terrace formation~~the cutting of terraces~~ (Blom et al., 2017, 2016; Simpson and Castelltort, 2012; Slingerland and Snow, 1988; Wickert and Schildgen, 2019). In addition, most numerical models for river-profile evolution rely on equations derived for the steady-state case. As such, they may not accurately simulate transient responses, which are important for capturing terrace formation and modifications of $Q_{s,in}$ in the transfer zone. ~~). Hancock and Anderson (2002) modeled bedrock strath terrace formation, a partially analogous process, but their erosional stream-power-based approach cannot be easily translated to transport-limited systems, where slope and long-profile evolution result from both sediment and water inputs.~~

Physical experiments provide an alternative approach to studying the dynamics of the transfer zone, including terrace formation (Baynes et al., 2018; Frankel et al., 2007; Gardner, 1983; Lewis, 1944; Mizutani, 1998; Schumm and Parker, 1973; Wohl and Ikeda, 1997) and the evolution of $Q_{s,out}$ (Bonnet and Crave, 2003; Van den Berg van Saparoea and Postma 2008). ~~).~~ Most experimental studies have tested the cutting of terraces due to base-level fall (Frankel et al., 2007; Gardner, 1983; Schumm and Parker, 1973), ~~)~~ or explained their cutting~~formation~~ through autogenic processes (Lewis, 1944; Mizutani, 1998). Only one experimental study by Baynes et al. (2018) investigated terrace formation as a response to changes in ~~sediment supply (~~$Q_{s,in}$~~)~~ or ~~water discharge (~~Q_w~~)~~, but this study focused on vertical incision into bedrock and strath-terrace cutting. Van den Berg van Saparoea and Postma (2008) and Bonnet and Crave (2003) investigated~~performed experiments to investigate~~ the effects of variability~~pulses~~ in Q_w and $Q_{s,in}$ on topograpic~~the~~ evolution ~~of longitudinal channel profiles~~ and ~~sediment discharge at the basin outlet (~~$Q_{s,out}$~~)~~, but neither considered how these processes may be linked to ~~they did not focus on~~ terrace formation. To our knowledge, there are no experimental studies that consider the combined evolution of two records of landscape evolution – ~~systematically compare how~~ fill terraces in the transfer zone and sediment discharge to~~formed through various mechanisms may differ from one another, or investigate~~ the deposition zone – in response to environmental perturbations~~impacts of terrace formation on downstream sediment discharge~~.

In this study, we present results from seven physical experiments of the transfer zone, represented by a single braided channel~~channels~~ in non-cohesive sediment, in which we perturb Q_w, $Q_{s,in}$, and base level. We investigate the timing and geometrical response (slope, width) of the alluvial channel in the transfer zone (with a particular focus on ~~to test three potential mechanisms of~~ fill-terrace cutting) and patterns and response rates of $Q_{s,out}$, with a particular focus on how the records may be linked and if a combination of both records can be diagnostic of specific changes in boundary conditions.

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2 Methods

To test the impact of different~~ due to~~ external forcings on fill~~-perturbations: (1) an increase in Q_w, (2) a reduction in $Q_{s,in}$, and (3) a fall in base level. We furthermore monitor our experiments for terrace cutting related to autogenic processes. Subsequently, we discuss: (1) channel responses to perturbations in external forcing and conditions for~~ terrace formation in the transfer zone~~, (2) differences in lag-times between the onset of the perturbation and the timing of terrace cutting and consequent differences in terraces profiles, (3) the relationship between terrace surface slope and the terrace formation mechanism,~~ and ~~(4) the effects of fluvial aggradation or bed incision on~~ sediment export to ~~discharge at the outlet of~~ the deposition zone, we performed seven experiments at the Saint Anthony Falls Laboratory in Minneapolis, USA, in 2015 (Table 1). The experimental setup consisted of a wooden box with dimensions of 4 m x 2.5 m x 0.4 m (Fig. 2A) that was filled with quartz sand with a mean grain size of 144 µm. At the inlet, we supplied sand and water~~river system ($Q_{s,out}$).~~

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~~2 Formation of fluvial fill terraces~~

~~Fluvial terraces form in response to perturbations that happen either upstream ($Q_{s,in}$, Q_w), or downstream (base-level changes) along the river. Such perturbations may be the result of environmental changes (external or allogenic perturbations), or the result of internal (autogenic) dynamics within the system. For each external or internal forcing mechanism, we summarize below observations from field studies, numerical models, and physical experiments.~~

~~2.1 Sediment to water discharge ratio ($Q_{s,in}/Q_w$)~~

~~Alluvial rivers adjust their slopes and widths such that, in a graded (steady) state, the incoming water discharge (Q_w) can transport the incoming sediment ($Q_{s,in}$) downstream (Buffington, 2012; Gilbert, 1877; Lane, 1955; Mackin, 1948). Scherler et al. (2015) referred to terrace formation related to changes in Q_w as the '*discharge-driven model*'. In this model, a reduction in Q_w leads to valley aggradation due to deposition of sediment on the riverbed. A subsequent phase of increased Q_w can then cause incision. In contrast, the '*hillslope-driven model*' requires variability in $Q_{s,in}$. When an increased $Q_{s,in}$ exceeds the sediment-transport capacity of the river, the excess sediment is deposited. Deposition of sediment elevates the channel bed, increases its slope, and thereby increases the sediment-transport capacity of the river until it matches the incoming sediment supply, $Q_{s,in}$. If $Q_{s,in}$ is reduced such that the sediment-transport capacity exceeds the sediment supply, the river tends to incise. The incision both supplements $Q_{s,in}$ with material from the channel bed and lowers the channel slope, thereby decreasing its transport capacity towards an equilibrium with the new $Q_{s,in}$.~~

~~Terrace formation due to variability in Q_w has mainly been related to climatic changes, such as those caused by glacial-interglacial cycles (Penck and Brückner, 1909). Field studies favor this model when times of valley aggradation coincide with drier conditions and incision coincides with wetter conditions (Hanson et al., 2006; Scherler et al., 2015;~~

~~Schildgen et al., 2016; Tofelde et al., 2017). Variability in $Q_{s,in}$ to river channels can have a variety of causes, including climatically driven changes in regolith production rates on hillslopes (Bull, 1991; Norton et al., 2015; Savi et al., 2015), climatically driven vegetation growth that stabilizes sediment on hillslopes (Fuller et al., 1998; Garcin et al., 2017; Huntington, 1907), and exposure of regolith following glacier retreat (Malatesta et al., 2018; Malatesta and Avouac, 2018; Savi et al., 2014; Schildgen et al., 2002). Landslides also deliver sediment to rivers, and the rate of landsliding can vary in response to changes in tectonic rock uplift rates or precipitation (e.g., Bookhagen et al., 2006; McPhillips et al., 2014; Scherler et al., 2016; Schildgen et al., 2016). Increases in precipitation can mobilize additional sediment from hillslopes until the climate returns to a drier state (Dey et al., 2016) or until hillslopes are stripped bare (Steffen et al., 2010, 2009). All of the above interpretations are based on a temporal link between the formation of fill terraces and climate proxy data, and suggest that variability in Q_w and/or $Q_{s,in}$ can drive terrace formation.~~

~~Numerical models have been developed to investigate the evolution of fluvial terraces in response to variable Q_w and $Q_{s,in}$ (Boll et al., 1988; Veldkamp and Vermeulen, 1989; Veldkamp and Van Dijke, 1998), and model results have been compared to different terrace sequences in Europe (Meuse River: Bogaart and van Balen, 2000; Tebbens et al., 2000; Maas River: Veldkamp and Van Dijke, 2000; Allier River: Veldkamp, 1992, Veldkamp and Van Dijke, 1998). Similarities between modeled terraces and field observations support the conclusion that terraces can form in response to variable Q_w and/or $Q_{s,in}$.~~

~~2.2 Base-level changes~~

~~Fluvial terraces can also be the product of changes in base level at the downstream end of the river. A drop in base level locally creates a steeper channel gradient at the downstream end. To return to a steady-state profile, the channel typically incises into its bed through an upstream-propagating knickzone, which, in the case of alluvial channels, can be highly diffuse (Begin et al., 1981; Grimaud et al., 2015; Whipple and Tucker, 1999; Wickert and Schildgen, 2018). A rise in base level leads to a local reduction in channel slope at the downstream end. To return to a steady-state profile, the channel deposits sediment upstream of the location of base-level rise to increase the slope again. Fluvial fill terraces can thus be formed in response to alternating phases of base level rise and fall.~~

~~Although either tectonic or climatic forcing can lead to changes in base level, alternating rises and falls are most commonly associated with climatic forcing. Early observations in the Lower Mississippi Valley (USA) related valley aggradation to a glacio-eustatic sea-level highstand and marine transgression, whereas valley incision and consequent terrace cutting was linked to sea-level fall (Fisk, 1944; Shen et al., 2012). Other field studies have related terrace formation to climatically driven alternations of sea level (Merritts et al., 1994) or lake level (Farabaugh and Rigsby, 2005). Sediment aggradation associated with sea-level rise followed by incision during sea-level fall has also been shown by a numerical model that aimed to model the evolution of the Meuse terrace sequence in Europe (Tebbens et al., 2000; Veldkamp and Tebbens, 2001). In addition, terrace cutting following base-level drop and upstream knickzone migration has been produced in flume experiments (Frankel et al., 2007; Gardner, 1983; Schumm and Parker, 1973).~~

~~2.3 Complex response and autogenic processes~~

~~In addition to external (i.e., allogenic) forcing described above, internal dynamics can also drive terrace formation. Internally-driven terrace formation can result from internal feedbacks in response to a change in boundary conditions ('complex response') or due to purely internal dynamics with constant boundary conditions ('autogenic' processes). Below, we distinguish between complex responses and autogenic processes, and we discuss how they may lead to terrace development.~~

~~A non-linear response within the channel system to a linear external change can be considered a complex response (Schumm, 1979, 1973). For example, field observations (Faulkner et al., 2016; Schumm, 1979; Womack and Schumm, 1977), physical experiments (Gardner, 1983; Schumm and Parker, 1973), and numerical models (Slingerland and Snow, 1988) indicate that several terrace levels may be cut in response to a single drop in base level. Schumm (1979, 1973) observed that incision of the main stem lowered the base level for the tributaries, which consequently started to incise and transport additional sediment to the main stem. The elevated sediment supply in turn exceeded the transport capacity of the main stem, triggering deposition in the formerly incised channel. Once the tributaries were adjusted to the new base level, sediment supply decreased, which triggered renewed incision of the main stem into the recently deposited material. Whereas the initial, externally-driven base-level drop created a first terrace level, all subsequent terraces were formed in response to internal feedbacks within the fluvial system and therefore cannot be directly linked to an external perturbation.~~

~~In contrast to a complex response, we consider autogenic terraces to be those that are formed in response to non-linear processes within the fluvial system under constant external boundary conditions. One example is a meander cut-off, which can occur without any external perturbation and leads to a local increase in channel slope. The resulting increase in bed shear stress triggers incision and subsequent terrace formation. This phenomenon has been observed in the field (Erkens et al., 2009; Gonzalez, 2001; Womack and Schumm, 1977) and has been replicated using numerical models (Limaye and Lamb, 2016). Another example is local storage and release of sediment, which results from and feeds back into locally non-uniform sediment transport rates. By storing or releasing sediment, each section of the channel changes the local boundary condition on the segment directly downstream (~~$\cancel{Q_{s,in}/Q_w}$~~) or upstream (bed elevation and thus slope). Consequently, sediment deposition, channel incision, and terrace formation can happen simultaneously in different parts of the channel (Lewis, 1944; Patton and Schumm, 1981).~~

~~3 Methods~~

~~To test the dynamics of fill-terrace formation in response to different external forcing conditions and the impact of terrace formation on sediment transport across the transfer zone of a source-to-sink system, we performed seven experiments at the Saint Anthony Falls Laboratory in Minneapolis, USA, in 2015. The experimental setup consisted of a wooden box with~~

~~the dimensions of 4 m x 2.5 m x 0.4 m (Fig. 1A) that was filled with quartz sand with a mean grain size of 144 μm. At the inlet, sand and water were supplied~~ through a cylindrical wire-mesh diffuser filled with gravel to ensure sufficient mixing of sand and water. Water discharge (Q_w) and sediment supply ($Q_{s,in}$) could be regulated independently of one another.~~separately.~~ At the downstream end, water and sand ($Q_{s,out}$) exited ~~the basin~~ through a 20 cm-wide gap that opened onto the basin floor below. This downstream sink was required to avoid deltaic sediment deposition that would, if allowed to grow, eventually raise the base level of the upstream fluvial system. At the beginning of each experiment, we shaped an initial channel ~~was shaped~~ by hand (Fig. 2A~~1A~~) and ran the experiment~~experiments were run~~ under reference conditions ($Q_{w,ref}$ = 95 ml/s, $Q_{s,ref}$ = 1.3 ml/s) for 240 minutes. This runtime was sufficient to reach a quasi-steady state in which the average $Q_{s,out}$ approximately equaled $Q_{s,in}$. After this ‘~~“~~spin-up’~~up”~~ phase, the channel had a uniform equilibrium slope of approximately 7%.

Every 30 min, we stopped the experiments to measure topography using~~perform a scan with~~ a laser scanner mounted on the railing of the basin that surrounded the wooden box. Digital elevation models (DEMs) created from the scans have a horizontal and vertical resolution of 1 mm (Fig. 2B~~1B~~). Using those DEMs, we measured the evolution of channel cross-sectional profiles, longitudinal channel profiles, and surface slopes. Long profiles were calculated by extracting the lowest elevation point in each cross-section at 1 mm increments. By plotting elevation against the distance down the long axis of the box rather than against channel length, resulting slopes are slightly overestimated due to the minor sinuosity of the channels. To directly compare terrace and channel slopes, we extracted 5 cm wide swath profiles along the terrace surfaces and the equivalent stretch of the modern channel. Where~~The width of swath profiles had to be reduced on~~ terraces ~~of the DQ_w_IQ_w and the $IQ_{s,in}$_$DQ_{s,in}$ experiments because terraces in these runs~~ were narrower than 5 cm, we reduced this swath width. Slopes were calculated based on a linear fit through the mean elevation profiles. To assess uncertainties, the root mean square error (RMSE) was calculated between the linear model and the observed data.

Overhead photos were taken every 20 s with a fish-eye lens (Fig. 2C~~1C~~). Distortions of the photos were ortho-rectified in Adobe Photoshop and photos were resampled at 1 mm horizontal resolution to directly overlap with the laser scans. Photos were turned into binary images with values of 1 for wet pixels and 0 for dry pixels. This binarization was performed by transforming the *rgb* (red, green, blue) images into *hsv* (hue, saturation, value) images and then manually defining a hue cut-off for each experiment that best separates wet and dry pixels in the image (Fig. 2D). To distinguish wet and dry pixels by color, the supplied water was dyed blue (Fig. 2C). ~~1D).~~ From the binary images, the number of wet pixels in each cross-section (perpendicular to the basin margin and therefore to the average flow direction) were counted. Analyses were restricted to the areas within the orange box (Fig. 2C~~1C~~, D), because terraces mainly developed in this part of the channel and because we considered this sector at the upstream side of the basin~~channel~~ to be least affected~~unaffected~~ by the fixed location of the outlet. To calculate average channel width, the average number of wet pixels in 1200 cross sections ~~perpendicular to the basin margin (therefore perpendicular to the average flow direction)~~ were counted and are reported with one standard deviation. No overhead photos were taken for the *Ctrl_1* experiment~~,~~ because of an error in the camera installation.

We manually measured $Q_{s,out}$ at 10-minute intervals by collecting the discharged sediment in a container over a 10-second period and measuring its volume. This approach allowed us to estimate whether the system had returned to steady state

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($Q_{s,in} \approx Q_{s,out}$) during the runs. At the same 10-minute interval, we measured bed elevation at the inlet and at the outlet to estimate the spatially-averaged channel slope. We interpreted a constant slope for over more than 30 minutes as additional evidence for a graded (steady state) channel. The data can be found in the supplementary material.

We ran seven experiments to monitor how ~~test the impacts of~~ changes in $Q_{s,in}$, Q_w, and base level affect~~on the~~ channel adjustment, the evolution of fill terraces along the main-stem and sediment discharge at the outlet ($Q_{s,out}$) through time.~~.~~ The experiments are summarized in Table 1. To investigate the effect of Q_w, we ran two separate experiments: in one experiment we doubled Q_w (IQ_w = increase discharge) to 190 mL/s at 240 min (end of the 'spin-up' ~~up~~ time) and in the other experiment we first halved Q_w to 48 ml/s at 240 min and then returned to the initial 95 mL/s at 480 min ($DQ_w\_IQ_w$ = decrease discharge, increase discharge). To test the effect of $Q_{s,in}$, we ran one experiment in which we reduced the $Q_{s,in}$ by 83% to 0.22 ml/s ($DQ_{s,in}$ = decrease sediment supply) at 240 min and another one in which we first doubled $Q_{s,in}$ to 2.6 ml/s at 240 min and then halved $Q_{s,in}$ again to the initial 1.3 ml/s at 480 min ($IQ_{s,in}\_DQ_{s,in}$ = increase sediment supply, decrease sediment supply). All $Q_{s,in}$ and Q_w changes were imposed instantaneously, resulting in a step function in the forcing (Table 1).~~.~~ Immediately before imposing these changes, we covered the near-channel surface with a thin layer of red sand to optically identify the area that was~~is~~ reworked after the change. This method allowed us to distinguish visually between fill-top (covered in red sand) and fill-cut terraces (red sand removed due to continuous overwash). We ran one experiment in which we dropped the base level by 10 cm gradually over 20 min starting at 240 min, resulting in a base-level lowering rate of 0.5 cm/min (*BLF*). For this experiment, we started with a base level higher than in the initial setting by flooding the basin surrounding the wooden box (Fig. 2A~~1A~~). The final base level equaled that~~those~~ of the other experiments. In this experiment, the red sand was applied immediately before the onset of base-level lowering. Additionally, we performed two control experiments in which we made no changes to the initial conditions in ~~order~~ to investigate whether terraces would form in our experiment without any change in external forcing (*Ctrl_1*, *Ctrl_2*).

3~~4~~ Results

Fluvial terraces were cut in the experimental runs IQ_w, $DQ_w\_IQ_w$ (in the IQ_w phase), $DQ_{s,in}$, $IQ_{s,in}\_DQ_{s,in}$ (in the $DQ_{s,in}$ phase) and *BLF* (Fig. ~~2,~~ 3, 4). No terraces were formed after the 'spin-up' time of *Ctrl_1* and *Ctrl_2*. The terraces visible in the cross-section of *Ctrl_2* formed in response to incision during the 'spin-up' phase and did not substantially develop after 240 min (Fig. 4B, red line). We named the terraces to the left of the channel (in downstream direction) T_L and the terraces to the right T_R. In all terrace-forming experiments, both fill-top (red sand) and fill-cut terraces (red sand removed) formed (Fig. 3). Only in the IQ_w, $DQ_w\_IQ_w$ and $IQ_{s,in}\_DQ_{s,in}$ experiments were the fill-top terraces preserved as the most extensive terrace surface, at least on one side of the channel (Fig. 3A, B, D). In the $DQ_{s,in}$ experiment, only a fraction of the fill-top terrace (T_L) survived the transient channel adjustment phase (Fig. 3C). In all experiments that experienced upstream perturbation, fill-top and fill-cut terraces formed only in the upstream half of the sandbox. In contrast, in the BLF experiment, terraces formed in

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the downstream channel reach immediately after the onset of base-level drop, but were mostly destroyed within 30 min (Fig. 3E, G). Later during the BLF experiment, terraces formed in the upstream portion of the sandbox (Fig. 3F).~~3B, red line).~~

Fill-cut terrace cutting lagged minutes to hours behind the onset of the imposed perturbation (Fig. 3). We determined lag times from overhead photos, defined as the time interval between the onset of the perturbation (at minute 240 or 480) and the final time that the future terrace surface was occupied by water. In the two experiments during which we changed Q_w and in the $IQ_{s,in}\_DQ_{s,in}$ experiment, the cutting of~~To form~~ fill-cut terraces began within 6 minutes after the change in boundary conditions (Fig. 3A, B, D). In the IQ_w experiment, for example, the majority of the T_L terrace is a fill-top terrace (0 min lag-time) and only a small part at the downstream end was occupied until 6 minutes after perturbation (Fig. 3A, H). In the $DQ_{s,in}$ experiment, however, several fill-cut terraces formed successively with lag-times between ~14 min and 289 min (Fig. 3C). This experiment was the only one in which a sequence of terraces, instead of a single major surface, developed. In the *BLF* experiment, terrace cutting in the upstream part of the basin began 112 and 117 min after the onset of base-level lowering (Fig. 3F).

Fill-terrace formation requires~~,~~ changes in the channel-bed elevation and ~~channel~~ width of the active floodplain~~are required~~. In our experiments, ~~channel-elevation changes occurred by~~ sediment deposition (aggradation) or erosion (incision) altered the channel-bed elevation (Fig. 5~~4~~). However, these ~~bed-elevation~~ changes were not uniform along the channel.~~ reach (Fig. 4).~~ In the runs *Ctrl_1* and *Ctrl_2,* the longitudinal profiles were stable over time and experienced only minor lowering in bed elevation (max. 4 cm) ~~occurred~~ at their~~the~~ upstream ends~~end~~ (Fig. 5A~~4A~~, B). A sudden increase in Q_w (IQ_w, and the IQ_w phase of $DQ_w\_IQ_w$) or a decrease in $Q_{s,in}$ ($DQ_{s,in}$, and the $DQ_{s,in}$ phase of $IQ_{s,in}\_DQ_{s,in}$) both led to channel~~river~~ incision (Fig. 5C, D, G, H). This incision~~, which~~ was most pronounced at the upstream end, near the changing boundary condition, but ~~(Fig. 4C, D, G, H) and was, in most cases,~~ not recognizable at the downstream end (Fig. 5D~~4D~~, G, H), where the channel-bed elevation was fixed due to the steady base level. Sediment deposition in the channels followed a decrease in Q_w (DQ_w phase of $DQ_w\_IQ_w$) or an increase in $Q_{s,in}$ ($IQ_{s,in}$ phase of $IQ_{s,in}\_DQ_{s,in}$), which was, again, most recognizable at the upstream end of the channel~~reach~~ (Fig. 5E~~4E~~, F). The drop in base level, however, caused maximum incision at the downstream end, and the incision wave migrated upstream as a diffuse knickzone (Fig. 5I~~4I~~).

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Channel ~~The evolution of~~ slope and width changes~~of the active channel~~ were observed in the absence of external perturbations. Channel slopes in the~~tracked through time (Fig. 5). The~~ *Ctrl_1* and *Ctrl_2* marginally decreased ~~experiments only showed a marginal decrease of channel slopes~~ after the 240 min 'spin-up' time from ~0.074 and 0.071, respectively, to around 0.070 and 0.067 (~6 % reduction; Fig. 6A~~5A~~). As such, we consider any change in slope after the 'spin-up' time that is on the same order as those observed in *Ctrl_1* and *Ctrl_2* as ongoing adjustment to the reference condition as opposed to the result of an external perturbation. Channel width in the control experiments varied slowly between ca. 20 cm and 35 cm.

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External perturbations in water and sediment inputs forced the channel width and slope to evolve. An instant doubling of Q_w (IQ_w; Fig. 6B~~5B~~) resulted in a rapid~~, exponential~~ decrease in channel slope that decayed exponentially as the channel approached a new graded state. After approximately 480 min, the slope was reduced from ~0.072 to ~0.043 (40% reduction), and new stable conditions were reached. The doubling of Q_w also triggered an instant narrowing of the channel from ~35 cm to ~15 cm (~57 % decrease), followed by subsequent slow widening.

In contrast, suddenly reducing ~~A sudden reduction in~~ Q_w to half its initial value ($DQ_w\_IQ_w$; Fig. 6C) increased channel~~5C) resulted in an increase in~~ slope from ~0.072 to ~0.085 (18% increase) between 240 and 480 min runtime, and caused~~a widening of~~ the channel to widen from about 25 cm to about 45 cm (~80% increase) during the same time period. The subsequent doubling in Q_w back to its initial value triggered a rapid ~~(nearly exponential)~~ reduction in slope back to the initial ~0.072 (~15% reduction, again following an exponential decay) and an instantaneous narrowing of the channel (~45% reduction) followed by slow widening.

Reducing~~A reduction in~~ $Q_{s,in}$ by 83% ($DQ_{s,in}$; Fig. 6D~~5D~~) triggered ~~a decrease in~~ channel slope to decrease at a slower~~. The~~ rate ~~of decrease was lower~~ than in the IQ_w run;~~,~~ and the new slope stabilized around 0.060~~06~~ (24% reduction from the initial 0.079). An instantaneous decrease in channel width also occurred, but this change was again less pronounced than what we observed in the IQ_w experiment (~33% reduction). We detected no~~No~~ subsequent widening of the channel~~ was detectable~~.

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Finally, increasing ~~An increase in~~ $Q_{s,in}$ ($IQ_{s,in}\_DQ_{s,in}$; Fig. 6E~~5E~~) led to ~~an increase in~~ channel steepening~~gradient~~ from a slope of~~about~~ 0.070 to about 0.078 (11% increase) and increased~~an increase in~~ channel width from about 30 cm to about 55 cm (~83% increase). The subsequent reduction in $Q_{s,in}$ decreased ~~led to a decrease of~~ the channel slope and caused an instantaneous channel narrowing to < 30 cm, followed by subsequent widening back to the initial width of ~30 cm.

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During~~For~~ the base-level fall experiment (*BLF*; Fig. 6F), mean~~5F),~~ channel slope instantly and rapidly increased after the onset of base-level fall from about 0.047 to 0.073 (55% increase), and continued to increase~~it increased~~ at a slower rate ~~further~~ to about 0.08, before decreasing~~lowering~~ back to 0.072. These mean~~However, these slope~~ values, however, average over~~ are simply calculated based on the height difference at the inlet and outlet, ignoring~~ any spatial variability in incision, meaning~~slope along the experiment reach~~ that they do not resolve the details of the diffusive~~is, in the *BLF* experiments, significant due to knickzone~~ propagation of the knickzone. Beyond impacts on slope, the~~. The~~ drop in base level resulted in a sudden decrease~~drop~~ in channel width, followed by three cycles of channel widening and narrowing. In summary, we observed that an increase in Q_w and a decrease in $Q_{s,in}$, resulted in an immediate decrease in channel slope (through upstream incision) and an instant reduction in channel width, whereas a drop in base level caused an increase in channel slope (through downstream incision) and a reduction in channel width (Fig. 6~~5~~).

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~~The time of terrace cutting lagged minutes to hours behind the onset of the perturbation (Fig. 5). Lag times were determined from overhead photos and are defined as the time interval between the onset of the perturbation (at minute 240 or 480) and the last time the future terrace surface was occupied by water. The times given in Fig. 5 refer to the last occupation of the areas for which swath profiles were extracted (Fig. 6 right panel). In the two experiments in which we changed Q_w and in the $IQ_{s,in}\_DQ_{s,in}$ experiment, terrace cutting in the upstream reach of the channel (Fig. 3 right column, Fig. 4; dashed arrows) began within ~5 minutes after the change in boundary conditions (Fig. 5; black arrows). In the IQ_w experiment, for example, the majority of the T_A terrace was cut instantly (no removal of red sand) and only a small part at the downstream end was occupied again until 6 minutes after perturbation (Fig. 2A, B). In the $DQ_{s,in}$ experiment, however, the T_A and T_B terraces were~~

~~cut 297 and 144 min after the perturbation (Fig. 5D). In the *BLF* experiment, terraces in the downstream channel reach were cut immediately after the onset of base-level drop, but were mostly destroyed within 30 min (Fig. 2C, D). Terrace cutting in the upstream part of the basin began 112 and 117 min after the initial perturbation (Fig. 5F).~~

To analyze how well terrace surfaces represent~~the~~ channel slopes~~ bed profiles~~ immediately preceding the time of perturbation~~ were preserved by the terraces~~, we compared the elevation profiles of the~~ two~~ terraces on each side of the channel (yellow and orange lines) with the channel that existed at the onset of the perturbation (red line) (Fig. 7). We sampled elevations across the most extensively preserved terrace surface, regardless of its lag time, in a way that is similar to terrace mapping in the field.~~6).~~ In experiments with increasing Q_w (IQ_w, IQ_w phase of $DQ_w\_IQ_w$) or base-level changes (*BLF*), the elevation profiles of the terraces are similar to the initial floodplain~~channel~~ profile (Fig. 7A~~6A~~, B and E). In cases of changes in $Q_{s,in}$ ($DQ_{s,in}$, $DQ_{s,in}$ phase of $IQ_{s,in}\_DQ_{s,in}$), the terraces were cut at lower elevations than the former channel (Fig. 7C~~6C~~, D). In the $DQ_{s,in}$ experiment, fill-cut terraces on either side of the channel formed at different elevations, with one surface~~terrace~~ about 3 cm below the other (Fig. 7C, 4E).~~6C, 3E; unpaired terraces).~~ In contrast, terrace surfaces~~terraces~~ in the other four experiments are at approximately the same elevation (~~paired terraces) (~~Fig. 4, 7~~6~~). Despite similar elevations~~being paired~~, the slope differences ~~slopes of the two terraces differ from each other by~~ between T_L and T_R range from about 5% ($IQ_{s,in}\_DQ_{s,in}$) to~~and~~ 33% (IQ_w). When comparing terrace slopes to the active channel slopes ~~(blue lines)~~ at the end of each run (blue lines),~~,~~ terrace slopes are steeper (by 20–122%) in all experiments in which upstream conditions (Q_w, $Q_{s,in}$) ~~were~~ changed (Fig. 7~~6~~ A-D). In contrast, the slopes of the terraces and the active channel in the *BLF* experiment are similar to one another (within 11%)~~each other~~ (Fig. 7E~~6E~~).

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Changes in boundary conditions also affected $Q_{s,out}$~~sediment discharge at the outlet~~ (Fig. 6~~5~~, lowest panels). An instantaneous doubling of Q_w (IQ_w; Fig. 6B~~5B~~) resulted in an instant increase in $Q_{s,out}$ to more than 20 times $Q_{s,in}$. This rapid increase was followed by an exponential decay down to the initial $Q_{s,out}$ value. A sudden reduction in Q_w to half its initial value ($DQ_w\_IQ_w$; Fig. 6C~~5C~~) resulted in a decrease in $Q_{s,out}$. The subsequent doubling in Q_w back to its initial value triggered a rapid increase in $Q_{s,out}$ that decayed over time. In contrast, neither the instantaneous reduction in $Q_{s,in}$ by 83% ($DQ_{s,in}$; Fig. 6D~~5D~~) nor the doubling in $Q_{s,in}$ ($IQ_{s,in}\_DQ_{s,in}$; Fig. 6E~~5E~~) triggered a measurable change in $Q_{s,out}$. For the base-level fall experiment (*BLF*; Fig. 6F~~5F~~), $Q_{s,out}$ could not be measured before and during the base level drop~~,~~ because the basin surrounding the wooden box was flooded for this experiment. $Q_{s,out}$ was only measured from minute 280 onwards, which corresponds to minute 40 after the 'spin-up' of the base level fall. At that time, $Q_{s,out}$ was still about 10 times higher than $Q_{s,in}$, and $Q_{s,out}$ decreased approximately linearly from that time onwards.

4~~5~~ Discussion

When attempting to use geomorphic or depositional records to reconstruct paleo-environmental conditions, we face a range of challenges. One challenge is to understand how the information on environmental boundary conditions is translated into the sedimentary record, considering the potential modification of sediment signals during fluvial transport. A second, related challenge is that depositional records and fluvial terraces often cannot unambiguously be associated with a particular forcing mechanism. In the following, we will discuss these two challenges, and we will focus on both records that we monitored in our experiments – fill terraces in the transfer zone and sediment discharge to the deposition zone ($Q_{s,out}$). Further, we will discuss the use of an integrated set of observations to address these challenges, the limitations that arise when comparing the experimental work to natural settings, as well as potential implications of our observations for future field studies.

4.1 Terrace formation in the transfer zone

4.1.1 Conditions of terrace formation, lag times, and the ~~5.1 Channel response to perturbations and conditions of terrace formation~~~~The~~ preservation of pre-perturbation channel profiles

The cutting of fluvial fill terraces requires ~~that~~ vertical incision and a simultaneous reduction~~outpaces lateral erosion on one or both sides~~ of the active floodplain width.~~channel.~~ Whether this occurs depends on the response of alluvial channels to changing boundary conditions, which can include~~occur through~~ adjustments of~~to their~~ slope, wetted perimeter (width and depth), and/or bed-surface texture (grain-size distribution) (e.g., Blom et al., 2017; Buffington, 2012 and references therein; Wickert and Schildgen, 2019).~~(Blom et al., 2017; Buffington, 2012 and references therein).~~ Because the grain-size distribution in our experiments remained constant, we focus our discussion on the externally forced adjustments of channel slope (S) and width (w) during terrace formation.

In our experiments, river incision (with terrace cutting) was driven by an increase in Q_w, a decrease in $Q_{s,in}$, or a fall in base level (Figs. 3 - 6). In the case of base-level fall, incision began at the downstream boundary and diffused upstream, producing a transient steepening. Enhanced Q_w or reduced $Q_{s,in}$, on the other hand, decreased channel slope. The evolution of longitudinal channel profiles in our experiments is in agreement with earlier flume studies that investigated channel response to upstream (van den Berg van Saparoea and Postma, 2008) and downstream (Begin et al., 1981; Frankel et al., 2007) perturbations, as well as with numerical models that predict the evolution of longitudinal profiles following variations in $Q_{s,in}$, Q_w or base level (Blom et al., 2017; Simpson and Castelltort, 2012; Wickert and Schildgen, 2019). In all experiments, incision and terrace cutting coincided with an instantaneous decrease in channel width, while aggradation corresponded to an increase in channel width (Fig. 7).

A common application of fluvial-terrace mapping is to reconstruct paleo-longitudinal channel profiles from terrace remnants (e.g., Faulkner et al., 2016; Hanson et al., 2006; Pederson et al., 2006; Poisson and Avouac, 2004). These profiles are thought to be representative of the former channel profiles, ideally reflecting their geometries immediately prior to a perturbation. However, morphological adjustments of a channel to external perturbations require time, such that the geomorphological response can lag behind the changes in environmental parameters (e.g., Blum and Tornqvist, 2000; Tebbens et al., 2000; Vandenberghe, 2003, 1995). The lag time between an external perturbation and the onset of terrace cutting determines how much time the fluvial system has to modify the terrace sediments before their abandonment. In the following,

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we first discuss the relationship between lag-times and the preserved terrace profiles related to upstream perturbations ($Q_{s,in}$, Q_w), followed by those related to downstream perturbations (*BLF*).

In our experiments,~~In our experiments, river-bed aggradation and channel steepening occurred after a decrease in Q_w and after an increase in $Q_{s,in}$, whereas river incision (with terrace cutting) and channel-slope lowering were driven by an increase in Q_w, a decrease in $Q_{s,in}$, or a fall in base level (Figs. 2, 3, 4). In the case of base-level fall, incision began at the downstream boundary and diffused upstream, producing a transient steepening. The evolution of longitudinal channel profiles in our experiments is in agreement with earlier flume studies that investigated channel response to upstream (van den Berg van Saparoea and Postma, 2008) and downstream (Begin et al., 1981; Frankel et al., 2007) perturbations, as well as with numerical models that predict the evolution of longitudinal profiles following variations in $Q_{s,in}$ or Q_w (Blom et al., 2017; Simpson and Castelltort, 2012; Wickert and Schildgen, 2018). In addition to slope changes, channels can also adjust to external forcing by changing their width (Fig. 5; Buffington, 2012; Church, 1995; Curtis et al., 2010; Dade et al., 2011). In all experiments, an increase in channel width occurred during aggradation (reduced Q_w, increased $Q_{s,in}$), and an instantaneous decrease in channel width occurred at the start of incision (increased Q_w, reduced $Q_{s,in}$, *BLF*; Fig. 5).No terraces were formed during the two control experiments after the 'spin-up' time. However, this finding does not imply that autogenic terraces do not exist in natural systems, as meander bend cut-off (Erkens et al., 2009; Gonzalez, 2001; Limaye and Lamb, 2016; Womack and Schumm, 1977) could not be tested with our experimental setup. We observed internal variability in sediment storage and release, for example in the form of bank collapse due to lateral channel migration during the experiments. However, local lateral sediment input through bank collapse did not trigger terrace formation in our experiments. Our experimental set-up also precluded terrace formation in response to internal feedbacks between the main stem and tributaries (Schumm, 1979, 1973, Gardener 1983, Schumm and Parker 1973, Slingerland and Snow 1988).~~

~~In order to link drivers and response, we turn to the work of Wickert and Schildgen (2018), who coupled equations for flow, sediment transport, and channel morphodynamics to solve for long-profile changes in transport-limited rivers. From this work, in which channel width is allowed to self-adjust following Parker (1978), we distill the following relationships between channel width (w), slope (S) and either $Q_{s,in}$ or Q_w:~~

$$\cancel{Q_w \propto \frac{w}{S^{7/6}}} \qquad \cancel{(1)}$$

~~and~~

$$\cancel{Q_{s,in} \propto w} \qquad \cancel{(2)}$$

~~Eq. 2 predicts the observed reduction in channel width after a decrease in $Q_{s,in}$ (Fig. 5, eq. 2). Eq. 1 predicts that slope should decrease as water discharges increases, which is consistent with the observed decrease in slope from about 0.072 to 0.043 (Fig. 5B) in the IQ_w experiment, in which water discharge doubled. However, this amount of slope decrease should be matched by an 8% increase in channel width, which runs contrary to the observed instantaneous reduction in channel width by ~57% followed by gradual widening. This response is transient, whereas Wickert and Schildgen (2018) assume an equilibrium~~

~~width; the relationship between time-evolving slope, width, and basal shear stress is the most likely cause of this discrepancy. The equilibrium-width solution used by Wickert and Schildgen (2018) assumes a constant ratio between the basal shear stress at bankfull discharge (τ_b) and the critical shear stress for the initiation of sediment motion (τ_c), which can be described by (Parker, 1978):~~

$$\cancel{\tau_b = (1+\epsilon)\tau_c} \qquad \cancel{(3)}$$

~~Parker (1978) suggested that the fraction of excess shear stress at bankfull flow (ϵ) is about 0.2 for self-formed gravel-bed rivers with equilibrium widths. Empirical measurements have confirmed an epsilon of 0.2 in a large number of rivers across the US (Phillips and Jerolmack, 2016), but Pfeiffer et al. (2017) illustrated that ϵ increases in tectonically active regions. It could be that rapid uplift is analogous to incision in our experiment during its transient-response phase, causing the channel to narrow and τ_b to increase, which further accelerates incision. Our experimental results demonstrate that accurately simulating long-profile evolution may require an improved understanding of the transient response of channel width.~~

~~5.2 Preservation of channel profiles~~

~~A common application of fluvial-terrace mapping is to reconstruct paleo-longitudinal channel profiles from terrace remnants (e.g., Faulkner et al., 2016; Hanson et al., 2006; Pederson et al., 2006; Poisson and Avouac, 2004). Reconstructed longitudinal profiles from terrace remnants are thought to be representative of the former channel profiles, ideally of conditions immediately prior to perturbations. However, morphological adjustments of a channel to external perturbations require time, such that the geomorphological response can lag behind the changes in environmental parameters (e.g., Blum and Tornqvist, 2000; Tebbens et al., 2000; Vandenberghe, 2003, 1995). The lag-time between external perturbations and the onset of terrace cutting determines the degree of reworking of terrace material. Consequently, the shorter the lag-time, the better the preservation potential of environmental conditions that existed prior to the time of perturbation.~~

~~In our experiments, the terrace surfaces preserve the former channel elevation profiles in the two increased Q_w experiments and in the *BLF* experiment (Fig. 6A, B and E). In contrast, in the decreased $Q_{s,in}$ experiments, terrace-elevation profiles are lower than the river channel immediately preceding the perturbation and, in case of the $DQ_{s,in}$ run, the terraces are also unpaired (Fig. 6C, D). Focusing on~~ the upstream-perturbation-related terrace surfaces (fill-top and fill-cut terraces) following an increase in Q_w had ~~experiments first, we observed~~ short lag times (≤ 6 min; ~~between perturbations and terrace cutting in all Q_w-related experiments (~~Fig. 3A, B and 6B~~5B~~, C) and preserved~~), which ensured good preservation of~~ the channel elevation profiles~~profile~~ prior to perturbation well (Fig. 7A~~6A~~, B). Similarly, terrace cutting in the $IQ_{s,in}\_DQ_{s,in}$ experiment was characterized by short (T_R~~T_B~~) or no (T_L~~T_A~~) lag times (Fig. 6E~~5E~~). The small discrepancy between terrace slopes and initial channel slopes in this experiment (Fig. 7D) is a result of slope variations between the center of the channel belt (where initial and final channel profiles were measured), and the sides of the channel belt, where the terrace slopes were measured.

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In contrast, terrace cutting in the $DQ_{s,in}$ experiment occurred with a delay of several hours and the terraces were also cut successively (Fig. 3C, 6D).~~hour delay.~~ The difference in lag ~~-~~times between the T_L~~T_A~~ and T_R~~T_B~~ terrace of about two and a half hours resulted in different terrace elevations on both sides of the channel~~unpaired terraces~~, with elevation profiles several cm below the channel profile prior to perturbation (Fig. 7C). These results illustrate how short lag times are critical to enable accurate reconstructions of the pre-perturbation channel-profile, which can potentially be used to reconstruct paleo-environmental conditions. But what determines the duration of the lag-time?~~6C).~~

The length of the lag time between the perturbation and the abandonment of a terrace surface is expected to depend on the ratio of vertical incision versus lateral erosion. Bufe et al. (2018) and Malatesta et al. (2017) demonstrated that the rate of lateral channel migration scales inversely with the height of valley walls (elevation difference between a terrace surface and the active channel). As such, higher incision rates after a perturbation lead to faster wall-height growth and greater reductions in lateral mobility. Accordingly, fast incision should result in short lag times between the onset of the perturbation and terrace cutting, guaranteeing good preservation of the channel profile that existed prior to the perturbation. In contrast, slow river incision and enhanced lateral channel movement can lead to long lag times, with terrace profiles that reflect a channel profile at some (unknown) phase of adjustment. The incision rate, on the other hand, is thought to be a function of the excess sediment transport capacity, and sediment transport capacity should be directly proportional to Q_w (Wickert and Schildgen, 2019): doubling Q_w should correspondingly double the excess sediment transport capacity, whereas halving $Q_{s,in}$ should increase the excess sediment transport capacity by a factor of 0.5. Therefore, increases in Q_w should, in theory, cause more rapid incision, shorter lag-times, and a higher preservation potential for the pre-perturbation channel profile than a proportionately equal reduction in $Q_{s,in}$. However, while one of the two experiments with a reduction in $Q_{s,in}$ ($DQ_{s,in}$) is consistent with this theory (Fig. 6D), in the other one ($IQ_{s,in}\_DQ_{s,in}$), we observed relatively short lag-times (Fig. 6E). These unexpectedly short lag times might be related to how the incision phase was preceded by an aggradation phase (due to an increase in $Q_{s,in}$). Possibly, the system rapidly settled back to the initial conditions because it had not completely adjusted to the preceding increase in $Q_{s,in}$.

~~The length of the lag time between the perturbation and the abandonment of a terrace surface depends on how effectively vertical incision outcompetes lateral erosion. Bufe et al. (2018) have shown that the rate of lateral channel migration scales inversely with the height of valley walls (elevation difference between a terrace surface and the active channel). As such, the higher the incision rate after perturbation, the faster wall-heights grow and the more lateral mobility is reduced. Due to this positive feedback, rapid incision after a perturbation should result in short lag-times between the onset of the perturbation and terrace cutting and a good preservation of the channel profile that existed prior to perturbation. In contrast, if the river incises more slowly, terraces may be cut long after incision initiates, and the terrace profile will not directly reflect the channel profile prior to perturbation.~~

The lag time between the onset of base-level fall and the cutting of terraces in the upstream reach~~part~~ of the channel~~valley~~ is about ~~~~~115 min (Fig. 6I~~5I~~), which was the time required for the knickzone~~knickpoint~~ to propagate upstream. As such, for ~~base-level-fall-related~~ terraces related to base-level fall, the temporal lag between the onset of the perturbation~~base-level fall~~ and terrace cutting increases with ~~increasing distance to the terrace~~ upstream distance. Hence~~. In~~

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~~other words~~, terrace surfaces created through upstream knickpoint migration are diachronous, become progressively younger upstream despite being a physically ~~a~~ continuous unit. ~~Faulkner et al. (2016) found decreasing OSL ages with upstream distance in a fill terrace along the Chippewa River, USA that formed in response to base-level fall.~~ Similar results have been reported from field studies (Faulkner et al., 2016; Pazzaglia, 2013).~~conclusions were also reached by Pazzaglia (2013).~~ In comparison, incision ~~was~~ initiated near-synchronously along the entire experimental channel~~reach~~ when ~~incision was~~ triggered by a change in upstream boundary conditions (*IQ*$_w$, *DQ*$_{s,in}$; Fig. 5C~~4C~~, D). In summary, lag-~~-~~times between the onset of a~~the~~ perturbation and terrace cutting depend on the combination of local incision rates after the perturbation and the trigger for incision (base-level fall vs. a change in upstream conditions).

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~~Lag-times between the perturbation and the onset of terrace cutting can be important when dating the surfaces of fluvial fill terraces in the field. Common methods to date the onset of river incision include the dating of terrace surface material with cosmogenic exposure dating (e.g., Schildgen et al., 2016; Tofelde et al., 2017), dating sand or silt lenses with optically stimulated luminescence close to the terrace surface (OSL; e.g., Fuller et al., 1998; Schildgen et al., 2016; Steffen et al., 2009) or dating embedded organic material with ^{14}C (Farabaugh and Rigsby, 2005; Scherler et al., 2015). When transferring our observations to a field scenario, the ~2h or more of channel material reworking before terraces were cut within the upstream part of the reach in the *BLF* and the *DQ*$_{s,in}$ experiment would result in terrace ages that are younger than the time of perturbation. The best temporal correlations between the perturbation and the terrace surface ages are achieved by those formed by changes in Q_w due to the fast onset of vertical incision and minimal reworking of terrace surface material. To assess the significance of this time-lag in natural systems requires more work on how to scale the experiment to larger channels.~~

~~5.3 Differences in terrace surface slopes~~

~~To reliably use fluvial terraces to reconstruct paleo-environmental conditions (i.e., changes in base level, $Q_{s,in}$ or Q_w), the identification of the terrace formation mechanism is important. We found that for $Q_{s,in}$ or Q_w related terraces, the slopes of terrace surfaces are always steeper than the active channel (the new steady state channel after the perturbation), whereas the slope of terraces formed due to downstream perturbations is very similar to that of the active channel (Fig. 6). Similar observations have been made in the field. Poisson and Avouac (2004) measured a reduction in channel slope between terraces due to deeper incision at the upstream end of a flight of terraces in the Tien Shan. They related the changes in longitudinal profiles (inferred from the terraces) to changes in Q_w. In contrast, Faulkner et al. (2016) measured terraces in the Chippewa River, a tributary to the Mississippi River, which were created in response to base-level fall and upstream knickpoint migration due to incision of the Mississippi channel bed after deglaciation. They observed no major slope change between the longitudinal profile reconstructed from the terrace and the modern channel. According to Wickert and Schildgen (2018), the relationship between slope S, $Q_{s,in}$ and Q_w, for alluvial rivers taking self-adjusting channel width and channel roughness into account, can be described as:~~

$$\cancel{S \propto \left(\frac{Q_{s,in}}{Q_w}\right)^{6/7}} \qquad \cancel{(4)}$$

~~According to this relationship, a decrease in $Q_{s,in}$ or an increase in Q_w results in a lower channel slope. A drop in base level should, after the signal has propagated upstream, result in a slope similar to the channel before the perturbation because the $Q_{s,in}/Q_w$ ratio is unchanged. Hence, our findings suggest that slope comparisons between the terrace surfaces and the active channel could indicate whether an upstream or a downstream perturbation caused the cutting of the terraces. However, such comparisons are only informative if the active channel is still graded to the boundary conditions that initiated incision and terrace cutting. In addition, this approach to identifying the terrace-formation mechanism requires negligible tectonic tilting of the terraces after cutting.~~

~~**In tectonically active regions, both strath and fill terraces have been used to infer tectonic deformation rates (e.g., Hu et al., 2017; Lavé and Avouac, 2000; Litchfield and Berryman, 2006; Peters and van Balen, 2007). Variability in slopes over time, derived from reconstructed longitudinal channel profiles, have been used to infer local deformation rates (e.g., Hu et al., 2017; Lavé and Avouac, 2000). The observed slope differences between terrace surfaces and the active channel after upstream perturbations in our experiments (Fig. 6), however, imply that slope differences observed in the field can only be used to infer tectonic deformation rates if one can either rule out (Lavé and Avouac, 2000) or quantify slope changes related to changing Q_w and/or $Q_{s,in}$ (Pazzaglia, 2013). Because the slope changed in our experiments of upstream perturbations, incision rates were not uniform along the channel (Fig.**~~

4.1.2 Terrace geometry as an indicator of perturbation type

Because fluvial fill terraces result from changes in Q_w, changes in $Q_{s,in}$, or a drop in base level, their presence alone does not indicate which of the parameters changed over time. However, our experimental results revealed differences in terrace geometry between changes in upstream (Q_w, $Q_{s,in}$) versus downstream (*BLF*) conditions. For terraces related to changes in $Q_{s,in}$ or Q_w, the slopes of terrace surfaces are always steeper than the active channel (the new steady state channel after the perturbation), whereas the slope of terraces formed due to downstream perturbations is very similar to that of the active channel (Fig. 7). These observations concur with predictions from theoretical work that suggest a positive scaling of slope and $Q_{s,in}$ and a negative scaling of slope and Q_w, while a drop in base level should, after the signal has propagated upstream, result in a slope similar to the channel before the perturbation because of a constant $Q_{s,in}/Q_w$ ratio (Lane, 1955; Mackin, 1948; Malatesta and Lamb, 2018; Meyer-Peter and Müller, 1948; Wickert and Schildgen, 2019; Wobus et al., 2010). Similar observations have been made in the field. In the Tien Shan, Poisson and Avouac (2004) found a successive reduction in slope within a terrace sequence, which they related to changes in Q_w. In the Central Andes, Pepin et al. (2013) explained downstream-converging terraces (and thus a reduction in terrace-surface slopes) on a piedmont through variability in climatic drivers. In contrast, along the Chippewa River in the USA (a tributary to the Mississippi River), where terrace cutting is linked to base-level fall, Faulkner et al. (2016) found no substantial slope change between the longitudinal profile reconstructed from the terraces and the modern channel.

Our findings support earlier observations that slope comparisons between the terrace surfaces and the active channel could indicate whether an upstream- or a downstream-sourced perturbation caused the cutting of the terraces (Faulkner et al., 2016; Pepin et al., 2013; Poisson and Avouac, 2004; Wobus et al., 2010). However, such comparisons are only informative if the active channel is still graded to the boundary conditions that initiated incision and terrace cutting. In addition, this approach to identifying the terrace-formation mechanism requires negligible or quantifiable tectonic tilting of the terraces after cutting.

4.2 Sediment discharge to the deposition zone

4.2.1 Q_s-signal modification in the transfer zone

In steady-state, the transfer zone experiences no net sediment deposition or removal of sediment. Hence, when averaged over a certain time, $Q_{s,in}$ equals $Q_{s,out}$ and the signal can be considered as faithfully transmitted to the deposition zone (e.g. Romans et al., 2016). During geometrical channel adjustments, however, when sediment is either deposited or eroded to adjust the channel slope, $Q_{s,out}$ differs from $Q_{s,in}$ (Fig. 6). This is schematically shown in figure 8 by the divergence between the solid line ($Q_{s,in}$) and circles ($Q_{s,out}$). Thus, Q_s-signals can be considered as modified as long as the transfer zone is in a transient state. The total time of Q_s-signal modification depends on two variables: (1) the time a certain transfer zone requires to reach graded conditions again, i.e. the channel response or equilibrium time (Castelltort and Van Den Driessche, 2003; Howard, 1982; Métivier and Gaudemer, 1999; Paola et al., 1992) and (2) the frequency at which boundary conditions (Q_w, $Q_{s,in}$ base level) change. Consequently, if the period of the forcing is shorter than the required response time of the channel reach, the Q_s signal will never be faithfully transmitted (Paola et al., 1992).

4.2.2 Observable changes in sediment export to the deposition zone ($Q_{s,out}$)

Regardless of whether the Q_s-signal is faithfully transmitted or modified, we observed changes in $Q_{s,out}$ in our experiments, which would likely be reflected by changes in sedimentation rates within the deposition zone. Enhanced $Q_{s,out}$, for example, was generated both by an increase in Q_w (Fig. 6B) and by a drop in base-level (Fig. 6F). Hence, from the $Q_{s,out}$ record alone, the driving mechanism cannot be identified. The temporary $Q_{s,out}$ peak in the IQ_w experiment (Fig. 6B and schematically in Fig. 8C) resembles the observations of earlier numerical (Armitage et al., 2013, 2011; Tucker and Slingerland, 1997) and experimental work (Bonnet and Crave, 2003; van den Berg van Saparoea and Postma, 2008). In both this earlier work and ours, the peak in $Q_{s,out}$ was generated during the transient phase of slope adjustment. According to equation 1, an increase in Q_w will decrease channel slope and, therefore, trigger river incision. Because $Q_{s,in}$ was held constant during the experiment, the additional sediment that reached the outlet was remobilized from within the channel, in particular from the upstream part (Fig. 5C, G; Castelltort and Van Den Driessche, 2003; van den Berg van Saparoea and Postma, 2008; Wickert and Schildgen, 2019). In contrast, a decrease in Q_w requires a steeper channel slope, which is achieved through sediment deposition within the channel (Fig. 5E). In our experiments, this adjustment appears as a reduction in $Q_{s,out}$ relative to the upstream sediment supply during the transient slope-adjustment phase (Fig. 6C and 8D).

A decrease in $Q_{s,in}$ should, following the achievement of a graded channel profile, reduce $Q_{s,out}$, whereas an increase in $Q_{s,in}$ should increase $Q_{s,out}$ (Allen and Densmore, 2000; Armitage et al., 2011; Bonnet and Crave, 2003). According to equation 1, reducing $Q_{s,in}$ will trigger temporary incision because a lower slope is required to transport less sediment with the same Q_w. Conversely, increasing $Q_{s,in}$ without changing Q_w will require a steeper transport slope and thus trigger aggradation. Channel incision and slope reduction occurred in the $DQ_{s,in}$ experiments (Fig. 5D, H and 6D, E), whereas aggradation and slope increase followed an increase in $Q_{s,in}$ (Fig. 5F and 6E). However, in none of the experiments with variable $Q_{s,in}$ was a clear change in $Q_{s,out}$ recognizable during the transient phase of slope adjustment (Fig. 6D, E and 8E, F). We consider the negative feedback between $Q_{s,in}$ and the bed-elevation change during the transient channel-adjustment phase as the main reason for this lack of response (Simpson and Castelltort, 2012; van den Berg van Saparoea and Postma, 2008). The additional sediment supplied upstream is deposited within the channel, resulting in aggradation, and is therefore not detectable at the outlet. When less sediment is supplied upstream, the channel incises and complements the supplied upstream sediment with remobilized sediment from within the channel, such that once again, no change in $Q_{s,out}$ is detectable at the outlet during the adjustment phase. We did not run the experiments long enough to analyze the adjusted steady-state phase, but we would expect that once the channel has adjusted to new equilibrium conditions, $Q_{s,out}$ will eventually equal $Q_{s,in}$ (Fig. 8E, F; Allen and Densmore, 2000; Armitage et al., 2011; Bonnet and Crave, 2003; Wickert and Schildgen, 2019).

Internal channel dynamics can lead to variability in $Q_{s,out}$ even without external forcing. In the *Ctrl_1* and *Ctrl_2* experiments, scatter in the $Q_{s,out}$ signal was up to 5 times the value of $Q_{s,in}$ (Fig. 6a). This variability is due to continuous lateral movement of the channel and subsequent bank collapse, which results in stochastic contributions of additional sediment. Lateral channel mobility of a stream varies with water and sediment discharge (Bufe et al., 2018; Wickert et al., 2013). However, if the volume of sediment mobilized from valley walls due to lateral migration is much larger than the change in $Q_{s,in}$, then no clear signal in $Q_{s,out}$ might be recognizable, even after channel adjustment. The channel instead will continually adjust to the stochastic lateral input of sediment.

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Regarding $Q_{s,out}$ signals, we conclude that terraces, floodplains, and the channel itself act as a temporary storage space where sediment can be deposited or remobilized when boundary conditions change (Coulthard et al., 2005; Simpson and Castelltort, 2012; van den Berg van Saparoea and Postma, 2008). The consequence of sediment deposition or remobilization during transient response times is that $Q_{s,in}$ differs from $Q_{s,out}$, such that the Q_s-signal can be considered as modified during transient phases of channel adjustment. Our data also support earlier findings by Simpson and Castelltort (2012) and van den Berg van Saparoea and Postma (2008), who concluded from their respective numerical model and physical experiments that Q_w variability creates an amplified, substantial response in $Q_{s,out}$, whereas changes in $Q_{s,in}$ create a dampened response in $Q_{s,out}$ due to the a negative feedback between $Q_{s,in}$ and channel slope. Our experiments, illustrated schematically in Fig. 8, also suggest that Q_w-driven $Q_{s,out}$ changes are temporary, and that as the channel slope adjusts to the new input Q_w, $Q_{s,out}$ evolves back to its initial steady-state value. In contrast, $Q_{s,out}$ signals driven by changes in $Q_{s,in}$ may not be observable during transient channel adjustment, but will occur and persist once the channel has adjusted to new steady-state conditions.

4.3 Combining the records

Both fill terraces in the transfer zone and changes in sedimentation rates in the deposition zone record changes in boundary conditions. Our experiments provide an opportunity to investigate the links between terrace formation and sediment export to the deposition zone, as well as how both of these records, if available, can disambiguate changes in past $Q_{s,in}$, Q_w, or base level.

Sediment discharge at the outlet ($Q_{s,out}$) is a function of (1) upstream sediment supply ($Q_{s,in}$) and (2) sediment deposition or mobilization within the transfer zone. Doubling Q_w while holding $Q_{s,in}$ constant triggers river incision to archive a lower channel slope. This incision leaves behind terraces that are steeper than the modern, graded channel (Fig. 9). The sediment mobilized in the transfer zone during the transient incision phase produces a transient peak in $Q_{s,out}$. Reducing $Q_{s,in}$ also reduces the equilibrium transport slope, causing channel incision and terrace abandonment. However, no peak in $Q_{s,out}$ during the transient phase is visible, as the extra sediment remobilized from the transfer zone is compensated by the preceding reduction in $Q_{s,in}$ (Fig. 9). Finally, incision due to a fall in base level also generates a temporary peak in $Q_{s,out}$ due to the additional sediment remobilized within the transfer zone. The terraces left behind following this base-level-driven incision, however, have slopes parallel to that of the modern, graded channel. In summary, $Q_{s,out}$ reflects a combination of $Q_{s,in}$ and the geometrical adjustment of the transfer zone, which in turn is recorded by fill terraces.

Coupling the two records, if available, provides the opportunity to unambiguously identify the forcing mechanism, which is not possible using either the fill terraces or the deposits alone (Fig. 9). For example, the presence of terraces whose slopes are steeper than the present-day channel, combined with a simultaneous but transient peak in $Q_{s,out}$, points towards a change in Q_w as the main driver. In turn, terraces whose slopes are steeper than the main channel in combination with a lagged reduction in $Q_{s,out}$ point towards a change in $Q_{s,in}$ as the main driver. Finally, a temporary increase in $Q_{s,out}$ in combination with channel-parallel terraces that young in the upstream direction indicates past base-level fall. Complications may arise when the forcing includes a combination of changes in $Q_{s,in}$ and Q_w. Nevertheless, our results point to the potential of combining terrace records with sediment-export data for the reconstructions of paleo-environmental conditions.

4.4 Limitations of experiments

Physical experiments allow for investigations of the isolated influence of individual key parameters on landscape evolution. However, a number of limitations arise when attempting to compare the experimental results to natural settings.

First of all, in natural sediment-routing systems, the three distinct zones of erosion, transfer and deposition (Fig. 1) are coupled to one another (Allen, 2017). Erosion processes on the hillslopes, for example, determine the amount of sediment provided to the transfer zone, i.e. $Q_{s,in}$ (e.g., Dixon et al., 2009; Tofelde et al., 2018). In turn, changes in channel-bed elevation in the transfer zone can affect hillslope-erosion processes (e.g., Hurst et al., 2012; Roering et al., 2007). In our experimental setup, however, we investigate the response of the transfer zone as an isolated feature and can thus not account for any hillslope-channel feedbacks that might lead to additional variations in sediment supply to the channel.

Second, we varied Q_w and $Q_{s,in}$ separately, forcing them to remain independent of one another. In natural systems, however, they are commonly coupled. For example, changes in precipitation can alter both Q_w and $Q_{s,in}$ – directly though changes in rainfall-driven sediment-transport rates from hillslopes to the channel (Bookhagen et al., 2006; Dey et al., 2016; Steffen et al., 2010, 2009) and indirectly through long-term changes in hillslope-stabilizing vegetation types (Garcin et al., 2017; Langbein and Schumm, 1958; Schmid et al., 2018; Torres Acosta et al., 2015; Werner et al., 2018). Those feedback mechanisms between different key parameters (Q_w, $Q_{s,in}$) and between sub-zones of sediment-routing systems will likely complicate the forcing-response behavior of natural systems.

Third, we have only investigated a single, braided channel. Therefore, our experimental set-up does not allow us to investigate channel-geometry adjustments related to feedbacks between the main stem and adjacent tributaries (Schumm, 1979, 1973), and related terraces forming at channel junctions (Faulkner et al., 2016; Larson et al., 2015; Schildgen et al., 2016). Furthermore, we can draw no conclusions on terraces forming in meandering rivers, such as those related to meander-bend cut-off (e.g., Erkens et al., 2009; Gonzalez, 2001; Limaye and Lamb, 2016). As such, the lack of terrace formation in the two control experiments after the 'spin-up' time does not imply that autogenic terraces do not exist in natural systems, because several potential mechanisms of autogenic or complex-response terrace formation like meander-bend cut-off (Erkens et al., 2009; Gonzalez, 2001; Limaye and Lamb, 2016; Womack and Schumm, 1977) or internal feedbacks between the main-stem and tributaries (Schumm, 1979, 1973, Gardener 1983, Schumm and Parker 1973, Slingerland and Snow 1988) could not be tested with our experimental set-up.

Fourth, apart from the step changes in input parameters, $Q_{s,in}$ and Q_w were held constant through time. As the experiments exhibit geomorphically effective flow conditions at all times (intermittency equals 1), we assume that the experiments integrate over a number of large floods in natural channels. Natural rivers in turn, experience a wide range of intermittencies. This variability in natural systems complicates any attempts to scale channel response times and lag-times from experiments to real systems, but also complicates the comparison of real systems with each other.

Finally, we performed a limited number of experimental runs, with only the control experiments being repeated. Although we did not repeat the experiments that included external perturbations, we consider the last phase of the two experiments during which we performed two changes ($DQ_w\_IQ_w$ and $IQ_{s,in}\_DQ_{s,in}$) as repetition of the experiments with only one perturbation (IQ_w and $DQ_{s,in}$), albeit with different absolute values of Q_w and $Q_{s,in}$. Comparing those experiments reveals similar trajectories of channel evolution (longitudinal profiles, slope, width; Fig., 5 and 6). In addition, the same boundary conditions ($Q_{s,in}$, Q_w) persisted during the 'spin-up' phases as well as at the end of the two experiments during which we performed two changes. During those conditions, the channel slopes always evolved to a value of ~0.07. Although not being exact repetitions of the same experiments, the evolution to the same equilibrium conditions indicates that the results are reproducible. But, we acknowledge that further repetitions would improve our ability to quantify variability that is internal to each system.

Despite the above limitations, the performance of physical experiments under controlled conditions allows us to directly link causes and their effects in fluvial systems. As the key parameters ($Q_{s,in}$, Q_w, base level) can be varied independently, physical experiments provide the opportunity to isolate the influence of different environmental parameters on the evolution of landscapes and to test theoretical models (Paola et al., 2001). As many processes acting in landscapes are scale-independent, experimental observations can help to decipher process behavior in natural systems (e.g., Cantelli et al., 2004).

4.5 Implications for field studies

Despite the restrictions when comparing the experimental work to natural systems, the general patterns that we observed do have implications for field studies. First, lag times between the perturbation and the onset of terrace cutting can be important when dating the surfaces of fluvial fill terraces in the field. Common methods to date the onset of river incision include the dating of terrace surface material with cosmogenic exposure dating (e.g., Schildgen et al., 2016; Tofelde et al., 2017), dating sand or silt lenses with optically stimulated luminescence close to the terrace surface (OSL; e.g., Fuller et al., 1998; Schildgen et al., 2016; Steffen et al., 2009) or dating embedded organic material with ^{14}C (Farabaugh and Rigsby, 2005; Scherler et al., 2015). When transferring our observations to a field scenario, the ~2h or more of channel material reworking before terraces were cut within the upstream part of the channel in the *BLF* and the $DQ_{s,in}$ experiment would result in terrace ages that are younger than the time of perturbation. The best temporal correlations between the perturbation and the terrace surface ages are achieved by those formed by changes in Q_w due to the fast onset of vertical incision and minimal reworking of terrace surface material. To assess the significance of this time lag in natural systems requires more work on how to scale the experiment to larger channels.

Second, in tectonically active regions, both strath and fill terraces have been used to calculate river incision rates to infer tectonic uplift rates (e.g., Litchfield and Berryman, 2006; Maddy et al., 2001; Schildgen et al., 2012; Wegmann and Pazzaglia, 2009). Because the slope changed in our experiments of upstream perturbations, incision rates were not uniform along the channel (Fig. ~~4). Litchfield and Berryman (2006) also measured variable fluvial incision rates based on terrace heights at several locations along 10 major rivers located along the Hikurangi Margin, New Zealand. Accordingly, gradients in incision rates along rivers should be interpreted in the context of potential changes to the shape of the longitudinal profile.~~

~~5.4 Signal propagation and implications for stratigraphy~~

~~Alluvial rivers adjust their channel geometry (slope, width, and depth) with regards to incoming Q_w and $Q_{s,in}$ (Lane, 1955; Mackin, 1948). Consequently, a change in input parameters leads to an adjustment in channel geometry through the~~

deposition or remobilization of sediment until new equilibrium conditions are reached (transient phase). The required adjustment time is referred to as the response time of the channel (Paola et al., 1992). We expect that a change in Q_w will trigger a transient response in $Q_{s,out}$ during that adjustment phase, but $Q_{s,out}$ is expected to return to the initial value once the new steady-state channel geometry is reached (Armitage et al., 2013, 2011). In contrast, a change in $Q_{s,in}$ will result in a permanent adjustment of $Q_{s,out}$ once the channel geometry is adjusted to the new conditions (Allen and Densmore, 2000; Armitage et al., 2011).

According to Eq. 4, an increase in Q_w is expected to result in a lower channel slope and, therefore, to initiate river incision. In our IQ_w experiment, we observed an up to 20-fold increase in $Q_{s,out}$ after the perturbation, followed by a return to previous $Q_{s,out}$ values at about 300 min after the perturbation (equivalent to 540 min runtime; Fig. 5B). As such, the $Q_{s,out}$ signal is generated during the transient phase of slope adjustment. This pattern is schematically shown in Fig. 7C. Because $Q_{s,in}$ was held constant during the experiment, the additional sediment that reached the outlet was remobilized from within the channel, in particular from the upstream part (Fig. 4C, G). This result corroborates previous observations from physical experiments (van den Berg van Saparoea and Postma, 2008) and numerical models (Armitage et al., 2013; Simpson and Castelltort, 2012). In contrast, a decrease in Q_w requires a steeper channel gradient, which is achieved through sediment deposition within the channel (Fig. 4E). In our experiments, $Q_{s,out}$ was reduced relative to the upstream sediment supply during the transient slope-adjustment phase (Fig. 5C and 7D).

A decrease in $Q_{s,in}$ should, following the achievement of a graded channel profile, also produce a reduced $Q_{s,out}$, whereas an increase in $Q_{s,in}$ should result in enhanced sediment discharge at the outlet (Allen and Densmore, 2000; Armitage et al., 2011). According to Eq. 4, a reduction in $Q_{s,in}$ will trigger temporary incision, because a lower slope is required to transport less sediment with the same amount of Q_w, whereas an increase in $Q_{s,in}$ will require a steeper slope and thus trigger aggradation. We observed channel incision and slope reduction in the $DQ_{s,in}$ experiments (Fig. 4D, H and 5D, E) and aggradation and slope increase following an increase in $Q_{s,in}$ (Fig. 4F and 5E). However, in none of the experiments with variable Q_s is a clear signal in $Q_{s,out}$ recognizable during the transient phase of slope adjustment (Fig. 5D and E, 7E and F). We consider the negative feedback between $Q_{s,in}$ and the bed-elevation change during the transient channel-adjustment phase as the main reason for this lack of response (Simpson and Castelltort, 2012; van den Berg van Saparoea and Postma, 2008). The additional sediment supplied upstream is deposited within the channel, resulting in aggradation, and is therefore not detectable at the outlet. When less sediment is supplied upstream, the channel incises and complements the supplied upstream sediment with remobilized sediment from within the reach, such that once again, no clear reduction in $Q_{s,out}$ is visible during the adjustment phase. We did not run the experiments long enough to analyze the adjusted steady-state phase, but we would expect that once the channel has adjusted to new equilibrium conditions, the changes in $Q_{s,in}$ will eventually become visible in $Q_{s,out}$ (Allen and Densmore, 2000; Armitage et al., 2011).

Internal dynamics within the channel can lead to variability in $Q_{s,out}$ even without external forcing. In the *Ctrl_1* and *Ctrl_2* experiments, scatter in the $Q_{s,out}$ signal was up to 5 times the value of $Q_{s,in}$ (Fig. 5a). This variability is due to continuous lateral movement of the channel and subsequent bank collapse, which results in stochastic contributions of additional sediment.

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~~Regarding $Q_{s,out}$ signals, we conclude that terraces, floodplains, and the channel itself act as a temporary storage space where sediment can be deposited or remobilized when boundary conditions change (Coulthard et al., 2005; Simpson and Castelltort, 2012; van den Berg van Saparoea and Postma, 2008). Our data support earlier findings by Simpson and Castelltort (2012) and van den Berg van Saparoea and Postma (2008), who concluded from their respective numerical model and physical experiments that signals of Q_w variability create an amplified signal in $Q_{s,out}$, whereas changes in $Q_{s,in}$ create a dampened signal in $Q_{s,out}$ due to the a negative feedback between $Q_{s,in}$ and channel gradient. Our experiments, illustrated schematically in Fig. 7, also suggest that Q_w-driven $Q_{s,out}$ signals are transient, and that as the channel slope adjusts to the new input Q_w, $Q_{s,out}$ evolves back to its initial steady-state value. In contrast, $Q_{s,out}$ signals driven by changes in $Q_{s,in}$ may not be observable during transient channel adjustment, but will occur and persist once the channel has adjusted to new steady-state conditions.~~

5). Litchfield and Berryman (2006) also measured variable fluvial incision rates based on terrace heights at several locations along 10 major rivers located along the Hikurangi Margin, New Zealand. Accordingly, incision rates reconstructed from terraces should be interpreted in the context of potential changes to the shape of the longitudinal profile in addition to tectonic changes.

Third, despite information stored in variable sediment deposition rates in the transfer or deposition zone, information on landscape evolution can also be preserved in the chemical composition of the deposited sediment. Our findings have implications for geochemical signatures of sediment, for example the concentration of cosmogenic ^{10}Be, which is commonly measured to infer catchment mean denudation rates (Bierman and Steig, 1996; Brown et al., 1995; Granger et al., 1996). In cases of channel aggradation, $Q_{s,out}$ is reduced compared to $Q_{s,in}$ due to deposition within the channel (Fig. 8B, D, F). The exported sediment could be sourced from incoming sediment that is not deposited (grey circles) and/or mixes with remobilized sediment within the channel (yellow circles). In general, net deposition along the channel leads to the majority of the grains at the outlet being freshly delivered from local hillslopes, thus carrying the contemporaneous catchment chemical composition at the time of transport. In contrast, during incision, older material stored within the channel, floodplain, and/or terraces is remobilized and contributes to the temporary peak in $Q_{s,out}$ (Fig. 8A, C, E). Shortly after the perturbation, most of the remobilized sediment will be stratigraphically high and relatively young (yellow circles), but older material from deeper layers (orange and red circles) will be progressively remobilized and mixed with young material from upstream. Cosmogenic nuclide analyses along the eastern Altiplano margin (Hippe et al., 2012) and in the Amazon basin (Wittmann et al., 2011) indicate that sediment can be stored within the fluvial system over thousands to millions of years. Remobilization of formerly deposited material and subsequent mixing with fresh hillslope material (incoming sediment) can temporally modify signals stored in the geochemical composition of detrital river sediments (e.g., Tofelde et al., 2018; Wittmann et al., 2016, 2011). The degree of modification is thought to be a function of the ratio between fresh and remobilized material exported at a certain time as well as of the difference in geochemical composition between the fresh and remobilized material. We conclude that modern chemical signals are more likely to be transmitted through the system during aggradation phases, whereas local sediment that has been transiently stored may strongly overprint the signal of modern sediments during times of incision.

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~~6~~ Summary and Conclusion

We performed seven physical experiments to investigate the effects of changing boundary conditions ($Q_{s,in}$, Q_w, base level) on channel geometry and related fill-terrace cutting as well as on sediment discharge ($Q_{s,out}$). To reliably reconstruct paleo-environmental conditions from terraces in the transfer zone or sedimentary deposits in the sedimentation zone, it is important to understand (1) how information on environmental conditions may be modified and eventually transferred into the geologic record, and (2) whether the geomorphic characteristics of terraces or the patterns of sedimentation rates are specific to the forcing mechanism.~~, fill-terrace formation and signal propagation in fluvial sediments. In particular, we recorded the evolution of channel slope and width during adjustment to new boundary conditions. Furthermore, we explored the conditions under which fill terraces form and how well they preserve the channel profile prior to perturbation based on lag-times between the onset of perturbation and terrace cutting, synchronicity of incision along the length of the channel, and the relationship between terrace-surface slopes and terrace-formation mechanisms. In addition, we examined the implications of changing~~

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~~boundary conditions on signal propagation through the sediment-routing system. Our experimental findings can be summarized as follows:~~

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Our experiments provided important insights into how sediment discharge to the deposition zone ($Q_{s,out}$) and the formation of terraces in the transfer zone are coupled. The amount of sediment discharged to the deposition zone ($Q_{s,out}$) is a combination of sediment supply to the transfer zone ($Q_{s,in}$) and its modification in the transfer zone through sediment deposition or remobilization. Deposition or remobilization of sediment within the transfer zone occurs mainly during the transient response phase after a perturbation. Hence, during transient response times, $Q_{s,out}$ does not equal $Q_{s,in}$. One consequence of this finding for field studies is that the geochemical composition of sediment sampled during transient phases does not represent the modern hillslope-conditions. For example the ^{10}Be concentration in detrital sediment can be greatly modified, especially during incision phases, when older sediments within the transfer zone are remobilized.

The same modifications (sediment deposition and remobilization) that alter the Q_s-signal during transient times also form fill terraces in the transfer zone. Increases in Q_w trigger channel incision to archive a lower equilibrium slope. The resulting temporary peak in $Q_{s,out}$ coincides with the cutting of terraces whose slopes are steeper than the main channel. Reducing $Q_{s,in}$ also reduces the equilibrium transport slope, causing channel incision and terrace abandonment. However, no substantial increase in $Q_{s,out}$ occurs during the transient phase, as the extra sediment remobilized from the transfer zone is compensated by the preceding reduction in $Q_{s,in}$. Finally, a drop in base-level causes a temporary peak in $Q_{s,out}$ and the formation of terraces parallel to the modern channel. Hence, if both records are available, the combination of the two can unambiguously identify the main forcing mechanism of channel adjustment. The identification of the mechanism can be important, for example, when using the height of terraces to infer channel incision rates. As upstream perturbations cause greater incision at the upstream end than at the downstream end, incision rates inferred from terrace heights are thought to vary along the profile.

~~1.~~ The cutting of terraces following an upstream perturbation ($Q_{s,in}$, Q_w) requires a period of time (lag-time) expected to be a function incision rate, which in turn is thought to be a function of the excess transport capacity of a channel (Wickert and Schildgen, 2019). Indeed, our experiments showed that greater excess transport capacity leads to faster incision and shorter lag-times, which ensures a better preservation potential of the channel profile that existed prior to perturbation. These lag-times can also be critical for field studies that attempt to link the ages of terraces surfaces to the timing of perturbations, as long lag-times may lead to substantial temporal mismatches.~~An increase in Q_w, a decrease in $Q_{s,in}$, or a drop in base level triggered river incision and terrace cutting, combined with an instantaneous reduction in channel width.~~

~~2. The observed reduction of channel width after an increase in Q_w runs contrary to the expected channel widening under equilibrium conditions. This finding indicates that the transient response of the fluvial system – not captured in the equilibrium relationship between channel width (w), discharge (Q_w) and slope (S) from the coupled equations of Wickert and Schildgen (2018) – may be significant. We suggest that the~~

~~transient channel-width response may lead to an excess shear stress at bankfull flow (ε) that differs from the commonly assumed and encountered value of ~0.2τ_c (Parker, 1978; Phillips and Jerolmack, 2016).~~

3. ~~The lag-time between an external perturbation and terrace cutting determines (i) how well terraces preserve and record the pre-perturbation channel longitudinal profile, and (ii) the degree of reworking of terrace-surface sediment. We found that rapid incision creates terraces that effectively track external forcing and record the pre-perturbation channel profile, whereas slower incision enables lateral migration of the channel, with terraces cut during the transient phase that lag behind the timing of forcing and do not preserve the pre-perturbation channel profile.~~
4. ~~In comparison to incision triggered by changes in upstream conditions ($Q_{s,in}$, Q_w), which occurred near synchronously along the entire channel reach, incision triggered by base-level fall created the upstream migration of a knickzone. Consequently, the lag-time between the drop in base level and the cutting of a terrace surface increased with distance upstream. Due to increased surface reworking with distance upstream, the preservation potential of the channel surface prior to perturbation decreases with distance upstream.~~
5. ~~Terraces related to upstream perturbations ($Q_{s,in}$; Q_w) were always steeper than the active channel at the end of the experiment. In contrast, the final, adjusted channel slope was similar to the initial channel slope in the base-level fall experiment. This difference can help to identify the terrace-formation mechanism in field settings, but complicates the interpretation of terraces as tectonic deformation markers.~~
6. ~~Changes in Q_w caused a measurable signal in $Q_{s,out}$ during the transient phase of channel adjustment, whereas $Q_{s,out}$ signals related to changes in $Q_{s,in}$ were not detectable during the transient phase due to buffering (sediment storage or release) of $Q_{s,in}$ as the channel adjusted its gradient. Changes in $Q_{s,in}$ are thought to become more recognizable once the channel has adjusted to new steady-state conditions. Because Q_w-driven signals generated an amplified $Q_{s,out}$ signal during the transient channel response phase, they have a higher potential to be preserved in the stratigraphic record than do changes in $Q_{s,in}$ if upstream conditions are changing periodically with a period that is shorter than the channel response time.~~
7. ~~Signals extracted from the geochemical composition of sediments are more likely to represent modern-day conditions during times of aggradation, whereas the signal will be temporally buffered due to mixing with older, remobilized sediment during times of channel incision.~~

~~We experimentally demonstrated that fluvial fill terraces can form due to changes in water discharge (climate), sediment supply (climate or tectonics), or base level (climate or tectonics). We demonstrated major differences in lag-times between the onset of perturbation and terrace cutting, and consequently in the resulting terrace elevation profiles and slopes. Therefore, information on the initial channel and environmental conditions that existed prior to the time of perturbation are not always well preserved in the terraces. We conclude that identifying the mechanism of fluvial fill terrace formation is~~

~~necessary to reconstruct past climatic or tectonic forcing accurately and that sediment storage and remobilization of sediment in alluvial channels can influence signals stored in the discharge ($Q_{s,out}$) or chemical composition of sediment.~~

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Author contribution. S.T., S.S. and A.W. designed and built the experimental setup. S.T. and S.S. performed the experiments. S.T. analyzed the data with the help of S.S., A.W. and A.B. All authors discussed the data, designed the manuscript and commented on it. S.T. designed the artwork.

Competing interests. The authors declare that they have no conflict of interest.

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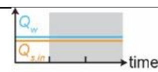
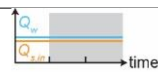
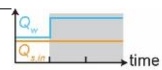
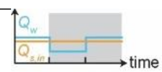
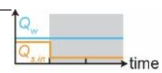
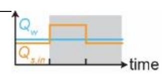
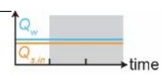
Table 1. Water and sediment inputs to the experiments.

Experiment	0 – 240 min (reference conditions)		240 – 480 min		480 min until end		Graphical description
	Q_w (ml s^{-1})	$Q_{s,in}$ (ml s^{-1})	Q_w (ml s^{-1})	$Q_{s,in}$ (ml s^{-1})	Q_w (ml s^{-1})	$Q_{s,in}$ (ml s^{-1})	
Ctrl_1	95	1.3	95	1.3	95	1.3	
Ctrl_2	95	1.3	95	1.3	95	1.3	
IQ_w	95	1.3	190	1.3	190	1.3	
DQ_w_*I*Q_w	95	1.3	47.5	1.3	95	1.3	
D$Q_{s,in}$	95	1.3	95	0.22	95	0.22	
I$Q_{s,in}$_*D*$Q_{s,in}$	95	1.3	95	2.6	95	1.3	
BLF	95	1.3	95	1.3	95	1.3	

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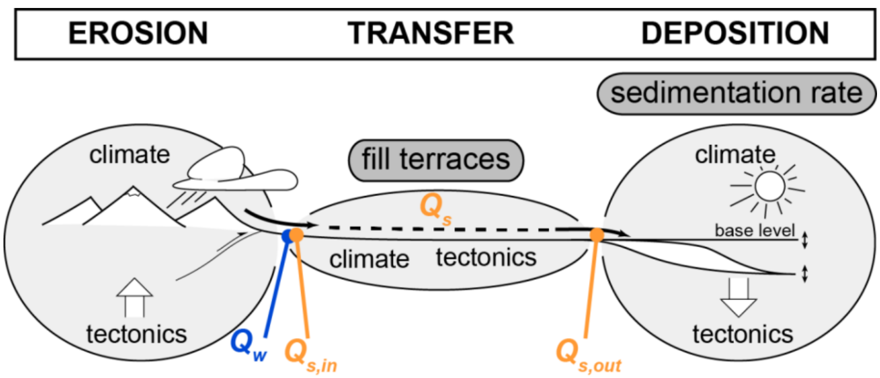
EROSION
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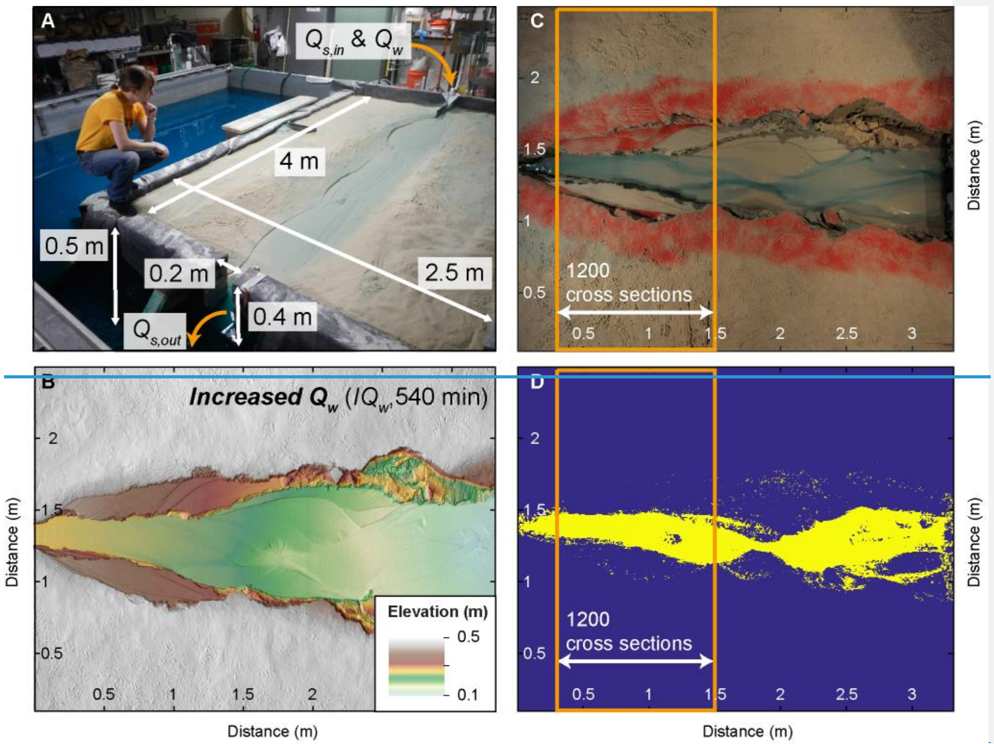

Figure 1. Schematic summary of a sediment-routing system and records of landscape evolution. Sediment-routing systems are typically subdivided into three zones: sediment production, sediment transfer, and sediment deposition (Allen, 2017; Castelltort and Van Den Driessche, 2003). As sediment production is thought to vary with environmental conditions, changes in those conditions might be preserved in sedimentary records in the transfer zone (e.g., fill terraces) or deposition zone (e.g., sedimentation rates). Complications arise, however, as alluvial rivers within the transfer zone continuously adjust their channel geometry to the incoming water discharge (Q_w) and sediment supply ($Q_{s,in}$) through sediment deposition or remobilization, and thus modify the sedimentary signal. The amount of upstream sediment supply combined with additional sediment remobilized within the transfer zone minus the amount of sediment deposited in the transfer zone determines how much sediment is discharged from the transfer zone to the deposition zone ($Q_{s,out}$). Figure modified from Castelltort and Van Den Driessche (2003).

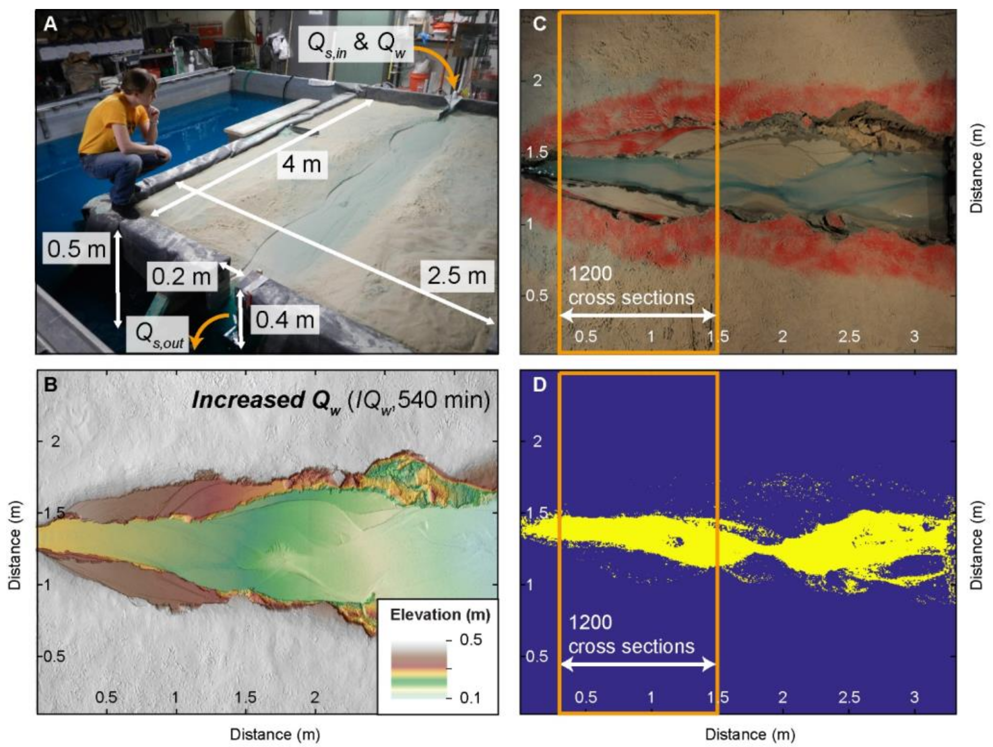

Figure 2.~~:~~ Experimental setup, data collection, and analysis. (A) Overview of experimental setup. Sediment supply ($Q_{s,in}$) and water discharge (Q_w) can be regulated separately. For all but the base~~-~~level fall (*BLF*) experiment, the base level was fixed. Water and sediment fell off of an edge at the outlet. For the *BLF* experiment (shown in the picture), the base level was controlled through the water level in the surrounding basin. (B) Digital elevation model (DEM) derived from laser scans showing the final topography of the increased water (IQ_w) experiment. (C) Overhead photograph of the IQ_w experiment taken directly before the scan shown in B. The surface was covered with a thin layer of red sand before the instant increase in Q_w~~discharge~~ was performed. The remnants of red sand on the terraces indicate no further reworking after the onset of increased discharge. (D) Overhead photographs were turned into binary (wet, dry) images from which the average channel width within the analyzed area (orange frame) can be calculated.

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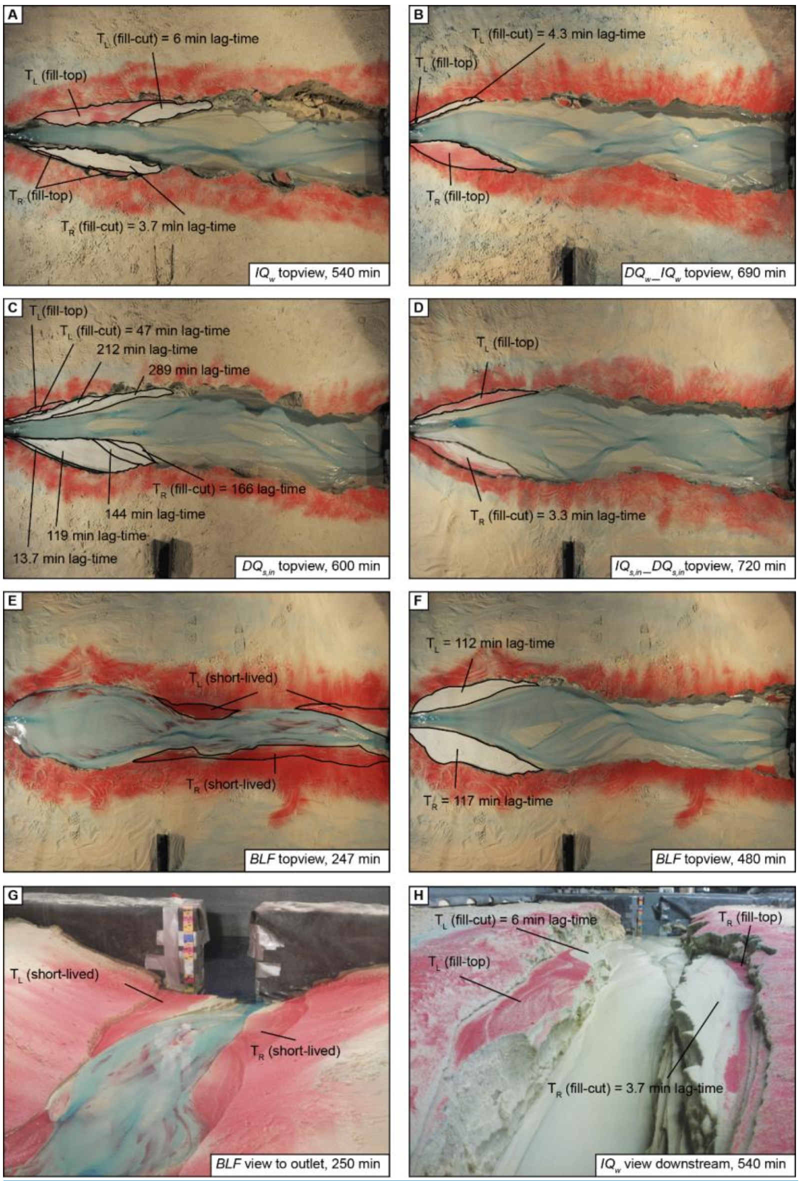
A
T_L (fill-cut) = 6 min lag-time
T_L (fill-top)
T_R (fill-top)
T_R (fill-cut) = 3.7 min lag-time
IQ_w topview, 540 min
B
T_L (fill-cut) = 4.3 min lag-time
T_L (fill-top)
T_R (fill-top)
DQ_w—IQ_w topview, 690 min
C
T_L (fill-top)
T_L (fill-cut) = 47 min lag-time
212 min lag-time
289 min lag-time
T_R (fill-cut) = 166 lag-time
144 min lag-time
119 min lag-time
13.7 min lag-time
DQ_s,in topview, 600 min
D
T_L (fill-top)
T_R (fill-cut) = 3.3 min lag-time
IQ_s,in—DQ_s,in topview, 720 min
E
T_L (short-lived)
T_R (short-lived)
BLF topview, 247 min
F
T_L = 112 min lag-time
T_R = 117 min lag-time
BLF topview, 480 min
G
T_L (short-lived)
T_R (short-lived)
BLF view to outlet, 250 min
H
T_L (fill-cut) = 6 min lag-time
T_R (fill-top)
T_L (fill-top)
T_R (fill-cut) = 3.7 min lag-time
IQ_w view downstream, 540 min

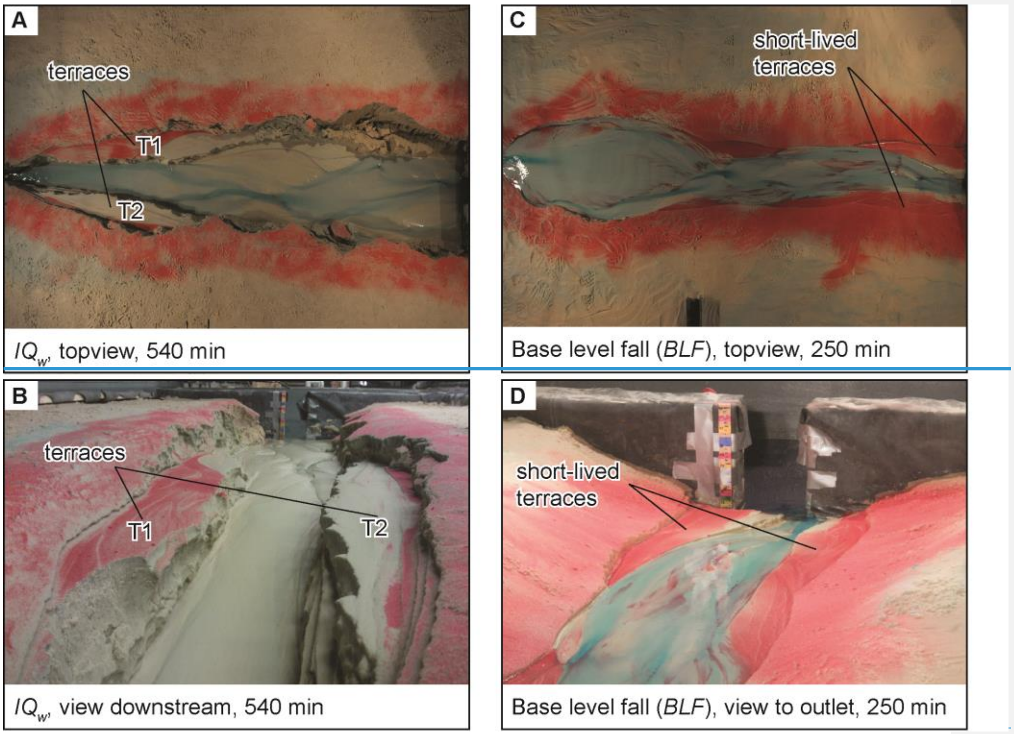

Figure 3. Fill terraces formed during experimental runs. (A-D) Top-view of ~~Paired~~ terraces that~~were~~ formed due to upstream perturbations.~~in the Increase Q_w (IQ_w) experiment and are shown from top (A) and looking in the downstream direction (B) at the end of the experiment (540 min = 300 min after spin-up time),~~ Remnants of red sand on the terrace surfaces indicate ~~that those~~ areas that have not been flooded after the change in boundary condition was performed (i.e. fill-top terrace). The other terrace surfaces were cut with the indicated lag-times (fill-cut terraces). (E-F)~~instant doubling in discharge,~~ During the base-level fall (*BLF*) experiment, terraces at the downstream end were abandoned instantly after the onset of base-level fall (250 min = 10 min after onset of *BLF*). ~~Terraces are shown from above (C) and looking in the downstream direction (D).~~ Those terraces were destroyed shortly~~shorty~~ after they were cut. A new set of terraces ~~was~~ formed in the upstream part ca. 112 and 117~~120~~ min after the onset of *BLF*. (G-H) Downstream view for the *BLF* (G) and IQ_w (H) experiments.

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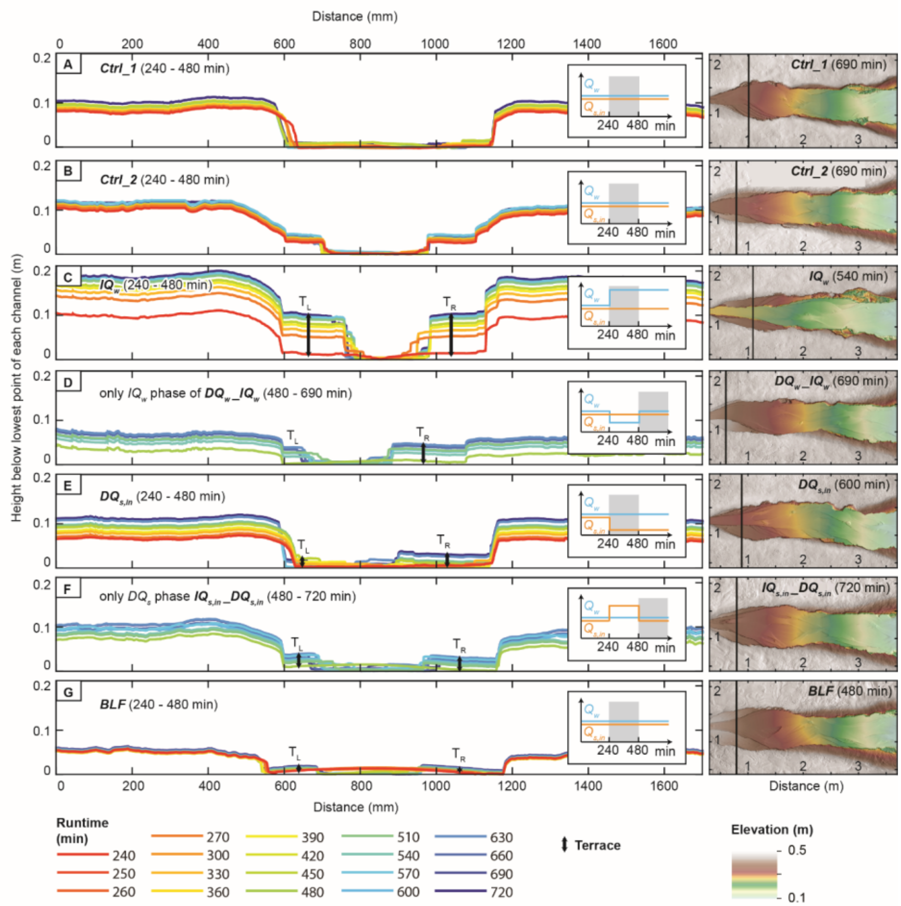

Distance (mm)
Height below lowest point of each channel (m)
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B Ctrl_2 (240 - 480 min)
Ctrl_2 (690 min)
C IQw (240 - 480 min)
IQw (540 min)
D only IQw phase of DQw_IQw (480 - 690 min)
DQw_IQw (690 min)
E DQs,in (240 - 480 min)
DQs,in (600 min)
F only DQs phase IQs,in_DQs,in (480 - 720 min)
IQs,in_DQs,in (720 min)
G BLF (240 - 480 min)
BLF (480 min)
TL
TR
Qw
Qs,in
240 480 min
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Runtime (min)
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Terrace
Elevation (m)
0.5
0.1

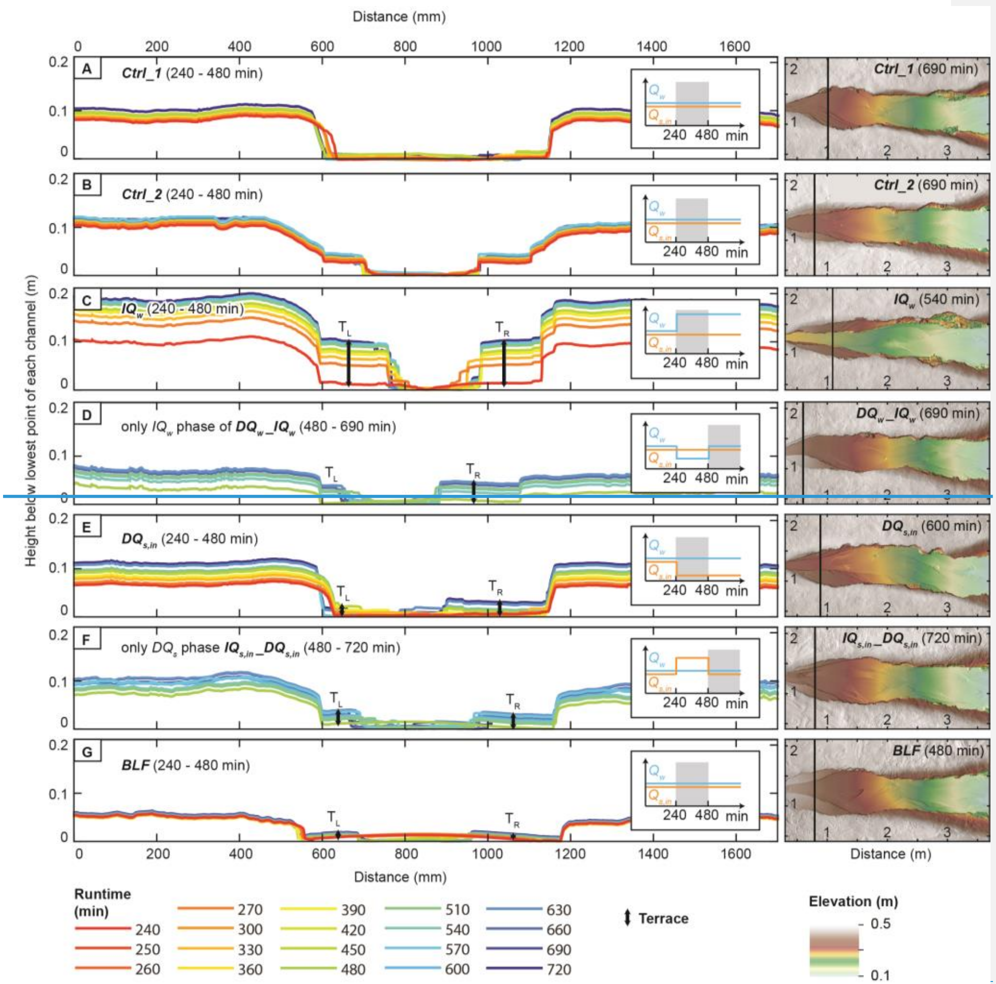

Figure 4. Evolution of cross ~~-~~sections in the upper part of the channel~~reach~~ (left panel). In each cross section, the lowest point is set equal to zero to track incision. The color scheme represents time after~~since~~ the ‘spin-up’ phase (i.e.~~last change in boundary conditions (equivalent to either~~ 240 min~~ or 480 min experiment time~~). For better comparison, we plot a maximum of 240 minutes for all experiments, despite longer recordings for some of the runs. Exact location of cross sections are indicated by the black lines in the DEMs displaying the last scan of each experiment (right panel). Cross-sections haven been chosen at the terrace midpoints and thus vary slightly between the experiments. The times given in parentheses are the absolute experiment runtimes.

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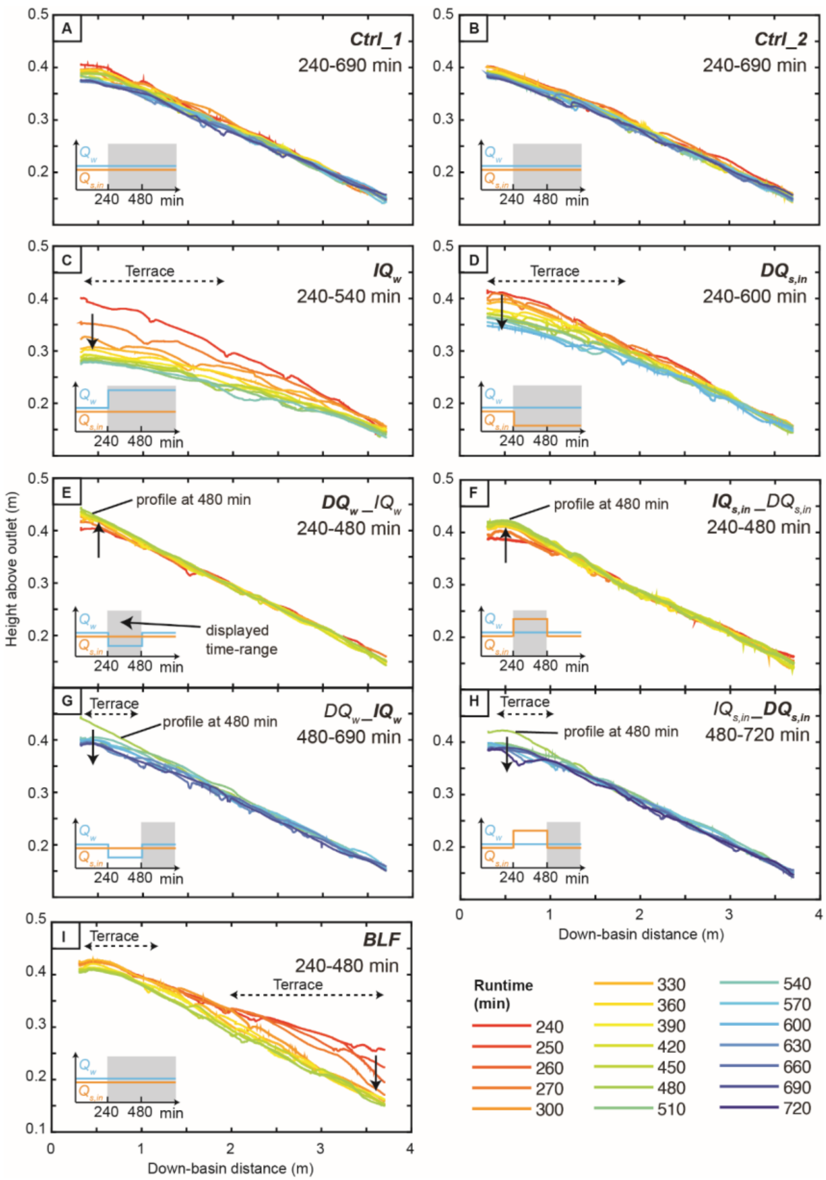

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C
Terrace
IQ_w
240-540 min
D
Terrace
DQ_s,in
240-600 min
E
profile at 480 min
DQ_w_IQ_w
240-480 min
displayed
time-range
F
profile at 480 min
IQ_s,in_DQ_s,in
240-480 min
G
Terrace
profile at 480 min
DQ_w_IQ_w
480-690 min
H
Terrace
profile at 480 min
IQ_s,in_DQ_s,in
480-720 min
I
Terrace
BLF
240-480 min
Terrace
Q_w
Q_s,in
240 480 min
Height above outlet (m)
Down-basin distance (m)
Runtime (min)
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270
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360
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690
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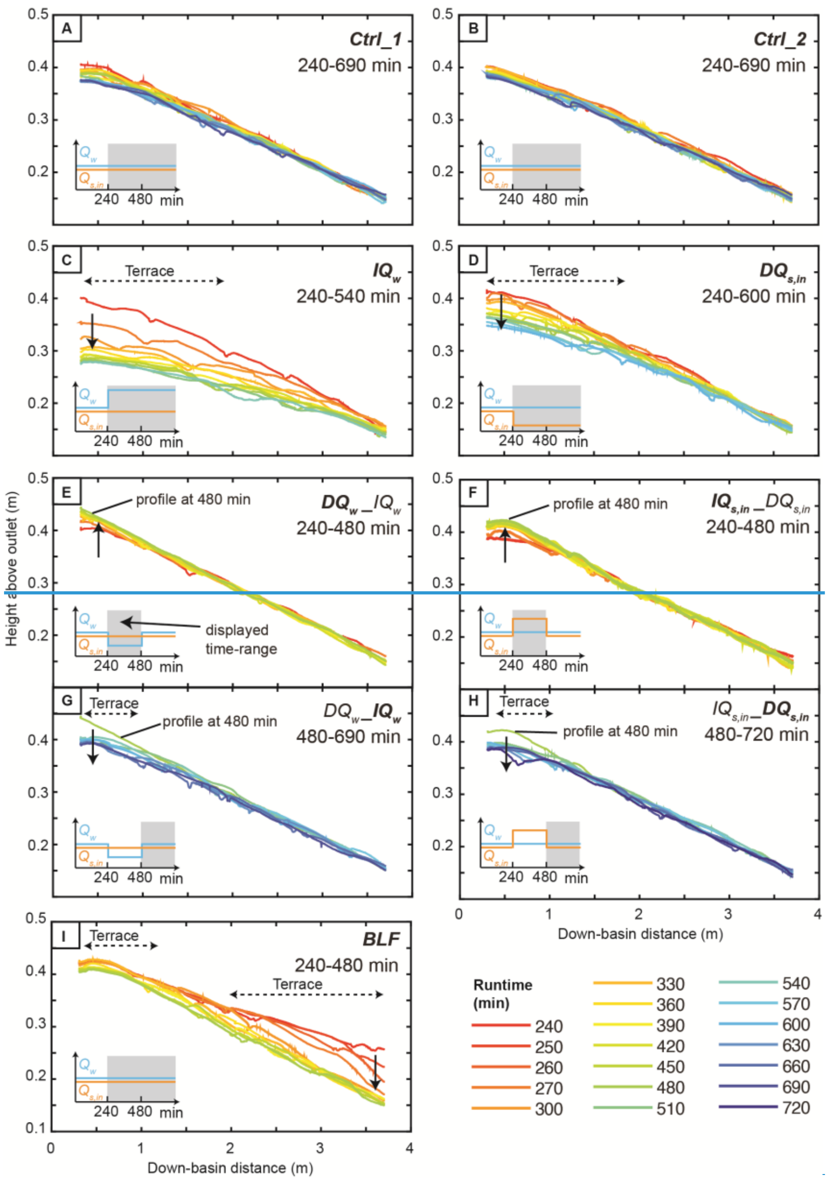
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Figure 5. Evolution of longitudinal river profiles from minute 240 (end of 'spin-up' phase) onwards. River profiles were extracted from the laser scans. Laser scans were recorded every 30 min, and an additional two scans at 10 and 20 minutes after the initiation of the base-level fall were conducted during the *BLF* experiment. Dashed arrows indicate down-basin distance along which terraces formed. Solid arrows indicate modes of aggradation or incision. Note that the $DQ_w\_IQ_w$ and $IQ_{s,in}\_DQ_{s,in}$ were split into two panels each, with one panel representing each phase.

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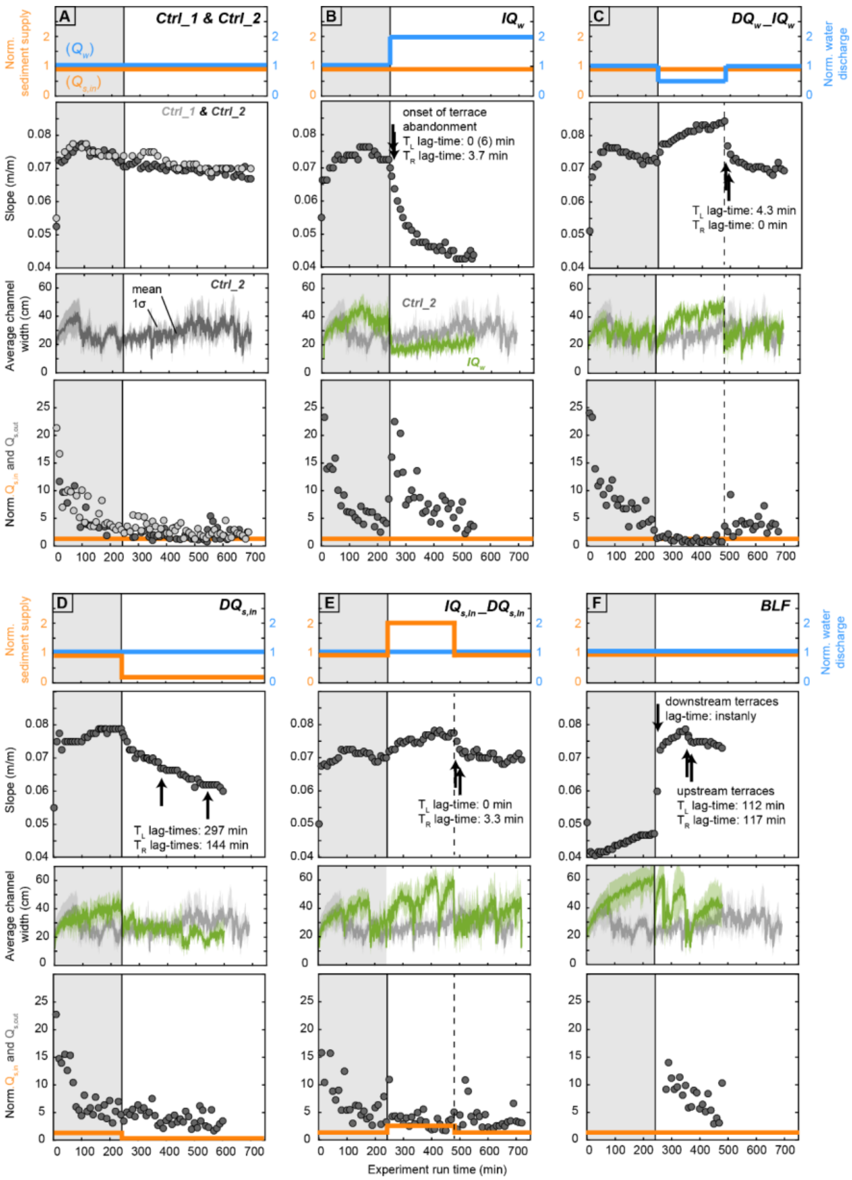


A
Ctrl_1 & Ctrl_2
B
IQw
C
DQw_IQw
D
DQs,in
E
IQs,in_DQs,in
F
BLF
Norm. sediment supply
Norm. water discharge
(Qw)
(Qs,in)
Slope (m/m)
Average channel width (cm)
Norm Qs,in and Qs,out
Ctrl_1 & Ctrl_2
mean
1σ
Ctrl_2
IQw
onset of terrace abandonment
TL lag-time: 0 (6) min
TR lag-time: 3.7 min
TL lag-time: 4.3 min
TR lag-time: 0 min
TL lag-times: 297 min
TR lag-times: 144 min
TL lag-time: 0 min
TR lag-time: 3.3 min
downstream terraces lag-time: instanly
upstream terraces
TL lag-time: 112 min
TR lag-time: 117 min
Experiment run time (min)

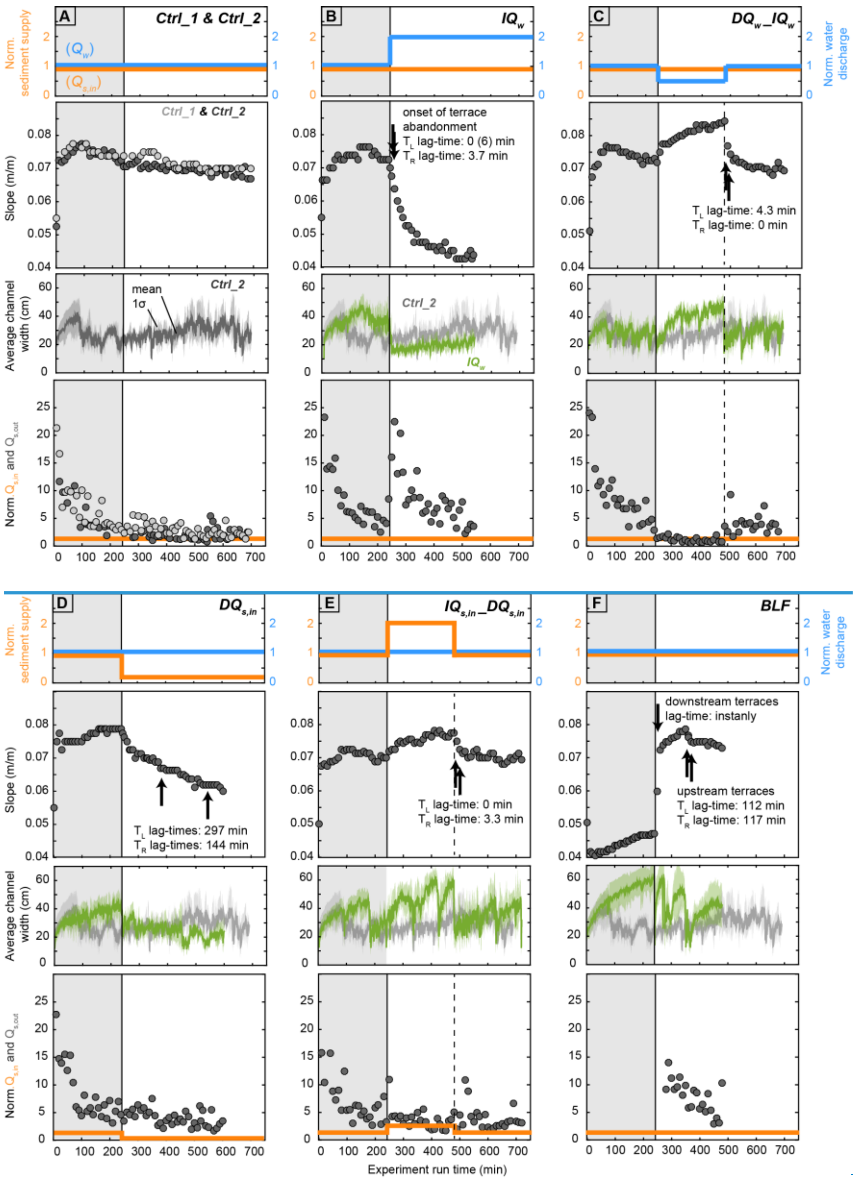
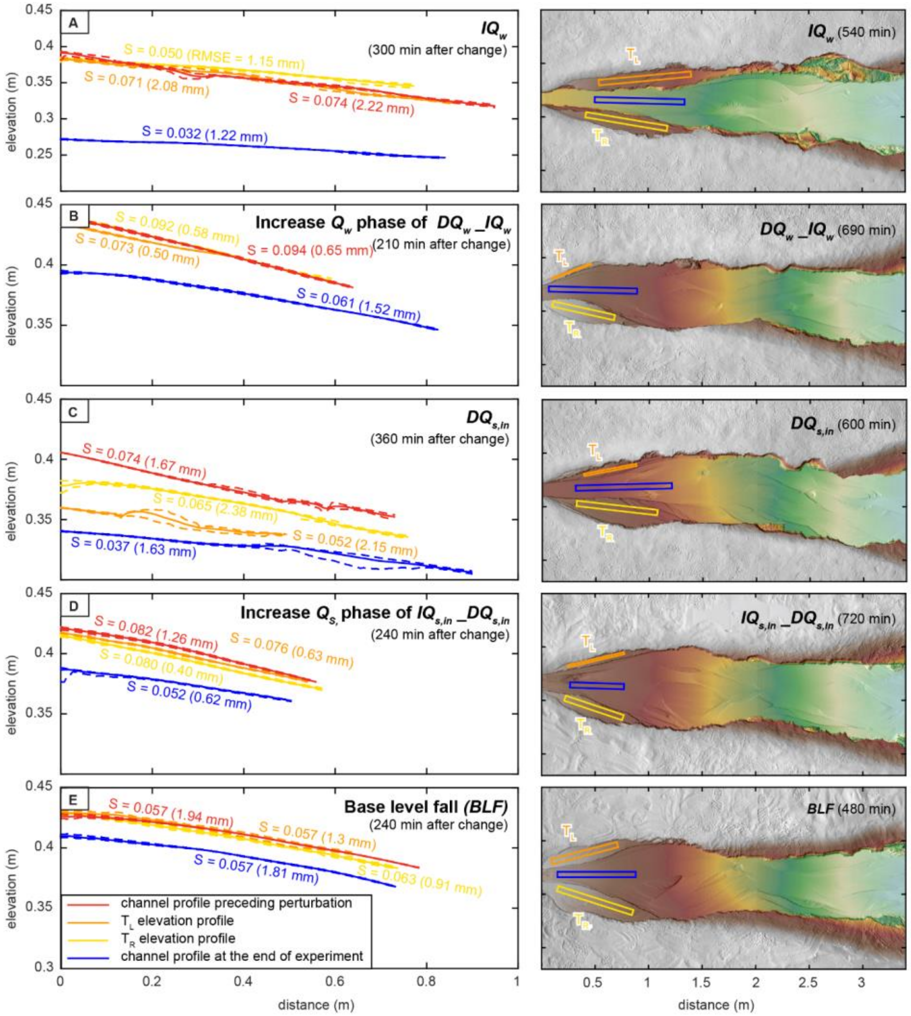

A
Ctrl_1 & Ctrl_2
B
IQ_w
C
$DQ_w\_IQ_w$
Norm. sediment supply
Norm. water discharge
(Q_w)
($Q_{s,in}$)
Slope (m/m)
onset of terrace abandonment
T_L lag-time: 0 (6) min
T_R lag-time: 3.7 min
T_L lag-time: 4.3 min
T_R lag-time: 0 min
Average channel width (cm)
mean
1σ
Ctrl_2
IQ_w
Norm $Q_{s,in}$ and $Q_{s,out}$
D
$DQ_{s,in}$
E
$IQ_{s,in}\_DQ_{s,in}$
F
BLF
T_L lag-times: 297 min
T_R lag-times: 144 min
T_L lag-time: 0 min
T_R lag-time: 3.3 min
downstream terraces
lag-time: instanly
upstream terraces
T_L lag-time: 112 min
T_R lag-time: 117 min
Experiment run time (min)

Figure 6. Input parameters and evolution of channel slope, ~~and~~ channel width and $Q_{s,out}$ during the experiments. Input sediment ($Q_{s,in}$; orange solid line) and water (Q_w; blue solid line) discharge were normalized to the reference input values ($Q_{s,ref}$ = 1.3 ml/s and $Q_{w,ref}$ = 95 ml/s). Slope (*S*, grey circles) was calculated based on the bed elevation difference between the inlet and the outlet divided by the length of the system. Channel elevation measurements for slope calculations were performed manually during the runs. Black arrows indicate times when terraces in the upstream part of the sandbox started to be cut. Channel width was calculated as the mean number (solid lines) of wet pixels in each of 1200 cross section within the box indicated in Fig. 2C~~1C~~, D. The colored shaded areas around the curves indicate the standard deviation of the 1200 measurements. The evolution of width without any external perturbation (*Ctrl_2*) is plotted for comparison with each other experiment in which external conditions were changed (B-F). Note that no measurements are available for the *Ctrl_1* experiment due to issues with the installation of the overhead camera. Sediment discharge at the outlet ($Q_{s,out}$; grey circles) during the experimental runs is compared to input sediment ($Q_{s,in}$; orange solid line); both were normalized to reference input values (Q_{s_ref} = 1.3 ml/s). Note that no $Q_{s,out}$ measurements are available for the first 280 min of the *BLF* experiment, as no sample collection was possible during the flooding of the surrounding basin during base-level regulations (Fig. 2A). The first 240 min of each experiment were adjustment to the reference settings (grey box) and were not included in the analyses. ~~Black arrows indicate times when terraces in the upstream part of the sandbox started to be cut.~~


A
IQ_w
(300 min after change)
S = 0.050 (RMSE = 1.15 mm)
S = 0.071 (2.08 mm)
S = 0.074 (2.22 mm)
S = 0.032 (1.22 mm)
B
Increase Q_w phase of $DQ_w\_IQ_w$
(210 min after change)
S = 0.092 (0.58 mm)
S = 0.073 (0.50 mm)
S = 0.094 (0.65 mm)
S = 0.061 (1.52 mm)
C
$DQ_{s,in}$
(360 min after change)
S = 0.074 (1.67 mm)
S = 0.065 (2.38 mm)
S = 0.052 (2.15 mm)
S = 0.037 (1.63 mm)
D
Increase Q_s phase of $IQ_{s,in}\_DQ_{s,in}$
(240 min after change)
S = 0.082 (1.26 mm)
S = 0.076 (0.63 mm)
S = 0.080 (0.40 mm)
S = 0.052 (0.62 mm)
E
Base level fall (BLF)
(240 min after change)
S = 0.057 (1.94 mm)
S = 0.057 (1.3 mm)
S = 0.057 (1.81 mm)
S = 0.063 (0.91 mm)
channel profile preceding perturbation
T_L elevation profile
T_R elevation profile
channel profile at the end of experiment
elevation (m)
distance (m)
IQ_w (540 min)
$DQ_w\_IQ_w$ (690 min)
$DQ_{s,in}$ (600 min)
$IQ_{s,in}\_DQ_{s,in}$ (720 min)
BLF (480 min)
T_L
T_R
distance (m)

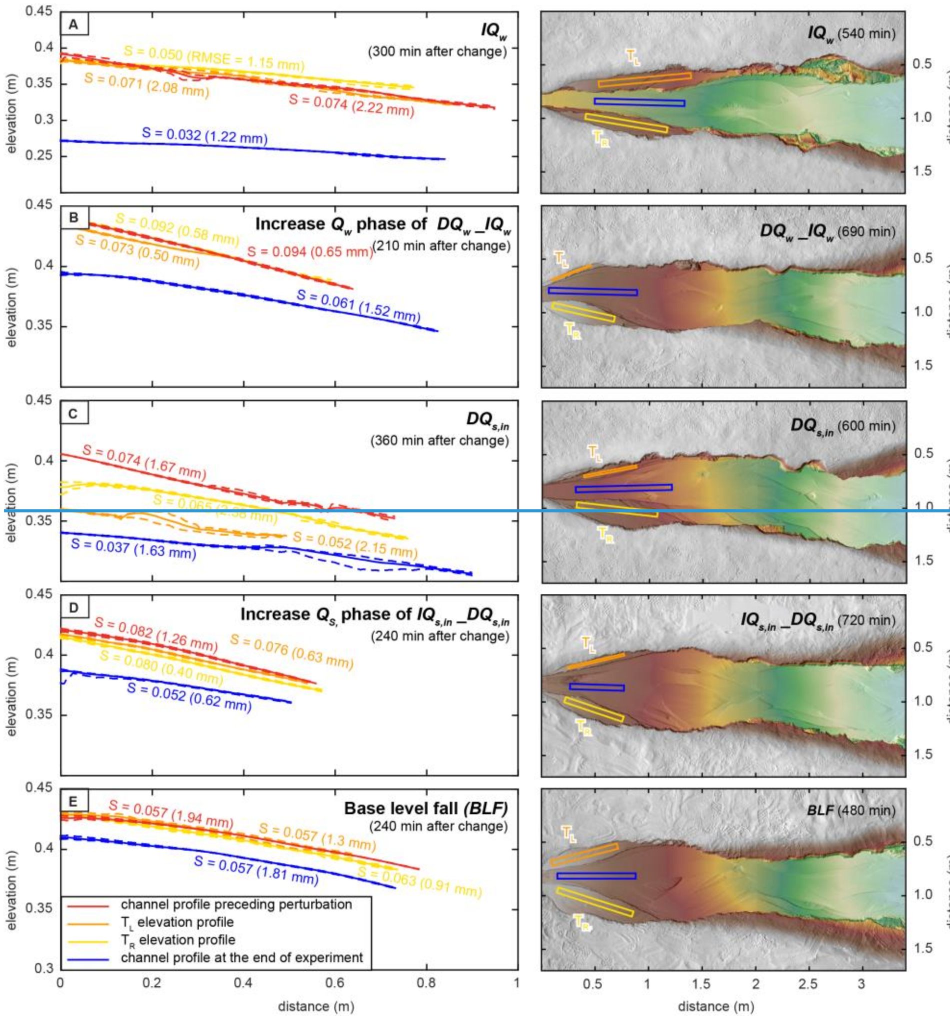


Figure 7. Elevation profile and slope comparison of terrace surfaces and active channels. Elevation profiles are given as mean (solid lines) and minimum and maximum values (dashed lines), extracted along a 5 cm wide swaths as indicated on the right panel. Swath width was reduced in two cases of too narrow terraces to 1 cm (*$DQ_w\_IQ_w$* T_L~~T_A~~ terrace) and 2 cm (*$DQ_{s,in}$* T_L~~T_A~~ terrace). T_L~~T_A~~ and T_R~~T_B~~ indicate terraces on one side each and refer to labels of lag-times given in figure 6~~Fig.5~~. Slopes were calculated based on a linear fit through the mean elevation profiles. Numbers in parentheses give the RMSE between the linear fit and the measured data. For the four experiments

in which upstream conditions ~~were~~ changed (A-D), the slopes of the terraces are steeper than those of the active channel at the end of the experiment. In contrast, in the *BLF* experiment, slopes of the terraces and the active channels are about the same. Note the different *y*-axis for the IQ_w run that was necessary to display the deep incision.~~for better visibility.~~ Colors for elevations~~of elevation~~ in the right panel are the same as those in figure 4~~Fig. 3~~.

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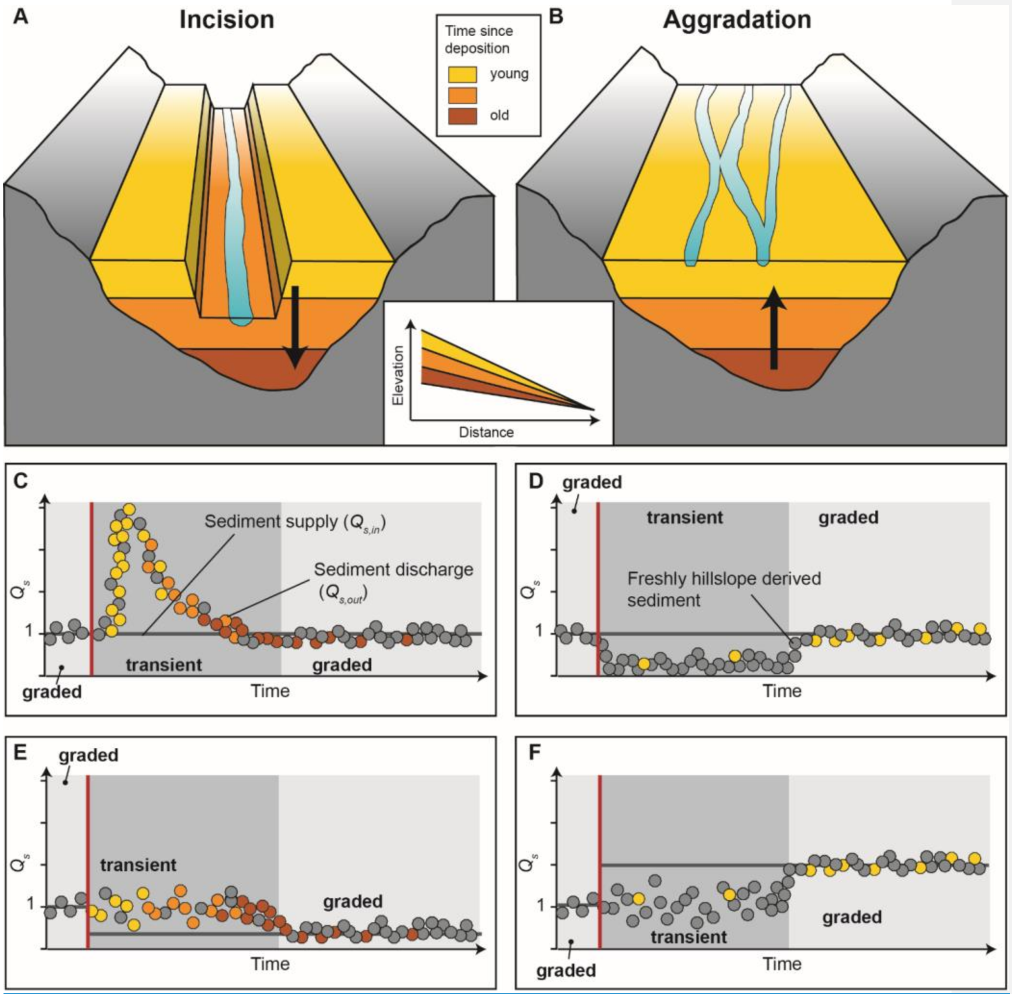
A
Incision
Time since deposition
young
old
B
Aggradation
Elevation
Distance
C
Sediment supply ($Q_{s,in}$)
Sediment discharge ($Q_{s,out}$)
Q_s
1
transient
graded
graded
Time
D
graded
transient
graded
Freshly hillslope derived sediment
Q_s
1
Time
E
graded
transient
Q_s
1
graded
Time
F
Q_s
1
transient
graded
graded
Time

Figure 8. Schematic model of the evolution of $Q_{s,out}$ including inferred ages of the sediment. (A-B) Channel geometry and approximate distribution of sediment deposition ages during phases of incision and aggradation. (C-F) Evolution of $Q_{s,out}$ (circles) compared to $Q_{s,in}$ (grey solid line) during the transient response phase after perturbation (dark grey) as well as after channel adjustment (light grey). The colours of the circles indicate the age (i.e. storage times before export) of the discharged sediment according to the panels A and B.

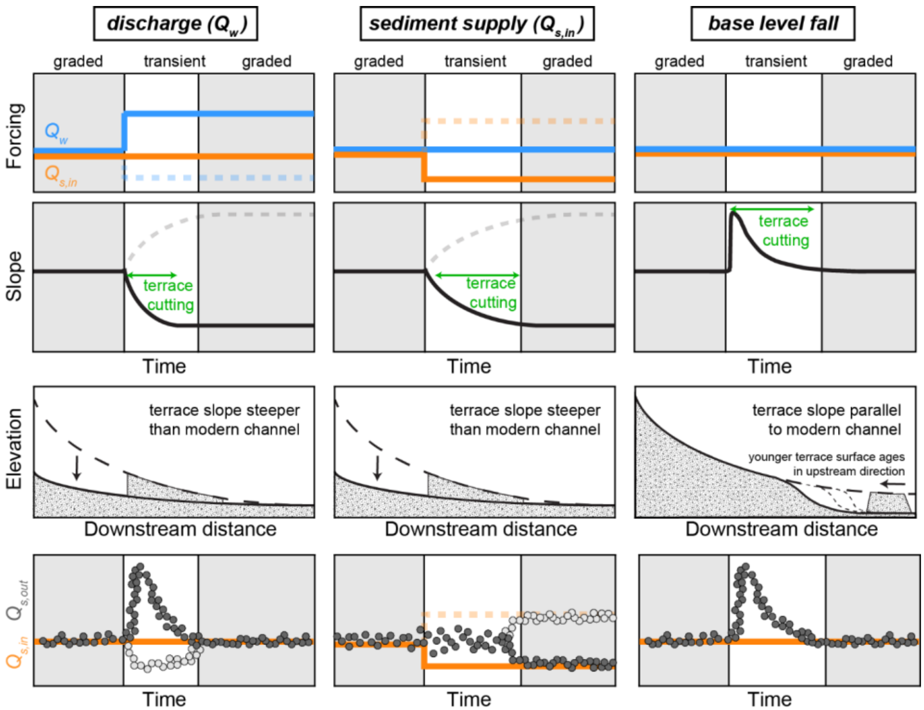

Figure 9. Combining two records of landscape evolution to overcome ambiguity. Terraces in the transfer zone and sediment discharge to the deposition zone ($Q_{s,out}$) as a proxy for sedimentation rate. The presence of terraces whose slopes differ from that of the main channel, combined with a simultaneous but transient peak in $Q_{s,out}$, points towards Q_w as the main driver of long-profile evolution. Terraces with steeper slopes compared to the modern channel combined with no immediate peak but an eventual reduction in $Q_{s,out}$, points towards $Q_{s,in}$ as the main driver. A temporary increase in $Q_{s,out}$ combined with terraces that parallel the modern channel profile and become younger in the upstream direction indicate past changes in base level.

~~. Schematic model of the evolution of signals at the outlet stored in either sediment volume or the chemical composition of the sediment.~~

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