Available online at www.sciencedirect.com
ScienceDirect
Geomorphology 83 (2007) 152–182
GEOMORPHOLOGY
www.elsevier.com/locate/geomorph

The timing and magnitude of coarse sediment transport events within an upland, temperate gravel-bed river

Simon C. Reid [a], Stuart N. Lane [b,*], Jessica M. Berney [a], Joseph Holden [a]

[a] *School of Geography, University of Leeds, Leeds, LS2 9JT, UK*
[b] *Department of Geography, Durham University, Durham, DH1 3LE, UK*

Received 5 July 2005; received in revised form 19 April 2006; accepted 28 June 2006
Available online 4 October 2006

Abstract

This paper describes the application of a new instrument to continuously measure bedload transport, an impact sensor, to a 72 km^2 test catchment in the Yorkshire Dales, northern England. Data from a network of impact sensors are linked to repeat surveys of channel morphological response, to get a better understanding of the conditions that lead to sediment generation and transfer. Results suggest certain areas of the catchment act as key sediment sources at the annual time scale, with material being quickly delivered to the lower parts of the catchment along the steep bedrock channel. Sediment transfer within the tributaries occurs in significantly smaller magnitudes than within the main channel; but it moves more frequently and at different times of the year, with transfer rates being strongly conditioned by larger-scale valley geomorphology. The lower 5.6 km reach sees a significant reduction in gradient and a widening of the valley. This permits significant accumulation within the channel, which has persisted for many years. This lower reach is very sensitive to changes in sediment supply and there is good agreement between changes in bedload transport data and the surveyed channel response. These observations have major implications for how river management projects should be developed in upland environments, especially those where large-scale geomorphological controls have a major impact upon the sediment transfer process. Evidence suggests that where river management restricts lateral movement of the channel and transfer of sediment into floodplain storage, changes in sediment supply can lead to areas of severe accumulation, acceleration of bank erosion and exacerbated flood risk.

Keywords: Coarse sediment transport; Connectivity; Cross-sections; Gravel-bed river; Impact sensor; Climate change

1. Introduction

The high rate of coarse sediment delivery to upland river systems is an ongoing problem. It is leading to problems of river channel management, notably associated with the channel instabilities that follow from high sediment delivery rates. In-channel sediment fill also leads to exacerbated flood risk. The spatial location of sedimentation will be related to the ease with which sediment can be moved through the drainage network. It is recognised that under moderate to high flow conditions, bedload particles will only move to the next bar downstream of the erosion site (e.g. Church and Hassan, 1992; McLean and Church, 1999; Pyrce and Ashmore, 2003). At very high flows, travel distance becomes controlled more by the magnitude and duration of sediment-transporting flow conditions: particles may become mobile and step length may exceed that to the next bar downstream

* Corresponding author.
E-mail address: s.n.lane@durham.ac.uk (S.N. Lane).

0169-555X/$ - see front matter © 2006 Elsevier B.V. All rights reserved.
doi:10.1016/j.geomorph.2006.06.030

(e.g. Hassan et al., 1991, 1992). Despite recognition of the step-like nature of coarse sediment transfer, we have traditionally assumed that all coarse material supplied to a reach will eventually be transported through it (Hooke, 2003). Sediment that is not transported will either be broken down through abrasion during temporary storage until it can be transported, or stored until the channel morphology adjusts such that the partitioning of bed shear stress within the reach is competent to transport this size fraction (e.g. Baker and Ritter, 1975; Wilcock and Southard, 1989). This is the traditional basis of river engineering schemes where sediment transfer is managed through river channel engineering (e.g. channel straightening) to change the partitioning of shear stress and increase sediment transport rates, so resulting in sediment evacuation. However, this overlooks the longer-term geomorphological response of the river-floodplain system to sediment delivery, and the fact that upland river reaches may be long-term storage zones for coarse sediment.

The need to understand the controls on sediment sources, as well as the transfer of sediment to and within the drainage basin, is being amplified by concerns over environmental change. Projected increases in precipitation associated with environmental change, which are severe in upland regions in the north of England, could significantly enhance river management problems if they result in increased sediment delivery to the river network. Work in this region (Reid et al., submitted for publication) has shown that by the 2080s, coarse sediment generation could increase by up to 68% over levels typical of the 1990s, largely associated with a greater propensity for events that are sufficient to cause significant hillslope failure, as opposed to channel adjacent. In practice, the traditional separation of channel and hillslope sources is artificial as they represent end members of a continuum defined by changing degrees of connectivity between the sediment source and the drainage network. Thus, the Reid et al. work shows that a prime reason for greater sediment delivery rates is not only an increase in the magnitude and frequency of hillslope failure, but also a rising probability that these failures actually connect with the drainage network.

The need to understand the sediment delivery and transfer process is supported by a considerable volume of theoretical, experimental and field-based research into how coarse sediment moves through river systems. However, there has been much less understanding of how this is manifest over spatial scales larger than the river reach. Such an understanding is difficult to acquire as it requires monitoring of sediment transfer and channel response at a frequency that matches that of sediment transport events and over a spatial scale that is sufficient to include both hillslope sediment sources and in-channel sedimentation zones. In this paper we use continuously recorded sediment transport data, coupled to repeat surveys of channel response, to get a better understanding of the conditions that lead to sediment generation and transfer within an upland river system. This includes measurement of sediment transfer at all flows, including extreme flood events. We use this to understand which sorts of events are responsible for delivering and transferring sediment within our case study system, and which provides a baseline understanding for thinking about how future environmental changes might impact upon sediment delivery and river response.

2. Background

Coarse sediment transfer within river channel networks is a four stage process: (a) the delivery of coarse material from hillslopes or river banks to a stream; (b) entrainment at a critical shear stress from the stream or river bed; (c) transfer downstream; and (d) deposition in a temporary store or permanent sink. We now have a substantial understanding of each of these processes in isolation (e.g. in terms of controls on entrainment) and, to some extent, in terms of how they couple (e.g. the implications of weak size selectivity at entrainment for downstream transfer of delivered sediment). In relation to the delivery of sediment into streams, there are two important processes: (1) the erosion of material from the landscape; and (2) its connection to the drainage network. There are now well-established models for predicting the location of coarse sediment sources within catchments. In temperate upland environments, most coarse sediment sources are linked to rainfall-triggered shallow translational landslides (e.g. Dietrich et al., 1982; Montgomery and Dietrich, 1994; Dhakal and Sidle, 2003). Some areas of hillslopes have the potential to fail repeatedly, generating sediment for significant periods of time as individual slides, until the supply becomes exhausted. Most shallow translational landslides are formed after wet antecedent conditions followed by a prolonged period of rainfall (e.g. Brooks et al., 2004) possibly triggered by a burst of higher intensity. Wet antecedent conditions provide reduced soil moisture deficits (e.g. Campbell, 1975), keeping the catchment relatively well saturated, with a perched water table in unstable sites (Dhakal and Sidle, 2003). Prolonged rainfall gradually increases pore water pressures, sometimes just above a drainage-impeding layer, commonly the B to C horizon (e.g. Brooks et al., 1993). This in turn reduces the shear strength of the soil (e.g. Hearn and Griffiths, 2001) whilst simultaneously increasing the downslope weight of the surface material. A slide may then be initiated by high intensity rainfall which causes pressures to rise quickly until a shear stress threshold is reached (Montgomery et al., 1997; Montgomery et al., 2002). Slides have also been seen following relatively short periods of intense rainfall (e.g. summer thunderstorms) with dry antecedent conditions (e.g. McEwen and Werritty, 1988). These intense storms deliver the volumes of water required to saturate the dry soil quickly and exceed the shear stress threshold in unstable areas. The issue of the connectivity of failed sediment sources has been explored to a lesser degree, although research by Harvey (e.g. Harvey, 1987, 1997a, b, 2000) is a notable exception. In upland fluvial systems, steep, easily erodible hillslopes may supply sediment to steep, often bedrock channels (e.g. Harvey, 2001). These systems tend to be strongly coupled, with sediment transfer rates limited by the delivery of sediment from slopes to streams, and the effects of environmental change are quickly transmitted downstream. In turn, sediment delivery depends on the extent to which active zones are coupled to the hillfoot or channel. This can lead to either temporary or long-term storage of failed sediment on hillslopes in sedimentation zones that are decoupled from the channel system for a period of time. Reid et al. (in press) modelled coarse sediment delivery connectivity with the drainage network and found that maintaining hydrological connection from source to tributary was important in delivering failed sediment, so constraining sediment coupling to tributary marginal zones where hydrological connection could be more readily guaranteed. In turn, this implies a seasonal effect in relation to sediment delivery, linked to both generation and connectivity effects. Harvey (1974) found that 74% of hillslope sediment delivery occurred during winter.

Our understanding of the controls on sediment entrainment is now particularly well-developed. In rivers and streams with a coarse-bed, significant departures from perfect size selectivity have been described, arising from (after Ryan et al., 2002): (i) differences in relative roughness (Church and Kellerhals, 1978; Bathurst et al., 1983); (ii) clustering of individual grains (Laronne and Carson, 1976; Reid and Frostick, 1984; Reid et al., 1985); (iii) particle geometry (Kirchner et al., 1990; Buffington et al., 1992); (iv) imbrication (Mantz, 1980); and (v) compaction or packing of surface grains (Church and Kellerhals, 1978; Reid et al., 1985; Hassan and Reid, 1990). Many coarse-bedded channels have an armoured surface layer that is thought to suppress rates of sediment transport until a critical threshold is reached (e.g. Carling and Reader, 1982; Parker and Klingeman, 1982; Gomez, 1983; Brayshaw et al., 1983; Andrews and Erman, 1986; Church et al., 1987; Chin et al., 1994).

There has been a substantial increase in our understanding of the net sediment transfer process. Two phases of coarse sediment transport have been noted in the literature: (i) mainly sand and small gravel moving at relatively low rates over a stable channel bed during low to moderate flows (Phase I); and (ii) coarser grain movement (Phase II), linked to an increase in transport rates, including material sourced from both the channel bed and subsurface during moderate to high flows (Ryan et al., 2002). Transitions between Phases I and II imply the initiation and termination of large transport events. This somewhat ‘fuzzy’ threshold is often poorly defined, although phases and phase changes in gravel-bed channels have been noted by Jackson and Beschta (1982), Carling (1988), Ashworth and Ferguson (1989) and Warburton (1992). The threshold between phases is suggested (Ryan et al., 2002) to begin at about 80% of the bankfull discharge, but studies have shown that this can range from 60 to 100% (e.g. Parker, 1979; Jackson and Beschta, 1982; Parker et al., 1982; Andrews, 1984; Andrews and Nankervis, 1995). Phase II may be comparatively short, occurring only during parts of larger flood events. A significant proportion of the total sediment transport may therefore be moved during Phase I (Andrews and Smith, 1992; Wilcock and McArdell, 1993; Lisle, 1995). Thus, the most effective discharge, the discharge which transports most bedload over a long period (Wolman and Miller, 1960; Andrews, 1980; Emmett and Wolman, 2001) may, therefore, not be the maximum bankfull discharge. Crucially, the relative discharge which causes the change in phase may be affected by local factors such as channel type or sediment delivery rate (e.g. Laronne and Carson, 1976; Dietrich et al., 1989; Andrews and Nankervis, 1995; Lisle, 1995). However, Ryan et al. (2002) found that the phase change at 80% of the bankfull discharge was remarkably consistent, even in sites of varying slope, grain size, surface roughness and bed topography. They suggest that this could indicate that gravel-bed channels are ‘adjusted’ such that the onset of coarse sediment transport coincides with flows that occur relatively frequently.

The different phases of sediment transfer in turn have implications for both sediment sorting and the downstream characteristics of deposited sediment. Pyrce and Ashmore (2003) show using flume experiments that smaller particles travel further and faster than larger ones at low flows. Event duration has a positive effect upon travel distance (after Church and Hassan, 1992; Hassan et al., 1992; Haschenburger and Church, 1998). During moderate flows, travel distance becomes more controlled by the morphology of the channel, rather than flow conditions. Particles only move to the next bar downstream of the erosion site. At very high flows, travel distance becomes controlled more by the flow conditions where all particles can become mobile and length may exceed that to the next bar downstream. The question of how particle displacement is linked to the development of bed topography has received relatively little attention (Church and Hassan, 1992; Ashmore and Church, 1998; Sear et al., 2000) but is very important in the sediment transport process. Pyrce and Ashmore (2003) used an inverse approach, to back calculate sediment transport rates from the measured displacement of bed material (after Einstein, 1937; Church and Hassan, 1992; Hassan et al., 1992; Goff and Ashmore, 1994; Lane et al., 1995; Ashmore and Church, 1998; Haschenburger and Church, 1998). Progress has been made in the measurement of particle transfer using painted, magnetic and radio-transferring particles, but the number of particles that can be traced is small, making it difficult to link spatial variability in deposition to the sediment transfer process.

Given the above, two emerging themes can be identified. First, each of these four processes involves a response that is dependent upon the availability of sufficient energy to mobilise and to transport sediment. However, the magnitude of energy required for each of these processes is dependent on event magnitude and duration as well as local partitioning of available energy into that available for transport. This means that different events may be more or less important for each stage in the process. To understand how different events are related to different stages of sediment transfer, it is necessary to move beyond laboratory experiments or at-a-point field sampling, especially where the latter is restricted to manageable flows. In terms of larger flows, the exception has been situations where permanent bedload transport recording devices have been installed that can record all flow events (e.g. Reid et al., 1985). There has been much less investigation of sediment transport processes across a range of sizes of flow events, where records are generated simultaneously for many points in the drainage basin. Such records are central to understanding the nature of sediment delivery, including the relationship between catchment hydrology and each of the four stages identified above. This is particularly important in relation to the potential impacts of future climate change as if there is variation between components of the sediment transfer system in terms of their sensitivity to different flow events, then if climate change impacts upon the magnitude and frequency of different flow events, there could be a marked change in typical sediment delivery and transfer regimes, which will, in turn, impact upon channel response. Second, research has yet to link fluctuations in coarse sediment delivery rates to channel response over anything except very short spatial scales (i.e. reach scales). Upscaling of this linkage is crucial as it will provide the means of coupling shorter-term and smaller-scale fluctuations in sediment transfer to longer-term and larger-scale geomorphological processes, especially in relation to long-term evolution of river-floodplain systems.

This paper aims to improve understanding of the short-term interaction between sediment delivery and channel response in a 72 km^2 upland catchment in relation to the conditioning effect of larger-scale and longer-term aspects of valley geomorphology. The paper provides results from continuous recording of coarse sediment transfer at 10 locations within the catchment and from repeated measurements of river channel response at sixty cross-sections.

3. Methodology

3.1. Study catchment

This research is based in the Upper Wharfe catchment (Yorkshire Dales National Park, northern England) from the headwaters of Greenfield and Oughtershaw down to the Cam Gill Beck confluence, just upstream of Starbotton (Fig. 1). The drainage area is 72 km^2 with an elevation range from 211 m above Ordnance Datum (OD) at the catchment outlet close to Starbotton to 701 m on the divide close to Buckden Pike. The catchment consists of shallow blanket peat on the hilltops, rough grass and moorland on the middle slopes with sheep grazing and pasture land on the lower

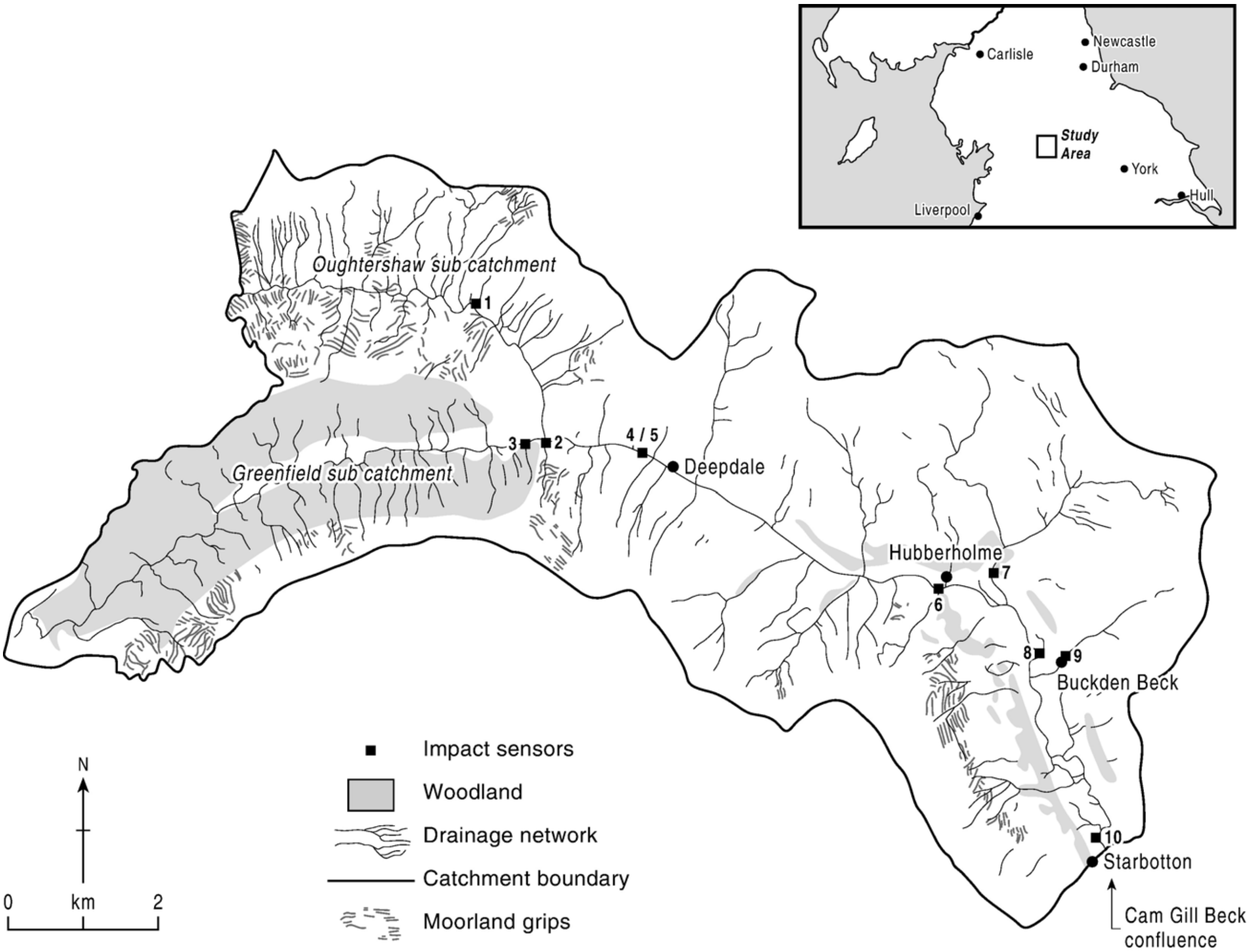

Fig. 1. Location of the study site and the impact sensors shown in Table 1. Stage was recorded continuously at each impact sensor location and transformed to discharge using a rating equation based upon spot gauging and rating curve extension.

floodplain. Coniferous forests were planted commercially at Greenfield during the 1950s, these are due to be removed within the next five years. Some of the moorland headwater areas have been drained using shallow open drains (grips), which are thought to have extended the channel network, leading to steeper hydrographs with both decreased time-to-peak and shortened recession limbs. This may have implications for both sediment generation and transport (Hey and Winterbottom, 1990). The catchment is dominated by prevailing westerly air streams and receives 1750–2000 mm of precipitation per year. The long-term average annual precipitation for Buckden is 1710 mm, with just 433 mm of annual evapotranspiration (Heritage and Newson, 1997). The area appears to be very sensitive to both localised convective summer thunderstorms and winter cyclones, which tend to produce high-intensity rainfall events (Merrett and Macklin, 1999).

Catchment geology comprises Carboniferous age Great Scar Limestone, Yoredale Series and Millstone Grit. Distinctive horizontal terraces have been formed in places by a combination of erosion and glacial processes. The headwaters of the catchment are, in many places, deeply incised into the Millstone Grit. Due to Pleistocene glacial activity and steep slopes, very thin soils overlie the bedrock, which can lead to ready sediment generation from the hillslopes. The main channel from the headwaters to Hubberholme consists mainly of exposed limestone bed and banks, with minimal sediment accumulation. This relatively steep reach permits rapid sediment transport due to low frictional resistance. Hence, generated sediment from the headwaters and upper tributaries can quickly be moved to the lower areas. Just upstream of Hubberholme, the channel gradient is significantly reduced as the topography changes from a steep narrow valley to a classic U-shaped, wide deglaciated floodplain. This permits a significant increase in sediment accumulation which exacerbates flood risk. There is also a significant increase in channel sinuosity, which itself may reduce particle step lengths and, hence, increase accumulation rates.

3.2. Hydrological measurements

Measurement of precipitation across the catchment was essential for this study due to the spatial heterogeneity of precipitation, especially in upland areas such as the Upper Wharfe. Ten tipping bucket rain gauges (Campbell ARG100s) were twinned with Hobo event data loggers and were located across the catchment. Precipitation data were collected with a 5 min time-step to allow direct comparison to discharge and sediment transport. In-stream sediment transport monitors were located at ten sites and discharge was estimated at the location of each monitor (Fig. 1 and Table 1). Stream stage was measured using Eijkelkamp/Van Essen pressure transducers (Divers) with automated data loggers in stilling wells. These were located at Buckden Beck, Cray Beck and Starbotton and recorded water depth every 5 min. Stage data were corrected using an atmospheric pressure Diver located between the three instruments. Stage data for Oughtershaw, Greenfield, Deepdale and Hubberholme were obtained from Environment Agency stilling wells as part of their HydroLog data system. Stage at the Beckermonds site was not recorded due to installation problems on the eroded bedrock but discharge was estimated by catchment-area scaling based on the Oughtershaw data upstream. Buckden Bridge stage was not recorded since there were no tributary inputs between it and Cray Beck and therefore the discharge here closely approximated to the sum of Hubberholme and Cray Beck discharges. Stage (m) was converted to discharge (in $m^3\ s^{-1}$) using rating curves derived from stream gauging. For the smaller tributaries at low flows, gaugings were based upon the salt dilution method, due to the turbulent nature of the channel. The tracer used was sodium chloride since it is hazard-free and has minimal adverse environmental impacts. For the rating curves, only results from triple-replicate measurements with a standard deviation of <10% were used. For the main channel a velocity-area approach was used with a Bray-Stoke Flowmeter. Discharge was measured using standard velocity-area methods (Buchanan and Somers, 1969; Nolan and Shields, 2000). Data for both rainfall and discharge were available from March 19th, 2003 until March 16th, 2004. In order to establish the representativeness of results obtained from only one year of data, Fig. 2 shows the flow data for Flint Mill, 70 km downstream from the study site. This is used as the benchmark site as data extend back to 1956. If we assume that the Flint Mill data scale down with catchment area, we can conclude that for the study reach, 2003–2004 was a relatively dry period in relation to flow, consistently below the median flow results, except for flows lower than the *Q80*, which fall on the median, and for flows greater than *Q5*, which were some of the lowest on record.

3.3. Options for measurement of sediment transfer processes

Path length, in-channel sources and deposition sites have been measured using a variety of techniques. Tracers have traditionally been used to study clast movement in gravel-bed channels. Painted natural pebbles (e.g. Leopold et al., 1966; Wilcock, 1997), magnets inserted into natural pebbles (e.g. Hassan and Church, 1992; Reid et al., 1992; Ferguson and Wathen, 1998; Haschenburger and Church, 1998; Stott and Sawyer, 2000) and radio-tracked particles (e.g. Habersack, 2001) are all commonly used techniques. Whilst most tracer studies have low recovery rates, are

Table 1
Characteristics of the rain gauge, discharge measurement and impact sensor locations, with site gradient within 10 m and 100 m of the sensor

Location	Site number	Type	Bankfull channel width (m)	Site slope (10 m)	Site slope (100 m)
Oughtershaw	1	Tributary	3.2	0.039982	0.084145
Beckermonds	2	Tributary	7.7	0.001598	0.020086
Greenfield	3	Tributary	5.1	0.002533	0.015270
Deepdale (2 sensors)	4/5	Main channel	10.4	0.005662	0.010666
Hubberholme	6	Main channel	9.5	0.029214	0.015919
Cray Beck	7	Tributary	7.1	0.006852	0.025330
Buckden Bridge	8	Main channel	9.3	0.004563	0.007581
Buckden Beck	9	Tributary	1.6	0.065092	0.107914
Starbotton	10	Main channel	11.1	0.023126	0.004351

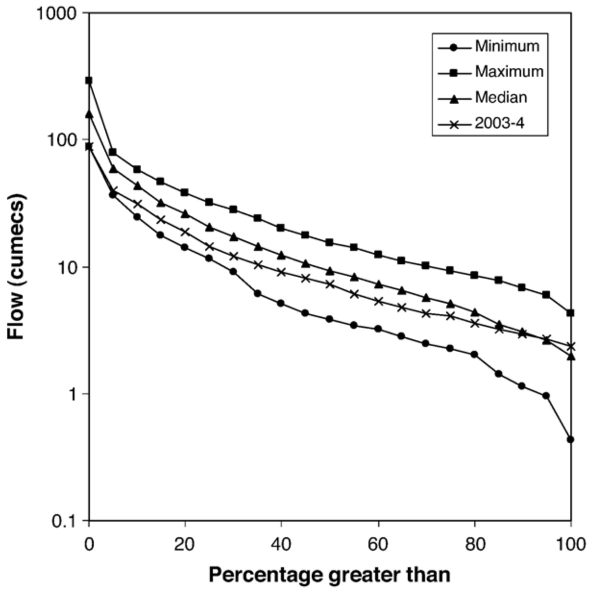

Fig. 2. The flow characteristics are based upon the long-term record for Flint Mill, ca. 70 km downstream.

applied across few and infrequent displacement events and have imprecisely known flow characteristics (Pyrce and Ashmore, 2003), they have revealed a great deal about the net changes from sediment transport and deposition within river reaches (Tunnicliffe et al., 2000). Wilcock (2001) suggests that for a true empirical relationship between sediment and water discharges to be gained, a large number of samples are needed over a complete range of flows (e.g. Emmett, 1980; Hubbell, 1987). Instantaneous and continuously logged observations of bedload transport are relatively uncommon (Tunnicliffe et al., 2000), but use several methods: (i) net and basket samplers (e.g. Helley and Smith, 1971; Emmett, 1980; Bunte, 1996); (ii) buckets and pit traps (e.g. Hollingshead, 1971; Church et al., 1991; Powell, 1992; Powell and Ashworth, 1995); and (iii) in-situ magnetic detection instruments for clasts with implanted magnets or the minor signals from naturally iron-rich clasts (e.g. Ergenzinger and Conrady, 1982; Ergenzinger and Custer, 1983; Reid et al., 1984; Custer et al., 1986; Spieker and Ergenzinger, 1987; Custer et al., 1987; Bunte, 1996; Tunnicliffe et al., 2000). These methods suffer from a range of problems. Fixed devices are often expensive, easily damaged and, in the case of pits and traps, may fill before the flood event is over (e.g. Powell, 1992). Net samplers can be dangerous at high flows and so a full range of conditions cannot be sampled using this method. Duizendstra (2001) sampled bedload transport from bridges using an enlarged Helley-Smith (BTMA-2) sampler but found that estimates of transport varied from direct channel measurements.

Given the above problems, it is not surprising that initial motion and bedload transport are hard to measure in alluvial and bedrock channel systems, especially across the full range of measured flow events. However, Carling et al. (2002) reported on a continuously logging impact sensor that is robust at high flows yet deployed and downloaded at low flows. It detects the acceleration of a steel plate fixed to a rock channel bed or large boulder, after the plate is struck by a clast (greater than 3 mm) in motion. The instrument is installed flush with the channel bed or gravel surface to ensure that it only detects true bedload motion. Carling et al. (2002) found that the results from a sensor matched contemporary theory. Above a critical shear stress for sediment movement initiation, further increases in shear stress could be related to an increase in transport rates. On the falling limb, transport rates show a gradual decline at first, followed by a more rapid decrease as the threshold is reached. Two main issues arise. First, the response of the impact sensor is non-linear (Carling et al., 2002). This is because the sensor can only record a maximum of three clasts per second. Hence, the likelihood of more than one clast hitting the plate in less than a third of a second increases with shear stress (discharge) and transport rate. Therefore, the results tend towards a lower boundary estimate as transport rates increase. This leads to a conservative estimate at higher flows and a more reliable estimate at low to moderate flows. Second, bedload is commonly routed along well defined and narrow pathways within upland rivers (Carling et al., 2002) and so installation location is vital to obtain representative data. These issues aside, the sensors can be used: (a) to obtain very reliable estimates of the timings of sediment entrainment and transfer cessation; and (b) a less reliable estimate of the relative intensities of transport through time. In theory, this can be done at many points in a drainage basin as the sensors are easy to install and relatively robust in the face of extreme flow events.

3.4. Sensor design and placement

The sensor was made of a Tinytag Plus TGPR1201 Count Input Data Logger adapted to include an accelerometer. All ten loggers used in the study were set to detect only grains greater than 20 mm in diameter. The logger was 80 × 60 × 35 mm in size and the steel plate was 150 × 130 × 6 mm (Fig. 3). The installed memory allows 55 days of logged information at 5 min intervals, with a 12 month battery life. Data used here range from March 19th, 2003 until March 16th, 2004 to allow comparison with the hydrological data.

Drake et al. (1988) found that only fine gravel (2 to 3 mm) saltated, whilst coarser particles rolled along the bed, with corners and edges contacting the bed on average about twice per particle diameter of travel. Therefore, the sensor plate (150 × 130 × 6 mm) is of the correct size to record a rolling particle only once (in this case), yet large enough to catch those saltating, if required. Given the need for careful sensor location, optimal location was gained through analysis of the first autumn floods in September 2002. In bedrock and gravel-bed channels, the low summer flows often lead to the gravels becoming covered with algae due to shallow water depths and minimal movement. The first autumn floods scour the gravels and leave 'tracks' of clean gravels which identify the most active areas of the bed (Carling et al., 2002).

Six sensors were installed within the main channel and four within the tributaries (Table 1). All were located next to the stage recorders described above (Fig. 1), to minimise error in relationships between hydrological records and sediment movement. There were nine individual locations, with two sensors located at Deepdale. This provided a comparison between loggers, situated next to each other in line with the flow direction, to allow testing of sensor reliability and variability under identical hydrological conditions. There were no significant discrepancies between the two sensors. The loggers can record between 0 and 255 impacts within a given time interval and a maximum of 3 impacts per second. All of the main channel sensors were often saturated during larger storm events, recording 255 on a number of occasions, out of a theoretical maximum of 900. The solution to this would be to lower the time-step to 1 min (from 5 min) but this would require a download every ten days, which was not practical. Hence, data were converted from actual records to a 'level of saturation' by dividing each record by 255 to scale data between 0 and 1. This is deemed the 'transport intensity'. Full saturation of 1 is equivalent to 255 clasts in 300 s, which gives a mean transport rate of 0.85 clasts per second. The data for each sensor were upscaled to the channel width to get a mean transport

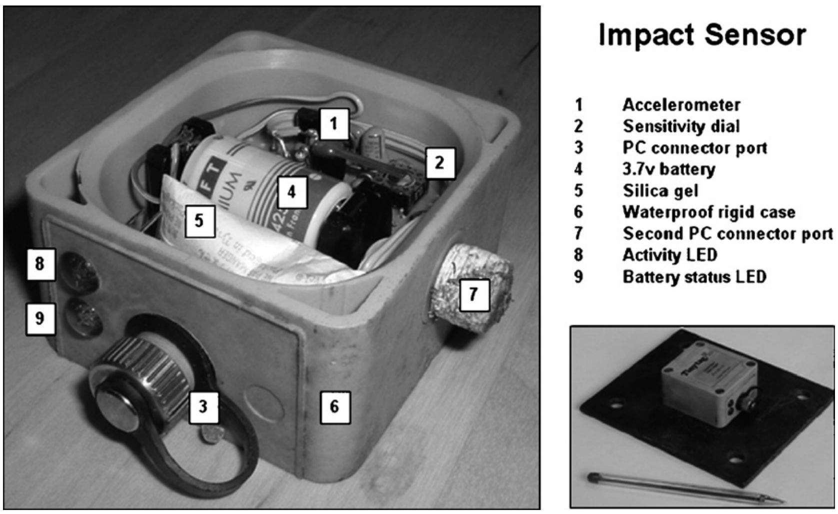

Fig. 3. Schematic diagram of an opened impact sensor and attached to the metal plate, ready for installation.

intensity at each sample interval which assumes an equal transport rate across the section. This is defined as the 'relative transport intensity' (r_n):

$$r_n = t_n \times \left(\frac{w}{0.13}\right) \tag{1}$$

where t is the transport intensity in clasts per second, w is the channel width (m) and 0.13 is the width (m) of the sensor plate (i.e. sampled width of the channel section). Use of Eq. (1) requires us to make a major (and probably incorrect) assumption: the sediment transport rate across the sensor is the same as across the full width of river channel. Our choice of sensor location, based on the first autumn storms of 2002, was such that the most active portions of the bed were identified. Subsequent visits showed that the width of the active bed did increase during the following winter, and it appears that the active width is related to event size. By assuming constant transport across the full width, we will be reducing variability between sediment transporting events (i.e. small events are scaled upwards by too great a degree), and this may make cross-comparison of relative intensity estimates between sites somewhat problematic. However, without sacrificing the goal of the project of obtaining a measure of spatial variability, this problem cannot be avoided. As we chose the most active sections of the bed, we are able to comment on when transport begins and ends, even though our relative intensity estimates are less reliable. Data periods where the sensors were recording no movement were removed, as were values less than 10% of the maximum value for each site. This provided a data set of larger, more extreme events, with low flow effects removed.

3.5. River system response

Repeat topographic surveying of the bed can provide information on the net volumetric change of bed material and the migration of bedforms (e.g. Ferguson and Ashworth, 1992; Lane et al., 1994; Martin and Church, 1995). The surveyed reach length must not exceed the step length, otherwise material may pass undetected. Cross-section surveys capture the bed topography at a certain moment in time and must be surveyed at a frequency similar to that of large flood events to be able to accurately link sediment transport to individual events and hence relate transport rates to effective discharge. However, they do provide a reliable means by which bedload transport volumes can be estimated and monitored (Davies, 1987), especially if done to a high level of precision, probably with an error less than the D_{50} (Lane et al., 1994). Use of specific surveys and a time-integrated approach avoids some of the complications caused by spatial and temporal variability of bedload measurements by treating the cross-section as a single entity (Hubbell, 1987; Fuller et al., 2003). Cross-sections can also detect areas of erosion and deposition which are important in assessing channel stability (Ashmore and Church, 1998). The majority of this work has concentrated on the reach scale (e.g. Neill, 1987; Ferguson and Ashworth, 1992; Ferguson et al., 1992; Warburton et al., 1993; Lane et al., 1994). The major instability zones within UK rivers are typically found at the inter-reach scale of 1 to 3 km (Fuller et al., 2003) and relatively few studies have addressed this (e.g. Brewer and Passmore, 2002), with transfers rarely quantified (Wathen et al., 1997). Many studies on gravel-bed rivers have used cross-sections to derive volumetric change information (e.g. Ferguson and Ashworth, 1991; Goff and Ashmore, 1994; Martin and Church, 1995; Paige and Hickin, 2000; Ham and Church, 2000; Fuller et al., 2002). The cross-section approach has an inherent weakness, in that it only provides valuable information at-a-point and has extrapolation weakness (Fuller et al., 2002). Furthermore, the section may not fully represent downstream changes in channel morphology through time and space (Lane et al., 1994). Additionally, the section may miss processes such as erosion and deposition which are occurring between sections but do not migrate downstream (Naden and Brayshaw, 1987; Wittenberg, 2002). Martin and Church (1995) showed an inaccuracy of 10 to 20% in predicting coarse sediment transport rates and volumes when an equal volume of sediment between cross-sections was assumed. There is therefore a need to consider the variability of sediment transport within and between reaches (Rice and Church, 1998; Harvey, 2002). Sediment budgets based on interpolation of cross-section data are subject to a high degree of uncertainty for this reason (Brasington et al., 2000). Those derived from Digital Elevation Model (DEM) subtractions (e.g. Lane et al., 1994, 1995; Milne and Sear, 1997; Stojic et al., 1998; Eaton and Lapointe, 2001; Chappell et al., 2003; Lane et al., 2003) have moved towards a 'distributed terrain-sensitive survey' (Ashmore and Church, 1998; Westaway et al., 2000; Lane, 2001), whilst others have used cross-sections extracted from DEMs for analysis (Fuller et al., 2003). Although the use of DEMs represents landform surface variability well (e.g. Zevenbergen and Thorne, 1987; Lane et al., 1994; Lane, 1998), Brasington et al. (2000) and

Fuller et al. (2003) noted the practical difficulty of topographic data acquisition at a resolution adequate to reflect channel morphology, especially using high resolution Global Positioning System (GPS) survey under riparian trees. Hence, DEM-derived changes in morphology are not often used in studies of size greater than the bar-scale, where cross-section data can be acquired precisely and processed quickly, especially using high resolution GPS survey (Brewer and Passmore, 2002). As a result, sediment transport rates and volumetric changes derived from cross-section surveys represent lower bound estimates (Fuller et al., 2003) since: (i) sediment may move through or within the reach without and surface expression at the specific cross-section site; and (ii) significant negative bias in volumes of erosion and deposition within the reach can be produced by localised changes at the survey site which may not represent changes in the reach to the next cross-section (Lindsay and Ashmore, 2002). Since sections remain fixed during the study period, irrespective of later channel changes, their initial sites must be chosen with considerable care, such that they: (a) fully represent the subreach under study; (b) represent the processes under study and the length-scale over which these processes operate; and (c) should not erode too severely during the study such that surveys would become impossible or dangerous. With these considerations, accurate error propagation is needed for estimates of sediment transport rates derived from cross-section analysis (Fuller et al., 2003).

In this study, significant channel response is only evident in the lower reach where there is a marked reduction in valley gradient (downstream of Hubberholme, Fig. 1). Thus, river channel response in the lower reach was quantified using repeated surveys of 60 cross-sections within the 5.6 km reach from Hubberholme Bridge to the Cam Gill Beck confluence at Starbotton. 72% of this reach is relatively straight and 28% has clear meanders. The sections were located between 30 m and 200 m apart (Fig. 4) to reflect channel morphological complexity. On straight reaches, the cross-sections had a mean spacing of 111.5 m, compared to 48.4 m on meandering reaches. This choice reflected observations that meandering reaches were much more dynamic in relation to sediment delivery events, as well as being morphologically more complex, both of which required closer cross-section spacing. The meander section spacing was such that it did not exceed the step length of particle movement, which can be similar to the meander wavelength (Pyrce and Ashmore, 2003).

Of the 60 cross-sections, 52 were surveyed with a Leica Geosystems SR530 RTK 24 channel dual-frequency real time Global Positioning System (GPS) receiver with a mean point precision of better than ± 0.005 m± 0.5 ppm. The remaining 8 were surveyed with a Leica Geosystems TPS1100 Professional Series Total Station with a mean point precision of better than ± 0.005 m± 2.0 ppm. The Total Station approach was used to survey cross-sections underneath dense tree cover where the GPS equipment could not get a fix to the required level of precision, even during the winter when there were no leaves on the trees. The Total Station was used as it communicated with the GPS reference station and surveyed points were obtained to approximately the same level of precision as using the GPS directly. Surveys were conducted in March and December from December 2001 until March 2004, but only the March 2003, December 2003 and March 2004 data are used here, since hydrological and impact sensor data were only available for this period. The mean number of survey points per cross-section for all surveys was 115, with a mean across-section spacing of 0.181 m between adjacent survey points, with all breaks of slope identified. The mean vertical error of individual points from all surveys was 0.016 m, with a mean horizontal error of 0.010 m. These errors are sufficiently low to not undermine the across-section point spacing. The detection of volume change and propagation of error was conducted using the end area method. The volume (V) between two cross-sections (l and $l+1$), separated by distance D is given by their end areas (A) as:

$$V_{l,l,+1} = D_{l,l,+1}\frac{A_l + A_{l+1}}{2} \tag{2}$$

Hence, the change in volume between any two periods is given by:

$$\Delta V^{t,t,+1}_{l,l,+1} = \frac{D_{l,l,+1}}{2}\left(A^{t+1}_l - A^t_l + A^{t+1}_{l+1} - A^t_{l+1}\right) \tag{3}$$

where t is time. If the upper surface of the cross-section is set so as to be horizontal and to run through the bank with the highest elevation, then any area can be determined from:

$$A^t_l = \sum_{i=1}^{n-1}\left({}^{i+1}h^t_l + {}^{i}h^t_l\right)\left({}^{i+1}x^t_l - {}^{i}x^t_l\right) \tag{4}$$

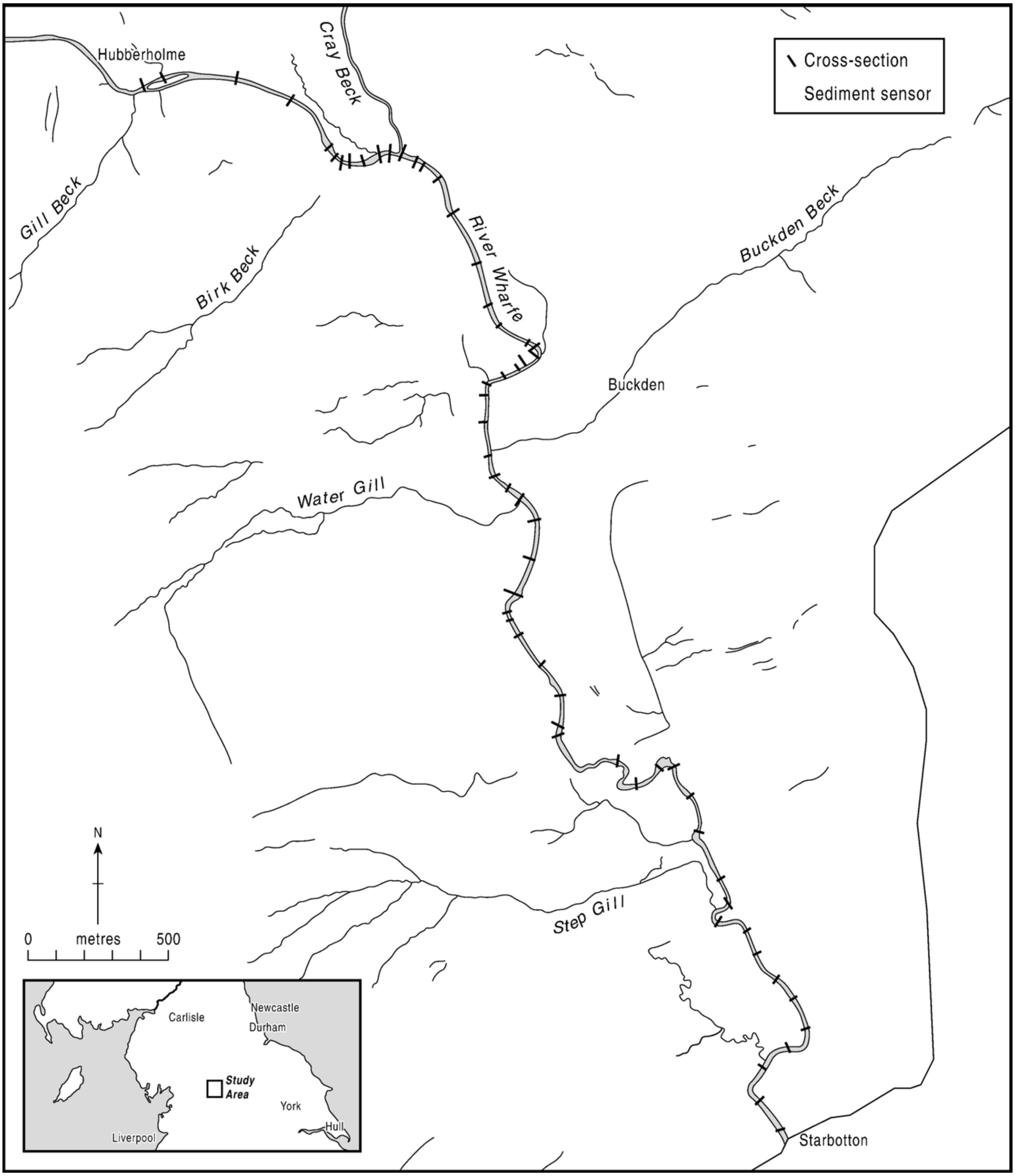

Fig. 4. Cross-section locations within the 5.6 km surveyed reach.

where h is the difference between bed elevation at position x and the bank with the highest elevation and i is the index position along cross-section l at time t. Thus, Eq. (3) becomes:

$$\Delta V_{l,l+1}^{t,t+1} = \frac{D_{l,l+1}}{2}\left(\sum_{l=l}^{l+1}\sum_{i=1}^{n-1}\left({}^{i+1}h_l^{t+1} + {}^{i}h_l^{t+1}\right)\left({}^{i+1}x_l^{t+1} - {}^{i}x_l^{t+1}\right) - \sum_{l=l}^{l+1}\sum_{i=1}^{n-1}\left({}^{i+1}h_l^{t} + {}^{i}h_l^{t}\right)\left({}^{i+1}x_l^{t+1} - {}^{i}x_l^{t}\right)\right) \tag{5}$$

Following Taylor (1997), the uncertainty in a parameter q derived from variables $x \ldots z$ measured with uncertainties represented by their associated standard deviation of error (σ) is given by:

$$\sigma_q = \sqrt{\left(\frac{\partial q}{\partial x}\sigma_x\right)^2 + \ldots + \left(\frac{\partial q}{\partial z}\sigma_z\right)^2} \tag{6}$$

provided the uncertainties in $x \ldots z$ are random and independent and based upon elevation errors that are Gaussian. The assumption of independence applies in this case. Under the assumption that there is uncertainty in D (σD), h (σh) and x (σx), then application of Eqs. (5) to (6) gives:

$$\sigma_{\Delta V_{l,l+1}^{t,t+1}} = 0.5\sqrt{\begin{array}{l}\left(\sigma_D^2\left[\left(\sum_{l=l}^{l+1}\sum_{i=1}^{n-1}\left({}^{i+1}h_l^{t+1} + {}^{i}h_l^{t+1}\right)\left({}^{i+1}x_l^{t+1} - {}^{i}x_l^{t+1}\right) - \sum_{l=l}^{l+1}\sum_{i=1}^{n-1}\left({}^{i+1}h_l^{t} + {}^{i}h_l^{t}\right)\left({}^{i+1}x_l^{t} - {}^{i}x_l^{t}\right)\right)\right]\right)^2 \\ + \left(D_{l,l+1}^2\sigma_h^2\sum_{t=t}^{t+1}\sum_{l=l}^{l+1}\sum_{i=1}^{n-l}\left({}^{i+1}x - {}^{i}x_l^t\right)^2\right) + \left(D_{l,l+1}^2\sigma_x^2\sum_{t=t}^{t+1}\sum_{l=l}^{l+1}\sum_{i=1}^{n-l}\left({}^{i+1}h_l^t + {}^{i}h_l^t\right)^2\right)\end{array}} \tag{7}$$

Given that uncertainties in elevation, position and cross-section separation can be assumed to be independent, Eq. (7) gives the associated uncertainty in volume estimates. It does not give the systematic error that will arise due to the effects of cross-sectional representation of the continuous variation in river bed morphology. The associated bias is minimised through the cross-section sampling strategy that we reported above.

4. Sediment dynamics

4.1. Sediment transport events

During the period March 19th 2003 to March 16th 2004 there were numerous distinctive sediment transport events (excluding minor events which were less than 10% of the maximum value at each site). Distinct events were defined as having at least 24 h between the peaks of consecutive sediment transport events, so they could not be confused with a single event of multiple pulses. Events have been classified based upon their peak relative transport intensity (RI) into: (i) small events, $\mathrm{RI} \leq 5.00$; (ii) moderate events, $5.01 \leq \mathrm{RI} \leq 25.00$; and (iii) large events, $\mathrm{RI} \geq 25.01$. The rationale for this division came from an exploration of data for all

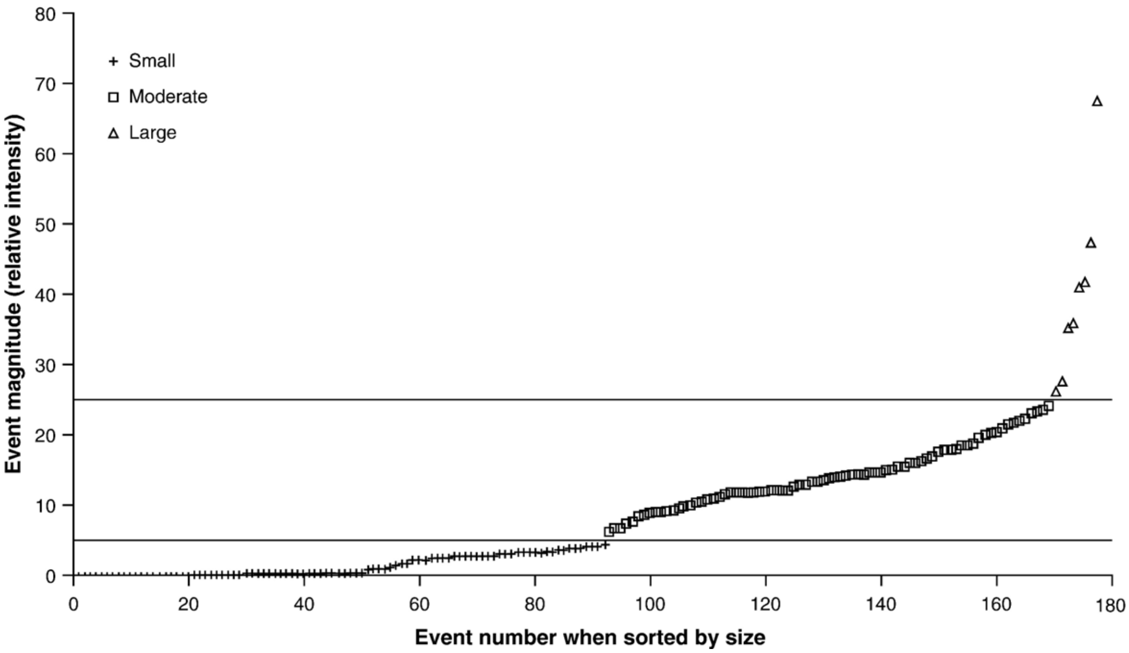

Fig. 5. Transport events from all 10 sites plotted by size to show three distinct categories of event magnitude.

sites. All 177 distinct transport events from the 10 impact sensors were collated and the data set sorted by relative transport intensity. These data were then plotted (Fig. 5) with three clear sizes of event evident suggesting breakpoints at RI = 5 and RI = 25. This division provides a means of comparing sites in terms of their relative transport intensities (Tables 2 and 3).

The tributaries and headwaters had many small events, but few moderate to large events, even though the total number of events may have been similar to the main channel. For example, Oughtershaw had 21 events during the year, all of which were small. The main channel sites had more moderate to large events, with fewer smaller ones. For example, Hubberholme had no small events, yet 26 moderate and 3 large. Some differences to this general pattern are seen. Greenfield, a headwater site and the most inactive site during the study, had just 2 small events, but 6 moderate and no large, which suggests that the annual transport regime here is dominated by a few, short duration events. Starbotton, the last main channel site had 29 small events, with just 4 moderate and 2 large, which implies that sediment moved past the site during small frequent events and that large flood events here do not transfer large volumes of sediment. In terms of total sediment transport (summed relative intensity), the drained peatland headwater area of Oughtershaw and the forested Greenfield cannot account for the majority of sediment production, even when uncertainty in determining the relative intensity measure is considered, as the differences in totals between sites were so large. Total activity (summed relative intensity) at Oughtershaw for the year was 156.5, with 90.4 for Greenfield, which indicates that: (i) these areas are not significant sources as compared with either remobilisation of in-channel sources or delivery of sediment from other sources; and (ii) mature forested areas appear to reduce coarse sediment generation since the Oughtershaw and Greenfield areas are of similar size, have similar soils and are subject to the same weather conditions. Just downstream of Oughtershaw, Beckermonds had a total activity of 6469.7, which reflects sediment generation in the reach upstream, associated with river erosion into late Quaternary deposits. The mean activity for Deepdale was 2867.3, which suggests that there may be large volumes of sediment transfer to storage downstream of Beckermonds. Downstream to Hubberholme, a series of smaller tributaries appear to have the ability to cut into the Millstone Grit and glacial deposits and generate great volumes of sediment which are easily transported by the bedrock channel, leading to a total activity at Hubberholme of 291675.2. Field observations suggest no storage within this reach. From Hubberholme downstream, there is a marked reduction in annual transport intensity: since only 10.1% of this sediment reached Buckden Bridge, which has a total activity of 29305.8. Only 3.0% of sediment passing Buckden Bridge reached Starbotton (889.3). This implies rapid deposition downstream from Hubberholme, which is expected given the associated reduction in valley slope. The tributary inputs of Cray Beck (885.2) and Buckden Beck (243.7), although they are sources, are not large enough to significantly alter the sediment transport characteristics of the Upper Wharfe. Both join the main channel within the depositional reach. However, the delivery rates from Buckden Beck are significant when compared with the rates recorded downstream at Starbotton.

Of all the main channel sites, Hubberholme had the greatest total transport time, accounting for 5.6% of the year, which was much higher than any of the other sites. Buckden Bridge had recorded sediment transport occurring for 0.67% of the time, with the least activity occurring at Starbotton which is at lowest end of our study catchment (0.13%). The latter value is similar to

Table 2
Characteristics of sediment transport events (low flow events removed) seen at the 10 sites during the study period: T indicates Tributary or Headwater

Location	Number of distinct events	Event sizes (small; moderate; large)	Total time in motion (h)	Percentage of time in motion	Total activity (sum of RI)	Maximum movement during largest event (clasts per second)
1. Oughtershaw (T)	21	21; 0; 0	468	0.09	157	9.9
2. Beckermonds	16	15; 1; 0	3392	0.65	6470	18.6
3. Greenfield (T)	8	2; 6; 0	96	0.02	90	1.0
4. Deepdale upstream	3	0; 3; 0	384	0.07	3002	8.8
5. Deepdale downstream	5	0; 5; 0	292	0.06	2733	15.7
6. Hubberholme	29	0; 26; 3	29252	5.59	291675	30.7
7. Cray Beck (T)	19	15; 4; 0	620	0.12	885	7.7
8. Buckden Bridge	31	0; 28; 3	3004	0.57	29306	40.3
9. Buckden Beck (T)	10	10; 0; 0	1264	0.24	244	2.3
10. Starbotton	35	29; 4; 2	700	0.13	889	57.5

Table 3
Distribution of event magnitude at each site during the study period

Relative intensity	Event size	1. Oughtershaw	2. Beckermonds	3. Greenfield	4. Deepdale upstream	5. Deepdale downstream	6. Hubberholme	7. Cray Beck	8. Buckden Bridge	9. Buckden Beck	10. Starbotton
<0.50	Small	16	0	0	0	0	0	0	0	6	19
0.51–1.00		3	0	0	0	0	0	0	0	2	4
1.01–2.00		2	0	0	0	0	0	0	0	1	3
2.01–5.00		0	15	2	0	0	0	15	0	1	3
5.01–10.00	Moderate	0	0	3	1	2	14	4	13	0	1
10.01–25.00		0	1	3	2	3	12	0	15	0	3
25.01–50.00	Large	0	0	0	0	0	3	0	3	0	1
50.01–100.00		0	0	0	0	0	0	0	0	0	1
>100.01		0	0	0	0	0	0	0	0	0	0

that for the tributaries with much smaller upslope contributing areas such as Oughtershaw (0.09%) and Cray Beck (0.12%). A greater duration of sediment motion was recorded at Beckermonds (0.65%), but storage downstream led to a low value at Deepdale (0.06%). The most inactive site was Greenfield with sediment transport recorded for just 0.02% of the time. This subcatchment had around 80% mature plantation forest cover and river bed had an extensive biofilm and algal development thereby supporting the sensor data on sediment inactivity. The forest cover is likely to have changed both the hydrology and the sediment delivery of the catchment. Only the sediment delivery is explored here, but the hydrological changes may have exacerbated the situation.

4.2. Event timings and seasonality

Precipitation across the catchment followed a similar annual distribution, although the yearly total in the headwaters (Greenfield and Oughtershaw) was slightly greater than the lower parts of the catchment. For example, for the study period March 19th 2003 to March 16th 2004 the total rainfall at Greenfield was 1624 mm, compared to 1258 mm at Buckden. The long-term average for Buckden (1961 to 1990) is 1710 mm, which suggests that this was a relatively dry year.

Fig. 6 shows precipitation at Greenfield during the study period. There is a seasonal variation (as with all sites) between the summer convective thunderstorms, which are of high intensity but short duration and the more prolonged winter rainfall events, some of which may be affected by snow melt. The summer of 2003 was relatively dry in this region with rainfall in August at 34% below average (CEH Wallingford, 2003). However, precipitation that did occur was associated with isolated storm events, which resulted in some sediment transfer. It was not possible to identify a 'critical' precipitation which caused a discharge large enough for the onset of sediment transport, due to the spatial complexity of the catchment and the great variability in rainfall throughout the year (Table 4). All sites had a large variability in rainfall–runoff lag times. For example, the catchment outlet at Starbotton had a lag time ranging from 7 to 17 h between peak rainfall and peak discharge, depending on rainfall magnitude and duration. Similar rainfall events did not generate the same magnitude discharge events. Furthermore, equal discharges at the same site did not cause the same amount of sediment transport. Therefore, a clear relationship between discharge and sediment transport cannot be defined. This is probably due to the

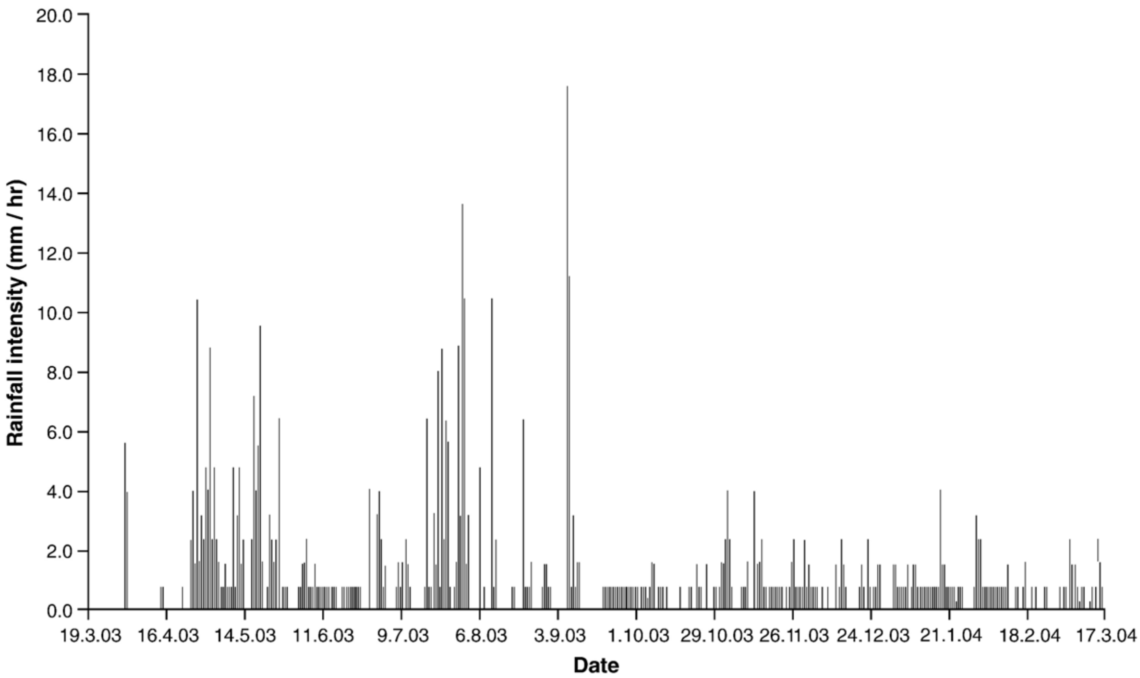

Fig. 6. Precipitation at Greenfield during the study period within 1 h periods.

availability of sediment from sinks upstream, and the relative importance of new sources (i.e. supply from the hillsides and channel banks) versus remobilisation of sediment delivered to the system during earlier events. Lag times between rainfall and discharge were reduced for all sites during the winter months when the catchment was wetter allowing rapid flow to be generated more readily and, presumably, sediment to be delivered to and transported within the drainage network earlier during the event. The majority of larger-moderate and large sediment transport events occurred during the autumn and winter months (Fig. 7). Smaller-moderate and small events occurred throughout the year, although these were more common during the summer months, possibly as a result of rapid sediment generation following high intensity thunderstorms.

Taking October to March to represent the winter months and April to September as the summer months, it was possible to identify some important differences in seasonal sediment transport. For each site, the proportion of total activity (summed relative intensity) was calculated for the summer and winter months. For the five tributary sites a mean of 36% of activity occurred during the winter. However, for the main channel sites 61% of activity occurred during the winter. Although the distribution of the percentage activities should be treated

Table 4
Main characteristics of the larger sediment transport events

Location	Peak rainfall to peak discharge lag (h:min)	Maximum discharge (m^3 s^{-1})	Mean discharge (m^3 s^{-1})	Critical discharge to initiate sediment transport	Peak discharge to peak sediment transport (h:min)
1. Oughtershaw	2:30 to 4:00	3.41	2.44	No relationship	No relationship
2. Beckermonds	0:15 to 6:00	5.28	2.73	3.2 m^3 s^{-1}	1:00 to 3:30
3. Greenfield	1:00 to 3:45	3.29	1.04	No relationship	No relationship
4/5. Deepdale (2 sensors)	1:00 to 4:00	14.52	6.65	8.1 m^3 s^{-1}	0:30 to 1:30
6. Hubberholme	1:25 to 5:30	18.86	7.86	3.0 m^3 s^{-1}	0:30 to 2:30
7. Cray Beck	1:00 to 6:00	2.22	1.72	1.8 m^3 s^{-1}	0:15 to 2:30
8. Buckden Bridge	2:30 to 8:15	22.57	7.43	6.0 m^3 s^{-1}	0:00 to 0:15
9. Buckden Beck	2:30 to 7:00	1.04	0.80	0.9 m^3 s^{-1}	1:15 to 4:30
10. Starbotton	7:00 to 17:15	29.89	10.12	No relationship	No relationship

All data relate to the identified 90% largest events and is not a yearly average.

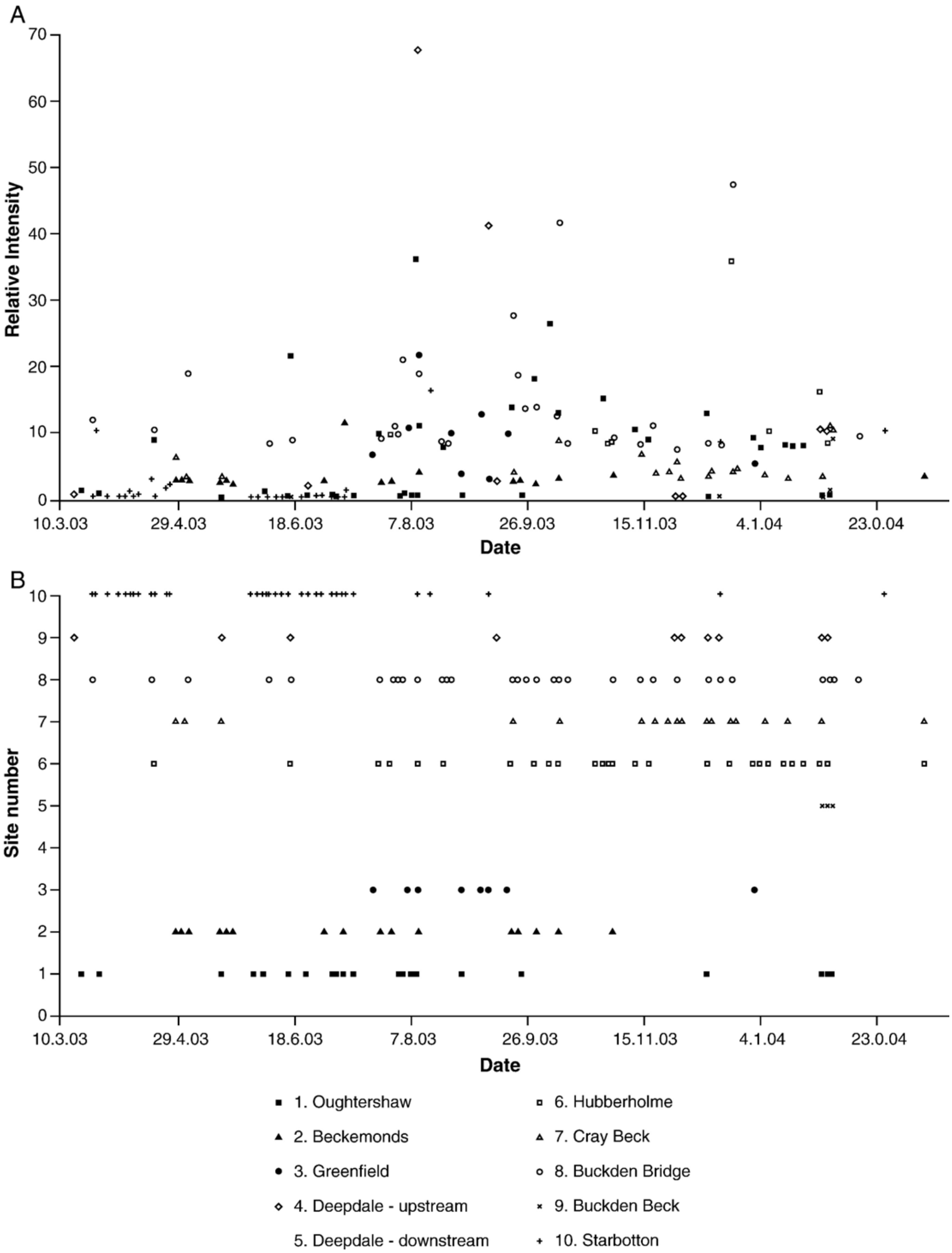

Fig. 7. (A) The relative intensity (magnitude) of distinct sediment transport events at the ten surveyed locations; and (B) the timings of distinct sediment transport events seen at the different locations during the study period.

with some caution, as it will be partly dependent upon the characteristics of the 2003–4 period, the summer 2003–4 period was associated with very few sustained high flows, but a number of short duration high flows linked to convective rainfall events. Thus, the characteristics of the 2003–2004 period allow us to tease out an important observation: the tributaries and the main channel may be responding to different sediment transport controls. The main factors are likely to be: (i) sediment delivery from upstream channel bed, banks and hillslopes; and (ii) channel discharge. The small tributaries, with steep channel slopes have the ability to move sediment very efficiently, especially if sediment is generated quickly from the hillslopes during summer thunderstorms (Reid et al., in press), without the need for progressive catchment wetting: provided the rainfall is sufficient to induce sufficient water table rises to cause hillslopes and channel banks to fail and to deliver sediment to the drainage network, then sediment generation may occur. Clearly, the probability of this will increase as the catchment-averaged saturation deficit falls, because less rain will be required for flows to become extreme. Therefore, the dominant signal appeared to be extreme rainfall events. The main channel sites had much greater volumes of sediment in storage or motion, the sediment was subject to sorting processes, and in combination this means that significantly more energy was required for entrainment and transport, due to low bed gradient and the development of bed sedimentary structures. It was not until the higher winter flows (e.g. 18.9 $m^3 s^{-1}$ on February 3rd, 2004), that there was sufficient energy to transport material through the river system. The 2003–2004 period was associated with flows that were generally below the median and this is likely to cause an imbalance between sediment delivery from tributaries and sediment transfer within the river, with the magnitude and frequency of the latter under-estimated.

Return periods of sediment transport are shown in Table 5. Small events occurred every 0.5 to 2.0 months at Starbotton, with larger events having return periods that filled the record completely (i.e. they were indeterminate with only 12 months of data). The most active sites of Hubberholme and Buckden Bridge showed near identical patterns in the return periods for all event magnitudes, which indicates that these sites may be acting in response to similar processes which govern sediment deposition and transport.

There appeared to be close agreement at Hubberholme between storm events and sediment transport during the autumn and winter months (Fig. 8A), a pattern seen at all the main channel sites. For example, at Starbotton (the least active main channel site), the largest two transport events, in terms of total sediment moved, seen during the study period (August 10th, 2003 and September 10th, 2003) occurred just after the discharge peaks (of 5.0 and 3.6 $m^3 s^{-1}$ respectively), following large prolonged storms (Fig. 8B for the August 10th, 2003 event). These two events accounted for 72% of all activity at Starbotton during the study

Table 5
Return periods (months) for events of different magnitude seen at each site during the study period

Relative intensity	Event size	1. Oughtershaw	2. Beckermonds	3. Greenfield	4. Deepdale upstream	5. Deepdale downstream	6. Hubberholme	7. Cray Beck	8. Buckden Bridge	9. Buckden Beck	10. Starbotton
<0.50	Small	0.75								2.00	0.63
0.51–1.00		0.63								1.50	0.52
1.01–2.00		2.40								4.00	1.71
2.01–5.00		6.00	0.80	6.00				0.80		6.00	2.00
5.01–10.00	Moderate		0.80	2.40	12.00	6.00	0.86	0.63	0.92	12.00	3.00
10.01–25.00			12.00	2.00	4.00	2.40	0.46	3.00	0.43		3.00
25.01–50.00	Large		12.00	4.00	6.00	4.00	0.80		0.67		3.00
50.01–100.00							4.00		4.00		6.00
>100.01											

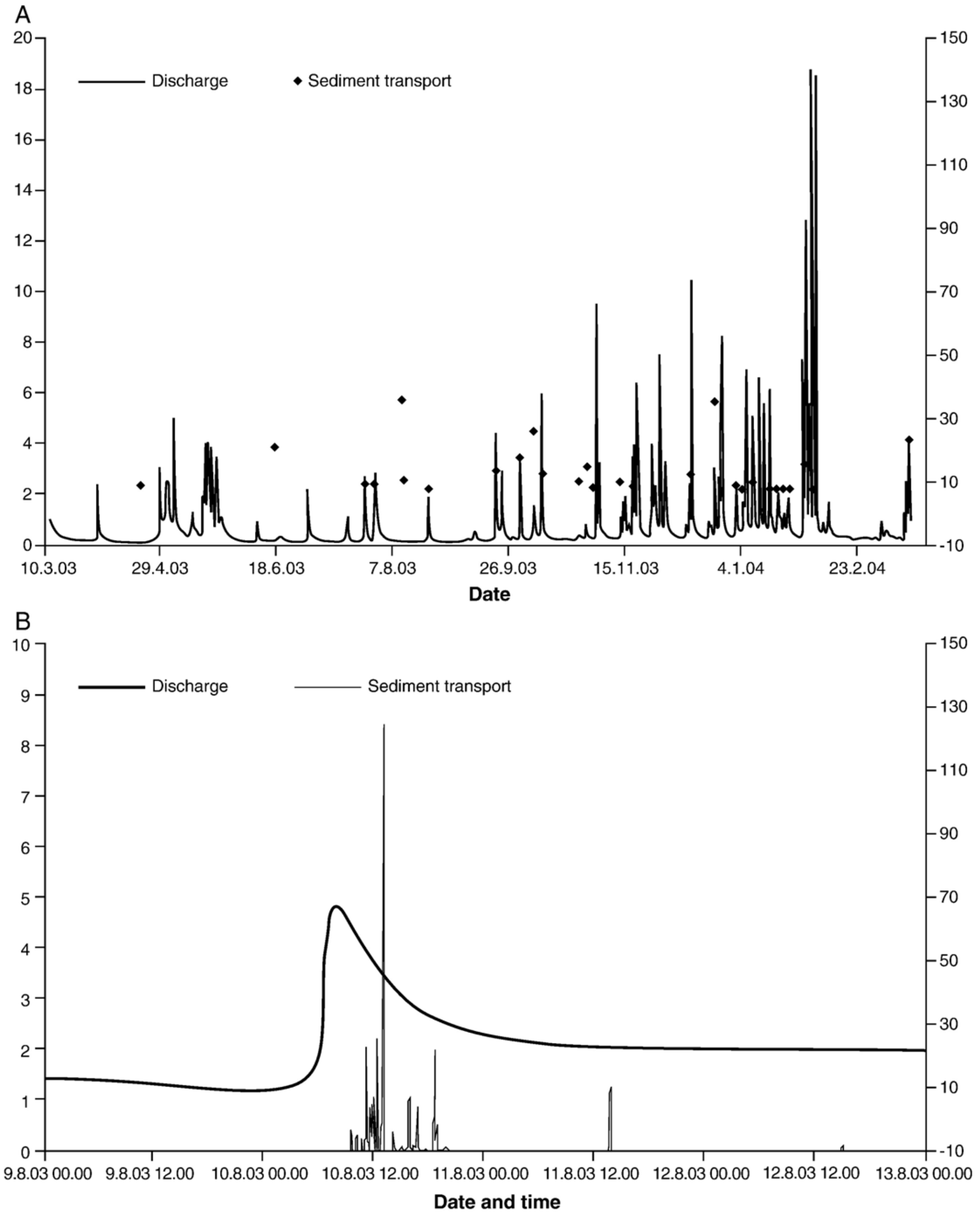

Fig. 8. The relationship between channel discharge and sediment transport at: (A) the most active site at Hubberholme; and (B) the catchment outlet at Starbotton for the event of August 10th, 2003.

year, which suggests that prolonged high discharges may flush out sediment temporarily stored within the sinuous reach upstream. Indeed, the period April to July accounted for just 11.1% of all activity, suggesting autumn and winter dominance in the transport at this site.

4.3. Surveyed river response

Due to the relatively short spacing between cross-sections, volume change data and mean bed levels could be estimated. The volume change data were derived directly from Eq. (5), whereas the mean bed levels were calculated as the end areas divided by the mean bankfull channel width. Data used here relate to: (i) March 2003 to December 2003; and (ii) December 2003 to March 2004 to allow comparison with the hydrological data. Derived changes between cross-sections have been divided by the spacing between the sections to give a volume change per metre of channel. This allows individual reaches to be compared in terms of their activity, irrespective of spacing. The mean bed level approach can be used to suggest zones of erosion and deposition within a larger reach.

Reaches where deposition has occurred are shown by positive values of volume change, with erosion shown by a negative volume change (Table 6). The net volume change (to the nearest m^3) during March to December 2003 (sum of all changes) was 6000 $m^3 \pm 45.9\%$, compared to 14400 $m^3 \pm 45.5\%$ during December 2003 to March 2004. This suggests that both periods were associated with significant deposition within the 5.6 km reach, probably aided by the relatively dry nature of the study period. This is supported by the mean bed level data which show that, for the year, the cross-sections between Hubberholme and Buckden Bridge saw a mean bed level change of 0.080 m ± 38.9%, with a 0.155 m ± 51.6% change between Buckden Bridge and Starbotton. The overall mean change for all sites during the year was 0.108 m ± 64.6%. The winter period (December to March) appeared to be associated with most deposition (71% winter, 29% spring, summer and autumn), as indicated by Fig. 9.

Two active zones of significant accumulation were seen (Fig. 9B). The first was around the site of the former Buckden gravel trap (located at cross-section 100 and removed in June 2002). This reach (cross-sections 070 to 170) had a mean net volume change per metre of channel for the year of 7.0 $m^3 \pm 25.2\%$ and a mean bed level rise of 0.122 m ± 25.2%. The second accumulation zone was where the channel sinuosity significantly increased between cross-sections 430 to 530, where the mean deposition was 0.186 m ± 27.1% and mean net volume change per metre of channel for the year of 4.8 $m^3 \pm 27.1\%$.

This active zone is immediately downstream of the individual reach which had the largest mean uncertainty for both periods (cross-sections 420 to 430) at ±8.7%, with uncertainties of ±9.4% for March 2003 to December 2003 and ±8.1% for December 2003 to March 2004. The two cross-sections within this reach were spaced further apart than most others (Fig. 4), since the channel here was too deep to survey safely, even at low flow. There is a clear increase in uncertainty for both data periods as cross-sections are spaced further apart (Fig. 10). Within this individual reach, estimates of volume change per metre of channel ranged from 0.88 m^2 to 1.07 m^2 (March 2003 to December 2003) and from 5.1 m^2 to 5.3 m^3 (December 2003 to March 2004). This maximum uncertainty for the study is not significant, since the predicted range of values are not outside those observed in the data (minimum during March 2003 to December 2003 was −5.71 m^2; maximum during December 2003 to March 2004 was 9.49 m^2). However, this uncertainty did cause a change in classification of the site in terms of mean bed level during March 2003 to December 2003 (Table 7), with the minimum estimate (−0.067 m) classifying the site as erosional, whereas the maximum estimate (0.121 m) indicating deposition, as shown by all volume change estimates.

The two active zones of significant accumulation (cross-sections 070 to 170 and 430 to 530) are associated with parts of the catchment where increased flooding has been witnessed in recent years (Fig. 11), since the channel volume (i.e. its capacity to hold water) is being dramatically reduced.

During the study there were no significant areas of erosion that were sustained over the year. Taking the net volume change for 2003–2004 as an indication of overall activity, only 9 of the 60 cross-sections showed erosional characteristics. This may be because of the dry nature of the study period coupled with the importance of sustained high flows for main channel sediment transport indicated by the sediment impact sensors. However, research presented elsewhere (Lane et al., in press) has shown that this has been an aggradational reach since at least 2001, with the exception of the period December 2002 to March 2003. Comparison with Flint Mill shows that 2002–2003 was a year in which flows were, throughout, higher than the long-term median, and this serves to emphasise the importance of sustained winter high flows for sediment reworking once delivered from the tributaries.

Within reach-averaged data, three areas of erosion could be seen: (i) cross-sections 170 to 190 (mean net change per metre of channel of −2.2 $m^3 \pm 12.8\%$; mean bed level change of −0.01 m ± 12.8%); (ii) cross-sections 260 to 290 (mean net change per metre of channel of −1.2 $m^3 \pm 13.3\%$; mean bed level change of −0.043 m ± 13.3%); and (iii) cross-sections 380 to 410 (mean net change per metre of channel of −1.8 $m^3 \pm 15.1\%$; mean bed level change of −0.063 m ± 15.1%). These erosional zones were all located on relatively straight sections of the channel, immediately downstream of large accumulation areas. The most erosional individual reach (i.e. between two adjacent cross-

Table 6
Volume and uncertainty estimates for cross-section changes during the study period

Cross-section reach	March 2003 to December 2003			December 2003 to March 2004			Cross-section reach	March 2003 to December 2003			December 2003 to March 2004		
	Volume change (m^3)	Uncertainty (%)	Volume change per metre of channel (m^3)	Volume change (m^3)	Uncertainty (%)	Volume change per metre of channel (m^3)		Volume change (m^3)	Uncertainty (%)	Volume change per metre of channel (m^3)	Volume change (m^3)	Uncertainty (%)	Volume change per metre of channel (m^3)
010 to 020	−95.96	5.1	−1.32	390.11	6.2	5.37	310 to 320	−95.88	5.8	−1.82	189.56	5.1	3.60
020 to 030	486.35	8.4	1.88	1209.13	8.3	4.67	320 to 330	−126.58	6.7	−1.43	369.79	5.1	4.18
030 to 040	1164.86	7.9	5.84	683.52	8.1	3.43	330 to 340	15.37	6.5	0.12	451.54	6.2	3.51
040 to 050	1279.73	8.1	6.03	444.57	8.2	2.10	340 to 350	−358.68	6.4	−2.84	1197.38	6.5	9.49
050 to 060	10.40	5.6	0.30	10.03	4.4	0.29	350 to 360	−377.55	4.8	−4.75	659.29	6.7	8.29
060 to 070	4.80	4.2	0.15	−1.18	4.8	−0.04	360 to 370	−152.36	4.5	−5.71	24.01	4.7	0.90
070 to 080	139.60	4.6	4.95	82.95	4.8	2.94	370 to 380	184.72	5.2	3.25	−7.78	5.8	−0.14
080 to 090	191.74	4.4	4.24	200.88	5.1	4.44	380 to 390	−237.56	6.4	−1.93	−281.07	5.7	−2.29
090 to 100	245.79	5.8	4.06	162.77	5.9	2.69	390 to 400	−142.21	6.3	−1.18	−1.38	5.7	−0.01
100 to 110	172.41	4.4	4.34	92.10	5.5	2.32	400 to 410	−149.49	6.5	−1.45	137.84	6.4	1.34
110 to 120	200.38	4.8	4.62	113.26	5.5	2.61	410 to 420	13.05	5.4	0.39	99.08	4.9	2.98
120 to 130	182.24	4.0	3.47	111.33	5.2	2.12	420 to 430	265.49	9.4	0.98	1422.02	8.1	5.22
130 to 140	133.45	5.5	4.17	73.28	4.2	2.29	430 to 440	2.43	6.0	0.02	722.29	7.0	4.78
140 to 150	352.60	5.3	4.82	202.07	6.0	2.76	440 to 450	233.54	6.2	2.22	300.37	6.9	2.86
150 to 160	569.45	6.8	4.77	353.84	7.1	2.97	450 to 460	156.75	5.9	2.47	228.07	4.8	3.59
160 to 170	646.24	7.8	3.39	316.20	7.8	1.66	460 to 470	31.00	6.5	0.28	472.11	6.0	4.28
170 to 180	−45.76	7.0	−0.33	−41.18	7.3	−0.30	470 to 480	278.65	6.6	2.18	502.53	6.2	3.93
180 to 190	−205.71	4.9	−2.55	−106.15	6.0	−1.32	480 to 490	335.38	7.4	1.82	710.42	6.6	3.86
190 to 200	264.47	6.2	2.26	103.51	5.7	0.89	490 to 500	−10.37	5.2	−0.12	368.70	5.3	4.27
200 to 210	104.21	4.4	3.90	31.91	4.2	1.20	500 to 510	38.73	6.0	0.43	408.03	6.9	4.53
210 to 220	14.74	5.9	0.41	85.71	4.6	2.39	510 to 520	−134.02	5.4	−1.17	612.50	5.3	5.35
220 to 230	−124.93	5.2	−2.44	287.06	4.8	5.60	520 to 530	−106.02	5.4	−1.24	342.58	5.1	4.00
230 to 240	−56.76	4.3	−1.88	133.29	4.6	4.41	530 to 540	59.80	5.8	0.54	98.02	6.1	0.88
240 to 250	85.88	4.9	1.71	12.28	4.7	0.24	540 to 550	109.11	5.6	1.32	66.17	5.8	0.80
250 to 260	196.33	5.5	2.83	−71.43	5.2	−1.03	550 to 560	65.88	6.4	0.58	185.47	6.0	1.64
260 to 270	12.12	5.6	0.33	−35.86	4.6	−0.97	560 to 570	34.90	6.0	0.32	214.13	6.2	1.97
270 to 280	−112.94	6.4	−1.22	−134.65	4.9	−1.45	570 to 580	−2.74	5.9	−0.02	170.68	5.5	1.50
280 to 290	−68.38	5.5	−0.61	22.36	5.5	0.20	580 to 590	154.42	6.7	1.18	−11.21	6.9	−0.09
290 to 300	−37.73	6.7	−0.53	55.56	6.0	0.78	590 to 600	174.04	6.4	1.40	−172.90	6.4	−1.39
300 to 310	30.25	4.1	0.47	111.95	5.1	1.73							

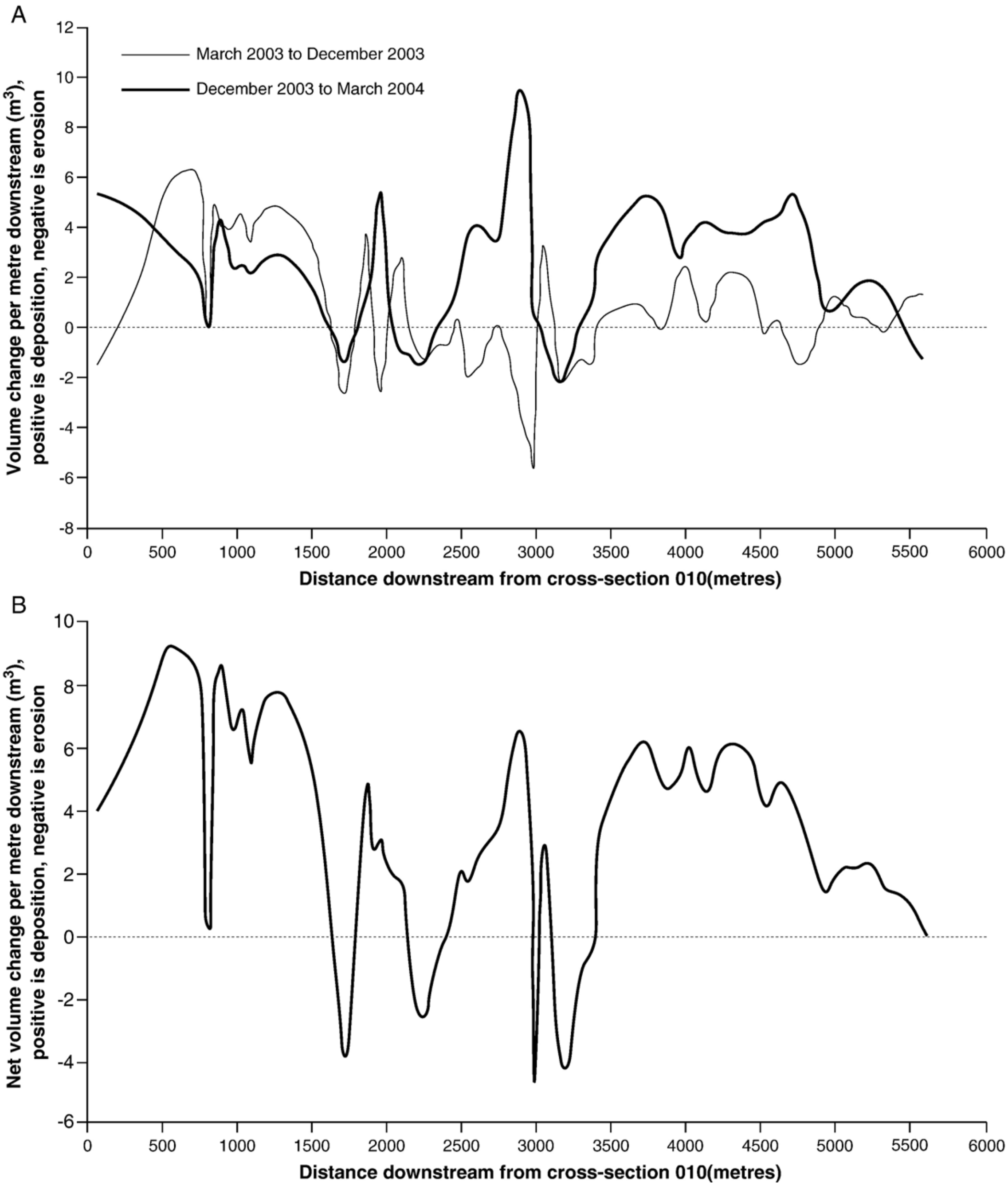

Fig. 9. (A) Volume change for all reaches during the two data periods; and (B) net volume change for all reaches during the study period.

sections) was not located within one of these erosional zones, but between cross-sections 360 and 370 (mean net change per metre of channel of $-4.8\ m^3 \pm 6.5\%$). This was just downstream of the widest area of sediment accumulation (cross-sections 340 to 360) and a significantly depositional individual reach (cross-sections 340 to 350) which had a mean accumulation of $0.256\ m \pm 12.3\%$ during the study year). The channel then narrows and steepens significantly, providing more energy for erosion of the bed and banks, especially due to the residual excess energy the channel now has after all of its transported material has been deposited in the accumulation zone upstream. Indeed, all identified erosional zones appeared to be just downstream of

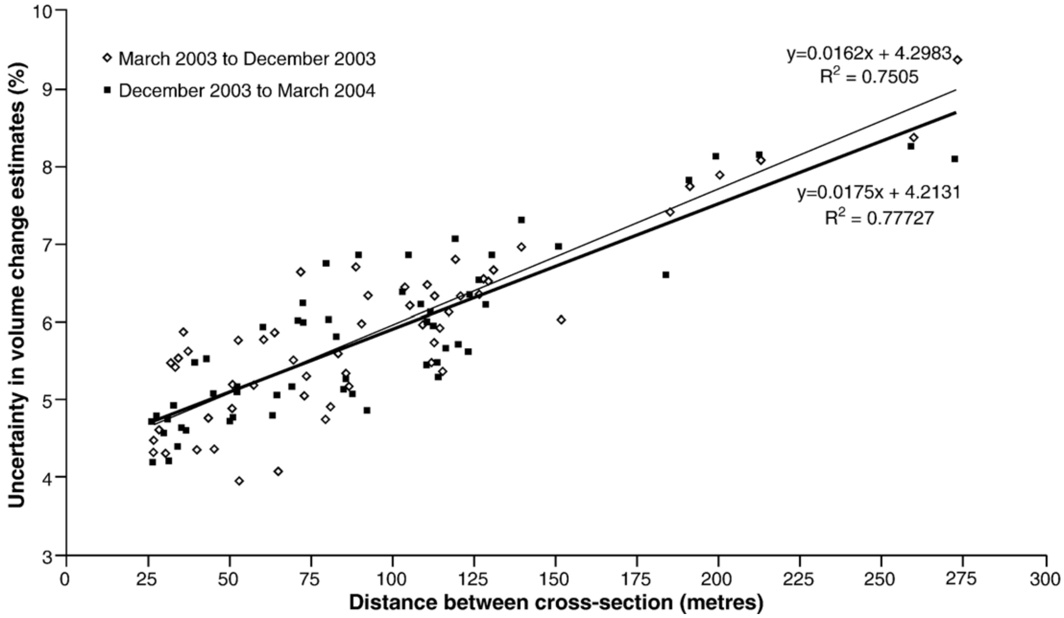

Fig. 10. Uncertainty in volume estimates with spacing between cross-sections.

large accumulation areas. Downstream of this reach, the channel once more widens, has a shallower gradient and an increase in sinuosity, which causes the largest overall accumulation area, in terms of total net volume, between cross sections 410 and 590. This 2.05 km reach had a mean net volume change per metre of channel of 3.8 $m^3 \pm 37.2\%$ and a mean bed level change of 0.124 m $\pm 37.2\%$ along its entire length during the study year, which included the most depositional individual reach within the study with a yearly mean bed level rise of 0.285 m $\pm 9.1\%$ between cross-sections 500 and 510. By contrast, the individual reach containing the impact sensor (cross-sections 590 to 600) had just 0.004 m $\pm 9.0\%$ of accumulation and 0.01 $m^3 \pm 9.0\%$ of net volume change per metre of channel during the year. This was the least amount recorded within the study indicating that sediment was not leaving the system in significant quantities.

Table 7
Estimations with uncertainty for volume change and mean bed level data for the individual reach (cross-sections 420 to 430) with the highest mean uncertainty during the study period

	March 2003 to December 2003	December 2003 to March 2004
Volume change (m^3)	265.49	1422.02
Uncertainty (m^3)	24.95	114.89
Uncertainty (%)	9.4	8.1
Volume change per metre (m^3)	0.97	5.22
Mean bed level change (m)	0.027	0.136
Min. volume change (m^3)	240.54	1307.13
Max. volume change (m^3)	290.44	1536.92
Min. volume change per metre (m^3)	0.88	5.13
Max. volume change per metre (m^3)	1.07	5.32
Min. mean bed level change (m)	−0.067	0.042
Max. mean bed level change (m)	0.121	0.230

4.4. The relationship between sediment transport data and surveyed channel response

The volume of change and mean bed level data support the impact sensor findings which suggest that sediment transport is significantly reduced between Hubberholme and Starbotton, resulting in the development of significant accumulation zones. However, due to the relatively infrequent topographic surveys compared to the frequency of sediment transport events, it was not possible to link individual transport events to surveyed volume change. The surveyed individual reach at Starbotton (cross-sections 590 to 600) had a net volume change per metre of channel of just 0.01 $m^3 \pm 9.0\%$ during the surveyed year. When compared to the mean net volume change per metre of channel for the upstream zone (cross-sections 410 to 590) of 1.9 $m^3 \pm 37.2\%$, this suggests that Starbotton is either a transfer reach or one that receives little sediment from upstream. When combined with the impact sensor

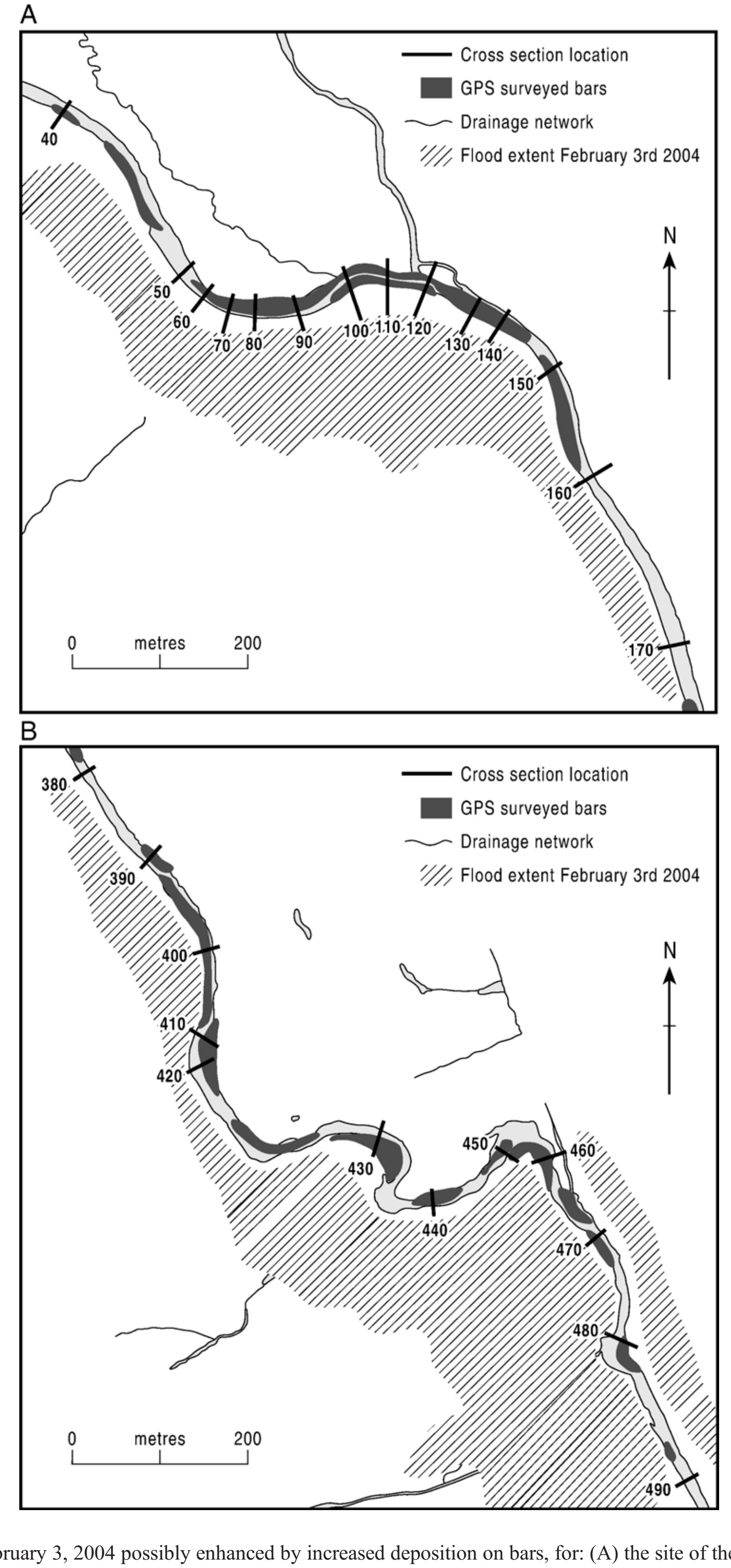

Fig. 11. Flooded areas on February 3, 2004 possibly enhanced by increased deposition on bars, for: (A) the site of the former Buckden Gravel Trap (cross-sections 070 to 170); and (B) cross-sections 430 to 530.

data, which showed very low transport rates, it appears that sediment does not leave the system in large quantities. Even the 2002–2003 year noted above, in which one survey period recorded net degradation, was associated with net aggradation. The surveyed volume changes also match the impact sensor results in suggesting: (i) significant sediment accumulation from Hubberholme to Buckden Bridge; (ii) even greater accumulation from Buckden Bridge to Starbotton; and (iii) that the tributary inputs of Cray Beck (just upstream of cross-section 130) and Buckden Beck (just upstream of cross-section 290) do not significantly impact upon sediment accumulation. Given that sensor impact data may provide an (uncertain) estimate of sediment transport rate, the obvious question is: can we use impact data combined with volume of change data to close the sediment transport budget? If so, are we likely to have measured the system at a resolution sufficient to detect process signature? This is especially important given the better temporal resolution of our impact data and the better spatial resolution of our cross-section data.

5. Discussion

The results suggest that sediment transfer within tributaries generally occurs in significantly smaller magnitudes than within the main channel (total activity of between 90 and 885 in the tributaries and 889 and 29166 in the main channel, Table 2). However, bearing in mind the study year was relatively dry, the number of distinct events during which sediment movement was detected was commonly similar to main channel sites (Table 2). This is reflected in typical event sizes (Table 2), with the tributaries biased towards small and moderate events. Thus, the tributaries act as a series of diffuse coarse sediment sources. In this catchment, they delivered sediment into mainly steep, bedrock channels which do not permit large quantities of generated hillslope sediment to be stored for significant periods of time. The steep slopes allow any generated sediment to be quickly transported to the main channel stem during even relatively small discharges, resulting in a high level of connectivity, as observed by others (e.g. Harvey, 2001). Tributary sediment delivery to the main channel is limited by hillslope supply and the connection of failed sediment to individual tributaries more than it is the transport of sediment in those tributaries. Reid et al. (in press) show that these connected-failures are typically associated with convective rainfall. The latter is more common in the summer, and the tributaries have just 36% of their activity in the winter and 64% in the summer. This reflects other observations that failures can occur following relatively short periods of intense rainfall (e.g. summer thunderstorms) even with dry antecedent conditions (e.g. McEwen and Werritty, 1988). There is clear variation in the relative intensity of sediment delivery between tributaries (Table 2).

The sensor data suggest that Phase II, the initiation and transport of coarse grains (mostly $>D_{16}$), is relatively short, occurring only during parts of larger flood events. This supports the findings of other studies (e.g. Jackson and Beschta, 1982; Beschta, 1987; Ryan et al., 2002). Transport of coarse material within the tributaries has been shown by the impact sensors to sometimes occur during relatively low flows. This is likely to be a result of the low friction of the bedrock channels and their steep slope. A significant proportion of the total sediment transport within these areas may therefore be moved during Phase I (Andrews and Smith, 1992; Wilcock and McArdell, 1993; Lisle, 1995).

The main channel sites were generally associated with larger transport events, with sediment in motion for much longer time periods, two orders of magnitude greater at Hubberholme than in some of the tributaries (Table 2). This was generally not through there being more transport events but through those events being able to sustain transport for longer (Table 2). As the winter was associated with flows able to sustain transport for longer periods, and bearing in mind the dry nature of the study year, it resulted in the main channel sites having 61% of their activity during the winter and 39% during the summer. Thus, the extent to which tributary sediment delivery results in aggradation, and the rate of that aggradation, will be conditioned by the nature of flows in the main channel, which will vary strongly with dominant climatic conditions. The nature of this particular study year aside, sediment generation and transfer within the main channel involves response to different types of events. Sediment generation is linked to extreme flow events that are sufficient to connect hillslope sources to the tributaries and streams. These can occur at any time. However, when these extreme flow events occur in the summer they appear to be of insufficient duration to move large amounts of sediment over long distances within the main river.

Within the main river, sediment transfer is strongly conditioned by larger-scale valley geomorphology. Sediment is quickly moved through the steep bedrock reach to Hubberholme, where a significant reduction in gradient and a widening of the valley results in a reduced transport capacity. The result is exceptionally high levels of sediment activity at Hubberholme (Table 2), in terms of both the time that motion is recorded for as well as the total levels of activity. Between Hubberholme and Buckden Bridge, sediment storage begins and only just over 10% of sediment passing Hubberholme and entering from Cray Beck reaches Buckden Bridge, suggesting that this is a major sedimentation zone. The sensor data are supported by the surveyed channel response which suggests that large amounts of sediment are entering storage within this zone. There are two major sedimentation reaches. The first is close to the site of the former Buckden gravel trap (immediately upstream of the confluence with Cray Beck, Fig. 4). Here, during this study period, the site was rebuilding the mid-channel and point bars evident before the channel was regraded when the gravel trap was removed in June 2002. There is even greater accumulation from Buckden Bridge to Starbotton, with just over 2% of the sediment entering at Hubberholme (including Cray Beck and Buckden Beck) passing Starbotton and 16% of the sediment passing Buckden Bridge (including Buckden Beck) making it to Starbotton. Both of these zones are where main channel coupling has broken down (e.g. Harvey, 2002), the channel is sinuous and actively migrating in response to the asymmetric nature of sediment deposition within individual cross-sections, and this process is probably encouraging these sites to remain as sedimentation zones. Sedimentation in the Hubberholme to Starbotton zone is reflected not only in the measured in-channel sedimentation during the short study period reported here (Fig. 9), but also the management practices adopted over the previous 30 years (e.g. gravel shoal removal from 1985 to 1987, as part of engineering work to reduce flood risk). The reduction in valley slope appears to be compounded by a major historical event within the valley: the Cam Gill catastrophic flood of 1686 (Coulthard et al., 1998; Coulthard, 1999). This generated the decadal sediment discharge from this catchment during just one storm (Coulthard et al., 1998), and resulted in an alluvial fan which extends across the full width of the valley. Although there has been some downcutting by the main River Wharfe into this deposit, it generally acts as a local rise in base level and has created a major sediment discontinuity. The river response to this would be in-channel sedimentation upstream, over the last 350 to 400 years. This would lead to the formation of point bars which over time, and through continued accumulation, would divert water towards the banks. This would lead to increases in channel sinuosity and the transfer of sediment into floodplain storage, through meander migration. One interpretation of this case-study is that it is a 'sedimentary bath tub' conditioned by a combination of the late glacial legacy of deposited till and valley morphology with a major historical event.

The main channel sediment activity generally rises steeply to Hubberholme and then falls rapidly down to Starbotton. The tributaries deliver sediment at variable rates but superimposed onto these main channel activity gradients. For instance, compared with Hubberholme, Buckden Beck is relatively unimportant (Table 2). However, compared with the residual transport after deposition in the reaches between Hubberholme and Buckden Bridge, the contribution from Buckden Beck is significant (about 42% of the material recorded at Buckden Bridge). Thus, in relative terms, the importance of individual tributaries changes as in-channel sedimentation occurs, itself impacted upon by upstream changes. This has implications for how we think about the impact of tributaries upon instream changes in grain-size: tributary impacts will scale with both tributary delivery rates and in channel sediment transfer rates.

These findings have a number of important implications. First, in the short-term, there are projected increases in rainfall which are likely to be most severe during the winter months (Hulme et al., 2002). In relation to sediment delivery, Reid et al. (submitted for publication) found a dramatic increase in the extent to which failed sediment connects with tributaries under these future scenarios, much of this associated with increases in winter activity. Thus, the seasonality of tributary sediment delivery events reported herein may well change, with winter characteristics becoming more dominant. In the main channel, these increases may lead to greater and more prolonged discharges which could permit more main channel transport and reworking of delivered sediment. In our case study, this probably translates into aggradation throughout the reach, as a result of base level control. However, the real issue may not be the effect of climate change upon channel flows, but the effect upon sediment delivery to the system. The greater transport potential of higher flows, coupled with suggested increases in delivery through climate change (Reid et al., submitted for publication), is likely to have a major effect upon accumulation within the lower reach, especially if the base level control at Starbotton continues to have such a major influence upon the sedimentation process. The nature of sediment transfer through the discontinuity at Starbotton appears to be linked to very high magnitude events: 72% of all yearly activity at Starbotton was seen in less than a month (August to September, 2003) during just a few high discharge events. Evacuation of sediment from the Hubberholme to Starbotton reaches may require very long duration flood events that are capable of producing event-by-event step lengths that extend into the reaches downstream of Starbotton. Thus, it is only possible to understand the effects of future climate changes upon sediment delivery and transfer through: (1) an explicit treatment of sediment sources, entrainment, transfer and deposition; coupled to (2) a full consideration of the character of individual climate events; and (3) analysis of the sequencing of events. Each component of the system has a different flow requirement for sediment mobilisation and this needs to be analysed along with changes in the frequency and duration with which that flow is met, in order to determine the impacts of future precipitation changes. This was reflected in the difficulty of identifying a 'critical' rainfall (Table 4) which causes a discharge large enough for the onset of sediment transport: similar rainfall events did not generate the same magnitude discharge events; equal discharges at the same site do not cause the same amount of sediment transport; and a clear relationship between discharge and sediment transport cannot be defined. Other studies (e.g. Laronne and Carson, 1976; Dietrich et al., 1989; Lisle, 1995; Andrews and Nankervis, 1995) also conclude that this relationship is hard to define. This study illustrated how the effectiveness of a given flow event depends upon both its size and duration in relation to whether or not the location under consideration is close to potential hillslope sources, influenced by high rates of sediment delivery from upstream or impacted upon by larger scale attributes of valley morphology.

Second, the results have major implications for how river management projects are developed in upland environments, especially those where large-scale geomorphological controls have a major impact upon the sediment transfer process. Traditionally, river management is based upon stabilisation to protect farmland, properties and footpaths, which prevents transfer of sediment into floodplain storage. Indeed, consultant reports for this reach of river in the late 1990s advocated focusing on stretches where eroding banks needed to be stabilised (e.g. RKL Arup, 1999a,b). This assumes that a channel can be designed to be robust, as with other engineering works (e.g. Gregory and Park, 1974; Petts, 1977; Kellerhals, 1982; Williams and Wolman, 1984; Andrews, 1986; Hey and Winterbottom, 1990; Andrews and Nankervis, 1995; Kondolf, 1997; Gilvear and Bradley, 1997; Werritty and McEwen, 1997; Surian, 1999); and to absorb changes in sediment supply with only minor adjustment in its form. However, Hooke (2003) has suggested that channel morphology and stability in fact reflect the net sediment budget, which in turn reflects: (i) the net erosion and deposition seen; and (ii) the supply and connectivity of sediment from the hillslopes and upstream reaches (e.g. Emmett and Wolman, 2001). Where river management restricts, (a) lateral movement and (b) transfer of sediment into floodplain storage, the channel bed becomes responsive (e.g. Schumm, 1979; Harvey, 2001; Hooke, 2003) with changes in sediment supply (which have not been adequately addressed) leading to areas of severe accumulation. In this river, responsive effects have been surveyed as: (i) acceleration of bank erosion within both zones of significant accumulation as flow is diverted around depositional bars; and (ii) exacerbated flood risk as the conveyance volume of the channel is severely reduced (Fig. 11). Since hydrological data were only available for the year (March 2003 to March 2004), older surveys of channel response have not been used within the analysis. However, inclusion of these data suggests that: (i) there has been a significant increase in accumulation within the 5.6 km reach, with the net volume change of 20400 $m^3 \pm 64.6\%$ for the study year much greater than the 3450 $m^3 \pm 34.2\%$ of March 2002 to March 2003; and (ii) the winter of 2003/2004 (14400 $m^3 \pm 45.5\%$) was the first with no net erosion since surveys began in 2001/2002 (-5300 $m^3 \pm 42.7\%$), possibly due to increased sediment transfer following the removal of the Buckden gravel trap which after its abandonment appeared to be causing a sediment transfer discontinuity. Clearly, this research has benefited from intensive data collection over a short period. In the longer-term, by continuing the repeat monitoring, it should be possible to obtain a more detailed understanding of the relationship between event size, step length and morphological response.

6. Conclusion

This paper has demonstrated the effective use of a new technique for monitoring bedload transport during the complete range of channel discharges and environmental conditions. This allowed distributed data to be intensively collected over a 72 km^2 upland catchment. Results suggest that there are high sediment delivery rates from the tributaries and upper reaches of the catchment. At present, the majority of tributary sediment delivery occurs during the summer months. However, it is not until the higher winter flows that there is sufficient energy to transport material through this part of the catchment, with the majority of main channel transfer occurring during the winter months. Bedload transport patterns within the lower 5.6 km reach show a close relationship to surveyed channel morphological response, with significant accumulation within this reach. High rates of summer supply and severe winter deposition have made many sections of the lower channel unstable. An increase in meandering is being observed at these accumulation sites, with a large increased flood risk, especially during the winter months. This paper has highlighted the importance within river management projects to work with the river, rather than against it. Restricting lateral movement and sediment transfer into floodplain storage may lead to an even more unstable channel, especially in upland areas with high rates of coarse sediment delivery from the headwaters and hillsides.

Acknowledgements

This research was supported by NERC Connect Grant NER/D/S/2000/01269 awarded to SNL, Mike J. Kirkby and Adrian T. McDonald, by Environment Agency R&D award E1-108 awarded to Adrian T. McDonald and SNL, and by the National Trust. Simon C. Reid is funded by a University of Leeds PhD studentship. Joseph Holden was supported by a NERC Research Fellowship NER/I/S/2001/00712. Christine Kuettner helped with the collection of some of the discharge and precipitation data. Adrian Harvey and Malcolm Newson provided very valuable comments on an earlier draft of this manuscript.

References

Andrews, E.D., 1980. Effective and bankfull discharges of streams in the Yampa River basin, Colorado and Wyoming. Journal of Hydrology 46, 311–330.

Andrews, E.D., 1984. Bed-material entrainment and hydraulic geometry of gravel-bed rivers in Colorado. Geological Society of America Bulletin 95, 371–378.

Andrews, E.D., 1986. Downstream effects of Flaming Gorge Reservoir on the Green River, Colorado and Utah. Geological Society of America Bulletin 97, 1012–1023.

Andrews, E.D., Erman, D.C., 1986. Persistence in the size distribution of surficial bed material during an extreme snowmelt flood. Water Resources Research 22, 191–197.

Andrews, E.D., Nankervis, J.M., 1995. Effective discharge and the design of channel maintenance flows for gravel-bed rivers. In: Costa, J.E., Miller, A.J., Potter, K.W., Wilcock, P.R. (Eds.), Natural and Anthropogenic Influences in Fluvial Geomorphology: The Wolman Volume. American Geophysical Union, pp. 151–164.

Andrews, E.D., Smith, J.D., 1992. A theoretical model for calculating marginal bedload transport rates of gravel. In: Billi, P., Hey, R.D., Thorne, C.R., Tacconi, P. (Eds.), Dynamics of Gravel-Bed Rivers. Wiley, Chichester, pp. 41–48.

Ashmore, P., Church, M., 1998. Sediment transport and river morphology: a paradigm for study. In: Klingeman, P.C., Beschta, R.L., Komar, P.D., Bradley, J.B. (Eds.), Gravel-Bed Rivers in the Environment. Water Resources Publications, Colorado, pp. 115–148.

Ashworth, P.J., Ferguson, R.I., 1989. Size-selective entrainment of bed load in gravel bed streams. Water Resources Research 25, 627–634.

Baker, V.R., Ritter, D.F., 1975. Competence of rivers to transport coarse bedload material. Geological Society of America Bulletin 86, 975–978.

Bathurst, J.C., Graf, W.H., Cao, H.H., 1983. Initiation of sediment transport in steep channels with coarse bed material. In: Sumer, B. M., Muller, A. (Eds.), Mechanisms of Sediment Transport. Balkema, Rotterdam, pp. 207–213.

Beschta, R.L., 1987. Conceptual models of sediment transport in streams. In: Thorne, C.R., Bathurst, J.C., Hey, R.D. (Eds.), Sediment Transport in Gravel Bed Rivers. Wiley, Chichester, pp. 387–408.

Brasington, J., Rumsby, B.T., McVey, R.A., 2000. Monitoring and modelling morphological change in a braided gravel-bed river using high resolution GPS-based survey. Earth Surface Processes and Landforms 25, 973–990.

Brayshaw, A.C., Frostick, L.E., Reid, I., 1983. The hydrodynamics of particle clusters and sediment entrainment in coarse alluvial channels. Sedimentology 30, 137–143.

Brewer, P.A., Passmore, D.G., 2002. Sediment budgeting techniques in gravel bed rivers. In: Jones, S., Frostick, L.E. (Eds.), Sediment Flux to Basins: Causes, Controls and Consequences. Special Publication, vol. 191. Geological Society, London, pp. 97–113.

Brooks, S.M., Richards, K.S., Anderson, M.G., 1993. Approaches to the study of hillslope development due to mass movement. Progress in Physical Geography 17, 32–49.

Brooks, S.M., Crozier, M.J., Glade, T.W., Anderson, M.G., 2004. Towards establishing climatic thresholds for slope instability: use of a physically-based combined soil hydrology-slope stability model. Pure and Applied Geophysics 161, 881–905.

Buchanan, T.J., Somers, W.P., 1969. Discharge measurements at gauging stations, USGS, Techniques of Water Resources Investigations, Book 3, Chapter A8.

Buffington, J.M., Dietrich, W.E., Kirchner, J.W., 1992. Friction angle measurements on a naturally formed gravel stream bed: implications for critical boundary shear stress. Water Resources Research 28, 411–425.

Bunte, K., 1996. Analysis of the Temporal Variation of Coarse Bedload Transport and its Grain Size Distribution, General Technical Report RM-GTR-288, USDA Forest Service, Fort Collins.

Campbell, R.H., 1975. Soil slips, debris flows and rainstorm in the Santa Monica Mountains and vicinity, southern California. USGS Professional Paper, p. 851.

Carling, P.A., 1988. The concept of dominant discharge applied to two gravel-bed streams in relation to channel stability thresholds. Earth Surface Processes and Landforms 13, 355–367.

Carling, P.A., Reader, N.A., 1982. Structure, composition and bulk properties of upland stream gravels. Earth Surface Processes and Landforms 7, 349–365.

Carling, P.A., Benson, I., Richardson, K., 2002. A new instrument to record sediment movement in bedrock channels. ICCE/IAHS: Erosion and Sediment Transport Measurement: Technological and Methodological Advances, Oslo Workshop.

CEH Wallingford, 2003. Monthly River Flow and Rainfall Archives, available online: www.nwl.ac.uk/ih/nrfa/monthly_summaries/archive.html.

Chappell, A., Heritage, G.L., Fuller, I.C., Large, A.R.G., Milan, D.J., 2003. Geostatistical analysis of ground-survey elevation data to elucidate spatial and temporal river channel change. Earth Surface Processes and Landforms 28, 349–370.

Chin, C.O., Melville, B.W., Raudkivi, A.J., 1994. Streambed armouring. Journal of Hydraulic Engineering 120, 899–917.

Church, M., Hassan, M.A., 1992. Size and distance of travel of unconstrained clasts on a streambed. Water Resources Research 28, 299–303.

Church, M., Kellerhals, R., 1978. On the statistics of grain size variation along a gravel river. Canadian Journal of Earth Sciences 15, 1151–1160.

Church, M., McLean, D.G., Wolcott, J.F., 1987. River bed gravels: sampling and analysis. In: Thorne, C.R., Bathurst, J.C., Hey, R.D. (Eds.), Sediment Transport in Gravel Bed Rivers. Wiley, Chichester, pp. 269–325.

Church, M., Wolcott, J.F., Fletcher, W.K., 1991. A test of equal mobility in fluvial sediment transport: behaviour of the sand fraction. Water Resources Research 27, 2941–2951.

Coulthard, T.J., 1999. Modelling Upland Catchment Response to Holocene Environmental Change, Unpublished PhD dissertation, University of Leeds, U.K.

Coulthard, T.J., Kirkby, M.J., Macklin, M.G., 1998. Modelling the 1686 flood of Cam Gill Beck, Starbotton, Upper Wharfedale. In: Howard, A., Macklin, M.G. (Eds.), The Quaternary of the Eastern Yorkshire Dales: Field Guide. Quaternary Research Association, London.

Custer, S.G., Bunte, K., Spieker, R., Ergenzinger, P., 1986. Timing and location of coarse bedload transport: Squaw Creek, Montana. Transactions-American Geophysical Union 67, 943–945.

Custer, S.G., Ergenzinger, P., Bugosh, N., Anderson, B.C., 1987. Electromagnetic detection of pebble transport in streams: a method for measurement of sediment transport waves. In: Ethridge, F., Flores, R. (Eds.), Recent Development in Fluvial Sedimentology. Society of Palaeontologists and Mineralogists, vol. 39, pp. 21–26.

Davies, T.H.R., 1987. Problems of bedload transport in braided gravel-bed rivers. In: Thorne, C.R., Bathurst, J.C., Hey, R.D. (Eds.), Sediment Transport in Gravel Bed Rivers. Wiley, Chichester, pp. 793–828.

Dhakal, A.S., Sidle, R.C., 2003. Long-term modelling of landslides for different forest management practices. Earth Surface Processes and Landforms 28, 853–868.

Dietrich, W.E., Dunne, T., Humphrey, N.F., Reid, L.M., 1982. Construction of sediment budgets for drainage basins. In: Swanson, F.J., Janda, R.J., Dunne, T., Swanston, D.N. (Eds.), Forested Drainage Basins. U.S. Department of Agriculture, pp. 5–23.

Dietrich, W.E., Kirchner, J.W., Ikeda, H., Iseya, F., 1989. Sediment supply and the development of the coarse surface layer in gravel-bedded rivers. Nature 340, 215–217.

Drake, T.G., Shreve, R.L., Dietrich, W.E., Whiting, P.J., Leopold, L.B., 1988. Bedload transport of fine gravel observed by motion picture photography. Journal of Fluid Mechanics 192, 193–217.

Duizendstra, H.D., 2001. Determination of the sediment transport in an armoured gravel-bed river. Earth Surface Processes and Landforms 26, 1381–1393.

Eaton, B.C., Lapointe, M.F., 2001. Effects of large floods on sediment transport and reach morphology in the cobble-bed Sainte Marguerite River. Geomorphology 40, 291–309.

Einstein, H.A., 1937. Bedload transport as a probability problem. In: Shen, H.W. (Ed.), Sedimentation (reprinted in 1972). Water Resources Publications, Colorado, pp. 105–108.

Emmett, W.W., 1980. A field calibration of the sediment trapping characteristics of the Helley-Smith bed load sampler. USGS Professional Paper, vol. 1139.

Emmett, W.W., Wolman, M.G., 2001. Effective discharge and gravel-bed rivers. Earth Surface Processes and Landforms 26, 1369–1380.

Ergenzinger, P., Conrady, J., 1982. A new tracer technique for measuring bedload in natural channels. Catena 9, 76–93.

Ergenzinger, P., Custer, S., 1983. First experiences measuring coarse bedload material transport with a magnetic device. In: Sumer, B.M., Muller, A. (Eds.), Mechanics of Sediment Transport, Proceedings of Euromech, Istanbul, vol. 156, pp. 223–227.

Ferguson, R.I., Ashworth, P.J., 1991. Slope-induced changes in channel character along a gravel-bed stream: the Allt Dubhaig, Scotland. Earth Surface Processes and Landforms 16, 65–82.

Ferguson, R.I., Ashworth, P.J., 1992. Spatial patterns of bedload transport and channel change in braided and near-braided rivers. In: Billi, P., Hey, R.D., Thorne, C.R., Tacconi, P. (Eds.), Dynamics of Gravel-Bed Rivers. Wiley, Chichester, pp. 477–495.

Ferguson, R.I., Wathen, S.J., 1998. Tracer-pebble movement along a concave river profile: virtual velocity in relation to grain size and shear stress. Water Resources Research 34, 2031–2038.

Ferguson, R.I., Ashmore, P.E., Ashworth, P.J., Paola, C., Prestegaard, K.L., 1992. Measurements in a braided river chute and lobe: I. Flow pattern, sediment transport and channel change. Water Resources Research 28, 1877–1886.

Fuller, I.C., Passmore, D.G., Heritage, G.L., Large, A.R.G., Milan, D.J., Brewer, P.A., 2002. Annual sediment budgets in an unstable gravel bed river: the River Coquet, northern England. In: Jones, S., Frostick, L.E. (Eds.), Sediment Flux to Basins: Causes, Controls and Consequences. Special Publication, vol. 191. Geological Society, London, pp. 115–131.

Fuller, I.C., Large, A.R.G., Charlton, M.E., Heritage, G.L., Milan, D.J., 2003. Reach-scale sediment transfers: an evaluation of two morphological budgeting approaches. Earth Surface Processes and Landforms 28, 889–903.

Gilvear, D., Bradley, S., 1997. Geomorphological adjustment of a newly engineered upland sinuous gravel-bed river diversion: Evan Water, Scotland. Regulated Rivers: Research and Management 13, 377–389.

Goff, J.R., Ashmore, P., 1994. Gravel transport and morphological change in the braided Sunwapta River, Alberta, Canada. Earth Surface Processes and Landforms 19, 195–212.

Gomez, B., 1983. Temporal variations in the particle size distribution of surficial bed material: the effect of progressive bed armouring. Geografiska Annaler 65, 183–192.

Gregory, K.J., Park, C., 1974. Adjustment of river channel capacity downstream from a reservoir. Water Resources Research 10, 870–873.

Habersack, H.M., 2001. Radio-tracking gravel particles in a large braided river in New Zealand: a field test of the stochastic theory of bed load transport proposed by Einstein. Hydrological Processes 15, 377–391.

Ham, D.G., Church, M., 2000. Bed material transport estimated from channel morphodynamics: Chilliwack River, British Columbia. Earth Surface Processes and Landforms 25, 1123–1142.

Harvey, A.M., 1974. Gully erosion and sediment yield in the Howgill Fells, Westmorland. In: Gregory, K.J., Walling, D.E. (Eds.), Fluvial Processes in Instrumented Watersheds. Institute of British Geographers. Special Publication, vol. 6, pp. 45–58.

Harvey, A.M., 1987. Sediment supply to upland streams, influence on channel adjustment. In: Thorne, C.R., Bathurst, J.C., Hey, R.D.

(Eds.), Sediment Transport in Gravel Bed Rivers. Wiley, Chichester, pp. 121–150.

Harvey, A.M., 1997a. The role of alluvial fans in arid zone fluvial systems, In: Thomas, D.S.G. (Ed.), Arid Zone Geomorphology: Process, Form and Change in Drylands, 2nd edition. Wiley, Chichester, pp. 231–259.

Harvey, A.M., 1997b. Coupling between hillslope gully systems and stream channels in the Howgill Fells, northwest England: temporal implications. Geomorphologie: Relief, Processus et Environnement 1, 3–20.

Harvey, A.M., 2000. Coupling within fluvial geomorphic systems: spatial and temporal implications. Journal of China University of Geosciences 11, 9–27.

Harvey, A.M., 2001. Coupling between hillslopes and channels in upland fluvial systems: implications for landscape sensitivity, illustrated from the Howgill Fells, northwest England. Catena 42, 225–250.

Harvey, A.M., 2002. Effective timescales of coupling within fluvial systems. Geomorphology 44, 175–201.

Haschenburger, J.K., Church, M., 1998. Bed material transport estimated form the virtual velocity of sediment. Earth Surface Processes and Landforms 23, 791–808.

Hassan, M.A., Church, M., 1992. The movement of individual grains on the streambed. In: Billi, P., Hey, R.D., Tacconi, P., Thorne, C.R. (Eds.), Dynamics of Gravel-Bed Rivers. Wiley, Chichester, pp. 159–175.

Hassan, M.A., Reid, I., 1990. The influence of microform bed roughness elements on flow and sediment transport in gravel bed rivers. Earth Surface Processes and Landforms 15, 739–750.

Hassan, M.A., Church, M., Schick, A.P., 1991. Distance of movement of coarse particles in gravel bed streams. Water Resources Research 27, 503–511.

Hassan, M.A., Church, M., Ashworth, P.J., 1992. Virtual rate and mean distance of travel of individual clasts in gravel-bed channels. Earth Surface Processes and Landforms 17, 617–627.

Hearn, G.J., Griffiths, J.S., 2001. Landslide hazard mapping and risk assessment. In: Griffiths, J.S. (Ed.), Land Surface Evaluation for Engineering Practice. Geological Society Engineering Geology Special Publication, vol. 18, pp. 43–52.

Helley, E.J., Smith, W., 1971. Development and calibration of a pressure difference bedload sampler. USGS Water Resources Division Report.

Heritage, G.L., Newson, M.D., 1997. Geomorphological Audit of the Upper Wharfe, University of Newcastle upon Tyne.

Hey, R.D., Winterbottom, A.N., 1990. River Engineering in National Parks: the case of the River Wharfe, U.K. Regulated Rivers: Research and Management 5, 35–44.

Hollingshead, A.B., 1971. Sediment transport measurements in a gravel river. Journal of the Hydraulics Division, ASCE 97, 1817–1834.

Hooke, J.M., 2003. Coarse sediment connectivity in river channel systems: a conceptual framework and methodology. Geomorphology 56, 79–94.

Hubbell, D., 1987. Bed load sampling and analysis. In: Thorne, C.R., Bathurst, J.C., Hey, R.D. (Eds.), Sediment Transport in Gravel Bed Rivers. Wiley, Chichester, pp. 89–118.

Hulme, M., Jenkins, G.J., Lu, X., Turnpenny, J.R., Mitchell, T.D., Jones, R.G., Lowe, J., Murphy, J.M., Hassell, D., Boorman, P., McDonald, R., Hill, S., 2002. Climate Change Scenarios for the United Kingdom: The UKCIP02 Scientific Report. Tyndall Centre for Climate Change Research, School of Environmental Sciences, University of East Anglia, Norwich, UK.

Jackson, W.L., Beschta, R.L., 1982. A model of two-phase bedload transport in an Oregon Coast Range Stream. Earth Surface Processes and Landforms 7, 517–527.

Kellerhals, R., 1982. Effect of river regulation on channel stability. In: Hey, R.D., Bathurst, J.C., Thorne, C.R. (Eds.), Gravel-Bed Rivers: Fluvial Processes, Engineering and Management. Wiley, Chichester, pp. 685–705.

Kirchner, J.W., Dietrich, W.E., Iseya, F., Ikeda, H., 1990. The variability of critical shear stress, friction angle and grain protrusion in water worked sediments. Sedimentology 37, 647–672.

Kondolf, G.M., 1997. Hungry water: effects of dams and gravel mining on river channels. Environmental Management 21, 533–551.

Lane, S.N., 1998. The use of digital terrain modelling in the understanding of dynamic river channel systems. In: Lane, S.N., Richards, K.S., Chandler, J.H. (Eds.), Landform Monitoring, Modelling and Analysis. Wiley, Chichester, pp. 311–342.

Lane, S.N., 2001. The measurement of gravel-bed river morphology. In: Mosley, M.P. (Ed.), Gravel-Bed Rivers V. New Zealand Hydrological Society, Wellington, pp. 291–320.

Lane, S.N., Chandler, J.H., Richards, K.S., 1994. Developments in monitoring and modelling small-scale river bed topography. Earth Surface Processes and Landforms 19, 349–368.

Lane, S.N., Richards, K.S., Chandler, J.H., 1995. Morphological estimation of the time-integrated bedload transport rate. Water Resources Research 31, 761–772.

Lane, S.N., Westaway, R.M., Hicks, D.M., 2003. Estimation of erosion and deposition volumes in a large, gravel-bed, braided river using synoptic remote sensing. Earth Surface Processes and Landforms 28, 249–271.

Lane, S.N., Tayefi, V., Reid, S.C., Yu, D. and Hardy, R.J., in press. Interactions between Sediment Delivery, Channel Change, Climate Change and Flood Risk. Paper forthcoming in Earth Surface Processes and Landforms.

Laronne, J.B., Carson, M.A., 1976. Interrelationships between bed morphology and bed material transport for a small gravel-bed channel. Sedimentology 23, 67–85.

Leopold, L.B., Emmett, W.W., Myrick, R.M., 1966. Channel and Hillslope Processes in Semi-arid Areas, New Mexico, USGS Professional Paper, 353-G.

Lindsay, J.B., Ashmore, P.E., 2002. The effects of survey frequency on estimates of scour and fill in a braided river model. Earth Surface Processes and Landforms 27, 27–43.

Lisle, T.E., 1995. Particle size variations between bed load and bed material in natural gravel bed channels. Water Resources Research 31, 1107–1118.

Mantz, P.A., 1980. Low sediment transport rates over flat beds. Journal of the Hydraulics Division, ASCE 106, 1173–1190.

Martin, Y., Church, M., 1995. Bed-material transport estimated from channel surveys: Vedder River, British Columbia. Earth Surface Processes and Landforms 20, 347–361.

McEwen, L.J., Werritty, A., 1988. The hydrology and long-term geomorphic significance of a flash flood in the Cairngorm Mountains, Scotland. Catena 15, 361–377.

McLean, D.G., Church, M., 1999. Sediment transport along lower Fraser River 2: estimates based on the long-term gravel budget. Water Resources Research 35, 2549–2559.

Merrett, S.P., Macklin, M.G., 1999. Historic river response to extreme flooding in the Yorkshire Dales, Northern England. In: Brown, A.G., Quine, T.A. (Eds.), Fluvial Processes and Environmental Change. Wiley, Chichester, pp. 345–360.

Milne, J.A., Sear, D.A., 1997. Modelling river channel topography using GIS. International Journal of Geographical Information Science 11, 499–519.
Montgomery, D.R., Dietrich, W.E., 1994. A physically based model for the topographic control on shallow landsliding. Water Resources Research 30, 1153–1171.
Montgomery, D.R., Dietrich, W.E., Torres, R., Anderson, S.P., Heffner, J.T., Loague, K., 1997. Hydrologic response of a steep unchanneled valley to natural and applied rainfall. Water Resources Research 33, 91–109.
Montgomery, D.R., Dietrich, W.E., Heffner, J.T., 2002. Piezometric response in shallow bedrock at CB1: implications for runoff generation and landsliding. Water Resources Research 38, 1274–1292.
Naden, P., Brayshaw, A.C., 1987. Bedforms in gravel-bed rivers. In: Richards, K.S. (Ed.), River Channels–Environment and Process. Blackwells, Oxford, pp. 249–271.
Neill, C.R., 1987. Sediment balance considerations linking long-term transport and channel processes. In: Thorne, C.R., Bathurst, J.C., Hey, R.D. (Eds.), Sediment Transport in Gravel Bed Rivers. Wiley, Chichester, pp. 225–239.
Nolan, K.M., Shields, R.R., 2000. Measurement of stream discharge by wading, USGS, Water Resources Investigation Report 00-4036.
Paige, A.D., Hickin, E.J., 2000. Annual bed-elevation regime in the alluvial channel of Squamish River, southwestern British Columbia, Canada. Earth Surface Processes and Landforms 25, 991–1009.
Parker, G., 1979. Hydraulic geometry of active gravel rivers. Journal of the Hydraulics Division, ASCE 105, 1185–1201.
Parker, G., Klingeman, P.C., 1982. On why gravel bed streams are paved. Water Resources Research 18, 1409–1423.
Parker, G., Klingeman, P.C., McLean, D.G., 1982. Bedload size and distribution in a paved gravel-bed stream. Journal of the Hydraulics Division, ASCE 108, 544–571.
Petts, G.E., 1977. Channel response to flow regulation: the case of the River Derwent, Derbyshire. In: Gregory, K.J. (Ed.), River Channel Changes. Wiley, Chichester, pp. 145–164.
Powell, D.M., 1992. Bedload Entrainment, Transport and Deposition in Braided Reaches, Unpublished PhD dissertation, University of Leeds, UK.
Powell, D.M., Ashworth, P.J., 1995. Spatial pattern of flow competence and bedload transport in a divided gravel bed river. Water Resources Research 31, 741–752.
Pyrce, R.S., Ashmore, P.E., 2003. Particle path length distributions in meandering gravel-bed streams: results from physical models. Earth Surface Processes and Landforms 28, 951–966.
Reid, I., Frostick, F.E., 1984. Particle interaction and its effects on the thresholds of initial and final bedload motion in coarse alluvial channels. In: Koster, E.H., Steel, R.J. (Eds.), Sedimentology of Gravels and Conglomerates. Canadian Society of Petroleum Geologists, Calgary, pp. 61–68.
Reid, I., Brayshaw, A.C., Frostick, L.E., 1984. An electromagnetic device for automatic detection of bedload motion and its field applications. Sedimentology 31, 269–276.
Reid, I., Frostick, F.E., Layman, J.T., 1985. The incidence and nature of bedload transport during flood flows in coarse grained alluvial channels. Earth Surface Processes and Landforms 10, 33–44.
Reid, I., Frostick, L.E., Brayshaw, A.C., 1992. Microform roughness elements and the selective entrainment and entrapment of particles in gravel-bed rivers. In: Billi, P., Hey, R.D., Thorne, C.R., Tacconi, P. (Eds.), Dynamics of Gravel-Bed Rivers. Wiley, Chichester, pp. 188–205.
Reid, S.C., Lane, S.N., Montgomery, D.R., Brookes, C.J., in press. Does hydrological connectivity improve identification of coarse sediment delivery in upland environments? Paper forthcoming Geomorphology.
Reid, S.C., Lane, S.N., Dugdale, L.J., Montgomery, D.R., Brookes, C.J., submitted for publication. Climate change effects upon coarse sediment sources and delivery within an upland catchment, paper submitted to Earth Surface Processes and Landforms.
Rice, S., Church, M., 1998. Grain size along two gravel-bed rivers: statistical variation, spatial pattern and sedimentary links. Earth Surface Processes and Landforms 23, 345–363.
RKL Arup, 1999a. Dynamic Assessment of Unstable Reaches of the Upper Wharfe, Environment Agency Final Report.
RKL Arup, 1999b. Upper Wharfedale “Best Practice” Project - Erosion Risk Assessment, Environment Agency Report.
Ryan, S.E., Porth, L.S., Troendle, C.A., 2002. Defining phases of bedload transport using piecewise regression. Earth Surface Processes and Landforms 27, 971–990.
Schumm, S.A., 1979. Geomorphic thresholds: the concept and its applications. Transactions of the Institute of British Geographers 4, 485–515.
Sear, D.A., Lee, M.W.E., Oakey, R.J., Carling, P.A., Collins, M.B., 2000. Coarse sediment tracing technology in littoral and fluvial environments: a review. In: Foster, I.D.L. (Ed.), Tracers in Geomorphology. Wiley, Chichester, pp. 21–56.
Spieker, R., Ergenzinger, P., 1987. New developments in measuring bedload by the magnetic tracer technique. Erosion, Transport and Deposition Processes. IAHS Publication, vol. 189, pp. 169–178.
Stojic, M., Chandler, J.H., Ashmore, P., Luce, J., 1998. The assessment of sediment transport rates by automated digital photogrammetry. Photogrammetric Engineering and Remote Sensing 64, 387–395.
Stott, T., Sawyer, A., 2000. Clast travel distances and abrasion rates in two coarse upland channels determined using magnetically tagged bedload. In: Foster, I.D.L. (Ed.), Tracers in Geomorphology. Wiley, Chichester, pp. 389–400.
Surian, N., 1999. Channel changes due to river regulation: the case of the Piave River, Italy. Earth Surface Processes and Landforms 24, 1135–1151.
Taylor, J.R., 1997. An Introduction to Error Analysis: The Study of Uncertainties in Physical Measurements, 2nd edition. University Science Books.
Tunnicliffe, J., Gottesfeld, A.S., Mohamed, M., 2000. High resolution measurement of bedload transport. Hydrological Processes 14, 2631–2643.
Warburton, J., 1992. Observations of bedload transport and channel bed changes in a proglacial mountain stream. Arctic and Alpine Research 24, 195–203.
Warburton, J., Davies, T.R.H., Mandl, M.G., 1993. A meso-scale field investigation of channel change and floodplain characteristics in an upland braided gravel-bed river, New Zealand. In: Best, J.L., Bristow, C.S. (Eds.), Braided Rivers. Geological Society Special Publication, vol. 75, pp. 73–87.
Wathen, S.J., Hoey, T.B., Werritty, A., 1997. Quantitative determination of the activity of within-reach sediment storage in a small gravel-bed river using transit time and response time. Geomorphology 20, 113–134.
Werritty, A., McEwen, L.J., 1997. The fluvial geomorphology of Scotland. In: Gregory, K.J. (Ed.), Fluvial Geomorphology. Chapman and Hall, London, pp. 21–32.
Westaway, R.M., Lane, S.N., Hicks, D.M., 2000. The development of an automated correction procedure for digital photogrammetry for

the study of wide, shallow, gravel-bed rivers. Earth Surface Processes and Landforms 25, 209–226.

Wilcock, P.R., 1997. Entrainment, displacement and transport of tracer gravels. Earth Surface Processes and Landforms 22, 1125–1138.

Wilcock, P.R., 2001. Toward a practical method for estimating sediment-transport rates in gravel-bed rivers. Earth Surface Processes and Landforms 26, 1395–1408.

Wilcock, P.R., McArdell, B.W., 1993. Surface based fractional transport rates: mobilisation thresholds and partial transport of sand-gravel sediment. Water Resources Research 29, 1297–1312.

Wilcock, P.R., Southard, J.B., 1989. Bed load transport of mixed size sediment: fractional transport rates, bed forms and the development of a coarse bed surface layer. Water Resources Research 25, 1629–1641.

Williams, G.P., Wolman, M.G., 1984. Downstream effects of dams on alluvial rivers. USGS Professional Paper, p. 1286.

Wittenberg, L., 2002. Structural patterns in coarse gravel river beds: typology, survey and assessment of the roles of grain size and river regime. Geografiska Annaler 84, 25–37.

Wolman, M.G., Miller, J.P., 1960. Magnitude and frequency of forces in geomorphic processes. Journal of Geology 68, 54–74.

Zevenbergen, L.W., Thorne, C.R., 1987. Quantitative analysis of land surface topography. Earth Surface Processes and Landforms 12, 47–56.