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BEDLOAD TRANSPORT PROCESSES IN A CHUTE-BAR UNIT OF A BRAIDED RIVER

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Abstract

Boundary shear stress and bedload fields were intensively mapped in a chute-bar unit of a proglacial braided river on four days during meltwater floods. Shear stress is highly variable reflecting the non-uniform flow associated with flow convergence and divergence and an irregular channel cross-section. Bedload transport rates vary systematically through the reach decreasing laterally from the talweg and downstream from the chute. This pattern is consistent from day to day despite changes in the amount of sediment transported. The poor correlation between bedload transport and shear stress can be explained by short term variations in sediment supply connected with channel changes, avulsions and the storage and release of sediment held in bars. Comparisons of the spatial distribution of bedload median grain size and maximum particle size shows that generally the same bedload population moves through all parts of the channel despite differences in capacity and local shear stress. Only the extreme coarse tail of the bedload grain size distribution is selectively transported. This has implications for understanding the sorting mechanisms in braided rivers and the choice of grain size for competence calculations.

Introduction

Braiding represents one way in which rivers may adjust their planimetric geometry to the imposed conditions of sediment supply, valley slope and discharge. Braided rivers have a characteristic form of constantly migrating and switching multiple channels and extensive rapidly aggrading and degrading gravel bar forms. Although favoured by high energy environments with variable discharges, abundant bedload and easily erodible banks (Cheetham 1979), braided rivers occur in a wide variety of environments from proglacial to semi-arid and at all scales from small streams to continental rivers.

Despite their widespread occurrence, processes operating in braided rivers have received comparatively little attention even though the hydraulics and bedload transport processes are of significant interest to several disciplines. Sediment transport processes are of particular interest to geomorphologists and sedimentologists because of the close connection with sedimentary form and process (Ashmore 1988). As active sediment transporting and depositing systems, braided rivers create characteristic sedimentary structures, which when preserved in the rapidly aggrading floodplain, are used by sedimentologists to reconstruct the nature of the fluvial environment that deposited them (Miall 1978). Until processes operating in modern braided rivers are understood more fully, the sedimentologist cannot make accurate and confident interpretations from ancient deposits (Bridge 1985, Reid and Frostick 1989). Sedimentary units produced by frequently avulsing and rapidly aggrading rivers also form important potential hydrocarbon reserves and mineral deposits (Miall 1978). An understanding of transport and depositional processes operating in braided rivers would greatly aid the prediction of oil reservoir size (Haldorsen *et al*. 1987), coal and gas reserves (Smith and Putnam 1980) and the likely location of heavy mineral enrichment (Slingerland 1984). Because of the close connection with lateral and vertical channel stability, it is important that engineers also gain an understanding of bedload transport mechanisms. Engineers building structures on braid plains frequently encounter pronounced bed scour at channel confluences, rapid bank erosion and bar migration inducing channel avulsions which have to be incorporated into design criteria (Neill and Hey 1982, Wang *et al*. 1987).

Few comprehensive field studies of bedload transport processes have been undertaken that provide data on transport rates and calibre in gravel bed rivers. This is largely due to the difficulty in making process measurements in channel forming flows, the wide range of sediment sizes, the low frequency and the short duration of transporting flows and the spatial and temporal variability in process, even in steady flow (Thorne *et al*. 1987). These difficulties are compounded when working in braided rivers by the difficulty of taking measurements with rapid channel change and the inherent variability in processes induced by non-uniform flow associated with flow convergence and divergence around mid-channel bars.

This paper presents results from the detailed study of hydraulics and bedload transport in a chute-bar unit of the braided Sunwapta River, Canada by a large research team with duplicate sets of equipment, essential for such spatially intensive fieldwork. Flow and bedload fields simultaneously mapped immediately upstream of an aggrading mid-channel bar illustrate the highly variable distribution of shear stress, the systematic cross-stream and downstream pattern of transport rate and the variability of bedload grain size distribution and composition. The importance of channel morphology in funnelling bedload through the chute to the downstream aggrading bars is considered and the inter-relationships between flow, bedload transport, and channel change discussed.

Previous Work

Early work on braided rivers derived from the need of sedimentologists to interpret coarse fluvial deposits in the geological record (Rust 1972, Miall 1978). The internal sedimentology of bars has been studied by Krigstrom (1962), Boothroyd and Ashley (1973), Bluck (1974, 1979) and Smith (1974). Work in the field by Hein and Walker (1977) traced the evolution of individual bars, developing the ideas of Leopold and Wolman (1957) and Smith (1974) to establish qualitative models describing bar growth, downbar fining and vertical grain size segregation of mid-channel bars. Mid-channel bars often exhibit a downbar decrease in grain size from barhead to bartail, the mechanisms of which are not fully understood. Suggestions include the winnowing out of fines from successive aggrading gravel sheets (Hein and Walker 1977), the preferential deposition of coarse sizes influenced by flow turbulence and bar pocket geometry (Bluck 1982), the addition of fining upwards avalanche faces at the bartail (Ashmore 1982) and the routing of fines to the bartail by flow divergence at the barhead and secondary circulations in the distributaries (Ashworth 1989).

Because of the difficulty in studying braided rivers in the field, several workers have taken advantage of flume models to investigate various aspects of the braided river environment (Ashmore 1982, 1988, in press, Ashmore and Parker 1983, Southard *et al.* 1984, Ashworth and Powell in review) using the principles of Froude modelling whereby the model stream is directly scaled to a field prototype. Ashmore (1982, 1988) modelled bar development and bedload transport at different values of constant discharge and slope. He found that mean bedload transport rates, averaged over the whole stream and over a fairly long time period could be predicted by conventional hydraulic variables - discharge, stream power and shear stress, but strong temporal variations were evident and channel patterns with different braiding indicies experienced different transport rates, even at constant stream power. Though the modelling approach has the advantages of allowing the direct observation of channel processes, a complete overview of the whole developing system and because of the compressed time scale, permits the study of the long term evolution of channel pattern, these must be weighed against the disadvantages. These include the accurate simulation of a coarse gravel bed river and associated hydraulics at such a small scale, the use of constant discharges and small depths and the need to allow the channel to develop freely, unhindered by the flume bed or walls (Davies 1987). Perhaps more importantly is the inability to quantify the detailed nature of hydraulics and bedload transport patterns in braided reaches at such a small scale and in such rapidly changing conditions (Ferguson 1987).

The field investigation of braided river dynamics at the reach scale received impetus from the work of Mosley (1982, 1983) and Southard *et al.* (1984) who identified the chute-bar unit as the fundamental spatial and morphological unit of braided rivers (although see Rundle (1985a, 1985b)). Convergent flow in the chute results in an acceleration of flow which decelerates as the flow diverges out of the chute. The curvilinear and accelerating flow field increases the shear stress in the convergence zone above that caused by hydrostatic pressure (Das and Townsend 1981, Richards 1982) such that it is strong enough to transport sediment from upstream and scour the chute bed and banks creating high transport rates. Out of the chute, the flow diverges, becomes less competent and shallower resulting in the deposition of a mid-channel bar with separate channels on either side. Distributary channels continue to transport sediment through the reach though not necessarily in equal amounts (Cheetham 1979) and in turn may develop into chutes, repeating the pattern.

Recent work on bedload transport has shown that even during steady flow, significant spatial and temporal variations in rates and sizes occur (Emmett 1975, Tacconi and Billi 1987, Whiting *et al.* 1988, Ashmore 1988, Kuhnle and Southard 1988, Iseya and Ikeda 1987, Dietrich *et al.* 1989). These have variously been attributed to the migration of low relief bedforms such as gravel sheets and local disruption of the armour layer, break up of sediment clusters and the interaction of coarse and fine sediments with reduced sediment supply. This implies that an accurate study of bedload transport processes at the reach scale require that the measurements be spatially distributed throughout the reach and rapidly taken to accommodate the apparent variability in process, creating obvious logistical problems. Davoren and Mosley (1986) were amongst the first to relate flow hydraulics, sediment transport and channel change in a chute-bar unit using jet boats to sample in channel forming flows. Using an uncalibrated basket sampler at cross sections across a braided reach of the Ohau River, New Zealand, they found that transport rates and hydraulic parameters were broadly consistent; high and low shear stresses and transport rates coincided with the fast flowing deep eroding chute and the shallower, depositional diverging flow zone respectively. Bedload transport rates were found to be spatially very variable. Confusing the issue however was the conclusion that local reach scale processes were affected by larger scale sediment transport variations since a similar reach with similar hydraulic characteristics behaved very differently when the upstream sediment supply was cut off.

Ashworth and Ferguson (1986) related simultaneous point estimates of shear stress and bedload transport in a braided reach of the proglacial Lyngsdalselva, Norway. They found that the development of channel bed and bar topography is both the cause of the spatially varied flow and transport and its consequence (Richards, 1987). Flow velocities and shear stress were consistent with flow convergence and divergence, creating variations in bedload transport rates which in turn helped explain channel changes and bar formation, despite a complex relationship with discharge.

Site Location and Description

The Sunwapta River, in Jasper National Park, Alberta, Canada (Figure 1a) is a proglacial tributary of the Athabasca River, flowing northwards from the Athabasca Glacier in the Colombia Icefield. The river flows from a small lake bounded by terminal moraines at the snout of the rapidly retreating glacier. The overall channel pattern of the upper 25 km is braided with two exceptions being when the river is confined within steep floodplain terraces and as it flows through a narrow gorge (Figure 1b). The study reach is located at Beauty Creek Flats, the longest and most extensively braided part of the Sunwapta, beginning downstream of the Sunwapta Gorge and continuing 11 km north where the channel becomes confined by alluvial fans and debris slides (Figure 1b). This part of the river has been the location of other work by Ashmore and Parker (1983), Rice (1982) and Dawson (1988).

The study reach, about 15 km from the river's source, consists of a well defined convergence zone, five metres wide, out of which the flow diverges asymmetrically around and over a lobate shaped medial bar (Figure 1c). At high flow the barhead is submerged by about 30 cm of water. The tributary feeding the right hand channel enters too far down the reach to affect processes operating at the barhead. Bed material is predominantly limestone and dolomite made up of coarse gravel up to 128 mm. Bed slope is 0.03 determined from levelling down the talweg.

The hydrology of the Sunwapta is typical of proglacial rivers with a clear diurnal cycle. Peak discharges occur in the evenings of hot sunny days when glacial melt reaches its maximum, although subsidiary peaks occurred on days with high rainfall despite relatively low temperatures. The discharge of the study reach was not measured but a chart recorder at the glacier snout (Figure 1c) indicted that the highest river discharges were on the 28 and 31 July although by the latter date, the reach had been abandoned.

Measurement Techniques

Proglacial rivers are ideal sites for examining bedload transport processes, hydraulics and bar deposition in braided rivers because the diurnal fluctuation in discharge permits the quantification of transport rates and hydraulics during the evening high flows and the associated bar deposition rates and sizes during the following morning low flow period when bars are emergent. A total of 126 point bedload samples were taken on the 26, 28, 29 and 30 July with estimates of flow strength taken at 114 of the sampling stations.

Channel form

Channel form was defined by levelling across 10 parallel cross-sections spaced 5 m apart in the direction of flow. Cross-sections are referred to by the distance downstream from the first section (see Figure 1c). After the initial survey on the July 22, repeated daily levelling from July 28-31 gave information on local scour and fill and bank erosion.

Bedload transport

Bedload transport rates were measured using 76 and 152 mm Helley-Smith samplers with 0.25 and 2 mm mesh bags. The coarser mesh had the advantage that the large amounts of fine material carried by proglacial rivers was not sampled. Despite the inherent difficulties in obtaining reliable estimates of transport rates with portable samplers when bedload is spatially and temporally variable (Hubbell 1987) using such devices is the most practical method for the direct measurement of bedload discharge, yielding reasonable results if used within a carefully devised sampling programme. In order to quantify the spatial pattern of transport rates and sizes, samples were taken at as many points along the cross sections as possible within the time constraints, typically 1 m apart. Intensive mapping of the bedload field on July 26, 28, 29 and 30 collected 27, 26, 38 and 39 samples from either 4 or 5 of cross-sections 5 to 25. Up to three Helley-Smith samplers were used simultaneously to enable the rapid measurement of the bedload field. Sampling times were for either 5 or 10 minutes depending on the sampling site. The largest Helley-Smith was used in the high transport/shear stress zones where the coarsest particles were expected to be in transport. Because the maximum particle size was always less than half the sampler aperture size and the collecting bag never more than half full, reduced trap efficiency was not a problem. It took up to three hours to quantify the total bedload field from chute to barhead during which observations indicated that the reach discharge had not changed appreciably. After drying, samples were sieved and weighed at half-phi class intervals and then truncated at 2 mm to calculate fractional and total transport rates and grain size distribution percentiles for gravel material. The b-axis of the maximum particle size in each sample was also recorded.

Bed material grain size was quantified to compare sediment transport to that available in the surface and subsurface. The spatial variability of river gravel size has been well documented (Church and Kellerhals 1978, Mosely and Tindale 1985) and methods for obtaining representative estimates of bed material sizes have been comprehensively reviewed by Church *et al.* (1987). Given time and logistical constraints it is difficult to work to the standards and recommendations of Church *et al.* so the spatial pattern of bed material was sampled by the quicker Wolman (1954) method. The D_{50} of the surface bed material, defined as the median diameter by number of 100 sampled surface grains, decreases from 30 to 22 mm from cross-sections 5 to 25 with a corresponding decrease in D_{90} of between 60 and 75 mm to 40 mm. Samples taken at four cross-sections on the 29th indicated that the reach had fined with a D_{50} of about 20 mm and a D_{90} of about 40 mm. Bar head material had a D_{50} and D_{90} of 27 and 44 mm respectively. This fining is consistent with the channel changes (see later).

Velocity and Shear Stress

Bedload transport rates were linked to the local flow field by three teams measuring vertical velocity profiles at sites immediately after bedload samples had been taken. Profiles were taken throughout the water depth using Pygmy Price, Swoffer 2100 and Ott current meters with single impellers mounted on top-setting rods so that the roughness height remained constant throughout the profile. Velocities were averaged over 30 seconds and taken at intervals of between 1 and 3 cm depending on depth and type of current meter.

Point shear stress (τ) and roughness height (z_o) were calculated by fitting the 'law of the wall' equation to the full velocity profile (Ashworth and Ferguson 1989) and mean velocity at 0.6 depth by integrating the profile over the whole depth. The standard error of the shear stress measurements, expressed as a percentage of the estimated shear stress were calculated using Wilkinson's (1984) equations and ranged from 5 - 44% with a mean of 17%. Roughness heights ranged from 3-32 mm with a mean of 8 mm. Shear stress measurements were only accepted as being reasonable estimates of flow strength if the semi-logarithmic height-velocity profile was close to linear (standard error < 30%) and z_o correlated well with local bed material size. 100 estimates met these conditions.

Results and analysis

Spatial pattern of flow and bedload transport

Pattern of flow strength

Although the data set for the flow velocity field is incomplete, some generalisations about the spatial distribution of velocity can be made. Similar patterns occurred over all four days, but results from July 28 and 29 will be used to illustrate the non-uniform nature of the flow. Channel morphology defines contrasting zones of converging accelerating and diverging decelerating flow at cross-sections 0-15 and 20-25 respectively. Where the flow is confined within the deep narrow chute mean velocity is at a maximum reaching 1.3 and 1.2 m s^{-1} on the 28th and 29th respectively. Velocity decreases laterally away from the talweg to about 0.8 m s^{-1} as the flow is retarded by the rough channel banks and bed in the shallower flows. Velocity also decreases with depth downstream of the chute at cross-sections 20 and 25 as the flow divides around the mid-channel bar. Mean cross-sectional velocity (calculated by averaging four flow estimates from the same four sampling stations at different cross-sections) decreases downstream from 1.12 to 0.78

m s^{-1} on the 26th (insufficient data is available for the 28th) and from 0.96 to 0.79 m s^{-1} on the 29th.

The spatial pattern of bed shear is not so well defined. Figure 2 shows the spatial pattern of shear stress for the study period. Daily estimates vary from about 5-60 N m^{-2} except on the 29th when the maximum is only 36 Nm^{-2}. Only 6 of the 100 shear stress measured over the four days are greater than 42 Nm^{-2}. Figure 2 shows there is no consistent pattern to the distribution of shear stress through the reach, although generally higher shear stresses tend to be associated with the talweg, decreasing laterally and downstream. The limited data of July 26 show measurements to be very variable with up to a seven fold difference between adjacent estimates. Flow stress is considerably lower at cross-section 25 than those upstream. The highest shear stresses of the four days coincides with the peak river discharge of the study period on the 28th. Shear stress decreases laterally from the talweg to the channel margins and downstream towards the barhead. Overall, shear stresses are generally lower and laterally more consistent on the 29th. The majority of estimates at cross-sections 5 and 10 are between 15-25 N m^{-2} before increasing at cross-section 20. Variability increases again on the 30th with a lateral decrease in shear stress estimates from consistently high values in the talweg. Estimates of shear stress in the talwegs either side of the bar at cross-section 25 are markedly lower than those in the chute.

Pattern of bedload transport rates

Mapping of the bedload field indicated that the sampling period coincided with the beginning and ending of activity in the reach (Table 1). A few bedload samples taken on the 23rd indicate that the reach was inactive with low transport rates of less than 10 g m^{-1} s^{-1}. Activity increased through to the 28th and 29th with peak transport rates approaching 300 g m^{-1} s^{-1} before decreasing to approximately 80 g m^{-1} s^{-1} on the 30th when the reach was abandoned due to an avulsion upstream (Table 1).

Over the four day period, transport rate shows a consistent and systematic variation downstream and laterally at individual cross-sections. Spatial maps of bedload transport rates > 2 mm are shown in Figure 3. July 28 and 29 were the most active days, with highest transport rates occuring in the chute talweg (cross-section 5-15) reaching peak values of 294 and 276 g $m^{-1}s^{-1}$ with all but the coarsest bed material in motion, decreasing to 102 and 35 g $m^{-1}s^{-1}$ as flow diverges around the bar at cross-sections 20-25 (see also Table 1). Bedload forms into two separate threads either side of the medial bar with relatively little sediment moving onto the barhead. More pronounced is the strong lateral gradient in transport rates at a cross section. Bedload transported through the chute is concentrated in a narrow zone, approximately 2 m wide with a rapid, systematic reduction in transport towards channel margins. For example on the 29th at cross-sections 10 and 15, rates vary from almost 0 to 276 and 135 g $m^{-1}s^{-1}$ and on the 28th from near 0 to 193 and 198 g $m^{-1}s^{-1}$ within a few metres. Of the four days, the 28th has the highest rates of bedload transport in the chute, associated with the migration of the talweg and the reworking of loose channel infill (see below). Out of the talweg and in the divergence zone little scour occurs and there is a change from an eroding to a depositional regime.

Relationship between bedload transport and shear stress

Several workers have attempted to correlate bedload transport rates with some measure of flow strength. The three most common are shear stress, (Einstein 1950, Ferguson *et al.* 1989), stream power, (Emmett 1976, Leopold and Emmett 1976, Reid and Frostick 1986) and mean velocity, (Thompson 1985). In all cases, relationships contained much scatter, attributed to experimental error and the complex nature of sediment transport. Comparison of Figures 2 and 3 show that the pronounced systematic lateral and downstream variation in transport rate is only broadly reflected by that of shear stress. A simple proportional relationship between bed stress and transport rate is well defined but has substantial scatter (Figure 4). A regression line fitted by the reduced major axis method (Mark and Church 1977) because errors are associated with both transport rate and shear stress measurements yields a highly significant relationship ($p>0.001$) but only a low R^2 value of 17% indicating that shear stress is a poor predictor of transport rate. Some of the scatter can undoubtedly be explained by measurement error since the degree of scatter increases at low transport rates, but some may be real, reflecting variations in transport rate associated with velocity or bedload pulses (Kuhnle and Southard 1988, Williams *et al.* 1989).

Channel change and the relation to bedload transport

Figure 5 shows the local pattern of erosion and deposition for July 22-30. From the 22nd to 28th, the chute becomes infilled as the flow is unable to transport all the sediment supplied from upstream and the channel widens by up to 3 m. Further deposition occurs across the full channel width at cross-section 20 and the existing bar aggrades, principally by accretion at its left flank. On the 28th the chute infill is reworked as a narrow zone, 3 m wide and up to 19 cm deep is eroded through the centre of the chute. Much of the sediment transported through the chute is deposited in the right hand distributary as a new medial bar. The reach is relatively stable on the 29th with only isolated areas of erosion and deposition. By July 30, channel changes upstream were creating an avulsion, which over the next 12 hours cut off the flow and sediment supply to the reach. Both erosion and deposition occur in the chute but downstream, sediment is deposited in the distributaries and bar margins are eroded.

The streamwise pattern in transport rates is broadly correlated with channel change. The relatively low transport rates on the 23rd and 26th (Table 1 and Figure 3) are associated with aggradation throughout the reach and decrease downstream as increasing amounts of sediment is lost to deposition. Higher bed stresses in the chute on the 28th create the highest transport rates as the loose deposits of the previous few days are eroded and the reduction in maximum transport rate of nearly 50% as the flow emerges from the chute is associated with deposition at cross-section 20 and the construction of a new bar at cross-section 25. However, a more detailed comparison between days and cross-sections indicate that significant discrepancies exist. The increasing amount of erosion between cross-sections 5 and 15 on the 28th (Figure 5) should be reflected by a corresponding increase in transport rate as progressively more sediment is transported downstream. However, transport rates decrease from cross-sections 5 to 10 before stabilising through cross-sections 10 to 15 (Table 1, Figure 3). Furthermore, on the 29th, despite only marginally lower transport rates (and shear stresses) to the previous day, the chute is relatively stable with minimal erosion compared to the 28th. Obviously sediment supplied to the head of the reach forms a substantial proportion of the bedload transported downstream and the chute acts as and effective corridor for all gravel supplied to it. Out of the chute, the maximum transport rate decreases by 80%, but there is no evidence of significant aggradation. Transport rates are much lower on the 30th despite little appreciable change in shear stress. This is due to the avulsion occurring upstream which began to cut off the flow and sediment supply to the reach. This highlights the role of the chute as an efficient transporting agent and shows that if the sediment supply is turned off, the bed soon stabilises and releases little further sediment to the bar downstream. These comparisons between channel change and sediment transport help explain the scatter in Figure 4 since point estimates of bedload transport rate may not be directly attributable to the shear stresses at that point but to those further upstream.

Bedload composition and grain size distribution

As well as total transport or capacity being spatially variable, the calibre or grain size of the bedload also varies throughout the reach. Figure 6 shows that the distribution of maximum particle size (D_{max}) is similar to that of transport rate with a clear lateral decrease from the talweg and downstream decrease from chute to barhead. D_{max} varies from 4 to 48 mm at cross-section 15 on the 29th and decreases from up to 48 mm to about 30 mm in the divergence zone, although a minority of samples have D_{max}'s up to 40 mm. The progressive lateral decrease in D_{max} is particularly prominent in the four upstream sections of the convergence zone (cross-sections 0-15) but where the narrow talweg dissipates and shallows in the divergence zone (cross-sections 20-25) the size segregation of D_{max} is not so clear cut.

The degree of correlation between D_{max} and total transport rate (Figures 3 and 6) is shown in Table 2. Stream capacity and competence are closely linked with a relationship significant at $p > 0.001$. This implies that entrainment and transport processes are strongly size selective. An increase in capacity seems to arise from the progressive reworking of existing sediment supplies rather than the input of new unsorted sediment. Comparisons of bedload grain size distributions with sizes available in the reaches' bed surface show that the bedload was always finer than the bed material (largest D_{max} in Helley-Smith 65 mm, bed D_{max} up to 90 mm). This may be due to the poorer sampling efficiency of the Helley-Smith with coarser grain sizes (in this case due to sampling time length rather than orifice size) but is more likely to be a real response to the flow strength being below the threshold required to entrain all sizes of sediment (c.f. Parker *et al.* 1982). This same pattern has been shown by Ashworth and Ferguson (1989, Figure 2) where at moderate flows entrainment was size selective and only at bankfull discharges did all sizes become mobilised. However, despite the high correlation between competence and capacity, Table 2 shows that the variations in D_{max} are only poorly related to changes in shear stress (significant at $p > 0.01$). Even when considering experimental error, the correlation coefficient is low, suggesting that other factors such as the nature of sediment supply and the pattern of channel change may be important influences on sediment transport.

Table 2 also shows that the coarsening of bedload with increasing transport rate is not reflected throughout the whole grain size distribution. The variation in the D_{90}, D_{75} and D_{50} percentiles becomes progressively less such that the D_{50} is relatively constant and almost independent of transport rate. This is clearly illustrated in Figure 7, a map of the distribution of D_{50} throughout the reach. Apart from the occasional anomalous sample, it is evident that there is little variation in D_{50} either downstream or laterally, with 80% of the total number of samples having D_{50} 's between 8 and 18 mm. With such consistency in the overall bedload grain size distributions, it is not surprising that the grain size percentiles in Table 2 show such poor correlations with shear stress (Figure 2), total transport rate (Figure 3), and D_{max} (Figure 6). The paradoxical situation whereby D_{50} is constant despite increases in D_{max} with capacity is discussed in more detail in Ashworth and Powell (in review).

Discussion

Mapping of the daily shear stress field shows that it is highly variable both within a few metres and throughout the whole chute-bar unit. In part, this may be due to errors in the shear stress measurements, but is also attributable to the complex interaction between channel morphology and the flow. Over 95% of the velocity profiles were log-linear indicating no need for zero height correction or a consideration of acceleration and deceleration effects. High shear stresses tended to be associated with the chute talweg, decreasing laterally and downstream. This pattern is what one would expect if shear stress is assumed to be directly proportional to the depth-slope product and slope shows little variation. This is supported by comparison of Figures 2 and 5 which show that the generally systematic decrease in shear stress towards the left bank is associated with the progressive decrease in depth from the talweg. Towards the right bank, shear stresses are more variable reflecting the undulating nature of the bed topography.

Transport rates show a much stronger and more systematic cross and down stream change. Estimates are almost certainly associated with some error, but the presence of a systematic pattern in transport rate and D_{max} gives confidence in the results and suggests that sampling times were long enough to catch the coarsest material in transport. Support for the representativeness of the Helley -Smith samples is also provided by a study of aggradation on the medial bar (Powell and Ashworth, in prep.) where sediment sizes in barhead deposition, integrated over each daily flow period, match those found in transport. Highest transport rates occur in a narrow strip through the chute where flow is deepest and fastest (less than 25% of the total bed area) with maximum cross-sectional transport rates up to 5 times the average. Either side of the talweg, transport rates decrease rapidly towards the banks. Such a narrow concentration of bedload has also been observed by Davoren and Mosley (1986) and Prestegaard (1989). As the flow emerges from the confines of the chute, the bedload field expands to occupy the full channel width and transport rates are reduced, becoming more consistent cross-stream. There was no evidence to suggest that bedload was being preferentially swept down either side of the distributaries and pebble tracer experiments showed that once deposited on the barhead, sediment was difficult to re-entrain. Despite the path of maximum transport rate approximating with the zone of high shear stress and the spatial pattern of transport rate broadly consistent with that of shear stress, the relationship between boundary shear stress and bedload fields is relatively poor.

The complexity of bedload transport is well known (see reviews by Klingeman and Emmett, 1982 and Bathurst, 1987) and it would be unusual if a simple correlation between shear stress and transport rate gave a well defined relationship in rivers with such non-uniform flow and bed topography. Some of the factors influencing bedload transport described in the literature are probably not as important in braided rivers. Because of the abundant sediment supply and the deposition of large amounts of gravel, cluster bedforms, imbrication and particle packing have little time to develop and bed material remains relatively loose, reducing the effects on the entrainment process as described by Laronne and Carson (1976) and Brayshaw *et al.* (1983). For similar reasons and because proglacial rivers experience regular diurnal floods in the summer, inhibition of transport by a coarse surface layer (Gomez 1983, Sutherland, 1987) is probably not too important in a braided river environment.

The nature of sediment transport and influence of sediment supply do appear to be the main driving forces governing the rate and size distribution of bedload. Evidence from the flume and field has shown that bedload transport can be unsteady with periodic pulses related to the migration of low relief bedforms such as gravel sheets (Iseya and Ikeda 1987, Whiting *et al.* 1988, Reid *et.al.* 1985). Results presented in Powell and Ashworth (in prep.) suggest that transport may occur in this form since bar deposition occurred by the successive aggradation of well sorted but independent gravel sheets. Although the origin and hydraulic stability of these bedforms is still not understood and their role in gravel transport only beginning to be quantified (Weir 1983), it appears that their migration may well be independent of the local bed shear stress. If this is true, then it is not surprising that there is such a poor correlation between bedload transport rate and shear stress (Table 2). In addition, at-a-point sampling of bedload and shear stress does not take into account conditions immediately upstream of the measurement site. With relict bars and uncohesive banks readily available to erode and collapse, sediment movement can be initiated at lower critical shear stresses than that required to entrain sediment from a tightly packed bed (Reid *et al.* 1985). Once in motion, particle weight and therefore momentum, is undoubtedly important in determining transport rates.

The rate of bedload transport and the distribution of sizes moving are important variables for the explanation of bar and downstream sorting. The mapping of the bedload field (Figures 3, 6 and 7) permits an overview of the changes in capacity and bedload composition through a confluence/diffluence unit and helps explain how different sizes are sorted into bed, bar and transient sediment. This aspect is elaborated in Ashworth and Powell (in review) but can be seen in general terms by comparing Figures 3, 6 and 7 and Table 2. The bedload D_{max} is highly variable, being particularly responsive to changes in channel capacity whilst finer percentiles become increasingly less variable such that the D_{50} is essentially constant. Neither shows a strong correlation with shear stress. The higher variability of D_{max} is to be expected since it is only a measure of one particle, and since samples were sieved at half phi intervals, it can vary by up to 26% without affecting the D_{50}. However, the difference in behaviour in D_{50} and D_{max} is quite striking since it occurs over a wide range of shear stresses and transport rates. Tacconi and Billi (1987) found a similar consistency in grain size distribution despite wide variations in capacity. The mechanism whereby the bedload composition is essentially kept constant despite wide variations in capacity and competence is unclear. However, the consistency is not caused by a compensatory change in the amount of fines since truncating the bedload at 11 mm produces a similar pattern. It is clear however that increases in competence is only reflected by a small increase in the coarse fraction, insufficient to affect the D_{50}. The implication from this is that sorting does not occur in transport except at the extreme coarse end of the grain size distribution. Bedload consists of essentially the same grain size distribution moving down the reach in a well defined pattern of varying amounts (see Ashworth and Powell in review).

The variability of D_{max} with respect to D_{50} is the opposite pattern described by Komar (in press). In a re-analysis of the data of Milhous (1973), he found that the D_{50} is most responsive to a change in flow strength and that the responsiveness decreases with progressively coarser percentiles. He concludes that variations in shear stress affect the whole grain size distribution such that many percentiles should be used to obtain multiple evaluations of flow competence. It is clear that these conclusions are not universally applicable to all rivers and this study supports the use of D_{max} in flow competence evaluations providing it can be representatively sampled.

Conclusion

This study has shown that given a sufficiently large, equipped research team, it is possible to intensively map the flow and bedload fields of natural rivers producing reasonably defined patterns of flow strength and bedload transport. The processes operating in a chute bar unit are dominated by the non-uniform flow associated with flow convergence and divergence. Bedload through the chute is concentrated in a discrete narrow zone, associated with the deep fast flowing talweg, and diverges out of the chute forming into two separate threads around the mid-channel bar. The link between boundary shear stress, transport rate, bedload grain size distribution and channel change is weak with the interrelationships complicated by variations in sediment availability and its source. Bedload entrainment in rapidly migrating braided rivers is largely independent of conventional sedimentological factors such as bed packing and hiding and protrusion since most particles originate from bank collapse and the reworking of loose bed sediment. Once sediment is in motion the spatial distribution of bedload transport rate and D_{max} is sensible being related to the downstream and cross-stream decreases in shear stress with flow depth. The overall bedload size population is independent of hydraulic controls and despite changes in capacity remains fairly uniform both across and downstream. If little sorting of bedload occurs at entrainment, and in rivers with abundant sediment supply, no sorting occurs during transport, sediment must be selectively sorted during deposition. The degree of sorting undertaken at each stage of a particle's journey through a river system needs quantification and further investigation.

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Tables

Table 1

Maximum daily bedload transport rates (>2 mm) in g m^{-1} s^{-1}.

N/S = not sampled

Date	Cross section Number				
July	5	10	15	20	25
23	8	10	8	6	4
26	N/S	31	40	16	11
28	294	193	198	102	N/S
29	82	135	276	43	35
30	38	79	29	11	57

Table 2

Table of correlation coefficients for D_{max}, D_{90}, D_{75} and D_{50} against total transport rate (n = 110) and shear stress (n = 90). Grain size percentiles are calculated for transport rates > 0.6 g $m^{-1}s^{-1}$ to ensure an adequate sample size.

Grain size parameter	Transport rate	τ
D_{max}	0.73 *	0.29 ~
D_{90}	0.50 *	0.21
D_{75}	0.37 *	0.08
D_{50}	0.23 +	-0.05

* significant at p > 0.001

~ significant at p > 0.005

+ significant at p > 0.01

Figure Captions

Figure 1 a-c. Location of the Sunwapta River and morphology of the study reach. Measurements were concentrated between cross-sections 5 and 25.

Figure 2. Spatial pattern of shear stress for July 26, 28, 29 and 30 estimated from velocity profiles. Area of circles is directly proportional to shear stress.

Figure 3. Spatial pattern of bedload transport > 2 mm for July 26, 28, 29, 30 as measured by 5-10 minute duration Helley-Smith samples. Note the logarithmic scale.

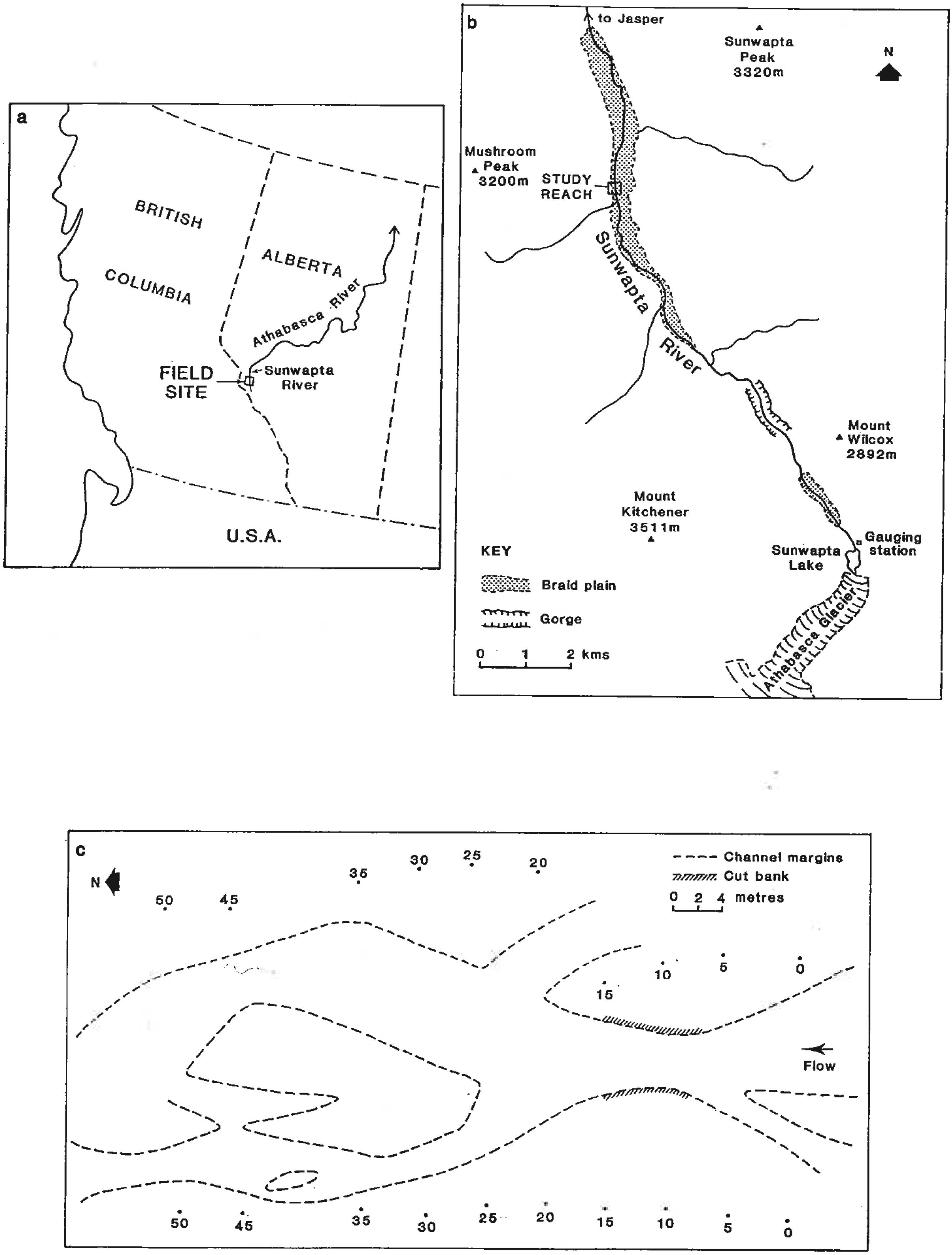
Figure 4. Relationship between point measures of bedload transport rate per unit width > 2 mm and associated estimates of point shear stress. The regression line is fitted by reduced major axis (see text).

Figure 5. Pattern of cross-sectional change from July 22 to 31 for cross-sections 0 to 25. Note the difference in interval dates between surveys.

Figure 6. Distribution of maximum particle size of bedload samples for July 26, 28, 29 and 30. Area of circle directly proportional to particle size.

Figure 7. Distribution of median particle size for bedload samples with transport rates $> 1\ gm^{-1}s^{-1}$ for July 26, 28, 29 and 30. Area of circle directly proportional to particle size.

Fig. 1


a
BRITISH
COLUMBIA
ALBERTA
Athabasca River
FIELD SITE
Sunwapta River
U.S.A.
b
to Jasper
Sunwapta Peak 3320m
N
Mushroom Peak 3200m
STUDY REACH
Sunwapta River
Mount Wilcox 2892m
Mount Kitchener 3511m
Gauging station
Sunwapta Lake
Athabasca Glacier
KEY
Braid plain
Gorge
0 1 2 kms
c
N
Channel margins
Cut bank
0 2 4 metres
Flow
50 45 35 30 25 20
15 10 5 0
50 45 35 30 25 20 15 10 5 0

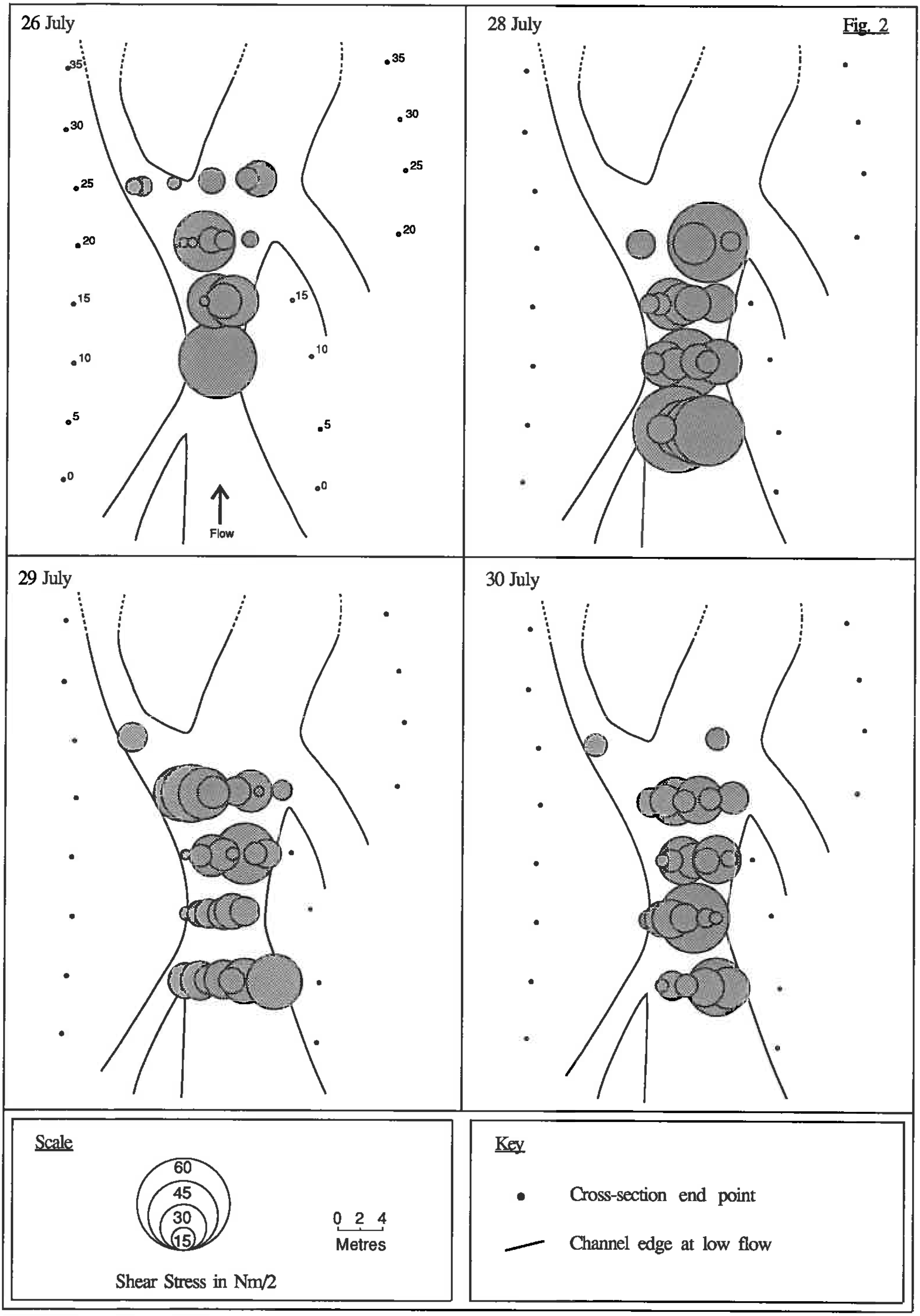


Fig. 2
26 July
28 July
29 July
30 July
35
30
25
20
15
10
5
0
Flow
Scale
60
45
30
15
Shear Stress in Nm/2
0 2 4
Metres
Key
Cross-section end point
Channel edge at low flow

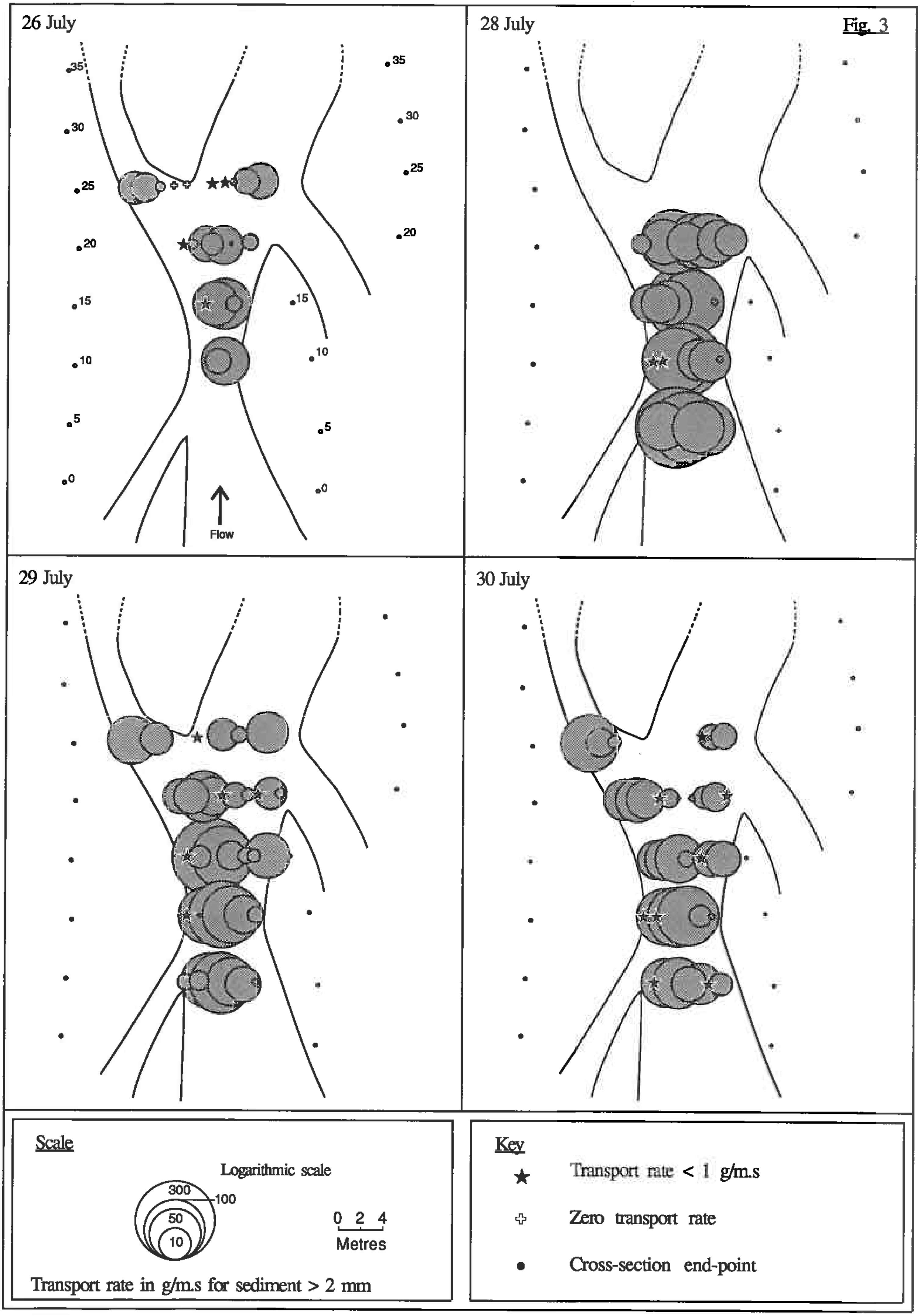
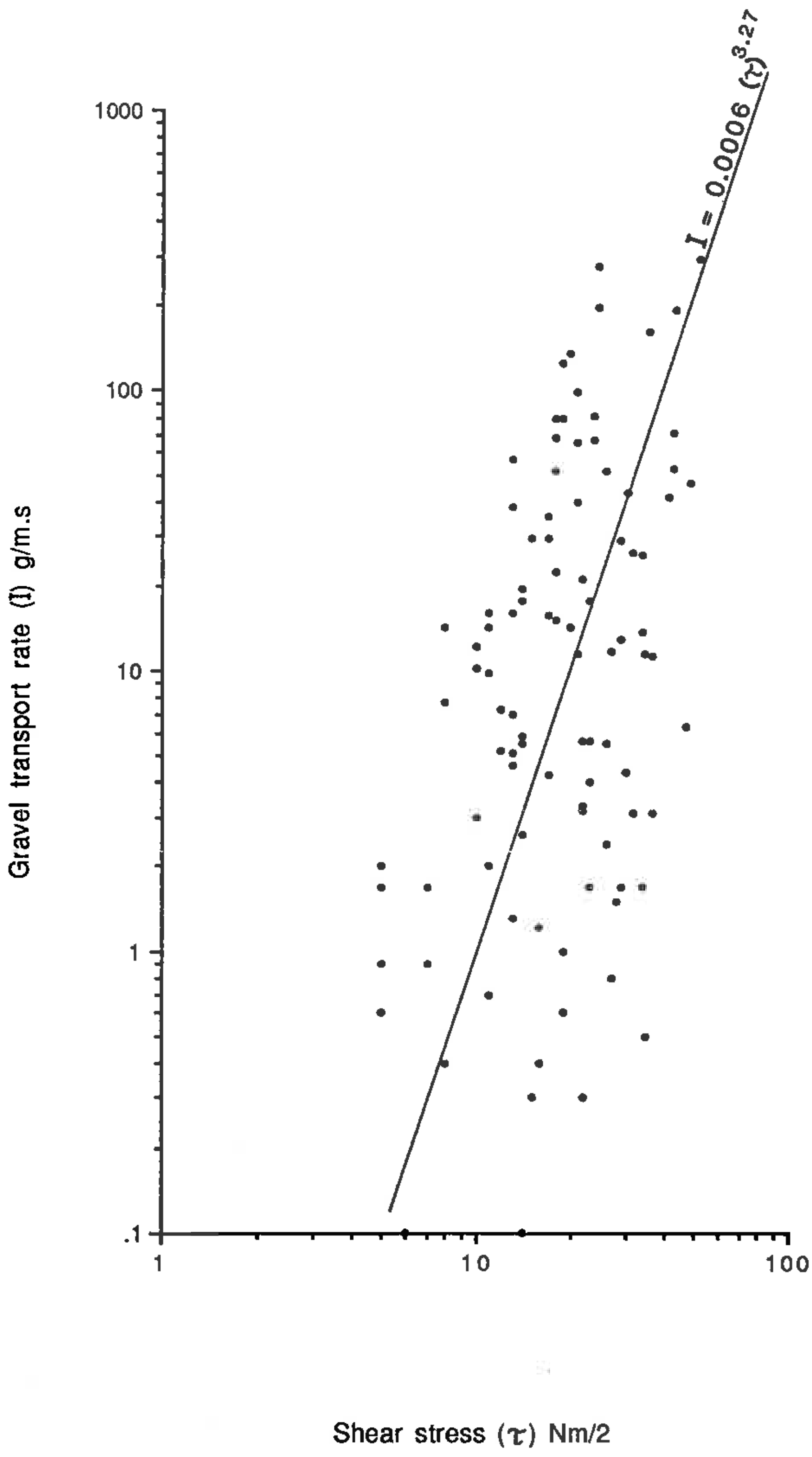


Fig. 3
26 July
28 July
29 July
30 July
35
30
25
20
15
10
5
0
Flow
Scale
Logarithmic scale
300
100
50
10
0 2 4
Metres
Transport rate in g/m.s for sediment > 2 mm
Key
Transport rate < 1 g/m.s
Zero transport rate
Cross-section end-point

Fig. 4



$I = 0.0006\,(\tau)^{3.27}$
1000
100
10
1
.1
1
10
100
Gravel transport rate (I) g/m.s
Shear stress (τ) Nm/2

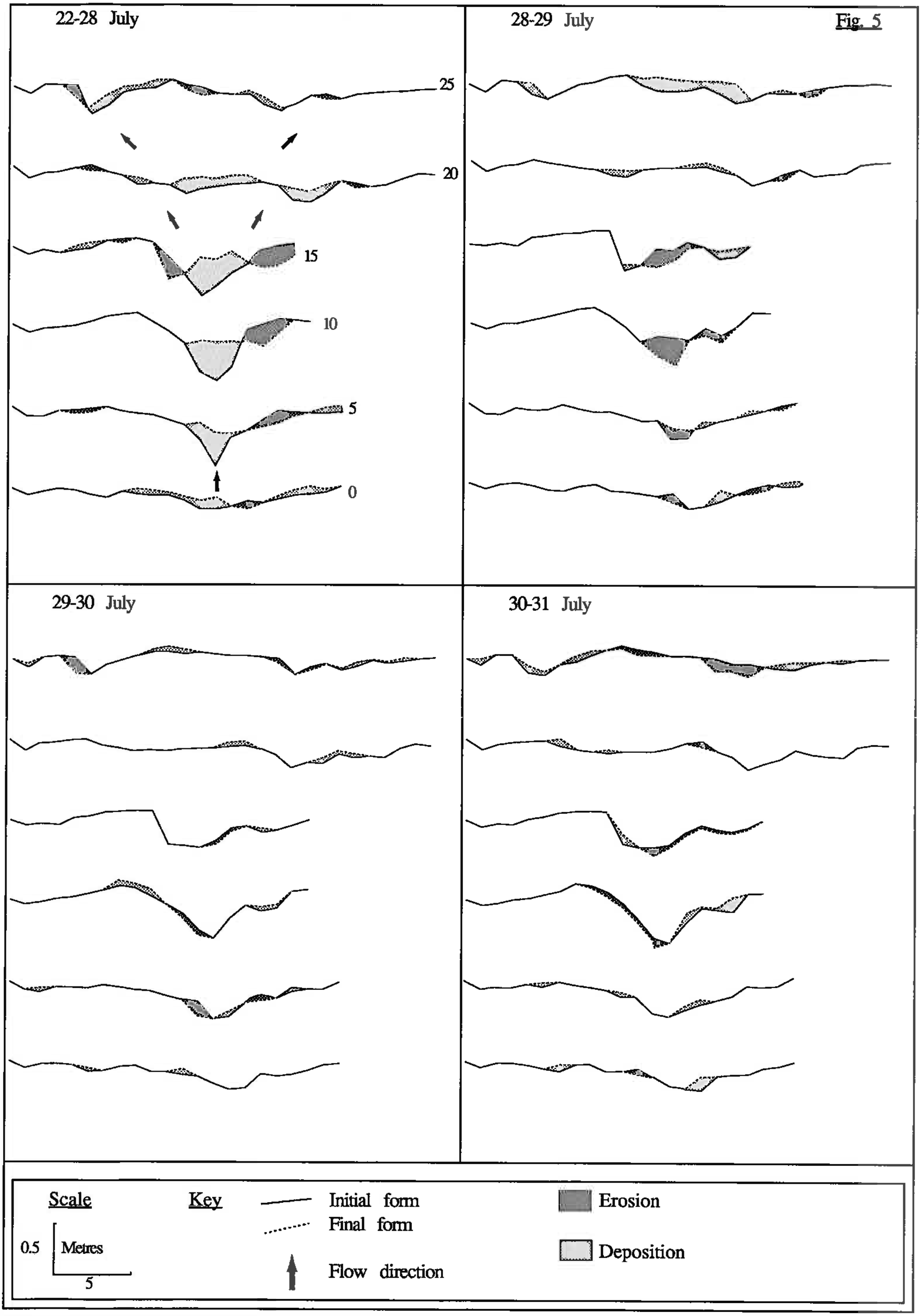
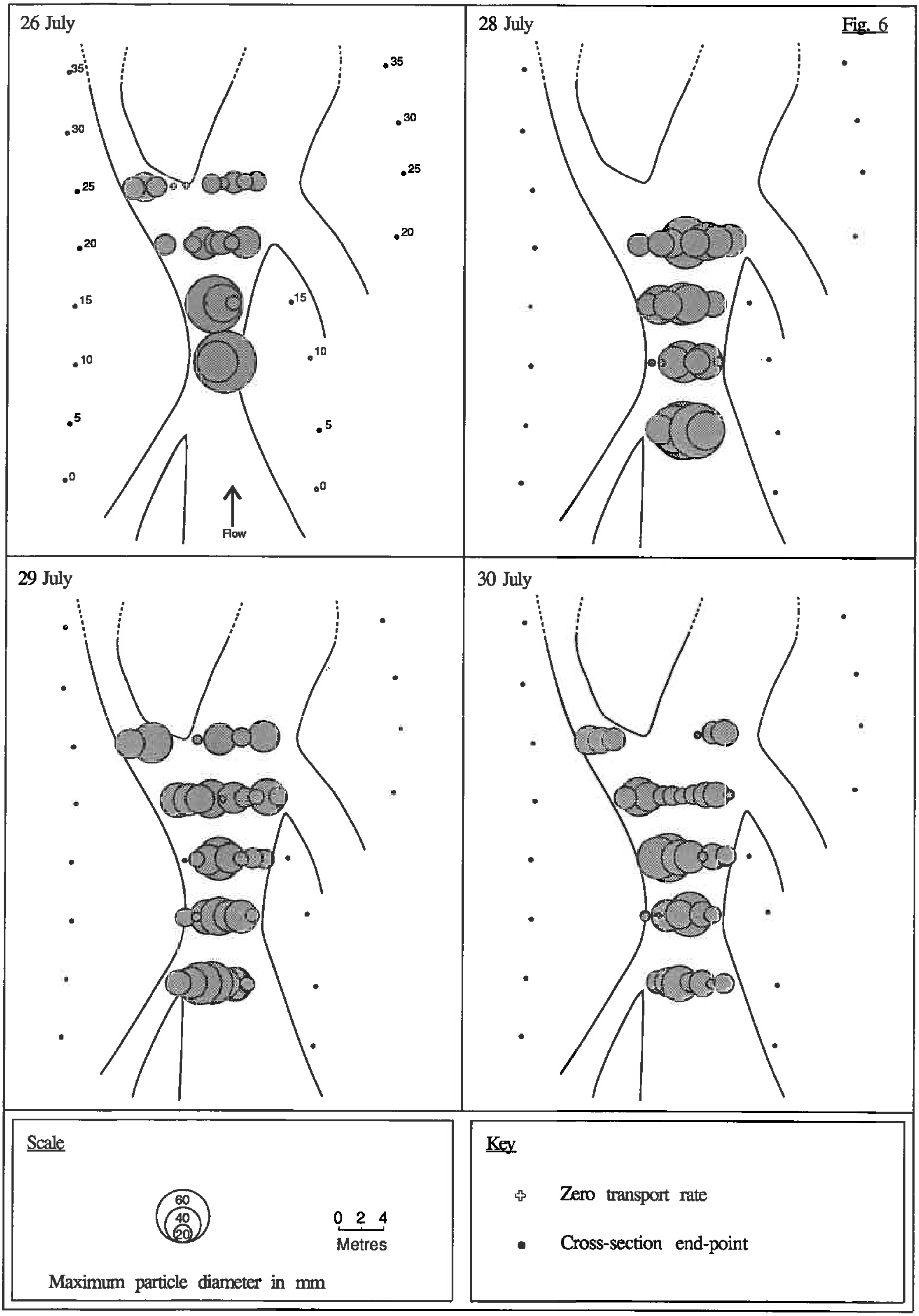


Fig. 5
22-28 July
28-29 July
29-30 July
30-31 July
25
20
15
10
5
0
Scale
0.5
Metres
5
Key
Initial form
Final form
Flow direction
Erosion
Deposition


26 July
28 July
Fig. 6
35
30
25
20
15
10
5
0
35
30
25
20
15
10
5
0
Flow
29 July
30 July
Scale
60
40
20
0 2 4
Metres
Maximum particle diameter in mm
Key
Zero transport rate
Cross-section end-point

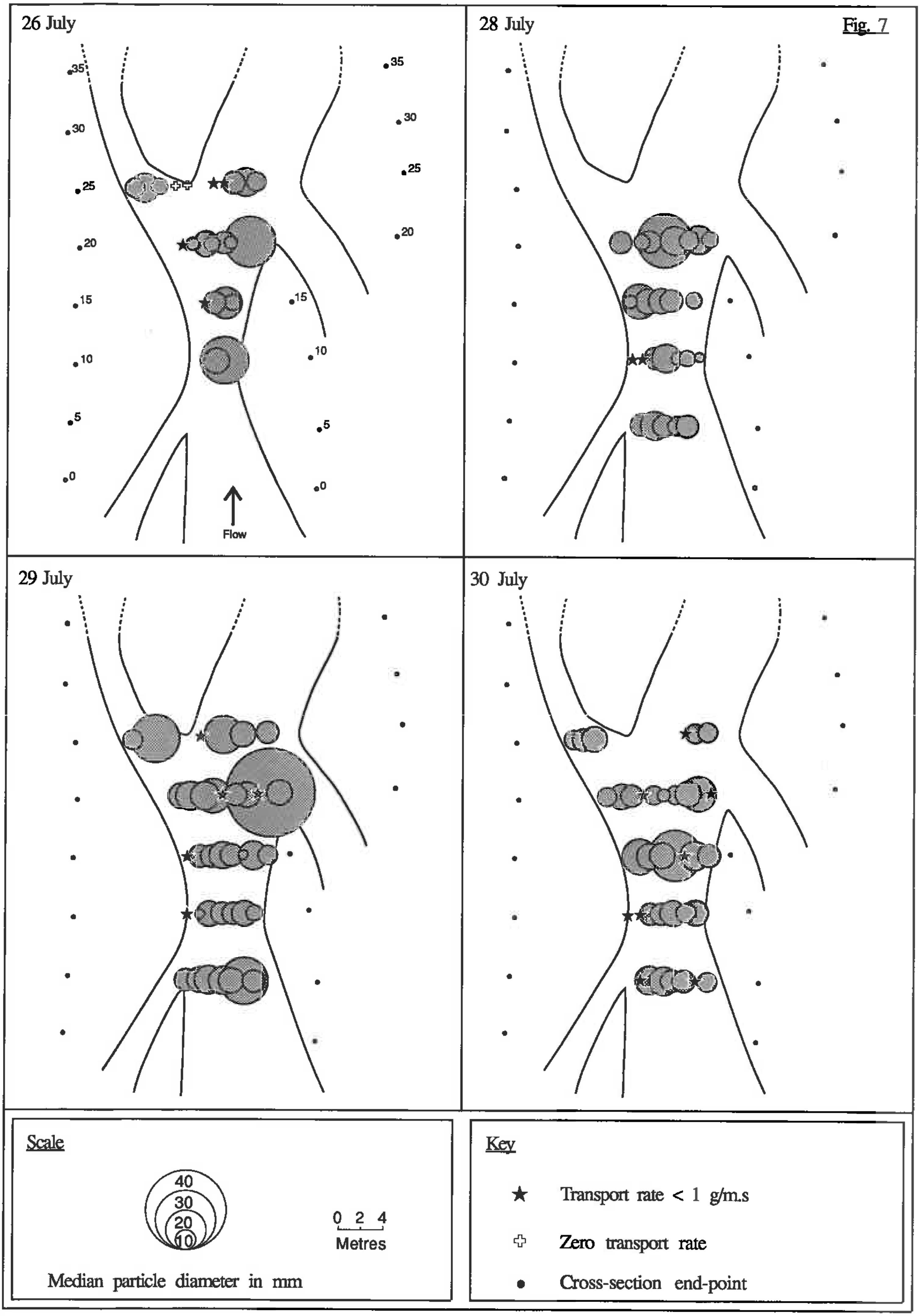


Fig. 7
26 July
28 July
29 July
30 July
35
30
25
20
15
10
5
0
Flow
Scale
40
30
20
10
Median particle diameter in mm
0 2 4
Metres
Key
Transport rate < 1 g/m.s
Zero transport rate
Cross-section end-point