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PATTERNS OF SEDIMENT STORAGE IN THE KOWAI RIVER, TORLESSE RANGE, NEW ZEALAND

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ABSTRACT

Changes in sediment storage in two reaches of the Kowai River, Torlesse Range, near Springfield, New Zealand, were investigated using aerial photographs, surveyed sections and palaeoflow reconstruction. The different timescales of the techniques enabled identification of a range of controls over sediment storage. The morphology of a lobe of sediment in the upstream reach, coupled with information on sediment supply from upstream (Hoey and Hockey, this issue), suggest the presence of bed waves moving down the river. Sediment transfer through the upper Kowai is affected by basin morphology, specifically bedrock-confined transfer reaches and relatively unconfined sedimentation zones. The response of river channels to an event is dependent on the magnitude, order and separation of sediment-supplying and sediment-transporting events, which may not coincide.

INTRODUCTION

Patterns of sediment storage in upland basins are highly variable (e.g. Beschta, 1983a; Macklin and Lewin, 1989; Richards, 1993). The dominant sediment supply processes and the degree of hillslope-channel interaction change downstream, causing changes in sediment storage patterns. Several studies of these changes have been conducted in the Kowai River basin (Ackroyd, 1986,1987; Ackroyd and Blakely, 1984; Beschta, 1983a,1983b; Blakely et al.,1981; Hayward, 1980). The present investigation focuses on waves of bed sediment moving through the river system. A bed wave is defined as a change in the amount of bed sediment in a particular reach relative either to that reach at different times, or to adjacent upstream and downstream reaches at a given time. In the Kowai River valley shape varies considerably due to local geology, so bed waves are defined with respect to time. Changes in sediment storage patterns in two reaches of the Kowai, about 9-12km from its source, are analysed for periods of 44 years (the length of the aerial photographic record), 12 years (length of the cross-section survey record), and 2 years (length of record reconstructed using palaeohydraulic methods).

LOCATION

The Kowai River is located on the eastern side of the Torlesse Range, an easternmost range of the Southern Alps (Fig. 1). Maximum elevation within the basin is 1998m (Castle Hill Peak), declining to 275m, 30km from the source, where the Kowai joins the Waimakariri River. The bedrock is predominantly greywacke sandstone of Upper Triassic age, but facies of siltstone, mudstone and volcanic rocks also are present in this basin as elsewhere in this area (Andrews, 1974). Annual precipitation exceeds 1000mm throughout most of the basin (Blakely et al., 1981).

Ackroyd (1987) recognised two geomorphic terrains within the upper basin above the confluence with the Foggy River. Above 1400m the landscape is dominated by erosional landforms from late Pleistocene glaciations, especially relict cirques now partially infilled with colluvium. Below 1400m, the terrain is fluvially dissected, with steep slopes covered by thin (<0.5m) colluvial deposits. Stream beds are 5-10m wide, confined by bedrock walls and debris fans. Below the confluence of the Kowai with the Foggy River (Fig. 1) valleys are wider (up to 4km wide in the vicinity of Springfield) and infilled by fluviglacial gravels. The present Kowai River has incised into these gravels producing several degradational terraces. Five glacial episodes were identified in the Kowai valley by Marden (1976).

Below about 9km from its source, the Kowai River can be classified as wandering type II in the scheme of Carson (1984), with some reaches being braided (having multiple channels at all but the lowest discharges). High flows can occur in response to heavy rain at any time of the year, and to spring snowmelt. Snowmelt is not always sufficiently rapid to produce flows transporting bedload. Peak discharges in the nearby Ashley catchment are greatest in March-September (NCCB, 1982) and a similar pattern would be expected in the Kowai River catchment. Porters and Bridge Reaches (Fig. 1) form the basis of the following discussion. A bedrock gorge is located at 2.3-2.5km, and a constriction occurs at 3.7km where the No. 2 road bridge is located.

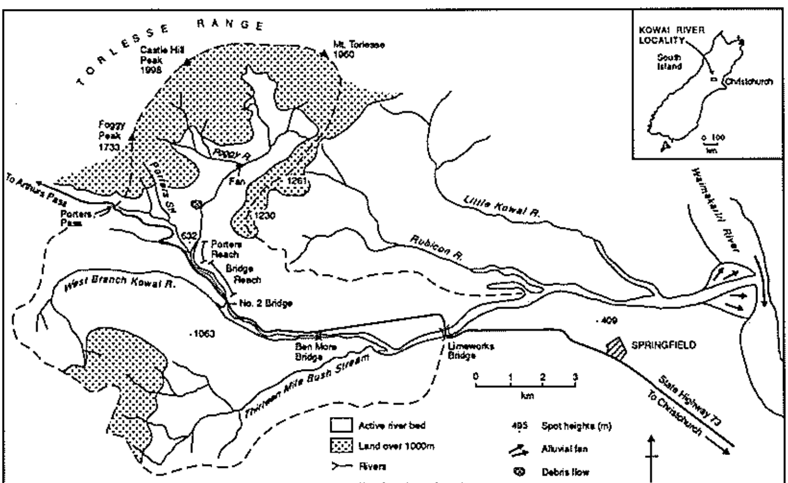

FIG. 1—Map of the Kowai River showing locations referred to in the text.

SEDIMENT STORAGE AT DIFFERENT TIMESCALES

Fluviglacial deposits and postglacial changes

Present river processes in the Kowai basin downstream of the Foggy River confluence are influenced by the locations of bedrock outcrops and fluviglacial deposits within the valley. Immediately upstream of Porters Reach, the Kowai passes through a bedrock constriction. Porters and Bridge Reaches are confined by fluviglacial terraces; Figure 2, based on maps prepared by Brunsden (1973) and Marden (1976) and on aerial photographs taken in 1987, shows the extent of these deposits in both reaches. Reconstructed valley cross-sections (Fig. 2) indicate the confining effect of the fluviglacial deposits and the magnitude of valley infill.

Blakely et al. (1981) suggested that widespread forest removal by fire between 500-1000BP induced a change from a meandering to a braided channel pattern, caused by accelerated sediment supply to the river and increased peak discharges. The process was repeated to a lesser extent following clearing of much remaining native forest during the 19th century. The active river bed was partially stabilised by tussock grasses and matagouri (*Discaria toumatou*) prior to c.1930, and since then by broom (*Cytisus scoparius*), blackberry (*Rubus fruticosus*) and gorse (*Ulex europaeus*).

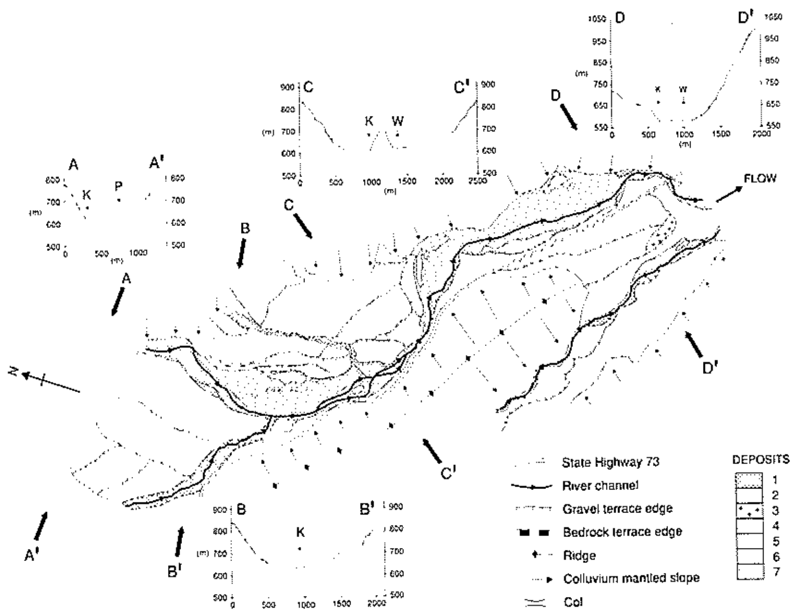

FIG. 2—Map of fluviglacial deposits in the vicinity of Porters and Bridge Reaches, based in part on maps by Brunsden (1973) and Marden (1976). Insets are valley cross-sections: their lines, but not precise limits, are shown by bold arrows. Letters on the insets refer to channel locations: K- Kowai R.; W- West Branch Kowai R.; P- Porters Stream. Inset vertical scale is m above sea level, and distance is m from the true left side of the valley. Solid lines are sections measured from 1:50 000 topographic maps augmented by field survey data. Dashed lines are best-fit parabolas for the rock slope parts of the cross-sections. Deposits are: 1) active river gravel; 2) vegetated floodplain composed of river gravels; 3) post-glacial terrace gravels; 4) Poulter gravels; 5) Blackwater 1 gravels; 6) Blackwater 2 gravels; 7) Otarama gravels.

Changes over a 44 year period, 1943-87

Beschta (1983b) used active channel widths determined from aerial photographs taken in 1943,1960,1972 and 1980 to assess channel response to changing sediment supply in this river. The active channel width is defined as the total distance across individual channels and exposed bed material, excluding vegetated islands within the channel. Assuming that increases in the active channel width indicate aggradation, Beschta identified zones of aggradation associated with valley constrictions, and in response to a storm of 150 year return period (Beschta, 1983a) in 1951. Measurements from 1965 and 1987 aerial photographs have been added to this data set.

Changes in the active channel width in Porters and Bridge Reaches over the period 1943-87 were standardised by division by the long-term mean value for each cross-section to remove the effects of variations in valley width. A paired t-test was used on the standardised data, following Beschta (1983b), to determine

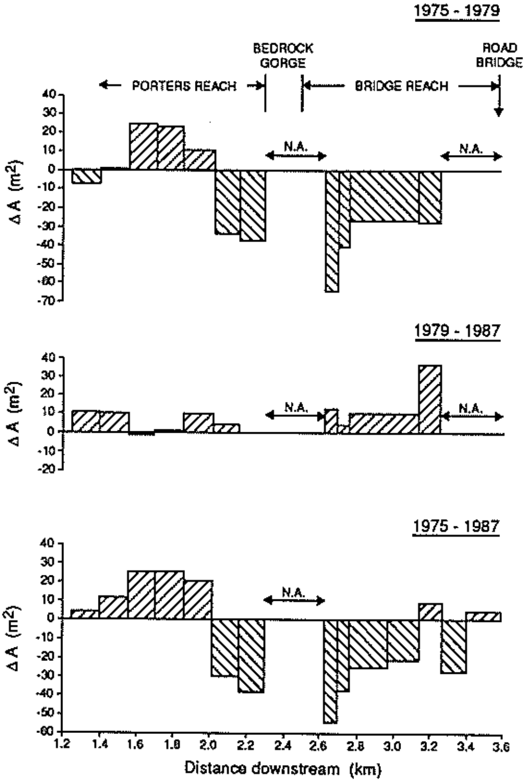

FIG. 3—Changes of area (A) of sediment stored in Porters and Bridge Reaches. Units are m^2. See text for details of calculations. N.A.- data not available. (a) 1975-1979. (b) 1979-1987. (c) 1975-1987.

significant reach-averaged width changes. In Porters Reach, significant changes were an increase in width of 42% (32m) during 1943-60 (α=0.01), and a decrease of 33% (38m) during 1965-75 (α=0.05). In Bridge Reach the only significant change was a widening of 37% (29m) during 1943-60 (α=0.01).

Changes in sediment storage since the 1951 flood were estimated by Beschta (1983a) from reconstructions of depositional surfaces formed in that event. This suggests net erosion of 59 000 m^3 in Bridge Reach over the period 1951-82, and net deposition of 13 000 m^3 in Porters Reach. The substantial erosion in Bridge Reach was not detected from changes in active channel width. Such changes are more rapid when they involve widening (aggradation) than narrowing (degradation), as the latter requires colonisation by vegetation of part of the bed. The 1965 aerial photographs show that much of the material deposited in Porters Reach during the 1951 flood was stabilised by vegetation during the period 1965-75. Since 1975 much of this material has been eroded, but there has been further deposition within the reach. In Bridge Reach, deposition in the 1951 event was followed by erosion of some of the accumulated material. Interpretation of this as a simple disturbance and relaxation cycle (Brunsden and Thornes, 1979) may be erroneous given the more complex behaviour of Porters Reach.

Changes over a 12 year period, 1975-87

Cross-sections were surveyed by staff of the Tussock Grasslands and Mountain Lands Institute in 1974/5 and 1979. Those within Porters and Bridge Reaches were resurveyed at least once during 1986/7. Volumetric changes between each adjacent pair of cross-sections were estimated by firstly calculating the change in cross-sectional area (ΔA, m^2) at each section within a given time interval. This enabled the cross-sectional averaged mean depth of aggradation/degradation to be calculated as $\Delta\bar{z} = \Delta A/w$, where w is section bank-bank width (m). Assuming that $\Delta\bar{z}$ and w both vary linearly between adjacent sections, and for sections i and j defining parameters $a = w_j/w_i$ and $b = \Delta\bar{z}_j/\Delta\bar{z}_i$, the change in volume of sediment stored between the sections (ΔV, m^3) can be calculated as

$$\Delta V = w_i \Delta\bar{z}_i L(2 + a + b + 2ab)/6 \qquad (1)$$

(Ferguson and Ashworth, 1992), where L is the distance between sections i and j. Where a and b differ from 1 this reduces bias in the result by about 15% compared with a simple end-area approach. The volume changes have been converted to equivalent mean area changes ($\Delta A = \Delta V/L$) for plotting (Fig. 3), such that the area of each bar on Figure 3 is proportional to the volume of aggradation/degradation between the bounding sections. Table 1 gives the reach totals of volume changes. Data are not available for some parts of the reaches (Fig. 3).

Net aggradation in Porters Reach between 1975 and 1987 (Table 1) occurred upstream of cross-section 136, and largely before 1979. The narrow sections (130 and 131) at the head of Porters Reach degraded during 1975-9 (Fig. 4a), but this is outweighed by aggradation at sections 132, 133 and 134. The surveys of sections 133 and 134 were truncated in 1975 at what was then the edge of a low vegetated terrace. By 1979 this terrace had been re-activated (it was no longer vegetated), necessitating the extended survey, and may have degraded following this re-activation. The degradation downstream of section 136 (Table 1) is due to general degradation at section 137 (Fig. 4a).

Table 1—Changes in sediment stored in Porters and Bridge Reaches, 1975-87 (units are m^3). All data rounded to nearest 100 m^3. Note that some Bridge Reach data covers the period 1975-87 directly without a 1979 survey. Summing 1975-9 and 1979-87 changes does not necessarily equal changes from 1975-87 calculations.

Reach	1975-79	1979-87	1975-87
Porters u/s section 136	+8500	+5300	+11300
Porters d/s section 136	-9700	+600	-9100
Porters -all	-1200	+5900	+2200
Bridge -all	-20700	+13300	-12000

Between 1979 and 1987 (Fig. 4b) most sections showed areas of both aggradation and degradation as channels were infilled and replaced by others. More general aggradation is apparent at section 134 where a lobe of sediment is most pronounced. The lobe also affected section 135, but no reliable comparison with 1975 or 1979 data is possible for this section. Downstream of section 136 there was little net change in this period, although there was a comparable pattern of channel infilling and new channel development. Similar patterns of cross-sectional change have been reported from both wandering and braided gravel-bed rivers (Laronne and Duncan, 1992; Werritty, 1984; Werritty and Ferguson, 1980).

In Bridge Reach, the two periods are associated with degradation and a lesser volume of aggradation respectively, although with some variability near the downstream end of the reach over the 1975-87 period (Fig. 3).

MORPHOLOGY OF A BED WAVE - PORTERS REACH 1987

During a flood in March 1987, the main low-flow channel switched from near to the left bank of the active bed to close to the right bank. This followed deposition of a lobe of sediment, approximately 500m long, in the upstream half of the reach. After the flood the leading edge of this lobe was 1-1.5m high, and the volume of sediment stored in it is estimated as 33 900m^3, equivalent to a mean aggradation of 0.82m. The bedform could be a wave of sediment moving downstream during the 1987 flood. The morphology and sedimentology of the lobe were investigated to examine its history and origin.

Bed morphology and sediments

Downstream of the leading edge of the bed wave are isolated areas of mature gorse and broom (Fig. 5). These lie less than 1m above the bed of the low flow channel in this area and appear unaffected by gravel transport. This suggests that bedload transport here was confined largely to the present low-flow channel during the event(s) which deposited the lobe upstream.

Surface and sub-surface sediment samples were taken from the constructional surface of the lobe and from other depositional surfaces within the reach. Surface samples were grid-by-number samples of 100 clasts > 2mm taken from within a 1m x 1m grid. Sub-surface bulk samples were collected from below the base of the

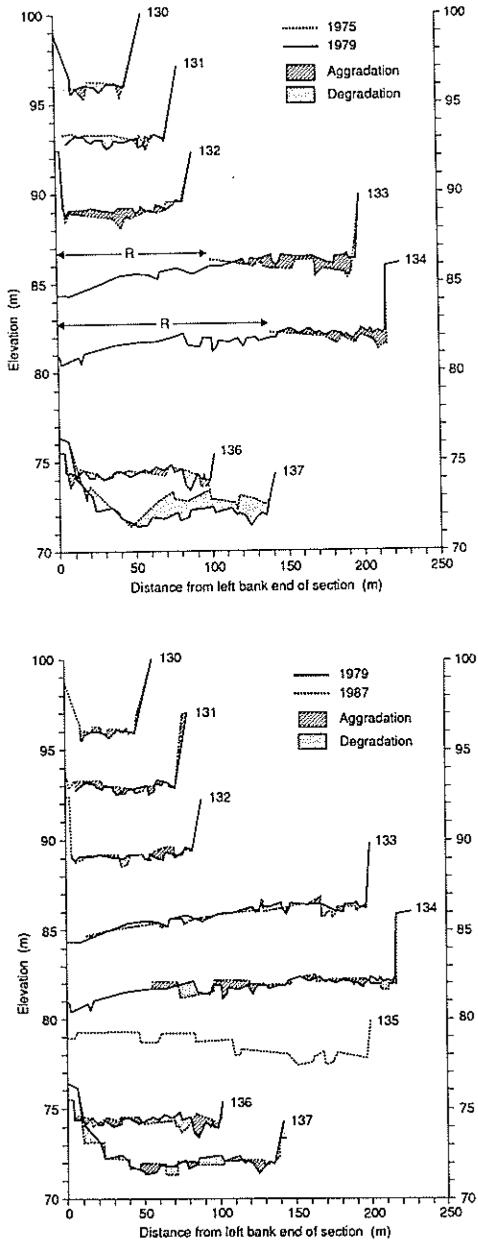

FIG. 4—Cross-section surveys from Porters Reach. (a) 1975 and 1979. (b) 1979 and 1987.

largest clast present in the surface material. No downstream changes in particle size were observed within the reach. Mean d50 for surface samples taken from the lobe was 19.9mm (standard deviation, s.d.=11.8 :N=14) compared with 25.2mm (s.d.=13.5 :N=15) for the non-lobe samples. Mean d90 values were 76.6mm (s.d.=32.4) and 93.2mm (s.d.=38.2), respectively. Neither d50 nor d90 values differed significantly for lobe and non-lobe samples (t-test). Bulk samples yielded d50 values of 10.2mm and 2.0mm respectively. This difference is not significant given that the largest clast in each of the two samples accounted for 5.3% and 3.7% of total sample mass. For adequate estimation of size parameters this percentage should be no greater than 1% (Church et al., 1987).

Depositional history of the sediment lobe

Clast orientation was measured at each of 18 sites, and the significance of the mean orientations assessed using Rayleigh's test. Orientation in the upper part of Porters Reach is broadly parallel to the nearest channel. Downstream of the Porters Stream confluence (Fig. 5) orientations were between 45° and 90° to the nearest channels, and are nearly perpendicular to the lobe front. These provide evidence of flow direction during deposition of the lobe.

Palaeohydraulic calculations of discharges associated with the deposition of unvegetated constructional surfaces within Porters Reach indicated that three events could have been responsible for their formation (Hoey, 1989). Two apparently stable surfaces downstream of the lobe corresponded to a discharge of 104 ± 26 m^3s^{-1}, which could be a flood in March 1986 or one later that year. Higher parts of the lobe and some other surfaces were associated with a discharge of 55 ± 23 m^3s^{-1}. Other surfaces, including some on the lobe, suggested a discharge of 22 ± 12 m^3s^{-1}. This latter material shows evidence of being deposited after the former but in close association with it, and may represent two stages of a single flood (Hoey, 1989). The third event is small (7 ± 1.5 m^3s^{-1}) and is likely to have affected only parts of the active bed close to the low-flow channel (Fig. 5).

These three events, coupled with lobe morphology and sediment data, can be used to postulate the stages of lobe deposition shown schematically in Figure 6. Prior to March 1987 the bed was a mixture of stable surfaces pre-dating the March 1986 flood, and dissected surfaces dating from that event and others later in 1986. Figure 6b shows the situation during the first part of the March 1987 event. The maximum downstream extent of bed disturbance is shown as the downstream limit of the lobe. In the latter part of the flood, flow directions altered (Fig. 6c) and the leading edge of the lobe was eroded. An absence of floods between March 1987 and the time when the reach was mapped suggests that the lobe was dissected during the latter part of the March 1987 event.

Sources of sediment for Porters Reach

There are few potential sediment sources between the confluence of the Kowai with the Foggy River and Porters Reach, about 2.2km downstream . The Kowai generally is less than 50m wide in this reach and is bounded by rock rather than gravel for over 50% of the reach. There are no major tributaries, but there is a single debris flow deposit (Fig. 1) of generally fine gravel about 1km upstream of Porters Reach. Most sediment entering Porters Reach is thus derived from upstream of the Foggy/Kowai confluence. Hoey and Hockey (this issue) calculated that bed

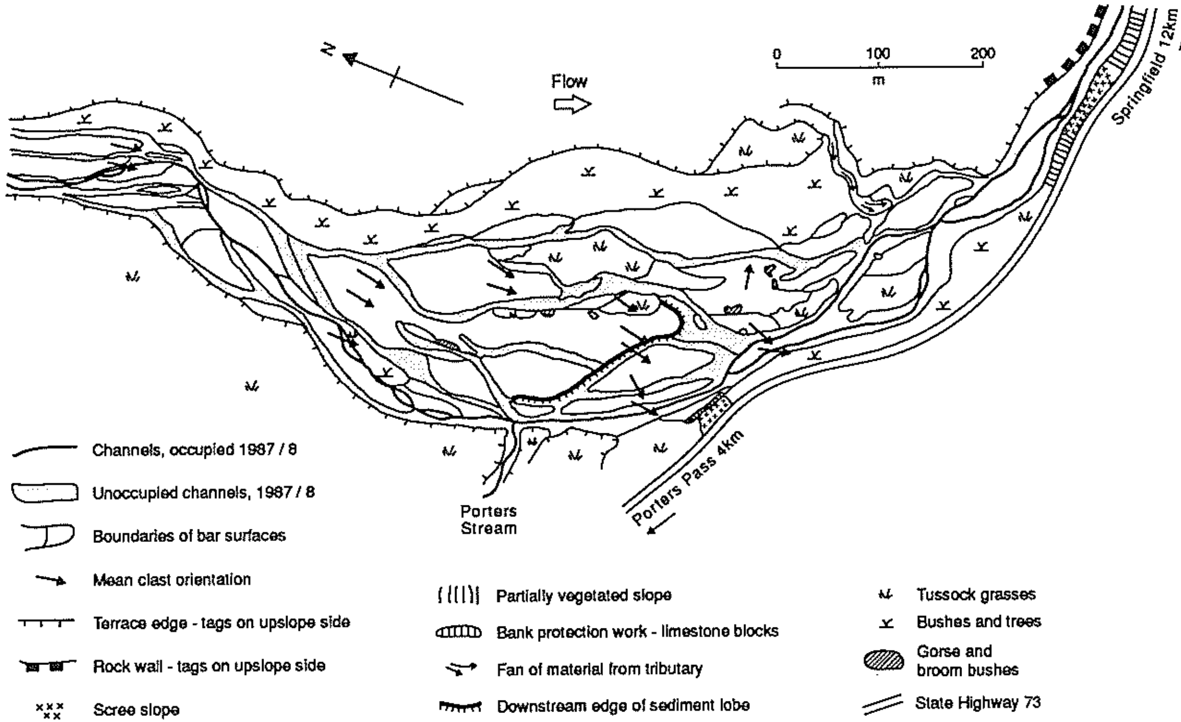

FIG. 5—Map of Porters Reach in late 1987, based on aerial photographs (August 1987) and field survey.

material discharges from the Foggy and Kowai Rivers are approximately equal. Previous work (Blakely et al., 1981; Beschta, 1983) has suggested that waves of sediment may enter the Kowai River system by a cyclical process of aggradation of the Foggy River fan, and its subsequent erosion by the Kowai. Given the relatively narrow active channel and therefore the absence of potential depositional zones between the confluence and Porters Reach, rapid transfer of any such wave to Porters Reach is expected. Evidence that the lobe of sediment observed in Porters Reach is derived from this mechanism is inconclusive. Firstly, sediment in the Foggy River fan is finer than that in the Kowai (Hoey and Hockey, this issue, Fig. 5), whereas lobe sediments are not significantly finer than non-lobe ones. Secondly, no jasper is observed in the Foggy River, but lobe sediments (Hoey and Hockey, this issue, Fig. 4, samples O,P,Q,R) contain significant jasper concentrations. This could be due to mixing between Foggy River sediment and Kowai sediment in the 2.2km between their confluence and Porters Reach. As an alternative, when the Foggy River fan is eroded, sediment accumulated behind it within the Kowai River may be released. The wave may therefore contain sediment from both catchments which mixes prior to entering Porters Reach. Either explanation could be supported by the observed increase in jasper concentration downstream of the confluence (Hoey and Hockey, this issue, Fig. 4).

The mechanism for release of waves of sediment into the Kowai downstream of its confluence with the Foggy River is plausible. Such waves will tend to become more mixed by both of the mechanisms explained above. On entering Porters Reach the wave is deposited (in whole or in part) as flow expands into the wider active channel bed in this reach.

DISCUSSION

Aerial photographs from 1966 show a sediment lobe in about the same location as in 1987. This suggests that the geometry of Porters Reach, which is confined by fluviglacial terraces and a bedrock gorge at the downstream end, determines the location of deposition. The active channel widens at the upper end of the reach and deposition may be promoted by decreased flow depth associated with the width increase. Blakely et al.(1981) classified Porters and Bridge Reaches as 'transition channels', differentiating them from 'transfer channels' both upstream and downstream. These are analogous to the sedimentation zones and transfer reaches of Church and Jones (1982). The bedrock gorge acts as a local control of base level and may limit aggradation downstream of section 136. The outlet from the gorge into Bridge Reach promotes renewed deposition. With another constriction at the road bridge at the downstream end of Bridge Reach, this reach also acts as a sedimentation zone.

Beschta (1983b) concluded that sediment loads influence active channel widths in the Kowai more than water discharges. Responses of the river to particular floods thus depend on the locations of stored sediment. The increased active channel widths in both Porters and Bridge Reaches between 1943 and 1960 reflect the influence of a very large flood in 1951. Subsequent responses have differed between the two reaches. The estimate from Beschta's (1983a) data of 13 000 m^3 aggradation in Porters Reach between 1951 and 1982 is only about twice as much as measured between 1979 and 1987 (Table 1). Vegetation subsequently stabilised low terraces formed during or soon after the 1951 flood, reducing the active channel width. The order of magnitude greater erosion from Bridge Reach

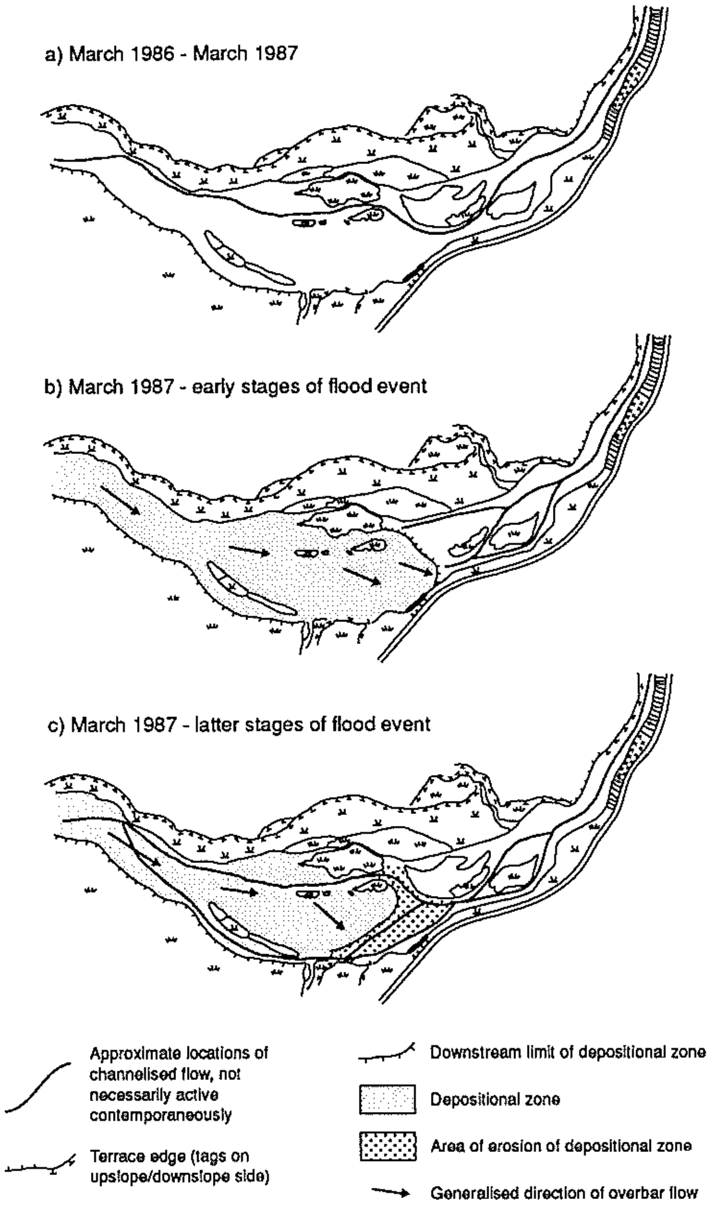

FIG. 6—Proposed evolution of the sediment lobe in Porters Reach during a flood in March 1987. Channel positions are approximate and their continuous occupation for the duration of the flood is not suggested.

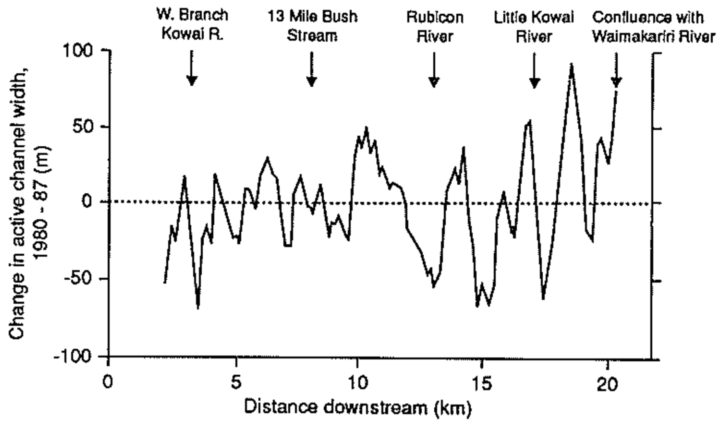

FIG. 7—Change in the active channel width of the Kowai between 1979 and 1987 aerial photographs.

(59 000m^3 or 0.39m of uniform aggradation; 1951-82), in the absence of significant changes in active channel width, indicates the magnitude of deposition in this reach following the 1951 event.

The effects of aggradation or degradation on a reach also depend on its location within the reach. Deposition within Porters Reach led to the sediment within the higher elevation deposits becoming immobile for at least 5 years (based on the dense tussock grass cover visible on 1975 aerial photographs). The deposition occurred towards the inside of the bend of Porters Reach where flood flow is less concentrated, due to reach geometry. Secondly, these terraces were sufficiently high to be colonised by vegetation relatively rapidly and thus partially protected from erosion. This is analogous to transferring from an 'active' to a 'semi-active' sediment storage reservoir in a scheme used by Kelsey et al.(1987) and Nakamura (1986).

Spatial controls on sediment delivery in the Kowai system include the location of the Foggy River relative to the Kowai and the greater availability of potentially mobile sediment within the Foggy River catchment. The different behaviour of Porters and Bridge Reaches and the importance of the rock gorge separating them are another example of spatial control. Sediment storage within low terraces in Porters Reach shows how reach geometry itself can influence sediment delivery from that reach. Responses to the large 1951 flood illustrate the importance of both the order and absolute magnitude of events in determining channel response. In a river where sediment availability largely controls channel response, this effect is exaggerated relative to situations where hydraulic controls dominate (e.g. Werritty and Ferguson, 1980). Sediment delivery thus is a response to time- and space-specific conditions, and the Kowai example lends support to Macklin and Lewin's (1989) rejection of the use of the concept of complex response to explain changes in sediment storage.

Locations of sedimentation zones in the Kowai are controlled by valley geometry, especially rock-confined reaches. Within this system, bed waves develop and propagate downstream. Beschta (1983b) identified a large bed wave introduced to the Kowai system as a consequence of the 1951 flood; it has subsequently migrated downstream and attenuated. Smaller waves have been identified at various locations within the catchment; the depositional lobe in Porters Reach is one example. This wave was introduced to the system upstream of Porters Reach, which is the furthest sedimentation zone upstream. Whether such a wave migrates downstream as a coherent form, attenuates, or dissipates depends on sediment storage patterns within the reach when it arrives. Wave generation and behaviour may alter through the catchment as a consequence of changes in the river channel. For example, between the Kowai/Foggy confluence and Porters Reach, the Foggy River fan and its interaction with the Kowai exerts a degree of control over wave development which declines downstream as the influence of rock gorges becomes more important. Griffiths (1989) has shown how bedload discharge rating curves can differ between adjacent reaches and produce differences in sediment storage. Over time these changes must be self-limiting if the river long profile is neither steepening nor declining. This is suggested as the mechanism by which adjacent sedimentation zones undergo cycles of aggradation and degradation. Because rock gorges inhibit upstream transfer of the effects of aggradation or degradation, waves introduced in Porters Reach will migrate downstream, with a tendency to attenuate as active channel widths and distances between successive transfer reaches increase.

The behaviour of waves further downstream in the Kowai is complicated by sediment supply from tributaries and erosion of fluviglacial deposits, and by the reduced frequency of bedrock-confined reaches. If changes in active channel width are closely correlated with sediment storage changes, the changes in widths between 1979 and 1987 (Fig. 7) suggest irregularly spaced bed waves in the Kowai, as far downstream as its confluence with the Waimakariri River.

Patterns of sediment storage in the Kowai are dependent on a relatively high sediment supply and interaction between sediment sources and river channels. Using the classification of Whitehouse (1988), the Kowai is located in the Eastern Alps region. The upper parts of the Kowai basin may provide a model for sediment transfer in the non-glacierized parts of the Axial Alps. The Kowai results should apply throughout the Eastern Alps, and especially the Basin and Range sub-region. Sediment transfer processes become more dependent on hydraulic conditions further downstream, so sediment storage patterns change. While bed waves have been recognised in rivers of the Canterbury Plains (Griffiths, 1979), their behaviour may not follow that of waves in the Kowai, since there is little bedrock control and different external influences (e.g. Wilson, 1985).

CONCLUSION

Variations in valley width, as a consequence of local geology and fluviglacial deposits, cause zones of sedimentation and transfer in the Kowai River. A variety of scales of bed waves develop within this system in response to varying catchment geometry and sediment transfer processes. The complexity and variability of channel responses to events of different magnitude is well recognised (e.g. Beschta, 1983a; Macklin and Lewin, 1989), and the Kowai example illustrates the range of channel response likely in a particular environmental setting.

The presence of bed waves at different scales in the Kowai is well documented. Beschta (1983b) showed how the largest of these attenuate as they migrate downstream. It is less clear what happens to smaller waves. Attenuation, dissipation and translation all are possible, depending on the order, magnitude and separation of sediment supply and sediment-transporting events.

In the present study, a range of techniques have been used to assess changes in sediment storage and the controls over them. While general conclusions for the Kowai River may be applied to other basins in similar environmental conditions, the particular responses of a basin will be site specific. The use of different spatial and temporal scales is a way to investigate basin sediment storage behaviour without long-term detailed monitoring.

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