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Flow behaviour, suspended sediment transport and transmission losses in a small (sub-bank-full) flow event in an Australian desert stream

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Abstract:

The behaviour of a discrete sub-bank-full flow event in a small desert stream in western NSW, Australia, is analysed from direct observation and sediment sampling during the flow event and from later channel surveys. The flow event, the result of an isolated afternoon thunderstorm, had a peak discharge of 9 m^3/s at an upstream station. Transmission loss totally consumed the flow over the following 7·6 km. Suspended sediment concentration was highest at the flow front (not the discharge peak) and declined linearly with the log of time since passage of the flow front, regardless of discharge variation. The transmission loss responsible for the waning and eventual cessation of flow occurred at a mean rate of 13.2% per km. This is quite rapid, and is more than twice the corresponding figure for bank-full flows estimated by Dunkerley (1992) on the same stream system. It is proposed that transmission losses in ephemeral streams of the kind studied may be minimized in flows near bank-full stage, and be higher in both sub-bank-full and overbank flows. Factors contributing to enhanced flow loss in the sub-bank-full flow studied included abstractions of flow to pools, scour holes and other low points along the channel, and overflow abstractions into channel filaments that did not rejoin the main flow. On the other hand, losses were curtailed by the shallow depth of banks wetted and by extensive mud drapes that were set down over sand bars and other porous channel materials during the flow. Thus, in contrast with the relatively regular pattern of transmission loss inferred from large floods, losses from low flows exhibit marked spatial variability and depend to a considerable extent on streamwise variations in channel geometry, in addition to the depth and porosity of channel perimeter sediments.

KEY WORDS suspended sediment; transmission loss; sub-bank-full flow; desert stream; Broken Hill Australia

INTRODUCTION

The observation of flow processes in ephemeral dryland streams is made difficult by the very low fraction of the year in which flow occurs. From a dryland stream in Israel, Reid *et al.* (1998) have recently shown this fraction to be about 2%, or roughly seven days per year. Monitoring is further hampered in many areas following rain, when roads may be cut or untrafficable. Masses of woody debris swept along within the flow also restrict gauging and sediment sampling work. Consequently, relatively few data exist on the behaviour of ephemeral flows (Reid *et al.*, 1994). Part of what is known has been derived indirectly from the study of bedforms and other sedimentary features that are preserved within the channel after flow has ceased (e.g. Williams, 1970), and has relied on the process of relating bedform character and size to flow depth and flow speed. This technique would often be unworkable for small flows, which may leave few or no bedform traces

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because the shear stresses they generate are inadequate to remould pre-existing, high-stage bed materials. Automatic recording devices offer one means of gathering data on ephemeral flows, but small events make even this approach difficult, owing to the fact that transmission losses may consume all flow before a sampling or gauging site is reached, resulting in a failure of the automated system to record the event. Indeed, in the face of transmission losses, which may cause marked spatial variability in flow characteristics, multiple recording sites are really essential so that a reasonable picture of the size of an event can be pieced together, but this makes the exercise costly. All of these obstacles, then, stand in the way of data collection during sub-bank-full flow events in ephemeral dryland streams.

This circumstance of logistically difficult data collection has repercussions in the investigation of ephemeral channel form and its interaction with flow dynamics. It is clear that our understanding of the way in which the geometry of ephemeral channels is influenced by flow events of varying magnitude (and especially bank-full flows) is less developed than in the case of perennial streams. In addition, there remains a need for a modified hydraulic geometry framework suitable for the study of channel form in streams where flow may decline (rather than increase) with catchment area, as a result of transmission losses. This lack of a suitable conceptual framework hampers the progress of dryland hydraulic geometry. Given the general shortage of reliable gauging data, one of the major uncertainties in the hydrology and channel hydraulics of ephemeral streams is how flow losses in small events compare with those in larger events. Dunkerley (1992) relied on large pieces of woody debris lodged in riparian and channel-bed trees, and debris lodged against shrubs from overbank flows beyond the channel margins, in order to reconstruct downstream trends in flood stage along a 15 km reach of Fowlers Creek, an ephemeral channel of the Australian inland. Over this flow path, a rapid decline in peak flood discharge was inferred and attributed to transmission loss. Very little debris of the kind used in that study is commonly available as a record of low flows. These move primarily smaller and lighter pieces of organic debris, notably fallen leaves and other litter. Materials of this kind are more prone to post-flood dispersal by wind from sites of deposition, and can easily be confused with local leaf fall from riparian and channel-bed trees. This makes it quite difficult to employ flood-transported debris to infer flow stage for small flow events. In the case of debris left by large floods, there is the additional benefit that the stranded detritus is elevated above the stage reached by more common flows, and is thus preserved for sufficiently long that the record it provides can conveniently be extracted.

All of the circumstances just reviewed have contributed to a notable dearth of data on small flow events in ephemeral channels. This paper reports observations made opportunistically during a discrete flow event that occurred in the course of unrelated fieldwork, and over the following 48 hours, while evidence in the channel could still be interpreted with certainty as being derived form the recent flow event. Attention is drawn to some factors that were observed to contribute to marked transmission loss from this flow event, and to others that acted to restrict such losses. Additionally, the nature and behaviour of the significant suspended load carried is reported from samples collected periodically during the flow. Though the available database is small, it enables some interesting hypotheses to be advanced.

THE FIELD AREA AND ORIGIN OF THE FLOW EVENT

The flow event studied occurred in Homestead Creek, an ephemeral stream located in the Barrier Range of western NSW, Australia. This stream is a tributary to a larger regional system, Fowlers Creek. Above their confluence, Homestead Creek has a catchment area of 30 km^2. Below the confluence, Fowlers Creek conducts the combined flows towards a shallow ephemeral lake, Lake Bancannia, which lies approximately 20 km away to the north-east. Flows are almost always consumed by transmission losses before reaching this basin. The Homestead Creek catchment is somewhat elongated in the N–S direction (Figure 1) owing to structural control exerted by a plunging syncline in the underlying Precambrian sediment and meta-sediment basement. High strike ridges flank the channel, and reach elevations of 288 m. The flanks of these ridges are covered by blocky bedrock rubble, and this retards overland flow and tends to promote infiltration (Brown, 1997). Lower slopes, in contrast, have only finer gravel veneers overlying thick silty soils. Vegetation canopies

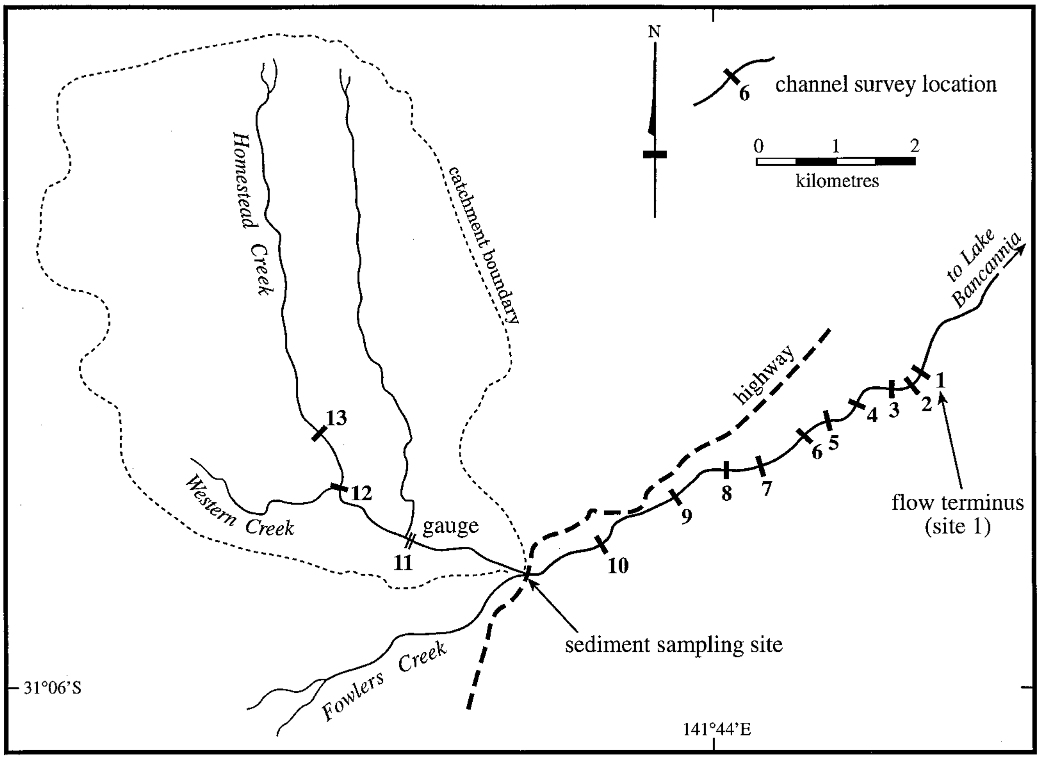

Figure 1. Location map of the Homestead Creek–Fowlers Creek area, showing locations of surveyed channel cross-sections and the location of suspended sediment sampling

cover only 25–30% of the surface, and bare areas have strongly developed rain-beat and other surface seals. Often, these seals are underlain by vesicular horizons (Brown and Dunkerley, 1996) and, in concert, the seals and vesicles provide a strong choke on infiltration. This is very marked and runoff is generated rapidly on lower slopes, arising from as little as a few mm of rainfall at intensities common in the area. The presence of thick silty soils over dominantly quartzite bedrock is explained by the late Quaternary deposition of dusts deflated from the continental interior, and laid down in blankets across the Barrier Range landscapes (Chartres, 1982). The channel of Homestead Creek is incised into reworked aeolian sediments, which were stored along the valley floors following their erosion from the hillslopes. The precise nature and chronology of these events remains unresearched.

Above the junction with Western Creek, the bed of Homestead Creek is frequently incised to bedrock, and carries only a veneer (5–30 cm) of mobile gravelly bed sediment. Occasional sediment bars, up to 0·5 m in thickness, occur in places. Below this junction, the bed is composed of increasing depths of alluvial sediment, dominantly of coarse sand and granule texture, but with accumulations of cobbles and boulders on bar tops or in lee-side accumulations behind bedrock outcrops, trees or large immobile particles. The bed is armoured in places, but equally often is surfaced by sands laid down during waning flow. The thickness of sediment overlying bedrock is quite variable, but the mean would be no more than 1 m. The channel of Fowlers Creek below the junction with Homestead Creek is similar, but displays higher bars of sand and gravel, which reach 3–4 m above the thalweg, and which divide the channel into multiple filaments at lower stages. A high proportion of the bed material is sand, but with local accumulations of gravel, cobbles and boulders. The largest sizes are progressively abandoned as flows traverse the channel and lose competence because of transmission loss, leaving only fine sands and silts in the lower reaches of the channel (Dunkerley, 1992).

The climate of the area is arid, the average rainfall being 190 mm. This is delivered from slow-moving synoptic depressions, and from brief thunderstorms such as those that resulted in the flood described here. There is no strong seasonal pattern to the rainfall, rain days being as likely in the cool winter as in the hot summer. Surface water is absent with the exception of shaded pools along some streams, where water may remain for many weeks following a flood event. Catchment vegetation is dominated by scattered chenopod shrubs, notably *Atriplex vesicaria* (bladder saltbush) and *Maireana* spp. (bluebushes), while larger streams support quite dense tree growth of the river red gum, *Eucalyptus camaldulensis*, both along the banks and on the channel bed.

The flow event

A flow event resulted from afternoon thunderstorms on 15 May 1998. These were observed from the lower part of Homestead Creek catchment, where it remained dry. Isolated thunderstorm cells were observed developing beyond the western perimeter of the catchment, and over its northern extremities, at about 4 p.m. Heavy rain persisted for about 45 minutes, following which the intensity fell and the storm cells dissipated. Rain from localized thunderstorm cells is notoriously difficult to measure, requiring very dense networks of rain gauges (Goodrich *et al*., 1995). Of 14 gauges located on the pastoral property, only records from rain gauges close to, but outside the Homestead Creek catchment are available to indicate the rain amounts delivered. The four closest gauges lie within 3 km and recorded totals of 2·4, 4·2, 5·6 and 23.0 mm of rain on the afternoon of 15 May, confirming marked spatial variability in rain totals. The fact that a substantial flow resulted in the channel suggests that a significant part of the upper catchment probably received totals nearer the 23 mm reading. Inspection of the ground surface on 16 May showed signs of runoff (fresh damage to unsealed farm roads, sediment deposited from sheet flow detained as it passed through dense shrub or grass patches, and freshly undercut banks along small tributary streams and gullies) over much of the central and northern parts of the catchment. South of Western Creek (Figure 1) no signs of runoff were found, confirming that the runoff originated in the northern part of the catchment.

Channel flow was observed moving towards a permanent gauging station on Homestead Creek (Figure 1 and Figure 2). The flow front was very low, and moving at a speed estimated at 1 m/s. Woody debris could be seen moving along within the flow, which was very turbid from suspended sediment. Monitoring was then moved to the junction of Homestead Creek with Fowlers Creek, in order to observe the arrival of the flow

Figure 2. Photograph taken on 15 May 1998 of the advancing flow front just upstream of he gauging structure on Homestead Creek

and to undertake systematic sampling of its suspended sediment load. The slow rate of advance allowed time to walk up the channel and once more observe the behaviour of the advancing flow front, which was easily out-paced on foot.

On the day following the flow, debris lodged on both banks at the gauging station was surveyed with a theodolite. The strand lines were at accordant heights above the gauge zero (left bank stage 60·4 cm, right bank stage 60·9 cm) and indicated a peak stage of 60 cm here. The rating table developed for this permanent concrete control structure (no longer used for continuous data collection) indicates that the discharge at this flow stage is 9·1 m^3/s.

BEHAVIOUR OF THE FLOW AND SUSPENDED SEDIMENT CONCENTRATION

Samples were collected by hand, standing in the flow, at the sealed bitumen road crossing located at the stream junction (Figure 1). This provided firm footing in the flow, and a stable datum against which depth could be judged. There is also a sight board indicating flow depth at this road crossing, for the advice of motorists. Flow depth rose rapidly after arrival of the front, and peaked at a depth of $\leqslant 30$ cm after about 10 minutes. Flow recession was then steady and continuous, and the flow was only a few cm deep when sampling was abandoned in darkness at 6:15 p.m. Flow stage at the sampling location remained low, owing to the broad and shallow engineered highway crossing. A total of 14 sequential samples was taken in washed, 500 ml wide-mouth polyethylene bottles, by submerging the bottle in the flow and filling it completely. The first sample, taken at 5:20 p.m., was of the very first flow to spill from the channel above the highway on to the bitumen surface. Thereafter, samples were taken at intervals of about 5 minutes until dark. The following day, after recording individual volumes, the water samples wee decanted into weighed aluminium dishes, dried in a laboratory oven at 105 °C and reweighed to the nearest 0·01 g.

The resulting time-series of suspended sediment concentrations (Figure 3) shows a remarkably smooth decline from a peak of sediment concentration in the very first flow. There was no prior flow whatsoever, so that there was no rapid rise in concentration; rather, the first flow carried the peak sediment concentration and a monotonic decline followed.

Figure 3 demonstrates that the sediment concentration declines linearly with the logarithm of time elapsed since first flow. The fitted relationship is

$$\text{sediment concentration (g/l)} = 26.2 - 5.12 \ln(\text{elapsed time since passage of flow front, minutes})$$

which is significant at $p < 0{\cdot}001$, $r^2 = 0{\cdot}99$.

This pattern of concentration decline appears not to have been reported previously. Frostick *et al.* (1983) noted that in one Kenyan data set, the peak sediment concentration was associated with the flood bore, and not with the discharge peak that arrived some minutes later. However, the decline in concentration after its early peak is much steeper in the Kenyan data, and Frostick *et al.* (1983) did not analyse the form of the relationships that they reported.

The peak suspended sediment concentration reached was 25·7 g/l; this declined to 5·8 g/l in the last sample collected. These concentrations are higher than those reported by Frostick *et al.* (1983) from a semi-arid site in Kenya, where peak concentrations were typically 10–15 g/l. They are, however, lower than those derived from the small arid Nahal Yael Basin in Israel (Lekach and Schick, 1982), where concentrations (including considerable amounts of suspended sand) quite often reached 50–100 g/l in small flow events (peak discharge 1–2 m^3/s or less). Concentrations of up to 90 g/l have been reported from Nahal Eshtemoa in the northern Negev (Reid *et al.*, 1994).

The observed pattern of concentration decline suggests a diffusion-controlled mechanism to us. A dependence of concentration on the square root of time would be suggestive of an infiltration-controlled loss of water volume without a corresponding reduction in sediment load. The root-t term in this case arises from the diffusive nature of the water entry and reciprocal air escape processes involved in the percolation of water

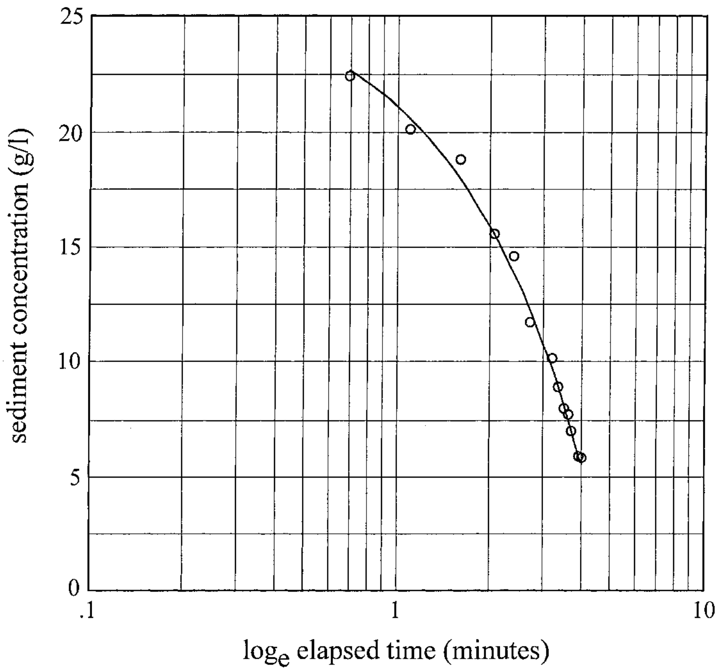

Figure 3. Behaviour of suspended sediment concentrations during the flow event of 15 May 1998. The data have been linearized by plotting concentration against the natural logarithm of elapsed time. The least-squares linear regression line is shown

into porous sediments (e.g. see Philip, 1957). However, while the fit of sediment concentration and root-t is good, the fit with log(t) is considerably better, showing very strong linearity. This log(t) dependence does not immediately suggest the nature of the mechanism responsible. It will, therefore, be necessary to build a model of the various processes involved in which to explore this further. The controls on sediment concentration almost certainly involve percolation of water with some form of filtering in the uppermost parts of the bed, leaving excluded sediment within the channel. The consequent progressive sealing of pores may account for the departure from root-t behaviour.

The regular decline in suspended sediment concentration after the concentration peak has been found in data on other flow events in this channel system. For example, a larger flow event, with peak discharge of about 18·6 m^3/s, was sampled on 17 October 1992. The flow was triggered by quite widespread rain, and the hydrography (observed at the Homestead Creek gauging structure already referred to) showed a rise time of about 1 hour and a recession of 4–5 hours. Suspended sediment concentrations rose rapidly with flood stage, peaking at 39 g/l. When the declining concentrations measured after the peak are plotted against time since the peak concentration, a linear decline with log(t) again results. The fit is very slightly better than for root-t, as noted for the 1998 flow event. For comparison with the relationship given above for the 1998 flow, that for 1992 is

$$\text{sediment concentration (g/l)} = 51{\cdot}9 - 9{\cdot}35 \ln \text{(elapsed time, minutes)}.$$

This relationship is significant at $p = 0{\cdot}0001$, with $r^2 = 0{\cdot}98$.

A good series of particle size analyses available for the 1992 flood (made with a commercial laser scattering particle size analyser) showed that in the case of mean volume particle diameter (d_{50} vol), the fit of declining size with the passage of time was actually better for root-t than for log(t). The root-t relationship (Figure 4), having $r^2 = 0{\cdot}97$ and significant at $p = 0{\cdot}0001$, is

$$d_{50}\text{vol } (\mu\text{m}) = 12{\cdot}7 - 0{\cdot}52 \text{ (square root of elapsed time, minutes)}$$

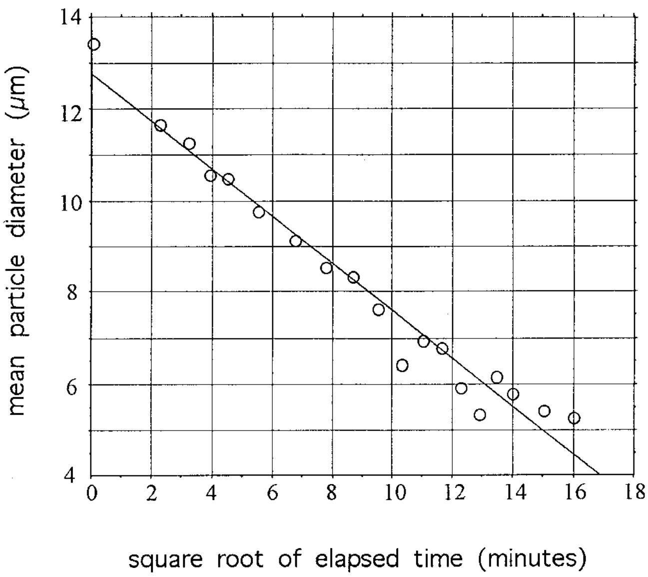

Figure 4. The relationship between mean volume particle diameter of suspended sediment (μm) and the square root of time elapsed since the peak concentration, Homestead Creek flood of 17 October 1992. The least-squares regression line is shown

This is strongly suggestive of a control involving diffusive percolation of flow into the bed, perhaps with time-dependent exclusion of suspended particles. The exploration of mechanisms of this kind controlling grain-size decline will again require the use of models incorporating bed texture and water uptake properties, together with the interaction of moderate suspended loads with the bed materials.

During the 1998 flow event, small amounts of sand and granule materials were detected moving as bedload at the highway crossing, but the amounts involved could not be quantified.

To the hypothesis just outlined can be added other mechanisms that might contribute to the observed patterns of monotonic decline in sediment concentration and particle size. A decline in the availability of sediments in the source area does not appear likely, as there are extensive areas flanking the channel along Homestead Creek, and these are known to release large amounts of sediment in all runoff events. A decline in the scale or intensity of mechanical turbulence within the flow in parallel with the discharge decline could contribute to the observed behaviour. However, recalling that the characteristics of the sediment load declined monotonically from the very first flow and were unaffected by the latter passage of the peak in discharge, this mechanism cannot be itself account for our observations. In other words, it is the peculiar insensitivity of the sediment load characteristics to the passage of the discharge peak that is suggestive of a mechanism linked instead to the physical progression of the flow front across the porous bed materials.

DOWNSTREAM TRENDS IN FLOW STAGE

The small 1998 flow provided the opportunity to observe flow stage at points successively downstream of the termination of flow. This was located on 16 May, at a point approximately 6 km below the highway crossing where sediment sampling was performed, and with a total channel length of 14·2 km from the northern end of the catchment. Over the flow path from the Homestead Creek gauge to the flow terminus (Figure 1), bed elevations fall from 175 m to 157 m, an average slope of 0·002. The termination of flow was not marked by any accumulation of sediment; the coarse sandy bed left by an earlier flow was planar and its surface clean for tens of metres upstream of the flow terminus. Rather, this location was indicated by wet sand, and by a low accumulation of leaves (primarily derived from the riparian and channel-bed trees). These leaves are ordinarily released in abundance and lie on the stream bed, where they are readily swept along by any advancing flow. The absence of significant mud overlying the bed suggests that only a small volume of flow arrived at this location. Upstream, the bed and banks in places were heavily cloaked in mud laid down from this flow event, often to thicknesses of 1–3 cm. Sediment concentrations may thus also have been lower at this extreme limit of flow, as a result of progressive deposition along the flow path.

At intervals running upstream from the flow terminus, the peak stage was identified from clearly delimited dampness of the banks, together with accordant lines of flood detritus, mostly leaves, and/or by mud set down below the water line and clearly contrasting with the sandier texture of the pre-existing bank materials. Wet bank materials (and the wet sand used to recognize the flow terminus) could only be employed in this way by virtue of the fact that no rain fell locally, all flow having been derived from the distant upper catchment area. The accordance of stranded litter and dampness of the banks confirmed that there had been negligible capillary rise of wetness in the banks. Cross-sections to define flow stage and cross-sectional geometry were surveyed with an electronic distance measuring theodolite, where they were wide, or with a measuring tap tightly strung across the channel to mark the water line, and hand measurement of verticals at 1 m intervals, where the bed was narrow. At some sites, existing mid-channel sand bars had not been overtopped by the flow, so that there had been two separate threads of flow. Both sub channels were then surveyed, and the determination of cross-sectional area and mean depth employed the data from each. In all cases, the selected sites were located away from major channel pools or obstructions, at locations that appeared to have carried uniform, free flow. All cross-sections were located with a 12-channel GPS receiver and later transferred to topographic maps at 1:25 000 scale. Channel lengths between stations were then determined by digitizing the stream course along the channel centre-line.

The resulting data showed a statistically significant pattern of decline in cross-sectional area of the flow with travel distance from the Homestead Creek gauge. The relationship, for which $r^2 = 0{\cdot}86$, was

$$\text{cross-sectional area of flow } (\text{m}^2) = 6{\cdot}2 - 0{\cdot}54 \text{ (distance below gauge, km)}$$

This relationship is statistically significant ($p = 0{\cdot}0001$).

The peak discharge recorded at the gauging station ($9{\cdot}1$ m^3/s) declined to zero over a channel flow path of 7·6 km. Strictly speaking, the calculating of transmission loss requires that a complete flood hydrograph be available for each site along the channel. Data of this quality are almost never available, and the working assumption made here is that the decline in discharge peak as assessed from strandline evidence may be taken as an acceptable general indicator of the pattern of transmission loss, though it is recognized that some lowering of the recorded peak stage may simply represent a broadening and lowering of the hydrograph shape. On this basis, the measured flow decline amounts to a mean transmission loss rate of 13.2% per km. Dunkerley (1992) estimated transmission losses of 12–13.5% per km for Sandy Creek (the range depending upon the method of estimation used), a considerably smaller stream located about 10 km away, but only 5–6·9% per km for Fowlers Creek. Both of these estimates were for large (bank-full or slightly overbank) flood events, and not for sub-bank-full flow. The lower 6 km of the flow-path of the small Homestead Creek flood is in fact in the Fowlers Creek channel, so that the estimated loss of 13.2% per km is more than twice the rate for large floods in the same channel. It is worthwhile, therefore, to consider those factors that contributed to this very high transmission loss in a sub-bank-full flow that was fully contained within the channel.

Factors affecting transmission losses in sub-bank-full flow

Generally speaking, in short-lived flow events, the major cause of transmission loss is seepage into bed and bank sediments. For larger and more prolonged events, losses to alluvial surfaces inundated by overbank flows also become significant.

Let us consider some of the controlling factors in the case of the particular sub-bank-full flow studied here. Many of these factors seem likely to be applicable to sub-bank-full flows in ephemeral channels.

First, at the stream confluence, the flow from Homestead Creek passed into the larger channel of Fowlers Creek, whose catchment area is 434 km^2, or nearly 15 times larger. The bed of Fowlers Creek was completely dry, the total rainfall in the 3 months prior to the date of the flood being only 4·8 mm. The probable previous flow event was on 7 February 1998, when 16·6 mm of rain was recorded in the catchment. Owing to the much greater channel capacity of Fowlers Creek, where flood stages can reach 4–5 m above the thalweg, the shallow flow debouching from Homestead Creek occupied only low-lying parts of the large channel. This had several consequences. Large sediment bars in the channel rise several metres above the thalweg. These are the product of high-stage events and are composed of gravel, cobbles and a matrix of coarse sand often deposited downstream of trees and associated debris barriers. Some are later colonized by trees and become semi-permanent. They may divide the channel into several filaments at lower stages, in the manner described as 'anabranching' by Nanson and Knighton (1996). In contrast, the lowest channel floors are commonly of better-sorted and more tightly packed, pebbly sands. They are indurated by fines deposited within the sediment during flows and also from the desiccation of turbid pools that remain in low points. In the flood of 15 May, flow in the lower 5 or 6 km of the path had been incapable of disrupting these materials, and had passed over them, leaving the bed undisturbed and depositing little or no mud. Thus, in parts of the channel where the sub-bank-full flow was confined to low-lying channel beds, transmission losses appear to have been quite low. This behaviour might well be promoted also by higher moisture levels in the bed sediments lying beneath the thalweg, where pools left by floods gradually dissipate by seepage. Higher parts of the bed would be submerged for shorter times because of relatively rapid stage rise and fall, so taking up less water, and would also lose water readily because of the hydraulic head provided by their elevation above the thalweg. A first conclusion, therefore, is that the sub-bank-full flow may have suffered lower transmission losses and hence passed further along the channel because of the textural properties and relatively higher water content of the bed materials along the thalweg.

A second characteristic of the large channel, however, had the opposite effect. Fowlers Creek is characterized by significant undulations in bed elevation, whose origin is unclear. Low-lying and relatively narrow zones of the channel, which occupy reaches of 100–200 m in length, end downstream with a zone of reverse bed slope. In other words, they constitute pools along the channel. These pools impound water to depths of 1–2 m (Figure 5) and this water remains, as noted earlier, for many weeks following a flow event. Pool volumes estimated from field survey lie in the range 1–3 Ml, so that even at the peak discharge at the gauging station (9 m^3/s), a pool would take 5–10 minutes to fill and begin overflowing downstream. During the hour of flow monitoring, less than 30×10^6 l (and possibly only $10–20 \times 10^6$ l) was discharged into Fowlers Creek, so that 25–80% of this could be consumed by the filling of the four large pools (taking total volume of these to be 8×10^6 l) which lie along the Fowlers Creek section of the flood course. These estimates cannot be improved owing to the lack of a hydrograph record for the entire flow from which to integrate the total flow volume. None the less, channel storage in low-lying pools was evidently a major contributor to transmission loss in this event. This appears not to have been highlighted in prior studies of transmission loss. Many smaller low points in the channel, such as scour holes cut in earlier floods, also acted to abstract volumes of the flow.

The large and multi-thread channel of Fowlers Creek promoted a second form of transmission loss that is peculiar to sub-bank-full flows in this system. At several places, a branch of the flow had spilled from the low-lying course, by way of a cross-channel, into another of the channel filaments. However, the relatively rapid decline in stage ensured that the exiting flow only persisted for a short time period. Evidence of these escaping flows was traced for some distance down the channel filaments into which they had spilled. Some had travelled for up to 100 m, but none had persisted for sufficiently long to rejoin the main flow. These spills were thus lost in their entirety into the bed sediments. In larger floods, all channel filaments are active, and such complete losses do not arise.

Mud deposition in slow-moving water, especially in deep pools, must have acted to restrict seepage losses. Very heavy mud drapes, of 1–3 cm thickness, were observed in many places along the channel, but especially upstream. Water has a significant residence time in pools of substantial volume, and mud deposition is

Figure 5. View of a large pool of turbid water just upstream of site 10 on Fowlers Creek. Sediment bars in the foreground are heavily cloaked in mud deposited by the flow of 15 May 1998. Photo taken on 16 May 1998 (view upstream). Most of the woody debris visible was transported by earlier, larger flow events

extensive. There would be less mud deposition during high-stage events, when the reverse bed slope at the downstream end of a pool is fully submerged and flow-through velocities are high. This has been observed in major floods and later confirmed by the presence of sandy dune bed forms left in these reaches after high-stage events. Therefore, flow-synchronous mud deposition must have acted to seal off pore spaces leading into the coarser channel bed sands and gravels, and so limited the rate of transmission loss. There was also some flow-synchronous mud deposition in areas of free flow away from pools, as indicated by extensive fields of mud ripples observed after flow recession.

A final characteristic of the sub-bank-full flow that contributed to a distinctive pattern of transmission loss was the load of leaves and other fine particulate organic matter (FPOM). Large flows in Fowlers and Homestead creeks carry considerable amounts of coarse particulate organic matter (CPOM), largely tree branches from the river red gums (Graeme and Dunkerley, 1993). The sub-bank-full flow remained at a stage below most of the debris jams left in trees where CPOM lodges. However, it was significantly affected by the behaviour of the FPOM. Obstacle clasts on the bed lead to stoss-side accumulations of leaves, which were packed tightly together, and turbulence created by these caused the excavation of small scour holes in the bed. Additionally, shallow flows spilling over wide planar sections of the bed swept up large quantities of FPOM that had accumulated by leaf fall on the dry bed, and this built up into extensive barriers on the stoss side of obstacles, such as growing trees or CPOM. In consequence, many small impoundments were created by the FPOM, and these served essentially to lower the rate at which the flow front progressed along the channel, thereby increasing the opportunity time for seepage losses into the bed.

DISCUSSION AND CONCLUSIONS

The limited data presented above indicated that the sub-bank-full flow had suffered relatively high rates of transmission loss on average along the channel, but has nevertheless persisted over many km at a very low stage (average depths through the last 4–5 km lay in the range 10–20 cm). These apparently contradictory findings can be resolved by interpreting the transmission loss in this event as displaying very marked spatial variability. All low-lying parts of the channel (including the major pools) that had to be filled and which then overflowed, must be reckoned as abstractions from the available flow volume. These losses would clearly amount to a larger fraction of the total volume carried by a sub-bank-full flow of 1–2 hours duration than from a bank-full flow of longer duration. The pools remain after the cessation of flow, and are consumed by evaporation and seepage over the ensuing weeks, leaving mud drapes along the bed and bank which clearly indicate the pool water level.

The other process that evidently promoted high transmission loss rates was the spillage of flows into channel filaments when low intervening bars were overtopped, presumably near the time of peak stage. Rapid flow recession ensured that these escaping flows were cut off soon after, providing insufficient discharge in the flanking channel thread to enable the flow to travel downstream and rejoin the main flow. These lateral spillages also provided, therefore, efficient and complete abstraction of a part of the available flow volume.

Elsewhere, though, the shallow flow, often following a stable, relatively moist and densely packed thalweg channel, commonly delimited by steep banks, appears to have been conveyed with relatively little loss. The flow stage shows remarkably small downstream decline in reaches free of pools. Clearly, no overbank losses were involved, and though affected by barriers of FPOM, the major channel barriers composed of CPOM did not significantly retard the shallow flows.

We therefore conclude that there were marked variations in transmission losses along the channel. For a bank-full flow event in which all channel filaments carried flow, and where all pool abstractions would rapidly be satisfied, a much more regular downstream pattern of decline would be expected, principally controlled by the rate of seepage into the bed and banks. This is indeed what the data of Dunkerley (1992) suggest. In other words, transmission losses in large events are governed primarily by the water uptake into the channel perimeter, and by the downstream fining of bed and bank materials, which progressively changes the porosity and hydraulic conductivity of these materials. This accords with the view of Reid and Frostick (1997) that the key determinants of transmission loss rate are the porosity and depth of the channel fill. In contrast, the evidence reviewed here suggests that in sub-bank-full events, local aspects of channel geometry, and especially two processes of water abstraction (pool abstraction and interfilament spillage abstraction), play a proportionally much large role. Indeed, in these low-stage events, bed and bank properties become relatively less important for two reasons: first, very little of the banks is wetted by the shallow flow (typically only the lowermost 20 cm in the present study), and secondly, because heavy mud drapes often restrict seepage into the bed and bank materials. It seems, therefore, that sub-bank-full flows are influenced by one set of factors tending to limit transmission loss, and another set that acts to accentuate loss. The aggregate outcome in the flow event studied here was an accentuation of losses, with a much higher rate of transmission loss (13.2% per km) for the sub-bank-full flow than has been estimated for bank-full and shallow overbank flows along the same channel (5–6·9% per km). Thus, localized abstractions from the flow at critical locations along the channel contribute greatly to the overall transmission loss, while in intervening reaches, losses may indeed occur no faster than in high-stage events, and may indeed proceed more slowly owing to the lower extent of bank submergence at low stage.

These findings have clear implications for the understanding of variability in transmission loss behaviour from event to event. They also suggest that convenient assumptions built into models simulating transmission loss may be unrealistic. For example, Faulkner (1992) adopted the following simple expression for the volume rate of transmission loss in Alkali Creek, in a semi-arid part of western Colorado

$$Tr = Q_t \, k \, t^{-1/2}$$

where Tr is the transmission loss rate (m^3/s) at time t after the start of flow, Q_t is discharge at time t and k is a proportionality factor, thought to be controlled by the channel width/depth ratio. This model was employed for a range of model discharges without the inclusion of any terms for the nature of channel flow and abstractions from available flow volume, including the role of bed form and sediment bars, whose affects were outlined earlier. It seems unlikely that any such simple expression will successfully embody the variation in factors determining transmission loss rates in events reaching different channel stages. Indeed, Faulkner (1992) presents a photograph in her paper that shows the flow front of a sub-bank-full flow bifurcating around a local high point on the channel bed. Given that the rate of loss of water into the bed will be maximized at or near the flow front, spatial variations in flow geometry of this kind cannot be ignored.

The dependence of transmission loss rate on flow stage appears not to have been systematically researched. Keppel and Renard (1962) demonstrated that at Walnut Gulch, in SE Arizona, transmission losses from channel flows increase approximately linearly with flow stage, at least until the whole of the bed surface is inundated. This was attributed to a steady increase in the wetted area of bed and banks as stage rises to this point. Beyond this, although they had no data, Keppel and Renard surmised that the rate of loss might flatten out because only the increasing pressure head would then be promoting faster water percolation. Presumably, at even higher overbank stage, losses are elevated further in the zone of the floodplain that becomes active as a sink. In this view, bank-full flows have no special attribute in terms of transmission loss behaviour. However, it has been shown here that sub-bank-full flows may also experience high rates of transmission loss, and in the case of the very small sub-bank-full flow investigated the rate is about double that for a bank-full flood in the same channel. It therefore appears possible that flows near bank-full may display lower transmission loss rates than either overbank or sub-bank-full flows. The reliable gauging data that would be required to test this hypothesis are unavailable but, if it is correct, bank-full flows may hold a key position in the sculpture of ephemeral channels akin to that known from perennial streams of the humid zone. In particular, lower transmission losses near bank-full stage would enable such floods to reach further down the stream system, and in this way be the most common agent of channel sculpture, at least in the distal reaches of the channel system, and perhaps along the entire stream course.

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