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CHANNEL CHANGES FOLLOWING STORM-INDUCED HILLSLOPE EROSION IN THE UPPER KOWAI BASIN, TORLESSE RANGE, NEW ZEALAND

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ABSTRACT

Channels of the 37 km^2 Upper Kowai Basin, Torlesse Range, New Zealand, aggraded from hillslope erosion during a 150-year storm in April 1951. These storm-induced deposits were reworked by streamflow and formed sequential deposits farther downstream. As distance downstream increased, the widths and lengths of the deposits increased and their depths decreased. The volume of sediment eroded since 1951 from channel storage provides conservative estimates of sediment yields attributable to the storm; these estimates ranged from 42 m^3 km^{-2} a^{-1} for the Torlesse Stream to 650 m^3 km^{-2} a^{-1} for the headwaters of the Foggy River. Sediment yields were influenced by the interaction between hillslope erosion and changes in sediment storage.

INTRODUCTION

Studies in New Zealand show that the linkages between hillslope erosion, sediment transport, and the resultant channel characteristics are complex and highly variable (O'Loughlin, 1969; Bennett and Selby, 1977; Grant, 1977; Mosley, 1978; Griffiths, 1979; Schumm, 1980; Blakely, *et. al.*, 1981). Even within a channel system, the identification of cause and effect is generally clouded by a large number of interacting factors (Leopold and Wolman, 1970; Schumm, 1977, 1980; Lisle, 1981; Pickup, 1981). This investigation of the Upper Kowai Basin was undertaken to identify primary sources of sediment, patterns of sediment movement through the stream system, and associated changes in channel morphology for the 30 years following a large storm in April 1951. It includes the analysis of selected debris flow and channel characteristics, and determination of changes in channel storage.

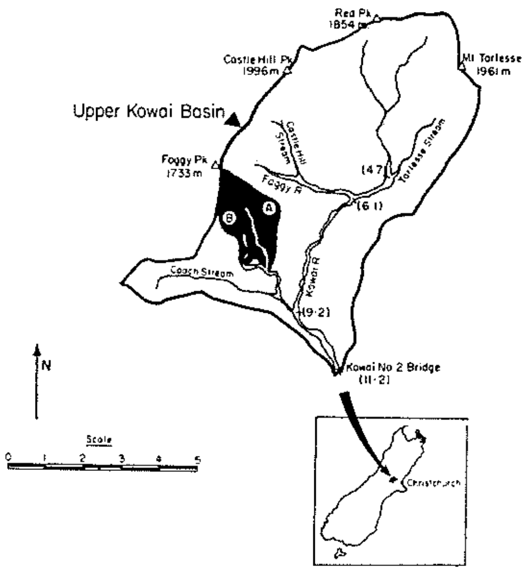

FIG. 1—Upper Kowai Basin, South Island, New Zealand. Downstream distances, in kilometres, from the drainage basin divide, are shown for selected locations along the Kowai River.

UPPER KOWAI BASIN

The Upper Kowai Basin (Fig. 1) drains the steep mountain slopes and hill country of the Torlesse Range near the north-western edge of the Canterbury Plains. Basement rocks of the Torlesse Supergroup are extensively fractured and sheared, and consist primarily of lithologically uniform, highly jointed, highly indurated, massive sandstones and mudstones of low metamorphic rank (Marden, 1976). The upper basin is cut by the historically active Porters Pass Fault; within a kilometer of its trace much of the basement rock consists largely of fault breccia. Gently sloping deposits of glacial outwash are found at lower elevations and have been incised by the Kowai River and its tributaries. Elevations range from 580 m at the mouth of the basin at the Kowai No. 2 bridge, up to nearly 2000 m on Castle Hill Peak.

Precipitation

Precipitation records for Mt. Torlesse Station were previously used to index precipitation trends for the Torlesse Stream area (Hayward, 1980). The 73 years of precipitation data available for this station (1909-1981) are also assumed to index precipitation in the Upper Kowai Basin.

Average monthly precipitation ranges from 70 mm in June to 100 mm in January; coefficients of variation for monthly amounts range from

45% to 70%. Annual precipitation averages 1032 mm a^{-1} at Mt. Torlesse Station with a coefficient of variation of 17%. There is no long-term trend in annual precipitation during the 73-year period of record. Autocorrelation analysis indicates precipitation during any given year is independent of precipitation during the previous year, but a cyclic pattern in a moving average of annual amounts is present (Beschta, 1983).

Most erosion occurs during discrete storms often lasting several days. A partial duration frequency analysis of the largest 73 storms between 1909-1981 showed that the difference in precipitation between 48 and 72-hour periods, for a given recurrence interval, was relatively small in comparison to amounts that occur within 24 hours or between 24 and 48-hour periods (Fig. 2). These results indicated that major storms are best indexed by precipitation that falls in 48 hours or less.

The seven largest storms—those with recurrence intervals greater than 10 years for both 24 and 48-hour storms—occurred between 1920 and 1959. The largest recorded storm occurred in April 1951 and had an estimated return period of between 125 and 160 years at Mt. Torlesse Station; however, it is likely that the return period was greater in parts of the basin. The storm was centered near the Torlesse Range and probably had a strong orographic component in the Upper Kowai Basin because it was an easterly storm. This large storm had a major impact on hillslopes and channels.

Mass Erosion

Various types of mass movements, such as those described by Cuff

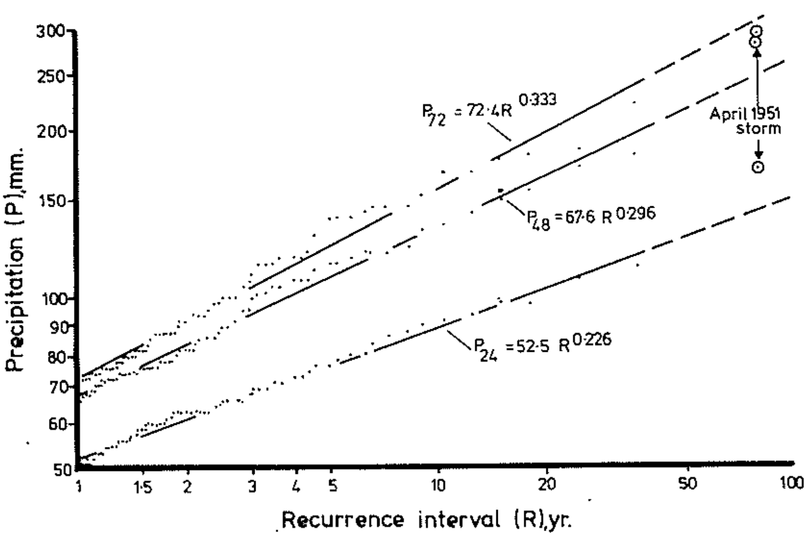

FIG. 2—Partial-series frequency analysis of the 24 (P_{24}), 48 (P_{48}) and 72 (P_{72}) hour precipitation for Mt. Torlesse Station, using 73 years of record from 1909 to 1981.

(1981) for the Upper Orari Catchment, occur throughout the Upper Kowai Basin. However debris flows during major storms are the dominant mechanism by which sediment is supplied to channels (Hayward, 1980). Some mass failures on hillslopes have moved sediment directly into the Kowai River, but in other instances, sediment has been carried downslope only to be deposited on alluvial fans, in topographic depressions, or on river terraces, and thus has had little or no effect on river sediment loads. Where channel aggradation has occurred, additional sediment has become available as the river has cut laterally into formerly inaccessible soils, parent materials, glacial deposits, and scree slopes (Blakely *et al.*, 1981). Thus, the temporal and spatial addition of sediment to the stream system has been extremely variable. Since 1951, the Kowai River and its tributaries have extensively reworked the deposits of the 1951 storm.

Within the Upper Kowai Basin, hillslopes in the Castle Hill Stream and Foggy River catchments have been actively eroding. Although aerial photographs from 1948 indicate that mass movements—predominately debris flows—were common in the Castle Hill Stream drainage prior to 1951, the storm undoubtedly increased the volume of sediment available to the channel system. Hillslopes in the Foggy River catchment have also actively eroded, but the earliest aerial photographs of this area are for 1960, so it was not possible to determine whether sediment movement in the Foggy River catchment was largely initiated by the 1951 storm or had been previously under way.

During the April 1951 storm, large debris flows occurred in two small basins, on the south-east side of Foggy Peak (Fig. 1). Basin A has a drainage area of 1.7 km^2; Basin B covers only 0.7 km^2. In both basins, deposition extended downslope to Porters Stream. Aerial photographs indicate that the erosional and depositional portions of the debris flows have remained relatively unchanged during the last 30 years. Because these debris flows are also typical of those occurring in nearby Castle Hill Stream and Foggy River catchments during 1951, detailed field measurements were undertaken in 1982 to estimate the volumes of soil and rock transported downslope. Cross-sections of the erosional and depositional zones were surveyed, along with hillslope topography immediately adjacent to the debris flows, so that a pre-1951 topographic surface could be estimated for each debris flow. By comparing surfaces, changes in cross-sectional area and volume were estimated.

Slopes in the erosional portion of each debris flow are generally steeper and more variable than those where deposition occurred (Fig. 3). This variability apparently reflects the uneven depth of fractured and weathered rock available for downslope movement. Surface gradients of the deposits are typically less than 15° and tend to decrease uniformly downslope.

The debris-flow deposit in Basin A consists of about 425,000 m^3 of material; slightly more material (about 5%) was estimated to have been deposited than removed from upslope. This discrepancy probably reflects uncertainties in estimating the pre-1951 topographic surface rather than a change in bulk density of the materials. Approximately 4,000 m^3, or only one percent of the total flow volume, has subsequently been eroded from the toe of the deposit by Porters Stream.

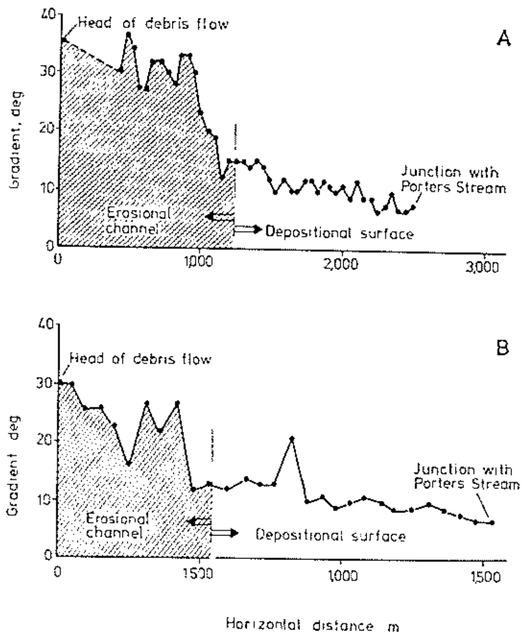

FIG. 3—Gradient of erosional channel and resultant depositional surface for two debris-flows (A and B) which occurred on the southeast slopes of Foggy Peak, April 1951.

In Basin B, the debris flow deposited 45,000 m³ of sediment; another 6,000 m³ were estimated to have been deposited in Porters Stream. Very little of the coarse sediment involved in either debris flow was carried to the Kowai River in 1951 or during the following 30 years.

At the downstream end of the erosional segments of the debris flows, 405,000 m³ and 57,000 m³ of sediment had moved downslope from Basins A and B, respectively. These volumes translate into basin average sediment yields of 500,000 and 160,000 m³ km^{-2}. If the point of reference used for erosion estimates is the downstream end of each basin where it joins Porters Stream, calculated average yields were less than 10,000 m³ km^{-2}. Farther down channel, Porters Stream joins Coach Stream and subsequently empties into the Kowai River. Where Coach Stream enters the Kowai River, a basin sediment yield of probably less than 1,000 m³ km^{-2} occurred as a result of the April 1951 storm. This example illustrates that high spatial and temporal variability is a major problem of interpreting sediment yields associated with mass movements.

Although numerous debris flows were initiated throughout the Upper Kowai Basin during the 1951 storm, factors other than precipitation,

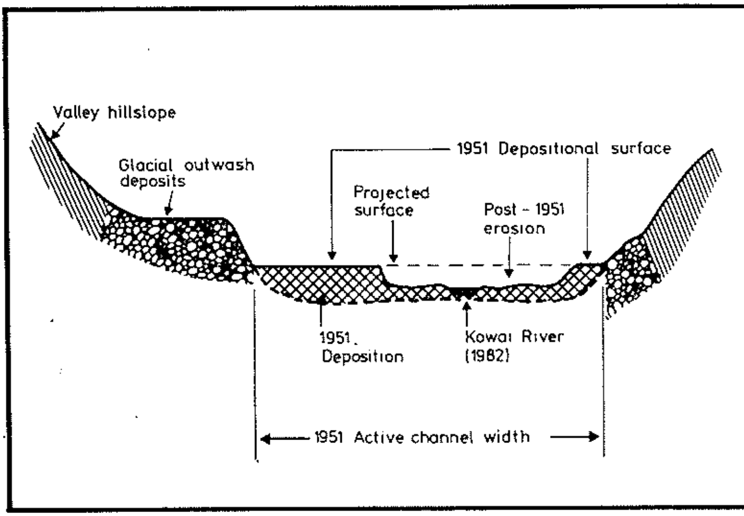

FIG. 4—Schematic diagram of 1951 depositional surface.

such as geologic faulting and vegetation changes, may also have contributed to slope failure. Numerous flows are located along Porters Pass Fault, where faulting has produced shattered and crushed bedrock prone to massive failure. The geologic conditions present in 1951 had existed for a long time; so, even though these subcatchments may previously have experienced as much precipitation as in April 1951, failure had not occurred. Many earlier failures are evident on adjacent portions of the Upper Kowai Basin; some of these may have been seismically triggered.

Major changes in the type of vegetation found on hillslopes and riparian zones of the Upper Kowai Basin and other mountains basins in central New Zealand occurred in the last 1,000 years (Molloy, 1963, 1964, 1967, 1969; and Molloy *et al.*, 1963). Prior to the arrival of the Polynesians, the Upper Kowai River Basin had a continuous forest cover of beech (*Nothofagus* spp.) up to an elevation of about 1300-1500 m, with alpine vegetation higher. Heavily weathered soil profiles provide evidence of a lengthy period of soil stability. The forest cover was converted to tussock grass in two stages. Early fires, 500 to 1,000 years ago, replaced beech forests with scrub species (principally *Dracophyllum* spp.) and tall tussock grasses (*Chionochloa* spp.). Repeated burning and grazing during the last 130 years of Eoropean influence reduced the scrub and increased the short tussocks and exotic sward-forming species (Molloy, 1967; Dick, 1978; O'Connor, 1982). Thus, trees and scrub with relatively deep and woody roots were replaced by low-growing vegetation with shallow, fibrous roots providing little mechanical strength to soil profiles. Today, only isolated pockets of beech forest remain in protected guts and gullies of the Upper Kowai Basin.

Conversion from forest to scrub to grass would not change infiltration capacities, unless the soil surface became exposed. Burning can cause bare soils and hydrophobic conditions, but such effects would be temporary. Thus, alteration of the hillslope hydrology from predominately subsurface flow to surface flow is unlikely. Hayward's (1980) data indicate that, even for bare ground, wet-soil infiltration rates generally exceed extreme rainfall intensities. Accelerated surface runoff is an improbable mechanism for initiating the debris flows experienced in 1951.

However, reduced root strength, caused by conversion of hillslope vegetation from forest to grass, could increase the potential for mass soil movements (Pierson, 1980). If the results of Ziemer's (1981) work in North American forests are applicable to New Zealand, scrub species can maintain soil strength (based on *in situ* shear tests) at or even above that experienced under native forest vegetation. Thus, when the forests and scrublands were converted to tussock grasslands, a lowering of geomorphic thresholds related to mass soil movements might have occurred. Reduction in evapotranspiration and concurrent increases in levels of saturation in the deep regolith could also reduce shear strength.

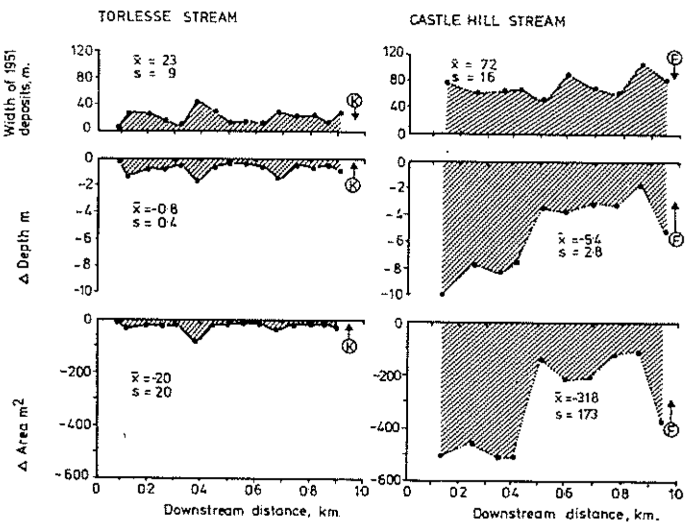

FIG. 5—Width of 1951 channel deposits along the Torlesse and Castle Hill Streams. Channel erosion from 1951 to 1982 shown as changes in average depth and cross-sectional area from the projected 1951 depositional surface. Torlesse Stream joins the Kowai River at "K"; Castle Hill Stream joins Foggy River at "F".

CHANNEL CHANGES

Extensive channel aggradation occurred during the storm of April 1951. The active channel commonly aggraded across the entire valley bottom, being laterally constrained only by valley slopes, bedrock or glacial deposits. The 1951 depositional surface (Fig. 4) was still identifiable in 1982. This remnant 1951 surface provided a reference from which to evaluate changes in channel characteristics and sediment storage during the last 30 years. The 1951 surface was formed by debris flow deposits in the steeper tributaries; fluvial processes dominated in the lower-gradient channels and the Kowai River. Subsequent stream erosion of these deposits left matched terraces along channels. The change in cross-sectional area since 1951, and hence the change in storage, could be determined from field surveys. Similarly, the amount of deposition could be determined where aggradation occurred on top of, but did not completely cover, the 1951 surface. In 1982, channel cross-sections were surveyed at 100 and 200 m intervals along the Kowai River and at 50 to 100 m intervals along Torlesse Stream, Castle Hill Stream, and Foggy River tributaries.

Torlesse Stream

The width of active channel during the 1951 storm averaged 23 m along the lower 0.9 km reach of Torlesse Stream and its true right-hand branch (Fig. 5). Because the active channel was confined by adjacent hillslopes, a large percentage of bed material was transported through the reach into the Kowai River.

By 1982, Torlesse Stream had largely eroded its 1951 deposits down to a coarse boulder lag. Lenses of finer materials remain at various locations along the channel. The changes in depth shown in Figure 5 represent the average depth of degradation of eroded portions of the 1951 deposits. The maximum depth of degradation at individual cross-sections would often be twice as great as the average values shown in Figure 5. Apparently less than 50% of the sediment deposited by the 1951 storm has since been excavated by the stream and routed to the Kowai River.

Cross-sectional surveys along the downstream 0.9 km of the Torlesse Stream and its true right-hand branch indicate that approximately 16,000 m^3 of sediment have been eroded from the 1951 deposits and transported to the Kowai River. A 1951 surface was not apparent farther upstream of this reach. Approximately 25,000 m^3 of sediment (including 9,000 m^3 from the true left-hand branch) have been transported out of the Torlesse catchment by erosion of the 1951 deposits. Much of this volume was probably removed during the 1951 storm after maximum aggradation had been attained. Substantial fan deposits at the mouth of the Torlesse Stream, which are evident on the 1960 aerial photos, indicate rapid erosion of the channel deposits must have occurred during the 1951 storm, or during several subsequent storms. Based on the size of the fan, 25,000 m^3 appears to be a reasonable estimate of sediment yield as a result of the 1951 storm.

For 1972 through 1977, Hayward (1979) found an average sediment yield of 30 t km^{-2} a^{-1} (15 m^3 km^{-2} a^{-1} assuming a density of 2 t m^{-3}) from the Torlesse catchment. During the six-year period, storms generally had recurrence intervals of less than five years. However, in April 1978, a storm with a return period of about 20 years occurred, producing 400 m^3 (800 t) of sediment in 70 hours. If the yield from the 1978 storm were distributed over a period equivalent to its 20-year return period, the estimated sediment yield would increase to about 20 m^3 km^{-2} a^{-1}. A sediment yield of this magnitude is similar to that measured from forested catchments in the South Island (O'Loughlin et al., 1978) and to bedload yields for many North Island rivers (Adams, 1979).

If the 25,000 m^3 of sediment eroded from the 1951 depositional surface along the Torlesse Stream were uniformly distributed over a 150-year period (the approximate recurrence interval of the 1951 storm), a sediment yield of 42 m^3 km^{-2} a^{-1} would result. This is an underestimate of total yield because it does not include sediment transported directly through the Torlesse channel during the 1951 event, and which did not contribute to the 1951 depositional surface. In addition, a large portion of all sediment mobilized during the 1951 storm still remains in storage along the Torlesse Stream and at other locations on hillslopes. Thus, the effects of storage (similar to that shown for the debris flows from Basins A and B on Foggy Peak) had a major influence on sediment yields from the mouth of the drainage basin. Pearce and O'Loughlin (1978) indicate that present rates of sediment yield from many regions of New Zealand probably more closely reflect the rate that sediment is being reworked from storage, rather than the rates of slope erosion. In this study, estimated sediment yields are less than rates of hillslope erosion because of storage effects.

The estimates of sediment yields for Torlesse Stream also raise the problem of how to distribute these volumes in time. I have chosen to use the storm recurrence interval as a basis for calculating average sediment yields from episodic events; thus, the effects of hydrologic events are weighted according to their frequency of occurrence. Without such weighting, measured or estimated sediment yields are difficult to interpret and compare, because they are strongly influenced by the sequence and magnitude of events during the study period.

Castle Hill Stream

The 1948 aerial photographs indicate that mass erosion was common in the Castle Hill catchment prior to 1951. Several debris-flow deposits, from 100 to 400 m in length, existed along the valley bottom in 1948. Aerial photographs taken in 1960 indicate extensive reworking of these deposits and other material supplied by the 1951 storm. For example, more than 1 km upstream of the confluence of Castle Hill Stream with the Foggy River, a large debris-flow deposit was extensively reworked by Castle Hill Stream between 1948 and 1960. These sediments passed through a gorge approximately 1 km upstream of the confluence. Another large debris-flow deposit, in the 0.5 km immediately below the gorge, is shown on the 1948 photograph. The 1951 storm added to that accumulation, but during the storm these deposits were breached by the stream and rapid incision followed. Another sequence of deposits then formed in the lower 0.5 km of valley, with the aggraded surface having an average slope of approximately 14°. The slope decreased to about 10° where the surface joined with similar deposits in the Foggy River. Thus, the 1951 depositional surface along Castle Hill Stream is a complex one, involving at least two depositional sequences during the storm. By extending the 14° slope of the second deposition zone through the upstream deposits (Fig. 6), I estimated that 120,000 m^3 of sediment had been removed from the upstream zone by Castle Hill Stream when aggradation in the lower depositional zone reached its maximum. Much of this sediment was deposited in the lower gradient valley immediately downstream of the first depositional zone; some may have been carried into the Foggy River. An estimated 142,000 m^3 of material has been removed from the second depositional zone, mostly during the 1951 storm itself. Distributing this later volume over a 150-year period gave a sediment yield of 550 m^3 km^{-2} a^{-1}. This value is somewhat greater than the estimated long-term erosional rate for this area of 400 m^3 km^{-2} a^{-1}, based on landscape evolution during the last 2.5 million years (Hayward, 1979).

Because the valley bottom along Castle Hill Stream aggraded in 1951, a condition which is still present in 1982, the 550 m^3 km^{-2} a^{-1} of sediment yield represents an underestimate of erosion rates. However, unlike deposits along Torlesse Stream, where the 1951 depositional surface resulted from one major phase, at least two sequences of aggradation occurred along the Castle Hill Stream channel. The sequential progression downstream of depositional zones made it more difficult to estimate yields than if the 1951 surface had formed simultaneously throughout the Upper Kowai Basin.

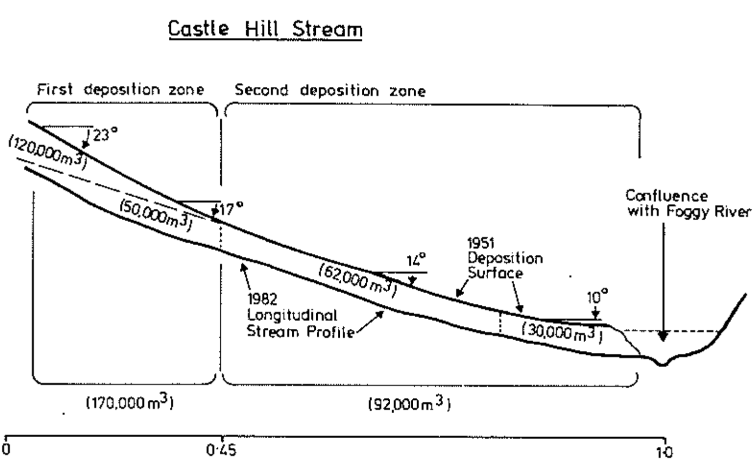

FIG. 6—Schematic diagram of net sediment volumes (m^3) eroded from a 900 metre section of Castle Hill Stream from 1951 to 1982.

The sequential formation of deposits along the channel gives the appearance of a "wave" of sediment working its way down the channel system. The dimensions of this wave are several orders of magnitude larger than those of the waves or slugs of bed material (Griffiths, 1979; Hayward, 1980; Beschta, 1981) that typically pulse through a channel system during periods of high flow. However, the individual slugs of bedload sediment represent the transport mechanism by which a deposit is initially formed and then subsequently reworked when upstream supplies are reduced. As a result, the zones of deposition do not migrate continuously as would be expected of a truly wavelike phenomenon; instead, they tend to form, be reworked by the stream, and then reform farther downstream.

The depth of aggradation within these zones decreases downstream. This is shown along Castle Hill Stream (Fig. 5) by the 8.4 m and 3.4 m depth of scour for the upstream and downstream deposits, respectively.

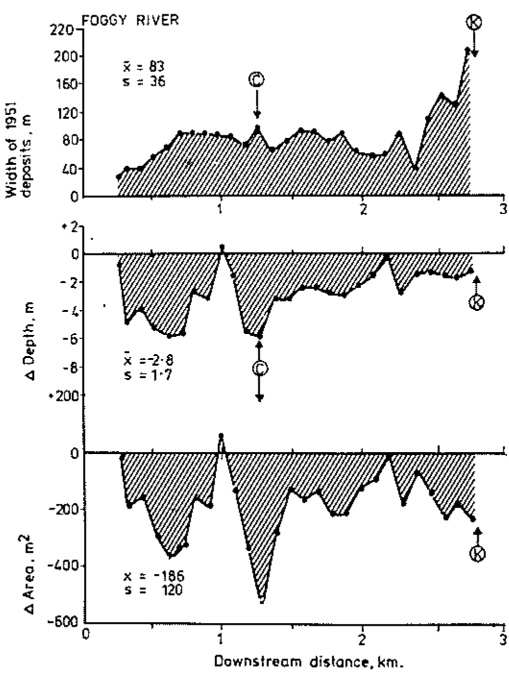

FIG. 7—Width of 1951 channel deposits along the Foggy River. Channel erosion and deposition from 1951 to 1982 shown as changes in average depth and cross-sectional area from the projected 1951 depositional surface. Castle Hill Stream joins the Foggy River at "C"; Foggy River enters the Kowai River at "K".

In general, the depth of degradation since 1951 also indexes the depth of aggradation that occurred in 1951.

Whereas the 1951 deposits along the Torlesse Stream covered a relatively narrow riparian zone to a depth of generally less than 2 m, aggradation along Castle Hill Stream was more extensive and much deeper, even though the catchment has less than half the area. Average width of the 1951 depositional surface along the lowermost 0.9 km of channel was over three times that found along Torlesse Stream (Fig. 5).

Foggy River

Deposition along the Foggy River channel in 1951 was also extensive. In addition to sediment from Castle Hill Stream, numerous mass failures from adjacent hillslopes fed soil and rock materials directly to the valley bottom. The 1951 deposits covered the valley bottom to an average width of 80 m; the maximum width of 210 m occurred where the Foggy River joins the Kowai River (Fig. 7). Because of a lack of pre-1951 aerial photographs, conditions in the Foggy River valley prior to the storm are not known; however, hillslope erosion prior to 1951 must have been extensive.

Interpretation of events along the Foggy River was hampered by the junction of Castle Hill Stream, approximately 1.5 km upstream of the Foggy River confluence with the Kowai River. Nevertheless, three depositional zones were identified. The first zone of sediment accumulation lies immediately upstream of the confluence of Castle Hill Stream and Foggy River. Since the 1951 surface formed, 195,000 m^3 of sediment stored along the channel have been removed and routed downstream. If this volume is distributed over a 150-year period, it represents a sediment yield of at least 650 m^3 km^{-2} a^{-1}. A minimum of 337,000 m^3 of sediment (142,000 m^3 + 195,000 m^3) passed the Castle Hill Stream and Foggy River confluence since the maximum level of deposition was attained in 1951.

The second zone of deposition is bounded at its upstream end by the junction of Castle Hill Stream and Foggy River and at the downstream end by a channel-constricting bedrock outcrop which causes a narrowing of the 1951 depositional surface at about 0.5 km from the catchment mouth (Fig. 7). From this 1 km reach, 210,000 m^3 of sediment have been removed by fluvial action since 1951. This volume is approximately two-thirds of the amount estimated to have entered the upstream end of the reach. Because the pre-1951 channel system and flood plan topography are largely buried by sediment laid down in 1951, the amount of deposition attributable to the 1951 storm could not be determined.

The third zone occupies approximately 0.5 km of Foggy River channel immediately upstream of its confluence with the Kowai River. Since this zone was formed, nearly 85,000 m^3 have been removed. Large volumes of 1951 sediment still remain in all three depositional zones.

Because the 1951 surface of the second and third depositional zones cannot be traced continuously through the gorge, it is not known whether they formed sequentially or concurrently. If both attained their maximum level of accumulation concurrently during the storm, then a minimum of 295,000 m^3 of sediment (210,000 m^3 + 85,000 m^3) was transported into

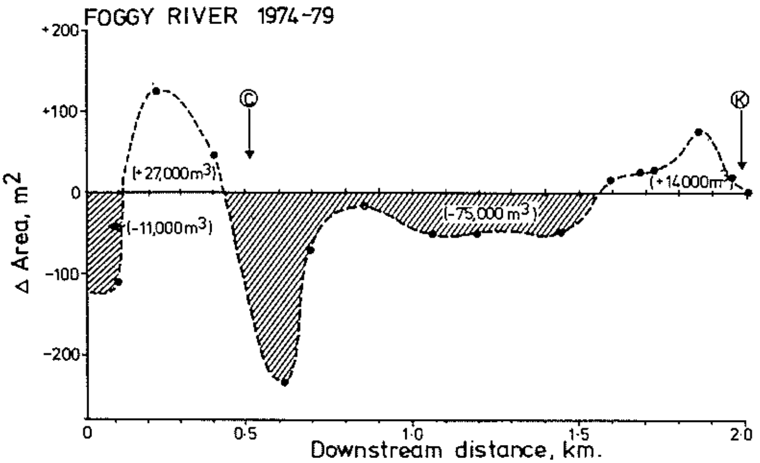

FIG. 8—Net changes in cross-sectional area of the Foggy River Channel deposits from 1974 to 1979. Volumetric changes in storage shown in parenthesis. Castle Hill Stream joins the Foggy River at "C"; Foggy River enters the Kowai River at "K".

Kowai River from Foggy River since 1951. This represents, over a 150-year period, an average sediment yield of 360 m^3 km^{-2} a^{-1}. However, if the second zone was the source of sediment for the third zone, the estimated minimum sediment yield is greatly reduced.

Between December 1974 and December 1978, the largest storm in the Foggy River catchment occurred in April 1978. Hayward (1979) estimates the storm had a return period of 20 years at the nearby Torlesse Stream, and slugs of bed material were observed to move through the lower reaches of the Foggy River (Hayward, 1980; Blakely *et al.,* 1981). Changes in surveyed channel cross-sections between 1974 and 1979 (R. J. Blakely and D. Adamson, unpublished data, Centre for Resource Management, Lincoln College) indicate that both aggradation and degradation occurred. Upstream of Castle Hill Stream, 27,000 m^3 of sediment accumulated, primarily from reworked 1951 deposits, of which only 11,000 m^3 of degradation appears in Figure 8. Immediately downstream of the confluence, degradation occurred. Along the 0.5 km reach upstream of the junction with the Kowai River there is aggradation, but volumes are only one-fifth of the influx from upstream channel sources. Thus, during the 1974-79 period approximately 60,000 m^3 of reworked channel deposits were released into the Kowai River from the Foggy River catchment. These data again illustrate the widely differing changes in storage associated with adjacent channel reaches.

As previously indicated, 27,000 m^3 of sediment were deposited along the valley bottom of the Foggy River upstream of Castle Hill Stream during the 1974-79 period (Fig. 8). However, only 5,000 m^3 of net deposition was found between 1951 and 1982 (Fig. 7). This would indicate that the Foggy River incised into the 1951 deposits along this reach prior to 1974, but then began net deposition during the period 1974 to 1979. The total volume of sediment stored along the Foggy River, upstream of Castle Hill Stream, probably exceeds 200,000 m^3. When the Foggy River breaches this depositional zone during high flow, another wave of sediment is expected to be released as the stream reworks the accumulated material.

Upper Kowai River

The 1982 cross-sectional surveys were also used to identify channel changes along the Kowai River associated with the April 1951 storm. However, because of the complexity of sediment movement through the Kowai River from its headwater and tributary channels, no attempt was made to estimate sediment yields at various locations along the channel.

The width of the active channel and resulting 1951 depositional surface was greatly influenced by bank characteristics. For example, where the Kowai River (Fig. 9) flows through a bedrock gorge at 4.2 to 4.6 km, widths of the 1951 deposits average only 25 m. Deposits widened to over 100 m immediately downstream of the gorge and occupied the entire valley bottom. Farther downstream at 7 to 9 km, active channel widths were approximately 60 m, again reflecting bedrock control. However, between 9 to 11 km, widths of deposits increased to 200 m, indicating

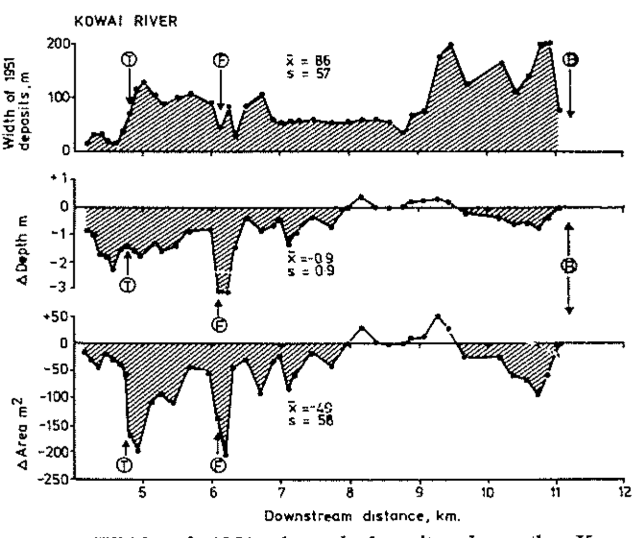

FIG. 9—Width of 1951 channel deposits along the Kowai River. Channel erosion and deposition from 1951 to 1982 shown on change in average depth and cross-sectional area from the projected 1951 depositional surface. Torlesse Stream and Foggy River join the Kowai River at "T" and "F", respectively; the Kowai No. 2 Highway Bridge is located at "B".

lateral migration of the braided Kowai River during the 1951 storm. Even though increased width is an expected consequence of an aggrading channel (Schumm, 1977), the magnitude of widening is also influenced by the characteristics of the riparian zone and channel banks.

Surveyed changes in depth and cross-sectional area, in relation to the 1951 surface (Fig. 9), indicate erosion of channel deposits between channel distances of 4 to 8 km, aggradation on top of the 1951 surface from about 8 to 9.5 km, and erosion from 9.5 to 11 km. Recent deposition has occurred at the 8 to 10 km reach of channel immediately upstream of the 9 to 11 km reach that showed extensive widening, hence deposition, during the 1951 event. Thus, the sediment eroded from hillslopes and reworked from the 1951 channel deposits during the last 30 years, appears to be accumulating upstream of the reach that showed the greatest widening *during* the 1951 event.

Whereas depths of degradation along Castle Hill Stream and Foggy River were generally 2 to 10 m, values for the Kowai River were typically less than 1 m. These results illustrate the tendency towards a progressive downstream decrease in depth of deposition, even though the total volume of material stored in each depositional zone is dependent upon a variety of factors, such as the amount of bed material available, magnitude of streamflow, and characteristics of the channel and valley. In addition, the relationship between depositional zone length (measured parallel to the channel) and catchment area (Fig. 10) indicates a lengthening of successive depositional zones downstream. Drainage area represents a surrogate variable in Figure 10, because it is usually strongly correlated with other catchment attributes and hydraulic factors related to sediment transport, such as channel slope and streamflow. By extrapolation of Figure 10, depositional zone lengths of the order of 40-60 km are estimated for the

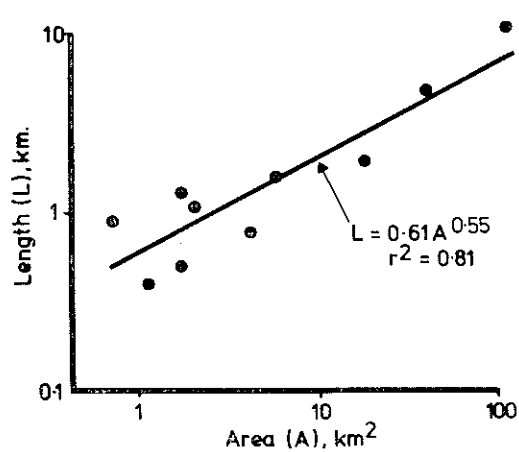

FIG. 10—Lengths of aggradational zones along the Kowai River and Upper Basin tributaries in relation to drainage area.

lower Waimakariri River, below the Kowai River. These distances would encompass most of the channel length of the Waimakariri River on the Canterbury Plain.

The widths associated with various depositional zones (Figs. 5, 7 and 9) also tend to increase in a downstream direction. As streams rework sediments from a particular depositional zone and route them downstream, a large proportion of the sediment may remain in storage. Unless the flux of sediment is augmented by bank erosion or sediment from adjacent hillslopes, each successive depositional zone will have less sediment. This may partially indicate why a downstream decrease in deposit depth might be expected, but does not explain a downstream increase in width or depth. Perhaps a change in particle size due to sorting or a decrease in channel gradient influences these dimensions. The dimensions of these zones may also be affected by local base levels, the spatial distribution of channel constrictions, influxes of sediment from bank erosion, and additions of both water and sediment from tributaries.

SUMMARY AND CONCLUSIONS

A downstream progression of depositional zones was created in the Upper Kowai Basin during the 1951 storm. Initial deposits were formed by debris flows and other mass failures. These sediments, along with those from adjacent hillslopes, incised glacial deposits and tributaries, were then reworked by high flows to form additional deposits farther downstream. Knowledge of this sequential development of depositional zones is fundamental for estimating sediment yields and understanding bed material transport through mountain stream systems. Although flow magnitude affects movement of bed material, sediment availability is an even more significant factor influencing the channel morphology of the Kowai River and its tributaries.

Results of this study extend to a relatively large basin the concepts of geomorphic thresholds and complex responses advanced by Schumm (1971, 1973, 1977) and demonstrated by Bergstrom and Schumm (1981) and Lyons and Beschta (1983). Although geomorphic thresholds are perhaps most often associated with hillslope processes, this concept may also be important in stream systems (Howard, 1980) where spatial discontinuities or transitions in channel characteristics occur. If the linkages between channel characteristics and fluvial processes can be clarified, predictive models of channel response may be developed, similar to stochastic and supply-based models for sediment transport (Griffiths, 1980; Van Sickle and Beschta, 1982).

Even though severe hillslope erosion in the Upper Kowai Basin was occurring prior to 1951, the April 1951 storm initiated wide-spread mass movements. Hillslope stability thresholds were obviously exceeded. Such thresholds can vary with rock type, faulting, weathering rates of soil and parent materials, and land use. The magnitude of the deposits and channel features created during the 1951 storm in the Upper Kowai Basin were exceeded only by the outwash deposits and terraces created thousands of years ago by fluvial-glacial processes. If hillslope erosion and associated channel deposition had been accelerated by the original burning of the forest vegetation some 500 to 1,000 years ago (Molloy, 1964, 1967, 1969), then the expected deposits and river terraces have been either obliterated or covered by erosion products of more recent origin.

Large spatial differences occurred in the amount of sediment moved during the 1951 storm. Channels aggraded and only a small proportion of the deposits have since been transported out of the Upper Kowai Basin by fluvial processes. Supply was further augmented along many sections of channel by erosion of streamside glacial deposits and scree slopes (Blakely *et al.*, 1981). Storage of sediment along valley bottoms and adjacent channel area reduced the export of bed material at the outlet of various basins. Because of this storage, estimates of erosion based on sediment yields would generally be too low for a basin undergoing a cycle of hillslope erosion, such as the Upper Kowai Basin. Conversely, when hillslopes stabilize, bedload yields measured at the mouth of a catchment would inflate the estimated rate of hillslope erosion because of sediment being removed from channel storage.

Depths of aggradation decreased downstream for most depositional zones, whereas lengths and widths increased. The flattening and lengthening of depositional zones resembles a spreading waveform of sediment moving downstream, and results from sequential aggradation and degradation of bed material as it is routed through the channel system.

This study provided additional perspectives regarding channel responses to sediment movement in New Zealand's mountain catchments. Although channel characteristics represent only state variables, an evalution of their changes over time can provide valuable clues regarding the importance of events that produced the existing channel system. It is imperative that measurements along the channel system be included in future studies, so that the significance of various factors influencing channel characteristics can be accurately assessed. The depositional process has been largely ignored in many sedimentation studies and represents an area of research that could provide important insights into channel responses from areas of active erosion.

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