Micro-scale Physical modelling as a tool to investigate Braided River Evolution

Michael M.M.[1]

[1]Bahir Dar University, Institute of Technology, School of Civil and Water Resources Engineering, Bahir Dar, P.O.Box-26, Ethiopia, email: micky_mehari@yahoo.com

Abstract

The findings from a series of micro-scale laboratory experiments designed to examine the effects of aggradation and degradation in the hydraulic geometry and sediment transport capacity of braided channels are discussed. The experiments were conducted in the University of Exeter Sediment Research Facility using a 5 m long and 2.7 m wide laboratory flume; hence situations are considered where channel width is unrestricted by the experimental setup. The experimental model is generic and is not scaled to a real world prototype. The evolution of river morphology was recorded using high resolution laser profiling to quantify channel changes (fill, incision and lateral erosion) and section geometry. During all the runs the evolution of the channel was recorded continuously using a Canon HG10 digital video camera and still imagery was collected at 2 minute interval using Canon EOS10D digital cameras. All cameras were mounted overhead. The quantitative observations from five experimental scenarios are discussed and evaluated. Rapid incision in the upstream portion of the channel resulted in the development of a single channel. However, the downstream reach remained braided as a result of continued delivery of sediments from the incising reach. The experimental work presented here has shown that despite the lack of dynamic similarity conditions and simplification of overall similarity criteria, fairly consistent results could be obtained which can be interpreted in a generic sense. The similarity between the laboratory channels from this experiment and those previously investigated by different researchers with or without a field prototype utilizing the Froude modelling principle suggest that the laboratory channels are at least qualitatively transferable to the field-scale.

Key words: micro-scale modelling, aggradation, degradation, river morphology

1. INTRODUCTION

Although a vast amount of research has been directed toward understanding braided rivers, we still do not completely understand the relationship between controlling independent variables (such as changes in sediment load) and braided river morphology. Relatively little has been done regarding the effects of sediment load on braided river pattern and evolution of longitudinal profile. Recent studies on bedload pulses in laboratory braided rivers (Ashmore, 1991, Hoey and Sutherland, 1991) provides further insight into some of the morphological changes attendant on changes in sediment load in rivers that are fundamentally in equilibrium with overall sediment load.

From a practical perspective, insight into the response of a braided system to changes in sediment load would be desirable for interpreting or predicting channel responses to climate change, upstream flow regulation by dams or any other hydraulic structure, land use change or river management.

Laboratory studies of this type range from true scale models, in which a valid model must exhibit geometric, kinematic and dynamic similarity between model and prototype as expressed by dimensionless ratios (Shen, 1979), to a more liberal approach where models are viewed as small prototype channels (Leopold and Wolman, 1957). Moreover, Hooke (1968) argues that geological models could meet similitude criteria based upon similarity of processes, provided that gross scaling relationships are met and the model reproduces morphologic characteristics of the prototype, which are produced by similar processes in both model and prototype.

The formal Froude modelling procedure in engineering and geomorphology in which the theorems of similarity mechanics observed in studies performed as a rule with water, offers certain advantages. A great number of river problems have been solved using this approach during recent decades. The major advantage of this type of model is that the results can, to some extent, be scaled to compare to the field prototype to validate the findings. However, this does restrict the applicability of the model, as it is limited to one field prototype and the results cannot easily be scaled to other areas. Nevertheless, an increasing number of problems have been encountered for which the solution obtained by this method is either insufficiently accurate, too time consuming, or too expensive. A typical example is modelling of braided rivers. Models usually need to be verified using prototype data. One of the problems with this is total braided river flood plain width. Although part of the problem may be attributed to difference in bank strength and difference in flow history (time of model rum)(Warburton, 1996), laboratory models are often limited by flume walls to describe the complex geometry of braided rivers or it should be accepted that lateral development of braids will be limited by the flume width.

This may not by itself pose a big problem for some braided rivers as flow rarely fills the full width of the braid plain and the active width, where much of the changes is occurring, is much smaller than the full braid plain width(Warburton, 1993). However, in extreme discharges which are very important because of their high transporting capacity, the braided rivers usually occupy the full bed width (Warburton, 1993). In most of these cases even if the planform geometry is to some extent statistically similar to the prototype, it is hardly possible to exactly reproduce the geometry of a given braided river reach (Young, 1996). Morphological modelling or assessing changes in downstream river morphology as a result of flow obstruction or any hydrological disturbance and long profile evolution data are very difficult to compare.

Undistorted Froude scaled models require similar slopes in models and prototype. Practically there is limitation in flume size, sediment management and cost because of the requirement of rough turbulent flow in Froude scaled models. Comparison of sediment transport processes is also another challenge in Froude scaled models. The considerable variability of bed load transport in braided rivers made it less meaningful to compare instantaneous bed load transport in model and prototype (Young, 1996). Direct measurement of bed load transport in the field is an almost impossible task due to the inherent temporal variability and errors associated with the sampling devices although it is easier in laboratory flumes. This often limits model scales to a linear scale of not more than 1:50, restricting modelling to typical reach of smaller braided rivers (Young, 1996). All this often limits the application of the Froude scaled models and in recent days the desire to at least reduce the cost and time of model testing methods, with reasonable accuracy comparable to Froude scale modelling, has been one of the basic motivations to develop methods of investigation in a smaller scale (micro-scale) without much attention being given to most of the dimensionless numbers.

With this in mind, the purpose of the present laboratory study was two-fold: to investigate the application of micro-scale physical model in braided river studies and to use this modeling approach and provide some insight into the effects of changes in sediment supply on braided river morphology.

2. EXPERIMENTAL DESIGN

This project uses a generic micro-scale hydraulic model to observe changes in channel morphology and sediment transport with changes in controlling variables. The model is not scaled to any specific prototype. Such experimental systems are based on the assumption that aspects of a natural system can be reproduced in a laboratory setting and that the processes responsible for producing features in nature will be similar to those operating in the laboratory environment.

The basic requirements for these experimental systems are that a number of gross scaling relationships be met, that the system reproduces some morphological characteristic of the landform in question, and that the processes that produced this characteristic in the experiment can logically be assumed to have the same affect on the original landform (Le Hooke and Rohrer, 1979).

Although the experimental River is based on the 'similarity of processes and performance' approach, to assist interpretation with respect to natural rivers it is useful to review briefly the initial and boundary conditions employed. In case of application of micro-scale modelling for a specific site, the approach to be followed as given by (Gaines and Maynord, 2001) involves adjustment of channel bed slope and water discharge to attain a chosen level of sediment transport intensity in the model. Once the model has reached this level of transport, it will be calibrated and a base test will be conducted. Deviation of all successive runs from the base test is assumed to occur in the prototype (Maynord, 2006).

The approach taken in this study entails a programme of flume experiments that does not follow a formal framework of similitude consideration. Although this experiment is based on the use of generic micro-scale physical model and not fully scaled to a prototype, an utmost effort was put to ensure realistic representations of sediment grain size and water discharge in terms of capability to transport sediments and formation of important morphodynamic phenomena. Based on this, simple scaling relations were used to test and select model input conditions. Published data on grain-size distribution of gravel-bed Rivers is considered to have an order of magnitude of the model grain-size (Ashmore, 1991, Young and Davies, 1991).Several trial experiments were also run to arrive at those experimental conditions and ensure their suitability.

The parameters for this project are calculated for sand with grain size distribution varying from 0.25mm-0.71mm (d_{50}~0.47mm). This size range is relatively narrow compared with the sediment size characteristics of natural gravel bed Rivers, which often have a range of grain sizes extending from boulders to silt. The decision was taken to exclude the silt and clay sized fractions, which would be required to represent finer (sand) sizes present in natural Rivers, because inter-particle cohesion in these fractions limits the capacity for these fractions to represent non-cohesive sediment. In other studies, this problem has been addressed by using light-weight sediment like crushed coal and other non-cohesive fine sediment to represent the finer non-cohesive fractions (Whipple et al., 1998) and (Sheets et al., 2002), however, it was considered that omitting fine sediments was an adequate solution for this research. The experimental sediment grain size and water discharge relationship is established mainly using the Shield's theory (Shields, 1936) for initiation of bed materials.

As we are not simulating a particular river, there is no strict similarity between values of sediment feed rate and water discharge. Even if there is debate on the scatter of the data and the interpolation of the trend line in the original Shields diagram (Peakall et al., 1996), former research has shown that it is possible to create a completely movable bed for τc (critical shields parameter) more than 0.056. The selected critical bed shear stress will vary depending on the hydraulic conditions in the flume. So, to create a moderately moving bed for d_{50} = 0.47mm, critical bed shear stress (τ_c) values are chosen (for water at 20^{o}c) from the Shield's diagram (Shields, 1936). If a given critical bed shear stress (τ_c) for a particular average grain size is less than the applied bed shear stress then movement of the bed material is expected.

Flow depth is then calculated using the uniform flow approach and equivalent roughness is estimated using a Keulegan type roughness estimator. Based on this and a preliminary channel width of 0.2m, the threshold discharge is calculated. The channel dimension is chosen to just accommodate the imposed discharge and keep the influence of the flume walls to a minimum later in the experiment when the river braids and expands in width. This approach gives order of magnitude estimates of model parameters to be used under a specified condition of grain Reynolds number. Experimental discharge values are selected based on this reference and slightly refined based on prevailing experimental conditions and characteristics of available water pump.

To achieve the study objectives, five different experimental run (RunS1-S5) were designed and formed at constant discharge and varying sediment input to induce successive scenarios of aggradation and degradation (Table 1). Experiment Run S1 was run until no appreciable difference was observed in sediment feed and collection rate in which case it could be assumed that the channel was in approximate equilibrium, though the channel itself was not monitored for changes in slope or elevation. Once a stable channel is formed the same timing was followed for the rest of the experiments with some modifications depending on the situation. In Table 1 below Q_w is the water discharge and Q_s is the sediment input used in the experiments.

Table 1: Sequence of the experimental series.

Run	Q_w *(10^{-5}) (m^3s^{-1})	Q_s (gs^{-1})	Duration (mins)	Remark
S1	5.8	0	1320	Stable ch.
S2	5.8	0.2	1620	Aggradation
S3	5.8	0	1320	Degradation
S4	5.8	0.2	1320	Aggradation
S5	5.8	0	1320	Degradation

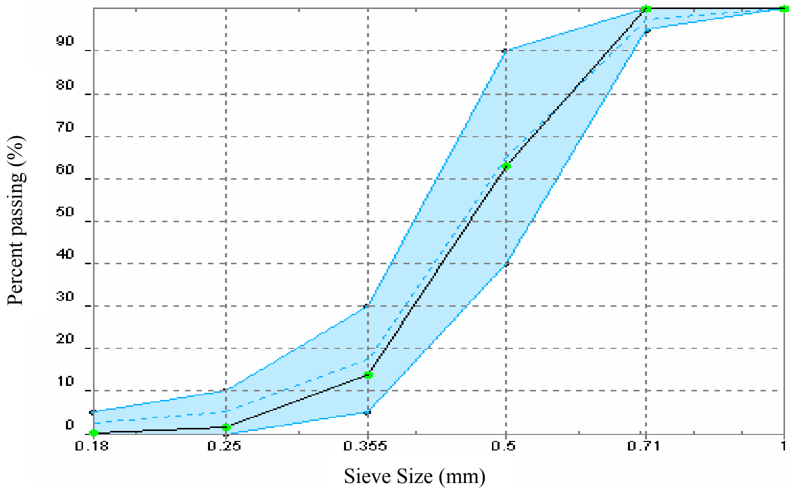


Figure 1: Gradation curve of the sand used in the experimental model runs.

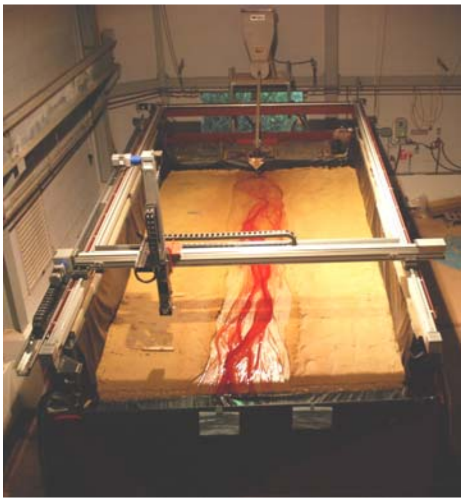


Figure 2: The experimental apparatus

The work for this project was carried out in the University of Exeter Experimental Landscapes Facility in a multi-purpose contemporary terrain modeller (sedimentation tank). The purpose-built sedimentation tank has dimensions of 5m length, 2.7m clear internal width and 1m depth. Rails running longitudinally on either side provide location and measurement datum to within +/-3mm under any loading condition. A false floor was constructed up to 0.75 m depth to reduce the depth of sand required to 0.25 m. The tank is pivoted at its lower edge and can be angled by a maximum of 15^0 from the horizontal. The tank is equipped with four 150mm wide cuts to simulate lateral flows if required and if not there are plates to cover the cuts. To facilitate base level change studies, the elevation of both the upstream run-on plane/inlet channel and the vertical downstream weir can be adjusted remotely using the hard-wired computer controller with adjustable vertical height for base level change studies (see Figure 2).

Bed topography can be measured by a laser micro-topographic scanner which can traverse the whole 5m length of the tank along a carriage mounted on a high accuracy rail system. The scanner measures output bed topography to a resolution of 1mm in cross stream (y) and downstream (x) directions and 0.1mm in elevation (z). The carriage and laser were both automatically controlled by software. Water is pumped to the upstream inlet from a smaller water tank connected with an online tap avoiding the labour work of filling the tank. The tank has a float controlled inlet valve so that a constant pumping head could be maintained. Sediment is supplied at a controlled rate to the upstream inlet using a variable speed gravity-fed sediment hopper. The sediment drops into the inlet groove made of steel where it mixes with the water, being fed at a specified discharge, before reaching the basin. At the point where the mixed water and sediment drop vertically onto the flume surface, smaller gravel is placed on the surface to dissipate the energy of the incoming water and thereby to prevent bed disturbance. Potassium permanganate solution was added regularly to the flow to facilitate flow visualization and identification in imagery.

At the downstream end of the basin a plastic gutter conveyed both the liquid and sediment flux to a constant head overflow tank. The tank rested on a high precision laboratory scale (accuracy 0.1g) which was used to measure the cumulative weight of exported sediment at 5 minute intervals. These data were used with observations of channel width at the downstream flume to derive sediment transport per unit river width. But both sediment and water is not recirculating in the system.

At the start of each experiment the sediment and water discharge rates were set to the predetermined values and held constant throughout the experiment, unless the experiment dictated otherwise. An order of magnitude of sediment feed rate to use in subsequent experiments was arrived at after a period of time during which the sediment collected in the overflow tank was monitored. The aim was to achieve and maintain a balance in feed and transport rate and to avoid long term aggradation at the entrance of the river channel. This experiment was carried out with a discharge of 3.5 l/min ($5.8*10^{-5}$ m^3s^{-1}) and gave a feeling of the amount of sediment transport to be expected in subsequent micro scale experiments. However, this does not imply that the channel is in absolute equilibrium as there was no long term monitoring of either longitudinal slope or channel configurations. There seems to be a small variation in sediment feed rate in the long term which proved to be very difficult to quantify. Accordingly, an effort was made to ensure that it is as constant as possible by checking the feed rate regularly and top-up continually to keep the level of sediment in the hopper constant.

3. RESULTS AND DISCUSSION

3.1. Variability and Changes in Sediment Output and Channel Storage

Sediment transport measurements taken at the outlet of the flume are displayed in Figure 3. When sediment output was first measured during the period when a stable braided channel was being established, a time lag of about 50min to 60min was observed before the output rate began to increase. Once this occurred, sediment output immediately increased and the rate was never constant. For the remainder of the run, sediment output in RunS1 fluctuated with greater amplitude until approximately 500min. but after that it became relatively consistent with minor fluctuation. Excluding the initial phase and from the exponential fit in Figure 3, there seems to be a gradually decreasing average transport rate during the development of the braided network.

At the beginning of the first aggradation cycle (from 0 to 275min in Figure 3, RunS2), the transport rate was generally higher than the transport rates observed in the later stages of the previous degradation run, with a mean of 0.205g/s, which was nearly equal to the sediment feed rate of 0.2 g/s and statistically indiscernible with the mean transport rate of RunS1 (samples within the same standard deviation). Channel storage was almost constant during this time. However, as channel aggradation progressed, sediment outputs reduced gradually and channel storage suddenly stepped-up and began to increase continuously (Figure 4). Observation of the changes in the long profile and cross section of the main channel (Figure 9C, D, L) shows that at the beginning of the first aggradation cycle (S2), the sediment input was probably stored in the actively incised upstream reach (section between 4m and 5m upstream of flume outlet). Around the middle of the flume, there was some erosion but this is intensified further downstream close to the flume outlet (1m upstream of flume outlet).The mean sediment output rate for RunS2 excluding the first 275min. of the flume run was 0.148 g/s, which is well below the feed rate.

When sediment feed ceased during the second degradation (S3), channel storage reduced abruptly. Similar sediment output continued despite cessation of sediment input at the upstream end of the flume. The mean output rate for RunS3 is 0.138 g/s, which is not very different from 0.148 g/s, as measured in RunS2 excluding its initial phase. With an increase in sediment feed again to 0.2 g/s, sediment output reduced for the first 340 min. and varied with lesser amplitude having a mean of 0.094 g/s. But, when sediment feed continued, the output started to increase slightly with a mild fluctuation around a mean of about 0.105 g/s and standard deviation of 0.042 g/s. Moreover, as channel aggradation continued, sediment storage increased at a steeper rate than during the previous aggradation cycle (RunS2). During the last degradation cycle (RunS5), similar sediment output continued for about 360 min (around 5300 min in Figure 3). The mean output rate in this period was 0.088 g/s. With continued degradation and cessation of sediment feed from upstream, a slight increase in sediment output is observed. The overall mean of this last degradation cycle is 0.129 g/s.

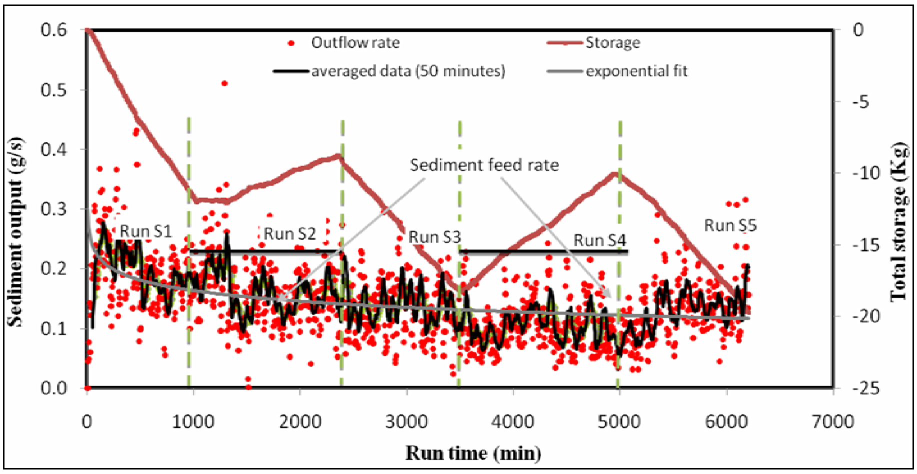


Figure 3: Measured sediment transport as a function

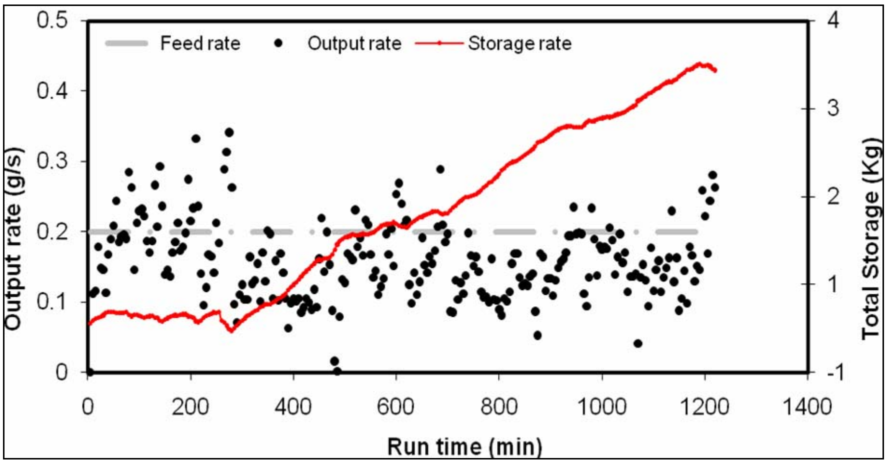


Figure 4: Sediment output-Storage relations for RunS2. of time throughout all runs.

Overall, a positive relationship between sediment output from flume and channel storage was observed through the experiment. Although the relations between sediment output and channel storage in RunS2 and RunS4 displayed similar styles, the magnitude and patterns of response differed early in the cycle. At the start of RunS2, when sediment feed was increased, storage was almost constant while output was steadily increasing. In RunS2, storage then increase steadily while output rates fluctuated around 0.15 g/s. When feed was stopped in RunS3, storage continuously decreased as output remained in a similar state as in the previous run. There is a large loss of storage volume of around 10,000g during RunS3. In RunS4, storage increased greatly as a slight reduction in output was observed up to around 3900min in Figure 3. After this time, sediment output remained fairly high until the end of this run even though storage continued to increase. The sediment storage in the second aggradation cycle (RunS4) increased at a much steeper rate than the first aggradation cycle (RunS2). This may be due to the fact that sediment storage may persist long after aggradational events and channel responses may be affected as a result of repeated cycles of aggradation and degradation (Madej et al., 2009).

At the beginning of RunS5, transport rates were the same as in RunS4, as storage declined and the channel bed started to degrade. This decline in storage continued up to the end of the run and transport rates also appeared to increase slightly towards the end of the experiment.

When sediment input ceased in the degradation runs, the channel actively incised and channel morphology changed to more of a single thread near the flume inlet and remained braided close to the outlet. However, there was not significant variation between the mean sediment output rates for the degrading runs (S1, S3 & S5) and mean sediment output rates of aggradation runs (S2 and S4). This suggests that transport rates respond weakly to sediment supply variations and hence sediment transport appears to be more closely related to transport capacity than sediment supply. This is in agreement with flume experiments of (Germanoski and Schumm, 1993) but counter to the experiments of (Madej et al., 2009), who observed high sediment output rates for aggradation runs when the sediment feed rate was high. In fact, the mean sediment output for the first aggradation cycle (S2) was higher than the next degradation cycle (S3), but this is again changed and mean sediment output rate went further down for the second aggradation cycle (S4). The sediment output rate was partly restored in the last degradation cycle (S5). There is also a tendency for transport rates to remain similar for sometime despite a change in sediment supply conditions.

Temporal variability in bed load transport rates has been observed in various laboratory studies of braided river dynamics (Ashmore, 1988, Ashmore, 1991, Hoey and Sutherland, 1991, Warburton and Davies, 1994, Young and Davies, 1991) and field studies e.g. (Goff and Ashmore, 1994, Griffiths, 1979). This temporal variability in sediment transport was attributed to the downstream migration of bedforms at different temporal scales, migration of large bars or groups of bars and localized incision (Gomez et al., 1989, Hoey, 1992) and has often led to difficulty in measuring both total and local bed load transport in braided rivers e.g. (Bridge, 1993).

Assessment of the time series shown in Figure 3 confirms that bed load transport rates are variable and shows a series of fluctuations. Although within-run transport rates sometimes vary from close to zero to just twice the mean rate, the extent of variability observed here is much lesser than in previous investigations e.g. (Ashmore, 1988, Hoey and Sutherland, 1991, Warburton and Davies, 1994, Bertoldi et al., 2009). This is probably due to the fact that most previous investigators used much higher water discharge leading to higher transport capacities and hence possibilities of large scale incision or bank erosion. At smaller discharges (similar to the one used in this experiment), the sediment transport capacity will be lower with less variability (Thomas, 2003).

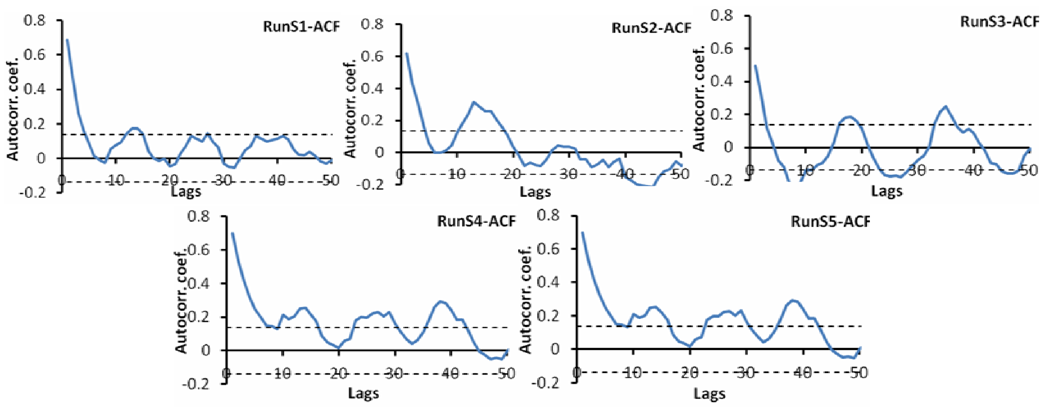


Figure 5: Auto-correlation functions of the original sediment transport time series. Dashed line represents 95% confidence limits.

Similarity of sediment output rates over a period of time despite a changing input condition (as was observed in almost all of these experiments) is an indication of ‘’persistence’’ (a tendency for a system to remain in the same state from one observation to the next). This was found to be a very common characteristic of the bed load transport processes in braided river models (Warburton and Davies, 1994). The best way to describe this statistically is using an autocorrelation function, because visual assessment from the time series is too subjective. The autocorrelation function provides a measure of the correlation between transport rates in the series at positions separated by a time interval along the series.

Auto-correlograms of the time series of bed load output rate shown in Figure 5 revels strong and positive autocorrelation with significant periodicity in the fluctuations of bed load transport rates. This is observed in almost all runs. The interpretation of this is that there is a greater persistence and some statistical dependency on the previous transport rates of the series as positive departure from the mean is followed by positive departure from the mean and vice versa. The frequency of periodicity varies between runs. However, a peak in the autocorrelation function at lags of about 1 hour is common to all runs. Periods ranging from 2 to 10 hours were observed in braided river models of (Ashmore, 1988).

3.2. Changes in Channel Pattern and Morphology

Figure 6 shows photographs taken towards the end of each run. Although quantitative evaluation of the intensity of braiding proved difficult due to the narrower nature of the channels, visual inspection of the channel in each experiment and investigation of the DEMs demonstrated that aggradation and sediment storage changed the channel morphology in each aggrading run by increasing the number of braid bars and widening the active channel itself. Similar morphological changes have been reported to some sections of laboratory braided channels as a result of bed load pulses moving into the reach (Ashmore, 1991). The magnitude of aggradation and degradation varied throughout the length of the flume, but it was greater near the flume entrance and there were minimal changes near the outlet due to the fact that bed level is fixed at the downstream end, hence damping out any major elevation changes.

Figure 7 shows typical cross-sectional profile at various distances down the flume during all the runs and illustrates the changes that occurred throughout the experimental run, and how this varied with time. Generally, the further downstream the cross section is located, the less channel change in terms of degradation and aggradation occurred. Figure 7 (A, B, K) represents cross sections taken at the channel bed before sediment feed began and a braided channel had formed. Major changes in the first run include considerable channel widening, degradation of the channel near flume inlet and deposition of eroded materials in the channel midway down the flume.

The channel were transformed to single-thread system in the upstream reach and remained braided in the downstream reach. This may be attributed to the fact that the amount of sediment supplied from the rapidly degrading and widened upstream reach resulted in an increase in sediment discharge downstream, where the channel was unable to transport all the sediment supplied to the reach.

Table 2: Channel morphological characteristics of the experimental series.

Run	*Bed load transport (g s)*		*Mean bed slope*	*Channel width (m)*		*CBRI (mm)*
	Mean	Std. dev.		Mean	Std. dev.	
S1	0.193	0.063	0.0227	0.59	0.07	20.1
S2	0.156	0.058	0.0274	0.74	0.07	11.7
S3	0.138	0.063	0.0252	0.49	0.1	19.4
S4	0.105	0.042	0.0259	0.86	0.15	14.2
S5	0.129	0.049	0.0203	0.54	0.13	26.1

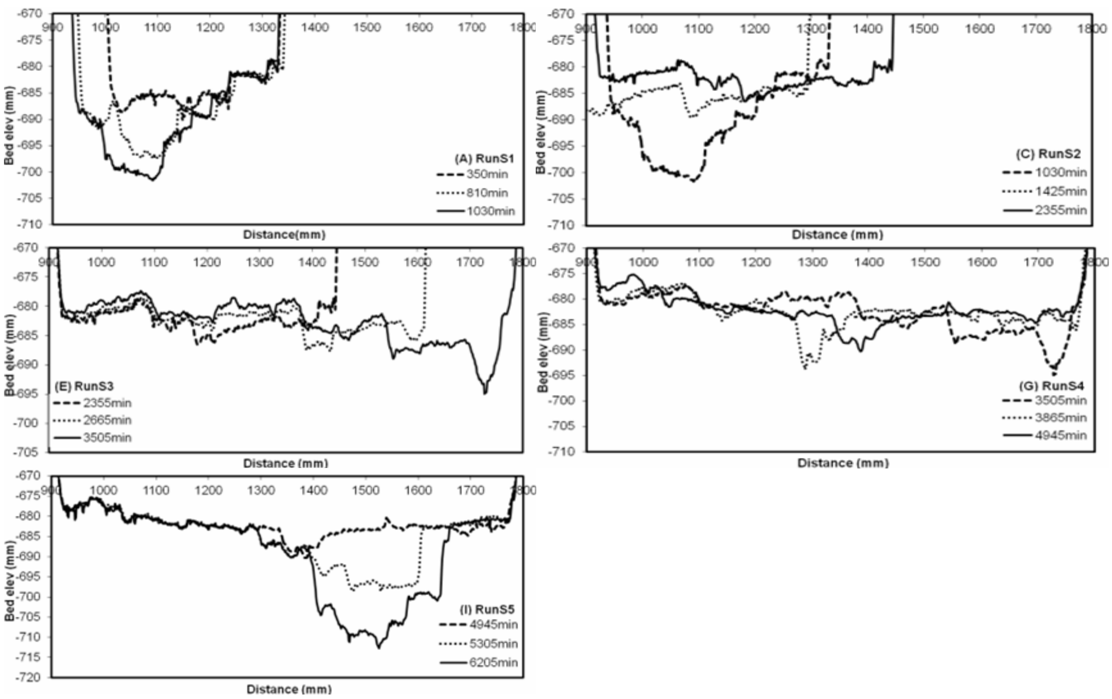
Consequently, despite the absence of sediment feed to the upstream end of the channel, the downstream part continued to aggrade. During the second run with sediment feed (RunS2), the thalweg and active channel bed filled at the upstream cross sections and the channel widened by more than 10cm (Figure 7c). Channel fill and widening midway down the flume at this time was insignificant (Figure 7d). Moreover, as can be seen from the plot of long profile evolution Figure 8, thalweg elevations in the downstream section remained well below bed elevations during the previous degradation run.

Aggradation was greatest near the inlet, associated with development of new braid bars which forced the flow into the channel banks and resulted in lateral erosion and an increase in overall channel width by more than 25% to 0.74 m. This value was more or less consistent throughout the length of the channel. The CBRI (cumulative bed relief index) dropped from nearly 20.1mm at the end of RunS1 to 11.7mm because of deposition in the active flow channels apparently reducing the relief between channel beds and bar tops.

This is in agreement with (Hoey and Sutherland, 1991, Madej et al., 2009) but against the laboratory braided channel experiments of (Germanoski and Schumm, 1993). Germanoski and Schumm (1993)attributed the different responses measured between their experiment and those of others e.g.

(Hoey and Sutherland, 1991) as a reflection of different rates and magnitudes of aggradation. It is obvious that the bed relief index increases when a braided channel is degraded as a result of a rise in the difference between the incising channel bed and stable bar tops but (Germanoski and Schumm, 1993) also illustrated the greater number of braid bars that formed during their aggradation experiment contributed to an increase in BRI. They found that the bed relief index could increase if aggradation occurred over the entire cross section and produced new bars. The BRI may remain constant or decrease if agradation is confined to the channels thalwegs. Germanoski and Schumm (1993) concluded that both aggradation and degradation can increase the BRI, although for different reasons.

Figure 6: Photographs of channel pattern taken towards the end of each Run.



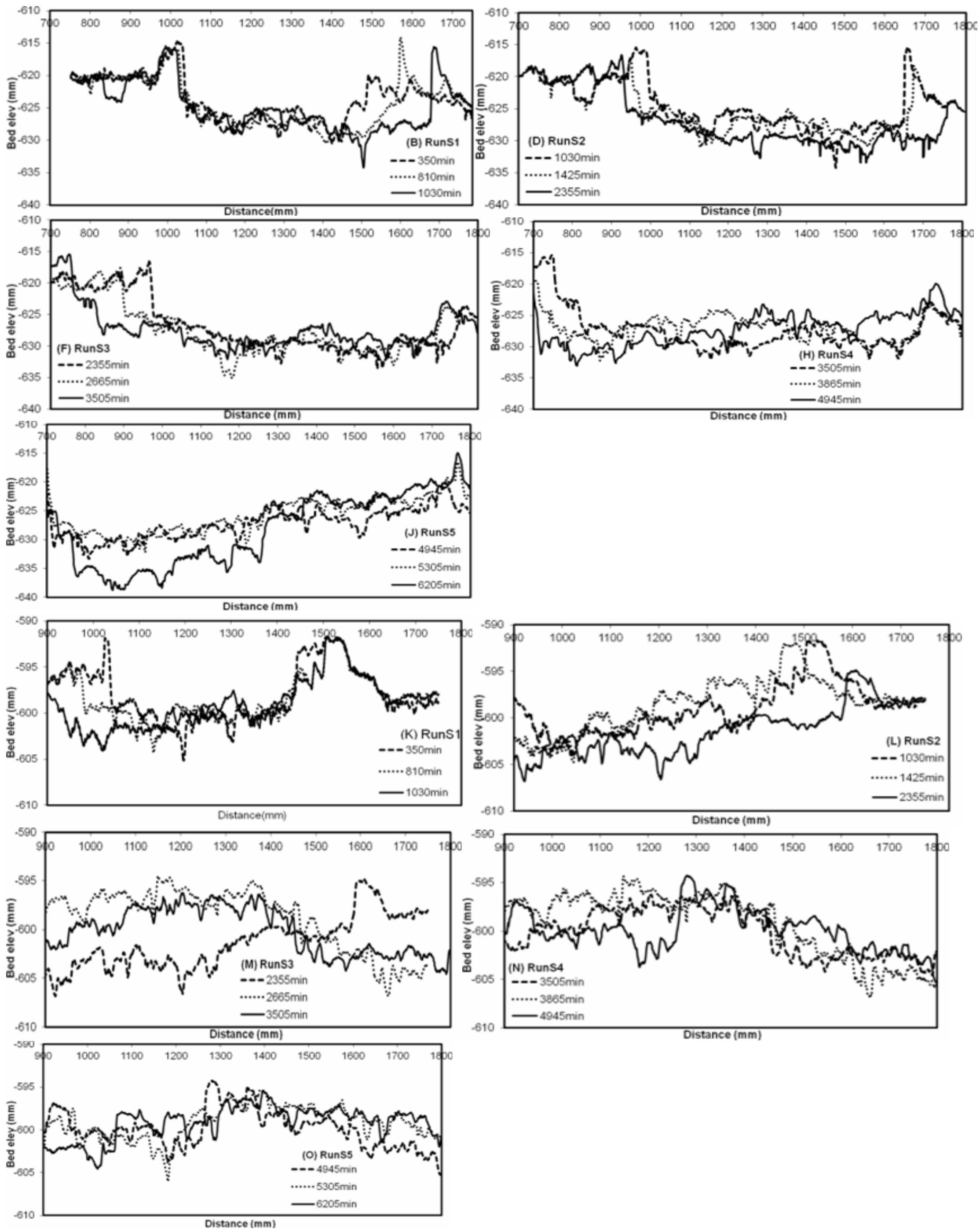


Figure 7: Typical cross-sectional changes during aggradational and degradational runs in the flume, measured at 1000mm (K,L,M,N,O), 2500mm (B,D,F,H,J) and 4500mm (A,C,E,G,I) upstream of outlet.

During RunS3 (Figure 7, E, F and M) the main channel continued to migrate and shifted to the right in the upstream parts of the channel. By the end of the run, it had migrated by more than 35cm, while at the same time the thalweg deepened by more than 1cm close to the upstream end of the flume. Degradation appeared to have been concentrated in the upper reaches of the channel and terraces appeared to have been formed. Towards the middle of the flume the channel migrated slightly to the left and the thalweg somewhat filled up. However, further downstream thalweg bed elevations remained below those in RunS2 for most of the reach. Mean morphologically active channel width, which is defined as the parts of the channel actively changing its elevation by erosion and deposition, declined and CBRI increased to almost pre-RunS2 conditions. As in the case of degradation in the first run, the channel here also experienced a transformation towards a single thread pattern in its upstream reach. More than half the length of the channel remained braided in the downstream part of the flume.

Initially, the channel response to the second period of sediment feed was similar to that during the first, with a small amount of widening and minor channel fill (Figure 7, G, H and N). But as sediment feed progressed, the channel completely filled in. A relatively large mid-channel bar formed near the middle of the flume (from 1000mm- 2500mm downstream the flume entrance), dividing the channel and hence the flow. The CBRI declined from 19.4mm at the beginning of RunS4 to 14.2mm at the end of RunS4. The number of channels appeared to increase throughout the flume, although many those channels were relatively narrow and shallow. The most significant channel change occurred in RunS5. Incised channels formed on the true left of the flume towards the upstream end and on the true right towards the middle of the flume. The depth of incision in this channel decreased in the downstream direction as can be seen from the long profile evolution. Upstream the channel was actively incising (Figure 7I) and by the end of the experiment it had incised by more than 2cm. Mid-flume degradation appeared to be concentrated on the true right side of the channel with a reduced magnitude of about 1cm. However, further downstream channel changes did not appear to be very significant owing to the effects of the fixed bed elevation at the flume outlet. The CBRI rose to its highest value in the experiment of 26.1mm. Channel bed slope was reduced significantly, but mean sediment transport rate increased, although it maintained smaller than in the corresponding degradation runs without sediment feed.

Differences in channel form exist between the five experimental runs, with the most noticeable changes occurring in the upstream one-third of the flume. Channel pattern in downstream areas remained similar for all the runs. In runs with sediment feed, cross section elevations clearly demonstrate the aggrading nature of the channels. In RunS2, the magnitude of aggradation varied from around 1.7cm (measured close to flume inlet at X=4500mm) to almost nothing (measured around mid-flume at X=2500mm). Small amounts of degradation (about 0.6cm) were measured close to flume outlet at X=1000mm. Overall, amounts of aggradation were almost three times as great as amounts of degradation. The trend was the same for the second aggradation cycle (RunS4) though with a lesser magnitude. Absence of sediment feed essentially reversed the trend. Most parts of the upstream flume reach were degraded; the highest degradation of around 3cm being for the last experiment (RunS5), measured at X=4500mm. Further downstream changes were reduced in magnitude. The notable difference between RunS5 and RunS3 was that in RunS5 degradation was highest and continuous throughout the length of the flume except at the very downstream end. Moreover, degradation was concentrated over less than 50% of the channel bed occupied by flow. During aggradation runs, a greater proportion of the bed accommodated changes in sediment volume.

3.3. Development of Longitudinal Profile and Channel Bed Slope

Figure 8 shows the development of the longitudinal profile throughout the course of the experiment. The channels observed here typically exhibited convex upward profiles marked by several distinct inflections. The slope of the channel declined as the sediment feed rate ceased for RunS1, RunS5 and to some extent RunS3. Longitudinal profiles of rivers are most commonly concave upwards. However, many dry land rivers are either less concave than those located in humid temperate areas (Parsons and Abrahams, 2009) or upward convex (Schumm, 1961). This may be attributed to the fact that the ratio of sediment to stream flow often increases downstream due to transmission losses (Parsons and Abrahams, 2009). This means that sediment is transported by progressively less flow, leading river bed aggradation and development of convex profile. Likewise, when more sediment flows out of the reach than is fed in, the channel is forced to degrade. This transient period of degradation at the upstream boundary assisted by the transmission losses and inability of flow to transport sediment, will force the profile of the channel to be upward convex. In the current experiments, incision was very rapid and intense for degradation runs in the upstream part of the channel. Thus, even when sediment feed stopped, the downstream reach continued to aggrade because of continued delivery of sediment from the rapidly degrading reach in the upstream portion of the channel (Germanoski and Harvey, 1993). This is probably what happened especially in RunS1, RunS3 and to a lesser extent in RunS5. Snow and Slingerland (1987) also modelled and analysed the effects of major variables like changes in water discharge, sediment discharge, sediment calibre on the long profile evolution of both sand and gravel beds by holding the other variables constant.

They observed that downstream variation in sediment discharge relative to flow discharge has strong potential to influence profile form and produce convex profiles.

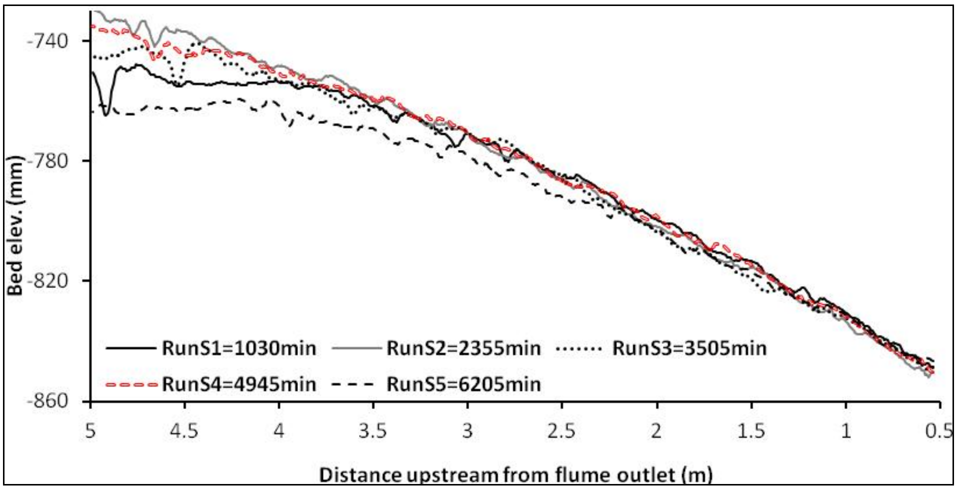


Figure 8: Longitudinal profile of channel thalweg measured at the end of each Run.

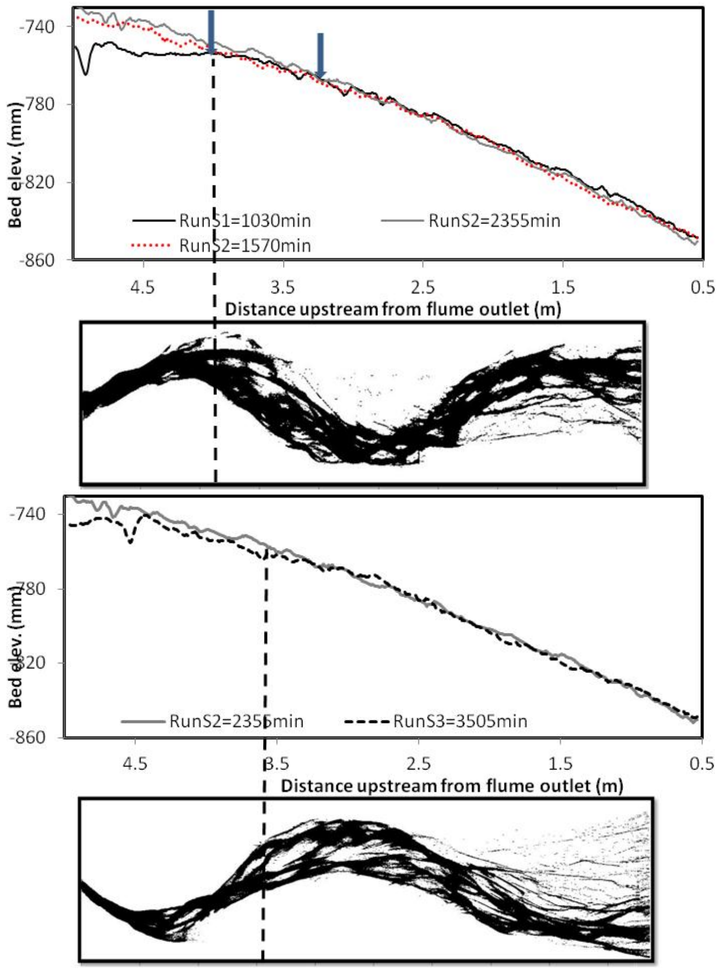
Fill and scour of the bed varied for short lengths of the flume during aggradation and degradation runs. Most changes were accommodated between the flume entrance and middle point. Changes in thalweg elevation from 2.5m to the outlet were very minor. During the first aggradation cycle (RunS2) bed elevation increased and most changes in elevations were accommodated at the entrance of the flume (from 3.5m up to the inlet) while the rest of the flume exhibited little change. During all the degradation cycles erosion largely contributed for most of the changes in bed elevation especially from entrance to mid-flume. At the end of the final degradation cycle (RunS5), the bed elevation remained lower than the original for most of the flume length indicating that the channel is unable to recover from the disturbances (change in sediment supply).

The continued delivery of sediments from the actively incising upstream reach during the degradation experiments reduced the capacity of the channels to transport sediments further downstream. This has kept bed elevations in the downstream portion of the channel to remain high although sediment feed has stopped (Figure 9). Examining the evolution of the longitudinal profile over the course of the experiment on Figure 13 two distinct areas can be noted (indicated by a dashed line in Figure 9 A, B and C). Area 1 represents the section where the long profile of the degrading channel lies completely below the following or previous aggrading channel (in most of the cases this is located close to entrance of the flume). Area 2 corresponds to the section where the long profile of the degrading channel lies above and/or below the following or previous aggradation phase.

The two regions are separated by an inflection point: the point that separates the rapidly degraded upstream area from the braided downstream reach as reported by laboratory experiments of (Germanoski and Harvey, 1993, Germanoski and Schumm, 1993). This point also seems to migrate through time as illustrated in Figure 9. Initially, the division between the rapidly degrading upstream zone and the aggrading braided zone occurred close to the flume inlet (4m from the flume outlet as indicated in Figure 9A). The channel was then rapidly filled by sediments during the first phase of sediment feed. Intensive filling of this rapidly degraded zone has occurred for the first 500min and after that only minor changes happened in this reach. The inflection point seems to move further downstream with cessation of sediment feed in RunS3, though it had only advanced not more than 50cm from its former location (Figure 9B).

However, as the gradient continued to decrease in the upstream degrading zone (see upstream section of RunS5 in Figure 8) bed load transport also reduced (Table 2), leading to migration of the inflection point further downstream closer to the outlet. Very similar results have been reported by (Germanoski and Harvey, 1993). In the experiments of (Germanoski and Harvey, 1993) there is a clear trend that indicates the downstream migration of the inflection point as the channel continued to degrade. However, in this case it appears that phases of aggradation and degradation are alternating in the downstream section, with the exception of RunS1-S2, where in this case the long profile of RunS2 lies entirely above that of RunS1 below the inflection point (see Figure 9D). This small difference between the results is due to the fact that (Germanoski and Harvey, 1993) reported continuously degrading channel where as in this case continuous degradation is interrupted by supplying sediment and initiating aggradation in between degradation experiments. Moreover, there are certainly differences in controlling variables (water discharge, sediment feed rate and grain-size) between the two sets of experiments.

In Figure 9D profiles are plotted relative to an arbitrary datum with the flume gradient removed to illustrate the inflection point more clearly. The profiles do not show the true channel gradient. It clearly shows the inflection point between the upstream degrading reaches and the downstream aggrading reaches in the degrading channel. It suggests that degradation is rapid and high in the upstream reach and this has resulted in aggradation of the downstream reach of a degrading channel that has even buried the aggrading channel.



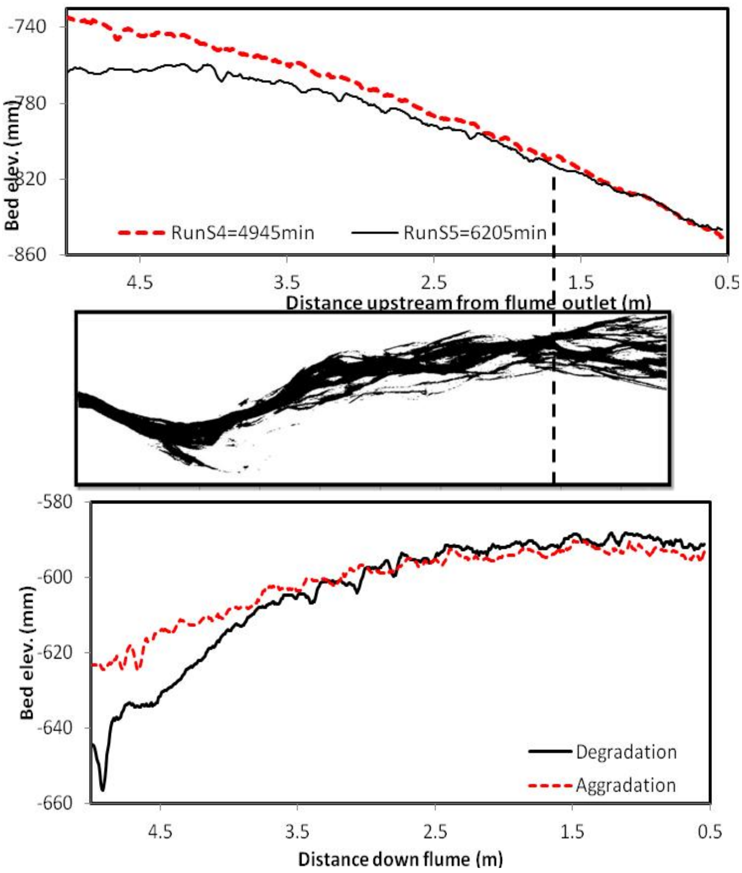


Figure 9: Relationship between channel pattern, longitudinal slope and migration of inflection point.

Several researchers have studied relationship between channel pattern and longitudinal slope both quantitatively and qualitatively (Lane, 1957, Leopold and Wolman, 1957, Ackers and Charlton, 1970, Osterkamp, 1978). Physical model studies have also been used in the past to demonstrate the existence of pattern threshold related particularly to slope (Schumm and Khan, 1971). In general, for a certain discharge, braided streams are characterized by greater slopes than meandering streams. Different channel patterns can also occur at different sections of the same river depending on longitudinal slope and sediment characteristics. Lee and Henson (1977) observed an abrupt change in channel pattern of the Red river from non-braided to braided while in that same section the longitudinal slope doubled from 0.034% to 0.068%. Similarly, (Leopold and Wolman, 1957) also observed a change in channel pattern in Horse Creek from single thread to braided where the longitudinal slope changed from 0.0022 to 0.0073. In experimental series 1, channel bed slope is calculated from regression of thalweg elevations of cross sections taken every 8mm in the stream-wise direction.

Furthermore thalweg elevations of channel cross sections acquired at 25cm longitudinal interval are also used to calculate temporal variation of longitudinal slope within an experiment. Channel slopes calculated by regression of thalweg elevations range in value between 0.020 and 0.027; the maximum being attained in the first aggradation cycle Runs2.

Since bed elevation is fixed at the downstream end of the flume, significant increase in bed elevation occurred in the upstream end of the flume. For all degradation runs a break in slope occurred together with a change in channel pattern. This was greatest in RunS1 and RunS5. In RunS1 there was a significant change in slope (from 0.002 to 0.028) when the channel pattern changed from single thread to braided although it was only in the upstream 1m of the channel that this occurred (Figure 9A). In RunS3, however, the change in slope was reduced (from 0.0065 to 0.028) owing to interruption of degradation by sediment feed from upstream. Generally, absence of sediment feed and channel degradation throughout the experimental series was associated with the channel moving towards a single thread, meandering system (Figure 9) with milder slope close to the inlet. During periods of sediment feed the slope of the channel was generally steeper and the pattern was braided.

The most abrupt change in gradient occurred in the area near the flume inlet or from 3.5m to 5m upstream of the flume outlet. When the upstream reach underwent a transition from braided to single-thread (RunS2 to RunS3), it experienced a reduction in longitudinal slope from 0.017 to 0.0065. When sediment feed was restarted in RunS4, the channel pattern in the same reach changed to braided and its longitudinal slope increased by more than double to 0.014 before finally attaining the minimum slope of 0.0046 in the course of the experimental run. In real terms, the increase in slope translates into aggradation and development of multiple threads at least for this experiment. In response to the cessation of sediment feed from upstream, an increase in transport capacity is observed which resulted in vertical incision with different magnitude and channel migration through bank erosion in all degradation runs; both of which ultimately reduced the channel slope.

4. CONCLUSION

The morphological changes reported in this experimental series were mostly similar to those seen in previous laboratory studies of braided channels e.g.(Ashmore, 1991, Hoey and Sutherland, 1991, Germanoski and Harvey, 1993, Germanoski and Schumm, 1993, Madej et al., 2009). In most of these studies the braiding intensity, number of braid bars, pattern complexity and morphologically active braid plain width all increase when a channel experiences aggradation. The same relationship is recorded in this experiment. However, unlike other similar laboratory experiments the bar size observed here do not show significant differences between aggrading and degrading channels. Degradation essentially reversed the morphological changes associated with aggradation. Initiation of degradation by stopping sediment feed resulted in vertical incision close to the flume inlet that sometimes extended to middle of the flume. Flow became concentrated into a single channel flanked by erosional terraces that ultimately left a greater proportion of the active channel exposed.

This is the most prominent feature of the evolution of sand bed laboratory channels. However, the downstream portion of the channel did not degrade due to the resultant influx of sediment arriving from the actively incised reach upstream. This is also in agreement with the laboratory experiments of (Germanoski and Harvey, 1993) and (Germanoski and Schumm, 1993). The experiments of (Germanoski and Harvey, 1993) have also been compared with Ash Creek, a degrading ephemeral-flow braided channel, on the east flank of the Mazatzal Mountains in central Arizona. The stream was degraded due to a reduction in sediment delivery as a result of re-vegetation of the drainage basin. The channel responses to this reduction in sediment yield included incision progressing from upstream to downstream and continued aggradation downstream, maintaining the braided pattern. These trends are very similar to the laboratory experiments reported here.

The experimental work presented in this paper has shown that despite the lack of dynamic similarity conditions and simplification of overall similarity criteria, fairly consistent results could be obtained which can be interpreted in a generic sense. The similarity between the laboratory channels from this experiment and those previously investigated by different researchers with or without a field prototype utilizing the Froude modeling principle suggest that the laboratory channels are at least qualitatively transferable to the field.

5. ACKNOWLEDGMENTS

This work formed part of a PhD project at the Department of Geography, University of Exeter, United Kingdom and Michael would like to thank the department for providing financial support for this research through an internal scholarship.

6. REFERENCES

1. ACKERS, P. & CHARLTON, F., 1970, *Meander geometry arising from varying flows,* Journal of Hydrology, 11**,** 230-252.
2. ASHMORE, P., 1991, *Channel morphology and bed load pulses in braided, gravel-bed streams*, Geografiska Annaler. Series A. Physical Geography**,** 37-52.

3. ASHMORE, P. E., 1988, *Bed load transport in braided gravel bed stream models, Earth Surface Processes and Landforms,* 13**,** 677-695.
4. BERTOLDI, W., ASHMORE, P. & TUBINO, M., 2009, *A method for estimating the mean bed load flux in braided rivers*, Geomorphology, 103**,** 330-340.
5. BRIDGE, J. S., 1993, *The interaction between channel geometry, water flow, sediment transport and deposition in braided rivers*, Geological Society, London, Special Publications, 75**,** 13-71.
6. GAINES, R. A. & MAYNORD, S. T., 2001, *Microscale loose-bed hydraulic models,* Journal of Hydraulic Engineering, 127**,** 335.
7. GERMANOSKI, D. & HARVEY, M. D. 1993, *Asynchronous terrace development in degrading braided channels*, DTIC Document.
8. GERMANOSKI, D. & SCHUMM, S., 1993, *Changes in braided river morphology resulting from aggradation and degradation,* The Journal of geology**,** 451-466.
9. GOFF, J. R. & ASHMORE, P., 1994, *Gravel transport and morphological change in braided Sunwapta River, Alberta, Canada*. Earth Surface Processes and Landforms, 19**,** 195-212.
10. GOMEZ, B., NAFF, R. L. & HUBBELL, D. W., 1989, *Temporal variations in bedload transport rates associated with the migration of bedforms,* Earth Surface Processes and Landforms, 14**,** 135-156.
11. GRIFFITHS, G., 1979, *Recent sedimentation history of the Waimakariri River*, New Zealand. *Journal of Hydrology(New Zealand),* 18.
12. HOEY, T., 1992, *Temporal variations in bedload transport rates and sediment storage in gravel-bed rivers*, Progress in physical geography, 16**,** 319.
13. HOEY, T. B. & SUTHERLAND, A. J., 1991, *Channel morphology and bedload pulses in braided rivers: a laboratory study*, Earth Surface Processes and Landforms, 16**,** 447-462.
14. LANE, E. W., 1957, *A study of the shape of channels formed by natural streams flowing in erodible material*, US Army Engineer Division, Missouri River.
15. LE HOOKE, R. B. & ROHRER, W. L., 1979, Geometry of alluvial fans: Effect of discharge and sediment size, Earth Surface Processes, 4**,** 147-166.
16. LEE, L. J. & HENSON, B. L., 1977, *The interrelationships of the longitudinal profiles and channel pattern for the Red River,* Journal of Hydrology, 35**,** 191-201.
17. LEOPOLD, L. B. & WOLMAN, M. G., 1957, *River channel patterns: braided, meandering, and straight*, US Government Printing Office Washington (DC).
18. MADEJ, M. A., SUTHERLAND, D. G., LISLE, T. E. & PRYOR, B., 2009, *Channel responses to varying sediment input: A flume experiment modeled after Redwood Creek*, California. *Geomorphology,* 103**,** 507-519.
19. MAYNORD, S. T., 2006, *Evaluation of the micromodel: An extremely small-scale movable bed model*, Journal of hydraulic engineering, 132**,** 343.
20. OSTERKAMP, W., 1978, *Gradient, discharge, and particle-size relations of alluvial channels in Kansas, with observations on braiding*, American Journal of Science, 278.
21. PARSONS, A. J. & ABRAHAMS, A. D., 2009, *Geomorphology of desert environments*. Geomorphology of Desert Environments**,** 3-7.
22. PEAKALL, J., ASHWORTH, P. & BEST, J., 1996, *Physical modelling in fluvial geomorphology: principles, applications and unresolved issues*, The Scientific Nature of Geomorphology**,** 221-253.
23. SCHUMM, S. & KHAN, H., 1971, *Experimental study of channel patterns*.
24. SCHUMM, S. A, 1961, *Effect of sediment characteristics on erosion and deposition in ephemeral-stream channels*. Geological Survey professional paper, 352.
25. SHEETS, B., HICKSON, T. & PAOLA, C., 2002, *Assembling the stratigraphic record: Depositional patterns and time scales in an experimental alluvial basin,* Basin research, 14**,** 287-301.

26. SHIELDS, A., 1936, *Anwendung der Ähnlichkeitsmechanik und Turbulenzforschung auf die Geschiebebewegung*. Mitt Preuss Versuchsanstalt für Wasserbau und Schiffbau, No 26, Berlin.
27. SNOW, R. S. & SLINGERLAND, R. L., 1987, *Mathematical modeling of graded river profiles*, The Journal of geology**,** 15-33.
28. THOMAS, R., 2003, *Development and evaluation of a cellular model to simulate braided river dynamics*, PhD thesis, University of Exeter.
29. WARBURTON, J., 1996, *Active braidplain width, bed load transport and channel morphology in a model braided river*, Journal of Hydrology New Zealand, 35**,** 259-286.
30. WARBURTON, J. & DAVIES, T., 1994, *Variability of bedload transport and channel morphology in a braided river hydraulic model*, Earth Surface Processes and Landforms, 19**,** 403-421.
31. WARBURTON, J., T. DAVIES AND M. MANDL, 1993, *A meso-scale field investigation of channel change and floodplain characteristics in an upland braided gravel-bed river*, New Zealand, Geological Society, London, Special Publications 75(1): 241.
32. WHIPPLE, K. X., PARKER, G., PAOLA, C. & MOHRIG, D., 1998, *Channel dynamics, sediment transport, and the slope of alluvial fans: Experimental study,* The Journal of geology, 106**,** 677-694.
33. YOUNG, W. & DAVIES, T., 1991, *Bedload transport processes in a braided gravel bed river model*, Earth Surface Processes and Landforms, 16**,** 499-511.
34. YOUNG, W. A. J. W., 1996, *Principles and practice of hydraulic modelling of braided gravel-bed rivers*, Journal of Hydrology (NZ) 35(2): 175-198.

AUTHORS BIOGRAPHY

Michael M.M is a hydraulic engineer with over 12 years of professional, research and teaching experience in design of drinking water supply projects and design of small scale dams for irrigation. He is a staff member at the school of civil and water resources engineering of Bahir Dar university, Ethiopia. Starting from January 2009, he is doing his PhD at the University of Exeter, UK. The area of his PhD is experimental and numerical investigation of sediment transport in Braided Rivers. He is interested in design of sustainable river engineering infrastructures, development and management of new technologies to control sedimentation in irrigation and hydropower systems, flow modelling in open channels, modelling of sediment transport in rivers, assessment of climate change impacts in river basins.