Fluvial Geomorphology and Paleofloods in Arid Central Australia.

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March 1998

A thesis submitted for the degree of Doctor of Philosophy of the Australian National University.

Heavitree Gap, Alice Springs during the recession of the 1988 flood, the largest event in the gauged record.

Not in grow position! A Coolabah tree (Eucalyptus microtheca) indicates the magnitude of the paleofloods.

I certify that this thesis does not incorporate without acknowledgment any material previously submitted for a degree or diploma in any university; and that to the best of my knowledge and belief it does not contain any material previously published or written by any other person except where due reference is made in the text.

Mary Bourke

Mary Bourke
6/3/98

Acknowledgments

Many people have contributed to the supervision of this research. Firstly I would like to thank my 'official' supervisor John Chappell, who has in the last four months been tremendously supportive in getting the final thesis product together. I have enjoyed our many meetings to discuss in meticulous detail the ideas and concepts which have ended up in this thesis and indeed the other ideas which are reserved for later exploration. I have come to appreciate his hands off approach to supervision which facilitated my own need to explore the topic and discover the 'story' in central Australia. John was very generous in financially supporting the remote area field work and the dating program and also overseas conference trips. It has been a privileged experience to be one of his students and (unfortunately) one of the last Bio & Geo students.

Included in that experience has been the help and support of the technical staff. Eugene Wallensky supervised the organisation of the early desert trips and Damien Kelleher helped to get the equipment together for all six trips. Others include Jim Neale and Carlo Martinello. Keith Fitchett produced the detailed drainage maps from aerial photographs. The Help of Dave Buckle, and the photographers Bob and Darren is also appreciated. Marcia Murphy was a great support to me throughout my time here.

My unofficial 'supervisor' Geoff Pickup is thanked for writing the 1991 paper which inspired me to undertake this research. Geoff was generous with his time, and resources at CSIRO and introduced me to the paleoflood landscape in his back yard and the incredible beauty of the centralian desert. I have also benefited from discussions with Prof. L. Starkel and Dr. M. Garcon and Dr. S. Tooth on trips to the study area.

Others whom I have had good discussions of my work with include Jacky Croke, Bob Wasson, Laurie Olive, Dave Gillieson, John Jansen, John Magee, Pauline English, Vic Baker, Martin Williams, and Geoff Hope.

Field assistance was invaluable and Phil Ward and Blue Dugless (although he may dispute it!) were fantastic, digging, driving, cooking, and generally putting up with heaps from a stressed out student - fantastic desert memories.

In Alice Springs the field work was assisted by base support from CSIRO in particular Mr. G. Pearse, and T. Hobbs. The staff at PAWA, Peter MacDonnell, John Vischusen, Scot Balfore and Allen Russ. CCNT provided surveying equipment and R. Grant is thanked in particular.

Ian Lovegrove, the manager of the Todd River Station was a fantastic support - providing showers and a bed and some great meals in the middle of the long hot and dusty field seasons. He is thanked in particular for coming to the rescue during the '95 flood.

Jimmy Hayse and Kevin Pick are thanked for allowing access to sites and for the informative discussions on their experience living in the central Australian Landscape.

The Department of Geography and Oceanography provided the facilities to undertake the suspended sediment analysis of the '95 flood samples. N. Hill Dr. Spooner and Ms Questiaux are thanked for their great help and many hours of work on the dating program as are A. Murray, P. Wallbrink and J. Olley.

Dr. Jean Kennedy is thanked for her random act of kindness in proof reading this thesis.

A special thanks to my many friends who have provided the very necessary distractions to keep me sane. In particular Soren Blau is thanked for her positiveness and friendship, others include Gillian Aitken, Belinda Barnes, Sophie Bickford, Brendan Brooks, Judith Cameron, Jacky Croke, Chopsticks, Matt Duvall, John Magee, Nick Gill, Geoff Hunt, Al Paterson, Paul Rainbird, Bev Johnson, Janelle Stevenson, Niki Van Dijke, Mem Wilson, Eugene Wallensky. My housemates, Nicholas Hall, Janette Gilbert and Lisa Sedger - kept the home front running and were a tremendous support when needed. Thanks Guys!

My family may have wondered many times when I would ever finish and get a real job but never let on. They have been unconditionally supportive of me throughout the whole process and continue to be so. Go raibh mile maith agat.

ABSTRACT

This thesis examines the fluvial geomorphology of the Todd River (9,300 km^2) in arid central Australia. It describes the morphology, morphostratigraphy, sedimentary characteristics and chronology of the modern flood plain and paleoflood complexes, and discusses the geomorphic effects of high magnitude floods and the subsequent recovery and adjustment of the river.

The Todd River has two kinds of flood plains: confined and relatively unconfined. Both have a complex morphostratigraphy which reflects the variable discharge regime of this semi-arid to arid, summer rainfall region. The flood plains are episodically eroded and reconstructed during flood events and both processes may occur in different parts of the system during a single event. The principal mechanisms of flood plain destruction are channel widening, flood plain stripping, swirl pit, flood and back channel scour. The principal mechanisms of flood plain construction are the deposition of insets, flood plain veneer, channel and swirl pit fill, channel abandonment and the formation of simple or compound bars which can become islands. Channels in the relatively unconfined reaches tend to shift position by avulsion; channels in confined reaches change width and migrate laterally. The flood plains have stepped surface morphologies with prominent flood channels on their surfaces and back channels often occur between surface elements at different elevations. Swirl pits develop around large River Red Gums (*Eucalyptus camaldulensis*). The principal morphostratigraphic units include surface and back channel fills, channel and flood plain insets, surface and buried paleoflood and flood plain remnants, veneer and swirl pit fill. Morphostratigraphic models of confined and relatively unconfined flood plains are presented. The mud, sand and gravel flood plain sediments frequently are not deposited in simple fining upward sequences, as coarse sediments are transported across flood plains during floods. Buried erosion surfaces are common within flood plain deposits. Rippled bar-top and dipping bar face sand and gravel, flood couplets, sand sheets mud drapes and mud balls are typical sedimentary structures. Dating by radiocarbon and optically stimulated luminescence shows that sediment is stored in the active flood plain for only a few hundred years before being transported and redeposited. Reaches of temporary sediment storage sometimes are subject to extreme channel widening, often located upstream of bedrock constrictions or downstream of tributaries, or where the channel boundary is composed of erodable paleoflood sediments.

Paleofloods greater than observed historical floods flowed from gorges through the MacDonnell Ranges as sheet floods depositing transverse bar fields, and channels flowed across fans forming braided 12 km splays that include large longitudinal bars. Flows were diverted around bedrock ridges forming erosional moats, macroturbulent scour trains and 4-8 km wide, braided expansion bars with marginal channels, downstream from constrictions. High longitudinal dunes were eroded and the most northerly region of the Simpson Desert dune field remains as streamlined remnants. Slack water sediments were deposited in caves, tributary mouths and back water sediments were deposited in the longitudinal dune swales. Paleoflood deposits are widespread owing to repeated channel avulsion and inundate a larger area than if channels had remained fixed. The terminal floodout of the catchment has been repositioned at least twice. Dating indicates that the highest magnitude flows occurred between 14,000 and 4,000 BP decreasing in magnitude since the mid Holocene. The dating of paleoflood sediments by optically stimulated luminescence and radiocarbon has extended the paleoflood record of central Australia back to 27 ka BP but there is no indication of large scale flooding during the last major arid phase 24,000 to 15000 BP, and floods have occurred randomly since ~14,000 BP. Some paleofloods may be coeval with events in northern Australia but dating is not sufficiently precise to prove this.

The adjustment and recovery of the Todd River to the most recent paleofloods is ongoing today and the recovery time appears to exceed the paleoflood recurrence interval.

Notes and Errata

1. References to Slayter should read Slatyer (R.O. Slatyer).

2. All morphostratigraphic data in this thesis are derived from auger holes, hand-dug pits, natural exposures and a few excavator pits, as stated in the text. Typical pit exposures are illustrated in the Plates; stratigraphic columns from these sources are shown where relevant, eg., Figures 4.36 - 4.40. Morphostratigraphic sections (eg., Figure 4.8 - 4.27; 4.30 - 4.34) are based on these data but boundaries between sedimentary units are both interpolated between pits or auger holes and sometimes are extrapolated beyond them (eg. Figure 4.27), usually on the basis of observations at sites not on the section line, including morphologic features and stream or gully exposures. Thus, "hard line" boundaries at several metres depth should be regarded as inferred, unless explicitly crossing an auger hole or pit.

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CHAPTER 1

Introduction

1.1. Arid Zone Rivers

Arid and semi-arid rivers remain one of the least studied and understood fluvial environments (Reid and Frostick, 1997). This may be explained by the expense of field research, the remote location of study areas, the infrequency of river flows, the paucity of instrumented sites and the generally harsh and arduous working conditions. However, in the past two decades there has been a notable trend from a field dominated by descriptive studies (*e.g.* Mabbutt, 1977) to one of process investigations (*e.g.* Laronne and Reid, 1993) and detailed landform studies (*e.g.* Croke *et al.*, 1996). These investigations provide a greater understanding of the complex and diverse nature of arid landscapes in addition to providing insights into fundamental geomorphic processes that can be applied outside desert regions (Lancaster, 1994). The shift in research focus from descriptive studies to process investigations has been facilitated by the valuable data output from arid zone research stations such as Fowlers Gap in Australia (*e.g.* Pilgrim *et al.*, 1979: Dunkerley, 1992; Fanning, 1994), the Negev Desert in Israel (*e.g.* Schick, 1977; Lekach and Schick, 1982; Laronne *et al.*, 1992) and Walnut Gulch in south-western United States (*e.g.* Lane *et al.*, 1994; Nichols *et al.*, 1994).

Given that over three quarters of the Australian continent is classified as arid to semi-arid (Croke, 1997), it is striking that little is known of the geomorphology of fluvial systems in these regions. While some research has been done on catchments in the Lake Eyre basin (*e.g.* Williams, 1971; Nanson *et al.*, 1986,1988; Croke *et al.*, 1996, Pickup, 1991) and the semi-arid western New South Wales (*e.g.* Dunkerly, 1992) and Western Australia (*e.g.* Wyrwoll and Milton, 1976), large areas remain uncharted both geographically and conceptually. By way of example, previous work has found that a low rate of landscape evolution in the older more stable parts of arid Australia precludes the major landforms showing significant influence of the present climatic regime and that even the minor forms are inherited (Mabbutt, 1977). While this is still held true for the larger landscape elements this thesis will show that in the piedmont and desert margins, extreme rainfall events during Late

Pleistocene and Holocene times have left a profound geomorphic imprint on the fluvial landscape.

A study of an Australian fluvial system will yield an understanding of a different style of arid zone river from those studied elsewhere in the world because the controls on river behaviour in semi-arid Australia are unusual: long-term denudation rates and sediment yields are very low, the region was not affected by Quaternary glaciations or periglacial processes and the fluvial systems have low gradients.

1.2. Semi-arid Flood Plains

Accepted models of river behaviour, predominantly developed in temperate and humid areas, view rivers as being in a state of dynamic equilibrium between channel form, discharge and sediment input (Chorley and Kennedy, 1971). It is difficult to apply this concept to arid systems which only flow occasionally, with considerable variation between flow events, and may not display an obvious 'mean' or quasi-equilibrium. Arid zone rivers are more appropriately described as disequilibrium systems (Graf, 1988a). A recent classification of flood plains (Nanson and Croke, 1992) identified unconfined vertical accretion sandy flood plains as a common type in semi-arid zones but until now little was known of this flood plain type, its stratigraphy, morphology or formation. This study serves to redress this situation and presents a detailed description of the morphology, morphostratigraphy, sedimentary characteristics and morphodynamics of the Todd River flood plains in Chapters 3 to 5.

1.3. Paleofloods

The major impacts of climate on the Australian arid and semi-arid environment result from extremes: severe droughts, large floods, extensive fires and intense winds. It is likely that many aspects of this environment are adjusted to these extremes, and that geomorphic processes have little effect between major events. In arid Australia, high magnitude rainfall and floods profoundly change river channels, flood plains, lakes, wetlands and riparian vegetation and as extremes dominate much of the hydrology and fluvial morphology, it is essential that we have a clear understanding of their effects upon river systems.

Catastrophic processes have traditionally been considered as quite different from uniform processes and since the debates of the 17^{th} and 18^{th} century, Uniformitarianism has generally reigned as a geological paradigm (Mayer and Nash, 1987). Data presented in this thesis (Chapters 6,7, and 8) show that catastrophic paleofloods dominate the fluvial geomorphology in arid and semi-arid central Australia.

The ability of catastrophic floods to completely alter the geomorphology of a landscape has been well documented. A classic example is the floods of glacial Lake Missoula which formed the Channelled Scabland of eastern Washington, eroding extensive anastomosing channels into bedrock, streamlining the loessal topography and depositing large-scale gravel wave trains (*e.g.* Baker, 1978a). Estimated peak discharges were as great as 17×10^6 $m^3 s^{-1}$ (O'Connor and Baker, 1992). Similarly, much attention has been paid to the role of floods on river channels and flood plains (*e.g.* Schumm and Lichty, 1963). While floods have been recognised as dominant forces in arroyo channel development in the semi-arid south-west of the U.S. (Graf, 1983a), much of the research into the relationship between high magnitude floods and flood plains has focused on the immediate geomorphic response to a single event or clustering of events (Schumm and Lichty, 1963; Schick, 1974; Nanson, 1986). Their findings have shown that high magnitude events, although geomorphically effective (Wolman and Gerson, 1978), are episodic disturbances in the more gradual development of flood plains. While the conventional concept views flood plain formation as a result of gradual, incremental aggradation by both vertical and lateral accretion, in some environments (*e.g.* those dominated by high intensity rainfall) catastrophic events may be more important (*e.g.* Baker, 1977). In this thesis a series of flood plain characteristics is presented which indicates that flood plains in central Australia are formed predominantly by interplay between lateral stripping and vertical accretion during high magnitude floods (Chapters 4 and 5).

Central Australia contains one of only a few examples so far described (*e.g.* Bull, 1991) of large-scale arid zone fluvial landscapes deposited by rainfall-triggered floods downstream from confined gorge reaches, and presents a rare opportunity to study the characteristics and processes of these high magnitude events through an investigation of their morphology and stratigraphy. The orographic effect of the MacDonnell Ranges generates extreme, though rare, rainfalls which are routed through large drainage basins onto piedmont and alluvial plains. The low frequency of extreme flows increases the likelihood of paleoflood deposit preservation. Chapters 6 and 7 present the results of the geomorphic investigations and the chronology of these high magnitude floods, the oldest of which date to 27,000 BP.

1.4. System Recovery and Adjustment

In addition to the immediate geomorphic effects of extreme floods, the degree to which the paleofloods have generated the contemporary drainage systems is of relevance. Bourke (*in press a*) has found that channel planform and gradient are still recovering from the geomorphic impacts of events that occurred over one thousand years ago. Little is known of the nature, the degree and the spatial and temporal variability of continual adjustment of these central Australian streams. This aspect will be explored in Chapter 8.

1.5. Aims

1. To describe the morphology, morphostratigraphy, sedimentary characteristics and morphodynamics of the Todd River in arid central Australia.

2. To describe the morphology, morphostratigraphy, sedimentary characteristics and morphodynamics of the Todd River paleofloods and their deposits.

3. To determine the age of the Todd River paleofloods in order to infer patterns of Late Pleistocene and Holocene extreme rainfall variability from the Todd River paleoflood record.

4. To describe the geomorphic responses and adjustments of an event-driven arid zone river to impacts of high magnitude floods.

1.6. Criteria For Selection Of The Study Area

The Todd River in Central Australia was chosen for this research for the following reasons:

1. The Todd River rises in the MacDonnell Ranges which are subject to the southerly incursion of cyclonic depressions from the north Australian monsoon. Runoff from the rugged high relief catchment is relatively

efficient and high magnitude floods deposit sediment across the piedmont and the downstream alluvial plains. Preservation of the fluvial landscape is good as high magnitude flows occur infrequently and human impact is low.

2. The fluvial morphodynamics of modern arid rivers in Australia and globally have been neglected.

3. The potential for preservation of paleoflood deposits is high. Others have found that the patterns of erosion and deposition generated by high magnitude events take considerable time to become obliterated in the arid zone (Wolman and Gerson, 1978; Pickup, 1985); in particular Pickup (1991) has reported on the preservation of catastrophic flood deposits and landforms close to the MacDonnell Ranges.

4. The Todd River system was suspected to be dominated by event-driven disequilibrium behaviour and therefore suitable for the aims and objectives of this study.

5. Although it is a remote area, access is relatively easy along pastoralists' tracks.

1.7. Thesis Presentation

The thesis contains nine chapters. Chapter 2 introduces the study area. Chapters 3 to 5 are concerned with arid zone flood plains and present the data, interpret the processes and propose two models of arid zone flood plain formation. Chapters 6 and 7 describe the geomorphic effects of the Todd River paleofloods and their chronology. Chapter 8 describes the adjustment of the Todd River to the paleoflood effects. Chapter 9 is the conclusion to the thesis. Appendix 1 presents the chronology results and discusses the methods. Appendix 2 contains additional data on the Todd River flood plains and is the companion appendix to chapters 4 and 5. Appendix 3 is the companion appendix to the paleoflood chapters (Chapters 6 and 7). Appendix 4 contains twenty colour print-outs of enhanced satellite images.

CHAPTER 2

Introduction to the Study Area

2.1. Location, Geology and Soils

The Todd River drains from the MacDonnell Ranges in central Australia (133° 50', 23° 40') passing through the town of Alice Springs (Fig. 2.1). It is one of a number of ephemeral streams draining into the Lake Eyre Basin. The basin contains a wide range of river types and catchment sizes. These include large systems such as Cooper Creek and the Diamantina River (for detailed description see Rust and Nanson, 1986; Knighton and Nanson, 1993) which rise in the tropical region of Queensland. Smaller systems include the Neales River (Croke *et al.*, 1996) rising in the more arid western basin. The Lake Eyre playa is the depocentre for many of these internally-drained river systems (Magee *et al.*, 1995).

The Todd catchment has a drainage basin area of approximately 9,300 km^2 with two large tributaries draining from the north, the Ross River (1,260 km^2) and Giles Creek (1,000 km^2) (Fig. 2.2, Table 2.1). Elevations in the catchment do not exceed 1,000 m and local relief is less than 300 m.

The headwaters of the Todd and its tributaries rise in Proterozoic crystalline and metamorphic rocks of the MacDonnell Ranges (Shaw and Wells, 1983). Within the complex set of ancient fold structures in which granites, metamorphic and more recent sedimentary rocks are exposed (Quinlan and Forman, 1968; Jennings and Mabbutt, 1986) some rocks are heavily weathered and others are more resistant, providing a sediment load ranging from cobbles to clay (Pickup, 1991).

Soils in the ranges are predominantly shallow sandy lithosols with flanking outwash plains capped by red clays. South of the ranges, the soils are dominantly siliceous sand (Northcote and Wright, 1983). Soils in arid Australia are highly weathered and poor in nutrients with less than half the mean level of phosphorus and nitrogen observed in other similarly arid regions (Stafford-Smith and Morton, 1990). Several authors have recognised a wide diversity in the age of different soils in this part of arid Australia (Jackson, 1962; Litchfield, 1969; Mabbutt, 1967). Litchfield (1969) recognised several alluvial systems on the plains adjacent to the MacDonnell Ranges (Table 2.4) which illustrate the periodic development of this landscape and its soils. This history is summarised in Table 2.4 and is subsequent to an earlier period of deep weathering, laterization, and an irregular development of silcrete crusts. Litchfield suggests that the Todd valley has been deeply dissected and refilled at least twice in its subsequent history. The two stages of infill are related by Mabbutt (1966) to the building of piedmont terraces (the 'intermediate' terraces) and the subsequent etching on them of flood plains (the 'low' terraces) which are shallowly incised by present drainage, terminating in small floodouts. These cyclical landforms are accounted for by the Hermannsburg, Stuart and Amoonguna surfaces (Table 2.4). An extended multicycle history is evident from the additional older Yambah and Burt surfaces.

2.2. Climate and Surface Hydrology

The area north of 25°S experiences a summer-wet winter-dry climate. This is comparable with other tropical monsoon regions of Asia and Africa where the seasonal migration of the equatorial trough dominates the alternation of wet and dry seasons. Similarly, the seasonality of rainfall in tropical and subtropical latitudes in Australia can be explained by movements of the equatorial trough (Hobbs, 1973; 1975; Riehl, 1979). In the southern hemisphere summer, monsoonal weather, which is associated with violent thunderstorms, tropical depressions and severe cyclones, brings rainfall to the northern half of Australia when the equatorial trough lies at the extreme southern limits (Suppiah, 1992). Rainfall decreases rapidly towards the inland and reaches its minimum, less than 150 mm over the Simpson Desert in South Australia. Mean annual rain days over the northern half of Australia vary from less than 30 days in the Simpson Desert to more than 160 days on the eastern flanks of the Great Dividing Range in Queensland. On average, 50 to 80% of total rainfall between 20 and 25°S is received within the four month period December to March.

The general weather pattern of central Australia is controlled by the seasonal march of the pressure systems and its location in the centre of the continent (Slayter, 1962). Mean annual rainfall about the ranges is 274.4 mm (median 238 mm) and displays a high inter-annual variability (Fig. 2.3a). Mean diurnal temperatures range between 5.1°C and 19.4°C in winter and 20.7°C and 35.4°C in summer. During the summer months sporadic heavy rain may fall from convectional thunderstorms and comprises most of the annual rainfall (Slayter, 1962). Rainfall from tropical cyclones originating in Queensland, the north coast and western Australia occur less than once a

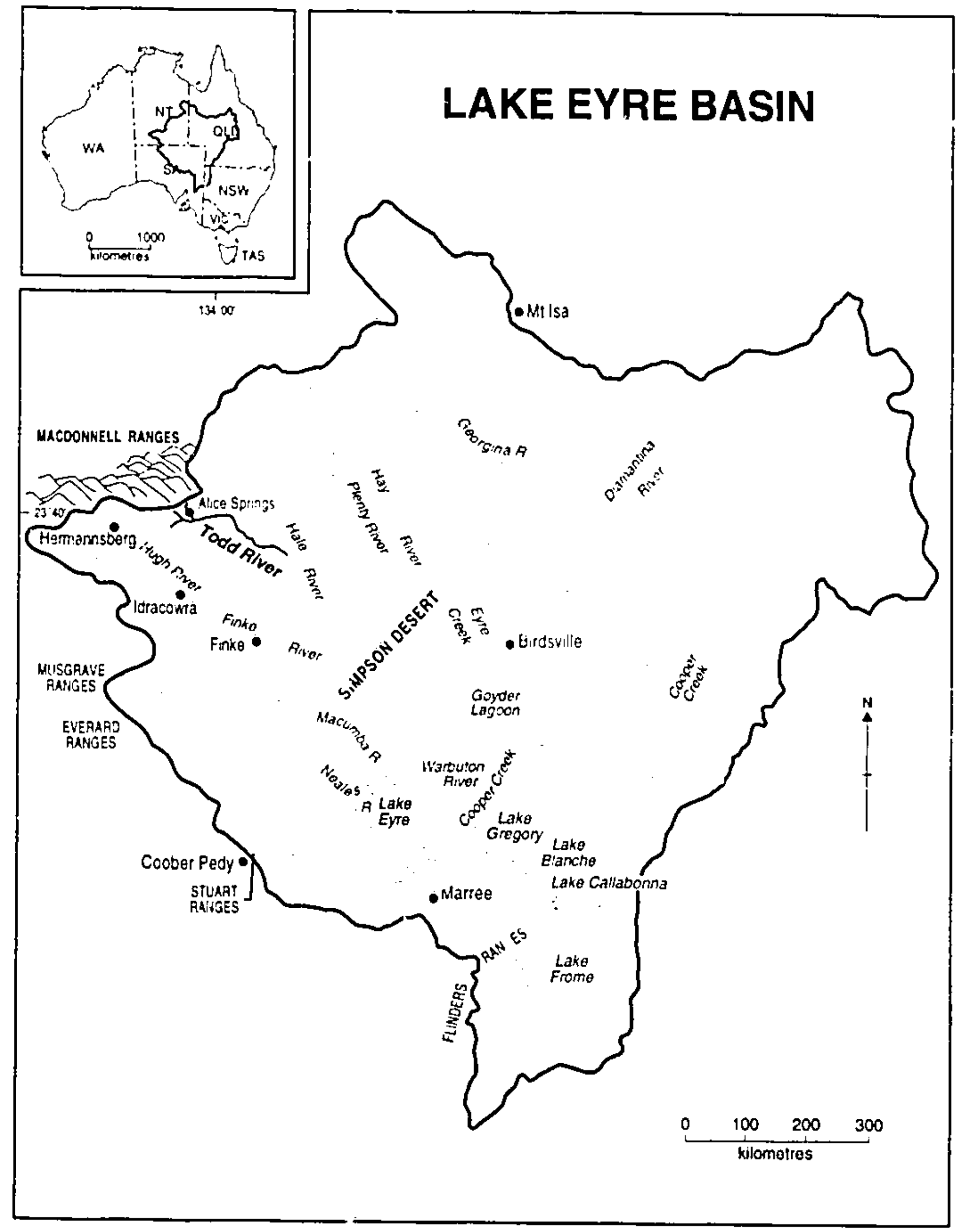


Figure 2.1. The Lake Eyre Basin.

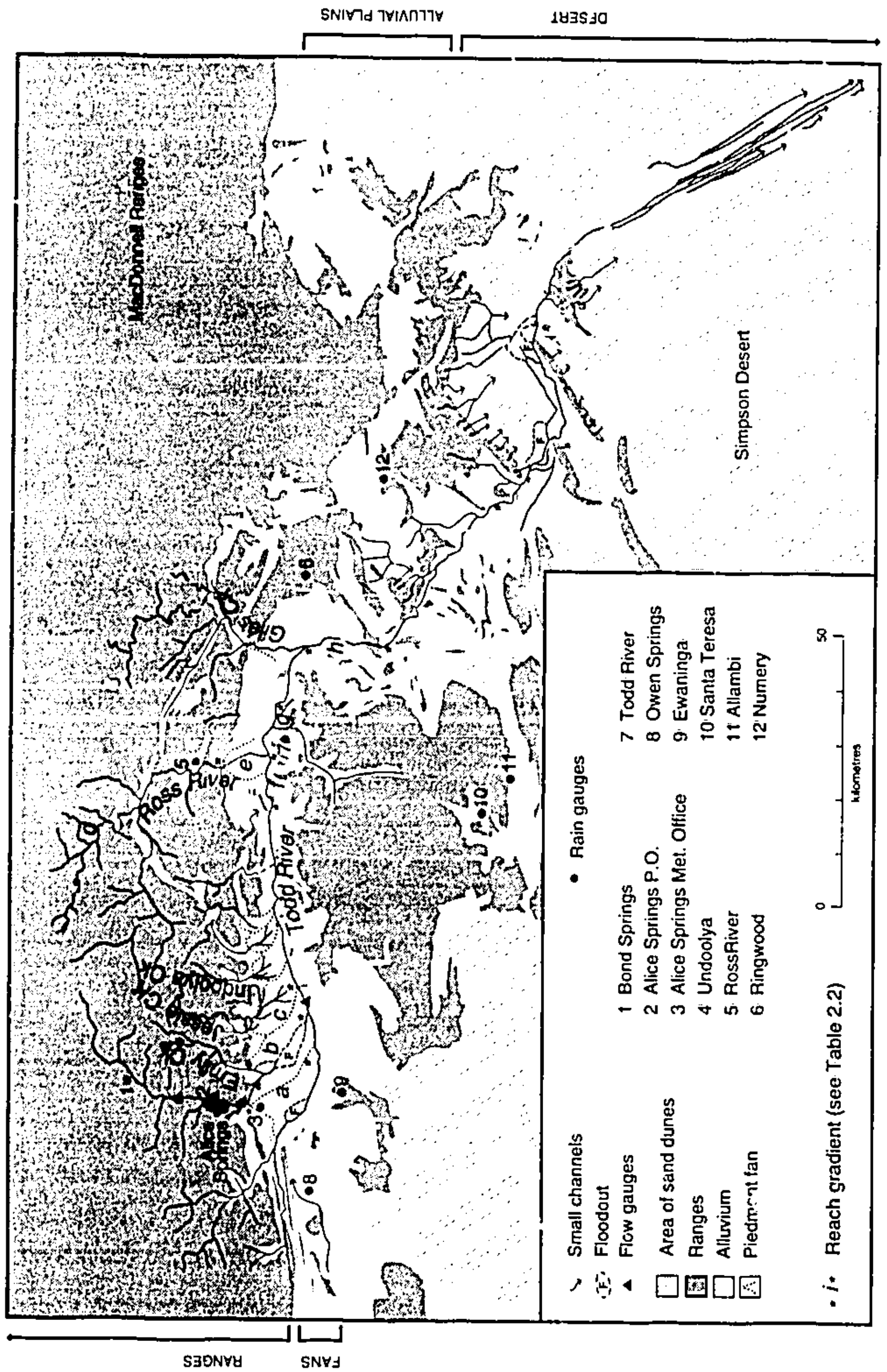


Figure 2.2. Todd River catchment.

year but bring rainfalls of between 100-200 mm in 24-48 hour periods that fall as light, steady rain interspersed with heavy thunderstorm showers (Slayter. 1962).

While all arid regions have spatially and temporally variable rainfall, the Australian year to year variability is high on a world scale for regions of comparable aridity (Gentilli, 1971; Griffin and Friedel 1985). Yearly totals are skewed towards lower values, but there are years of very high rainfall. Large incursions of cyclonic and monsoonal depression occasionally cause heavy rainfall over large areas. For example the gauge at Alice Springs recorded 80% of the median annual rainfall in 24 hours in March 1988. Some years may contain several of these high magnitude events, *e.g.* 3.4 times the median annual rainfall fell in 1973 (Stafford-Smith and Morton, 1990). Most of the summer rains are from high intensity storms which produce high runoff.

The Todd River is ephemeral with infrequent and sometimes large flows. The hydrologic regime is best described as 'flashy' as the riverbed is dry 98% of the time and flow events rise and recede rapidly. There are few discharge gauges on the Todd (Fig. 2.2) and runoff at the Wills gauge displays high inter-annual variability like the rainfall record (Fig. 2.3b). The largest discharge event since systematic records were commenced, considered to be the second largest event since European settlement of the region over 100 years ago, occurred in 1988 and had a peak discharge of 1190 m^3s^{-1} for a catchment of 450 km^2. This is regarded as a 1-in-50 yr event, on the basis of gauging since 1952. Flow hydrographs for the event (Figure 2.4) indicate the rapid rise, particularly at Wigley Gorge (2 m/hour), the very short time between peaks at the three stations and the short duration of the event (rainfall lasted about 6 hours and the flood about 12 hours) (Barlow, 1988).

There is little available data on sediment loads for central Australia. Pickup (1991) has presented the sediment transport rate for a given discharge (Fig. 2.5) and shown that it is considerably higher during drought conditions than after a wet period has restored vegetation cover. Runoff for a given rainfall is also higher, with the result that total sediment discharge can be up to ten times higher during a drought than it is during a wet period even though the total rainfall is less (Pickup, 1991).

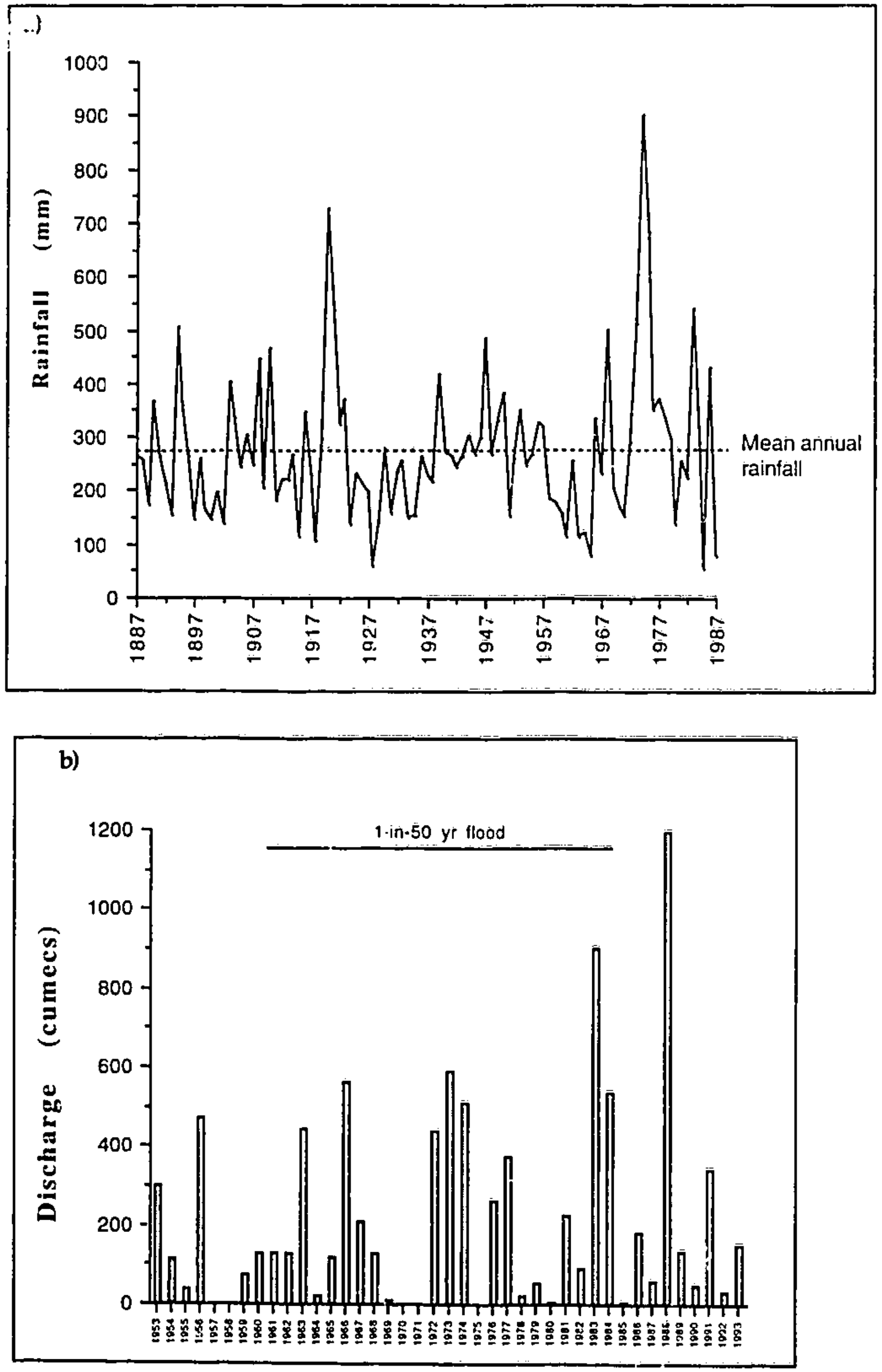


Figure 2.3. a) Total annual rainfall recorded at Alice Springs Post Office between 1887-1987. b) Maximum annual discharge recorded at Wills Terrace Gauge, Alice Springs 1953 to 1993.

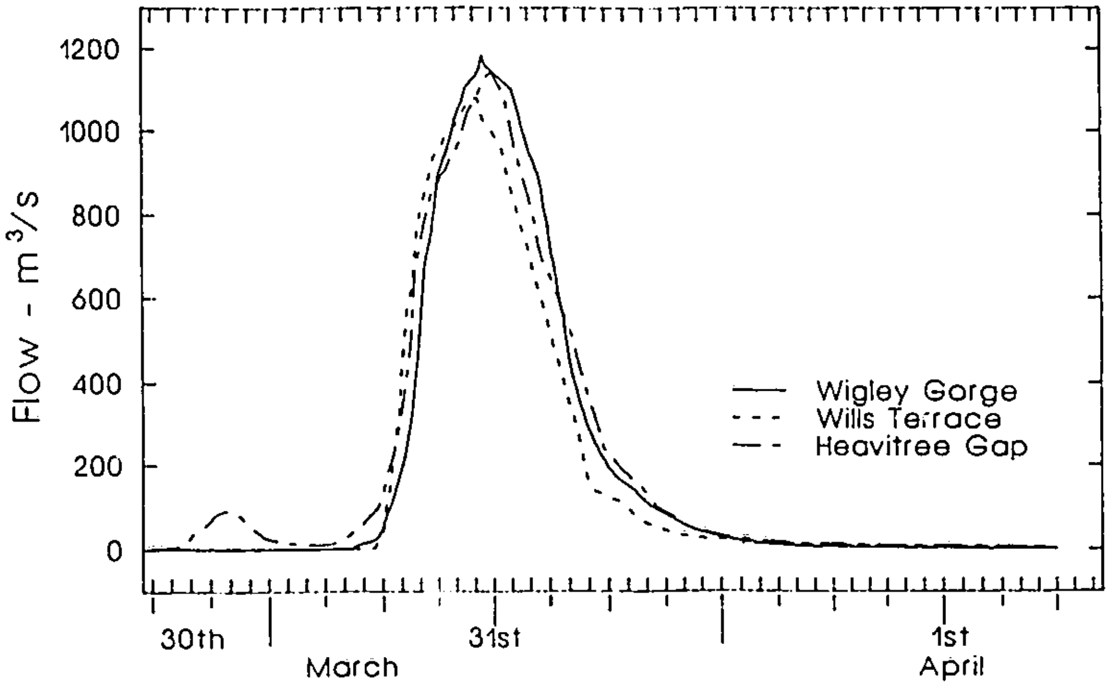


Figure 2.4. Todd River Flow Hydrographs for 1-in-50 year flood, 1988 (Barlow, 1988).

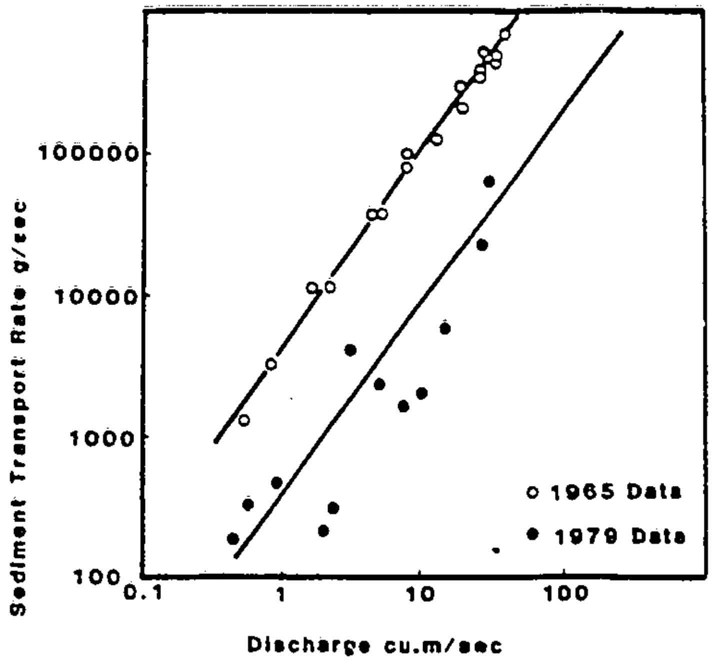


Figure 2.5. Sediment transport rates (Pickup, 1991) calculated from 1965 (dry period) and 1979 (wet period) floods for the Todd River, Alice Springs (from Department and Works, 1979). While the presentation of sediment data as a transport rate rather than concentration is open to criticism, the data illustrate just how great the effect of reduced vegetation cover is on sediment yield.

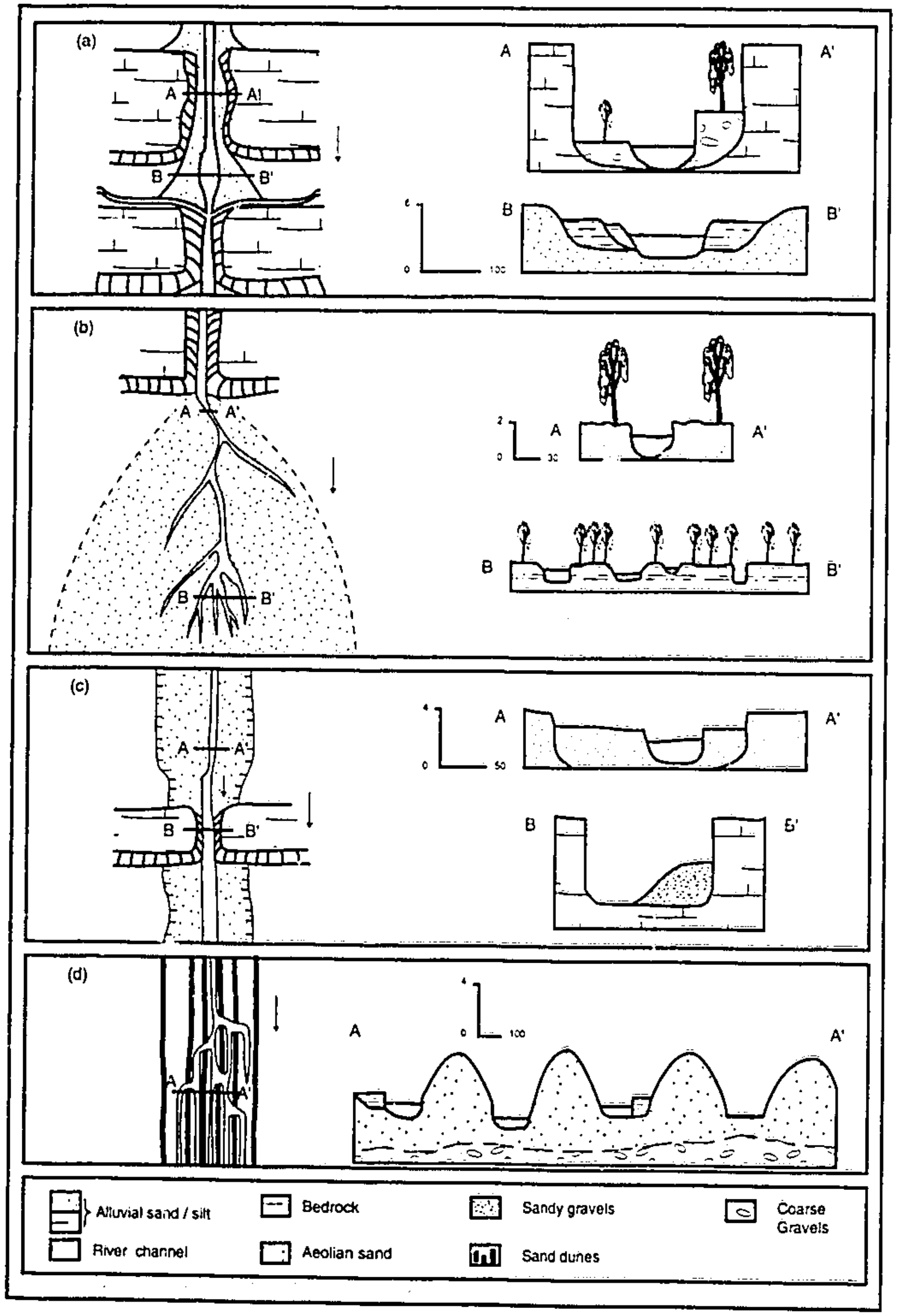


Figure 2.6. Schematic diagrams of Todd River planform and cross sections in each topographic domain (see Fig. 2.2). Arrows indicate flow direction.

a) Gorge/strike valley. A - A' Cross section through the gorge reach. A confined channel with coarse lateral deposits. B - B' Cross section through the strike valley reach. A relatively unconfined channel inset in alluvium with well defined channel and flood plain.
b) Piedmont reach. A - A' Single thread channel inset in piedmont fan alluvium.
B - B' Multiple channel network with reduced channel capacity and dense vegetation. Note distributary channel pattern in floodout.
c) Alluvial plains with outcropping ridges. A - A' Single thread channel inset into alluvium. B - B' Confined channel with asymmetrically channel cross section indicating a deep bedrock pool and lateral sandy mound.
d) Desert reach. A - A' Cross section through terminal floodout as channel flow along dune swales. Note trellised style channel pattern.

2.3. Vegetation

Vegetation patterns in central Australia display distinct zonation, as sand dunes and sand plains support different species (*e.g. Micromyrtus flaviflora, Duboisia hopwoodii, Plectrachne schinzii*) from the alluvial soils. Mahney (pers. comm.) has recognised three distinct alluvial surfaces in the Ross River gorge and floodout on the basis of vegetation patterns, frequency of inundation and flood processes which are applicable throughout the catchment. The oldest alluvial surface supports a low open woodland with *Hakea eyreana* and some *Acacia estrophiolata*. The second surface is a transverse bedform field deposited during extreme floods where vegetation is sparse due to flood scalds and damage by cattle, with isolated *Acacia murrayana* and *A.victoriae*. The third surface is the low area close to the channel that is inundated approximately once every ten years. These surfaces support an open shrubland of *Acacia victoriae, A. murrayana* and *Hakea eyreana*. The alluvial flats in the gorge support closed shrubland composed of *Acacia victoriae* and *A. murrayana* regrowth from the 1970's and 1980's floods. Older (approximately 100 yr) emergent *Acacia estrophiolata* are also present (Mahney, pers. comm.). The channel banks are lined by *Eucalyptus camaldulensis. Eucalyptus microtheca* is also associated with inundated surfaces (Kimber, 1996).

These areas contrast markedly with the vegetation on sand dunes and sand plains which support *Micromyrtus flaviflora, Duboisia hopwoodii*, and *Plectrachne schinzii*. The vegetation of the Simpson Desert forms distinct zones corresponding to crests, slopes and swales or inter-dune corridors. Cane grass (*Zygochloa paradoxa*) is one of the few plants that can colonise the coarse highly mobile crests. The spinifex *Triodia basedowii* grows on the stable slopes and in the sandy inter-dune corridors. A range of shrubs such as *Acacia, Eremophila* and *Grevillea*, and seasonally abundant herbs are also associated with these slopes and corridors. The less sandy corridors show a wide range of soil type, such as old alluvial floodouts, saltpans and gibber. This variety is reflected in the vegetation which ranges from low open woodlands or tall open shrub lands of *Eucalyptus microtheca, Acacia georginae, A. aneura, A. kempeana* and *Hakea* to low open shrubland of *Atriplex vesicaria, A. nummularia, Maireana aphylla, Holosarcia spp.* and *Muehlenbeckia cunninghamii*, to sparse short grasslands with scattered low trees and shrubs (Van Oosterzee, 1991).

2.4. Geomorphology

The channel pattern of the Todd River is essentially single thread, straight or gently winding with localised anabranching and it changes to a distributary pattern in the intermediate and terminal floodouts. Channel gradients are moderate to low (Table 2.2) and drainage basin areas of sub-catchments are in Table 2.1.

Tributary Catchment	Drainage Basin Area (km²)	Drainage basin area of Todd catchment to tributary confluence (km²)
Todd North of Alice	450	NA
Emily Creek	200	505
Jessie Creek	145	2350
Undoolya Creek	90	2,650
Ross River	1,260	3,600
Jinker Creek	215	NA
Giles	1,000	4,800

Table 2.1. Drainage basin area of tributary catchments and total area to confluence junctions.

Reach Name	Reach ID (see Fig 2.2)	Gradient (m/m)	Elevation (m)/ Distance (km)
Heavitree Fan	a	.0024	40/17
Emily Fan	b	.0048	60/12.5
Jessie Fan	c	.0036	40/11.2
Trephina to Shannon Bore	d	.0027	120/45
Shannon Bore to confluence	e	.0019	20/10.5
Triangle dune to confluence	f	.0013	20/15
Confluence to Stud Bore	g	.0014	20/14
Giles confluence to MBD	h	.0015	60/40
MBD to D	i	.0015	60/41

Table 2.2. Regional gradients measured from 1:50,000 maps

The alluvium along the Todd River is divided into three broad units displaying different sedimentary characteristics (Table 2.3). The youngest unit (A_0),which constitutes the present flood plain can be traced throughout

the catchment and is found in nearly all reaches of the channel. These flood plains are described in Chapters 4 and 5. The second alluvial unit A_1 is sand and gravel deposited by high magnitude floods and laid down in splays, distributaries transverse bar fields and channel complexes, which are described in Chapters 6 and 7. The third unit is the oldest. It is relatively indurated and considered to be Pleistocene. It both underlies the A_0 and A_1 units and forms high terraces. In places the A_2 alluvium is overlain by aeolian dune fields/dune remnants. This indurated unit retards active migration of the channel and provides an alluvial envelope within which the more recent alluvial units are set. Thus the alluvial stratigraphy in this catchment, particularly away from the ranges, may be broadly described as a thick(~100 m) aeolian and alluvial Pleistocene unit overlain (and incised) by paleoflood deposits and incised by the modern Todd River.

PROPERTY	UNIT A_0 Flood plains	UNIT A_1 Paleoflood	UNIT A_2 Pleistocene
Morphology	Stepped	Some steps	No steps
Colour, Munsell	10yr7/4	5yr5/6	2.5yr5/8
Texture	Clay to gravel	Clay to gravel	Clay to gravel
Pedogenesis	Rare	Common	Abundant
Bioturbation	Occasional	Abundant	Abundant
Oxidation	Absent	Absent	Abundant
Induration	Absent	Occasional	Abundant
Secondary Gypsum	Absent	Occasional	Abundant
Sedi. Structures	Well preserved	Poorly preserved	Not preserved
Charcoal/organics	Abundant	Occasional	Absent
Recently flooded	Yes	Yes	Yes
Chronology	Late Holocene	Holocene -Pleistocene	Pleistocene >59 Ka (Patton *et al.* 1993)

Table 2.3. Properties of Todd River alluvial units.

The channel cross sections are wide and shallow (average w/d is 90). The channel is usually well defined. Banks are composed of erodable sandy alluvium but occasionally, older cemented Pleistocene alluvium and aeolian deposits form more resistant channel boundaries. The channel carries a coarse sandy load, deposited as large scale ripples and tabular bars. These may be interspersed by finer silt and clay deposits which settle out in localised channel lows. Channel width is highly variable and averages 80 m; mean flood plain width to channel width ratio is 3.4:1 with a minimum of 0:1 and a maximum of 9.4:1.

The landscape through which the Todd River and its tributaries flow is characterised by many topographies, from the strike-ridge-dominated MacDonnell Ranges, through Pleistocene piedmont fans and wide alluvial plains until the terminal floodout in the longitudinal dunes of the northern Simpson Desert. Mabbutt (1966) inferred that the stability of the Australian continent and the absence of Tertiary marine incursions resulted in an importance of the inherited landscape in central Australia which is essentially a deeply weathered peneplain formed under a humid climate, with extensive lateritic and siliceous duricrusts. It forms low relief plains in northern, central and western Australia and elsewhere survives as plateau summits and bevelled crests or in the stony surfaces of low tablelands, where the duricrusts give rise to mesas and breakaways. Near the ranges the older surfaces of the plains commonly have a well-preserved persistent soil cover leaving a more extensive outer pattern of aeolian soils on sandplain and dune fields surrounding the ancient drainage foci (Jackson, 1962). Mabbutt (1966) has shown that there has been a general retreat of drainage to the inner alluvial tracts through the Quaternary. The broad topographic domains are reviewed below and a brief schematic description of the changing landform assemblage in each domain is shown in Figure 2.6.

2.4.1. The headwaters

In the headwaters, the Todd River and its tributaries flow through narrow and sometimes meandering gorges. Further south, folding of the Pre-Cambrian quartzite, sandstone and accessory carbonate rocks has formed low, sharp crested, east-west trending ranges separated by narrow plains. Trellised rivers drain these parallel strike ridges and tributary gorges, passing through short gorges superimposed on the strike ridges and alluvial or sand plain reaches in the strike valleys. Fluvial features in the steep gradient gorge reaches (Fig. 2.6a) are similar to the narrow and wide gorge morphologies described by Pickup *et al.*, (1988) and include an assemblage of high energy gravel bar deposits. Where they cross intervening strike

valleys, the flood plains widen locally (Fig. 2.6a). Slackwater deposits from high magnitude flows are sometimes located in the narrower strike valleys, at tributary junctions and in areas where gorge width increases.

2.4.2. Piedmont reach

Extending from the foot of the MacDonnell Ranges is a series of low angle fans (approximate gradients average .0036 m/m). On leaving the ranges, the Todd River and its tributaries cross these unconfined piedmont fan systems. Fluvial landforms in this reach reflect the comparatively lower energy conditions and, with the exception of the Ross River and Giles Creek, tributary channels terminate in a pattern of densely vegetated, low gradient distributary channel systems as transmission losses into the stream bed lead to a reduction in channel capacity downstream (Fig. 2.6b).

This reach is unconfined in that it does not follow a bedrock gorge. However, the flood plain is inset into alluvium and low angled Pleistocene colluvial aprons, which limit the width and lateral extent of flood plain processes. Pickup (1991) and Patton *et al.* (1993) have described the geomorphology of the piedmont reaches of the Todd River. The geomorphic effect of extreme floods on the low angled piedmont fans is seen as essentially aggradational although normally the fans are incised by small, ephemeral flows. Paleoflood forms include sand sheets, sand threads, ripple fields, overflow channels (Pickup, 1991), large scale paleo-braid channels, levee deposits and broad low relief bars (Patton *et al.*, 1993).

2.4.3. Alluvial plains

Downstream from the fans, the Todd River traverses a broad, strike-trending intermontane lowland, occupied by a wide alluvial plain bordering a broad sand plain interspersed with minor rounded hills and discontinuous strike valleys. These dissected plains and tablelands also contain small aeolian dune fields which are outliers of the main Simpson Desert (Fig. 2.2). In places, the Todd River has incised into carbonate-rich red earth which is developed on the Pleistocene alluvial surface. During higher magnitude flows, this relatively resistant sediment confines the high stage channel and facilitates the stripping and infilling of the flood plain. The resultant morpho-stratigraphy is complex, composed of a series of erosional surfaces and inset flood plain deposits (Chapters 4 and 5).

The variable composition of channel boundaries also influences the channel bed morphology especially where the river is laterally confined by narrow gaps in outlying bedrock ridges (Fig. 2.6c). Here the channel morphology abruptly changes from a relatively undulating sandy channel bed upstream of the confinement to a channel marked by a deep scour pool which occupies up to one-half of the channel width.

2.4.4. Desert reach

In the lower reaches of the Todd River, the channel occupies the 300-500 m wide, inter-dune corridors of the longitudinal north-west trending dunes of the northern Simpson Desert. The dune fields are fossil features fixed by vegetation (Mabbutt, 1966) and can be several kilometres in length; swale width is approximately 300 m. The crests of these dunes are actively mobile and overlie an indurated core at depth which limits the lateral migration of the channel. The river maintains a well defined channel for 10 km then the channel crosses a longitudinal dune and bifurcates. In this reach, incipient flood plains form spatially discontinuous benches composed of fine sand overlying aeolian sediment. Approximately 20 km downstream, the channel splits again, occupying several parallel swales (Fig. 2.2). Channels support dense stands of Coolabah (*Eucalyptus microtheca*) and channel morphology is subdued. Where the channel crosses dune orientation, fine textured back-flood deposits are emplaced. This pattern continues for a further 40 km until flow dissipates in a terminal flood basin. The spatial pattern of the longitudinal dunes controls the prevailing inverted trellised channel pattern in this reach (Fig. 2.6d).

2.4.5. Paleofloods

In addition to these broad topographic domains, river morphology is influenced by a flood plain topography made up of a series of Holocene flood deposits (Pickup, 1991; Patton *et al.*, 1993; Bourke and Pickup, in press) which extend from the gorge reach to the Simpson Desert. The modern channel is, in places, inset within the flood deposits, which provide an abundant source of coarse sediments, variable channel boundary resistance and topographic highs to confine flow. Paleoflood forms extend several kilometres from the active channel and contain a variety of morphologies including paleochannels, high-level bars and transverse bedform fields. Where the modern channel and the paleoflood channel(s) intersect, channel bed sediment may increase in size from gravelly sand to sandy gravel and

cobbles. This may be accompanied by a localised increase in channel width by up to 450 m. This 'beaded' channel planform is a marked feature of the Todd River (Chapter 8).

2.5. Land Use

The town of Alice Springs (population ~28,000) is located in the headwaters of the Todd River. Since European settlement of the region over one hundred years ago, land use within the catchment has become predominantly pastoral, in particular cattle grazing. Prior to this, mobile hunting and gathering were practised by Aboriginals (Spencer, 1896). The tourism industry is the main employer in Alice Springs with upward of 300,000 visitors per year.

2.6. Late Quaternary Paleoclimate In Central Australia

Although the past decade has seen an increase in the number of studies with a focus on paleoclimatic conditions in Australia, the arid centre has for the main part been neglected. Coupled with the poor preservation of pollen assemblages, little is known of the late Quaternary paleoclimate of central Australia. The identification of regional, broad climatic phases is inferred from geomorphic (*e.g.* Bowler, 1976, 1981; Wasson, 1984, 1986) and palynological data (*e.g.* Singh, 1981; Singh and Luly, 1991) and is synthesised elsewhere (*e.g.* Chappell, 1991; Ross *et al.*, 1992; Harrison and Dodson, 1993, Kershaw and Nanson, 1993; Kershaw, 1995). A brief summary is given here. Prior to 30,000 BP conditions were cool and wet with high lake levels in many regions. This was followed by a cooling and increasingly dry climate towards the last glacial maximum, the effects of which extended to about 12,000 BP. The post glacial phase saw a rapid warming followed by an increase in precipitation. By 6,000 BP temperature and precipitation were generally higher than present. The late Holocene is cooler and drier with some sites indicating a slight warming during the last 1,000 yr. Recent research indicates that these broad paleoclimatic trends were punctuated by extreme rains and floods (Pickup *et al.*, 1988; Pickup, 1991; Patton *et al.*, 1993; Bourke and Pickup, in press).

Williams (1974) was the first to establish an absolute chronology of fluvial and pedological events in the Todd River valley (See Table 2.5) which indicates the onset of the last arid phase around 26,000 BP and its persistence until the mid Holocene.

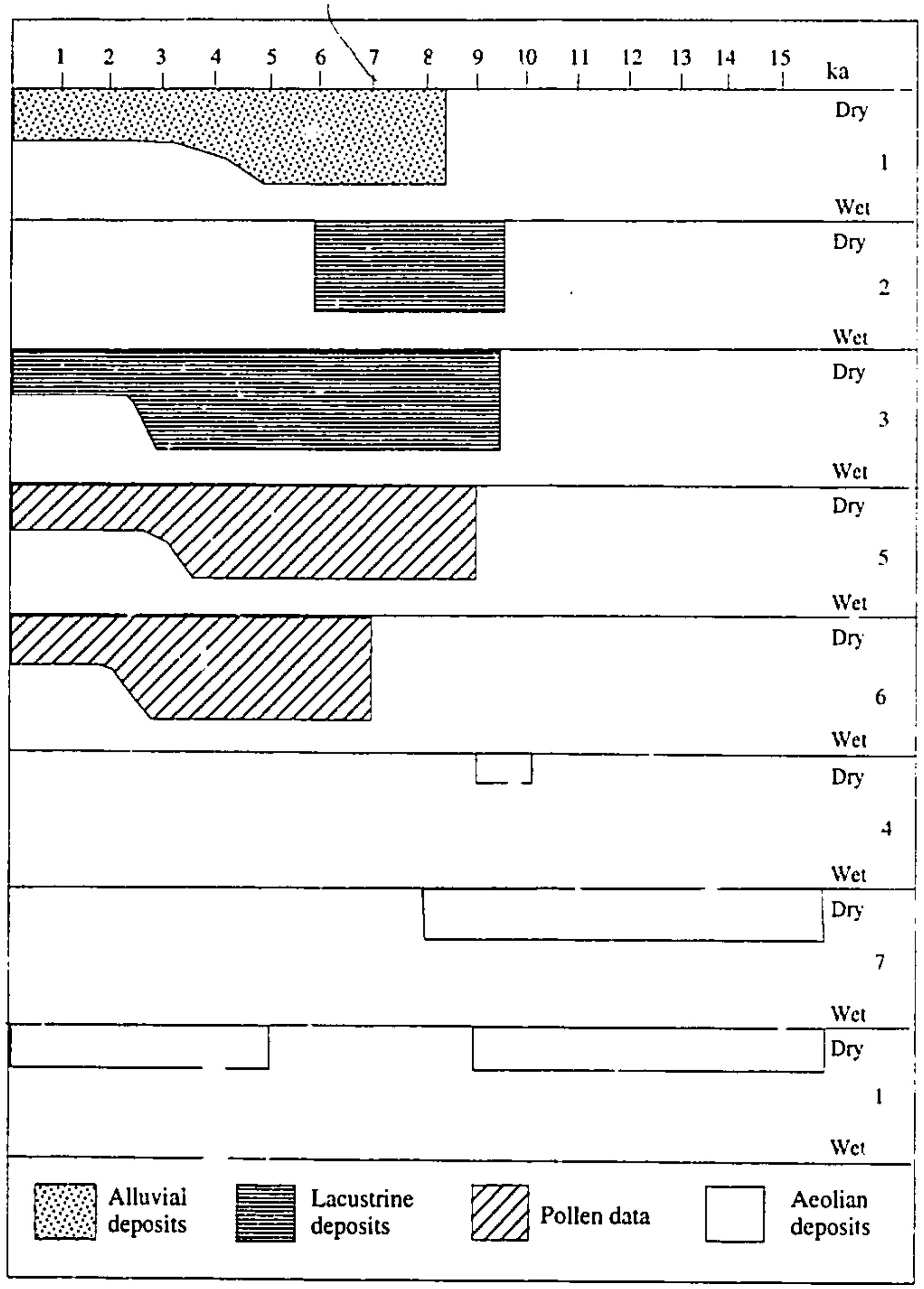


Figure 2.7. Late Pleistocene and Holocene paleohydrology in central Australia. 1. Finke River (Nanson *et al.*, 1995), 2. Lake Amadeus (Chen *et al.*, 1991), 3. Lake Eyre (Gillespie *et al.*, 1991), 4. Ross River (Patton *et al.*, 1993), 5. Lake Frome (Singh, 1991), 6. Lake Tyrrell (Teller *et al.*, 1982), 7. Strzelecki Desert (Wasson, 1984).

The paleohydrological reconstructions for the central Australian region (Fig. 2.7) employs geomorphic evidence including channel dimensions in the Finke River (Nanson *et al.*, 1995), floods on the Ross River (Patton *et al.*, 1993), phases of dune building in the Strezelecki and western Simpson Deserts (Wasson, 1984; Nanson *et al.*, 1995) and lacustrine and beach sedimentation in Lake Amadeus and Lake Eyre (Chen and Barton, 1991; Gillespie *et al.*, 1991). The palynological evidence is taken from Lake Frome (Singh, 1981) and Lake Tyrell (Teller, 1982).

Phase	Geomorphic and Pedogenic Activity
>27,000 BP	Valley flood alluviation and deposition of marginal screes
~27,000 - 24,000	Widespread pedogenesis. Nodular calcareous soils and massive calcretes developed
~24,000 - 12,000	Dune formation near north margin of Simpson desert
~12,000	Pedogenesis. Nodular calcareous soils
>/~ 5500 - ~1,500	Aggradation of marginal fans and central valley floor
~1,500 - ~1,200?	Gullying and fan dissection
1,200 - </~500	Partial aggradation of gullies
Today	Gullying in fans and valley floor locally

Table 2.5. Phases of geomorphic activity in the Todd catchment, summarised from Williams (1974).

In the plains north of Alice Springs meandering paleochannels within the terrace alluvium suggest that the extensive alluvial deposits were laid down by large sinuous laterally migrating channels from some time prior to ~56 ka until 17 ka (Nanson and Tooth, in press) and there is some evidence of fluvial activity around the Last Glacial Maximum (LGM) (TL dates of ~25 ka and 17 ka, also found by Nott *et al.*, (1996) in north Australia) (Table 2.6). After ~15ka decreasing flows and a reduction in channel size on the rivers draining north of the range left the older Pleistocene alluvium as paired terraces (Nanson and Tooth, in press). Evidence suggests that the more arid conditions of the late Pleistocene extended into the early Holocene in central Australia, although the termination of this aridity is spatially and temporally varied. Wasson (1984) found that dune building in the Strzelecki continued until approximately 8 ka. This correlates with the evidence from the Finke region which dates pale boundary dunes to between 17 and 9 ka (Nanson *et al.*, 1995). A period of relatively high discharges in the Finke River, probably associated with enhanced rainfall from ~9 ka to ~5 ka reworked most of the previous valley fill and replaced it with a flood plain that has survived largely unaltered to the present (Nanson and Tooth, in press). Significantly there was a hiatus in the activity of the adjacent dunes during this period. Reduced fluvial activity probably due to reduced rainfall led to channel stability and reactivation of the adjacent dune fields from the mid to late Holocene and the tops of the dunes remain active today (Nanson and Tooth, in press). The general trend from early to mid Holocene where conditions become generally wetter than present is supported by the higher lake levels in Lake Eyre to about 3.2 ka (Magee *et al.*, 1995) and a higher regional groundwater table in Lake Amadeus (Chen and Barton, 1991). Palynological evidence is in general agreement with this pattern with taxa from Lake Frome (Singh, 1981) and Lake Tyrrell (Teller *et al.*, 1982) indicating wetter phases in the mid Holocene and drier conditions from approximately 4 ka.

It is difficult to establish whether the fluvial activity in the central ranges during the early to mid Holocene reflected increased summer rainfall resulting from more frequent monsoonal and cyclonic incursions, or increased winter rainfall resulting in the frequency and intensity of mid latitude frontal systems. However evidence for enhanced wetness and runoff in northern Australia during the same period (Lees, 1992; Nanson *et al.*, 1993, Nott and Price, 1994, Nott *et al.*, 1996; Wende *et al.*, in press) suggest that it is most likely that the rivers draining the central Australian ranges have responded to changes to the northern Australian monsoon (Nanson and Tooth, in press) (Table 2.6).

System	Drainage	Climate	Surface Characteristics	Sediments	Post Depositional Alteration	Soil	Age
Contemporary	pastoral occupation, accelerated cycle of water erosion	Present day, semi arid climate; sporadic high intensity rains, infrequent high velocity winds	Channeling and riling;		Minor deflation	No soil development	Contemporary
Amoonguna	Incompetent; discontinuous channelling, few distributaries.	Present day, semi arid climate, sporadic high-intensity rains; infrequent high velocity winds.	Stranded; discontinuous deposits	Sands, silts, minor fine gravels	Deflation but no massive drifting or piling. Negligible pedogenesis Raw sediments	Fluvial regosol and incipient grumusols.	Recent past, Postglacial
Upper Stuart	Vigorous drainage with continuous aggrading channels	Semi arid to sub humid; More frequent high intensity rains than today; infrequent high velocity winds	Only found beyond debouchment from steep watersheds. Gravelly loads and sheets	Coarse deposits; sands and gravels, minor silts	Raw to very slightly weathered	Incipient sandy red earths and solonetz	Late glacial/ early postglacial (14 - 10 Ka?)
Lower Stuart	Vigorous drainage across inner plains with continuous aggrading channels and numerous distributary systems with sand choking and sand piling in final development	Subhumid. Seasonal high intensity rains and high velocity winds.	Spread more widely than upper surface across flood plains and outwash plain. Surface drift sands and sand dunes	Coarse gravelly, sands clays, silt load	Raw to weakly weathered sediments, local sand drifting and piling	Different according to drainage variation. Immature red clayey sands, sandy red earths, solonetz and grumsols	Last glacial (20 - 15 Ka)
Burt	As above	As above	Surface drift sands and sand dunes	Sands and clays with gravels	Raw to moderately weathered sediments. similar wind sculptured	Red clayey sands, sandy red earths, and buried solonetz developed on wind sorted sediment	50 - 60 Ka
Hermannsburg	Very vigorous across inner plain;	Humid at least seasonally with high intensity rains. Period of widespread shallow ground waters	Extensive piedmont deposits/gravel sheets. local lacustrine conditions. Extensive shallow ground water. Dissected formation on the interfluves between the flood plains	Coarse gravels and clays	Weak to moderately weathered sediments Calcreted	Read earths and grumusols, also lithosols and calcareous earths	100 - 130 ka
Yambah	Not known, possible redistribution from local interfluves in addition to major watersheds. A separate earlier phase of calcrete formation.	Possibly subhumid tropical	survives on very subdued slopes on interfluves.	paucity of gravels other than fine ironstone and quartz, clays, calcrete	Moderately weathered, considerable development of calcrete with attendant deposits of palygorskite	red earth and minor grumusols	Mid Pleistocene
McGrath		period of planation, deep weathering, and laterization			Deep weathering profile, lateritic features, abundant calcrete.	gravelly red earth	Before the end of the Tertiary

Table 2.4. Quaternary climate cycles in central Australia and their deposits derived from igneous and metamorphic Rocks. After Litchfield, (1969).

Oxygen Isotope Stage/phase/age	Site	Monsoon	Geomorphic evidence	Dating Methodology	Reference
Stage 1 *12-0 ka*	Finke River	Monsoon	*Fluvial activity*: early to mid Holocene reworked flood plain near Finke township.		Nanson et al, 1995
	Top end		*Fluvial activity*. Plunge pool activity early to mid Holocene	8 TL dates from 10.6±1.5 ka to 5.0 ±1 ka.	Nott and Price 1994; Nott et al, 1996.
Stage 2 *24-12 ka*	Lake Eyre and Lake Blanche	Monsoon	*Lacustrine activity*: beach Ridges ~22 m above present lake floor indicating greater runoff from monsoon tropics	3 TL dates 22.8±2.3 ka to 26.2±3.7 ka.	Note 1
	Lake Eyre basin		*Arid*. Dune building phase	43 TL ages average at 24 ka	Note 1
	Sandover River		*Fluvial activity*	25 ka to 17 ka	Tooth, 1997
	Top End		*Fluvial Activity*. Plunge pools activated. Higher discharges than late Holocene. Decline in fluvial activity following the LGM but increase after 16 ka.	7 TL dates from plunge pools from 30.6±2.1 ka to 17.9±2.5 ka.	Nott et al. 1996
Stage 3 *60-24 ka*	Channel country		*Fluvial activity*. Early to middle (60-32) Most active in steeper headwaters streams and less so near the arid centre of the basin. Lower magnitude than Stages 7 and 5.	2 TL dates 35.8±2.7 ka and 41.5±3.4 ka	Nanson et al, 1992, Nanson et al, 1988, 1992,
	Lake Eyre South and Lake Frome		*Lacustrine activity*. Beach ridges indicate that lake filled to ~20 m above present level. A significant hydrological event but not as strong as on the rivers of south eastern Australia.	[illegible] TL dates from 54.9±9.2 ka to [illegible]8.8±11.2 ka.	Magee et al. 1995
	The Top End	Enhanced Monsoon	*Fluvial activity*. Plunge pool deposits. Not a major episode in northern tropics	2 TL dates: 43.2±1.7 ka and 58.2±5.6 ka	Nott et al. 1996
	Gulf of Carpentaria	Enhanced Monsoon	*Fluvial activity*. Shallow channel deposition. Not a major episode in northern tropics	[illegible] TL dates from 48.7±3.2 ka to [illegible].1±7.8 ka	Nanson et al 199
Stage 4 *75-60 ka*	Channel Country and Lake Frome		*Arid/ Fluvial Inactivity*. Aeolian sands	[illegible] TL dates from 78.2±8.7 ka to 6[illegible].6±6.1.	Nanson et al. 1992
Stage 5 *(Last Interglacial)* *130-75 ka*	Channel Country & Lake Eyre	Humid	*Fluvial activity*. (Katipiri) Reworked coarse sands and over bank mud with peak at 110 ka or slightly younger. Sub stage 5e and 5b-a appear less active than 5b an 5c. Pedogenic mineralisation. Hoof prints and megafaunal fossils.	17 TL dates from 87.6±8.7 ka to 1[illegible]2±24 ka (average of 107 ka)	Nanson et al. 1992 Magee 1997
	Arnhem land		*Fluvial activity*. Relatively early onset of fluvial activity at Sub stage 5e	8 TL dates from 80.8±15.5 ka to 1[illegible]±25 (5 cluster between 120-130)	Nanson, pers. comm.
	Gulf of Carpentaria		*Fluvial activity*.	TL and U/Th 85 ka, > 80 ka	Nanson et al., 1992
Stage 6 (Glacial) *190-130 ka*	Channel Country		*Generally Arid but variable with separate arid and humid phases*. eg with interstadial fluvial activity (~160-170 ka). (Katipiri)	3 TL dates from 158±14 to 169±14.	Maroulis, in prep
	Arnhem land			3 TL dates from 157±30 to 177±30.	R. Roberts, pers. comm
Stage 7(Penultimate Interglacial) *244-190 ka*	Channel Country	Monsoon	*Fluvial activity*. (Katipiri) River terraces in headwaters. Extensive buried alluvium along the middle and lower reaches. pedogenic mineralisation. Megafauna fossils sometimes present.	6 TL dates from 204±15 ka to 263±35 ka.	Nanson et al 1988, 1992, Nanson and Tooth, (in press)
	Arnhem Land	Monsoon	Fluvial activity. Sand aprons and flood plain deposition	3 TL ages from 210±40 to 232±43.	R. Roberts, pers. comm.

Table 2.6. Summary of geomorphic activity and paleoclimates in central and northern Australia

Note 1: Dune activity in the Lake Eyre basin is documented by Wasson (1986), Nanson et al. (1995) and Magee (1997). High lake levels in Stage 2 are claimed by Nanson et al. (in press) but are not supported by amino acid racemisation age estimates from the critical deposits, reported by Magee (1997).

CHAPTER 3

A Review of Arid Zone Rivers

3.1. Introduction

Reviews of desert environments (*e.g.* Mabbutt, 1977; Graf, 1988a; Reid and Frostick, 1997; Cook *et al.*, 1993; Thornes, 1994a, 1994b) have adequately described aspects of arid watershed processes including in particular spatial and temporal variation in rainfall, rapid runoff from contributing slopes, transmission losses, the flashy nature of flood hydrographs, asynchronous tributary activity, the generally high suspended sediment loads and the tendency for suspended sediment concentrations to increase downstream. An aspect not equally well documented is the morphological response of arid river systems to these dynamic processes. This chapter synthesises the literature on arid zone rivers, in order to provide a framework in which to discuss the data from the Todd catchment. Specifically, this chapter outlines existing knowledge of channel pattern, morphology and bedforms, and flood plain morphology of arid rivers, followed by a summary of central Australian fluvial systems.

3.2. Channel Patterns

3.2.1. Braiding and anastomosing

Although meandering is reported in desert rivers, a straight to braided channel pattern is most common (*e.g.* the Gila River in the US, (Graf, 1981)) where low sinuosity streams with one or more channels intersect and bifurcate around braid bar shoals, which typically are pointed upstream and downstream with coarser traction clogs prograding upstream. Braiding of desert streams often reflects flows that have high energy but flow infrequently, where streams are dominated by sand or gravel bedload and have steep valley gradients, occasional large discharges, and non-cohesive banks without stabilising vegetation (Richards, 1982). Braiding may occur where there is a sudden increase in coarse sediment load, such as downstream of major tributaries (Pickup, 1991) although this is not unique to arid rivers.It is also found at sections of localised channel widening where mid-channel shoaling and subsequent stabilising of bar surfaces by vegetation occurs (Mabbutt, 1986). Although sometimes confused with braiding rivers anastomosing rivers consist of multiple channels separated by islands which are usually excised from the flood plain and which are large relative to the size of the channels (Knighton and Nanson, 1993). Anastomosing reaches can grade into braided ones (*e.g.* in Canadian gravel bed rivers (Smith and Smith, 1980) but are distinguished from them by finer texture sediments, greater bank stability and channel relocation by channel avulsion rather than by migration (Knighton and Nanson, 1993) or within-channel avulsion.

3.2.2. Compound channels

Graf (1988b) noted that many arid rivers have both meandering and braided characteristics and may fall in the middle of the configuration scale which ranges from straight to meandering to braided. He called these compound channels, and noted that they are not restricted to arid rivers. The Salt River in Arizona, for example, has an outer braided configuration occupied during infrequent high discharges and an inner well-defined low or main flow meandering channel (Graf, 1988b). The Cooper Creek in Queensland has a braided channel system coexistent with the anastomosing channel system (Rust and Nanson, 1986). The braided channels are active during major floods (Robinove, 1979), whereas the anastomosing channels flow at moderate discharges. The two channel patterns are inferred to coexist because of the strong textural contrast between the mud of the more elevated braided surfaces which is converted by pedogenic processes into a porous aggregate of sand sized pellets, and that of the incised anastomosing channels which is highly compacted and cohesive (Rust and Nanson, 1986).

3.2.3. Distributary patterns

Ephemeral streams of the arid zone typically are subjected to transmission losses and evaporation downstream. Desert channels close to the terminal floodout change from bedload channels to suspended load channels and usually change to anastomosing and distributary channel systems (Mabbutt, 1977) with a resultant decrease in distributary channel width and depth (Schumm, 1961). As channel gradient decreases there is an increasing trend to overbank flow until channel form terminates in broad splays of sand or finer sediment (Pickup, 1991). The flow spreads out over a wide, flat and generally featureless fluvial plain with some minor discontinuous gully development. Tooth (in press) has noted the nature of channel breakdown; in particular he records marked fluctuations in channel width and depth,

the development of splay channels, large-scale channel avulsions and infills, and abrupt decay of channels. These floodouts are described below.

3.3. Channel Geometry

3.3.1. Channel width

In the sandy, bedload-dominated section of desert streams, the channel cross section tends to be wide and shallow (~100-200 m wide and ~1-4 m deep) with well defined banks (Leopold *et al.*, 1966; Baker, 1977; Frostick and Reid, 1979, Graf, 1983). Wolman and Gerson's (1978) data on drainage basin size and channel width (depth is not available) (Fig. 3.1) reveal two interesting aspects of arid channel width. Firstly, in contrast to humid systems, channel width increases rapidly with an increase in drainage basin area. In drainage basins ~10-100 km^2, channel widths tend to measure between 100 and 200 m, a width normally only achieved in perennial rivers with catchments >10,000 km^2 (Reid and Frostick, 1997). Second, the asymptote value of 100 to 200 m width is reached as drainage basin area approach 50 km^2. This has been attributed to the compensation of tributary flows by transmission losses and the assumption that the finite size of rainfall cells impose an upper limit on the discharge of an ephemeral drainage system regardless of its size (Reid and Frostick, 1997). Variations of local channel width controlled by percentage silt/clay in the wetted perimeter (Schumm, 1961) and by the nature of local bank sediments (Murphy *et al.*, 1972) may be superimposed on the regional trends.

3.3.2. Channel bed morphology

Ephemeral channels are noted for their subdued bed morphology with a predominance of low relief or planar beds, inferred to be due to their shallow flows and resultant suppression of secondary current cells (Reid and Frostick, 1997). For example, Sneh (1983) found that channel bars usually only about 20 cm high and 2 m long in the Wadi El Arish system of the Sinai Peninsula. Many central Australian sand bed ephemeral streams have a more amplified morphology particularly where the most recent bed-forming flow was of a sufficient magnitude to form dunes and mega-dunes (Williams, 1970) and the presence of vegetation in the channels enhances secondary cell circulation. This apparent discrepancy is more likely a function of incomplete information rather than significant regional difference.

Channel bed morphology may be separated into forms emplaced during high stage floods, often situated in the higher portions of the channel, and those emplaced during low stage flows (Mabbutt, 1977). High stage flow forms observed in central Australian streams include large-scale ripples or dunes ~0.4 m high and between 1 and 5 m wavelength and longitudinal bars 2-3 m high, tens of metres wide and extending hundreds of metres downstream. Higher flow velocities may erode these forms and emplace a plane bed. Williams (1971) described the bedforms in the Todd River after the flood of 1967, when peak flow at Heavitree Gap was in the order of 200 m^3/sec; at the recession large ripples (1-5 m long and 7-35 cm high) had been deposited on the more elevated parts of the channel bed associated with high stage flow. Other less common forms include: small ripples, upper-regime plane beds, flute marks, longitudinal, transverse and linguoid bars. The waning and low stage flow forms include large transverse bars with flat, often gravel-veneered crests tens of metres in extent and terminating down-channel in avalanche faces up to 1 m high. Small ripples 5-15 cm wavelength are generally associated with finer well-sorted sand. Silt and clay deposits from slackwater at the end of recession occur as veneers deposited in pools and cut-off distributaries and have low preservation potential (Williams, 1971). Sneh (1983) found lower stage ripples formed in the troughs between bars with very thin, cracked, mud layers and noted the tendency of low stage stream to breach previously built bars, resulting in a geometric discontinuity of channel bedforms.

3.4. Flood Plain Morphology

In humid areas the flood plain generally occupies a clearly defined zone with sharp boundaries but this is not always the case in the arid zone (Cook et al., 1993), where the area over which a river has flooded is a less clearly defined zone owing to aeolian reworking of fluvial sediment. There are few descriptions of arid flood plain morphology. Sneh (1983) divided the flood plains of the Wadi El Arish in Sinai Peninsula into confined, open and terminal flood plains. The narrow confined flood plains (~200 m wide) have ~3m high steep channel banks and contrast with the wide (~200 m - 2,000 m wide) open flood plains which have low flow depths (~20 cm) and gently sloping banks. The terminal flood plains at the river terminus may widen further and are flat featureless plains.

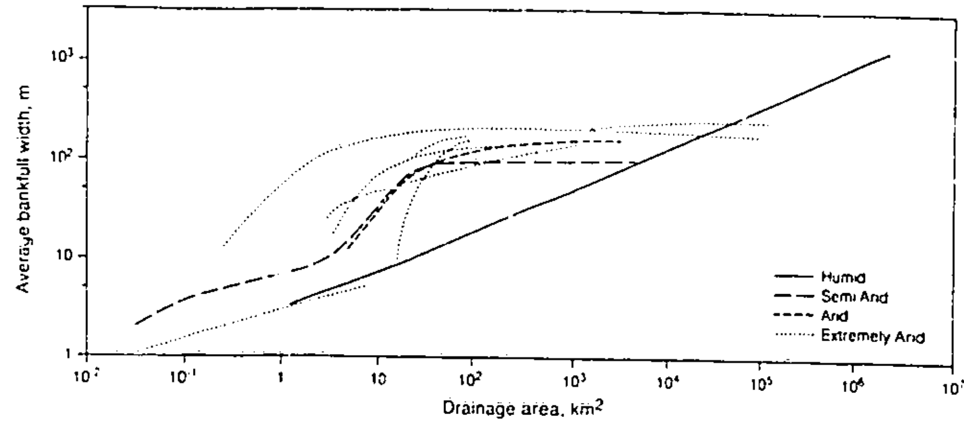


Figure 3.1. Relationship of bankfull channel width to drainage area for different climatic environments (after Wolman and Gerson, 1978).

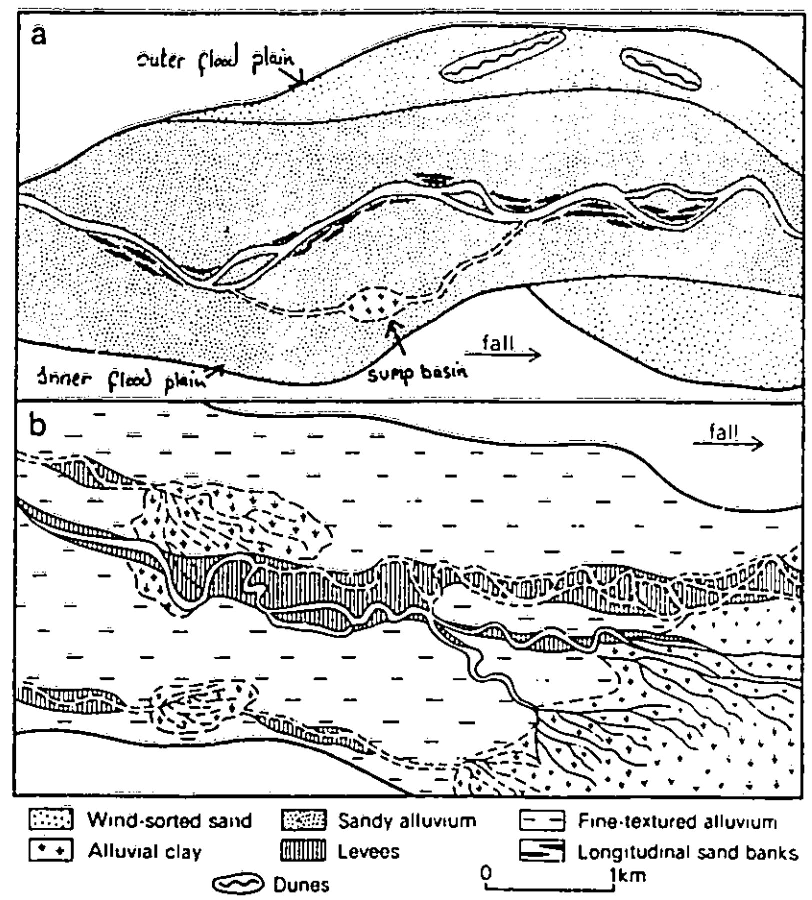


Figure 3.2. Characteristic features of central Australian flood plains. a) Sand bed channel, b) suspended load channel (Mabbutt, 1977).

Based on observations of Australian arid fluvial systems, Mabbutt (1977,1986) subdivided arid flood plains into two morphological zones: the inner and the outer flood plain (Fig. 3.2). The inner flood plain is subject to frequent changes and remoulding by floods, reflected in a lack of soil development in the layered sandy alluvium. It consists of longitudinal sand banks and intervening flood furrows formed during overbank flows. The flood plain may have distributary flood channels connected to small sump basins that are fed during floods (Fig. 3.2). According to Mabbutt (1977, 1986) the outer flood plain commonly stands at a slightly higher level, often bounded by a curved step marking a former course sometimes with unpaired low terraces, is slightly higher, composed of older alluvium, which may be reworked by wind and includes small aeolian dunes, partially stabilised by vegetation and sand moulding around shrubs (Mabbutt, 1977). The juxtaposition of flood plain surfaces of different age will be explored in the following chapters where it is proposed that the older surfaces are the result of several processes including deposition from large floods and the stripping of yet older alluvial surfaces.

By way of contrast, Pickup (1991) found the central Australian flood plains systems occur in a variety of types depending on catchment size, amount and type of sediment load, distance downstream and degree of lateral confinement, and suggested that they contain a mixture of fan and flood plain features and in many cases behave more like low-angled fans than flood plains. He identified three scales of activity: large-scale paleoflood forms, flood plains of the modern rivers and erosion cell mosaics. The flood plains of the current rivers consist of channel-levee complexes and terminal floodouts described below.

3.4.1. Floodouts

Floodouts are alluvial basins formed at the end point of the main or distributary fluvial system. With decreasing slope, channelled flows may pass into unconfined sheet floods (Davis, 1938, Tunbridge, 1984). Although the floodout zone is composed of a tract of distributary channels and floodout plains it is considered as part of the continuum of flood plains and defined as 'a site where channelled flow ceases and floodwaters spill across adjacent alluvial surfaces' (Tooth, in press).

Two types of floodout have been identified; intermediate and terminal (Pickup, 1991; Tooth, in press). Channels reform downstream of intermediate floodouts which are more common on larger drainage systems. They are areas of rapid and extensive sedimentation and alternating erosion/deflation, often located in localised basins where the river is not laterally confined (Pickup, 1991) or upstream of barriers such as alluvial, aeolian or topographic highs (Sneh, 1983; Mabbutt, 1986). Terminal floodouts are areas where flows finally dissipate and no channel exists downstream. Tooth (in press) identified two causes of floodout formation: downstream reduction of discharge and barriers to flow (aeolian, hydrological/alluvial and structural).

Pickup (1991) has described the wide flat and generally featureless floodout plains in central Australia as 'unchanneled flood plains'. Similar to Mabbutt's (1977) outer flood plains, fluvial landforms become mixed wi .i aeolian features and the fluvial environment gradually disappears. Aggradation of the suspended load deposits in these areas occurs slowly if at all due to the easy breakdown and aeolian removal of the thin mud crusts, organic material and salt deposits (Pickup, 1991).

3.5. Central Australian Flood Plains

In a review of Australian river systems Tooth and Nanson (1995) note that the body of research available on Australia's desert streams is relatively small and is focussed on four main themes. Firstly there are studies which attempt to unravel the Cainozoic and Quaternary history of alluvial surfaces (Mabbutt, 1967; Perry, 1962; Baker *et al.*, 1983; Pickup *et al.*, 1988; Grant, 1994; Nanson *et al.*, 1995 and Croke *et al.*, 1996). Secondly there are those describing the effects of scale floods on the landscape (Baker *et al.*, 1983; Pickup *et al.*, 1988; Grant, 1994; Pickup, 1991 and Patton *et al.*, 1993). Thirdly, work on the sedimentary structures of central Australian ephemeral sand bed streams includes that of Williams, (1969; 1970; 1971), Mabbutt, (1977; 1986), Zwolnski, (1985) and Croke *et al.* (in press). Fourthly, three papers stand alone in attempting to describe and model the morphodynamics of central Australian arid zone fluvial systems (Pickup, 1991, Bourke, 1994; Tooth, in press).

Descriptions of central Australian flood plains generally are brief with most accounts describing flood plains as precursory comments to other research concerns. These include the work of Patton *et al.* (1993) (Ross River), Nanson *et al.* (1995) (Finke River) and Pickup *et al* (1988) (Finke River). A more detailed investigation of central Australian flood plains was undertaken by

Grant (1994) (Todd River), and Pickup (1991) (central Australian rivers) and Tooth (in press) (Sandover, Bundy and Woodforde Rivers). The conclusions of these studies will now be summarised for each of the four geomorphic regions identified in Chapter 2: the ranges, subdivided here into headwater and gorge reaches, the piedmont zone, the alluvial plains and the desert reach.

3.5.1. The Ranges

3.5.1.1. Headwater Reach

Grant (1994) described the narrow and discontinuous flood plains of the rocky Todd River headwater area, upstream of Alice Springs where channels are commonly 8-10 m wide and 2 m deep, carrying coarse sandy bedload. Fresh sediments, which line the main channels, form low overbank deposits and occur as sediment splays or threads on the older flood plain surface emplaced during relatively high frequency and low magnitude floods. Hummocky levee surfaces indicate discontinuous overbank deposits and active flood scour depressions. Although the sandy channel banks are steep they are usually vegetated and stable. Rock bars and saprolite are often exposed along the channel. Buried soils may be found in the often loamy stratified flood plain alluvium, composed of envelopes of fresh sediment lying within an older, more extensive alluvial plain that remains inactive in all but the largest floods; a slightly higher flood plain terrace was observed. The older flood plain, where intact, is a level plain of low relief with a slight levee along the channel margin. It has highly weathered red-brown earth soils with a strong texture contrast between the A and B horizons. Grant suggests that channels have remained fixed against their valley walls during the Holocene.

Flood plain formation in the upper reaches of the Todd River is dominated by vertical accretion with some intersection point sediments from adjacent alluvial fans (Grant, 1994). Large areas of the flood plain (<1 m thick) have been stripped by extreme floods of their A and upper B horizon, exposing crusted, often cemented surfaces. The original surface is preserved as remnant islands and terrace-like forms along the margins of the flood plains. Gullies, scalds and sheet erosion are further degrading the scours adding complexity to the pattern of sediment redistribution. Infilling of scours is mainly from adjacent alluvial fans. Grant (1994) suggests that a limited sediment supply from upstream inhibits active flood plain development.

3.5.1.2. Gorge Reach

Pickup *et al.*, (1988) describe the channel morphology and bedforms of bedrock gorge reaches of the Finke River which they distinguish as 'wide gorge' and 'narrow gorge' morphology. In the long straight sections of narrow gorge reaches, the sand and gravel channel bed is composed of linguoid bars, dunes and sections of exposed bedrock. Above the main channel, <200 m wide planar sand sheets which may have occasional linguoid bars on their surface occur laterally or as mid channel bars. Poorly developed point bars are found on the inside of bends and linear bars up to five metres high divide the channel. Flood plains are transient with narrow flat-topped lateral bars composed of fine sediment at various elevations above the channel bed. These are undercut and the boundary with the gorge wall is exploited by small channels.

In wide sections of the gorge, the braided gravel bed channel includes gravel dunes and sand and gravel braid bars and laterally, sand levees form incipient flood plains. Higher, older discontinuous gravel surfaces may have been deposited by past, high magnitude floods and aeolian dunes are located on stable surfaces.

3.5.2. Piedmont Reach

On exiting the MacDonnell Ranges the rivers traverse broad fans, and channel ribbons occur and switch about. Pickup (1991) presents a 'channel-levee model of flood plain formation for piedmont reaches where there is greater sediment supply and runoff than in the headwaters described by Grant (1994). Channel-levee flood plains are described as thin threads of gravel and sand of variable length (from 1 km to 100 km) originating in the ranges and terminating in a floodout. The channels are established within an envelope of recent alluvium and have levees that decrease in thickness away from the channel eventually forming a thin discontinuous veneer. The sediment envelope is disconformably inset into older alluvium and is composed of channel fill, vertical accretion and a small proportion of lateral accretion deposits (Pickup, 1991). In places of increased sediment supply, the channels become discontinuously braided and the lateral extent of the envelope is increased. Pickup proposes that the channel-levee complexes

represent a continuing episode of recent sedimentation where threads of relatively coarse textured sediment produced by increasing erosion upstream are gradually advancing outwards from the ranges. This pattern of progradation in semi arid rivers differs from that of humid rivers which do not prograde downstream. The sequence of floods in the 1960's accelerated the development, entrenching the channels and aggrading the levees and floodouts.

Others have described the piedmont reaches of central Australian streams as incised in older sediment. Tooth (in press) observed the modern channels and flood plains of the Woodforde, Sandover and Bundy Rivers as laterally confined by highly indurated mottled red terraces of gravel, sand and silt often containing abundant pedogenic carbonate. Patton *et al.* (1993) described the Ross River as incised (5 m) in the paleoflood and cemented Pleistocene alluvium on the Ross fan system. The channel trends from a single thalweg that meanders about alternate bars along the channel margins to downstream reaches that are braided with large mid-channel diamond-shaped bars. Active flood plain formation is indicated by the burial of River Red Gums by over 1 m of sediment.

3.5.3. Alluvial Reach

Nanson *et al.* (1995) describe a 4 km wide flood plain near the township of Finke where the channel is between 100 - 150 m wide and is described as one which zigzags in a series of straight reaches deflecting off bedrock bluffs at the valley sides. The channel bed consists of sandy tabular bars with extensive mud drapes forming a thin (<1 cm) layer of dry mud curls on the surface and thin mud strata at depth. The channel stratigraphy consists of coarse and medium sand layers varying in colour from brown to yellowish red. Flood plain morphology is relatively flat with a few low sandy mounds and shallow flood channels (<1 m) and the boundary between the flood plain and adjacent aeolian dunes is sharp. Nanson *et al.* (1995) infer that it is a lateral accretion flood plain with a small amount of vertical accretion and aeolian deposition.

Tooth (in press) describes flood plain systems along the Woodforde River, where there is a shallow burial of terraces by sand and silt. The channel banks are breached by large splays exposing the abrupt contact with the older underlying alluvial unit. The channel is flanked by low relief alluvial plains. In these locations overbank flows spread for great distances and there is greater potential for channel migration.

3.5.4. Desert Reach

Tooth (in press) describes the intermediate and terminal floodouts on the Sandover, Bundey and Woodforde River systems along the northern MacDonnell Ranges where the interaction of aeolian and fluvial systems are important along the margins of the floodout and thin layers of alluvial silt and clays overlie aeolian sand. The rates of aggradation are slow and there is low preservation of sedimentary structures. Large areas of sand plain and source bordering dunes (5 m high) are isolated by the floodout and incorporated into the alluvial sedimentary sequence, and waterholes may be located between these higher areas. Isolated elliptical to round pans linked to floodouts through narrow swales are surrounded by a dense growth of *Eucalyptus microtheca* with gilgai soils and cracking clays in the centre.

CHAPTER 4

Todd River Flood Plains

4.1. Introduction

This chapter describes the main characteristics of the Todd River flood plains based on data collected from twenty cross sections and seventy excavations and auger holes. All figures and tables are at the back of the chapter. Figures 4.1-4.7 show the location of the study sites. The morphostratigraphic legend is an A3 pullout at the back of this chapter and the sedimentary profile legend is an A3 pullout at the back of chapter 6. References to flood plain surface numbers on the right bank are prefaced by an 'R' and on the left bank by an 'L'.

In this study flood plain is defined as the body of sediment bordering and deposited by the modern channel. While some recently inundated surfaces deposited by a former flow regime, such as stripped or lower elevation paleoflood surfaces and Pleistocene alluvium, constitute the hydraulic flood plain (Nanson and Croke, 1992) they are not included as flood plains here. Flood plains are distinct from the higher terraces which are composed of cemented Pleistocene alluvium and Pleistocene and Holocene paleoflood alluvium. Two flood plain types have been identified: those which are *confined* and those which are *relatively unconfined.* Confined flood plains form in areas where lateral movement of the channel is impeded by resistant channel boundaries such as bedrock, coarse textured paleoflood sediments, alluvial terraces cemented by pedogenesis and cemented aeolian dunes. Relatively unconfined flood plains are in reaches where the channel tends to exhibit lateral movement, generally by channel switching, facilitated by erodable channel boundaries. A further description of the study sites is given in Appendix 2 and Tables 4.1 and 4.2.

This chapter is divided into four parts. The first three parts describe the characteristics of the Todd River flood plains, and based on this data, present two morphostratigraphic models and a sedimentary profile model which synthesise the surface and sub surface characteristics of these arid zone flood plains. Part four presents the flood plain chronology determined by radiocarbon analysis and ^{137}Cs and ^{210}Pb concentrations.

4.2. Flood Plain Characteristics

The following is an overview of the Todd River flood plains as a whole. In the MacDonnell Ranges the Todd River and its tributaries (the Ross River and Giles Creek) drain steep headwater catchments where there is little alluvial storage and coarse sediment is supplied from colluvial fans and rock falls. The Ross and Giles Rivers meander in 15 km and 25 km long gorges through ridge and vale topography formed on strongly folded bedrock. In the gorges coarse rounded cobble and gravel fill in the gorges of unknown depth, probably deposited by paleofloods, is partly incised by the modern river, and 2-3 m deep flood plains comprised of a basal gravel and sand channel fill are overlain by vertically accreted flood plains composed of abandoned channels and large coalescing vegetated islands. The Todd River north of Alice Springs flows over bedrock through the comparatively short (500 m) Wigley Gorge bordered by low sand berms and the Telegraph Station constriction (2 km), where narrow flood plains and a terrace sit within a bedrock channel. The single thread, low sinuosity channel flows south over the heavily dissected vale through the town of Alice Springs to Heavitree Gap (Fig. 6.2).

On leaving the ranges the Todd River flows across the Heavitree fan in a low sinuosity channel bordered by the Amoonguna and Lower Stuart surfaces (Litchfield, 1969; Fig. 6.3) and terminates in an intermediate floodout south of the confluence with Emily Creek. The channel reforms 12 km downstream augmented by flow from Roe Creek and Jessie Creek (Fig 6.4) (catchment area: 145 km²) where it is entrenched in Pleistocene fan and bedrock with little alluvial storage. The low sinuosity trench-like Todd River flows east along the wide strike vale, through Pleistocene sediments from southerly draining piedmont fan systems on the left bank and a wide aeolian sand plain with low dunes on the right bank. Flow is augmented by contributions from small piedmont tributary basins with unchanneled confluences. Flood plains along this 48 km reach are composed of low benches and higher stepped surfaces with locally braiding reaches and vegetated islands; in places the channel erodes aeolian dunes.

On exiting the ranges the Ross River (catchment area: 1,260 km²) is incised (~5 m) in the Pleistocene piedmont fan and Holocene paleoflood sediment and transports large volumes of sand and gravel, stored in large vegetated islands and braid bars, to the confluence with the Todd (Fig. 4.2).

Downstream from the confluence the Todd River widens and braids around gravel and sand bars and islands (Fig 4.2) and meets Giles Creek (catchment area: 1,000 km²) 24 km downstream as it changes direction to flow south. The Giles Creek tributary is confined between high paleoflood and cemented Pleistocene terraces south of the ranges and the narrow and steep channel facilitates little flood plain development (Table 4.1).

South of the Ross confluence until the Rodinga floodout (Figs 4.2-4.7) 95 km downstream, the channel is incised in cemented Pleistocene terraces, Pleistocene and Holocene paleoflood terraces, confined in narrow gaps through outlying bedrock ridges or bordered on one bank by Pleistocene sediments and on the other by paleoflood complex deposits. The low sinuosity gravel and sand channel local locally and flow diverges around vegetated islands. Short reaches (1-2 km) of extreme channel width (200-450 m wide) alternate with longer reaches 80 to 100 m wide. Flood plain fills are in the order of 2-3 m deep, are of variable width (5 m to 1.3 km) or form a thin smear over adjacent older surfaces deposited by flood splays and distributaries; discontinuous channel benches are common.

Upstream of Rodinga Gap a second 5 km wide, densely vegetated intermediate floodout comprising an anastomosing and braiding channel network, grades into a bare, flat, featureless, clay, silt and fine sand plain south of the Range. The channel reforms 6 km downstream confined in the swale of longitudinal dunes (Fig. 4.7) with discontinuous, thin and narrow flood plains. The channel bifurcates along multiple swales, breaching dunes until the terminal floodout in a large clay pan 50 km downstream.

Three criteria are selected to describe the flood plains; surface morphology, morphostratigraphy and sedimentary characteristics. Typical characteristics of flood plains are presented in Figures 4.8 and 4.9.

4.2.1. Flood plain morphology

The morphological flood plain is defined by its three dimensional geometry. The description of the morphology of the Todd River flood plains is based on surveyed cross sections of the channel, flood plain and terrace. Five distinctive features of flood plain morphology have been selected for discussion: multiple levels, surface channels, overbank deposits, swirl pits and remnant Pleistocene surfaces. These are shown on cross sections of sites at locations indicated in Figures 4.1-4.7. The cross sections show morphostratigraphic elements described later, as well as morphologic profiles.

4.2.1.1. Multiple levels

A notable characteristic of these flood plains is their stepped morphology and abrupt increases of elevation with distance from the channel (*e.g.* Fig. 4.10, 4.11, 4.12). Relatively flat surfaces at different levels are separated by well defined, steep scarps generally less than 1 m in height (*e.g.* Fig. 4.11, between surfaces Lf5 and Lf6; Fig. 4.12, surfaces Rf3 and Rf4). The scarps are the result of lateral erosion either by the main channel or a back channel. In many locations the surfaces of individual levels are narrow, measuring only 1 - 2 m (Fig. 4.12). Plate 4.1 shows typical step morphology.

The number of levels on either side of the channel is variable, for example, at one site the left bank flood plain has six 'steps' (Fig. 4.11) while the right bank flood plain has one (Fig. 4.10). Another example has five 'steps' on the right bank and three on the left (Fig. 4.12). In addition to a variable number of levels, surfaces are also unmatched in elevation across the channel (Fig. 4.12). The downstream extension of individual steps is generally limited by local factors such as a tributary entrant, a change in local channel pattern or a bedrock outcrop.

4.2.1.2. Surface channels

While many of the flood plain surfaces are flat-topped (*e.g.* Fig. 4.13, Rf1; Fig. 4.14, Lf3), most tend to have been reworked by surface channels or recent overbank deposition of surface veneers or large migrating bars. Two types of flood plain surface channels are described, here referred to as back channels and flood channels.

<u>Back channels</u>

Back channels are single-thread, well-defined channels, bordering the scarp between flood plain steps or between the flood plain and terrace (*e.g.* Fig 4.11 and 4.12; Plate 4.2), which operate as flood channels. Their importance in the dynamics of channel switching in relatively unconfined flood plains warrants individual description here.

Plate 4.1. Left bank Late Holocene flood plain on Giles Creek set against Pleistocene terrace. The remnant steps in the flood plain are formed by successive lateral truncation. The lower flood plain step sediments have recently buried the bases of the mature *Eucalyptus camaldulensis*.

Plate 4.2. Right bank back channel near Anabranching site. Note the outcrop of red Pleistocene aeolian sediments to the left. Flow is towards viewer. A thin oblique smear of modern alluvium drapes over an erosion step in the Pleistocene sediments close to the back channel. Main channel is to the right and Plate 8.1 is aerial photo of the site.

Generally they are narrow features averaging less than 10 m wide but some exceed 50 m, usually where they are abandoned main channels (Fig. 4.15, surface Lf5 and Fig. 4.16, Lf1 is 125 m wide). Depths of the channels vary and some are punctuated by swirl pits, the most extreme example measuring 4 m deep (Fig. 4.34, surface Lf4). Benches (~0.75 m high) sometimes are located on the convex bank opposite where these channels laterally erode the scarp of the adjacent, higher level. Back channels occur at various elevations across the flood plains; some clearly are erosional and significantly alter the surface step morphology (*e.g.* Fig. 4.12, surface Lf1; Fig. 4.16, surface Rf1 and Rf2), others both incise and aggrade the flood plain surface (*e.g.* Fig. 4.11).

Flood channels

Well developed braided flood channels are common features on both the confined and relatively unconfined flood plain surfaces (Fig. 4.8 and 4.9). They are differentiated from back channels as they do not necessarily travel along the flood plain scarp, are multiple thread and tend to have smaller channel cross sections. There are two types: those which scour the pre-flood surface (*e.g.* Fig. 4.34, surface Lf3; Fig. 4.17, Rf1; Fig. 4.18, surface Lf1, Lf2) and those which divide around freshly deposited overbank bars (Fig. 4.19 surface Lf2). Individual braided channels have widths generally between 2 m and 20 m and although they have been observed to feed into or out of back channels, they tend to be independent of them. Flood channels branch from the main channel and travel parallel to it, often re-entering further downstream or terminating in a distributary flood plain sink. At Site 'A', the braided flood network changes across the flood plain surface from a set of relatively deep and narrow channels to a 300 m wide and shallow braided channel (see between 380 m and 680 m on surface Pf4 in Fig. 4.20) with large diamond-shaped bars (110 m wide) and hummocks. The distributary shallows rapidly and terminates against an aeolian dune, 800 m from the main channel in a flat mud floodout overlain by a thin sheet of aeolian sand, and upstream flows into the swale via a breach in the dune.

4.2.1.3. Overbank deposits

In addition to the surface channels recent overbank deposits interrupt the generally flat topography of the flood plain surface. These may be individual bars which extend downstream from trees or larger sand sheets and splays and bars which are often stepped (<60 cm) where an avalanche face has migrated along the flood plain surface.

On the right bank close to the Todd/Ross Confluence site, a series of 1 m by m streamlined hummocks built of fine red sand overlie the Pleistocene sediment. Although the surface is populated with Coolabahs and River Red Gums, these hummocks do not appear to have been deposited around prior vegetation. Similar streamlined hummocks occur at site 'A' (Plate 4.3, 4.4, and 4.5), located 430 m from the main channel along the bed of a flood distributary. They are between .9 and 2.5 m long and .35 and .45 cm thick, and are partly buried by the surrounding flood plain sediments.

4.2.1.4. Swirl pits

Swirl pits result from vortices in deep flows (Gardner, 1977) and are also associated with eddies around obstacles. They are found on the surface of flood plains (Fig. 4.14, Rf2; Fig. 4.34, Lf4) and in the main channel (Plate 4.6, Fig. 4.16 and 4.21). Where fully formed, they are elliptical in planform, deepest at their upstream end and can exceed 2 m depth. They usually form close to or around obstacles such as riverine vegetation (most commonly *Eucalyptus camaldulensis,* Plate 4.6) and adjacent to flood plain scarps. At site 'A', surface Rf2 (Fig. 4.22) is punctuated by a series of scours, similar to those shown in Plate 4.7, which developed around gidgee trees. At the No. 5 Bore site (Fig. 4.21) swirl pit erosion of the channel bed dominates bed morphology as does a similar feature on surface Lf4 (Fig. 4.34). These swirl pits are transient features and regularly infill with sediments.

4.2.1.5. Remnant Pleistocene surfaces

Reaches of the modern Todd channel are incised in cemented Pleistocene aeolian and alluvial sediment. Where the channel actively cuts the Pleistocene sediments, terraces up to 7 m above the channel are formed (Plate 4.8). In other locations stepped flood plains are inset against the Pleistocene terrace (Fig. 4.12 and 4.30). Where the flood plain has invaded a Pleistocene dune field, remnant, partially eroded and intact longitudinal dunes interrupt the relatively flat flood plain (*e.g.* Fig. 4.20).

Two sites have been selected here for further description, to illustrate confined flood plain morphology (Mosquito Bore 1 Fig. 4.27) and relatively unconfined flood plain morphology (Expansion Scour site Fig. 4.30).

Plate 4.3. This small hummock of fine sand (90 cm long by 25 cm high) is one of a series of such features across this wide flood plain close to the Rodinga Range.

Plate 4.4. Located at site 'A' this large hummock is partially buried by subsequent flood plain aggradation and is one of several along a flood distributary channel. Flow is towards the left of the photo.

Plate 4.5. Hummocky ground at the distal end of a flood distributary at site 'A' shows aeolian deposition around trees (a) underlain by a horizontal mud deposit (b). The source of aeolian sediment is a longitudinal dune to the right of the plate.

Plate 4.6. Swirl pit around River Red Gum (*Eucalyptus camaldulensis*). Note figure in right foreground of the scour for scale. Flow is towards viewer.

Plate 4.7. Looking upstream on surface of Rf2, Site 'A'. Swirl pits formed around the base of trees.

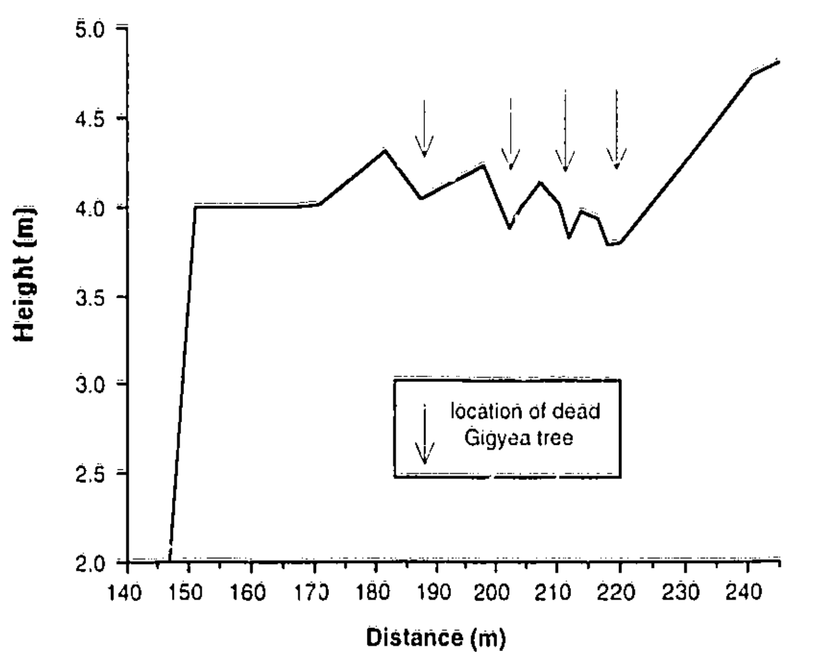


Figure 4.22. Swirl pit morphology, surface Rf2, site 'A'

4.2.1.6. Type site of confined flood plain morphology

The Mosquito Bore 1 site is located at Mosquito Bore (Fig. 4.4, Plate 4.9). The channel flows through a gap in an outcrop of Proterozoic Bitter Springs formation composed of dolomite, limestone, siltstone and sandstone. The channel is essentially straight with changes in channel direction at the bedrock ribs. Channel width is 102 m with a flood plain to channel ratio of 2:1 and the local channel gradient measured along a channel distance of 279 m is .00063 m/m. A bore on the left bank (see Plate 4.9) is a local watering point for cattle and well worn cattle and truck tracks cross the channel and flood plains.

A surveyed cross profile of the site is presented in Figure 4.27. The flood plains display a marked stepped morphology with steep and sometimes metre-scale risers. The flat or gently sloping flood plain surfaces are interrupted by surface channels. Surfaces such as Rf4 are 55 m wide while other surfaces are narrow such as Rf2 (3.5 m wide); the latter was truncated by an episode of channel widening. The right bank flood plain has five distinct alluvial surfaces. The four lower surfaces are covered by recent flood deposits and trash lines measured from trees in the centre of the channel confirm inundation over the lower three surfaces.

Two channels are incised into surfaces Rf4 and Rf5 (Fig. 4.28). Both are located at breaks in slope and activated by overbank flow and local runoff. The right bank alluvial units are inset in a bedrock embayment (Plate 4.9) and flow from the back channels abuts a bedrock ridge and drains into the main channel (Fig. 4.28, Plate 4.10). Surface channel 2 shows a small bench on the right bank (Fig. 4.27).

The left bank flood plain has four alluvial surfaces which do not match the elevations of the right bank flood plain surfaces (Fig. 4.27). They display the flat-topped and steep face morphology similar to the right bank but do not have prominent surface channels. The highest surface on the left bank is a Pleistocene aeolian deposit composed of well sorted fine red sand overlying bedrock. It grades into the terrace along an indistinct boundary.

4.2.1.7. Type site of relatively unconfined flood plain morphology

The location of the Expansion Scour site can be seen in Figure 4.5 and in Appendix 4 (Image). It is bounded on the right bank by an extensive sand and gravel paleoflood deposit and on the left by a cemented Pleistocene terrace (Fig 4.29). Located at the beginning of a 'beaded' reach, the modern channel width increases dramatically from a minimum of 50 m to a maximum of 450 m. The 50 m wide channel abuts the paleoflood terrace and the left bank flood plain is 417 m wide (flood plain to channel ratio is 8.4:1). Channel gradient is low measuring .00087 m/m.

The surface of the flood plain has three levels (Fig. 4.30, Lf1-f3) separated by steep risers ~1 m high. Back channels drain and incise the surface interface, the widest (25 m), on surface Lf2, is fed by two left bank tributaries (Fig. 4.29). The back channel on surface Lf1 fed by overbank flow from the main channel and widens downstream along the flood plain surface. The coarse sand and gravel on the surface of Lf1 appear to have been recently reworked by an anastomosing network of surface channels.

4.2.2. Flood plain morphostratigraphy

The morphostratigraphic unit is defined as:

> comprising a body of rock (or sediment) that is identified primarily from the surface form it displays; it may or may not be distinctive lithologically from contiguous units; it may or may not transgress time throughout its extent. (Frey and Willman, 1960, p. 7)

In meandering systems, conventional flood plains are comprised of laterally accreted, coarse textured channel sediments overlain by overbank, vertically accreted finer textured sediments (Lewin, 1978). In braided systems flood plains are predominantly composed of coarse textured channel sediments overtopped by a thin veneer of finer textured, vertically accreted, overbank sediments (Reinfelds and Nanson, 1993). The Todd River flood plains have characteristics of both but are closer to the braided river model in that they are formed predominantly by vertical accretion but have a thicker overbank component and a more complex morphostratigraphy.

The principal component of the Todd River flood plain is generally a remnant fill which has been stripped and/or laterally eroded and can be composed of cemented, structureless Pleistocene alluvium or coarse textured,

Plate 4.8. Vertical Pleistocene terrace bank profile 200 m downstream from Stud Bore 1 site. Note the oblique accretion and talus towards the base and colonisation by young River Red Gums. Flow is towards the viewer and tyre tracks for scale.

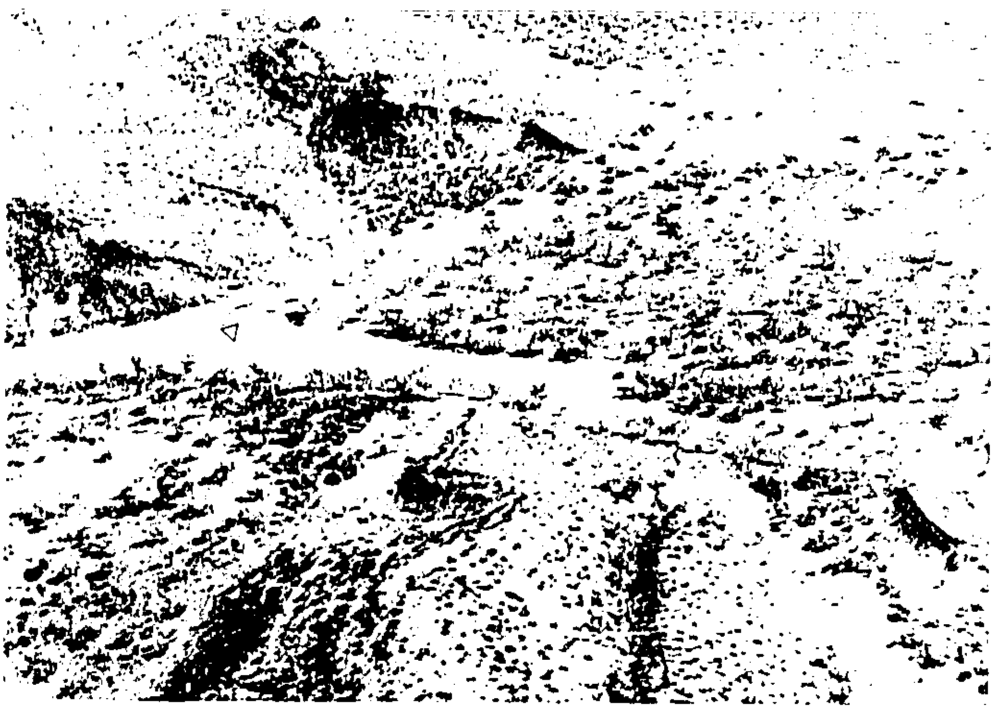


Plate 4.9. Oblique photo of Mosquito Bore 1 site confined between ridges of Proterozoic sandstone. Study site is indicated by arrow. Cross section roughly follows vehicular track to bore in lower left of photo. Flow is towards the left of the photo. Slack water deposits are located in caves in right bank bedrock ridge (a).

Plate 4.10. Mosquito Bore 1 site following the January 1995 flood. Flood plain surfaces Rf1 and Rf2 have been removed by lateral erosion. Freshly deposited fine red sand on the surface of Rf3 (a) extends obliquely up the face of surface Rf4 (b) and contrasts with the white channel sands. Note the flood plain surface channel fill (dashed line). Flow is towards the left.

Plate 4.11. Photo of right bank at Ross River Striped site. Flow is towards the left. River Red Gums are buried by 1m of sediment and their exposure indicates recent channel widening. The scarring of the trunk of the central tree may indicate a previous deeper episode of burial by sediment. Trees are growing on the surface of a coarse textured paleoflood deposit.

paleoflood channel fill sediments. Alternatively, it may be composed of a wide single or multiple channel fill. Examples include the paleoflood fill in Figures 4.19, and 4.24-4.26; the Pleistocene remnant underlying the right bank flood plain in Figure 4.14 or the left bank flood plain in Figure 4.11; and the channel fill underlying surfaces Rf2-4 in Figure 4.18.

The following sections focus on morphostratigraphic units which individually appear relatively minor in extent but together account for a substantial component of the Todd River flood plains, in addition to illustrating the complex fill associations. The Todd River flood plain is characterised by a complex assemblage of morphostratigraphic units, each of which is a body of sediment bounded by upper and lower surfaces with a defined geometry. Below is a description of the four principal morphostratigraphic units identified: channel fill, inset fill (channel and flood plain), flood plain veneer and swirl pit fill. Also described are remnants of morphostratigraphic units which have been post-depositionally altered by erosion, termed flood plain remnants (surface, buried and paleoflood). Two morphostratigraphic models (confined and relatively unconfined) are presented which summarise the main morphological and morphostratigraphic characteristics of the Todd River flood plains (Fig. 4.8 and 4.9).

4.2.2.1. Channel fill

Flood plain channel fills are elongated, sinuous bodies of sediment, dominantly sandy, usually flat-topped and always convex-down at the base. Their form derives from the flood channel or abandoned main channel which preceded them. Fill dimension varies, with larger flood channels up to 3 m deep and 30 m wide, typically found in relatively unconfined flood plains as back channel fills. Smaller flood channel features are generally less than 1 m deep and 3 m wide and are found in both types of flood plain locations. These will be described first.

Small flood channel fill

The surface of the flood plain at the Anabranching site (Fig. 4.18, Rf2) is reworked by a series of small, shallow, braided channels which run sub-parallel to the main channel and are perpendicular to the sub-surface channel fills. They fill channels which have eroded into the flood plain surface to a resistant mud layer. The central channel (Plate 4.12) is approximately 120 cm wide and 63 cm deep and the dip of the basal channel fill sediments is approximately 18° grading up to sub-horizontal layers. The channel fill sequence is composed of a series of large drape sheets which blanket the wider flood plain, and individual layers are traced continuously between the braid channels. By way of contrast the channel fills at Todd/Ross Left Bank (Fig. 4.11) consist of aggraded channel fills which are separate from the adjacent flood plain sediments. Some fills tend to be coarse, fining upwards from pebbly sand to sand (*e.g.* excavation C, Fig. 4.11); others are finer and are infilled with mud and sand (*e.g.* excavation F, Fig. 4.11).

<u>Large flood channel fills</u>

Large flood channels are more frequently found in relatively unconfined flood plain sites. The site in Plate 4.13 is located downstream from the Expansion Scour site where the channel narrows from 450 m to 100 m (Fig. 4.5 Appendix 4, Image 8). On the right bank is an exposure of a large-scale channel fill 2 m deep and 20 m wide (Plate 4.13). The channel is cut into planar bedded white channel sand and infilled with medium to fine textured red sand. The basal channel sediments gently dip along the former channel bed infilling towards the surface with horizontally bedded sediments. The upper layers are continuous across the wider flood plain surface. This channel drains perpendicular to the main channel and has no surface morphology. A larger back channel runs transverse to this buried channel and is actively dissecting the flood plain surface. Large scale channel fills are important morphostratigraphic units in relatively unconfined flood plains.

Another example of a large scale channel fill is at the Todd/Ross Confluence site (Plate 4.14). This channel fill unit lies 175 cm below the flood plain surface and is probably a former flood channel from the Ross River. The channel has incised a cemented gravelly sand unit and is approximately 117 cm deep and 8 m wide. Although the right bank of the channel fill is truncated and replaced by a flood plain inset the fill appears to have been symmetrical. The basal channel fill sediments dip steeply and grade into horizontally bedded sediments.

4.2.2.2. Insets

An inset fill is a elongate, roughly rectangular body of vertically accreted gravel, sand and mud sediment. Inset fills are usually flat-topped, but may dip channel-ward or have wavy eroded upper boundaries, and have vertical sides, and generally flat bases. Two types are identified here: the flood plain inset and the channel bank inset (bench and oblique).

<u>Flood plain insets</u>

Flood plain insets are composed of sediments deposited against the erosional scarp of a flood plain step or terrace away from the main channel. They unconformably overlie cemented Pleistocene aeolian/alluvial sediments, Pleistocene and Holocene paleoflood sediments and modern flood plain sediments and are the larger of the two morphostratigraphic inset units. At the Desert Reach site (Fig. 4.13) the right bank flood plain (65 m wide and 2 m deep) is inset against and deposited over an eroded bench sculpted from the cemented Pleistocene longitudinal dune flank. At the Stud Bore 1 site (Fig. 4.12) the flood plain unconformably overlies cemented Pleistocene sediment and is composed of a series of flood plain insets (~47-135 m wide and ~1-2 m deep) arranged in steps, laterally inset against one another with the boundaries exploited by back channels. Some of the insets are narrow and have been laterally eroded by a channel widening event (*e.g* Rf4). The bedding indicates that the fill is vertically accreted. At Mosquito Bore 1 site (Fig. 4.27) the insets are similarly set into one another. Here the erosional interfaces between the steps were confirmed by the abrupt truncation of sedimentary layers at steps. The depth of flood plain inset fill averages 3 m and can equate to channel depth.

<u>Channel bank insets</u>

Bench insets

Channel bench insets are composed of sediments deposited within the channel, are set unconformably against the channel bank and may overlie the channel bed sediments (Plate 4.21) or be conformable with them (*e.g.* Fig. 4.34). Channel bench insets are generally flat-topped, usually have a vertical face where exposed by a channel bank and can extend downstream for over 100 m or can be restricted by the local bank. For example, a channel bench deposited during the January 1995 flood was confined to the bank notch downstream of a large River Red Gum. It is 150 cm high and 2 m wide, extends 10 m downstream and has a vertical eroded face with some mud obliquely deposited towards the base. It is composed of fine red sand and organic debris in finely-bedded climbing ripples and unconformably overlies pebbly white coarse channel sand. At Todd/Ross Right Bank site (Fig. 4.10) surface Rf1 is a bench which is conformable with the underlying channel deposits. The 20 m wide bench is unconformably attached to the vertical face of the cemented Pleistocene terrace and younger channel deposits are inset against the vertical face.

Oblique Insets

Where the channel bank is stepped due to the stripping of less resistant layers in the flood plain sediments or oblique due to bank collapse of loose sand (*e.g.* Plate 4.8, Plate 4.13), sediments are deposited obliquely against the bank. While these morphostratigraphic units are not large-scale, they do contribute to the complex morphostratigraphy. At Anabranching site a thin smear of mud-rich sediment is obliquely inset over and against the irregular morphology of the channel bank (Plate 4.2). Oblique deposits are preferentially located towards the base of the channel bank as the top of the bank is generally vertical. Where the oblique deposits extend over the bank they are termed flood plain veneer, described below.

Centimetre-scale insets are deposited on minor steps in the bank morphology. One such example is shown in Plate 4.15 where thinly laminated fine red sand slopes obliquely up the eroded bench at the base of the Pleistocene profile. This channel inset has been laterally truncated by channel flow and channel sand emplaced against the eroded face, partially burying the inset. This combination of an eroded bench, channel bank inset and main channel fill inset results in a complex local morphostratigraphy at the small scale.

4.2.2.3. Flood plain veneer

Flood plain veneer fills are elongate irregular-shaped bodies of layered sediment, dominantly sand and mud, which derive their form from the morphology of the underlying surface which can be stepped or oblique. The larger flood plain veneers tend to have stepped lower boundaries and convex-oblique upper boundaries.

Plate 4.12. Typical sedimentary facies of island flood plain at Anabranching site. Three morphostratigraphic units: a basal channel fill (a) overlain by finer flood plain sediments (b) incised by braid channels (c) filled by sheet-drape deposits. Note the kingfisher nests in bank.

Plate 4.13. Large scale channel fill. Note evidence of lateral channel erosion and incision by the exposure of the root system of the large River Red Gum.

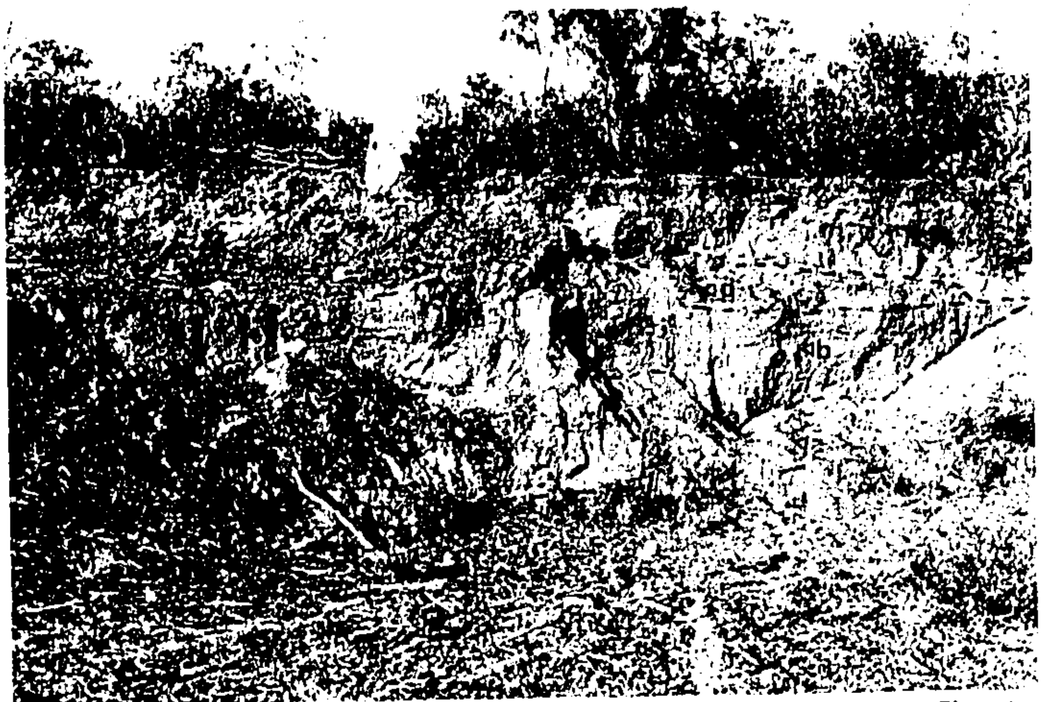


Plate 4.14. Exposure in back channel at Todd/Ross Confluence site. Flow is towards the right. The section is approximately 5 m high. Unit a underlies the channel fill and is covered by talus, Unit b is a channel fill incised into Unit a, Unit c is a large scale flood plain inset. Unit d is a small scale flood plain inset.

Plate 4.15. A small scale channel bank inset (a) deposited on eroded bench (b). Note the oblique style deposition up the bench, the eroded face of the inset and the channel bed sediment (c) partially burying the inset.

Plate 4.16. Flood plain veneer extends from the horizontal flood plain surface and dips towards the channel at Ross River Striped site. View looking downstream.

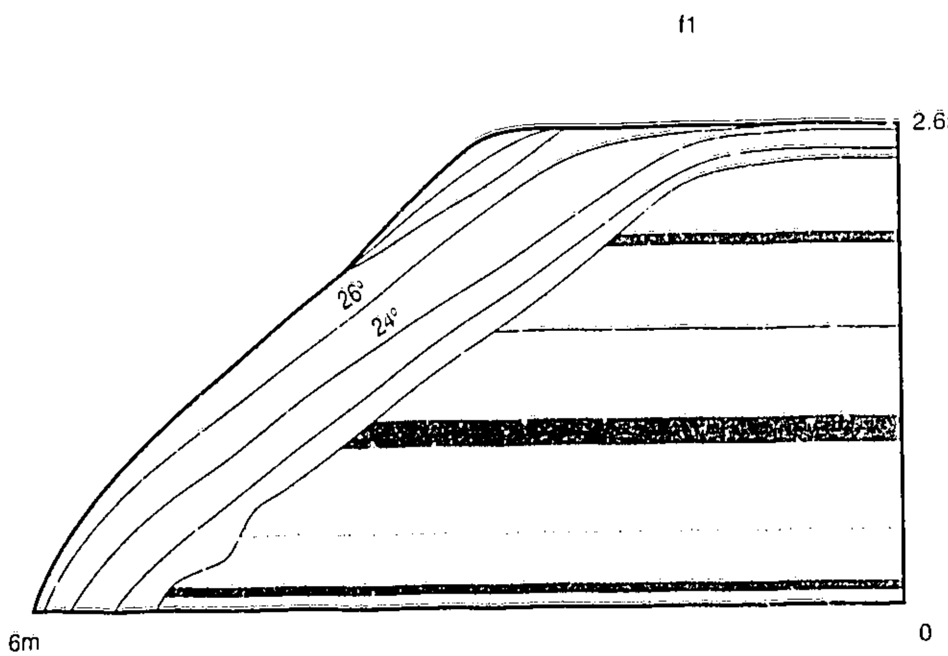


Figure 4.31. Morphostratigraphy of flood plain veneer overlying flood plain inset. Based on excavations at site 'A'.

The lapping up of younger alluvial sediments onto higher, adjacent surfaces was termed terrace veneer (Brakenridge, 1984). In this study the term is changed to flood plain veneer to indicate that this morphostratigraphic unit was only observed in the flood plain and not in adjacent terraces. As well as maintaining the horizontal surfaces, these veneers are found across steps on the flood plain and can extend over the channel bank. At confined sites the veneer may drape the bedrock boundary itself.

A morphostratigraphic model of flood plain veneer sedimentation based on excavation of the surface and bank of Rf1 at Site 'A' (Fig. 4.14) is presented in Figure 4.31. Within the channel, the veneer deposits lie at an angle of 24°-26°, overlapping the horizontally bedded, vertically accreted flood plain inset sediments, and extend across a horizontal distance exceeding 6 m. The oblique veneer is inferred to have accreted during four floods with two cut and fill layers towards the top of the sequence. These deposits subdue any small morphological irregularities on the flood plain. Individual layers of the veneer thin towards the base and top of the profile but appear uniform up the bank slope. Layer thickness is greatest on the channel-ward aspect and least on the flood plain surface. Veneer accretion is therefore more effective in laterally accreting the bank (80 cm) than vertically accreting the flood plain surface (30 cm).

Following the January 1995 flood at Mosquito Bore 1 site a layer of fine red sand was deposited on top of surface Rf3 and obliquely onto the face of surface Rf4 (Plate 4.10). The fresh sand was rapidly colonised by grass which may assist preservation of the sand veneer. Downstream from Mosquito Bore 1 site where the channel abuts the outcropping bedrock ridge, oblique smears of sediment extend up the rock face. The morphology of the rock wall is irregular and the thickness of the inset deposit varies according to the steeper and flatter portions of the rock slope, with the thicker and more horizontally bedded units located at the local break in slope. The deposit has two soil layers indicating stability.

4.2.2.4. Swirl pit fill

Swirl pit fills are elongate bowl-shaped bodies of sediment, dominantly sandy, usually flat-topped and always convex-down at the base; their form is derived from the scour which preceded them. Swirl pits along the Todd River are incised into cemented Pleistocene alluvium and unconsolidated sandy flood plain and channel sedime s. Dimensions vary between 1-3 m deep and 1-6 m wide. The following description is based on swirl pits partially and fully exposed at the surface and exposed in section.

After floods sand bars were observed to have migrated into the depressions, but more often they are partially infilled with structureless, or thinly laminated layers of desiccating mud up to 30 cm thick. One such swirl pit that was visited two days after the formative event at No. 5 Bore (Fig. 4.21) still contained 2 m of water with silty clay on the base and lower slopes of the swirl pit. Prior to this event there was no mud at the base of the swirl pit, perhaps due to desiccation and aeolian winnowing or burial by sediment slumped from the pit sides. Swirl pits also fill with organic debris such as tree branches and leaf litter (Plate 4.17). Swirl pits located away from the main channel can remain prominent features containing little fill.

A channel widening event that occurred at the Anabranching site during the January 1995 flood exposed a swirl pit morphostratigraphic unit formed in association with a Coolabah tree (Fig. 4.32). The tree is alive and in growth position but the former surface at the base of the tree has been scoured. The swirl pit sediments were deposited after at least 80 cm of alluvium was scoured from around the base of the tree perhaps several times. Past maximum scour depth excavated the planar bedded sand and exposed the tree root system. The boundary between layers is generally erosive and the minor depositional events were preceded by scour probably during the same event. At least eleven depositional events have filled this tree scour. Scouring tends to have been concentrated on the downstream side of the tree as indicated by the infill lenses not found in similar abundance on the upstream side (Fig. 4.32).

4.2.2.5. Flood plain remnant

A flood plain remnant is the eroded core of a previously more extensive flood plain. Remnants have unconformable contact with adjacent alluvial sediments and are of two types: those which are visible as truncated steps at the flood plain surface and those which are buried.

Surface remnants

Flood plain remnants visible at the surface have been truncated by lateral erosion of the main channel (Plate 4.1) or by back channels.

Morphostratigraphically, the remnant sediment body typically extends vertically through the full depth of the flood plain and is flat-topped with vertical or slightly concave lateral bounding surfaces. (*e.g.* Fig. 4. surfaces Lf3, Lf1, Rf1 and Rf2; Fig. 4.24, surface Rf1; Fig. 4.12, surface Rf4). Some of these are difficult to detect on the topographic profiles as they have minor morphological expressions, but have been shown by auguring to be >2 m thick. They are thus important indicators of flood plain destructive processes. Where the channel runs adjacent to and over resistant sediment of earlier flood plains, such as cemented Pleistocene alluvium or coarse textured paleoflood deposits, the sculptured surface remnants are preserved as benches in the channel bank (*e.g.* Fig. 4.12 surface Lf1a and Fig. 4.23, surface Lt1).

Buried remnants

Flood plain remnants may be buried by flood plain or channel sediments. Where Pleistocene sediment has been actively sculptured by the modern channel it is also termed flood plain remnant and most remnants fall into this category. Often, the buried remnants display a stepped face slope (*e.g.* Fig. 4.11, 4.30). In many locations the flood plain remnants extend under channel sediments (Plate 4.15).

4.2.2.6. Type site of confined flood plain morphostratigraphy

Although the Mosquito Bore 1 site (Fig. 4.27) includes examples of flood plain insets and surface flood plain remnants, it is not used as the morphostratigraphic type site as it does not have all the elements that can occur at a confined site and is not as well verified stratigraphically as the Todd/Ross site.

The Todd/Ross type site lies at the confluence of the Todd and Ross Rivers on a meander bend imposed by the fossiliferous siltstone and minor dolomite of the Arumba Sandstone formation (Figs. 4.2, 4.33, 4.34, Plate 4.18). The channel in this reach splits around broad vegetated shoals. Channel gradient downstream of the confluence is .00316 m/m and the flood plain to channel ratio is 1.7:1.

The Ross River is the dominant system in this reach as many of the flows from Alice Springs fail to reach this junction. On the outside of the meander bend (right bank) sandy cemented Pleistocene alluvium is periodically exposed in vertical cliff bank sections. The left bank flood plain consists of a series of alluvial units set against a cemented paleoflood terrace. The left bank flood plain/terrace boundary is a sinuous eroded scarp (Plate 4.18) and a series of younger alluvial units has been inset against this scarp. Because of the complex and variable flood plain plan form, three cross sections were examined along this reach (Fig. 4.33). The Todd/Ross Confluence site was measured between the higher Pleistocene surfaces on either side of the channel (Fig. 4.34) and the remaining two cross sections focussed on the flood plain morphostratigraphy (Fig. 4.10 and 4.11). The Todd/Ross Left Bank (Fig. 4.11) was measured across the left bank inset flood plain surfaces and deep back channel. The Todd/Ross Right Bank site (Fig. 4.10) was measured across the right bank surfaces. Comparison of the three sections indicates the downstream variability of flood plain morphology along this reach.

The Todd/Ross Confluence cross section (Fig. 4.34) shows three morphostratigraphic formations that comprise this site. The first is the cemented Pleistocene terrace. The second morphostratigraphic component is the main channel fill. The depth of the channel fill at this location could not be determined due to the free flowing nature of the pebbly coarse sand and mud ball channel sediments. Exposures show that the 90 m wide channel fill consists of well defined trough cross bedding and large sets of planar tabular cross bedding.

The third formation is the flood plain composed of surface channel, back channel, swirl pit (Plate 4.14) and flood plain insets. At the junction between the Lt1 and the flood plain is a deep back channel which is incised by a 4 m deep swirl pit. The flood plain is comparatively wide (155 m) and composed of a large sand and gravel basal horizon which has been unevenly eroded by channel widening and surface channel scour and rebuilt by the deposition of flood plain insets. There are five surfaces, four of which (Lf1, Lf2, Lf2a and Lf3) are inset unconformably into one another. The width of surface Lf1 indicates lateral erosion by the channel. Surface Lf4 underlies the flood plain and has been recently aggraded by a coarse white sand bar which is migrating down the slope of the large swirl pit. The lateral extent of this bar has been truncated by flows along surface Lf3. Surface Lt1 contains a channel fill unit unconformably overlain by vertically accreted sand and gravel sheets (Plate 4.14).

Plate 4.19 illustrates the relationship between alluvial units at the micro-scale in this reach and shows examples of erosional unconformities, flood

Plate 4.17. The gradual infilling of a swirl pit formed around a tree is assisted by the concentration of coarse woody debris on its upstream side.

Plate 4.18. An oblique photo of the Todd/Ross confluence (1994). The Ross River (R) flows from the top right to the bottom right, the Todd River (T) enters above the centre left.

plain veneer, step morphology and inset deposition. This complex flood plain morphostratigraphy indicates that processes of stripping, flood plain veneer and inset development occur at the small scale, that is, the scale of individual sedimentary layers, generally less than 40 cm. Thi complexity is masked by the surface morphology which appears to be a flat-topped low gradient flood plain form. The partially buried young Red Gum (*Eucalyptus camaldulensis*), approximately 20 years old, indicates the recent development of this morphostratigraphy.

Of equal complexity, but larger in scale (meso-scale) is the Todd/Ross Left Bank site (Fig. 4.11) where there are examples of flood plain insets, buried flood plain remnants, flood plain veneer and surface channel fill morphostratigraphic units (Fig. 4.11 a-d) The step-like morphology of the flood plain remnant (Fig. 4.11 a) is erosional in nature. the upper erosional step following a stratigraphic discontinuity in the sediments. A large flood plain veneer (Fig. 4.11 b) which appears at the surface as f6, underlies f5 and is composed of a vertically aggrading sequence dipping at approximately 15° over the eroded scarp of the buried flood plain remnant. This veneer has been subjected to lateral erosion and vertical stripping as evidenced by the stepped surface morphology. There are three back channels (Fig. 4.11 c), two of which have partially infilled with a fining upward sequence from structureless pebbly sand to planar bedded fine and medium red and white sand. The channel-ward part of surface f5 has aggraded by the deposition of a flood plain inset (Fig. 4.11 d) and the overbank migration of a bar. The morphostratigraphy at this site indicates the complex assemblage of fills and the episodic aggradation and erosion of the flood plains.

Figure 4.10 illustrates the relatively simple morphostratigraphy of the right bank surfaces. The channel inset (Rf1) is unconformable with Rt1 and the channel bed sediment is inset unconformably against the laterally eroded face of surface Rf1.

4.2.2.7. Type site of relatively unconfined flood plain morphostratigraphy

The Expansion Scour site (Figs. 4.5, 4.29, 4.30) morphostratigraphic figure indicates four formations: the left bank Pleistocene terrace (Lt1) the right bank paleoflood terrace (Rt1, t2), the flood plain and the channel. Excavation of the flood plain indicates that the morphostratigraphy is complex (Fig. 4.30).

As at the Todd/Ross Confluence site, the active flood plain is underlain by a buried flood plain remnant (i) comprised of a gravel and sand channel deposit which has been laterally truncated and vertically stripped displaying a stepped morphology. These steps are coincident with cemented sedimentary layers. The flood plain has five morphostratigraphic units, three of which are the eroded remnants of flood plain insets and flood plain veneers (ii-iv); the fourth is a channel inset (v) and the fifth a surface channel fill (vi). The sediments in units (ii) and (iv) dip channel-ward and are the surface flood plain remnants of incipient flood plain veneers which do not appear to have extended over the riser. Surface channel fill (vi) coarsens upward from a basal mud to medium and fine planar bedded sand. Unit (v) is a gravel and sand channel bar reworked by minor surface channels and a larger scale back channel. Unit (iv) was emplaced on an eroded step formed at the stratigraphic boundary between units (v) and (iii).

4.2.3. Morphology and morphostratigraphic models

Based on the data presented on fl[illegible] plain morphology and morphostratigraphy, two models are presen[illegible] which summarise the main characteristics of the Todd River confined and relatively unconfined flood plains (Fig. 4.8 and 4.9). The models share similar features, and differ principally in terms of the scale of the back channels, the presence of consolidated boundary sedimen[illegible] and the presence of large island features in the channel and attached t[illegible] the flood plain. The characteristics are summarised in Table 4.4.

Confined flood plains of the Todd River lie between cemented Pleistocene terraces, coarse textured or cemented paleoflood terraces, and bedrock outcrops. They tend to border a single thread channel and have low flood plain to channel width ratios. The flood plain surface has a pronounced stepped morphology and is reworked by surface channels and swirl pits (Fig. 4.8 2, 3). There are four surficial morphostratigraphic units found in confined flood plain formations: channel fills, flood plain veneer, insets (channel and flood plain) and swirl pit fill (Fig. 4.8, 13, 12, 4, 5). These morphostratigraphic units may be been post-depositionally eroded and survive as sculpted surface and buried flood plain remnants (Fig. 4.8, 6, 7). They can overlie channel deposits, be preserved as insets or as modified paleoflood or Pleistocene remnants.

Relatively unconfined flood plains of the Todd River are found in locations where lateral migration of the channel is possible *e.g.* between paleoflood terraces composed of erodable gravelly sand. They tend to border wide channels which locally braid or flow around well vegetated islands and generally have high flood plain to channel width ratios. The flood plain surface has a pronounced stepped morphology which is reworked by surface channels and swirl pits (Fig. 4.9, 2, 3). There are four surficial morphostratigraphic units found in relatively unconfined flood plain formations: channel fills, flood plain veneer, insets (channel and flood plain) and swirl pit fill (Fig. 4.9, 14, 11, 4, 5, 3). As in confined flood plains, the m[illegible]hostratigraphic units may have been post-depositionally eroded and survive as sculpted surface and buried flood plain remnants (Fig 4.9, 6, 7). In relatively unconfined flood plain reaches, back channels and islands can be significant morphostratigraphic units (Fig. 4.9, 1, 12). These surficial morphostratigraphic units generally overlie channel deposits, paleoflood deposits or cemented Pleistocene sediments preserved as buried and sometimes surface flood plain remnants.

4.2.4. Sedimentary characteristics of morphostratigraphic units

This section describes the sedimentary structures, stratification and bedding in sedimentary units of the morphostratigraphic bodies (symbols used in the figures are presented in Table 4.3). Sedimentary units are described as: one or more sets of strata deposited continuously under uniform or continuously varying conditions (Picard and High, 1973).

The sedimentary and morphological characteristics and depositional environments of morphostratigraphic units are summarised in Table 4.4. Sedimentary characteristics, which were not described in detail in the previous sections describing morphology and contextual features, are to a degree diagnostic of processes, and are now described.

4.2.4.1. Channel fill

Channel morphostratigraphic units which include the main channel, flood channels and back channels share common types of sedimentary characteristics including mud beds, sand beds, flood couplets, bar face and bar surface sedimentary units, but at different scales.

Small scale channel sediment characteristics

Small-scale channel fills described morphostratigraphically in section 4.2.2.1. are comprised of bedded to structureless medium and coarse sand or sandy matrix-supported gravels. Sometimes the channel fill may be comprised of one or more *flood couplets* in which laminated sand is conformably overlain by a mud layer; the tops of individual couplets may have desiccation cracks (Plate 4.20; Table 4.4)

Beds of uniform dense *mud* (Fm) represent the deposition of suspended load during flood recession and where the texture is coarser, silt laminations are found (Fl); otherwise the beds appear structureless. In some locations mud layers are rich in organic material (O) (Plate 4.21) and contain large leaf and wood fragments. Packing tends to be loose and the exposed faces of layers readily collapse. In flood couplets the mud beds are underlain by horizontally laminated, planar bedded *sand layers* which may grade into ripple stratification. In places the horizontal bedding is disturbed by bioturbation which may extend into adjacent layers making boundaries indistinct (Plate 4.22).

An excavation at the Anabranching site shows an example of a small scale braided channel fill (Fig. 4.35, Plate 4.12), The channel is 63 cm deep and forms part of a larger drape component which infills the adjacent surface channels and aggrades the wider flood plain surface. Basal sediments dip across the channel section at 18° and overlie a truncated, horizontal mud bed (not shown in Fig. 4.35). The basal sediments are an upward fining, inclined parallel stratified flood couplet sequence composed of medium and coarse red sand (Si) and mud (Fm). This is overlain by a second flood couplet composed of very fine red sand topped by a 4 cm layer of mud. A third flood couplet of organic-rich, fine red sand is also overlain by a 2 cm layer of silty mud. The remaining sub-horizontal fill is composed of medium to fine white and red sand which grades into fine red sand. These sediments have well preserved parallel laminae. The three flood channels were synchronously eroded into the surface of the flood plain and are inferred to have aggraded during four flood episodes. A second example is the surface channel at the Todd/Ross Left Bank site (Fig. 4.36), a shallow fining upwards channel fill where contact with the underlying clay-rich bed (Fm) is erosional and the basal matrix-supported gravels are internally graded and topped by alternating thinly laminated sand and silts ( ).

Plate 4.19. Complex morphostratigraphy of flood plain at Todd Ross Confluence site, including erosional unconformities and step morphology (a), truncated flood plain veneer (b), inset deposition (c) at the *microscale*.

Plate 4.20. Ross River Striped site. Note the thick mud beds, thin mud drapes, flood couplets and desiccation crack (a).

Plate 4.21. Channel bench unconformably overlying bar face deposit. Note the organic rich layers and the colour difference between the grey bedload and red and brown suspended load sediments.

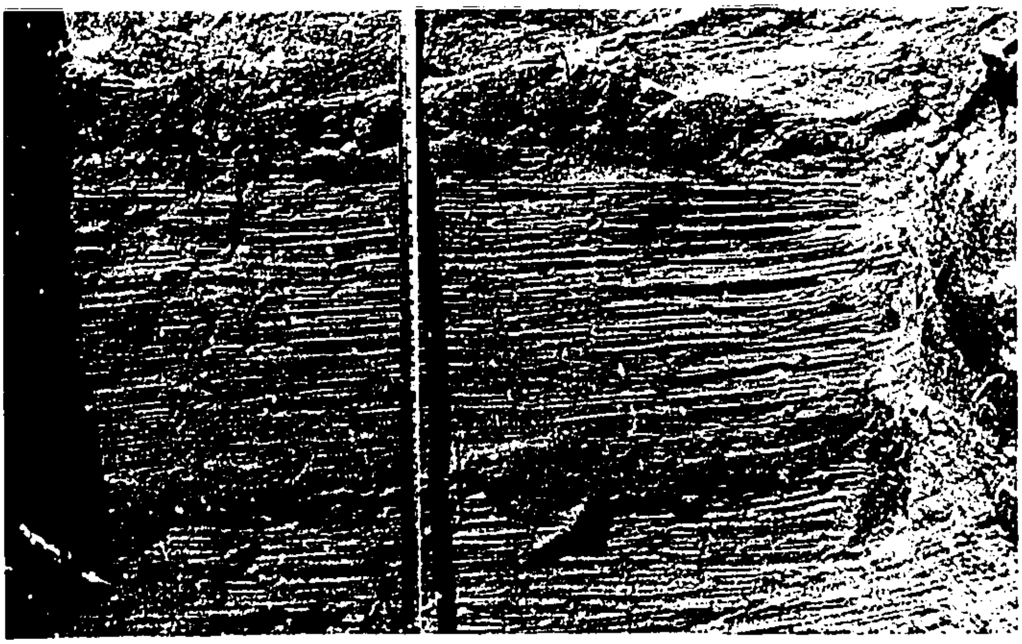


Plate 4.22. Conformable mud deposition overlying planar bedding (Sh) in fine red sand sheet and burrow into underlying strata.

Large scale channel sediment characteristics

Large-scale channel fills described morphostratigraphically in section 4.2.2.1 are comprised of bedded to structureless medium and coarse sand or sandy matrix-supported gravels in bar face and bar surface facies, sometimes with thin mud layers (Table 4.4). Migrating *avalanche faces* of longitudinal bars (Sh, St, Sp, Gm) have steeply dipping (15°) ~50 cm thick planar tabular sets comprised of alternating beds of poorly sorted pebbles and coarse sand and well rounded mud clasts (*e.g.* Fig. 4.37, Plate 4.21 and 4.23). While these sedimentary units generally occur towards the bottom of flood plain profiles they also occur at higher levels. *Bar surface* deposits (Gm, Sh) are generally structureless or poorly laminated coarse sand or gravel in a coarse sand matrix in 20-80 cm thick sedimentary units (*e.g.* Fig. 4.37). Similar to the bar face deposits they may occur in any location in the profile (Plate 4.23).

The large channel fill at the Anastomosing site (Fig 4.35) is a fining upward sequence from structureless medium gravels (Gm) to horizontally laminated white sands (Sh) overlain by a structureless layer of fine gravel in silty sand (Sm), and has an erosional boundary with the overlying inclined (Si) stratified medium and coarse sand layer. The long section through the river channel bed on the Ross River upstream of the confluence with the Todd River, shows a continuous lenticular stratification with only minor erosion between one of the sets towards the base. Two bar facies (Sp) are topped with mud drapes and three gravel sheet facies (Gm). The sequence is topped by a set (20 cm) of climbing ripples. Another section in the channel bed in Alice Springs town included a roughly horizontally layered, matrix-supported gravel fill with lenses of pebbly sand, plastic and aluminium.

Many surface channels contain hummocks and Figure 4.38 is a representative hummock profile mapped in a distributary flood channel, 170 cm high and superimposed on flood deposits. The basal 140 cm of structureless sand is overlain by 32 cm of climbing ripples composed of medium and fine sand, in turn overlain by a layer of inclined sand laminae overlain by a mud drape.

4.2.4.2. Channel insets

Channel insets generally comprise channel bar surface and bar face sedimentary units but may also include horizontal mud and sand units. A channel inset at the Todd/Ross right bank site (Fig. 4.10; 4.37) is composed of a sequence of parallel inclined gravelly coarse sand and mud balls topped with a planar horizontal sandy gravel. Sedimentary unit thickness varies between 2 cm and 80 cm and no erosional boundaries were detected in this profile.

4.2.4.3. Flood plain insets

Flood plain insets (Gm, Sp, Sh, Sr, Sw, Sl, O, Fl, Fm) are comprised of mud, sand, flood couplet, bar surface and bar face sedimentary units (Plate 4.23). The sedimentary characteristics of a flood plain inset at Mosquito Bore 1 site described in section 4.2.9. and will not be repeated here.

4.2.4.4. Flood plain veneer

Flood plain veneers are comprised of bedded to structureless medium and coarse sand, mud and flood couplet sedimentary units (Table 4.4). Strata inclination varies according to the morphology of the underlying sediments and is horizontal-parallel on the higher and lower surfaces and inclined-parallel at the break in slope between. Veneer sedimentation occurs over a series of depositional events which often result in cyclical flood couplet aggradation. Figure 4.39 was mapped from the downstream wall of a pit in a flood plain veneer unit at the Todd/Ross Left Bank site (excavation E). Four sedimentary units composed of finely laminated, inclined, sand and mud strata extend continuously from the upper to lower step.

4.2.4.5. Swirl pit fill

Swirl pit fills, described morphostratigraphically in section 4.2.4.4., are comprised of bedded to structureless mud, medium and coarse sand or sandy matrix-supported gravels in mud, sand and flood couplet sedimentary units (Table 4.4).

A swirl pit morphostratigraphic unit formed in association with a Coolabah tree (Fig. 4.32) was mapped and although the central portion was obscured by debris from the tree, this site contains rare evidence of sedimentary fill in a

swirl pit that developed around a tree. It includes three stratigraphic units: planar bedded white channel sands grade upwards through a sequence of pebbly sand beds to well-sorted medium and fine sand and mud. The truncation of sedimentary layers indicates erosion by succeeding events and layer thickness varies between 2? and 40 cm.

4.2.5. Representative sedimentary stacks

A fresh exposure was eroded during the January 1995 flood at the Mosquito Bore 1 confined flood plain site (Plate 4.24, Fig. 4.4, Fig. 4.40). The basal sedimentary unit (96 cm) in this 267 cm high profile is composed of three alternating red and white sand beds which display cross stratified, parallel discontinuous and structureless stratification. This is overlain by two flood couplets. The next layer is rich in ash and contains mud clasts towards the top. This is overlain by a mud layer and a silt-rich layer deposited during separate events. The next flood couplet sequence is overlain by a thick flood deposit (54 cm) which alternates twice from climbing ripples to horizontal stratification and is overtopped by a structureless sand unit with angular mud fragments. This is overtopped by four flood units. The lowst one is a fine textured flood couplet with good horizontal stratification in the lower silty unit. The next event deposited sand with good cross and horizontal stratification. The sequence is finished with two flood couplets, the upper one deposited during the January 1995 event (Plate 4.24).

A brief synthesis of the sedimentary characteristics at the Anabranching site (Fig. 4.35, Plate 4.12) illustrates the typical sedimentary units in a relatively unconfined flood plain site. The 280 cm deep section unconformably overlies cemented Pleistocene sediment. There are three morphostratigraphic components to this profile. The basal, and largest, unit is a 217 cm thick channel fill unit which underlies the 35 m wide island. This unit is composed of a structureless, poorly sorted, gravel/coarse sand unit which fines upward to a horizontally stratified coarse sand set. This is overlain by a silty/coarse sand and pebble set, capped by a thin silt drape, and overtopped by inclined parallel stratified medium to fine white and red sand. This sequence has truncated surfaces typical of migrating bar facies. The sediments in the bar facies dip at approximately 6° and the truncated upstream face also has an angle of 6°. These channel bed sediments are overtopped by the flood plain unit which is 63 cm thick and composed of two morphostratigraphic units. The lower unit is a sequence of horizontally bedded layers of alternating sand and mud (Sh, Fl, Fm). This unit was incised by surface braided channels and filled with alternating layers of sand and mud, described previously.

4.2.6. Vertical profile model

Figure 4.41 is a schematic vertical profile which summarises the main sedimentary facies components of the flood plains mapped in the Todd catchment. The characteristics of confined and relatively unconfined flood plain profiles are sufficiently similar that a single profile model is proposed.

The flood plains are composed of a series of vertically accreting layers emplaced during floods. The boundary between layers may be erosional as indicated by fill structures, rip-up clasts and truncated stratification. Textural composition varies from coarse gravel layers through to mud layers. There is no clear grading in the profile and mud layers alternate with sand and gravel layers in any order vertically through the profile. Flood couplets are common and mud balls are preserved in some of the sandy deposits. Bioturbation is common and disrupts layer boundaries and the internal stratification of the sets. Thin laminations are prominent as are horizontal stratifications. However parallel inclined stratification is often found in fill structures and veneer deposits. There are some buried soil layers and the roots from vegetation are often found in the sandy layers.

4.2.6.1. Sedimentary characteristics

Six *sediment size* samples were analysed in the laboratory from Mosquito Bore 1 site 100 m upstream of the profile in Figure 4.40. Samples were taken from sedimentary units. The mean size of sediment for the flood plain falls in the sand size category and the mean size of the sediment in each layer varies vertically, with an absence of an upward fining sequence. When examined in more detail (Fig. 4.42, a-e), the percentage silt is highest towards the base of the flood plain and the gravel content is higher towards the top of the flood plain. All samples are moderately sorted tending towards less well sorted towards the top of the profile (Fig. 4.42, e).

Seven samples were analysed at the Stud Bore 1 flood plain site (Fig. 4.12 excavation D). This flood plain is coarser than the previous site with the mean size of sediment falling in the medium sand category (Fig. 4.43). As in the previous example, there is no vertical trend in the size of sediment. Silt content is highest in the middle-to-top of the flood plain profile and gravel

Plate 4.23. This flood plain inset profile is located 275 m from the main channel and shows bar surface and bar face deposits overlying planar bedded fine red sand. Note the textural variation in the profile.

Plate 4.24. Mosquito Bore 1 site. Photo taken two days after January 1995 flood. A thin layer of fine red sand and the tree were freshly deposited on the flood plain surface. Flow was approximately 80 cm above this surface.

content increases away from the base. Samples are moderately and poorly sorted with a tendency for better sorting towards the top of the profile (Fig. 4.43, e).

There appears to be great variation in sediment size vertically in Todd River flood plain profiles with no linear trend apparent. Fine textured mud beds (Fm, Fl) may be in any position in the profile and although layers of gravel and coarse sand are generally positioned towards the base of profiles they can also be located higher. The flood plains typically exhibit abrupt textural change between sedimentary units.

Bioturbation of flood plain sediments is common. The most effective agents are ant colonies which have implications for the reliability of luminescence and radiocarbon dating in this region. Larger burrows formed by the witchetty grub disturb the boundaries between sedimentary units (e.g. Plate 4.22). Kingfishers also excavate their nests from the river banks (Plate 4.12) and the penetration of plant root structures disrupts the flood plain sediments.

Although most sets exhibit internal *grading*, there are examples of reverse grading, most commonly observed in thickly bedded, loosely packed, structureless sand and fine gravel.

The *colour* of sediments varies between sedimentary units and between strata. In the latter the finer textured laminae in Sh sedimentary facies are often red (10yr5/8) and the coarser laminae are very pale brown (10yr 7/4). At the larger scale the muddy organic-rich sediments tend to be yellowish brown (10yr 5/6), the bedload sediments pale brown and the finer suspended sediments red. This was observed after the January 1995 event where the sediments in the main channel were relatively clean and appeared whitish, the suspended sediments were pale red (Plate 4.10). This pattern was consistently observed throughout the catchment and may be explained by the differential rate of removal of the haematite coating of entrained Pleistocene sediments by differential abrasion rates of grains in the bedload and in the suspended load.

The *thickness* of sedimentary layers varies in the vertical profiles with no strong decrease with height in the profile: very thick layers (>30 cm) were mapped high in profiles. Similarly there does not appear to be a relationship between layer thickness and sediment size. Although the thinnest sets are mud drapes and there are some thinly laminated sandy beds the thickest layers are composed of gravel and coarse sand

4.3. Chronology of Flood Plains

The absolute age of the flood plains was determined by radiocarbon analysis (Tables 4.5 and 4.6) and radionuclide concentrations (Tables 4.7 and 4.8). Relative age was assessed by the preservation of sedimentary structures and the age of alive but buried riverine vegetation.

In the confined flood plains there are nine radiocarbon ages, from six sites and four ^{137}Caesium and ^{210}Lead determinations from two of those sites. Seven radiocarbon samples are younger than modern. In the relatively unconfined flood plains there are seven radiocarbon ages from three sites and five radionuclide determinations from two sites. As for the confined flood plains, the unconfined flood plains are young with five of the samples returning ages greater than modern (Table 4.6). The following is a description of each of the sample sites.

4.3.1. Confined flood plain chronology

Todd/Ross Left Bank

The age of alluvium was determined for three flood plain insets on the left bank. All samples were taken from depths in excess of 0.5 m and returned ages younger than modern (Figure 4.11). Samples from these flood plains were also examined for ^{137}Caesium and ^{210}Lead (Fig. 4.11, Table 4.7). Two samples indicate sufficiently high concentrations of both radionuclides and indicate an age greater than 1960. No significant levels were detected for the deeper sample.

Mosquito Bore 1 site

Two radiocarbon ages were determined for this site. Sample ANU 9276 was determined from a charcoal sample taken from 140 cm below the surface and returned an age of 320-0 cal BPand ANU 9280 from a charcoal sample taken from 79 cm below the surface (Fig. 4.27) and was younger than modern. No significant radionuclide concentrations were detected. The sediment in

excavations of all flood plains appeared fresh and unweathered, consistent with the relatively young age determinations.

No. 5 Bore site

A grab sample was taken from a backwater deposit in a left bank tributary mouth (Fig. 4.6). The sample age (ANU 9656, 753-421 cal BP) does not match the appearance of the sediments, which had well-preserved sedimentary structures and little bioturbation indicating recent deposition. The sample probably consisted of reworked charcoal.

Site 'A'

A radiocarbon sample was taken from the base of a swirl pit fill on surface Rf2 (Fig. 4.14, ANU 9652). The amount of charcoal was too small to be reliably dated but estimations indicate an age younger than modern. A second grab sample (ANU 9654) was taken from a left bank flood plain further downstream, at the top of a section of vertically stacked flood couplets, and 279-51 cal BP.

Desert Reach

The oldest age obtained for a confined flood plain deposit is 5936-5561 cal BP, (ANU 9659), from charcoal collected 180 cm below the surface (Fig. 4.13)from the site furthest downstream (Figure 4.7). This significantly older flood plain age may reflect reworking but it may also be a function of the reduced capacity to rework alluvium with distance downstream. Given the tendency for ephemeral floods to decrease in magnitude downstream and the lower frequency of events in these reaches, such flood plains are expected to be older in age.

4.3.2. Relatively unconfined flood plains

Ross River Striped site

Four samples were analysed by radiocarbon methods from this site. Three of these returned ages younger than modern (Fig. 4.19, ANU 8963, ANU 8970 and ANU 8968). The remaining sample age (ANU 8962) is 552-469 cal BP and may be erroneous as a sample from the flood plain across the channel, stratigraphically older, returned a younger than modern age. In addition the nature of the sediments from which the sample was taken is indistinguishable from the younger, dated, sediments in the profile. Four samples from this site were analysed for ^{137}Caesium and ^{210}Lead content (Table 4.8). No significant concentrations were detected. This is interpreted as just older than 1960 or due to the rapid aggradation of sediments.

The freshness of the sediment in addition to the burial of trees (estimated to be less than 100 years old) by 1 m of sediment indicates the young age of the flood plains in this reach (Plate 4.11).

Stud Bore 1

One sample was measured for ^{137}Caesium and ^{210}Lead content (Fig. 4.12). No significant levels of either radionuclide were found. However the surface of these flood plains is populated by mature River Red Gums which are buried by .5 to 1 m of sediment indicating a young age (<150 BP).

Anabranching site

Two radiocarbon ages were determined for flood plain deposits. One sample was from a surface channel fill unit (Fig. 4.18, ANU 9653) and its young age is verified by the burial of young Red Gums (approximately 10-20 yrs old) on the flood plain surface by 1 m of sediment. The second sample (ANU 9676) is from the upper section of a sequence of channel sand (Fig. 4.18) and is older (479-277 cal BP).

Jessie Gap fan

A sample was taken from the Jessie Gap fan (ANU 8971) at a depth of 48 cm. This sample returned an age of younger than modern.

Figure 4.1. Location of Ross River Striped Site and Shannon Bore Site. North is to the top of the figure and one square = 1 km.

Figure 4.2. Location of Todd/Ross Confluence site, Todd/Ross Left Bank site, Todd/Ross Right Bar site, Stud Bore 2 site and Stud Bore 1 site.

North is to the right of the figure and one square = 1 km.

Figure 4.3. Location of Giles Creek 1 site, Giles Creek 2 site, Giles Creek 3 site. North is to the top of the figure and one square = 1 km.

Figure 4.4. Location of Mosquito Bore 1 site and Mosquito Bore 2 site. North is to the top of the figure and one square = 1 km.

Figure 4.5. Location of the Expansion Scour site.
North is to th· top of the figure and one square ≡ 1 km.

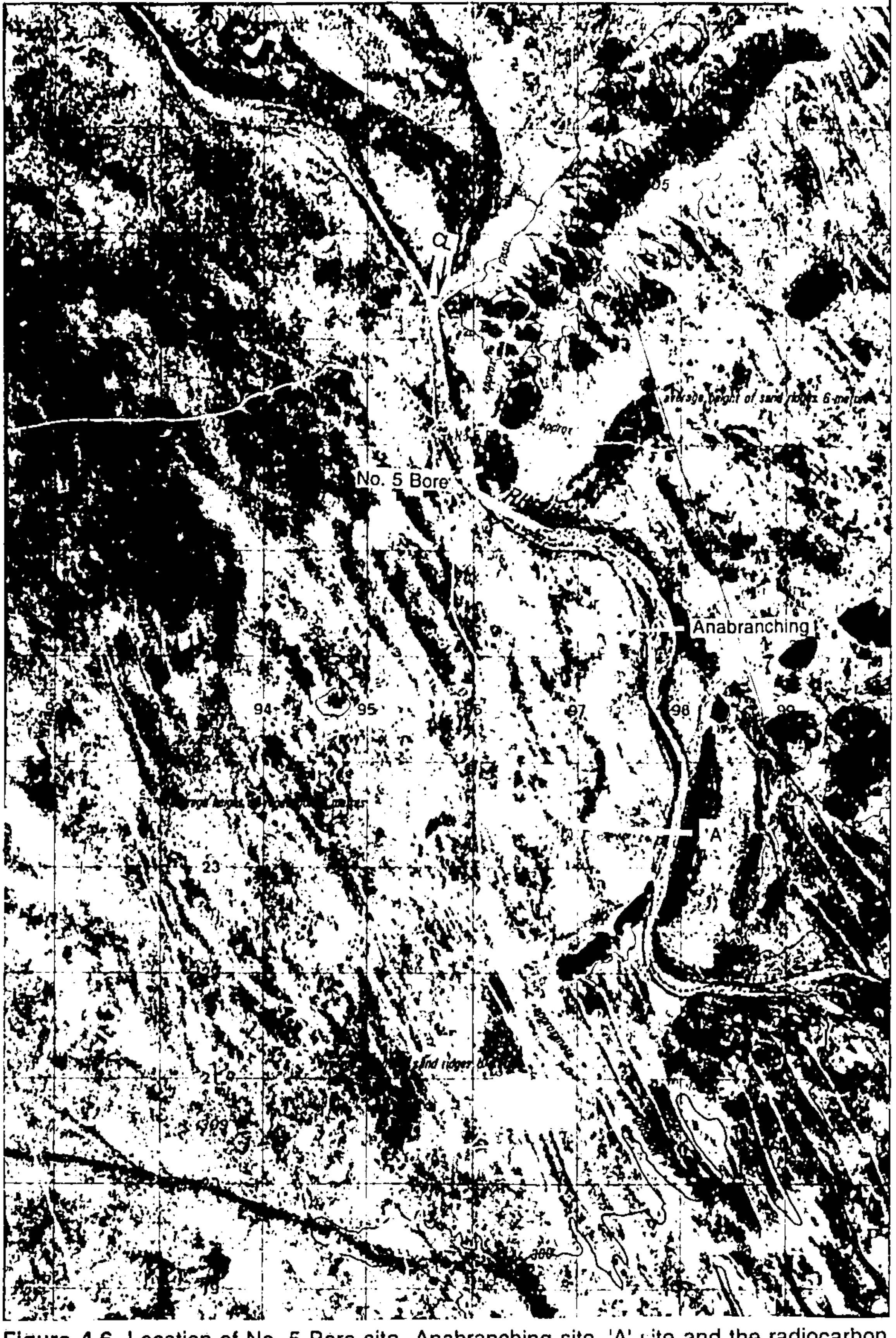


Figure 4.6. Location of No. 5 Bore site, Anabranching site, 'A' site and the radiocarbon grab sample CL. North is to the top of the figure and one square = 1 km.

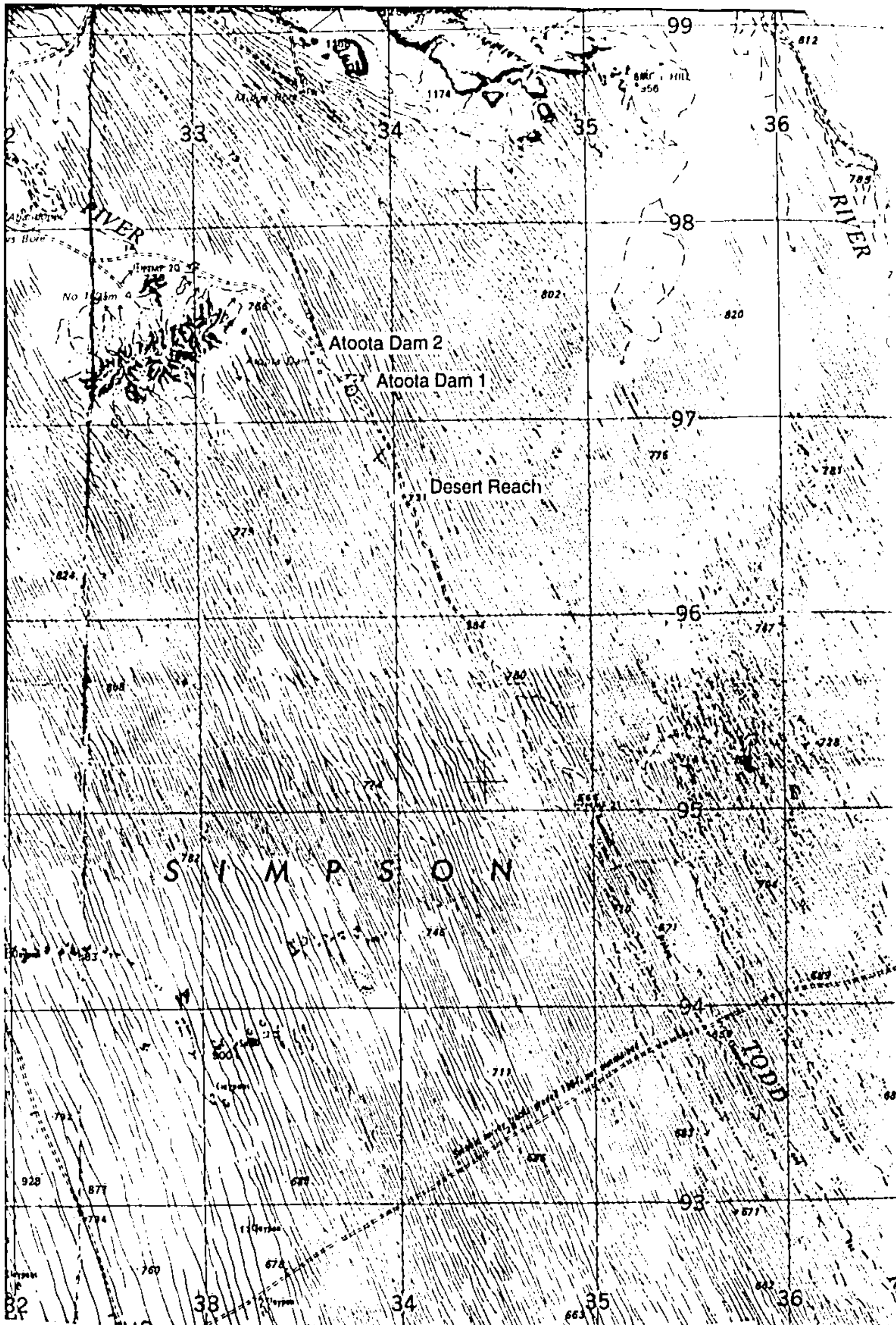


Figure 4.7. Location of Atoota 1 site, Atoota 2 site and Desert Reach site. North is to the top of the figure and one square = 9.2 km.

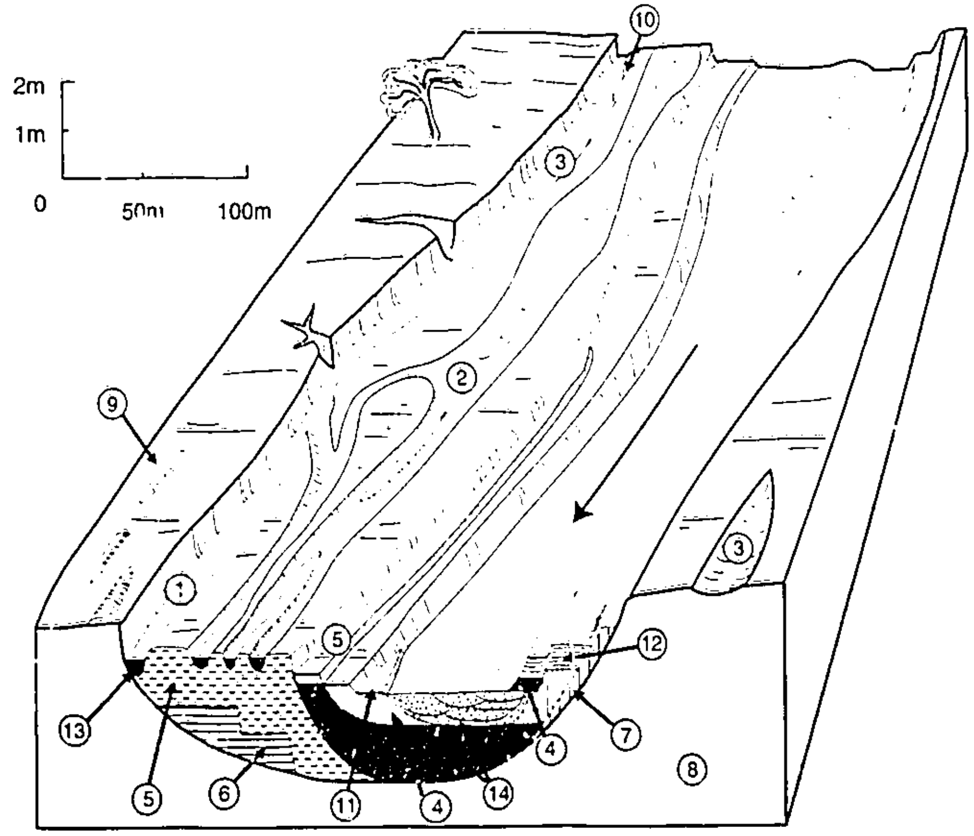


Figure 4.8. Morphostratigraphy of confined flood plains in the Todd River.
1. Back channel. 2. Flood channel. 3. Swirl pit. 4. Channel inset. 5. Flood plain inset. 6. Buried flood plain remnant. 7. Surface flood plain remnant. 8. Cemented Pleistocene or paleoflood terrace. 9. Bar. 10. Back channel bench. 11. Stripped surface. 12. Flood plain veneer. 13. Surface channel fill. 14. Large channel fill.

Morphostratigraphic cross sections (locations in Figs. 4.1-4.7). Flood plain and terrace elements are labelled on bar at top of figure and referred to in text prefaced by R or L indicating right or left bank. Legend is given in Figure 4.44 on A3 pullout. Capital letters o vertical bars indicate excavation.

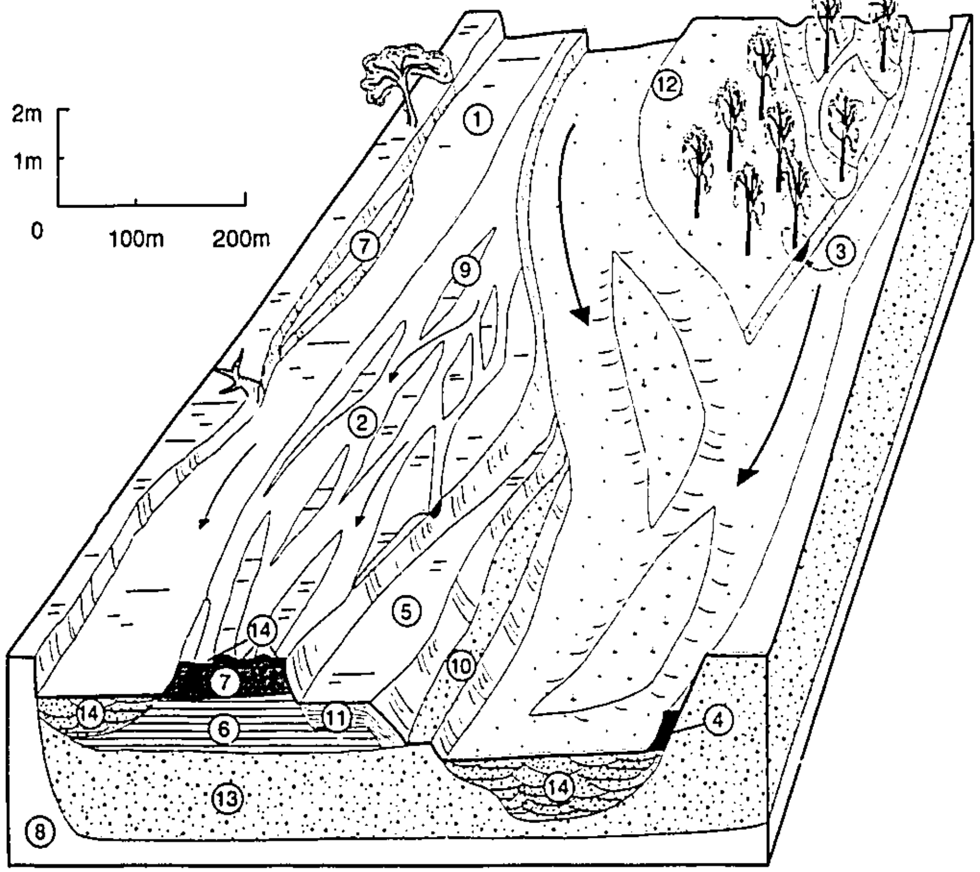


Figure 4.9. Morphostratigraphy of relatively unconfined flood plains in the Todd River.

1. Back channel. 2. Flood channel. 3. Swirlpit. 4. Channel inset. 5. Flood plain inset. 6. Buried flood plain remnant of bank attached island. 7. Surface flood plain remnant. 8. Pleistocene alluvium. 9. Overbank bars. 10. Stripped surface. 11. Flood plain veneer. 12. Island. 13. Paleoflood alluvium. 14. Chaññel fill.

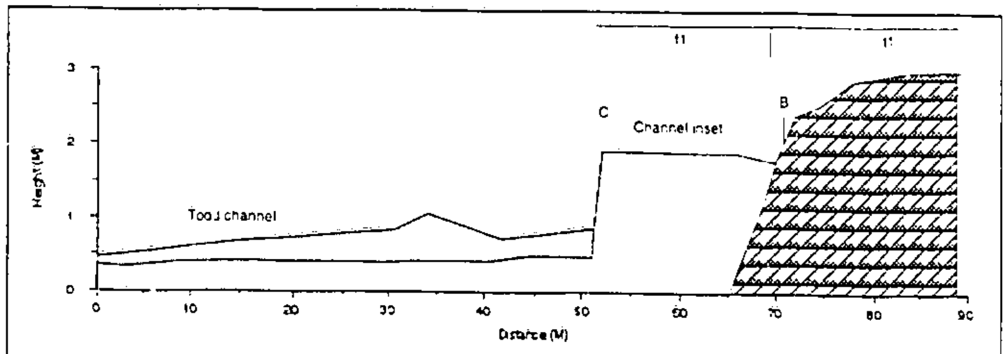


Figure 4.10. The Todd/Ross Right Bank site. For location see Figure 4.2.

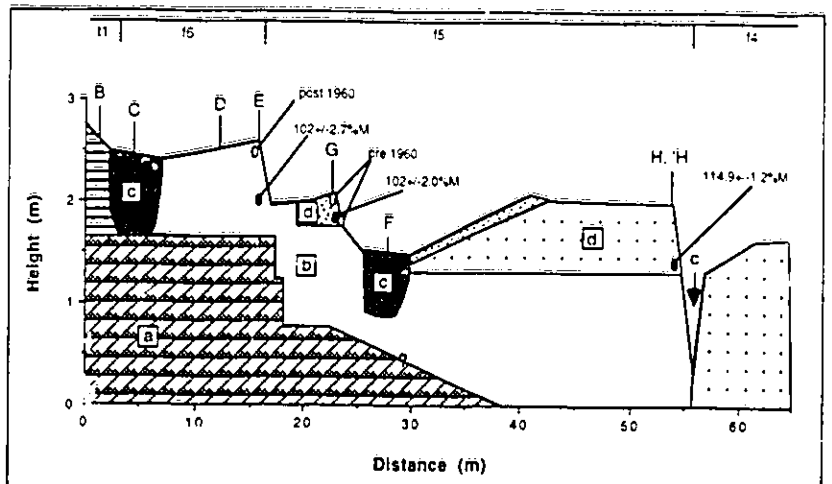


Figure 4.11. Todd/Ross Left Bank site. a: flood plain remnant; b: flood plain veneer, c: back channel, d: flood plain inset. For location see Figure 4.2.

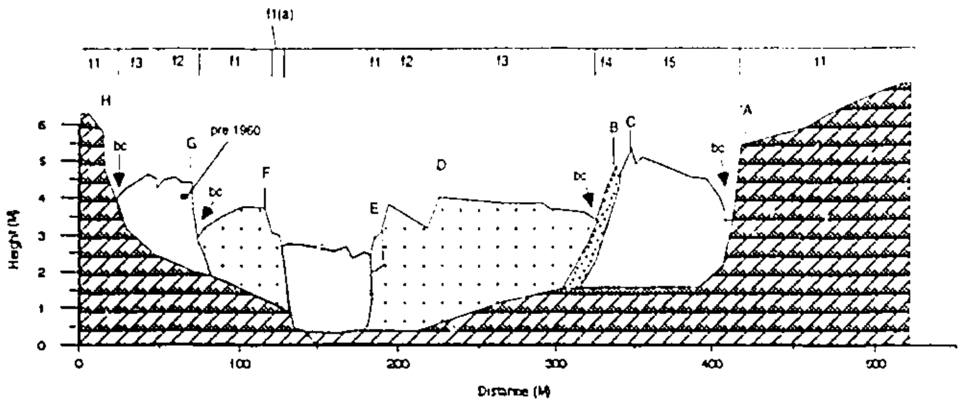


Figure 4.12. The Stud Bore 1 site. For location see Figure 4.2. bc, back channel.

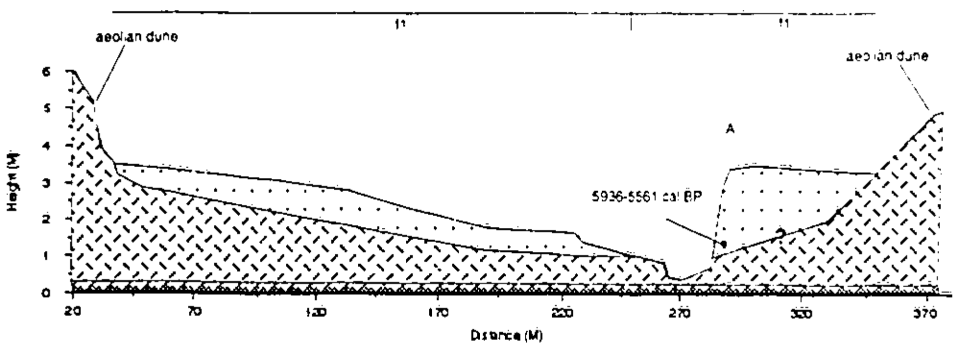


Figure 4.13. The Desert Reach site. For location see Figure 4.7.

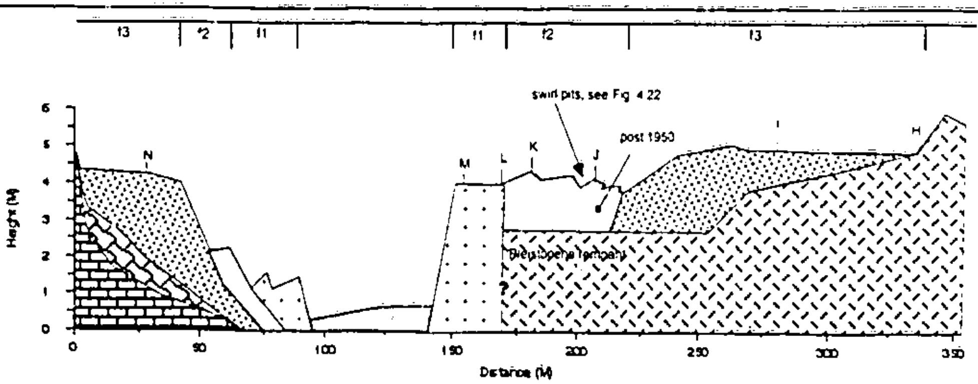


Figure 4.14. Site 'A'. For location see Figure 4.6

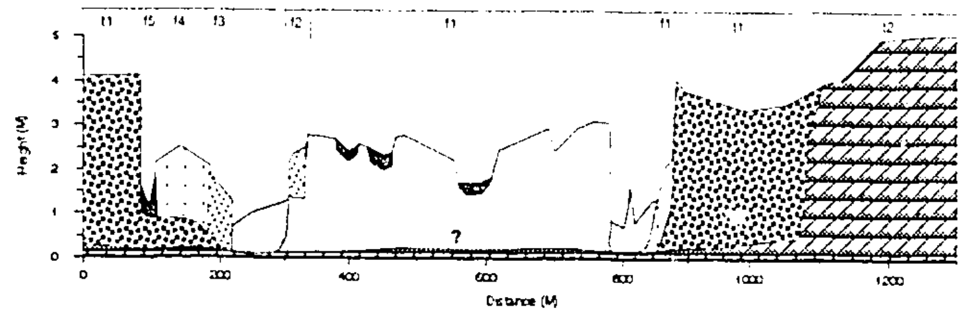


Figure 4.15. The Stud Bore 2 site. For location see Figure 4.2.

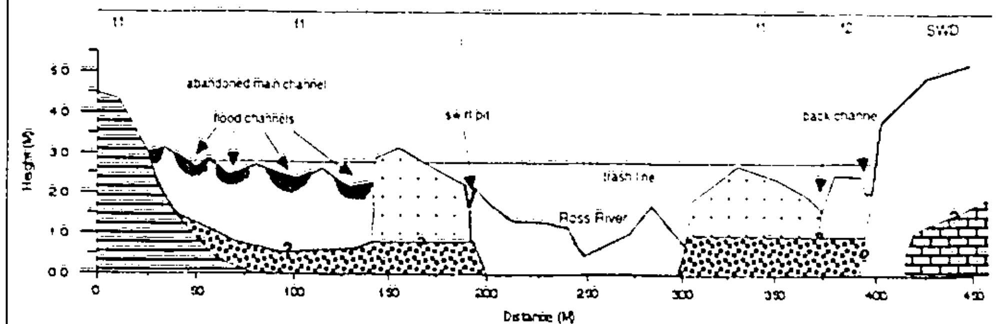


Figure 4.16. The Shannon Bore site. For location see Figure 4.1

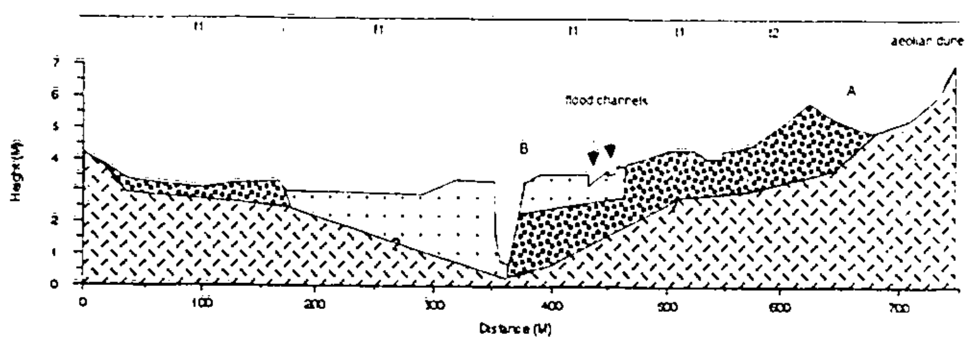


Figure 4.17. The Atoota Dam 1 site. For location see Figure 4.7

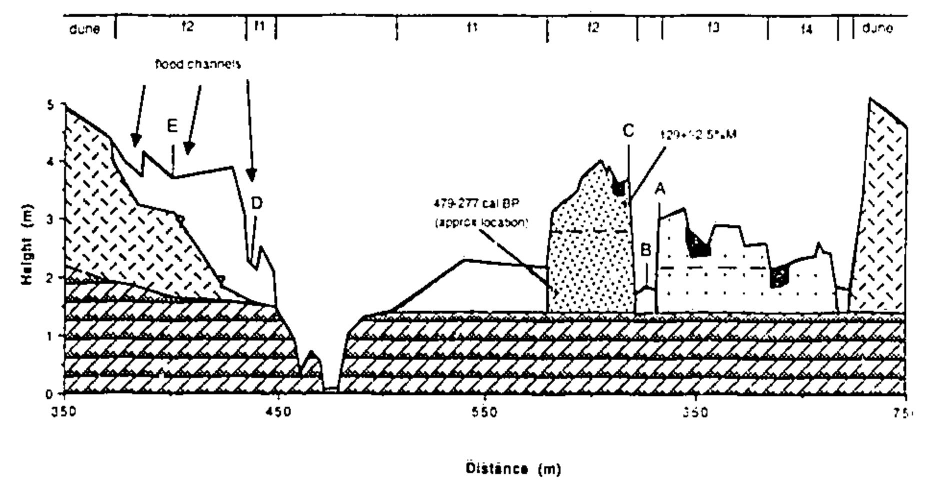


Figure 4.18. The Anabranching site. For location see Figure 4.6. Dashed line in right bank flood plain indicates top of channel fill.

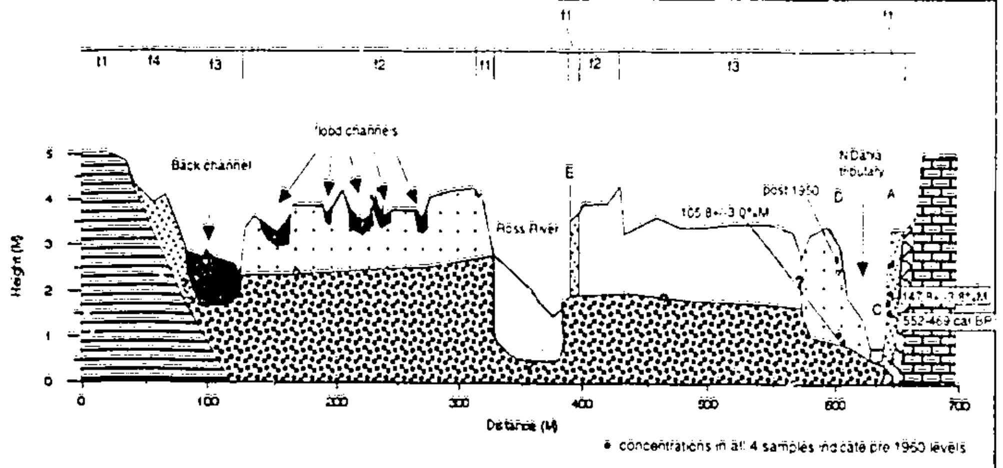


Figure 4.19. The Ross River Striped Site. For locatic i see Figure 4.1.

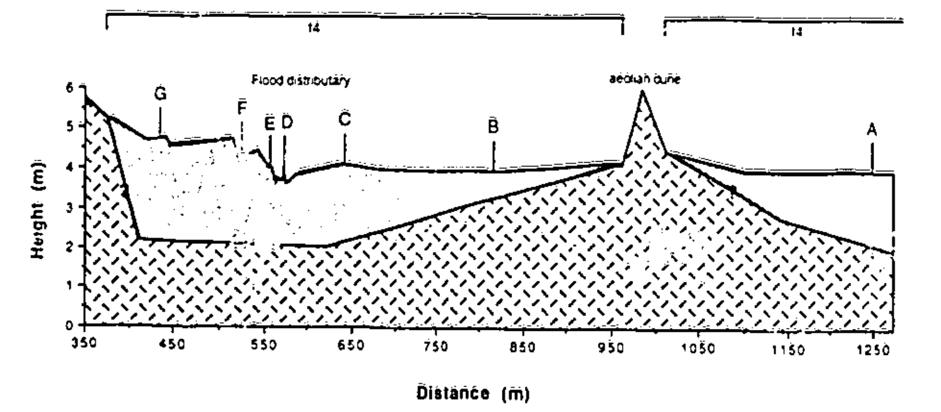


Figure 4.20. Site 'A' right bank flooJ distributary. For location see Figure 4.6.

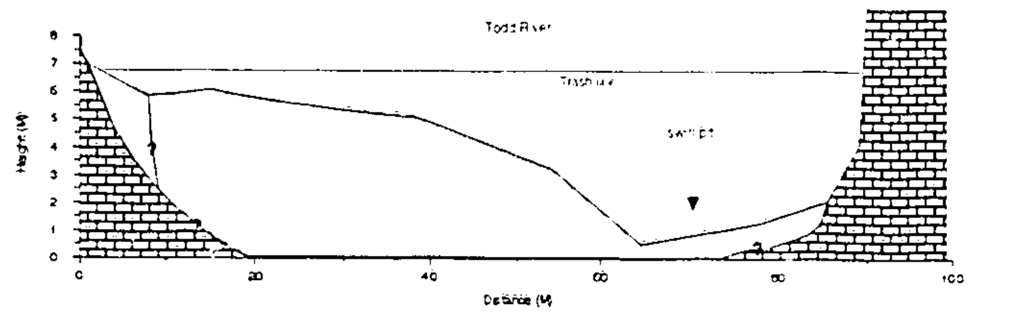


Figure 4.21. The No. 5 Bore site. For location see Fidure 4.6.

Figure 4.22. is with Plate 4.?

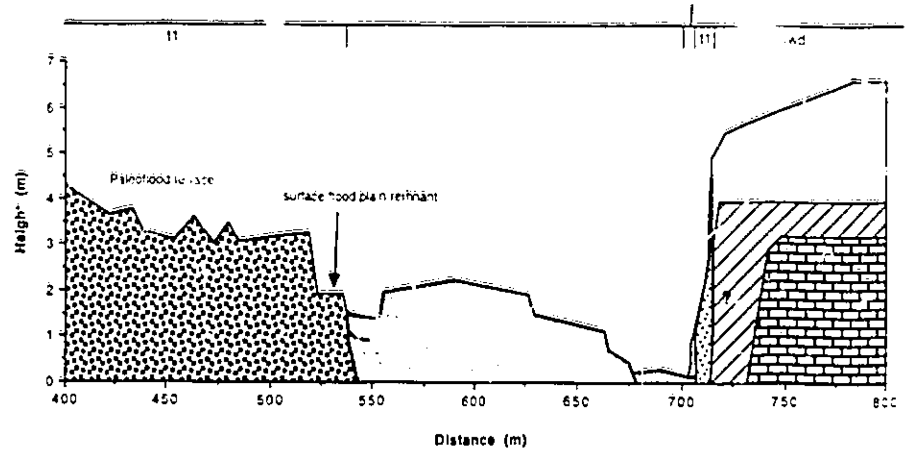


Figure 4.23. The Mosquito Bor· ? site. For location see Figure 4.4.

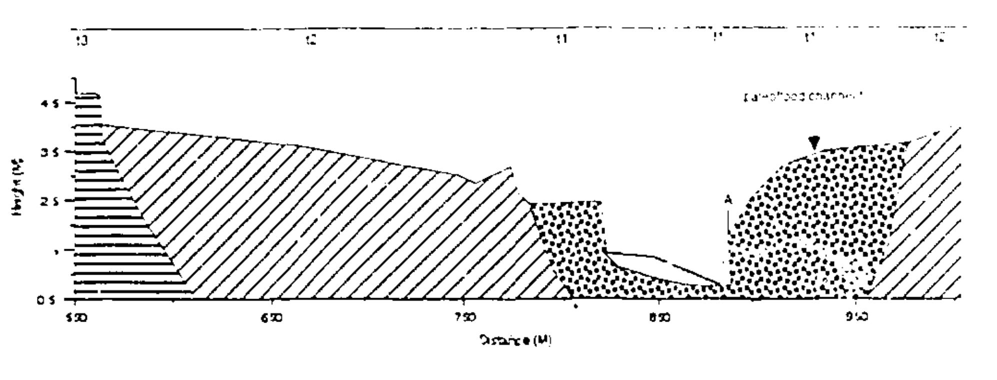


Figure 4.24 The Giles Creek 1 site. For location see Figure 4.3.

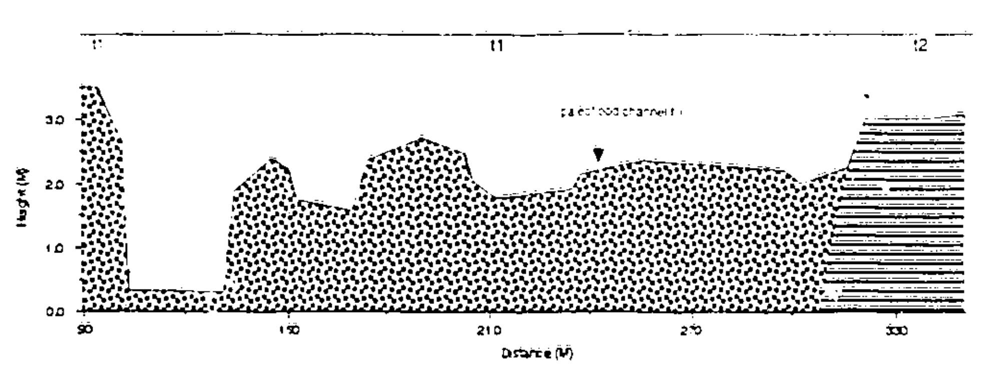


Figure 4.25. The Giles Creek 2 site. For location see Figure 4.3.

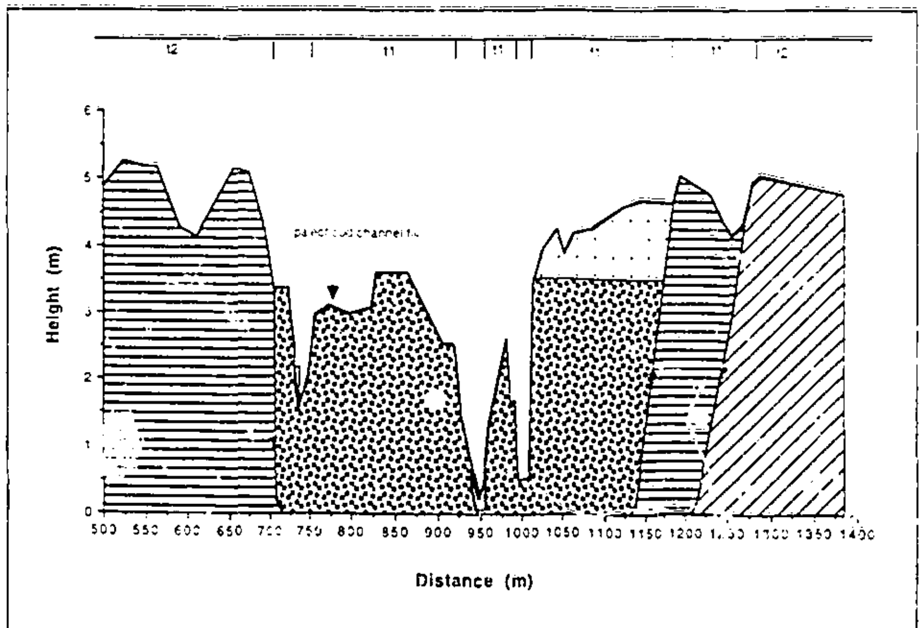


Figure 4.26. The Giles Creek 3 site. For location see Figure 4.3.

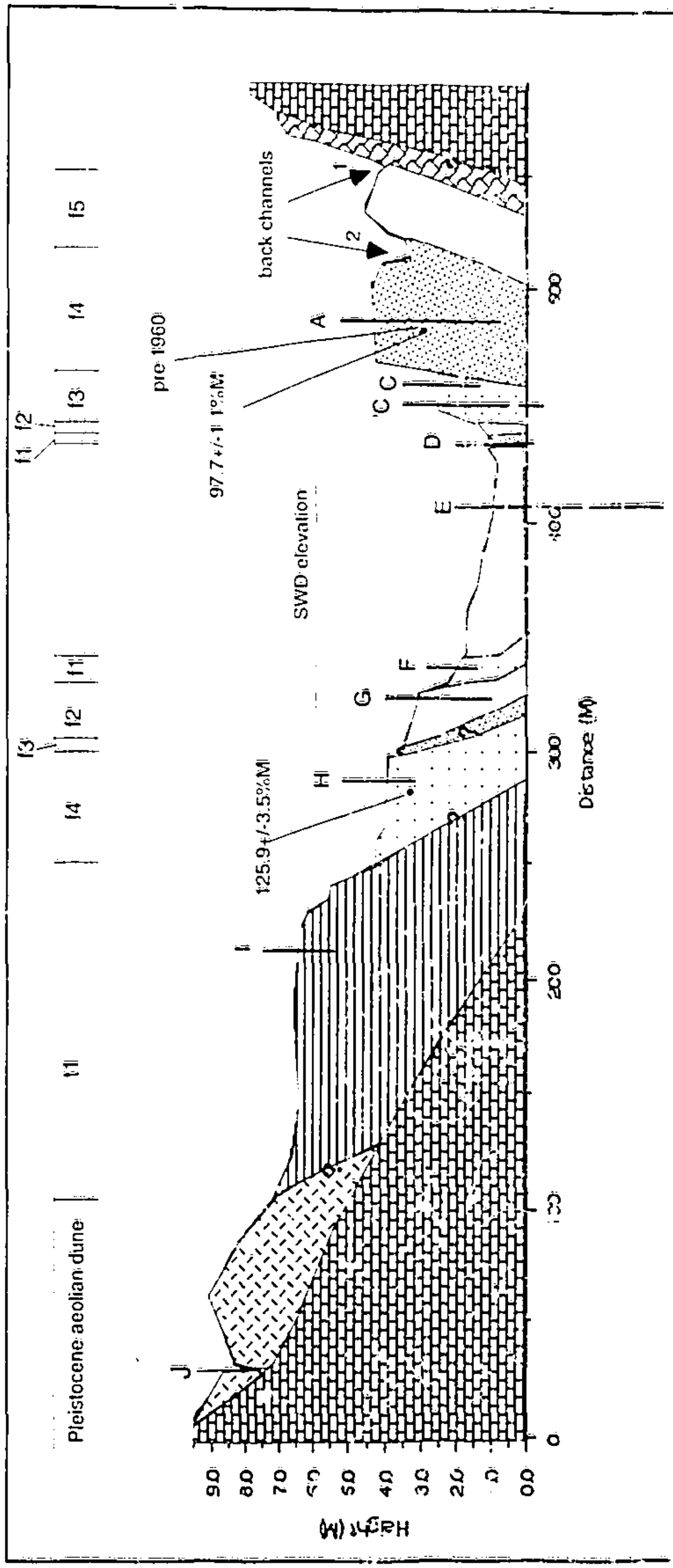


Figure 4.27. The Mosquito Bore 1 site. Type site of confined flood plain morphology. Depth of excavation is shown by length of line. For location see Fig. 4.4

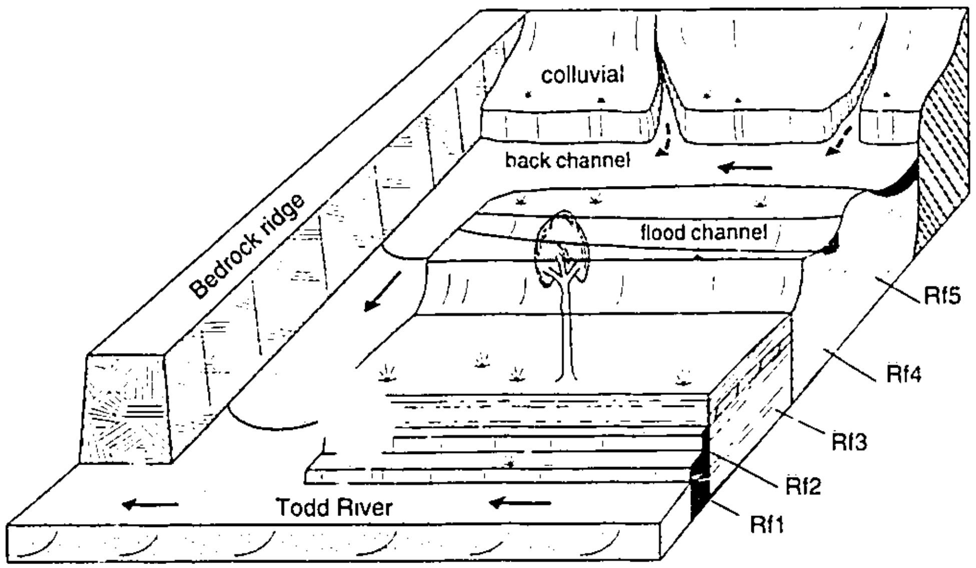


Figure 4.28. Schematic right bank flood plain, Mosquito Bore 1 site.

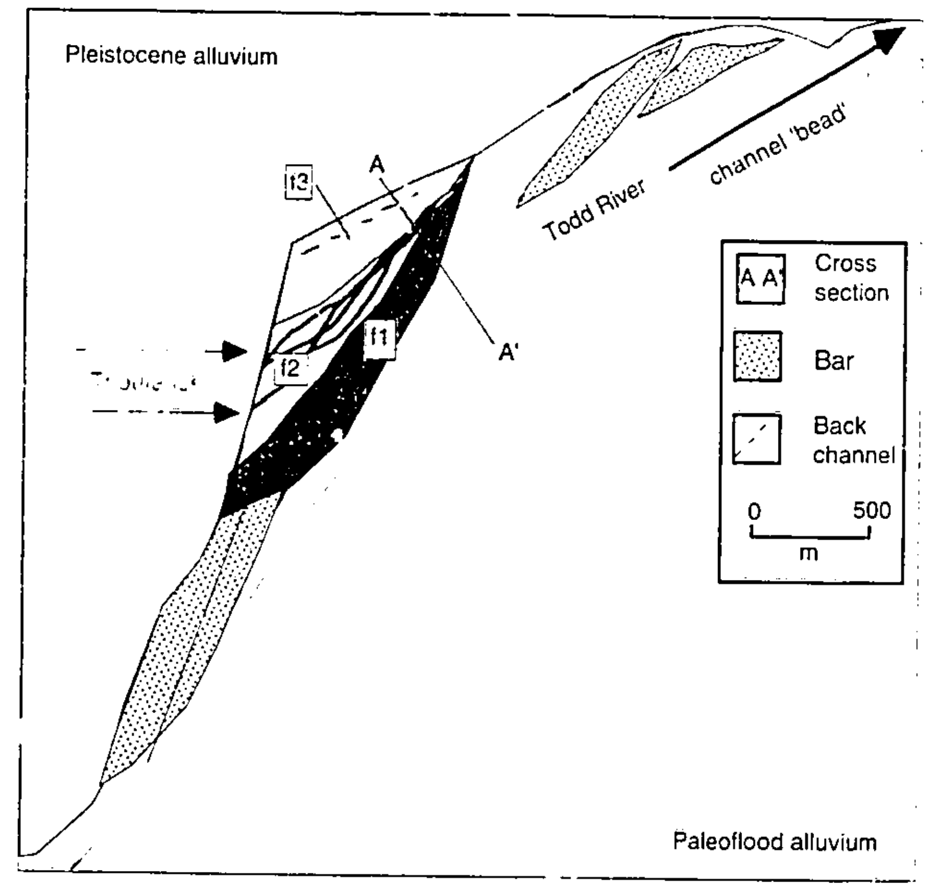


Figure 4.29. Schematic planform of Expansion Scour site.

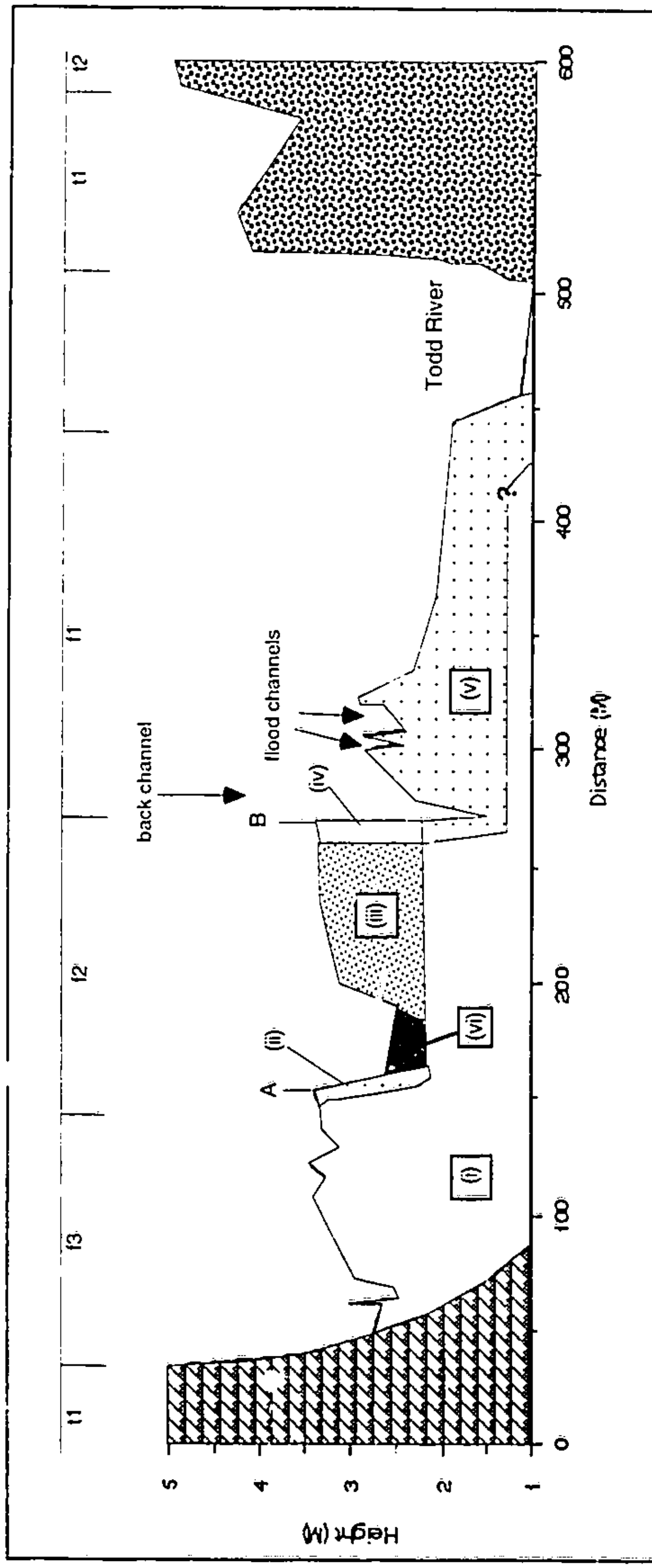


Figure 4.30. The Expansion Scour site. Type site of relatively unconfined flood plain morphostratigraphy. For location see Figure 4.5. (i) to (vi) are explained in text.

Figure 4.31. is with Plate 4.16.

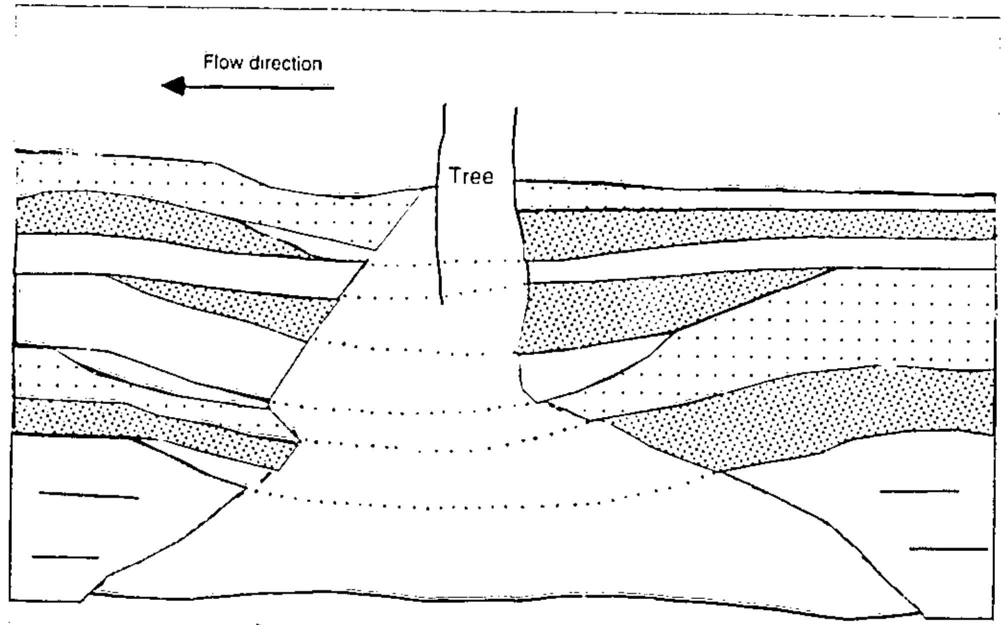


Figure 4.32. Morphostratigraphy of swirl pit fill around tree.

Central area is the tree and root exposure and associated debris. Dotted line is the inferred line between sedimentary layers upstream and downstream of tree.

Pre-existing channel fill planar bedded sands.

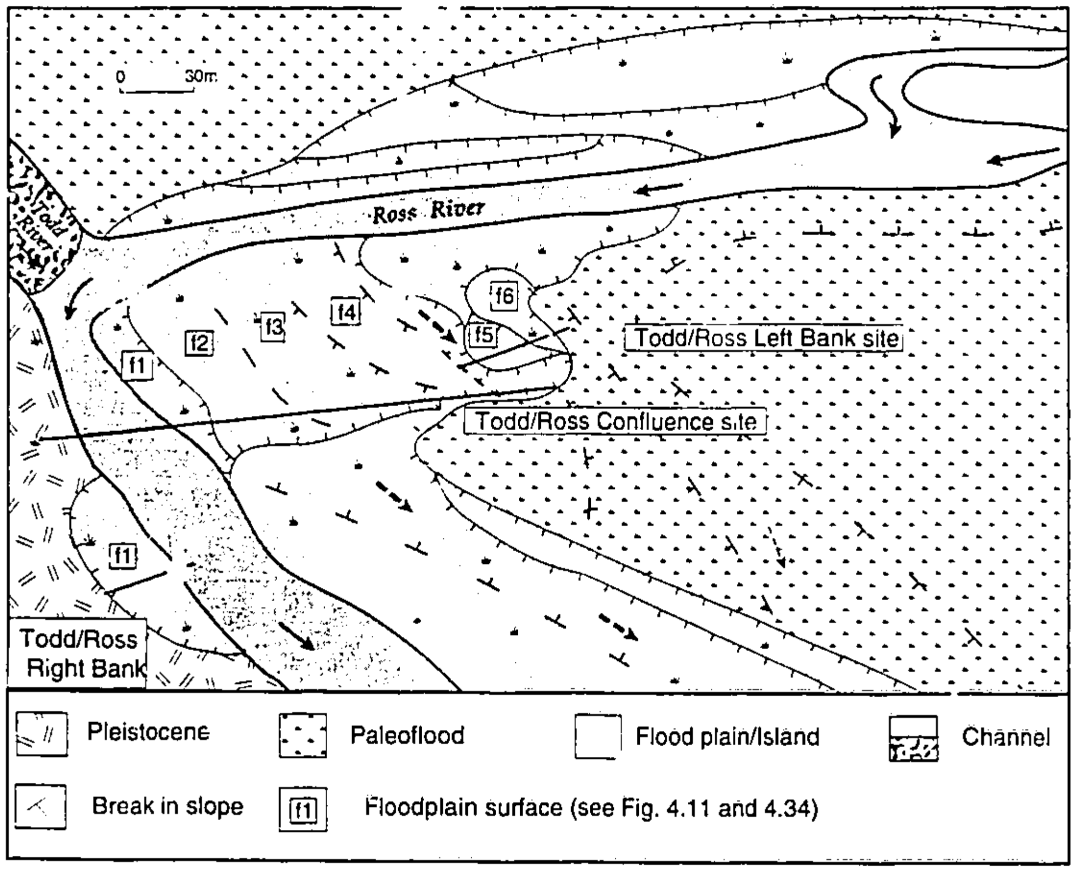


Figure 4.33. Schematic planform of the Ross/Todd confluence.

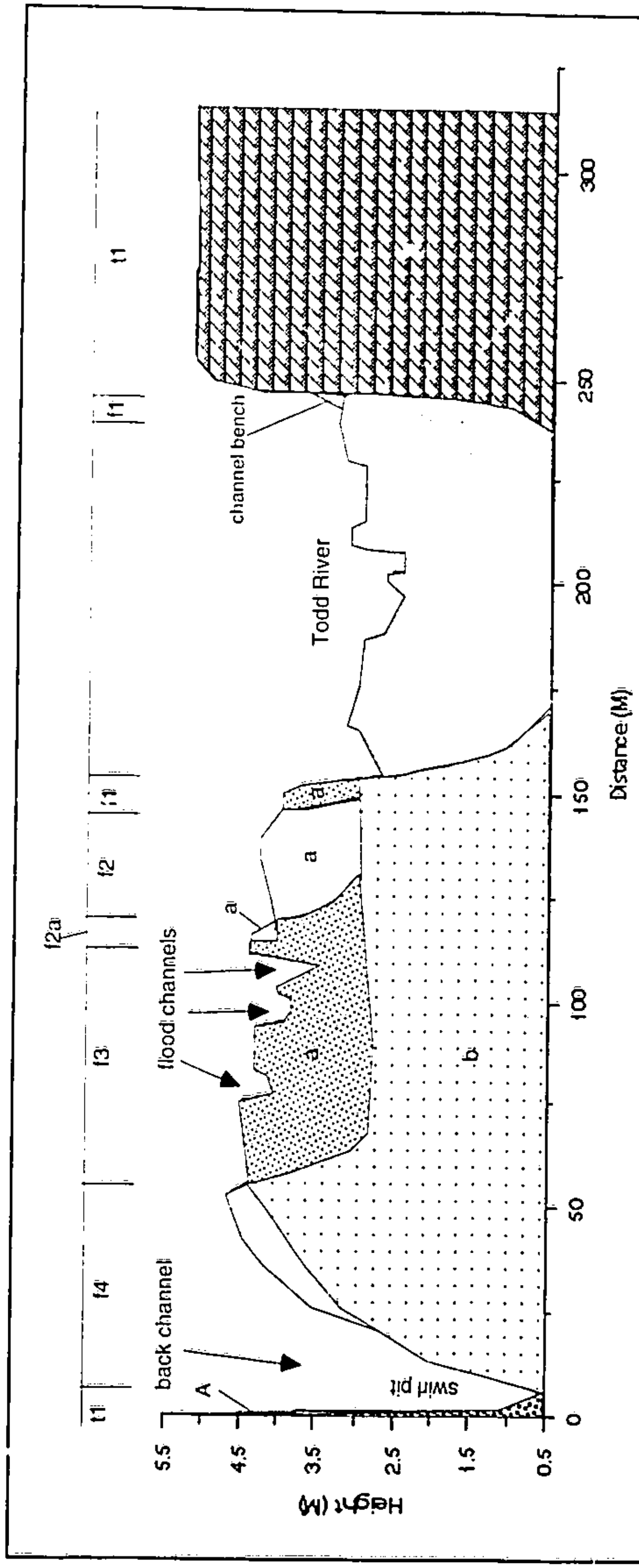


Figure 4.34. The Todd/Ross Confluence site. Type site for confined floodplain morphostratigraphy. a) flood plain inset, b) flood plain remnant (probably channel sediments). For location see Figure 4.2.

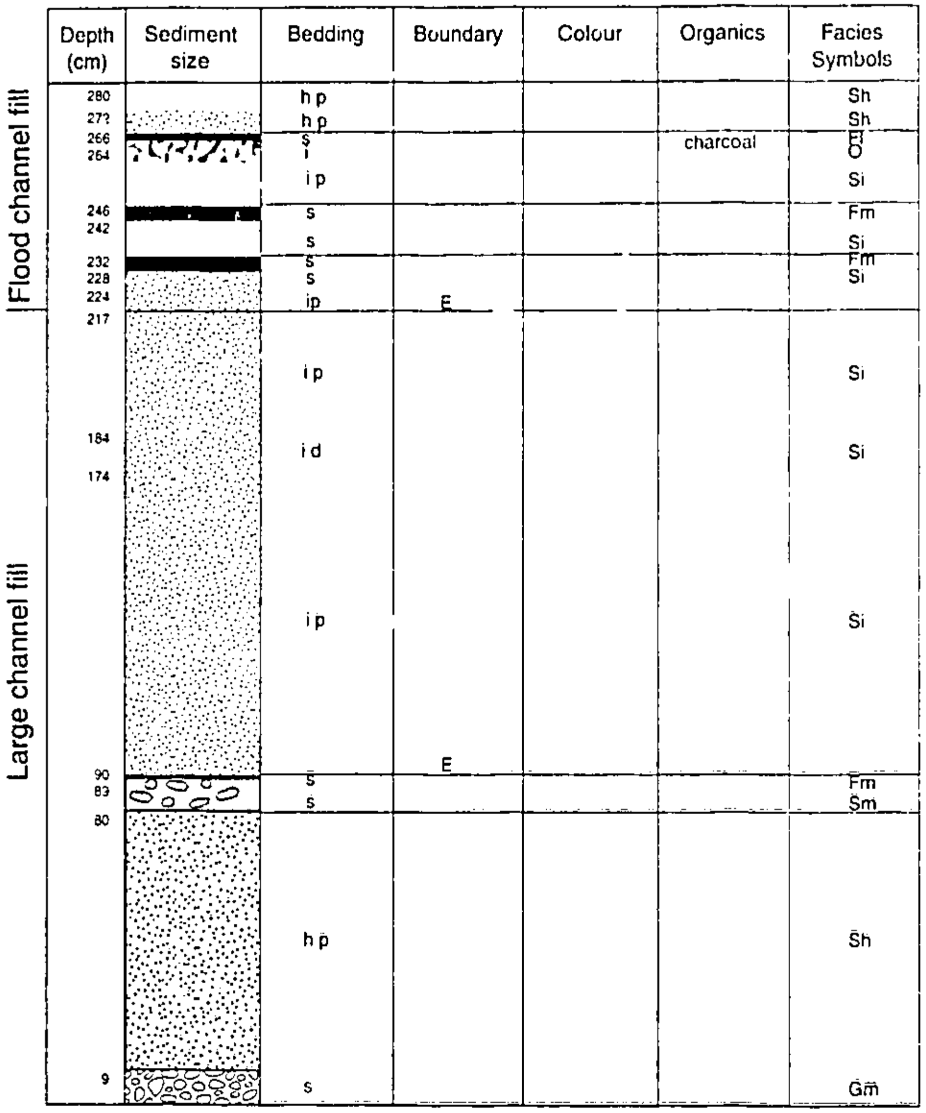


Figure 4.35. Profile C, Anabranching site (location in Fig. 4.18, Plate 4.12). A channel bank exposure in an island showing flood channel fill overlying a large channel fill.

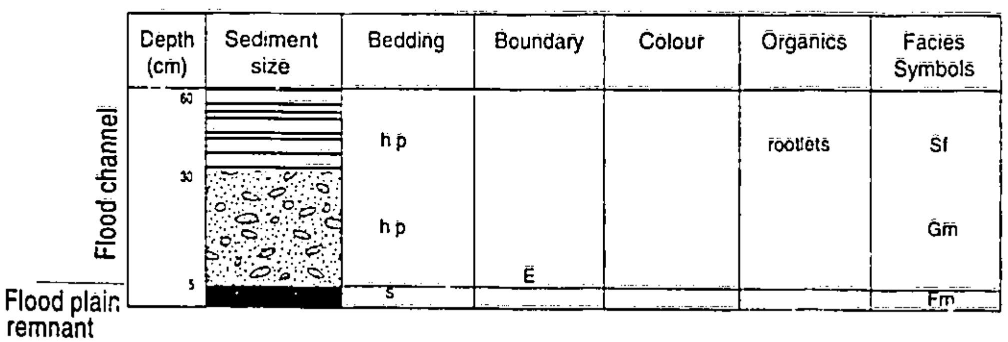


Figure 4.36. Profile C, Todd/Ross Left Bank (location in Fig. 4.11) A flood channel excavation showing flood channel fill overlying a floodplain remnant.

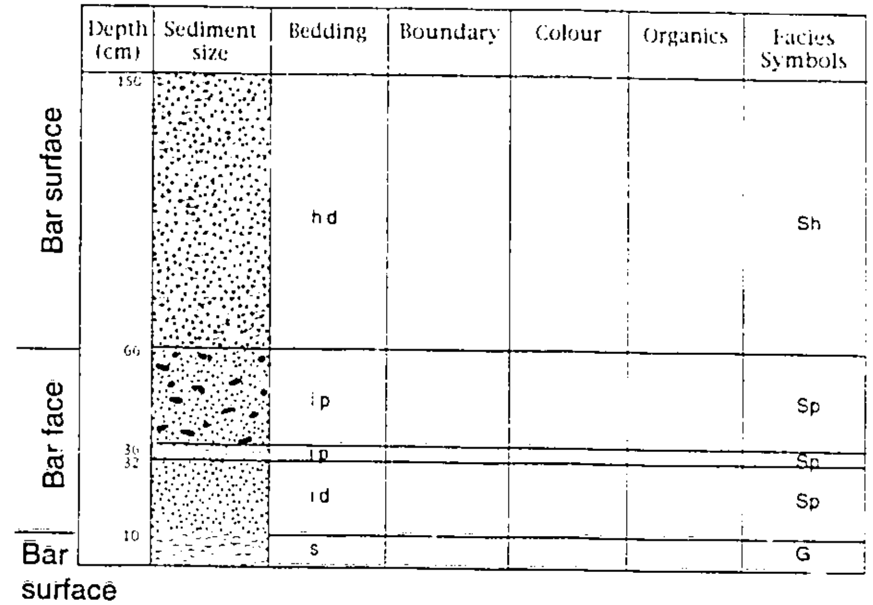


Figure 4.37. Profile C, Todd/Ross Right Bank (location in Fig. 4.10). A channel bank exposure of a channel inset showing bar surface and bar face units.

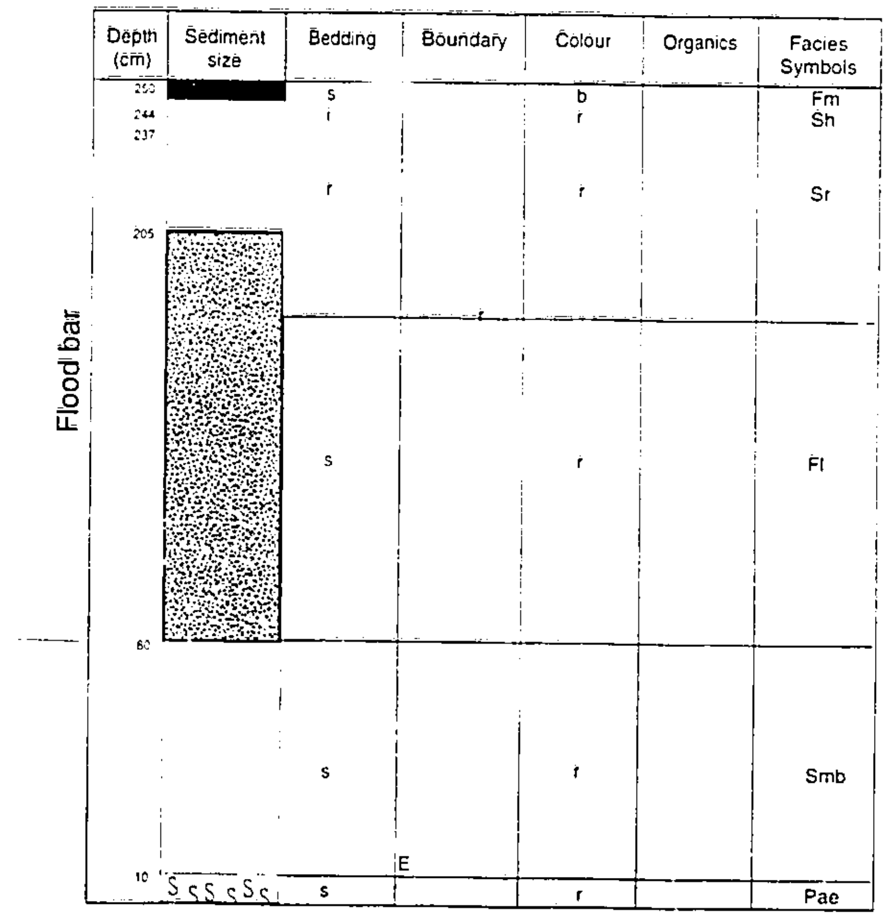


Figure 4.38. Profile E site 'A' (location in Fig. 4.20). Excavation through hummock in flood distributary sh[illegible]ing structureless and ripple laminations overlying cemented Pleistocene aeolian sediment.

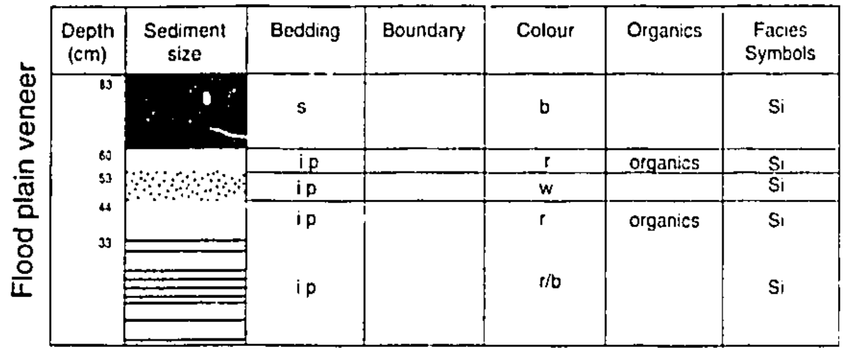


Figure 4.39. Profile E, Todd/Ross Left Bank (location in Fig. 4.11). Excavation through flood plain veneer.

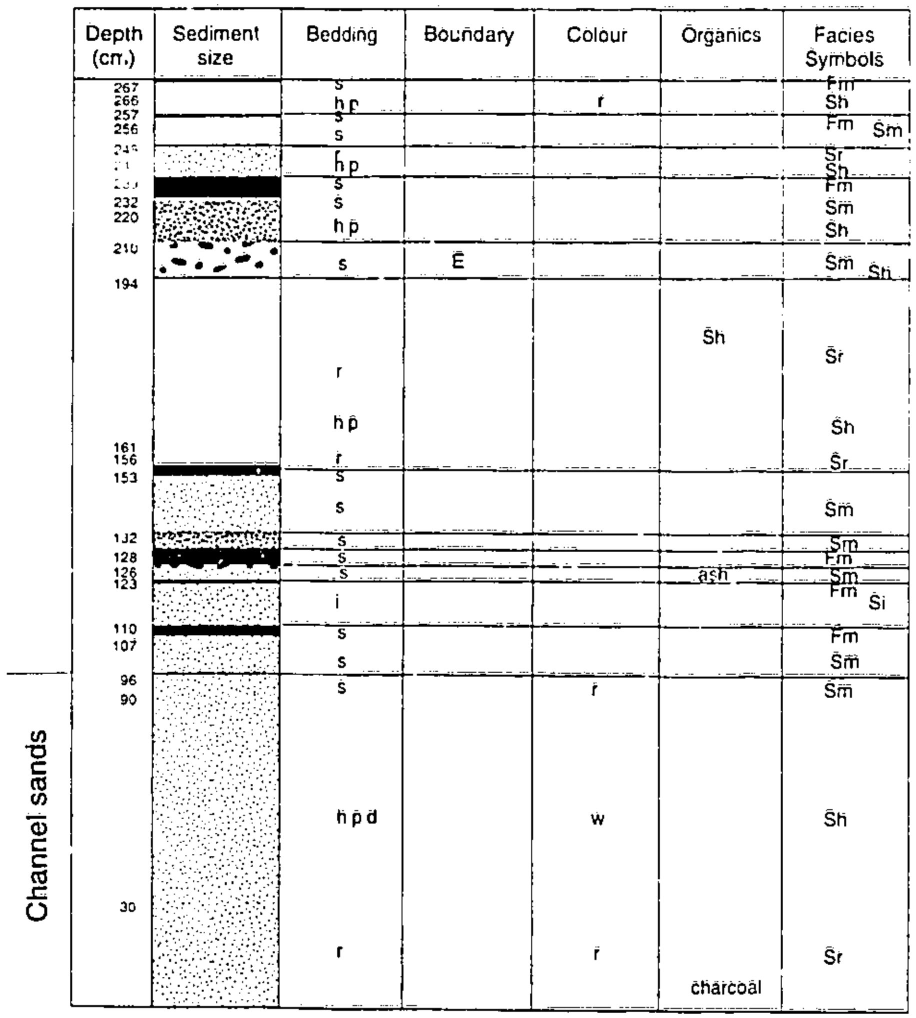


Figure 4.40. Profile C, Mosquito Bore 1, confined flood plain type site, (location in Fig. 4.27). Excavation in right bank showing vertically accreted flood couplets and sand layers over channel sands.

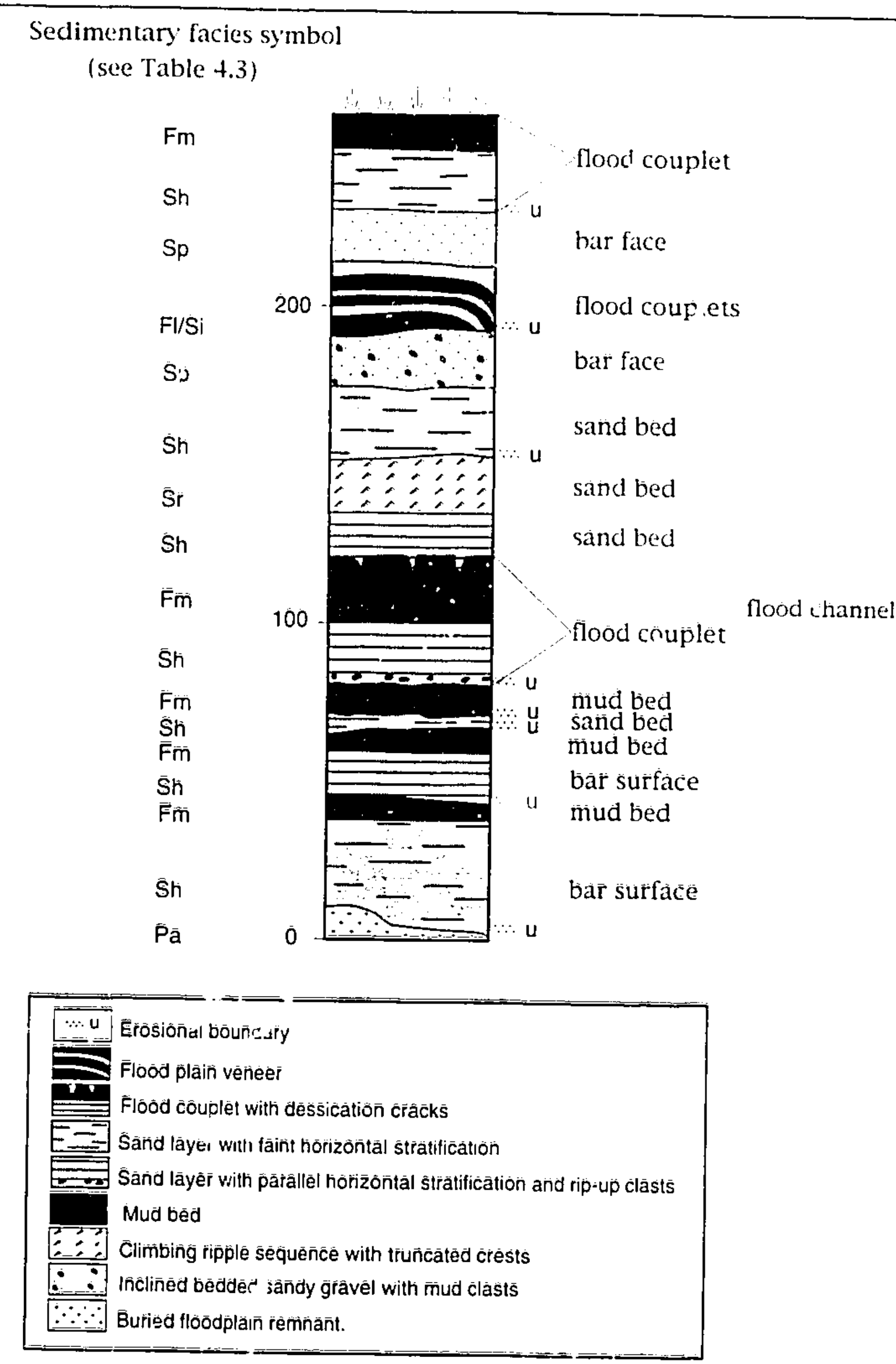


Figure 4.41. Vertical profile stack of Todd River flood plains. The typical thickness of flood plain sediment is 3 m. Any but usually not all of the above components can occur at any site in any order although large-scale channel fill and remnants occur towards the base.

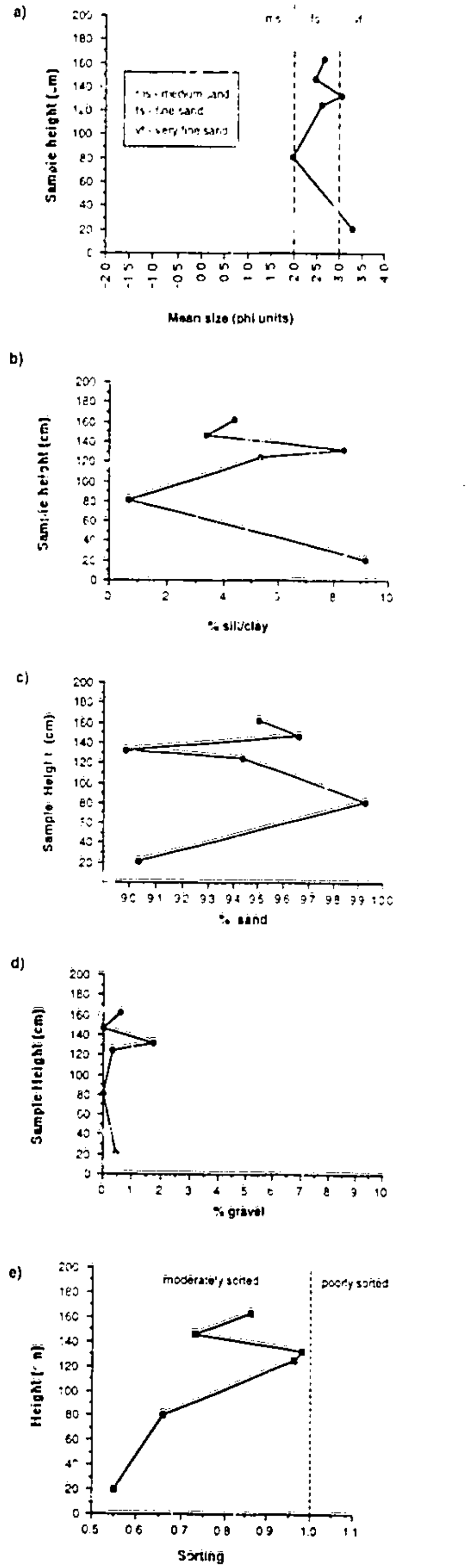


Figure 4.42 Sediment texture at the Mosquito Bore 1 site.
a) mean size b) % silt/clay c) % sand d) % gravel e) sorting

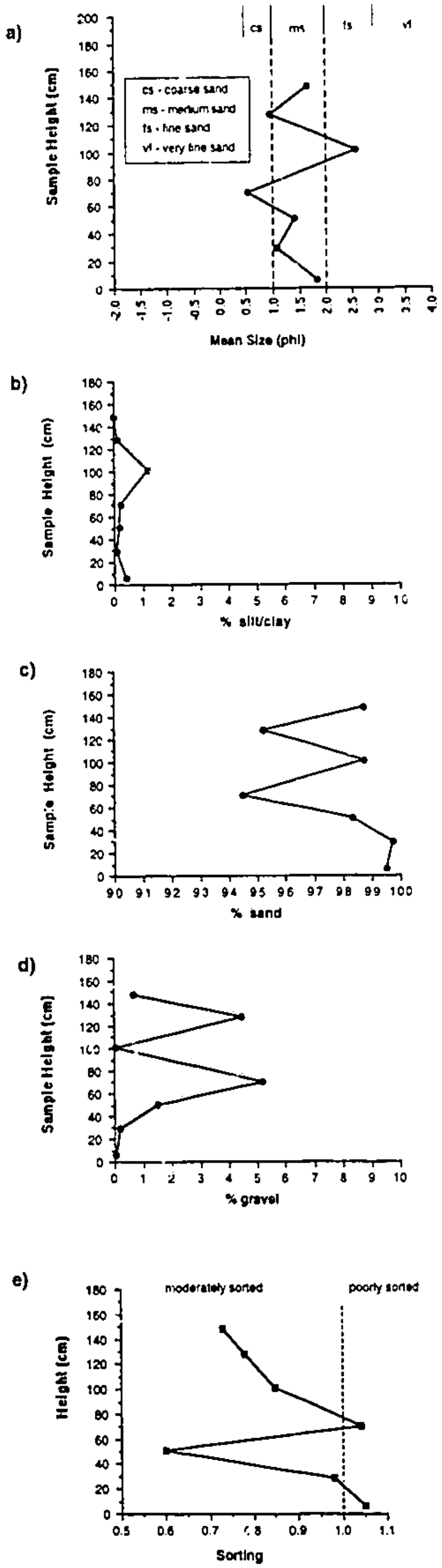


Figure 4.43. Sediment texture at Stud Bore 1site
a) mean size b) % silt/clay c) % sand d) % gravel e) sorting

Site	Channel Width (m)	Flood plain Width (m)	Total Width (m)	Other surface width (m)	Flood plain: Channel	Channel Gradient (m/m)
Todd/Ross Confluence	90.8	155.00	245.8		1.71:1	.00316
Stud Bore 1	69.8	331.5	401.3	115 (Pleistocene terrace)	4.75:1	.00225
Mosquito Bore 1	102.20	211.30	313.5	104 (dune) 156.5 (terrace)	2.07:1	.00063
Mosquito Bore 2	160	10	170	76 ([illegible]) 140 (paleoflood terrace)	0.6:1	NA
No. 5 bore	84.00	5.00	89.00		.06:1	NA
Site A	62.07	270	332.07	200.47 (dune)	4:1	.00164
Atoota Dam 1	28.00	262	290	220 (paleoflood terrace)	9.4:1	.00050
Desert Reach	155.5	155.5	311		1:1	NA
Giles Creek 1	62	10	72	154 (slope) 1467 (paleoflood terrace)	0.7:1	.00122
Giles Creek 2	30.5	0	30.5	809.5 (paleoflood terrace)	0:1	.00295
Giles Creek 3	54 (17, 12, 25)	140	194	1654.5 (paleoflood terrace)	2.6:1	.00481

Table 4.1. Confined flood plain and channel characteristics.

Site	Channel Width (m)	Flood plain Width (m)	Total Width (m)	Other surface width(m)	Flood plain: Channel	Channel Gradient (m/m)
Stud Bore 2	177 (82,95)	607.5	784.5	321.5 (paleoflood terrace) 152 (Pleistocene fan)	3.43:1	NA
Expansion Scour	49.9	417	466.9	83.9 (paleoflood terrace) 37.8 (Pleistocene terrace)	8.36:1	.00087
Anabranching	153.6 (9.9,10.7, 133)	1384	1537.6	25 (dune)	9:1	NA
Ross River Striped Site	112 (57.5, 54.5)	489	787	40 (tributary channel) 24 (terrace)	4.37:1	.00230
Ross River Shannon Bore	114	260	374	50 (SWD) 24.6 (terrace)	2.28:1	NA

Table 4.2. Relatively unconfined flood plain and channel characteristics.

Facies code	Texture/sediment properties	Sedimentary structures	Depositional Environment
Fm	Mud, silt. Generally <10 cm thick. Dark brown in colour	Structureless, desiccation cracks.	Suspended sediment in over bank, drape or waning channel flow and confluence back flooding.
Fl	Sand, silt, mud.	Finely laminated, very small ripples.	Suspended sediment in over bank, drape or waning channel flow and confluence back flooding.
O	Occurs as thin (<2 cm) litter layers in recent (< 5 years) deposits. Often with macro-organics (leaves, twigs) present.	Structureless.	Suspended sediment in over bank, drape or waning channel flow and confluence back flooding.
Sw	Wavy-bedded, medium-fine sands. Average thickness of 15-20 cm	May include sinusoidal ripples	Suspended sediments or lower-flow regime deposits laid down just beyond the threshold of motion.
Sr	Sand, very fine to coarse. Typically <20 cm thick	Ripples with variable internal structure. Typically <3 cm high, 10-15 cm long.	Ripples, lower flow regime.
Sf	Sand-mud	Horizontal parallel laminations	Over bank deposition
Sm	Sand-fine gravels	Structureless.	Rapid deposition over bank/channel
Sl	Sand, fine.	low angle (<10°) cross beds.	Scour fills, crevasse splays, antidunes.
Sh	Horizontally bedded fine-coarse sand/gravel. Often internally graded, in units 20-40 cm thick	Horizontal lamination	Upper flow regime plane bed (or lower flow regime for sands ≤0.6 mm)
Si	Medium-coarse sand, occasionally pebbly, internally graded units.	Inclined parallel stratification.	Veneer
Sp	Medium-coarse sand, occasionally pebbly. Internally grade[illegible]ts, generally >40 cm thick.	Planar-tabular cross beds, dipping at >15°	Foresets from avalanche faces of advancing subaqueous sand sheets.
Gm	Massive or crudely bedded matrix-supported gravels. Typically about 30 cm thick, with average B_{max} of 25 mm.	Occasional horizontal or subplanar bedding. Imbrication.	Bedload deposit: Longitudinal bars, lag deposits, sieve deposits.
G	Clast-supported gravels, with B_{max} up to 200 mm	Often imbricated	Channel framework lag gravels.
Sae	Sand, medium-fine	Thinly laminated	Modern aeolian
Sc	Gravel, sand	Massive, crude inclined stratification	Colluvium
Pa	Gravel lenses, Sand coarse to fine.	Massive, bioturbated, Carbonate rich, rhizomorphs	Pleistocene alluvium
Pae	Sand, medium to fine	Structureless, bioturbated, carbonate rich.	Pleistocene aeolian

Table 4.3. Facies coding scheme used in this study. From Brierley (1991) and Miall (1988b).

Element	Plate/Figure	Scale and geometry	Position in sequence[1]	Sedimentary Unit	Sediment facies (see Table 4.3 for code)	Texture and structures	Depositional Environment
Flood Channels *Small*	*Fig. 4.11* *Plate 4.12*	Elongated, sinuous, usually flat-topped and convex-down at the base. 1 m deep 3 m wide.	Middle-top	Mud beds, sand beds, flood couplets, bar face and bar surface sedimentary units	Sp with a dominance of Sh, Fm. Also O, Sr.	Bedded to structureless medium and coarse sand or sandy matrix supported gravels, mud.	Flood channels, generally braided.
Large	*Fig. 4.16, 4.19* *Plate 4.13, 4.14*	Elongated, sinuous, usually flat-topped and convex-down at the base. 3 m deep and up to 50 m wide.	Basal/top	Bar face and bar surface deposits, sometimes with thin mud layers	Sp, Sh, Gm. Sometimes thin layer of Fl.	Bedded to structureless medium and coarse sand or sandy matrix supported gravels	Abandoned main channel position
Insets *Channel*	*Fig. 4.10* *Plate 4.21*	Elongate, roughly rectangular, flat-topped but may dip channelward and vertical sides with flat base or gently sloping. Bench-like or oblique, generally small scale ≤1 m high also at cm scale.	Basal-middle.	Channel bar surface and bar face, may also include horizontal mud and sand units.	Gm, Sp, Sh, Sr, Fl, Fm, O.	Parallel inclined gravel, coarse sand and mud balls often topped with a planar horizontal sandy gravel.	Incised and laterally truncated channel bed deposit Oblique suspended sediment
Flood plain	*Fig. 4.12*	Elongate, roughly rectangular, flat-topped and vertical sides with flat base. Variable scale from cm to m.	Middle-top	Mud, sand, flood couplet, bar surface and bar face	Fm, Fl, O, Sw, Sr, Sl, Sh, St, Sp, Gm.	Horizontal, inclined and structureless, gravel, sand, and mud	Suspended sediment and migrating bedforms on flood plain surface.
Flood plain veneer	*Fig. 4.31* *Plate 4.16*	Elongate, irregular-shaped bodies which derive their form from the morphology of the underlying surface - stepped or oblique. From cm to m in thickness.	Top	Sand, mud layers and flood couplets	O, Fl, Fm, Sw, Sr, Sh (Ifr)	Bedded to structureless medium and coarse sand, mud.	Suspended load drape, migrating sheet.
Swirl pit	*Fig. 4.32* *Plate 4.17*	Semi-elongate, bowl-shaped. 1-3m deep and 1-6 m wide.	Middle to top	Mud, sand and flood couplet.	Sp, Sh, Sr, Sw, O, Fl, Fm.	Bedded to structureless mud, medium and coarse sand or sandy matrix supported gravels.	Elliptical scours generally associated with trees
Flood plain remnant *Pleistocene alluvial*	*Fig. 4.11* *Plate 4.1*	Truncated morphology	Bottom. May extend from bottom to top of sequence but be laterally discontinuous.		Sh, Sp, Gm, G.	Generally structureless sand, and gravel. Gravel may be in layers, rich in carbonate and rhizomorphs.	Truncated remnant of Pleistocene alluvial fill.
Paleoflood				Channel bar surface and bar face, may also include horizontal mud and sand units.	Sh, Sp, Gm, G.	Structureless to bedded coarse gravel and cobble, or sandy matrix supported gravels.	Truncated remnant of Pleistocene or Holocene alluvial fill.
Pleistocene aeolian					Sp, Sh	Generally structureless well sorted fine and medium sand, usually red, may be cross bedded, rich in carbonate.	Truncated remnant of Pleistocene aeolain dune.

Table 4.4. Summary of Todd River flood plain morphostratigraphy and sedimentary facies. [1]Position in sequence refers to position observed in profiles measured from basal surface.

Sample Name	Site	Depth (cm)	Age	Age (cal BP)	ANU Code	Material
	Todd/Ross Left Bank					
C19	H	88	114.9±1.2%M		9292	Wood, Charcoal
C20	G	66	102.1±2.0%M		9293	Charcoal
C21	E	59	102.4±2.7%M		9294	Charcoal
	Mosquito Bore 1					
C1	A	140	190±90 BP (97.7±1.1%M)	324-0	9276	Charcoal
C6	H	79	125.9±3.5%M		9280	Charcoal
	No. 5 Bore					
CI		20	600±130 BP	753-421	9656	Charcoal
	'A'					
CE	J	125	>modern		9652	Charcoal
CG		40	146±49 BP (98.2±0.6%M)	279-51	9654	Charcoal
	Desert Reach					
CM	A	180	4970±120 BP	5936-5561	9659	Charcoal

Table 4.5. ^{14}C of confined flood plains.

Sample Name	Site	Depth (cm)	Age BP	Age (cal BP)	ANU CODE	Material
	Ross River Striped					
RR 1.1g	A	208	490±70 BP	552-469	8962	Charcoal
RR 1.1M	A	190	147.8 ±3.8%M		8963	Charcoal
RR1.2b	D	306	105.8±3.0%M		8970	Charcoal
RR 1.2a	D	22	> modern		8968	Charcoal
	Anabranching					
'95 C1	A	120	300±50 BP	479-277	9676	Burnt wood
CF	C	34	129±2.5%M		9653	Charcoal
Jg 2.1 I	*Jessie Gap (Fan)*	48	104.5±3.7%M		8971	Charcoal

Table 4.6. ^{14}C age of relatively unconfined flood plains.

Sample Code	Site	CSIRO Code	Depth from surface (cm)	^{137}C Activity (Bq/kg)	^{210}Pb Activity (Bq/kg)	^{226}Ra (Bq/kg)	Age estimate
	Mosquito Bore 1						
MB6	A	[illegible].01	72-93	-.014±.241	50.293±3.205	53.543±.564	pre 1960
	Toss/Ross Left Bank						
T/R A1 5-10	G	AD236A.01	5-10	9.103±.292	161.523±4.061	41.668±.387	post 1960
T/R A1 75-82	G	AD234B.01	75-82	-.014±.241	50.293±3.205	53.543±.564	pre 1960
T/R A2 0-11	E	AD237A.01	0-11	7.578±.482	133.008±6.330	63.634±.711	post 1960
	Stud Bore 1						
SB LB 2@50	G	AB035M.01	50	1.508±.741	67.787±4.276	62.004±1.679	pre 1960

Table 4.7. ^{137} Caesium and ^{21}Lead concentrations in confined flood plain sediments.

Sample Code	Site	CSIRO Code	Depth from surface (cm)	^{137}C Activity (Bq/kg)	^{210}Pb Activity (Bq/kg)	^{226}Ra (Bq/kg)	Age estimate
	Ross River Striped Site						
RR1.1U	A	AD235A.01	80	.715±.132	41.575±1.926	46.245±293	pre 1960
RR1.1N	A	AD233M.01	160	.440±.664	39.840±7.120	37.797±.922	pre 1960
RR1.2 (v)	D	AD238M.01	65-69	-.305±1.346	58.331±13.555	56.477±1.818	pre 1960
RR1.2 (xvii)	D	AB036M.01	256-260	-.135±.093	36.455±5.430	40.483±1.676	pre 1960

Table 4.8. ^{137} Caesium and ^{21}Lead concentrations in relatively unconfined flood plain sediments. Note negative values are consistent with zero given the uncertanties.

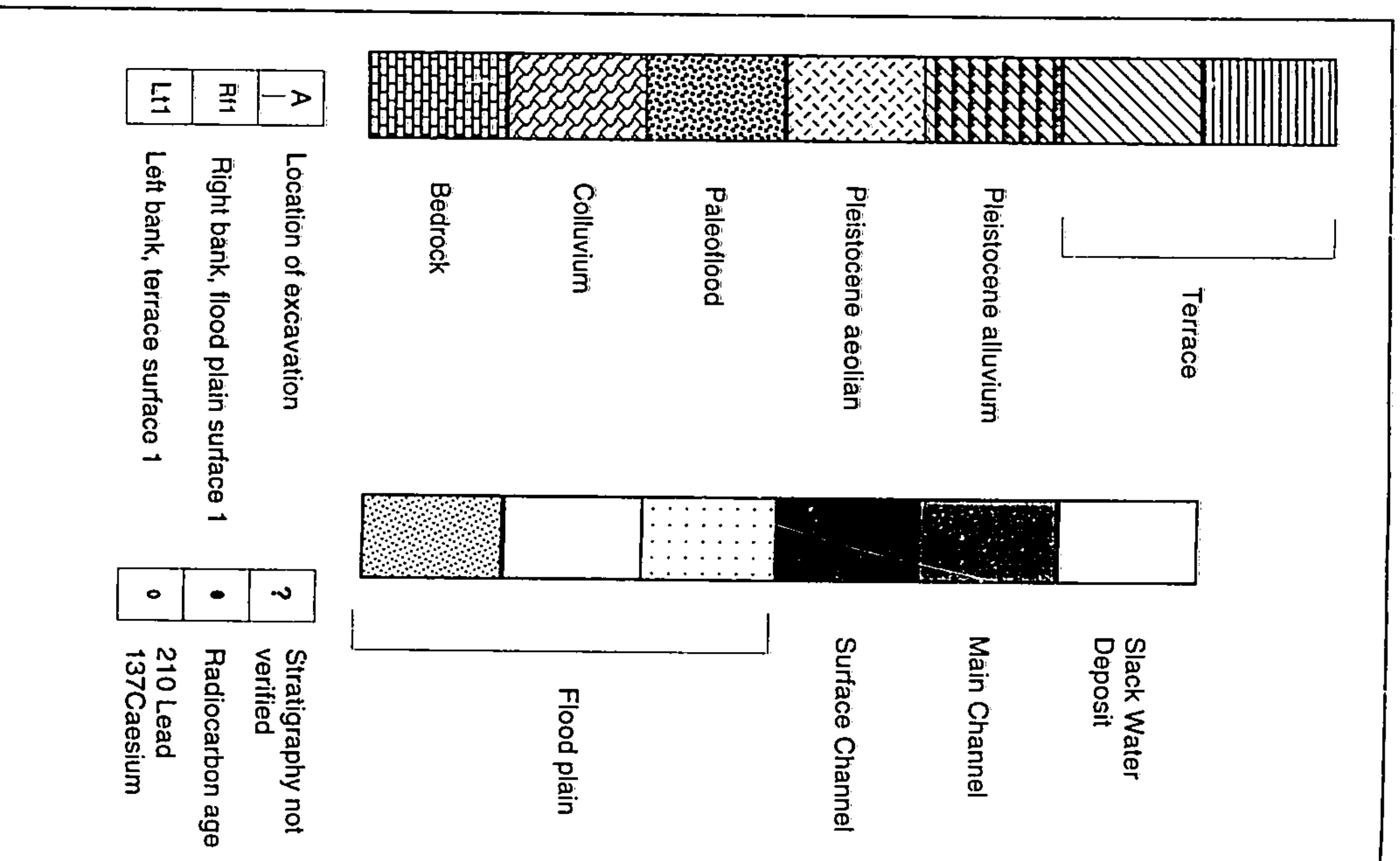


Figure 4.44. Legend for flood plain morphostratigraphic figures.

CHAPTER 5

Todd River Flood Plain Morphodynamics

5.1. Introduction

This chapter discusses flood plain morphodynamics in the Todd River. The focus is on the dynamic changes in form, inferred from observations of processes acting upon sedimentary bodies as well as on the time series of forms observed indirectly from morphostratigraphic relationships. The first section describes flood plain destruction and construction inferred from stratigraphic mapping and observations during and after floods, and a three scale model is proposed. A time series of confluence morphodynamics, switching channels and island formation indicating flood plain and channel change is presented. The next section proposes morphodynamic models for confined and unconfined flood plains which indicate the dominant role of flood magnitude and frequency.

5.2. Flood Plain Formation Processes

This section draws on data presented in Chapter 4 and examines the processes that form flood plains in the study area. The discussion is divided into two sections. The first section is concerned with processes of flood plain destruction and includes channel widening, flood plain stripping, swirl pit scour and surface channel scour. The second section discusses processes that build flood plains and includes the deposition of insets, flood plain veneer, channel and swirl pit fill.

5.2.1. Mechanisms of flood plain destruction

5.2.1.1. Channel widening

Channel widening is a significant process of flood plain destruction in the Todd River. The position of groves of River Red Gums (*Eucalyptus camaldulensis*) that formerly grew on the river banks and now stand within channel beds, sometimes 30m from the present banks, with their roots often exposed, indicates the retreat of the channel banks. In many locations the young age of trees on the cut banks suggests that flood plain aggradation and widening has been a recent phenomenon (Plate 4.11).

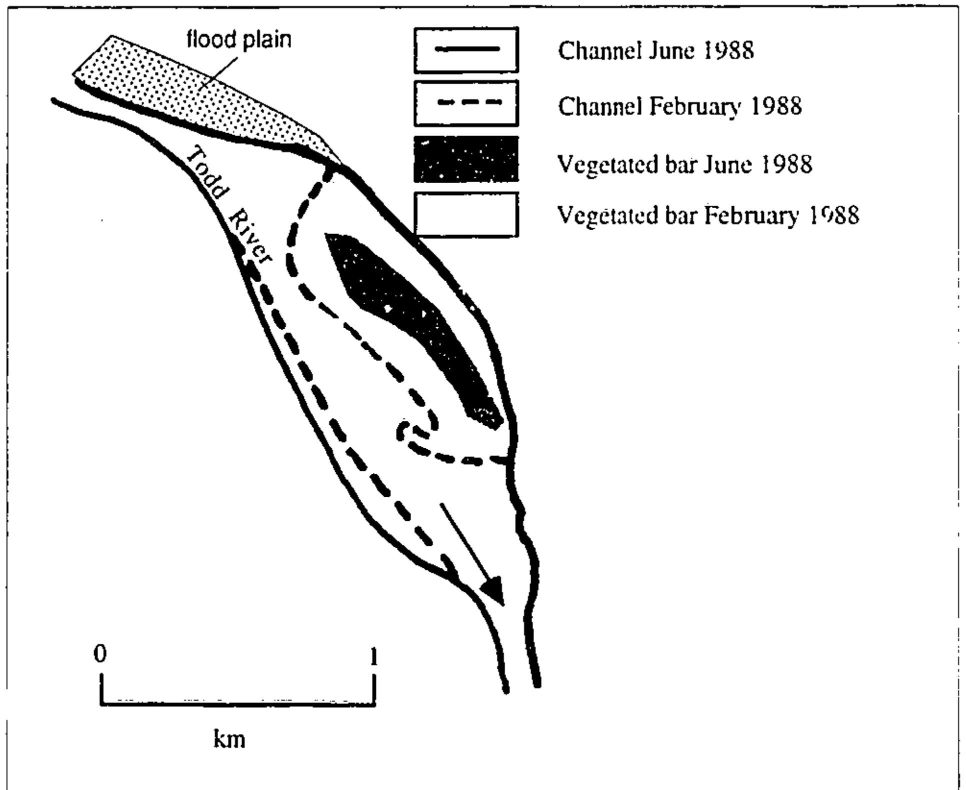


Figure 5.1. Width variation at the Expansion Scour channel bead before and after the 1988 flood.

The principal mechanism of channel widening is direct shear of the banks when shear stresses during floods exceed the bank strength (cf. Schumm and Litchy, 1963). Reaches of significant lateral erosion of the channel boundary are common in the Todd River where channel width increases from ~80 m to ~300 m. These channel widenings ('channel beads'), discussed further in Chapter 8, are inferred to result from local channel aggradation and flow deflection against the banks. However, channel beads are also widened during high magnitude flows. Figure 5.1 illustrates channel change before and after the 1988 flood at the Expansion Scour site where the channel width increased from ~200 m to ~450 m primarily due to the erosion of the large channel bar. Discharge is estimated for the 1988 flood through this reach at 1800 m^3s^{-1} assuming a flow depth of 2 m, flow width of 450 m and a velocity of 2 ms^{-1}. In the reach upstream, flow inundated the left bank flood plain.

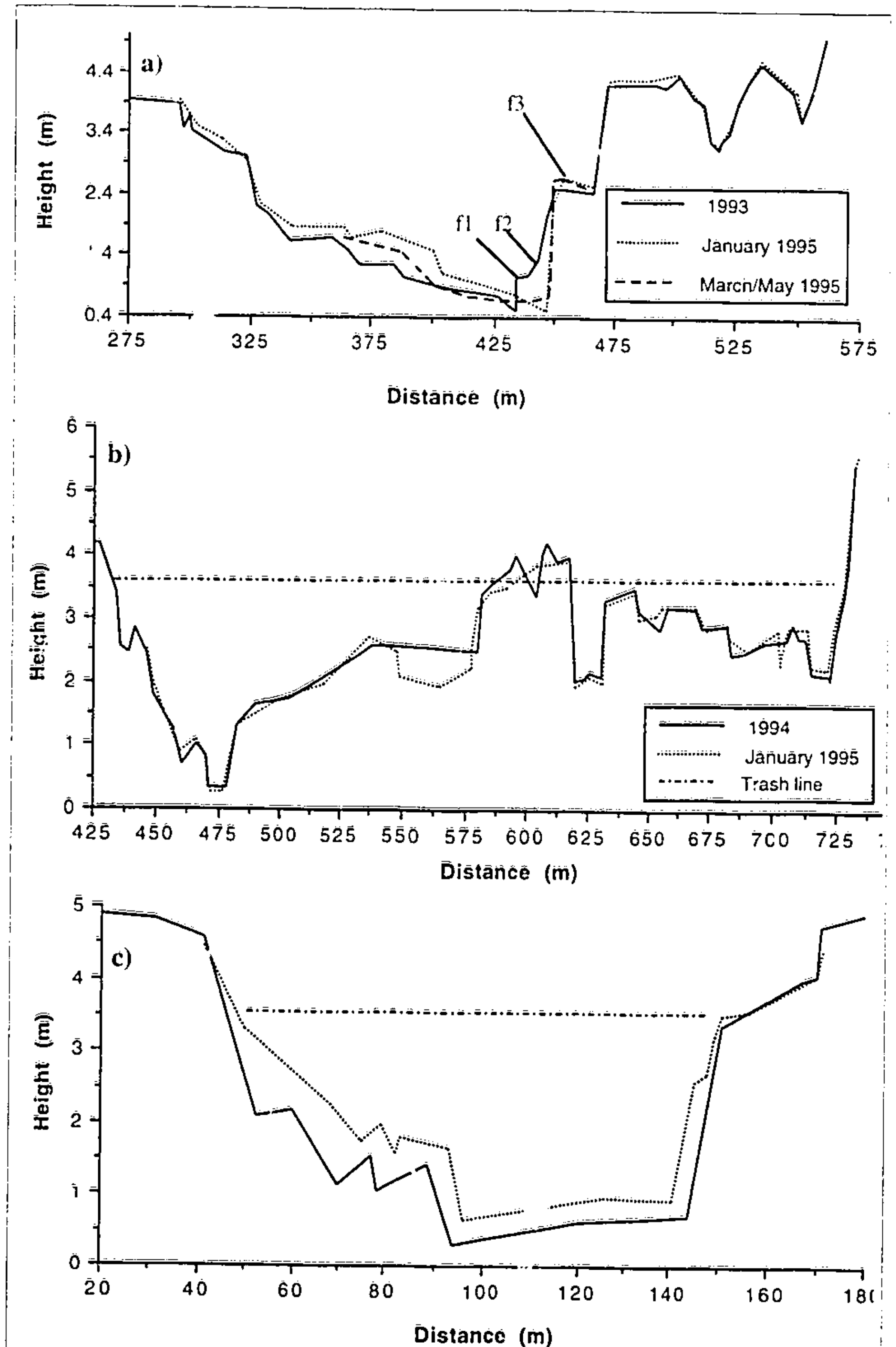


Figure 5.2. Channel changes following the 1995 floods.
a) Mosquito Bore 1, b) Anastomosing site and c) Site 'A'.

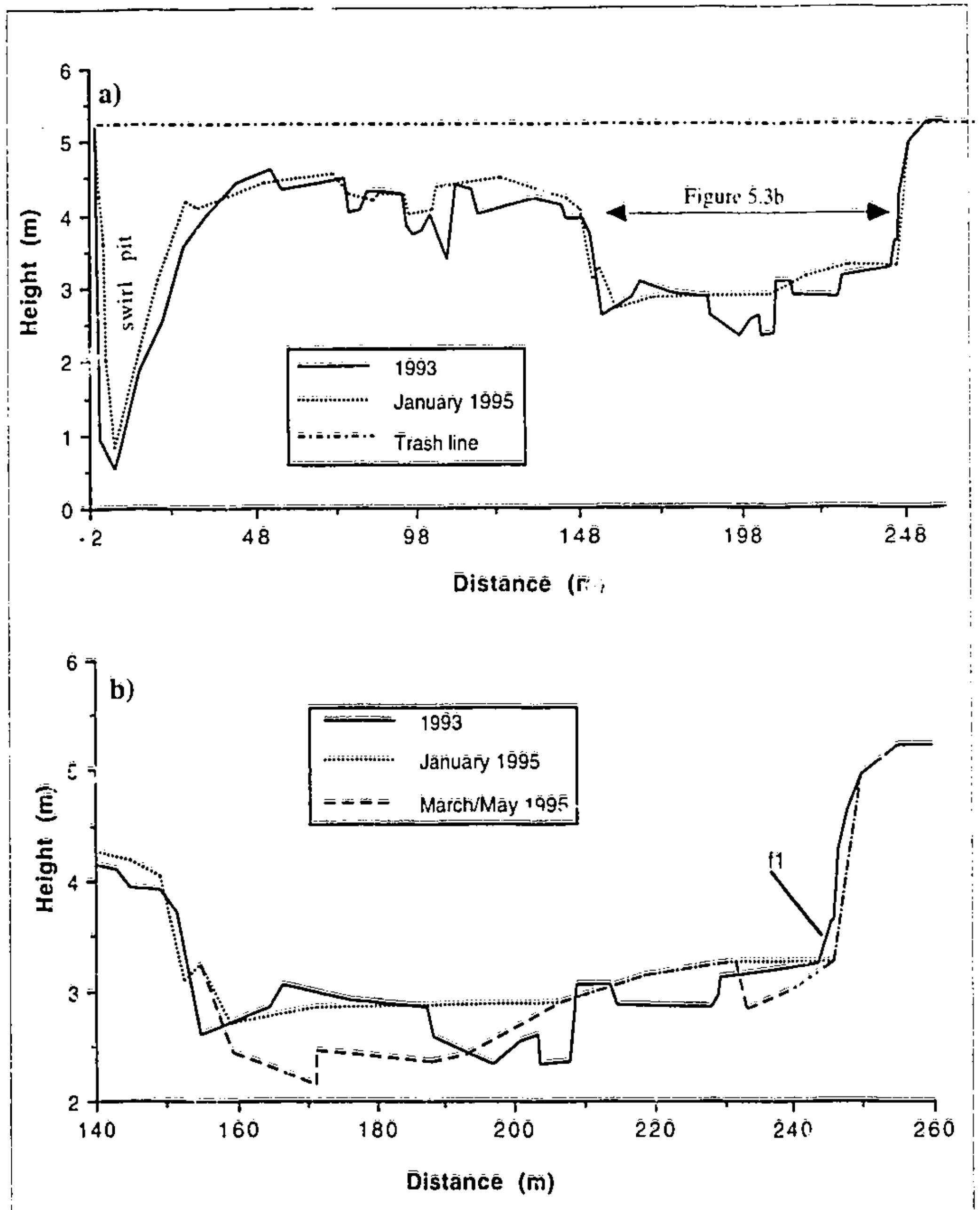
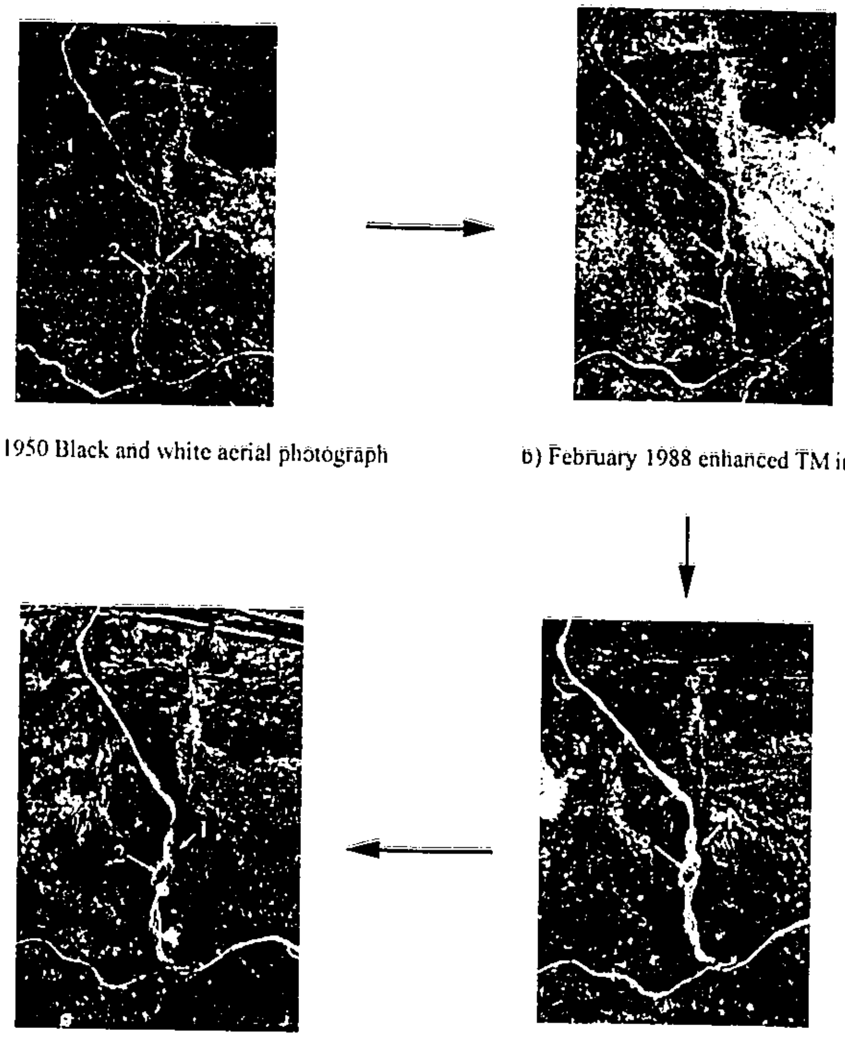


Figure 5.3. Channel change following the 1995 floods at Todd/Ross Confluence site. a) channel and flood plain, b) channel.

Pickup (1991) reported that channel width along reaches of the Ross River increased 300% as a result of the 1970's flows, which were reported to be larger in the Ross River than the recorded peak flow of 583 m^3 s^{-1} (in 1974) at Alice Springs (Barlow, 1988). Local catastrophic channel widening has often been a reported geomorphic effect of high magnitude floods. Scott (1973) observed lateral scour of up to 40 m, a doubling of the flood cross section in the Tujunga Wash, southern California in a large storm in 1969. Osterkamp and Costa (1987) estimated a 160% increase in channel width following the



a) 1950 Black and white aerial photograph

b) February 1988 enhanced TM image

d) 1989 black and white aerial photograph

c) June 1988 enhanced TM image

Figure 5.4. A Time series of channel change from 1950 to 1989.
Numbers are explained in the text and b and c are on Images 5 and 5a (Appendix 4).

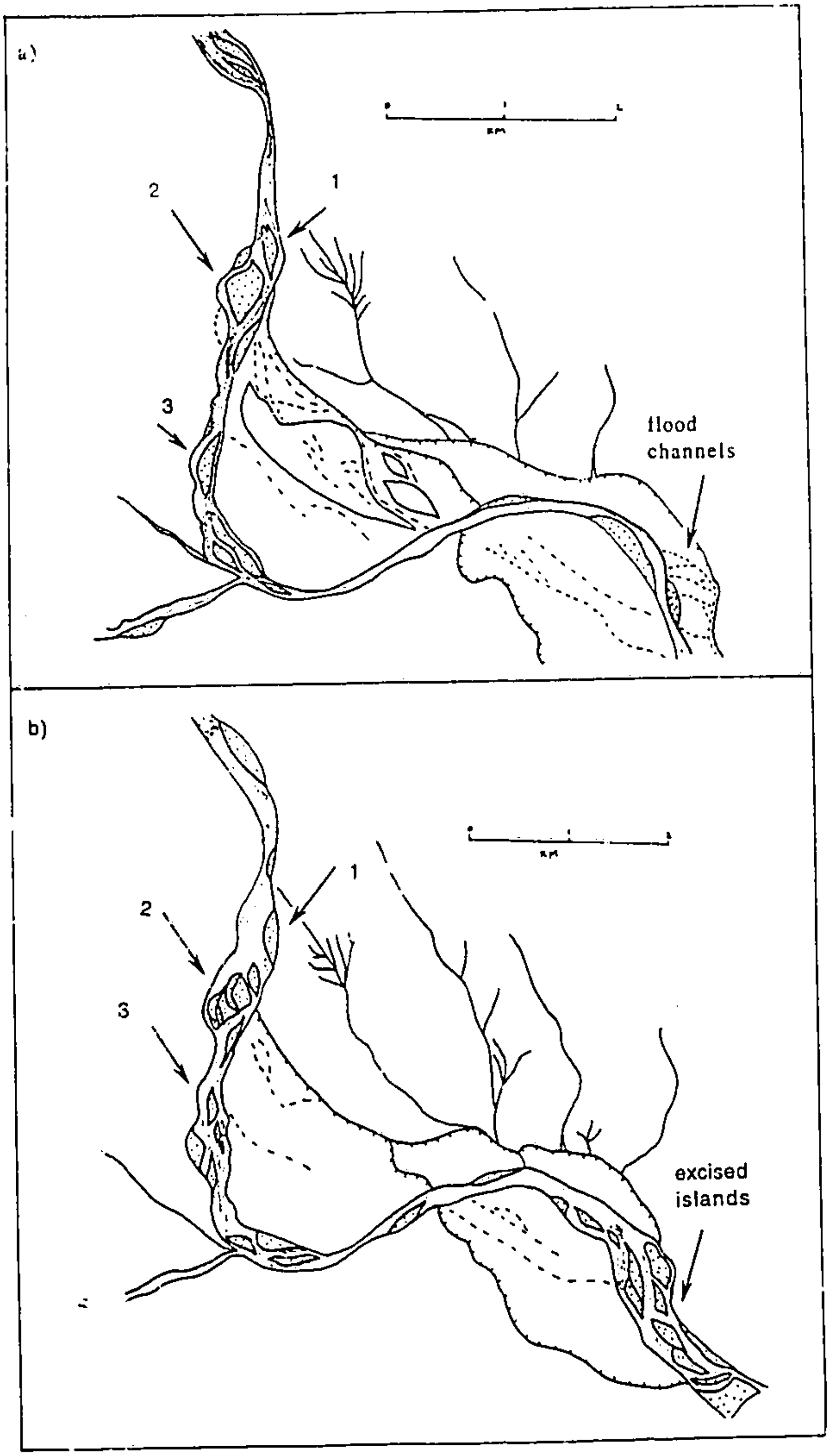


Figure 5.5. Time series of channel change and the excision of flood plains from the bank.
a) Drawn from 1950 aerial photograph. The Todd River left bank flood plain to the lower right of figure is partly incised by flood channels.
b) Drawn from 1989 aerial photograph. The flood plain has been excised from the bank by the exploitation of the flood channels. The arrows are those shown in Figure 5.4.

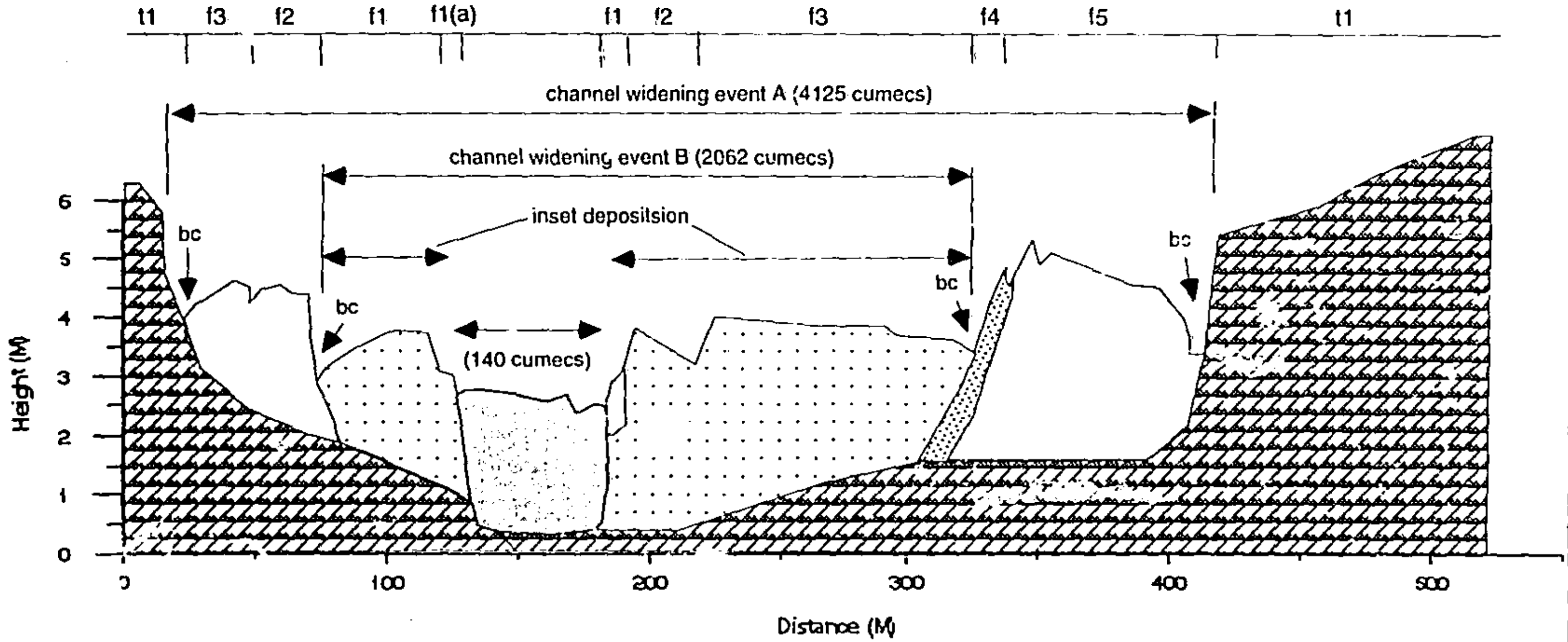


Figure 5.6. The Stud Bore 1 site. For location see Figure 4.2. bc, back channel. Peak discharge is estimated using Manning's equation and a velocity estimate of 2.5 m/sec.

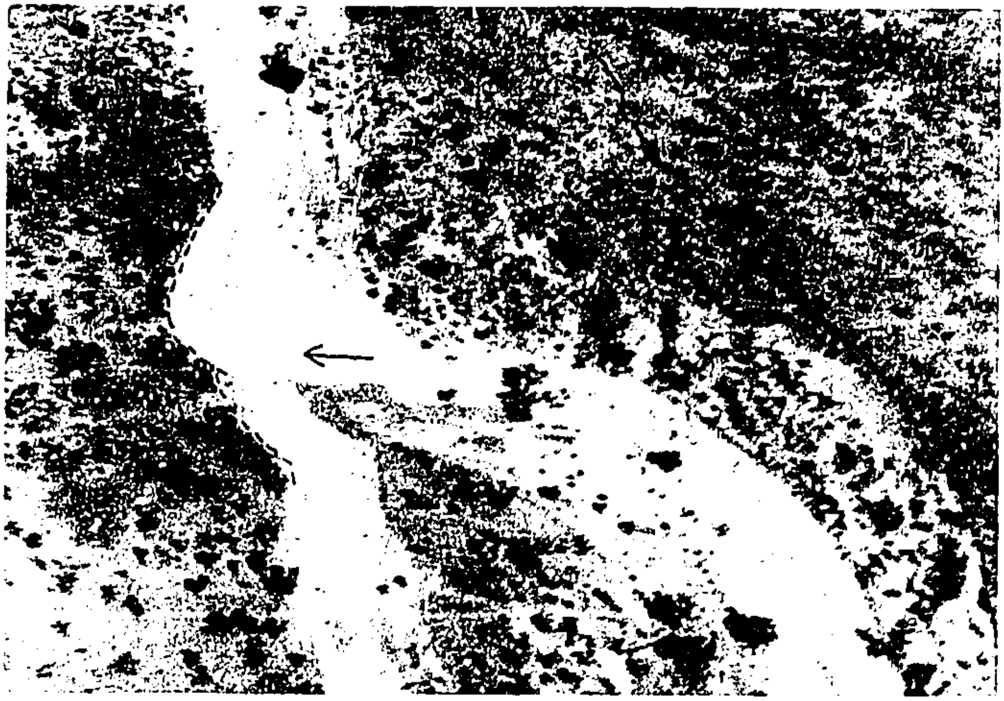


Plate 5.1. An oblique view of a notch forming by flow deflection downstream from a bar on the Ross River .

Plate 5.2. An oblique view of relict notching south of Giles Creek. Flow is towards the right of the plate.

1965 flood on the Plumb Creek, Colorado and Erskine (1994) reported that the Goulburn River was widened from 60 m to 300 m during the 1955 flood.

Measurements were made of channel changes after a series of relatively minor flows in 1995 at four sites: the Todd Ross Confluence, Mosquito Bore 1, the Anabranching site and Site 'A' (Figs. 5.2 and 5.3), which indicate changes in the channel and flood plain. The effects of the January flood were measured within days of the flow and the effects of the March and May floods were measured in September 1995. No further flows were recorded after May 1995.

Minor lateral erosion of the right bank occurred during the January 1995 flood at all four sites (Figs. 5.2 and 5.3). At Mosquito Bore 1 site, surfaces Rf1 and Rf2 were removed and part of surface f3 laterally eroded (Fig. 5.2a). The total width of the channel increased by 15.3 % (15 m) by the lateral shifting of the thalweg against the right bank. The thalweg was not deepened and the left side of the channel prc raded by a maximum of 60 cm. At the Todd/Ross Confluence site (Fig. 5.3b he right bank was laterally eroded by 2.5 m and the left bank by 1 m, a relatively minor 3.2 % increase in channel width as a result of the January flood. Surface Rf1 was removed during this flood and the Pleistocene terrace boundary was eroded.

The removal of stable channel islands results in an increase in channel width. Figure 5.4 is a time series of images of the Ross/Todd confluence from April 1950, February 1988, June 1988 and June 1989. A major flood in March 1988 was a 1:50 year flood at Alice Springs and was also very large on the Ross River. Three processes of island erosion are noted here. The first is the lateral erosion of the left bank lateral bar (arrow 1 in Fig 5.4). In 1950 the lateral bar was incipient with some surface vegetation and had grown and stabilised by February 1988. After the March 1988 flood the width of this bar was reduced by approximately 60%. The second process is exploitation of pre-existing surface channels on large islands and bars. The large diamond-shaped central channel bar (arrow 2) formed from coalesced bars, grew and changed shape between 1950 and February 1988. After the 1988 flood the central island was eroded and reduced in size by approximately 50%; exploitation and enlargement of surface channels appears to have been the principal mechanism (Fig.5.4c). If significantly enlarged, surface channels may be occupied by one of the threads of the main channel or the new location for a switched channel (second order avulsion, Nanson and Knighton, 1996).

Not only is this mechanism important for the erosion of islands but also for reworking flood plains attached to the channel bank. Figure 5.5 is drawn from aerial photographs. Figure 5.5a shows surface channels draining across the left bank flood plain. By 1989 (Fig. 5.5b) a multiple channel pattern was established by erosion of surface channels which separated the flood plain from the bank and divided it into islands and bars. The excision of flood plains from the bank has been described in rivers in south western Kansas and British Columbia (Schumm and Litchy, 1963; Deslodges and Church, 1993).

The third mechanism is the complete removal of islands (arrow 3, Fig 5.4b). The island was not in the channel in 1950 (Fig. 5.4a) but was well established by February 1988. The island was removed during the 1988 flood (Fig. 5.4c) and replaced by a series of smaller bars (Fig. 5.4d). In addition the reach of the Ross channel downstream from the large coalesced bar (arrow 2) appears to have widened during the 1988 flood (Fig. 5.4b, c).

One effect of channel widening is that remnants of laterally eroded flood plains appear as steps on the flood plain surface, and a number of surface flood plain remnants indicate destruction of flood plains by repeated channel widening. The lower magnitude events tend to erode one boundary preferentially but higher magnitude events cause significant lateral erosion of both channel banks. While surfaces that have been laterally truncated by channel widening are more frequently observed close to the channel, for example at the Mosquito Bore 1 site where the truncated surface Rf2 lies 10 m from the bank (Fig. 4.27), truncated surfaces also lie at a distance from the main channel, for example at Stud Bore 1 site where surface Rf4 is 150 m from the main channel (Fig. 5.6). Estimated peak discharge for the event (A) that widened the channel to the Pleistocene terrace is 4125 m^3s^{-1} assuming a cross sectional area of 1650 m^2 and a mean velocity of 2.5 ms^{-1}. Estimated peak discharge for the flood that widened the channel between Lf2 to Rf4 (Fig. 5.6) is 2062 m^3s^{-1} assuming a velocity of 2.5 ms^{-1} and a cross sectional area of 825 m^2, whereas the modern bankfull flow is estimated at 140 m^3s^{-1}. In extreme cases channel widening may completely remove Holocene flood plains during very high magnitude floods, as was reported for the Cimarron River by Schumm and Lichty (1963).

Bank notching is another, if smaller scale, channel widening mechanism in the Todd and occurs in three ways: notching associated with trees, flow deflection downstream of channel bars and and excising of locally erodable bank sediment. Where trees in the channel lie close to the bank (<1 m) a notch which extends the full height of the bank develops downstream of the tree. Osterkamp and Costa (1987) noted that scalloped channel banks were caused by enlarging depressions around fallen trees, and Graeme and Dunkerley (1993) proposed that trees and debris dams in Australian ephemeral channels initiate bank erosion by flow deflection, and that severe scour and erosion induced by riparian vegetation may assist the migration of the channel rather than stabilise it. Notching may also be initiated by flow deflection downstream of bars (Plate 5.1). Bank notching/elliptical scouring is important in beaded reaches which frequently have multiple scours (Plate 5.2). Notching has also been reported on braided rivers in New Zealand (Carson and Griffiths, 1987) and Leddy *et al.* (1993) produced notches in scaled physical models of gravel bed braided rivers from constriction avulsion where the blocking of a braid limb deflects flow against the banks. Notching may also reflect variations of bank sediment. Figure 5.14b shows where an abandoned channel fill is partially re-excised by a notch. Once notches are formed they may be occupied by large horizontal eddies which form in bays or areas where the channel is excessively wide (Matthes, 1947) and channel widening becomes self enhancing. While both caving and crumbling were observed in the study area their importance appears to be relatively minor.

5.2.1.2. Flood plain stripping

Nanson (1986) inferred that flood plains of the Clyde and Manning Rivers in NSW had been stripped during large floods, so that sandy flood plain sediments were stripped from the basal gravel at the downstream end of the flood plain and less than half the flood plain remained as a disjunct pocket of relict alluvium. The stripped zone was ~900 m long and the flood plains upstream and downstream remained intact. Nanson's (1986) model is presented in Figure 5.7.

In the Todd River, flood plain stripping operates at a variety of scales and differs from channel widening in that scour is not always to channel depth, although at the larger scale where Holocene alluvium has been entirely removed it is difficult to identify the specific erosion mechanism. In terms of depth, stripping ranges from the micro-scale (<20 cm) to the meso-scale (<5 m) to the mega-scale (valley bottom sequences) (Fig. 5.9). Micro-scale stripping was observed along Giles Creek when a strip of flood plain alluvium approximately 20 cm deep, 3 m wide and 300 m long was removed from the flood plain surface. At the meso scale flood plains are stripped to depths exceeding 2 m and widths of 30 m for distances of up to 500 m. Evidence described later shows that the entire valley bottom alluvial fill over distances of 10 km has been stripped in the past in mega-scale stripping events.

Flood plain stripping and refilling causes flood plains to be stepped with distinct vertical scarps. Similar to Nanson's (1986) account of the Clyde and Manning Rivers, flood plain stripping in the Todd River terminates at more resistant layers, *e.g.*, at dense mud beds or gravel layers often forming low benches. The erosion morphology of flood plain remnants buried under flood plains is direct evidence of flood plain stripping.

5.2.1.3. Swirl pit scour

The excavation of holes on flood plain surfaces and within channels has been reported from rivers in Ontario, Texas, and NSW (Gardner, 1977; Baker, 1978; Nanson, 1986). They are inferred to result from macroturbulent vortices associated with the flood peak and indicate deep high gradient flood flows (Baker, 1978). They are also associated with the development of vortices around obstacles and along irregular flow boundaries. In the Todd River asymmetric, elliptical scours are formed around River Red Gums (*Eucalyptus camaldulensis*), both in the channel bed and less frequently on flood plain surfaces. The size of the scour is proportional to the size of the tree and/or debris dam and larger scours tend to develop around the large trees (Graeme and Dunkerley, 1993). Scours have eroded through the modern channel bed into Pleistocene sediments at Stud Bore 1 site and have excavated the left bank back channel at the Ross/Todd confluence between the flood plain and the cemented paleoflood deposits to a greater depth than the current channel (Fig. 4.34). Trash levels found in trees above swirl pits indicate that they may form and be sustained by ~2 m deep flows and the tree may be destabilised by scouring and eventually uprooted. Graeme and Dunkerley, (1993) working on ephemeral channels in the Barrier Range, western NSW, found that scours around trees occur where flow velocities are relatively high, the flow is deep and there are few nearby obstructions. Fielding *et al.* (in press) noted scours up to 5 m deep associated with trees in the Burdekin River, Queensland and that the reclined form of *Melaleuca argentea* in the channel bed results in a reduction in scouring around the base of the trunk. Scour holes have also been reported on flood plains in meandering rivers in NSW and braided

rivers in Norway (Nanson, 1986; Nordseth, 1973). However, while swirl pits of various sizes do occur and appear, in some cases, to be long lasting and subjected to repeated excavation and fill, they are relatively minor instruments of flood plain destruction in the Todd system.

5.2.1.4. Flood channel scour

During floods, overbank flow is often concentrated in braid-like threads on the flood plain surface. In addition to braiding around fresh sand deposits, flood channels in the Todd River may also cut into the underlying flood plain sediment. Similar channels have been reported on braided river flood plains in New Zealand and Canada (Carson, 1984; Carson and Griffiths, 1987; Ashmore 1993; Reinfelds and Nanson, 1993) and are the principal form element on braided river flood plains in Norway (Nordseth, 1973). Flood plain channels of the Todd River have low sinuosity and often flow diagonally across the surfaces of low, broad islands. They erode to depths of 1 m and do not necessarily aggrade during waning stages of the flow. Surface channel scour has been reported by others to be a significant mechanism of flood plain erosion, *e.g.*, Reinfelds and Nanson (1993) found that 48% of the braided Waimakariri flood plain in New Zealand was reworked between 1948 and 1960 by the reactivation of abandoned channels within the flood plain, a process also noted by others (Werritty and Fergusson 1980; Carson, 1984b).

Back channels are more effective agents of both lateral and vertical erosion and are often observed to erode the scarp of the adjacent, higher flood plain level. Prominent back channels are often flood chutes while others form at the interface between steps in the flood plain surface and can result in significant vertical erosion. For example, on surface Rf5 at Stud Bore 1 site the back channel has lowered the surface by ~1 m (Fig. 5.6). Larger back channels may be abandoned main channel locations which may later take over the main channel (again) initiating significant flood plain erosion.

5.2.2. Mechanisms of flood plain construction

Flood plains along the study reach are formed predominantly by vertical accretion although lateral accretion is also important. During floods sediment is moved from erosion sites to deposition sites in pulses, and gravel, sand, silt and clay are deposited overbank. Flood plain construction is achieved by a number of mechanisms which include the formation of insets, channel fills, veneer sedimentation and swirl pit fill.

5.2.2.1. Flood plain and channel insets

The principal sequel to channel widening and/or flood plain stripping is the refilling of the stripped region. The insets may be vertically accreted flood channel deposits, vertically accreted overbank deposits or laterally attached channel bars.

Flood plain insets vertically aggrade over the eroded surfaces, *e.g.* following channel widening at the Stud Bore 1 site (Fig. 5.6), two wide (150 m, 50 m) flood plain insets composed of gravelly sand were deposited on both sides of the channel probably as braid bars, (surfaces M1-3 and Lf1/1a, Fig. 5.6) and later aggraded with fine and medium sand sheets and wide coarse sand overbank bars. Once reconstruction of the flood plain has begun, the surfaces of flood plain remnants and flood plain and channel insets appear as steps on the flood plain. Plate 5.3 illustrates sand deposition on a previously stripped bench following the January 1995 flood downstream of the Stud Bore 1 site.

Flood plain insets are formed by the attachment of channel bars in laterally migrating systems. The flood plain inset at the Ross/Todd confluence site (see Fig. 4.33) is inferred to have been initiated in this way and further aggraded by overbank deposition.

The Todd River flood plain is composed of a number of flood plain and channel inset units of variable size, textural composition and age. Deposition of flood plain insets has occurred on most of the flood plain steps and sometimes occurs on the higher steps with no deposition on the lower steps, reflecting the variable discharge regime.

5.2.2.2. Flood plain veneer

Flood plain veneer sedimentation is a process of predominantly vertical accretion, described by Brakenridge (1984) as the lapping up of younger alluvial sediments onto high adjacent surfaces. It effectively blankets the prior flood plain morphology and tends to be better preserved away from the channel. Where the veneer drapes over high, angular flood plain steps (>30 cm) it is generally underlain by oblique accretion deposits which act as depositional ramps (Fig. 5.8a), and on shallower steps the veneer deposits directly overlie the break in slope (Fig. 5.8b). Laminae are continuous over

Plate 5.3. Sand deposition on a stripped flood plain south of Stud Bore following the January 1995 flood. Note also the discontinuous flood plain veneer which extends from the adjacent flood plain step. Trash on fence indicated flow depth.

Plate 5.4. Thin veneer deposit from the January 1995 flood.

breaks in slope and contain flood couplets where the fine textured mud unit follows the depositional slope of the underlying sandy sheet drape. Veneer sedimentation appears to occur in conditions where there is no concentration of overbank flow into distinct threads and the variation from the fine mud to coarse sand in stacks of thin couplets suggests that it occurs continuously through a flood. Veneer sediments were found in confined and relatively unconfined sites. Plate 5.4 shows a thin flood plain veneer from the January 1995 flood.

5.2.2.3. Channel and swirl pit fill

Together, surface channel and swirl pits cause significant and repeated vertical erosion of the flood plain surface, alternating with episodes of fill. Surface channels such as at Site 'A' (Fig. 4.20) are filled by overbank bars and hummocks which tend to concentrate in clusters and are buried by wider flood plain deposits, usually mud and sand.

The irregular surface resulting from erosion by flood channels and swirl pits may be blanketed by more laterally extensive deposits such as sand sheets and veneers. This is found at the Anabranching site and was described in section 4.2.4.3. However, surface depressions may be only partially infilled by wider flood plain deposition, *e.g.*, at the Stud Bore 1 site the migration of a bar across the flood plain surface only partially blocked the surface channel and deflected back channel flow towards the terrace riser. Similarly a sand and gravel bar has only partially migrated down the side of the swirl pit at the Todd Ross Confluence site (Fig. 4.34).

Large channel fills in abandoned channels are important components in relatively unconfined flood plains and the morphodynamics are explored further in section 5.3.2. Volumetrically they may represent up to 30% of the flood plain (*e.g.* Fig. 4.16). Swirl pits observed after recent flows were found to be partly filled with mud which is sometimes stratigraphically preserved under sand fills.

5.2.2.4. Aeolian deposition

Aeolian deposits form minor elements in Todd River flood plains. Trees and shrubs on flood plains act as sediment traps for aeolian deposits and low mounds accrete around their bases, particularly close to aeolian dunes or where the trees have a low habit, or have recently died and collapsed to ground level (Plate 4.5). Aeolian sediments sometimes form a thin layer (<10 cm) on the flood plain surface. While broad dry river beds can be the source of aeolian dunes (Schumm, 1961; Fahnstock and Bradley, 1973; Nanson *et al.*, 1995), source bordering dunes are rare on the modern Todd flood plains but did form in association with paleoflood deposits (Chapter 8). During droughts vegetation cover decreases and sand begins to move; thus aeolian sand dunes 2 m high and 10 m long developed in the Rodinga floodout area in the 1960's (*pers. comm.*, K. Pick).

5.2.3. Three scale cyclical model

Morphologic and stratigraphic investigation in the study area has revealed the highly complex nature of these flood plains. Not only are there several different styles of flood plain destruction and re-formation but these processes operate at different scales (Bourke, 1994). A three scale morphodynamic model has been drawn up which conceptualises the processes of flood plain destruction and construction (Fig 5.9).

The evidence suggests that processes of flood plain construction and destruction have occurred repeatedly and enter the model at three scales. The *micro* scale concerns the removal and deposition of individual sedimentary layers, or part thereof (Fig. 5.9a) and occurs frequently along the channel bank and at the back channel margin (see Unit d, Plate 4.14). The *meso* scale describes processes involved in the erosion and accretion of several meters of alluvium, that is, the alteration of large sections of the flood plain (Fig. 5.9b). The 2 m deep left bank flood plain inset at the Todd/Ross Confluence site (Fig. 4.11) has been formed predominantly by meso-scale processes. The *mega* scale suggests that floods have removed entire valley bottom aggradation sequences in confined reaches of the Todd River (Fig. 5.9c) and large portions of flood plains in the relatively unconfined reaches (not schematically drawn). An example is the Stud Bore 1 site which has been stripped to the Pleistocene terrace in the past and reconstructed by inset deposition (Fig. 5.6).

The majority of radiocarbon ages of flood plains in the Todd River are younger than the 1950's in both the confined and relatively unconfined reaches. The oldest fills are preserved in relatively unconfined sites deposited approximately ~350-400 years BP (at 2 m deep, 552-469 cal BP, ANU 8962, and at 1.2 m deep, 479-277 cal BP ANU 9676) This coincides with the last large flood phase in the Todd (Chapter 7) and is inferred to be the age of the

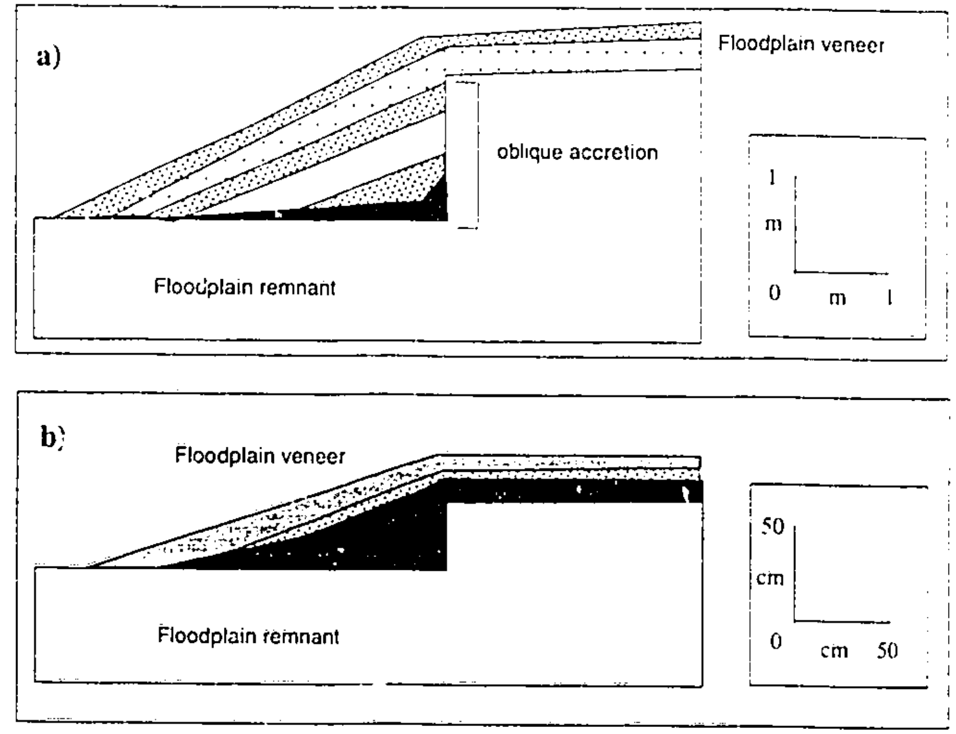


Figure 5.8. Flood plain veneer.
a) the deposition of floodplain veneer over a flood plain step (>30 cm) by the construction of oblique ramps.
b) the deposition of floodplain veneer over a flood plain step (<30 cm).

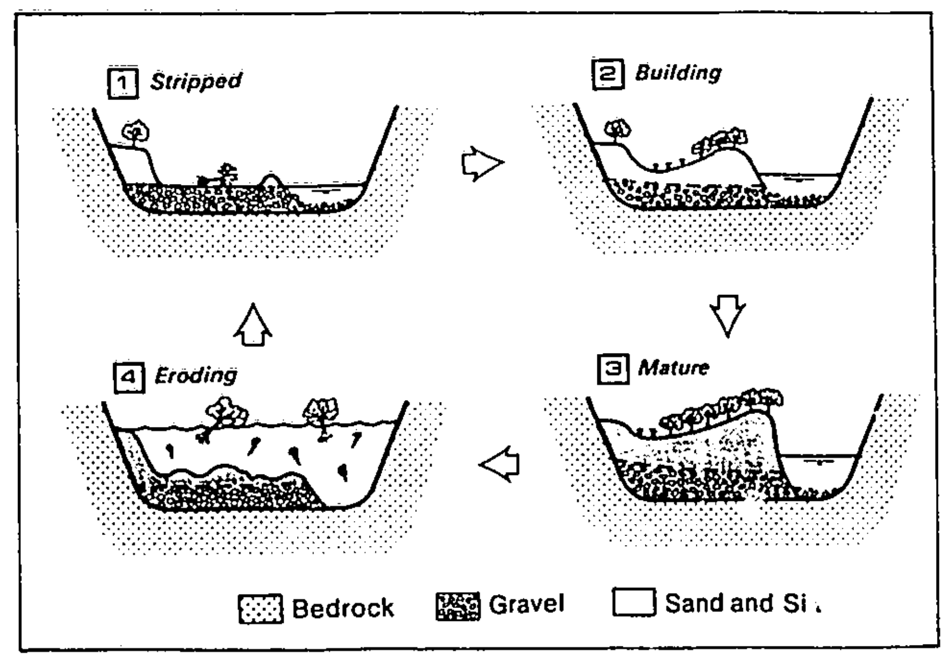


Figure 5.7. Nanson's (1986) episodic cycle of catastrophic stripping and gradual vertical accretion for flood plains along the Clyde and Manning Rivers.

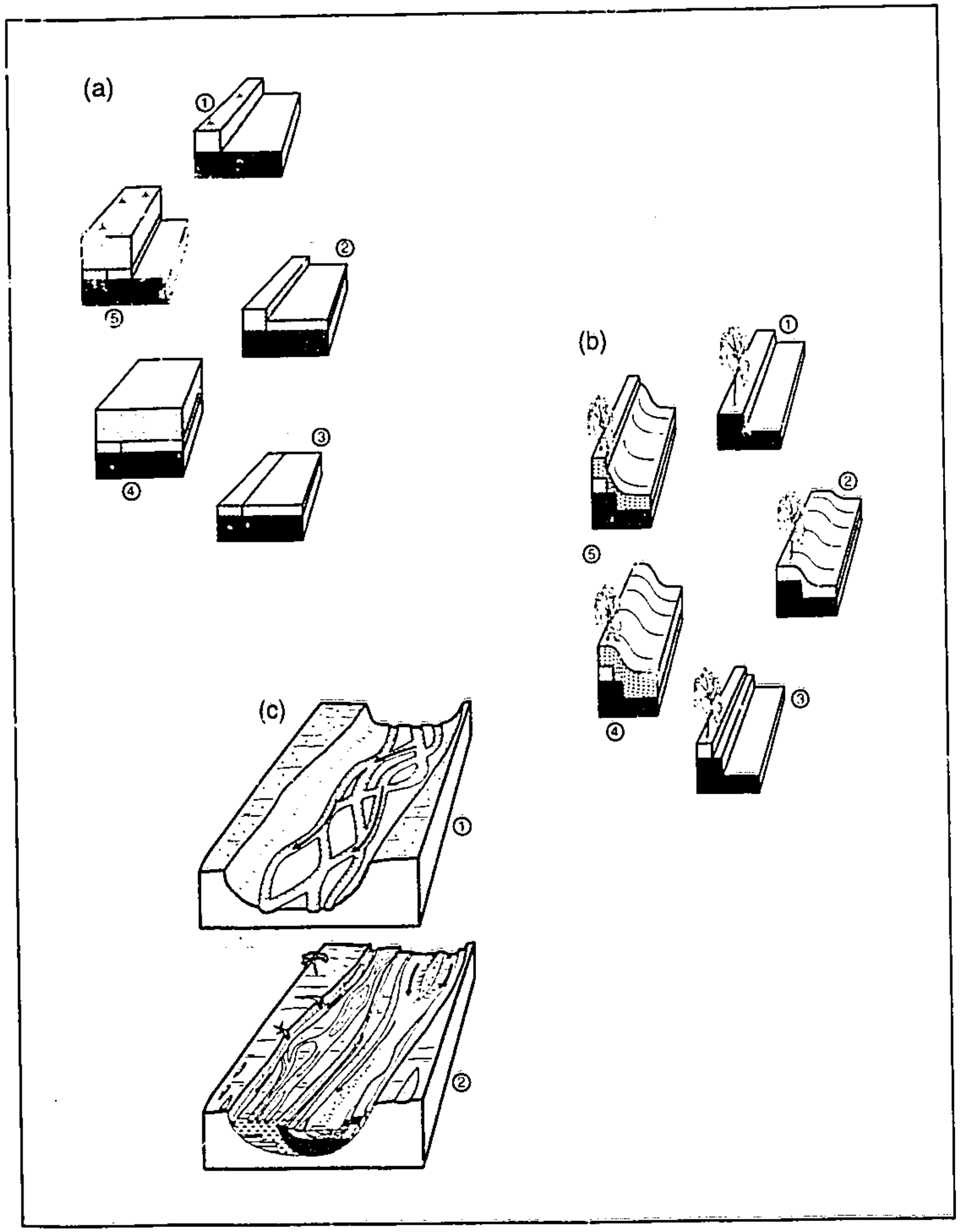


Figure 5.9. Models of flood plain morphodynamics
(a) Micro scale: < 20 cm, Sedimentary layer. 1. Partial stripping of sedimentary layer. Note the erosional scarp. 2. Inset deposition. 3. Stripping of surface. 4. Vertical accretion. 5. Stripping
(b) Meso scale: <3m, Flood plain 1. Stripping of flood plain. 2. Flood plain veneer deposition and burial of flood plain remnant. 3. Stripping of flood plain. 4. Flood plain veneer deposition. 5. Stripping of flood plain. Note: Micro scale processes operate within this scale.
(c) Mega scale: Valley bottom alluvial sequences. 1. Complete removal of valley bottom alluvial sequences to resistant Pleistocene boundary. 2. Rebuilding predominantly by (a) and (b).

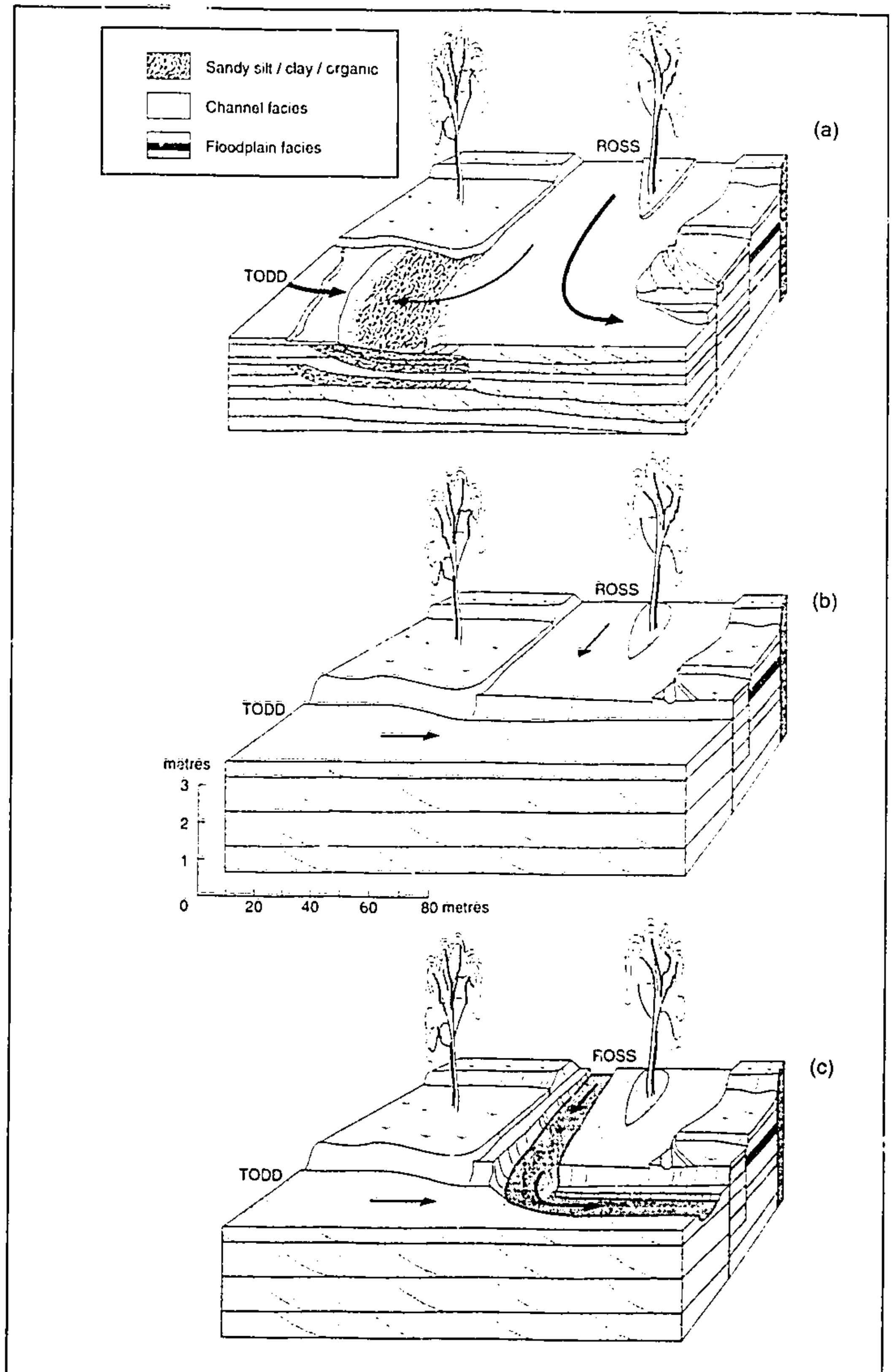


Figure 5.10. Todd/Ross confluence dynamics

a) Deposition from the Ross blocks the Todd.
b) High energy flow down the Todd channel in January 1995 cuts the prior Ross alluvial dam, leaving the Ross mouth hanging above the Todd.
c) Channel incision of the Ross River following the March 1995 event incises both the more recent Todd and prior Ross deposits.

Tributary	Channel distance to confluence (km)		Channel distance to confluence from range outlet (km)		Drainage basin area to confluence (km^2)		Accordant (A) Discordant (D)	Channel (C) Floodout (F)
	Tributary	*Trunk Stream*	*Tributary*	*Trunk Stream*	*Tributary*	*Trunk Stream*		
Emily	32.5	50	7.5	5	200	505	D	F
Jessie	37.5	72.5	10	27.5	145	2350	D	F
Undoolya	22.5	80	12.5	35	90	2650	D	F
Ross	62.5	120	10	77.6	1260	3600	A	Y
Jinker	25	127.5	23	82.5	215	3700	D	Y
Giles	65	145	12.5	100	1000	4800	A	Y

Table 5.1. Tributary and trunk stream length and drainage area at confluence October 1995. Note Accordant channels (A) join at the same elevation, discordant channels (D) join at different elevations.

last mega scale erosion event which catastrophically stripped many flood plains to a basal lag rebuilding has occurred episodically since.

Hughes (1994) noted that semi-arid rivers tend to rework their alluvial fill more rapidly than humid region flood plains. The combination of sandy non-cohesive perimeters and extreme events cause major changes in channel form (Patton and Baker, 1977) and depending on the sequence of recovery events flood plain rebuilding can also be more rapid (Schumm and Litchy, 1963; Wolman and Gerson, 1978). Hughes reported that the flood plain sediment storage time in many rivers in the north American prairies is thought to be around 200-300 years. In the Tana River in semi-arid south east Kenya, flood plain turnover may be as frequent as every 150 years, and a similar time has been postulated for the Animas River flood plain in Colorado (Baker, 1990), whereas in the meandering Beatton River in Canada it is about 700 years (Nanson and Beach, 1977; Hughes, 1994). Discounting the paleoflood deposits beyond the Todd flood plain described in Chapter 6, the present flood plains as mapped in Figure 6.1 appear to have a turnover time of a few hundred years on the basis of ^{14}C ages in Chapter 4, but this probably varies through the system.

5.3. Channel and Flood Plain Morphodynamics: Case Studies of Time Series

5.3.1. Tributary effects and confluence dynamics

In ephemeral streams, the pronounced variability of flow from contributing channels results in exaggerated and often rapid changes in channel morphology (Mabbutt, 1977; Reid *et al.*, 1997). The limited work on ephemeral confluences has noted the propensity for channels to aggrade at or below junctions (Schumm, 1961; Everitt, 1993; Thornes, 1994b) which has been attributed to asynchronous flow (Thornes, 1994b; Finley and Gustavson, 1983), transmission losses (Thornes, 1991; Thornes, 1994a), the disparity of flow magnitudes (Thornes, 1991) and hydrologic lag (Schick and Lekach, 1987).

In locations where the tributary sediment supply exceeds the channel transport capacity, the main channel may be blocked or deflected (Gerson, 1982; Finley and Gustavson, 1983; Schick and Lekach, 1987; Cook *et al.*, 1993; Everitt, 1993) forming barred junctions (Kennedy, 1984). A subsequent lateral migration of the channel away from the tributary (Finley and

Gustavson, 1983; Everitt, 1993) may lead to the development of cutoffs and channel relocation (Everitt, 1993). The tributary sediment subsequently may be moved in waves down the system (Finley and Gustavson, 1983) with the discrete tributary sediment contributions becoming increasingly mixed downstream (Reid *et al.*, 1997). The subsequent removal of the sediment dam may form discordant junctions (Kennedy, 1984), a common feature in ephemeral channels (Schumm, 1961; Cook *et al.*, 1993) resulting from a disparity in flow magnitude (Cook *et al.*, 1993) or a hydrologic lag (Schick and Lekach, 1987).

Step-changes in channel morphology were observed at the Todd and Ross junction. Upstream of the confluence, the Ross River incises the Ross A paleoflood channel (Figure 6.1c), channel banks are composed of loosely packed paleoflood deposits which overlie cemented Pleistocene sediments, and there is an abundant supply of sediment. By way of contrast the Todd River does not have access to a similarly abundant sediment supply. The higher sediment input from the Ross tributary in conjunction with its proximity to the ranges results in channel aggradation downstream from the confluence.

5.3.1.1. Confluence time series

The confluence of the Todd and Ross Rivers was monitored between 1993 and 1995 and illustrates the effects of disparate discharges from the two streams (Fig. 5.10). Before 1995 high flows from the Ross River deposited tabular bars of coarse and medium sand, backfilling the Todd so that laminated fine sand and silty clay was deposited upstream of the barred confluence (Fig. 5.10a, 5.11, Table 5.2). The upstream tabular bars in the Todd are not coeval with the mud.

During the January 1995 event flow from the Ross River peaked at the confluence earlier than flow down the Todd and the Ross River aggraded its bed in a manner similar to that illustrated in Figure 5.10a. The later arrival of the Todd flood incised the Ross alluvial dam leaving the channel bed of the Ross hanging 2.6 m above the thalweg of the Todd River (Figure 5.10b). The discordant junction morphology persisted until a lower magnitude flood during March 1995 incised a 1.5 m deep channel in the perched Ross bed (Fig. 5.10c, 5.12).

Variable	Upstream of Confluence	Downstream of Confluence
Gradient (m/m)	-.0007	.0032
Width (m)	40	180
Bed Material	Horizontally bedded clay-rich silts and fine sands, macro-organics	Crossbedded gravelly sands
Channel Pattern	Straight	Locally braiding

Table 5.2 Todd channel morphology and sediments upstream and downstream of the Ross confluence.

This example of confluence morphodynamics illustrates three important aspects of landform variability in ephemeral systems. Firstly, during high magnitude flows in ephemeral streams, it is the process that controls the forms. The forms created by such flows will control processes during subsequent smaller events (Graf, 1983b). In this way the persistent erosion and depositional patterns generated by paleofloods in the Todd modulate the landscape response to subsequent and smaller events. Secondly, the example highlights the importance of the order and magnitude of preceding events (Pickup and Reiger, 1979) on the geomorphic response to flow events. Thirdly, it emphasises that ephemeral streams are time dependent systems where forms and processes are rarely in equilibrium (Cook *et al.*, 1993).

5.3.1.2. Tributary spacing and length

The variability in the timing and magnitude of flow discharges between the contributing systems and their different sediment loads cause channel morphology and sedimentation to fluctuate above and below confluences in the Todd catchment. One factor affecting the magnitude of flows from contributing catchments is their spacing and length; the spatial and temporal variability of rainfall is also important.

The distance between significant tributaries influences the behaviour of floods downstream. In arid and semi-arid channels transmission losses through drainage diffusion, infiltration and evaporation/ evapotranspiration are significant (*e.g.* Schumm and Hadley, 1957; Schumm, 1961; Knighton and Nanson, 1994). While no transmission data exist for the Todd River, reports from the manager of the Todd River Station indicate that many large flows recorded at Alice Springs do not reach the Ross River confluence 78 km downstream (I. Lovegrove, *pers. comm.*), and flow in the system is augmented by only a few large tributaries, the most significant being the Ross River (1260 km^2) and Giles Creek (1000 km^2).

The proximity of tributaries to sediment sources in the MacDonnell Ranges is an important factor affecting channel morphostratigraphy at confluences. The Todd catchment is such that, south of Alice Springs, the trunk stream drains eastwards along a wide strike valley, joined by tributaries that travel relatively short distances from the ranges (Table 5.1). Accordingly, the sediment loads, textures, channel gradients and flow magnitudes in the larger tributaries are often greater than those in the trunk stream particularly where they are reworking paleoflood sediments. The distance from the ranges to tributary confluence is important, as transmission losses increase in the piedmont alluvial fans. Hence several of the small tributaries such as Jessie Creek have unchanneled junctions with the trunk stream (Table 5.1).

5.3.1.3. Rainfall variability

Asynchronous tributary flow may be the result of the variability in the location, timing and intensity of rainfalls. The event of March 1972 illustrates the typical characteristics of central Australian rainfall (Fig. 5.13). Firstly during the six day rain event, the Giles Creek catchment (Ringwood gauge, Fig. 2.2) recorded peak rainfall on the day prior to peak rainfall receipts in the Ross River catchment (Fig. 5.13). This temporal variability is attributed to the movement of storm cells up-catchment and would have led to downstream tributary systems being active before those upstream. Secondly, rainfall at the Alice Springs Meteorological Office (Fig. 2.2) was 94.2 mm in a twenty-four hour period on March 5th, while a gauge located 15 km away at the Alice Springs Post Office recorded no rainfall during that time period, indicating that rainfall was spatially variable across the catchment. Thirdly, most stations received rainfall over a five to six day period with a daily maximum receipt not exceeding 100 mm. However gauges located in the Ross and Giles catchments had close to the annual average in a twenty-four hour period. These rainfall characteristics would have contributed to asynchronous flow at tributary junctions in the Todd catchment.

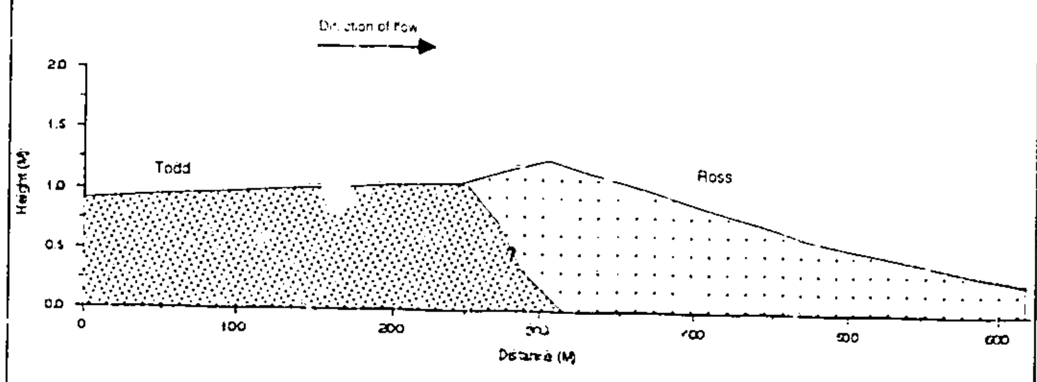


Figure 5.11. Long profile of Todd at Ross River confluence

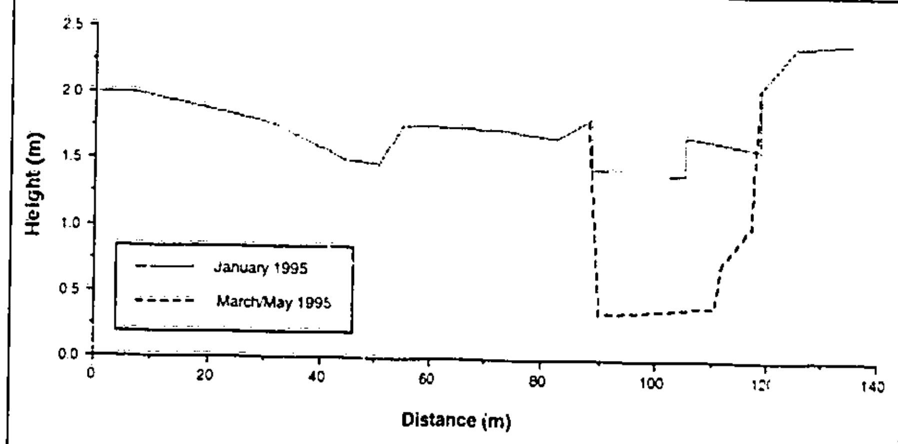


Figure 5.12. Incision of Ross River channel following the March 1995 event.

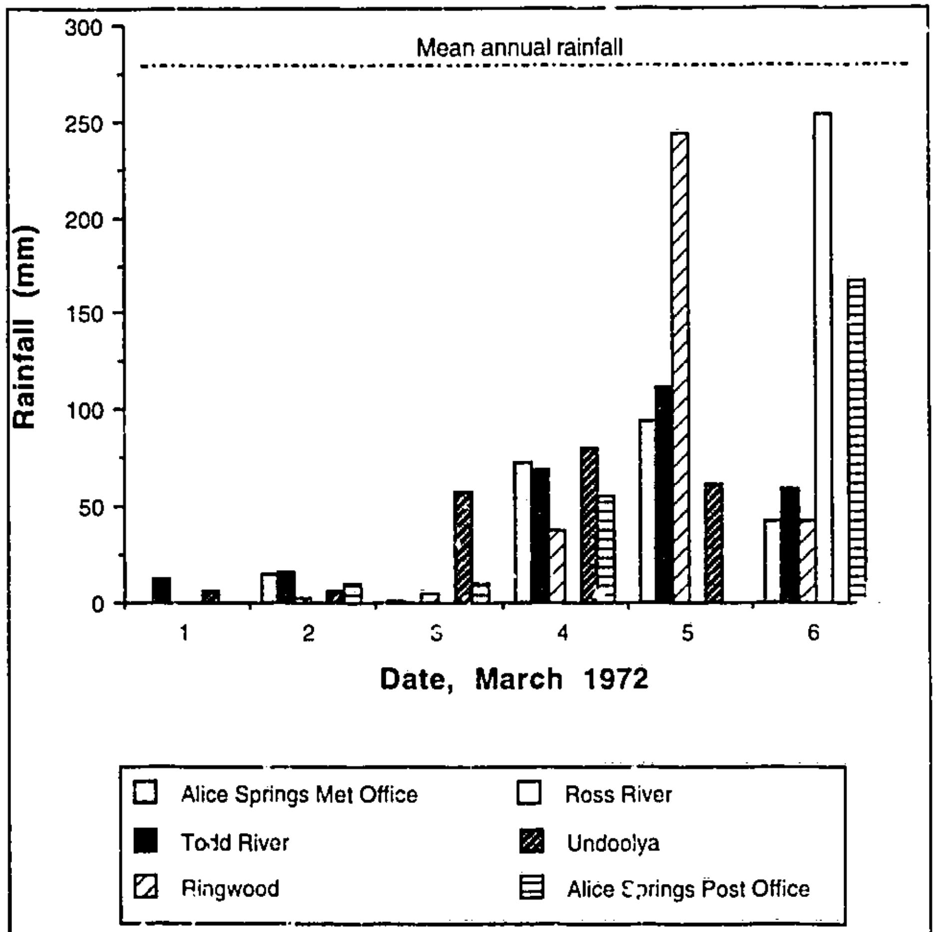


Figure 5.13. The variable nature of rainfall during early March, 1972.

5.3.2. Switching channels

Channel switching is important in the formation of relatively unconfined flood plains. The presence of wide (~50 m) back channels generally indicates former positions of the main channel, such as at the Ross River Striped site (Fig. 4.19), Shannon Bore site (Fig. 4.16) and Stud Bore 2 site (Fig. 4.15).

Three examples are given here of styles of channel switching in the Todd River. The first illustrates a site which is in a transient phase of switching. The second illustrates a reach where switching has occurred at least twice. The third is an example where the boundary topography of aeolian dunes influences the switching dynamics.

The first example is at the Stud Bore 2 site (Fig. 4.15) where large and small channel fills are important components in the flood plain morphostratigraphy. The study site is located on a meander bend, where prior to 1988 a single major channel meandered northwards and a relict chute crossed the meander plain to the south. Following the 1988 flood the flood chute widened (Appendix 4, Image 5, 5a) and when surveyed in late 1994 (Fig. 4.15) the channel thalwegs to the right and left of the central island were at approximately the same elevation and the channel widths similar (82 and 95 m). The river is inferred to have reoccupied a back channel and may switch the main flow to that location in the future, in which event the abandoned channel fill will be an important morphostratigraphic element.

The second example shows a series of changes between April 1950 and June 1988 at the sharp meander bend downstream from Mosquito Bore (Fig. 5.14) Downstream from the bend in 1950 (Fig. 5.14a) the channel is narrow and appears to have recently abandoned its left bank which has juvenile vegetation. By 1971 (Fig. 5.14b) the abandoned channel appears to be well vegetated and localised bank notching has occurred at the entrance of the abandoned channel. Prior to the 1988 flood (Fig. 5.14c) the meander migrated downstream forming a horse shoe bend, and it migrated even further after the 1988 flood (Fig. 5.14d). Since the 1995 flood the abandoned channel has been reoccupied by the main channel. The main channel appears to switch its location periodically by avulsion, perhaps assisted by notching at the abandoned channel entrance. Channel switching is often associated with channel incision, *e.g.*, at the Shannon Bore site (Fig. 4.16) the abandoned channel is 1.6 m above the main channel.

A schematic model based on observations at the Ross River Striped and Shannon Bore sites depicts flood plain morphostratigraphy in relatively unconfined reaches where channel switching is important (Fig 5.15). Initially, there is a single active main channel and a flood plain with surface channels (Fig. 5.15a). Avulsion, perhaps triggered by downstream movement of a sediment pulse, relocates the channel and the former channel aggrades perhaps with smaller flood channels (Fig. 5.15b). The abandoned channel continues to aggrade by overbank deposition of silt and clay and may occasionally carry [illegible] or gravel during higher energy flows which extend across the flood plain (Fig. 5.15c). A large flood may incise the abandoned channel fill (Fig. 5.15d) and may initiate another switching event, or both channels may operate simultaneously. Vegetation may stabilise abandoned reaches and enhance sedimentation (cf. Fig. 5.14a where the darker tone indicates juvenile vegetation).

The third example of channel switching occurs where the channel drains into the northern Simpson Desert and is controlled by the dune field topography. The 6-9 m tall Pleistocene longitudinal dunes have indurated cores and mobile crests which act as high levees but are sometimes breached by lateral migration of the channel. Plate 5.5 shows the Todd channel meandering across a swale and eroding the right bank dune. Downstream from this photograph the channel has breached the dune and switched to the adjacent swale (Fig. 5.16a). The abandoned downstream reach is now inundated only during higher flows when mud and fine sand are deposited, and the adjacent swale is similarly backfilled (Figure 16a and b). Reaches of the Todd channel along the swales are incised and this is inferred to be the result of the headward retreat of a minor nick point following the switching of the channel to the relatively lower local base level in the adjacent swale. Figure 5.16b shows a schematic cross-section indicating the incised channel and thin mud deposit in the adjacent backfilled dune. Further downstream (Plate 5.6), channel switching by dune breaching and backfilling of swales is more widespread, resulting in an inverted trellised channel pattern which may rejoin the original thread a short distance downstream.

Channel switching in dune fields has been described only rarely, but Adams and Hollis (1988, in Richards *et al.*, 1993) reported similar patterns in the Hadejia-Nguru wetlands of northern Nigeria, where rivers flow between

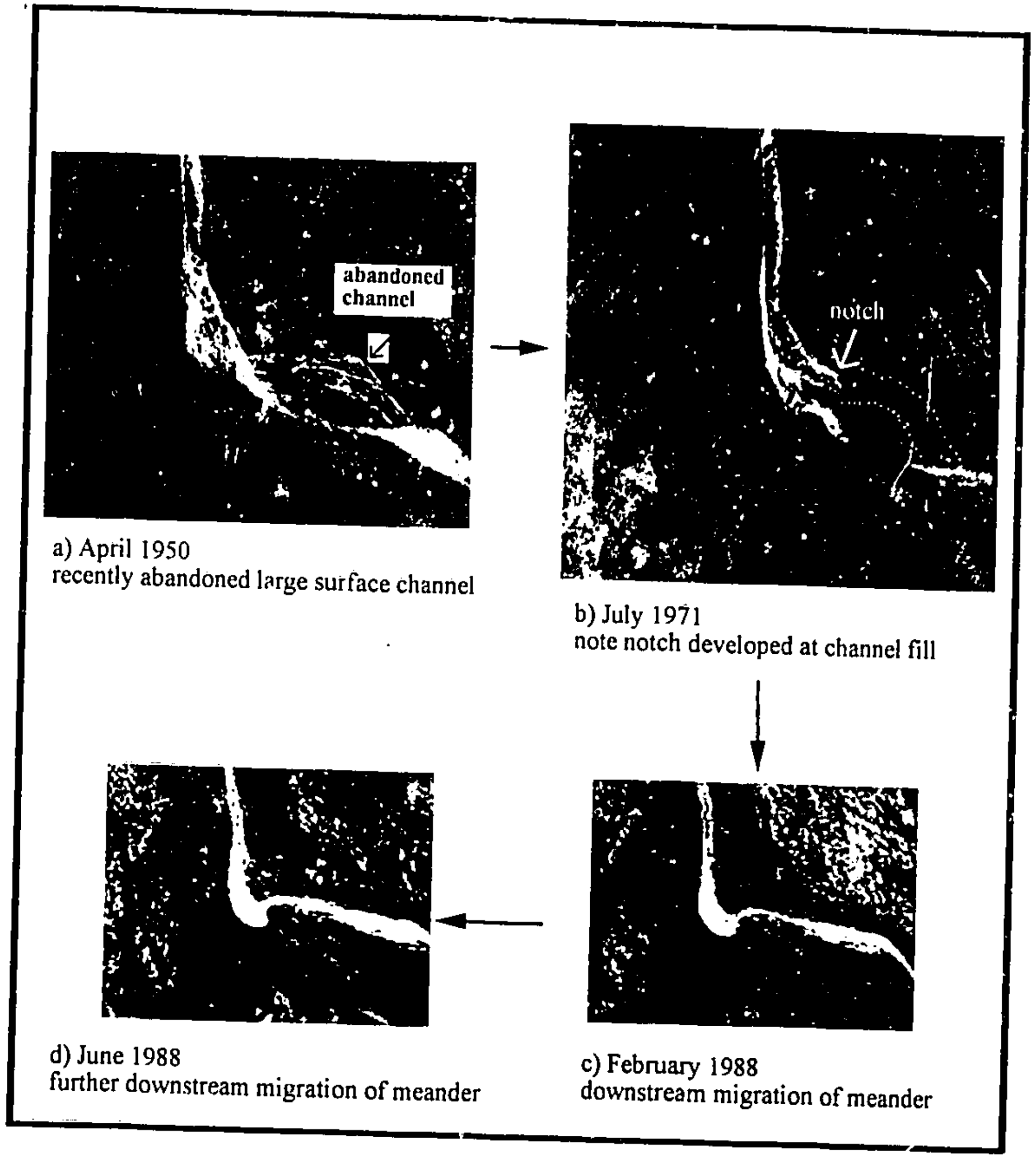


Figure 5.14. Time series of channel change south of Mosquito Bore showing channel avulsion and downstream migration of the meander bend.

Plate 5.5. Oblique aerial photo of Todd River meandering between longitudinal dunes in the northern Simpson Desert. Flow is towards the right of the photo. The adjacent left swale has been inundated by backflow as the channel has breached the left bank dune downstream.

Plate 5.6. Oblique aerial photo of the desert reach of the Todd River.

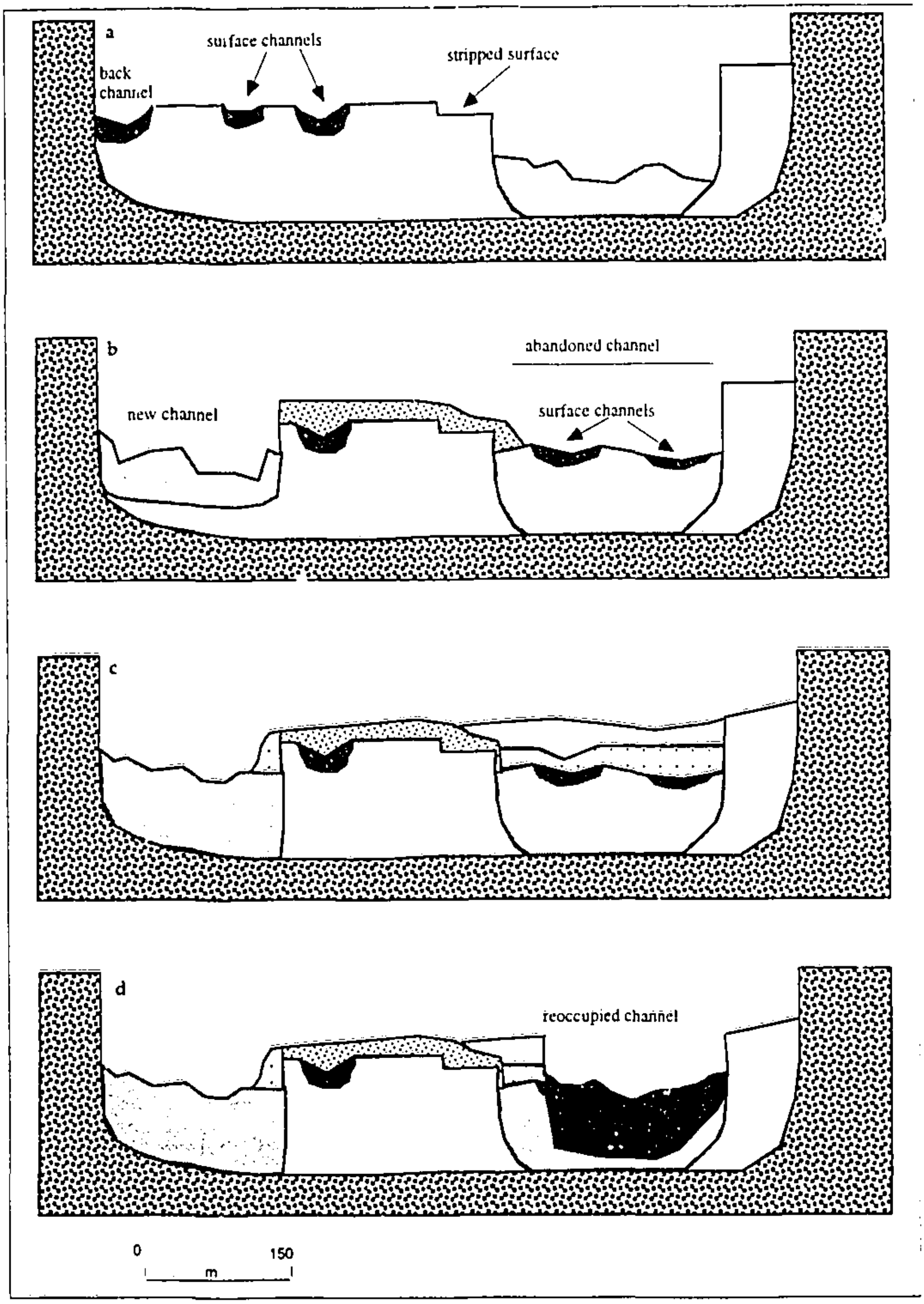


Figure 5.15. Schematic time series of channel switching in a relatively unconfined site.

a) The main channel is approximately 150 m wide and the wide flood plain is reworked by surface channels.

b) The back channel is incised and enlarged, the main channel is abandoned and inundated by surface channels during higher flows.

c) The continued aggradation of the abandoned channel and enlargement of the main channel.

d) Reoccupation of the abandoned channel. The channel may once again switch position or continue to drain through the two channels.

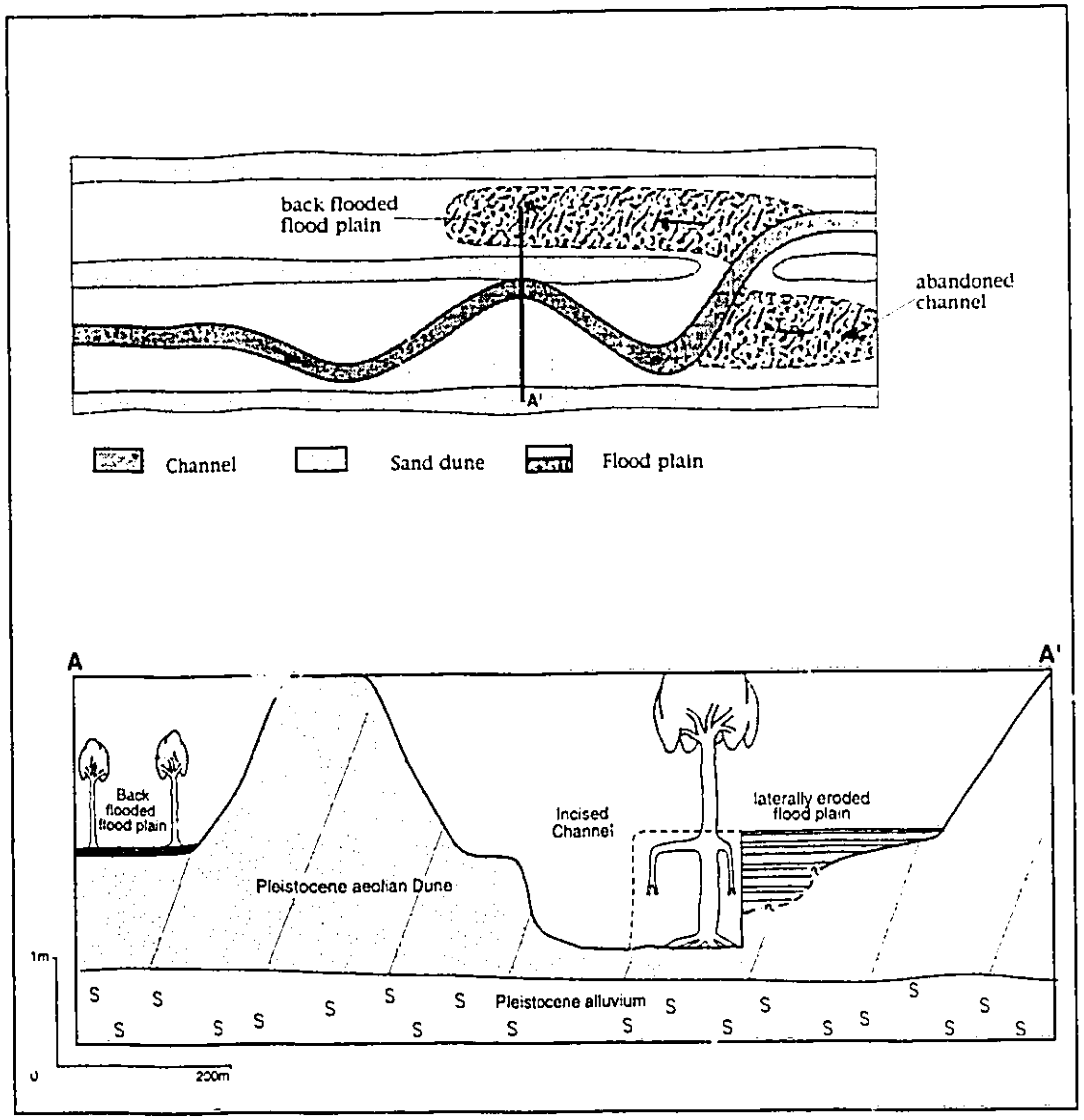


Figure 5.16.Flood plains and channel switching in the desert reach.
a) Schematic plan view of Todd River flowing in a dune swale (Plate 5.5).
The channel is incised in cemented Pleistocene swale sediments and meanders between the 6 m high dunes. Further downstream the channel has breached the dune most probably by lateral migration and avulsed to the adjacent dune swale. The abandoned downstream reach is now inundated only during higher flow and the adjacent swale is backfilled.
b) The schematic cross section shows two flood plain styles: insets and thin backwater deposits. The exposed tree roots indicate channel widening and incision, possibly due to the headward retreat of a local nick point initiated by channel switching.

degraded fossil dunes that are intermittently breached during floods. This autogenic change is caused by aggradation and made possible by the low elevation of the boundary dunes (Adams and Hollis, 1988). Channel switching processes are also important in wandering gravel-bed rivers (*e.g.* River Feshie, Scotland, Werritty and Ferguson, 1980)

5.3.3. Island formation

There are three ways in which islands form in the Todd River. Portions of flood plains can be excised by reincision of the surface flood channels (described above, section 5.2.1.1), braid bars can join to form a large compound bar, or single bars may simply grow by accretion aided by vegetation.

The large island in Figure 5.4 is an accreted compound bar formed by braid bar growth and lateral channel migration. The formation of secondary channels, the lateral migration of the channel and channel widening results in secondary channel abandonment and eventual attachment of the compound bar to the flood plain. Cadle and Cairncross, (1993) also found this a dominant process in the Permian sandy bedload Karoo sequence. Schumm and Litchy (1963) found that the flood plain construction in the Cimarron River in southwest Kansas was almost entirely by vertical accretion, and that island formation and the subsequent attachment of islands to the bank was by channel abandonment and accretion.

The colonisation of sediments by vegetation is important for the transition of mobile bar features to stable islands in the Todd River whereby vertical accretion and lee side deposition promote the attachment of islands to other islands and so develop flood plains. Others have also noted the importance of vegetation in promoting island and flood plain growth (Osterkamp and Costa, 1987; Thornes 1994 Sneh; 1983; Graf, 1988; Pickup *et al.*, 1988; Graeme and Dunkerley, 1993).

5.4. Models of Flood Plain Formation

5.4.1. Introduction

It is clear that flood plain morphostratigraphy and processes in the Todd River are complex, but there are two main types of flood plain: those formed in confined reaches and those formed in relatively unconfined reaches. The processes which form these types are now reviewed and compared with other models of flood plain formation.

Until relatively recently, it was considered that flood plains form mainly by lateral accretion (Wolman and Leopold 1957) with minor amounts of vertical accretion. This is reflected in the prominence of scroll bar deposits in migrating channel regimes (*e.g.* Schmudde, 1963; Blake and Ollier, 1971). While there is no doubt that lateral accretion is important, some flood plains are composed largely of vertical accretion deposits (Ritter *et al.*, 1973; Nanson and Young, 1981) and in general flood plains are composed of both lateral accretion and vertical accretion deposits (Nanson, 1986). Furthermore the flood plain may be complicated by crevasse splays, concave-bench accretion, channel infills, chute scour and fill, and the appending of channel islands to the flood plain (Schumm and Lichty, 1963; Burkham, 1972). The Todd River flood plain is of this complex type and is typical of variable discharge regimes where extensive flood plains may be destroyed by catastrophic floods and rebuilt by vertical accretion (Schumm and Lichty, 1963; Burkham, 1972; Nanson, 1986).

There have been three main approaches to classifying arid zone flood plains and they may be viewed as part of a downstream continuum, a flow magnitude continuum or as related to the degree of lateral confinement. Mabbutt (1977, 1986) proposed that flood plains vary laterally across the section and that there was a notable downstream trend: the upper plains sector, the lower plains sector and the floodout. Pickup (1991) approached flood plain classification from a scalar viewpoint detailing large-, meso- and small-scale flood plain processes. Sneh (1983), working on ephemeral streams in Israel, recognised the importance of the degree of confinement of the channel on the variability of arid zone flood plains. Like Mabbutt, he described flood plains at three points downstream which he termed confined, open and terminal flood plains. Confined and unconfined flood plains are clearly recognisable in the Todd system.

5.4.2. Confined flood plains

5.4.2.1. Introduction

Confined flood plains are located in high energy gorge reaches, entrenched into Pleistocene terraces or confined in short gaps through outlying strike ridges and between longitudinal dunes of the northern Simpson Desert. In

confined reaches the channel is single thread, generally straight and bounded by a markedly stepped flood plain with well developed lower benches in the channel. The most effective agents of flood plain destruction are channel widening, surface channel scour and swirl pit scour although the latter two are minor relative to channel widening. The principal constructive processes are the emplacement of insets, veneer deposition and surface channel accretion. Flood plain stores in confined reaches are shallow, averaging ~3 m deep; average total flood plain width is ~300 m and channel width is ~70-100 m. The lateral confinement of the channel inhibits the lateral migration and/or switching of the channel, and enhances the destructive potential of high magnitude events so that flood plain stores are relatively short lived. The following is a three stage model for the development of flood plains in confined reaches (Fig. 5.17).

5.4.2.2. The model

<u>Stage 1: Catastrophic erosion</u>

A high magnitude flood removes the entire valley bottom sequence (Fig. 5.17a) owing to the confinement of high energy flows. The principal effects are channel widening, and flood plain stripping where erosion is assumed to concentrate initially along surface channels. The flood may deposit a thin sheet of gravelly sand.

<u>Stage 2: Channel narrowing</u>

The next stage (Fig. 5.17b) is channel narrowing. It is expected that the extreme flood is followed by a series of ephemeral low to moderate flows which occupy and incise a narrow section of the flood channel, while floodwaters spread across the broad flood channel depositing overbank sediments.

<u>Stage 3: Construction and destruction of flood plains</u>

As the flood plain redevelops, vertical accretion will fluctuate, due in part to the arrival of sediment waves from upstream. At the local scale channel aggradation, bank erosion, stripping, formation of surface channels and the deposition of insets and veneer alternate with one another (Fig. 5.17c). The channel and flood plain will continue to form stepped morphology and complex morphostratigraphy and may be at any stage of recovery when the next high magnitude flood occurs. In this way the complex morphostratigraphy of confined flood plains is closely related to the timing, spacing and magnitude of ephemeral flows.

5.4.3. Relatively unconfined flood plains

5.4.3.1. Introduction

Relatively unconfined flood plains are located in wide gorge reaches, set within paleoflood terraces and located downstream of confluences and in channel beads where flood plains may be 500 to 1,000 m wide. The multiple channels are separated by high, well vegetated, stable islands and braid bars. The islands are typically diamond shaped with the long axis trending parallel or oblique to the flow. Some of the sub channels between islands and on the surface are perched above the channel bed and are only occupied during high stage flow. These relatively unconfined sites are similar to Nanson and Knighton's (1996) Type 5: gravel-dominated, laterally active anabranching rivers and to the wandering gravel-bed rivers that Church (1983) and Desloges and Church (1989) describe. The dominant channel commonly braid (Carson's (1984) wandering type 2 system) and there can be vigorous lateral activity.

Flood plain morphostratigraphy differs from more confined locations in two important ways. Owing to the presence of multiple channels, flood plains include broad islands and their morphostratigraphy includes large scale channel fills. Average fill depth is similar to confined flood plains (3.5 m) but average total width is greater at 560 m. In relatively unconfined flood plains, processes that operate laterally are more important than those operating vertically. Again the evolution of the flood plain can be seen in terms of a three stage model.

5.4.3.2. The model

Two possible scenarios are illustrated for relatively unconfined reaches; I applies when a large pulse of sediment moves downstream to a relatively unconfined site following stripping (Fig. 5.18) and II, when minor channel aggradation follows stripping (Fig. 5.19).

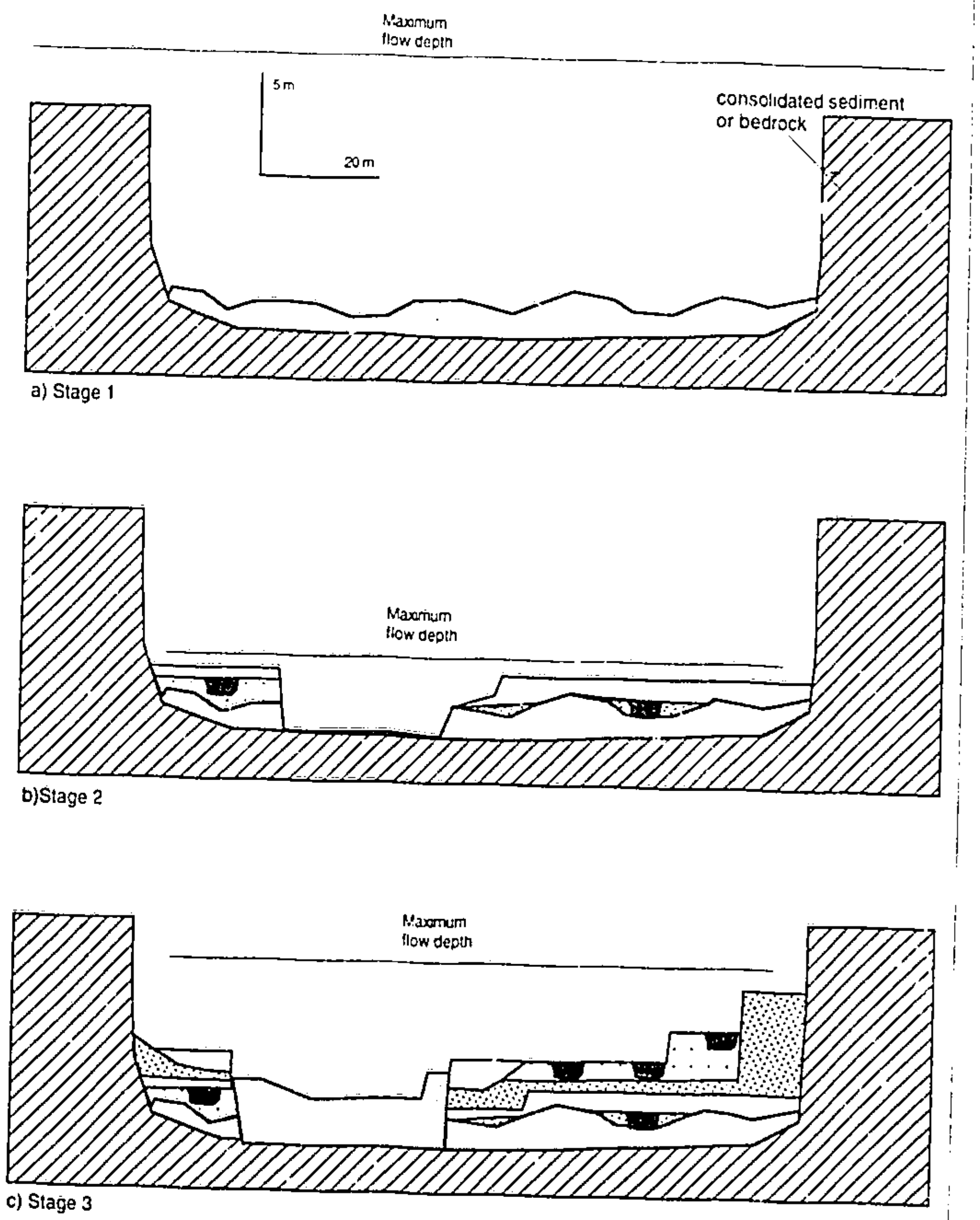


Figure 5.17. Confined flood plains.
a) Stage 1, Catastrophic erosion
b) Stage 2, Channel narrowing
c) Stage 3, Construction and destruction of flood plains

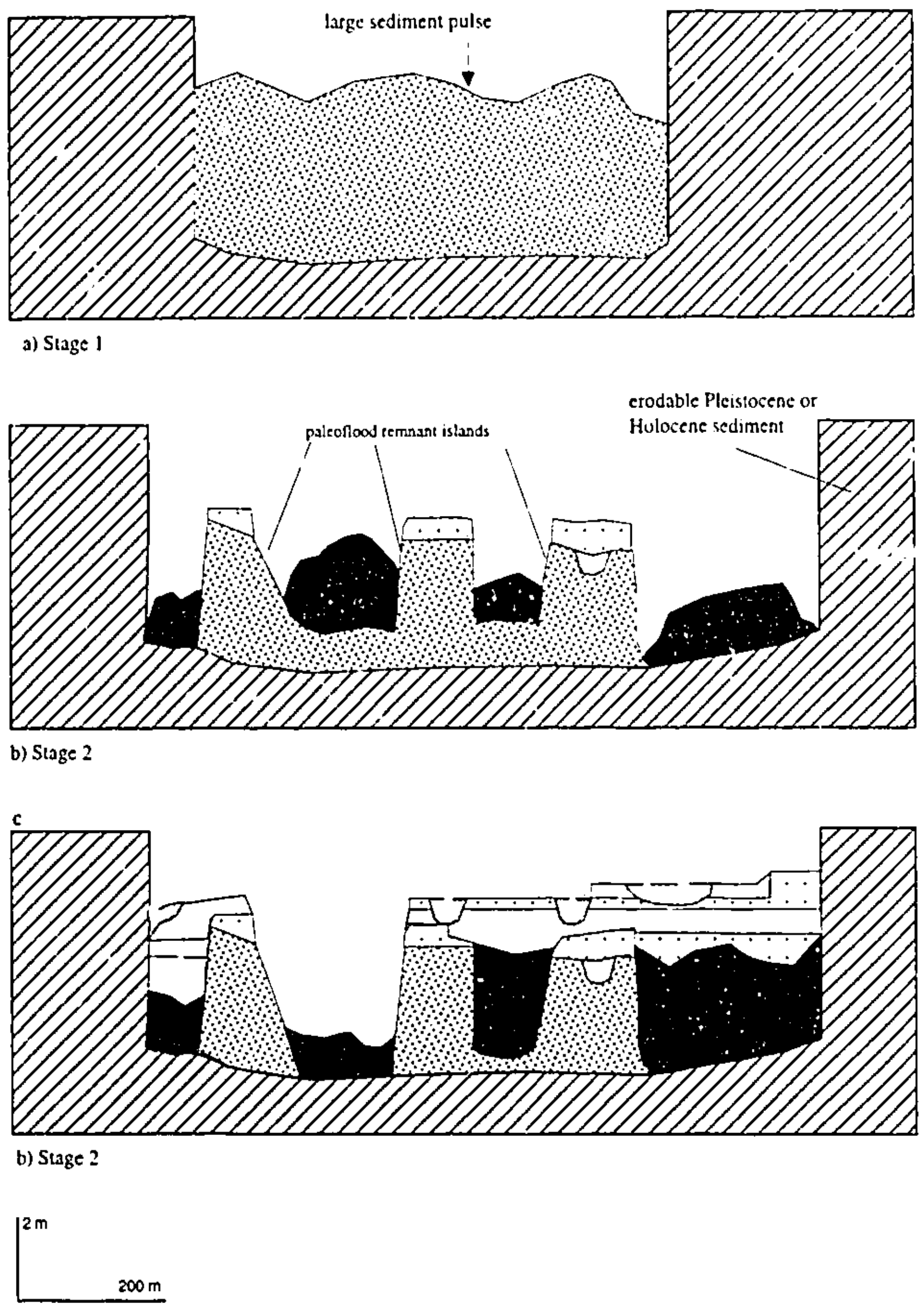


Figure 5.18. Relatively unconfined flood plains I
a) Stage 1, Catastrophic erosion and fill
b) Stage 2, Incision
c) Stage 3, Channel abandonment

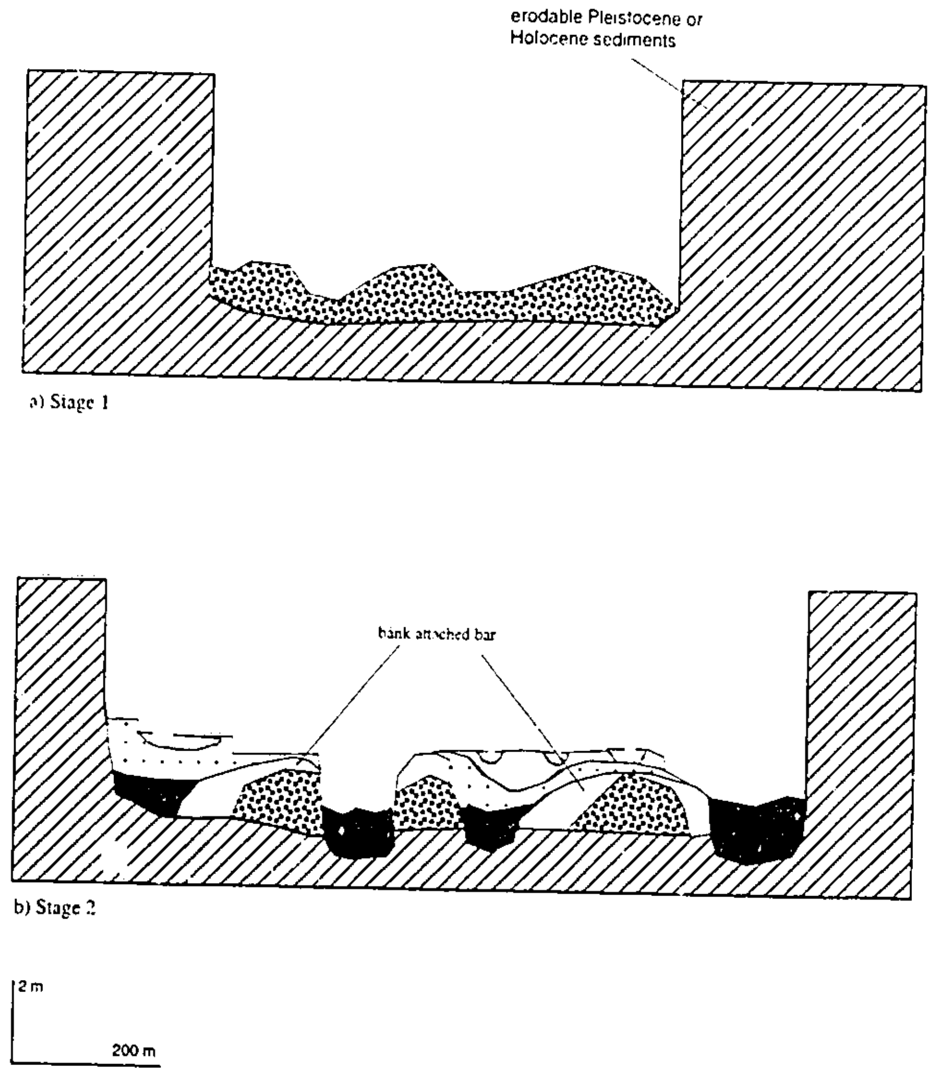


Figure 5.19. Relatively unconfined flood plains II.
a) Stage 1, Catastrophic erosion and minor fill
b) Stage 2, development of braid bars, channel widening and coalescing of islands to form bank-attached flood plains.

Stage 1: Catastrophic erosion and fill

Similar to stage 1 of the confined flood plain model, the effect of a large magnitude flood is to remove the valley bottom sequence within the channel train. This sediment may be entirely removed and if followed by the arrival by a sediment pulse from upstream, the channel may substantially fill with fine gravel in medium to coarse sand (Fig. 5.18a).

Stage 2: Incision

During this stage (Fig. 5.17b) the fill sediments dominate the processes in two ways. Firstly the remnant flood morphology dictates the position of the new narrower channel and secondly, the increase in channel gradient due to aggradation initiates incision of the channel fill along single or multiple threads. There is a concomitant phase of flood plain aggradation on perched remnants of the flood fill. During subsequent flows the channel bank is eroded by flow deflection around islands and sediment is moved from the banks to the channel and temporarily stored in large bars (Fig. 5.18b). These bars may become attached to the islands and are an important component of island formation (Fig. 5.19b). The deflection of flow around these bars increases bank erosion. Channel switching during high magnitude flows relocates the position of the main channel and creates a series of relatively stable vegetated islands separated by a multiple thread channel. An example of a site at this stage is the Anabranching site (Fig. 4.18).

Stage 3: Channel abandonment

A sequence of reduced flows may result in multiple channel abandonment and filling and lead to a phase of flood plain aggradation. The channel becomes a single thread with large surface channel as the flood plains become bank attached (Fig. 5.18c, 5.19b). These channels may be exploited at a later stage as higher magnitude flows reincise the surface channels. The evolution of relatively unconfined flood plains may alternate between stage 2 and 3 before the next catastrophic flood removes all or most of the alluvial fill. As the channel continues to widen the likelihood of complete cleanout decreases as wide channel sections (or beads) become areas of backflow preserving remnants and enhancing deposition. An example of a site at this stage is Expansion Scour site (Fig. 4.30).

5.4.3.3. Discussion

The morphodynamics of the Todd River flood plains indicate that their morphology does not represent the conventional flood plain/terrace sequence. Rather, the river has the morphological signature of an ephemeral, variable flow regime. Similar models presented by Schick (1974) and Nanson (1986) highlight the importance of event-based change and the impact of catastrophic floods. These will now be discussed in the light of confined flood plains in the Todd catchment.

Schick (1974) developed a conceptual model for the Nahal Yael research watershed (10 km^2) in southern Israel where he viewed valley floor aggradation operating quasi-continuously during relatively frequent minor floods. This pattern is interrupted by low frequency 'superfloods' which erode much of the channel and flood plain but leave behind remnant terraces at higher elevations. These are equivalent to the surface flood plain remnants in the Todd River flood plains (Fig. 4.8, 7). The period after the superflood is one where the flood plain and channel gradually aggrade until they overtop and bury the superflood terrace (buried flood plain remnants, Fig. 4.8, 6). Schick emphasised three aspects to this process. Firstly, terrace sequences do not correlate within and between desert drainage basins, chiefly because the flow events vary in space and time. Secondly, terraces do not reflect climatic changes unless the occurrence of superfloods is the response to climatic changes. Thirdly, the number of terraces is a reflection of both the magnitude and frequency of superfloods and the pattern of events between them. Schick's model is similar to the flood plain processes described for confined reaches of the Todd catchment.

Nanson's (1986) flood plain model for partially confined flood plains along high energy coastal rivers of NSW describes flood plains which form episodically by vertical accretion over periods of hundreds or thousands of years, following which catastrophic erosion by a single large flood or series of more moderate floods strips the flood plain to a basal lag deposit from which it slowly reforms (Fig. 5.7). He suggested that the periodic destruction resulted from progressive development of large levee banks and flood plain back channels. As the levee and flood plains vertically aggrade, overbank flow is displaced to the main channel and back channel with the resulting concentration of erosion energy. Eventually the channel boundary and flood plain are scoured by high flows which exceed erosion thresholds. Large back channels and surface channels are inferred to operate in the same manner

during high energy flows in the Todd River. Nanson (1986) noted that the four floods which catastrophically stripped the NSW flood plains had recurrence intervals of <9 yr and suggested that their combined effect was more important than their individual magnitudes suggest. In his model the progressive confinement of the channel increases the boundary shear stress of minor flow events. This is not applicable to the Todd where the boundary sediments are erodible and flood plain stripping occurs during high magnitude events.

Relatively unconfined flood plains are similar to flood plains on braided and anabranching rivers. It is not widely accepted that braided rivers readily form flood plains (Melton, 1936; Bridge, 1985). Graf (1988) describes braided streams as occupying the entire available space between low terraces, leaving no room for horizontal surfaces that are activated by present regime processes. However, Reinfelds and Nanson (1993) have identified processes of braided river flood plain formation in New Zealand which are similar to those described on the Todd River. They list three mechanisms which operate independently to form flood plains in the Waimakariri River in New Zealand: river bed abandonment, river bed aggradation and localised incision along primary and secondary channels.

In discussing compound channels, Graf (1988) noted that during the transition from braided to meandering patterns, many of the braid channels become infilled so that a new flood plain develops between the banks of the original braided channel and the banks of the new meandering or low flow channel. The surface adjacent to the new meandering channel becomes the flood plain and the meandering channel migrates across the floor of the braided system, reworking the deposits. While in principle this is applicable to the Todd River in that many of the flood plains have a core of flood deposited sediment, the channel tends to incise and switch rather than meander.

5.5. Conclusion

Flood plain construction and destruction is not synchronous through the catchment. In a manner similar to the findings of Nanson (1986) the flood plains are disjunct and have variable growth rates. During a single flood, erosion or deposition along a reach is a function of the local energy gradient and degree of confinement. The residence time of flood plain sediment varies substantially and so too do processes of mobilisation and deposition. Sediment pulses are generated during high magnitude flows and redeposited a short distance downstream with only minor sediment delivery at the terminus. This differs from rivers such as the Sepik in PNG which moves a high sediment load efficiently through to its delta (Chappell, 1993). The spatial variability in flood plain formation in the Todd River is reminiscent of Schumm's (1973) complex-response and critical threshold model of landscape change.

The variability of flood plains has implications for identifying ephemeral systems in the rock record. Downstream variation from single thread channels to multiple anabranching channels may lead to remarkably different stratigraphies. Thus, in a single system, variability must not be mistaken for regime shifts or climate change, as it is the autogenic response to ephemeral flow, sediment supply and degree of lateral confinement. The complexity of these flood plains indicates that paleoenvironmental reconstructive techniques using, for example, paleochannel dimensions or rates of flood plain aggradation (*e.g.*, Nanson, *et al.*, 1995) should not be attempted for this drainage basin without extensive stratigraphic interpretation. Larger channel dimensions or high rates of flood plain aggradation may simply be the geomorphic response of arid flood plains to high magnitude floods.

CHAPTER 6

Paleoflood Geomorphology

6.1. Introduction

This chapter presents the field observations of the Todd River paleoflood landscape. In the preface to the edited volume 'Catastrophic Flooding' Mayer and Nash (1987) define catastrophic flooding as:

> flooding of high magnitude and low frequency, devastating floods, or floods that result in significant changes in stream channel or stream valley characteristics.

The use of the term paleoflood here refers to the past occurrence of floods of high magnitude whose depositional and erosional effects remain visible in the landscape today. In the Todd catchment, paleofloods were greater in magnitude than the largest recorded floods.

Paleoflood landforms are difficult to recognise on the ground due to their subdued morphology. However, the coarse texture and extent of paleoflood sediments, vegetation patterns and the juxtaposition with contrasting fine red aeolian sand assist in the identification, and the landforms are easily recognisable on enhanced TM satellite images which are shown in Appendix 4.

Subtle differences in vegetation were used to map limits of deposition where morphology was not well defined. Flood deposits less than about 400 years old often have a higher density of Coolabahs (Kimber, 1996) whereas older flood deposits are populated with corkwood trees (*Hakea suberea*). Certain woody species, in particular whitewood trees (*Atalaya hemiglauca*), provide a good basis for distinguishing between slack water deposits (SWD) of different age and depth of surface layers in the Finke Gorge (Pickup, *pers. comm.*) where the abundance and height of whitewood and corkwood trees increases either with age of the surface sediment or a thinning of the 1974 flood deposits. Erect kerosene grass (*Aristida holanthera*), groundsel (*Senecio gregorii*), mallows (*Sida* sp) and buffel grass (*Cenchrus ciliaris*) appear to be insensitive to deposit age (Pickup, *pers. comm.*).

Nineteen cross sections and three long profiles of paleoflood complexes were surveyed using EDM and theodolite. Sixty-seven stratigraphic sections were recorded at river bank exposures, excavation pits and auger holes. Excavations using a 'bob cat' were undertaken at five sites. Aerial photo interpretations were undertaken by K. Fitchett and are presented in Figures 6.17 and 6.28. Sediment texture parameters are presented in Table 6.2 and Figure 6.22, paleoflood gradients in Table 6.7, paleoflood radiocarbon age estimates are in Table 6.3 OSL age estimates in Table 6.4 and site location maps are in Fig 6.2-6.16. All Figures, Tables and Plates are at the end of the chapter. The correlation of paleoflood surfaces is schematically drawn in Figure 6.56 and summarised in Table 6.8. Legends of symbols for morphostratigraphic figures (Fig. 6.57) and flood plain profiles (Fig. 6.58), on A3 pullouts, follow the Plates.

Geomorphological maps of paleoflood deposits of the Todd catchment are presented in Figures 6.1 a, b, c and d. The tendency for paleofloods to develop distributaries and splays has resulted in many discrete channel complexes of varying dimensions The mosaic of channel complexes has been subdivided into five principal systems: Heavitree/Undoolya, Ross, Giles Piedmont, Mosquito Bore Distributary and the Eastern Systems. The following section describes these paleoflood complexes in order of distance downstream and reviews previous research. The location of the sites is indicated on Figures 6.2-6.16.

6.2. Heavitree/Undoolya Paleoflood Complex

The Todd River drains through Heavitree gap onto a low gradient (.0024 m/m) piedmont fan system (Fig. 6.1b). The modern channel drains to the east of the fan into an intermediate floodout. Paleochannels across the fan surface indicate two prominent courses of the Todd River and its associated distributaries (Fig. 6.17). The more westerly course is associated with the position of the Todd flowing through the gap at Mt. Blatherskite. This appears to have been the penultimate course of the Todd until a flood in 1880 caused the channel to move to its present location (Wolley, 1966).

Litchfield (1969) identified three alluvial surfaces on the Heavitree floodout (see Figure 6.18a). He considered the oldest of these surfaces, the Lower Stuart surface, characterised by red earths, to date back to the Last Glacial Maximum (LGM) but Pickup (1991) has since correlated this surface with similar deposits at the Ross River dated at greater than 59 ka on the basis of a

saturated TL sample (Patton *et al.*, 1993) (Table 6.1). In this study it is referred to as the 'cemented Pleistocene alluvium'. The Lower Stuart surface passes beneath the Heavitree fan and deposits include gravel sheets and flood plain sandy alluvium with superimposed aeolian sand (Litchfield, 1969; Fig. 6.18b) which relate to a phase of vigorous fluvial activity, aggrading channels and numerous distributary systems. The end of the Lower Stuart phase was marked by aeolian activity represented by a 3 m high dune overlying the fluvial gravel in Litchfield's section A-A' (Fig. 6.18b).

The second surface referred to by Litchfield is the Upper Stuart surface (brown earths) which he considered to have formed in the late glacial and postglacial period. Pickup (1991) places it between 10,000 and >59,000 (Table 6.1). It is exposed as the surface of a low gravel mound in the upper fan region and passes beneath the younger Amoonguna surface (Litchfield, 1969). Close to the main channel (Fig. 6.18b) Upper Stuart deposits are over 5 m thick and form coarse gravel sheets up to 2 m thick and 1 km wide to the west beneath a sandy mantle. This was the last widespread phase of fan aggradation which eroded the Lower Stuart surface by 1 m (Litchfield, 1969).

The Amoonguna is the third and youngest surface recognised by Litchfield (Fig. 6.18a) and Pickup (1991) suggests an age of <10,000 with most deposits dating to <2,000 BP. Litchfield describes Amoonguna deposits as stratified gravel and sand flanking the present channel and its distributaries, occupying most of the head of the fan. The gravel becomes finer down the fan and grades to sandy fingers and finally to a network of silty sheets beyond the distributary termini, which continue for several kilometres and merge with surrounding floodouts of lesser creeks. The pattern of sedimentation indicates a shifting channel pattern (Litchfield, 1969).

Pickup (1991) described the large scale fluvial forms on the Emily Creek low gradient (.0048 m/m) fan to the east of the Heavitree fan (Fig. 6.19) The Undoolya bar field is 5 km long and 2.5 km wide and consists of a number of linear bars 2 m high and several kilometres long separated by swales 100-200 m wide. The spectral reflectance of the Undoolya bar field matches that of Litchfield's (1969) Lower Stuart surface (see Appendix 4, Image 2 for comparison) and is inferred to be part of that older unit.

6.2.1. Heavitree Gap

During road construction upstream of Heavitree Gap a trench (2.4 m deep and 41 m long) was excavated which revealed a lower cemented Pleistocene sand and gravel unit which correlates with the Lower Stuart deposits (Litchfield, 1969) and has an erosional upper boundary with scours 30 cm deep and 1.1 m long overlain by an upward fining sequence of sand and mud.

A 14 m core was drilled by the Power and Water Authority in Alice Springs in 1994, 100m downstream and 40 m east of the trench (site 2, Fig. 6.2) and was logged by the author (Fig. 6.20). No sedimentary structures were seen but the core varied in texture and colour (Plate 6.1). Basal Unit A overlies bedrock, is 2 m of grey/white sand with some pebble and silty layers, and is interpreted as a low energy channel deposit. Unit B is comprised of coarse clast supported gravel, 3.2 m thick and is interpreted as a high energy channel deposit. Contact with Unit C is sharp; Unit C is composed of silt and sand and is interpreted as a flood plain or backwater deposit created by an abandoned channel. Unit D is coarse pebbly grey and red sand 2.4 m thick, inferred to be a channel deposit. The uppermost unit (E) is a red and brown sandy silt and silt deposit (>2 m), probably a flood plain or backwater deposit. The high energy, coarse gravel sedimentary facies (Unit B) may correlate with Litchfield's Upper Stuart surface.

6.2.2. Heavitree ripple field

Pickup (1991) identified a large ripple field 1 km wide and 5 km long, at the end of distributaries which can be traced back to Heavitree Gap (west of the airport (Fig. 6.17). The site partly overlaps Litchfield's mapped area and appears to correspond with the Amoonguna surface. The area is comprised of transverse ripples >1 m high composed of red poorly sorted, gravelly sand with pale fine sand and silt in the swales.

A shallow pit (40 cm) was excavated on the surface of one of the bars (Fig. 6.3) and two radiocarbon and one OSL sample taken. The OSL sample ($ANU_{OD}170$) returned an age of 1030±420 BP and the first radiocarbon sample (ANU 8833) from a 4 cm layer of charcoal returned a very similar age of 1196-691 cal BP. A second ^{14}C sample was taken from between 9 to 30 cm in the profile returned a calibrated age of AD 1955, which is inferred to be

erroneous probably because the sample consisted of finely dispersed charcoal.

6.2.3. Jessie Quarry

A large source-bordering climbing dune east of Jessie Gap (23° 45', 134° 02') (Plate 6.2, Fig. 6.4), formed of sand derived from Jessie Creek, climbs the flank of the nearby range. The upper section of the dune is beige, 1 to 3 m deep and is underlain by white sand 30 m deep (R. Morley, *pers. comm.*). Stoss-face planar lamination and remnants of stumps in growth position are well preserved. The white dune is underlain by a red dune, which is laterally truncated by fluvial and colluvial deposits from an adjacent small steep catchment. The red aeolian sand resembles dated late Pleistocene dunes in the western Simpson Desert described by Nanson *et al.* (1995).

Charcoal from the white dune was sampled for radiocarbon analysis at the dune crest and at a shallow hearth, 75-88 cm below the surface on the dune flank. Charcoal from 86 cm at the crest site returned an age of 1874-886 cal BP (ANU 8835), and the hearth gave an age of 3087-2775 cal BP (ANU 8836). Thus the dune was active about 1000-3000 years ago. The construction of the late Holocene white dune is attributed to increased fluvial activity in Jessie Creek rather than increased aridity because the preservation of stumps in the dune indicate that there was sufficient available moisture for vegetation growth, which would have stabilised the dune and enhanced its aggradation rate by trapping sediment.

The Todd, Emily and Jessie piedmont systems are a mosaic of different alluvial and aeolian units dating probably back as far as 100 ka. They contain a range of paleoflood landforms and have been previously described by Litchfield (1969) and Pickup (1991). The fan surfaces are currently drained by ephemeral rivers which rapidly disintegrate into floodouts downstream. Paleoflood landforms are associated with land surfaces of different ages. There are two absolute ages (~1,000 BP) for paleoflood deposits associated with the Amoonguna surface and Pickup (1991) has identified large paleoflood forms on what appears to be the Lower Stuart surface in the Undoolya bar field.

6.3. Ross River Paleoflood Complex

The Ross River paleoflood complex which extends along the Ross River downstream of the ranges and piedmont fan (Fig. 6.1 c, Plate 6.3) was described by Pickup (1991) and Patton et al.(1993). These authors identify three paleoflood surfaces (Fig. 6.23): a large paleoflood complex (Ross A), a ripple field (in Ross C) and a large-scale bar system (in Ross B). Their findings will be discussed in conjunction with the observations described below.

6.3.1. Ross River Gorge

At Shannon Bore site a high mounded slack water deposit is located on the right bank 5 m above the channel bed, in the mouth of a small tributary catchment (Plate 6.4, Fig. 6.5, Fig. 6.21). A 4 m hole was augured into the surface and sediments taken for analysis. The sediment was a poorly sorted fine red sand throughout (2.08o) and is inferred to have been deposited during a single event. No event of greater magnitude has occurred since its deposition as no younger and higher deposits were found at this site. The deposit is currently being dissected by local drainage. One sample was dated by OSL from a depth of 2 m in the deposit and returned an age of 8350±950 L.' ($ANU_{OD}171$).

6.3.2. Ross A

A large paleoflood channel system immediately west and south east of the present Ross River was described by Patton *et al.* (1993; Fig. 6.23) as a braided channel which divides around longitudinal bars. Individual channels attain a maximum width of 50 m and the total channel width ranges to 800 m. The bars and channels have low relief (Fig. 6.24). The gradient of the paleochannel is inferred to be steeper than the modern channel as it is 2.4 m above the modern channel in the north and 0.3 m below in the south; sediment in the paleochannel and bars was 100% sand (Patton *et al.*, 1993). Based on the small number of sedimentary units present in exposures of the paleochannel alluvium, Patton *et al.* (1993) infer that the Ross A deposits represent five or fewer floods younger than 1535±85 cal BP (GX-15173), the latest occurring about 695±60 yr BP (GX-14022). Unfortunately detailed descriptions of sample locations and sedimentary sections are not available.

The Ross A paleochannel deposits were examined in a 5.2 m channel exposure along the left bank of the Todd 3.5 km downstream of the confluence (Fig. 6.23, 6.25) The deposit is a sequence of cobble, gravel, sand and compact mud beds. Charcoal from four horizons was dated by radiocarbon. The 2.7 m thick basal bed is a horizontally layered cobble and gravelly sand topped by structureless coarse and fine sand. Two charcoal samples from the finer sediments at the top of the gravel, which is interpreted as a paleoflood channel deposit, returned ages of ,67-790 cal BP (ANU 10063) and 985-724 cal BP (ANU 10062). The two overlying sand/silt couplets were not dated. An overlying 34 cm thick flood couplet containing mud balls is dated to 656-522 cal BP (ANU 1006[illegible]) and is overlain by a structureless medium gravel and mud flood couplet dated to 564-430 cal BP (ANU 10060). The top bed is undated but is similar to the dated sediments in weathering and colour and is assumed to be of similar age. The presence of erosional boundaries, root remnants and major textural contrasts between beds was used to infer a hiatus in deposition, which together with the chronology indicate that the Ross A deposits represent several (perhaps five) floods which occurred between ~1,000 and 500 yr BP. These results are consistent with those reported by Patton *et al.* (1993) and their youngest event (695±60 BP) may be equivalent to the 656-522 cal BP event dated here (Fig. 6.56, Table 6.8).

6.3.3. Ross B

This complex lies to the west and east of Ross A. It does not have pronounced channel form and has not been dated. Patton *et al.* (1993) describe it as an aeolian sheet (Fig 6.23); however satellite imagery (Appendix 4, Image 5a) indicates that the most easterly portion of this aeolian sheet is inundated by distributary channels from Ross A.

6.3.4. Ross C

Ross C extends from a gap to the east of the Ross River, eastward across the low gradient plain (.0013 m/m) and intersects with the Giles Creek alluvium 15 km downstream (Fig. 6.1 c). The southerly limit, near the Ross confluence, abuts against, and overlies the Pleistocene deposits on the flanks of the high bedrock ridge on the right bank of the Todd (Fig. 6.23; 6.26). The ripple field in the north west portion of Ross C was investigated by Patton *et al.* (1993) and extends from a water gap in the ranges, has a gradient of 0.0034 m/m, and expands downstream and curves to the east along the ranges (Fig. 6.23).

The ripple field is visible on aerial photographs and satellite images and is comprised of elongate low mounds composed of silty, very fine sand separated by strips of exposed paleosol manifest as light strips 100 m wide in the satellite imagery (Appendix 4 Image 5a). The average wavelength of the ripples is 122 m but decreases along the edge of the ripple field; the average height is 0.22 m. Patton *et al.* (1993) infer that the ripple field was formed by a shallow sheetflood that spread across the plain.

The stratigraphy of the Ross C channel deposits is exposed at the Stud Bore 2 site (Fig. 6.6, 6.26) The paleoflood sediments have the same elevation on both sides of the present channel indicating that the entire Ross C complex is syndepositional. The Ross C paleochannel deposits were examined in a 6.6 m left bank excavation (Fig. 6.27, Plate 6.5 a and b) which shows two units. The lower unit is a >5.2 m thick fining upward sequence from framework cobble, gravel and gravelly sand to a poorly sorted fine sand (2.01σ) and is overlain by a structureless, 1.6 m thick mud layer dated by radiocarbon and OSL. The mud bed indicates a high suspended sediment load and is inferred to be deposited in slow-moving water. The upper unit, inferred to be deposited during a separate flood, is represented by a 1.4 m sequence of structureless slightly gravelly red sand rich in carbonate. The age of the lower unit is estimated by OSL to be 12900±1800 BP ($ANU_{OD}161$) which is consistent with the degree of weathering (Plate 6.5a radiocarbon sample from this layer returned an age of 673-248 cal BP (ANU 10059) and is rejected because the charcoal was finely dispersed.

There have been at least two floods in the Ross C channel complex. Events 1 and 2 have been described above and a third event may be represented by the fine textured ripple field and/or the Shannon Bore slack water deposit in the Ross River Gorge (Fig. 6.56, Table 6.8).

6.3.4.1. Streamlined remnants

High (~2.3 m) streamlined remnant 'islands' occur on the floodout plain to the east of the modern Ross River. The remnant close to the confluence and south of Ross C is dissected by small channels (Fig. 6.23) which are inferred to have been formed during inundation of the remnant by high magnitude flows (Pickup, *pers. comm.*).

6.4. Giles Creek Piedmont Paleoflood Complex

The Giles Creek paleoflood complex drains south from Allua Gap to the northern Simpson Desert, through Wallaby Gap (Fig. 6.1 d and 6.28 a, b). Several flood tracts form wide braided channels and splay-like deposits with superimposed bars. The description of the Giles Creek paleoflood complex is subdivided into the Piedmont area and the Mosquito Bore Distributary.

The Giles Piedmont paleoflood complex reaches a maximum 8 km width, bordered on the west by the Ross C paleoflood complex and on the east by a sand sheet occupied by small ephemeral channels and floodouts. Mapping from aerial photographs (Fitchett, unpublished, Fig. 6.28a) indicates extensive channel patterns across the plain. The following is a description of five paleoflood complexes.

6.4.1. Giles A

Giles A extends from Allua Gap east of the modern Giles Creek to 25 km downstream (Fig. 6.1 d) and underlies more recent flood deposits. The braided, moderate gradient (.0022 m/m) complex is bordered to the east by an extensive sand sheet dissected by small ephemeral streams which have coarse textured longitudinal bars trending south west. The very poorly sorted, cemented, pale, carbonate nodule rich, medium sand (1.36σ) was exposed as the basal layer in a section through the Giles B surface (Fig. 6.29) and is estimated by OSL to be 26800±3000 BP ($ANU_{OD}16$ b). It is the oldest alluvial sample dated in this study and suggests that the Giles A deposits are coeval with Litchfield's Upper Stuart deposits. The most north-westerly section of the Giles A complex was included in a cross survey (Fig. 6.30) and is shown to abut a colluvial fan. It is dissected by shallow broad channels.

6.4.2. Giles B

Giles B is a 13 km long and 1 km wide gravel and sand splay which extends south east from the modern channel (Fig. 6.1 d) along a low gradient (.0002 m/m). The upstream reach is dissected by two wide (140 m, 90 m) shallow (0.6 m, 1 m) channels (Fig. 6.32). The downstream reach is superimposed by longitudinal gravel bars trending parallel to the larger splay. The bars range from 50 to 200 m wide and the swale width from 50 to 100 m (Fig. 6.33) often 1-2 m high. Plate 6.6 a and b indicate the relief and sediment texture of the Giles B longitudinal bars.

A 2 km long and 125 m wide bar with degraded transverse ripples superimposed on the surface was excavated (Fig. 6.7). Part of a larger bar complex which increases in elevation westwards, it has a well defined eastern side and less distinct western flank. A 20 m by 1.3 m transverse trench was excavated into the eastern side of the bar (Fig. 6.29, Plate 6.7a) and four stratigraphic units identified (Fig. 6.34) but no internal stratigraphy was preserved (Plate 6.7b). The horizontal basal layer (Giles A surface) which extends deeper than the trench floor is overlain by a poorly sorted fine sand (2.66ø) which dips approximately 15° along the bar flank (Fig. 6.34) and increases in thickness towards the surface. It is interpreted as a longitudinal bar deposited on a horizontal surface. This unit is draped by a thin (9 cm) moderately sorted, moderately indurated, fine sand layer (2.7ø) inclined parallel to the underlying bar. The top layer is a thick (47 cm) poorly sorted fine sand (2.41ø) with some gravel scattered on the winnowed surface, and is also inclined parallel to the underlying layers is estimated by OSL to be 9700±1500 BP (ANU_{OD}160a). The silty sand in the swale to the east of the bar may relate to subsequent lower magnitude flow(s) (Pickup *pers. comm.*). To summarise, the longitudinal bars on the Giles B paleoflood complex were deposited on the Giles A surface ~10 ka and are inferred to have accreted during three floods.

Giles C to Giles I were mapped from satellite images and included in cross sections (Figs 6.30-32) but were not examined in the field and are not described here. Giles J and K surfaces are too small to be shown on Figure 6.1d.

6.4.3. Giles J

The Giles J paleoflood deposit is incised in older Giles Creek paleoflood deposits (principally, Giles A, B and C) and is exposed along the modern Giles Creek banks as a cobble and gravel paleoflood channel fill (Fig. 6.30 to 6.32) which averages 350 m wide and >3.5 m deep (Table 6.5). At the Giles Creek 1 site a basal >1.6 m structureless cobble and coarse gravel layer of the 3.7 m channel fill is overlain by a fining upward sequence of alternating layers of dipping matrix-supported gravel and horizontally bedded coarse gravel. The second unit is a structureless medium and fine sand deposit. The lower channel fill deposit is inferred to have been deposited during a single event.

At least two events are identified in the Giles J fill at the Giles Creek 3 site (Fig. 6.32, 6.35). Event 1 at the base of the section is a >2.7 m thick cobble and coarse gravel layer overlain by a coarse and medium sand matrix supported gravel. Event 2 is identified as a structureless sand estimated by radiocarbon to be 3646-1933 cal BP (ANU 10068) and is overlain by an upward fining coarse gravel layer and an upward fining small gravel and silty sand layer. As these latter two events buried mature River Red Gums on the flood plain surface they are estimated to have occurred <150 yr BP.

6.4.4. Giles K

Giles K is located 3 km downstream from Giles Creek 3 site (Fig. 6.7). The stratigraphic section is drawn up in Figure 6.36 and the site is shown in Plates 6.8 and 6.9. Three stratigraphic units were identified (Fig. 6.36). Unit 1, composed of red cemented Pleistocene sand underlies the entire section (Plate 6.9) and is overlain by a 3 m thick unit (Unit 2) of alternating horizontal mud silt and fine sand layers perhaps deposited during six events. The fine texture sediments in Unit 2 are similar to those in the Ross C channel exposure and are inferred to have been deposited in a slow moving/stagnant water body. Unit 3 is a channel fill incised in Unit 2 and is composed of a fining upward coarse gravel to gravelly sand sequence, overlain by a 1 m thick structureless sand layer which extends across the plain. The channel fill is inferred to be an older Giles Creek which would have met with the Todd River 4 km upstream of the present confluence (Plate 6.8). No chronological data is available for this site but Unit 1 is similar to the Pleistocene sediment that is found throughout the catchment and is placed in Litchfield's Upper Stuart deposits and Units 2 and 3 are inferred to be older than Giles J unit (*i.e.* >3 ka) and it is possible that Unit 2 is the eastern extent of the Ross C complex.

6.4.5. Todd A

The Todd A surfaces are those which extend from the banks of the main channel but are higher than the modern flood plain. They have been identified in three locations and examined at a site in the Mosquito Bore area located 1.3 km downstream of Mosquito Bore (Fig. 6.1 d). The Mosquito Bore area is a key location in terms of the dynamics of paleoflood systems as flow was confined or diverted around the bedrock ridges. Three sites are described here, the distributary, the channel fill and the slack water deposit (SWD).

6.4.5.1. Site 1 The Todd A distributary

The Todd A paleoflood distributary flows southeast, exiting through a 40 m gap in the left bank bedrock wall of the modern channel across a wide plain with outcropping bedrock ridges and remnant streamlined aeolian deposits (Fig. 6.1 d, Plate 6.10, Appendix 4, Image 7a). A 550 m long profile (Fig. 6.8, 6.37) indicates a low elevation fine and medium sand SWD mound at the distributary intake followed downstream by a scour hole adjacent to the right bank bedrock wall that lines the first 150 m of the distributary and lies at shallow depth beneath the channel. Flood damaged vegetation and flood trash indicates the recent inundation of the distributary, most probably during the high flows of the 1970's. Approximately 500 m from the intake the distributary channel has a low width/depth ratio (1:0.08) (2.35 m deep and 28 m wide) which rapidly diminishes downstream forming a silt and mud multiple channel network (Plate 6.11), which floods out to a flat plain, reforms a channel when it abuts against a bedrock outcrop and eventually re-enters the Todd channel 6 km downstream from the intake, close to the Expansion Scour site.

A 3.7 m excavation of the right bank of the distributary (Fig. 6.39) indicates a vertical stack of flood deposits composed of poorly sorted fine sand beds, mud beds and flood couplets with a coarse 80 cm thick structureless pebble layer in mid section. Flood events were distinguished by textural contrast, roots, soil development and erosional boundaries. Two horizons in the sequence are dated by radiocarbon. The stratigraphically lower charcoal sample from a hearth returned an age of 463-340 cal BP (ANU 10065) and charcoal from the overlying layer returned an age of 511-312 cal BP (ANU 10064). The burial of (~150 year old) Coolabahs by >1 m of sediment (Plate 6.11) suggests the upper 275-378 cm of the profile accreted in the last 150 years.

6.4.5.2. Site 2, the channel fill

The structureless coarse gravel channel fill at the Mosquito Bore 2 site is similar in texture and morphostratigraphy to the Giles J channel fill identified 16.5 km upstream (Fig. 6.38). As it is likely that the Giles J fill extended from Giles Creek into the Todd River it is possible that remnants are preserved in pockets downstream of constrictions, such as the fill at the Mosquito Bore 2 site. The Mosquito Bore 2 channel fill is inferred to be coeval with the Giles J channel fill and placed at about 3,000 years old (Fig. 6.56).

6.4.5.3. Site 3, slack water deposits

Two slack water deposits are preserved in caves in the right bank dolomite ridge at the Mosquito Bore 1 site. The first is a thin layer (10 cm) of fine red sand, 4.7 m above the channel, which was dated by radiocarbon to 493-286 cal BP (ANU 9285; Fig. 6.40). A 1.1 m thick deposit in a lower cave (3.9 to 2.8 m above the thalweg) (Plate 6.12) contains a complex cut and fill, overlapping sequence of flood deposits preserved behind a roof-collapsed boulder (Fig. 6.40). It is composed of moderately well sorted very fine sand (3.20φ) with some angular roof fall fragments within and between deposits. Sedimentary structures were well preserved and included horizontal, inclined and oblique stratification (Plate 6.13). Deposits were differentiated from each other on the basis of the degree of bioturbation which stopped at sedimentary boundaries, the presence of angular rock fragments at sedimentary boundaries and the age determinations of the deposits. While there may be as many as twelve flood deposits, seven are identified with certainty on the basis of the above criteria and are labelled (i-vii) in Figure 6.40. Two of the three layers dated by radiocarbon returned modern ages and the third returned an age similar to the higher SWD at 507-313 cal BP (ANU 9657), which is also similar to the age of the Todd A distributary deposits. Assuming no change in bed elevation since 400 BP, the height of flood waters at the cave SWD site (4.7 m above the thalweg) is sufficient to inundate the Todd A distributary channel. The cave SWD record is similar in terms of the age of events and the number of events preserved in the distributary record.

One kilometre downstream a mound SWD is located in a bedrock embayment 6.25 m above the channel thalweg. An 3m hole was drilled through three sedimentary units to bedrock and revealed an upper 1.5 m thick aeolian well-sorted fine sand deposit underlain by a 90 cm thick SWD containing two flood units which overlie a pebbly sand (Fig. 6.38). The SWD elevation is similar to that of the distributary channel and the channel fill but because of the aeolian deposit is inferred to be related to the 3 ka channel fill.

6.5. Mosquito Bore Distributary Paleoflood Complex.

The Mosquito Bore Distributary (MBD) extends from the channel constriction at the Mosquito Bore 1 site south through Wallaby Gap and floods out into the northern Simpson Desert (Fig. 6.1 d). Jackson (1962) correctly inferred that the presence of the Todd River family of soils between Mt. Capitor Bore and

Wallaby Gap was related to a former course of the Todd. The paleoflood complex was examined along three reaches.

6.5.1. Reach 1, northern distributary plain

Reach 1 extends 10 km downstream from the sharp meander bend (Plate 6.14) to Mt. Capitor Bore (Fig. 6.9), is 2.5 km wide and bordered to the west by a colluvial and aeolian plain and to the east by a 70 m high bedrock ridge. The structureless sand matrix-supported gravel, braided channel is a convex-upward shape.

A 3 km long low alluvial bar extends from the meander bend to the colluvial fan upstream from Mt Capitor Bore and is bordered to the west by a fine sand and silt braid channel and a wider well-vegetated braid complex and to the east by colluvial fans. The convex morphology (Fig. 6.41) rises 70 cm above the surrounding alluvium and the winnowed surface has 10 cm high aeolian mounds. Four excavation pits along the crest of this longitudinal bar indicate a poorly sorted structureless gravelly silty sand between 60 cm and 150 cm thick overlying cemented Pleistocene sediment. An OSL sample from a pit in the bar crest returned an age of 6230±1370 BP ($ANU_{OD}162$) (Fig. 6.41).

6.5.2. Reach 2, Mt Capitor Bore to Wallaby Gap

Reach 2 of the Mosquito Bore Distributary extends 20 km between Mt Capitor Bore and Wallaby Gap (Figs. 6.1 d; 6.9; 6.10). Downstream from Mt Capitor Bore the paleoflood complex increases in width to a maximum of 4 km and has a gradient of .0016 m/m. It is bordered to the west by a laterally truncated aeolian dune field and to the east by the younger and lower elevation Mosquito Bore A paleoflood complex, described later. It is an expansion bar complex (Appendix 4, Image 8) crossed by braid channels, two of which support dense vegetation stands (Plate 6.15).

An excavation trench (MBD2, Fig. 6.9) on the 1.7 m high expansion bar complex (Fig. 6.42) indicates structureless, cemented poorly sorted medium sand (1.84ø) with some small pebbles. Relief on the expansion bar is low with braid bars rising 40 cm above braid channels in contrast to the 1.7 m deep and 135 m wide marginal channel. The morphological inflection on the eastern flank of the aeolian dune corresponds with the height of the paleoflood complex and is inferred to have been eroded by paleoflood flow. The paleoflood complex maintains a semi-consistent width of 3 km to the south splitting around a streamlined aeolian and clay pan remnant north of Wallaby Gap (6.10).

No slack water deposits were found at Wallaby Gap and a cross section across the gap indicates two possible entrants for paleoflood flow (Fig. 6.43). The 350 m wide eastern gap has a sand and silt fill of unknown depth but an Aboriginal rock painting on the left wall indicates the presence of a former waterhole at the gap entrance (Latz, *pers. comm.*); the narrower gap to the west is at a slightly lower elevation. Fifteen meters south of the gap a high streamlined fine red sand deposit may be a slack water deposit but is more likely to be an aeolian dune with a 3.5 m high step on the upstream side inferred to be channel-eroded (Fig. 6.44).

6.5.3. Reach 3, Camel Flat

Downstream from Wallaby Gap the paleoflood complex fans outwards to form an 8 km wide, low gradient (.00097 m/m) expansion bar complex (Fig. 6.1 d). The adjacent well sorted fine red sand aeolian dunes (6 m high) have been truncated by the flow (Appendix 4, Images 9, 9a). The moderately sorted deposit of fine sand (2.75ø) with small pebbles is arranged in longitudinal bars and shallow channels. The surveyed bar (MBD 3, Fig. 6.11) is 90 m wide and rises 3 m above the 70 m wide marginal channel and 2 m above the surface channel. (Fig. 6.45). The 1.6 m deep trench exposed an upward fining sequence from medium to coarse gravel in a medium sand matrix to fine gravel in a very coarse sand matrix and an OSL sample taken from a depth of 80 cm is estimated at 9920±1840 BP ($ANU_{OD}168$). This older age of the Mosquito Bore Distributary indicates inundation on more than one occasion. The paleoflood complex continues downstream around streamlined aeolian dune remnants, abutting against and deflected around bedrock ridges, and finally dissipates in the dune swales of the northern Simpson Desert (Fig. 6.46).

6.6. Eastern Systems Paleoflood Complex

East of the Mosquito Bore Distributary system, paleoflood complexes drain south from the modern Todd channel to terminal floodouts (Fig. 6.1 d, 6.46). The No. 5 Bore, Steele Gap-Rodinga Gap and modern Rodinga floodout were not covered by TM imagery and therefore are not included in Figure 6.1, but are indicated in the more general geomorphological map (Fig. 6.46).

6.6.1. Mosquito Bore A

Immediately to the east of the Mosquito Bore Distributary system and draining 17 km south from the Todd River, Mosquito Bore A abuts against a bedrock ridge and is diverted west and south around it (Fig. 6.1d). Downstream of the ridge and the streamlined colluvial and aeolian remnant, the 500-700 m wide cemented, poorly sorted fine sand (2.38σ) paleoflood complex is bordered to the west by the higher Mosquito Bore Distributary complex and to the east by Mosquito Bore B. It is estimated by OSL to be 5160±1340 BP (ANU_{OD}164).

6.6.2. Mosquito Bore B

Mosquito Bore B is of similar dimensions to Mosquito Bore A and abuts against the same high ridge but is diverted to the south east around it and floods out 19 km to the south between two high bedrock ridges in a small longitudinal dune field (Fig. 6.1 d). Channel sediment is cemented, poorly sorted fine sand (2.74σ with 2 % gravel) and an excavation showing at least two stratigraphic layers returned an OSL age of 2370±900 BP (ANU_{OD}165) from the younger event.

6.6.3. Todd B

East of Mosquito Bore B is a 2 km wide alluvial plain (Fig. 6.1 d). The cross section (Fig. 6.47a, b) morphology is based on field sketches as the instruments became faulty during the survey. The total depth of the coarse and medium sand matrix-supported small gravel paleoflood fill is unknown.

Section 1 lies close to the main channel, is composed of matrix supported gravel and fine sand and has a large channel form inset which suggests that the Todd has laterally shifted position. Section 2 is a 1.2 km wide sand paleoflood complex, dissected by three wide and shallow channels with low (24 cm) aeolian dunes. The third section is the Mosquito Bore B system described earlier.

Two radiocarbon and two OSL samples were dated for this reach. (Fig. 6.47b). The upper ^{14}C sample of a terrestrial shell is dated to 9883-8429 cal BP (ANU 9650). The underlying sample returned an age of 3712-3142 cal BP (ANU 9649) which is rejected as the charcoal was finely dispersed in the profile.

The older date is in good agreement with the OSL age (Fig. 6.54) 9780±1510 BP (ANU_{OD}163).

6.6.4. Allora B

The 200 m to 2,000 m wide Allora B flood distributary drains south from the right bank of the Todd channel along the low gradient (.0013 m/m) swales of the longitudinal dune system (Fig. 6.1d). A 2 m excavation indicates two poorly sorted fine sand (2.77σ) layers with a high (25%) silt/clay content overlying cemented Pleistocene sediment (Fig. 6.13). An OSL age from the younger event is 2110±760 (ANU_{OD}166).

6.6.5. No. 5 Bore

The No. 5 Bore paleochannel complex appears as a cutoff draining east from the modern channel through an outlying longitudinal dune field of the northern Simpson Desert (Fig. 6.46). Five sites on the abandoned course of the Todd channel and backwater area were examined (Figs. 6.14, 6.15).

6.6.5.1. Site 1, macroturbulent scours

Approximately 30 m from the cutoff, eleven scour holes are arranged in a semi-sinuous pattern along the thalweg of the paleochannel (Fig. 6.14, 6.48, Plate 6.16). They are closely spaced along a convex long profile, a higher density of scours associated with a steeper gradient (Fig. 6.49). The average depth is 0.88 cm with a maximum of 1.57 m and a minimum of 0.24 cm; the width varies from 74 m to 30 m with an average of 43.4 m and the length varies between 38 m and 16.5 m with an average of 23.6 m (Table 6.6). The downstream lip may be at a higher elevation than upstream (scour numbers 2, 8, Fig. 6.50a, b), at an equal elevation (scour numbers 1 and 6, Fig. 6.50a) but tend to be at a lower elevation (scour numbers 3, 4, 5, 7, 9, 10, 11, Fig. 6.50a, b). Cross sections tend to be asymmetric (Fig. 6.51), the walls steep and the floors scattered with coarse sand and cobbles.

A 1 m excavation in the side wall of one of these scours indicates a horizontally stratified medium sand and gravel basal layer truncated at the scour wall and overlain by a thin (15 cm) inclined stratified sand layer. This is inferred as the erosion of the scours into a horizontally bedded alluvium with subsequent colluvial movement of sediment down the scour slope. A radiocarbon age from charcoal sampled in the scoured alluvium returned an age of 1083-726 cal BP (ANU 9655).

6.6.5.2. Cross sections

The first of three cross sections (Fig. 6.15, 6.52) indicates a 100 m wide matrix-supported coarse gravel channel with a horizontally laminated medium and fine sand inset flood plain which has small braiding channels on the surface. The preservation of sedimentary structures in the flood plain indicates a young age. The adjacent left bank coarse sand and gravel unit is overlain by a structureless mud and fine sand levee. The right bank coarse white sand aeolian dune differs from the left bank red fine sand Pleistocene dune because the sediment in the former was locally sourced from the paleochannel in a manner similar to the dune at Jessie Gap, and is reworked by runoff from the ridge which transports angular weathered rock down slope (Plate 6.17).

The second cross section is located downstream (Fig. 6.15, 6.53) and two trenches were excavated by 'bob cat' (Fig. 6.55). The channel is underlain by red cemented Pleistocene sand which is overlain in the channel by a 2.5 m inclined gravelly sand inferred to be a channel bar and two structureless coarse and medium sand layers. A radiocarbon sample from the upper layer is dated at 1405-1135 cal BP (ANU 10058). On the left bank a mud rich coarse sand layer, incised by the channel is overlain by structureless sand and mud and a thin red fine sand aeolian deposit

A third cross section at No. 5 Bore (site 5.1) was excavated to the underlying cemented Pleistocene deposit (Fig. 6.15, 6.54). This is overlain by a moderately sorted fine sand (2.4ø) with some gravel dispersed throughout and increasing in concentration down the profile. Two layers of aeolian sand towards the top of the profile are interrupted by a 1 cm layer of mud from a lower magnitude flow probably after the channel avulsed. An OSL age from the channel fill is 680[illegible]±1850 BP ($ANU_{OD}169$) and indicates early Holocene fluvial activity of the Todd River. Depth to the aeolian base is unknown as auger refusal occurred at [illegible] 2 m. The right bank flood plain is a fine textured levee system similar to the left bank levee system upstream. A 4.5 m auger hole in the left bank terrace revealed a basal 50 cm mud layer overlying the Pleistocene sand and overlain by gravelly sand and silts.

6.6.5.3. Backwater areas

Marginal to the channel, the swales between longitudinal dunes were back filled with flood waters and numerous small lakes were formed in topographic lows. Today these lake beds are still apparent as small (<300 m^2) dry clay pans. Three excavation pits in the dune swales typically indicate 95 cm of silty fine sand which overlies the cemented Pleistocene alluvium and is overlain by a thin (5 cm) layer of modern aeolian sand. The stratigraphy of one of these paleo lakes however indicates a 5 m sequence of lacustrine mud overlying Pleistocene sediment. This lake sequence is interrupted by a 20 cm band of aeolian sand indicating a hiatus in deposition and inundation by floods on at least two occasions. The older basal lacustrine mud is mottled green indicating long-term high ground water conditions.

Therefore the geomorphology of the No. 5 paleoflood channel indicates that it was a previous course of the Todd which was active between at 8 ka and 1 ka. A large climbing aeolian dune was nourished from the channel sand and the swales of adjacent dunes backfilled with suspended load mud. The channel avulsed from this position to its present location approximately 1 ka most probably following the high magnitude event that formed the macroturbulent scours.

6.6.6. Steele Gap to Rodinga Gap

The 12 km (max) wide matrix-supported gravel expansion bar extends 30 km south of Steele Gap (Fig 6.46) through Rodinga Gap and is bordered to the west by truncated aeolian dunes in the south west by deposits [illegible]m the Mosquito Bore Distributary system a[illegible] the east by truncated aeolian dunes and the modern floodout. Paleoflo[illegible] channel gradients along this track are similar between Steele Gap and Rodinga Gap (.00164 m/m) and downstream of Rodinga Gap (.00163) but are less steep through Rodinga Gap (.00107 m/m) most probably as result of deposition upstream of the gap. The surface morphology is similar to the expansion bar complex at the Camel Flat site as some longitudinal bars are interspersed with smaller channels and some low aeolian dunes. An OSL sample from the structureless coarse and medium sand deposit returned an age of 3900±1270 ($ANU_{OD}167$).

On the northern side of the ridge to the west of Rodinga Gap an excavation revealed an alternating sequence of flood silts and aeolian sand overlain by 10 cm of red aeolian sand. Downstream from Rodinga Gap the expansion bar complex reaches a maximum of 6 km wide. It is bordered to the east and west by truncated longitudinal dunes. The coarse sand and gravel deposit returned an age of 12310±1400 BP (ANU_{OD}172).

6.6.7. Modern Rodinga floodout

There is a well-defined morphological and sedimentological break between the modern floodout and the Steele to Rodinga floodout, as the former is at a lower elevation and composed of mud and fine sand. South of Rodinga Range a braided distributary channel from the modern channel has inundated and eroded adjacent dunes (Fig. 6.46). Where it has breached longitudinal dunes an excavation pit reveals 80 cm of alluvial deposits from possibly four floods which overlie cemented red Pleistocene sediment.

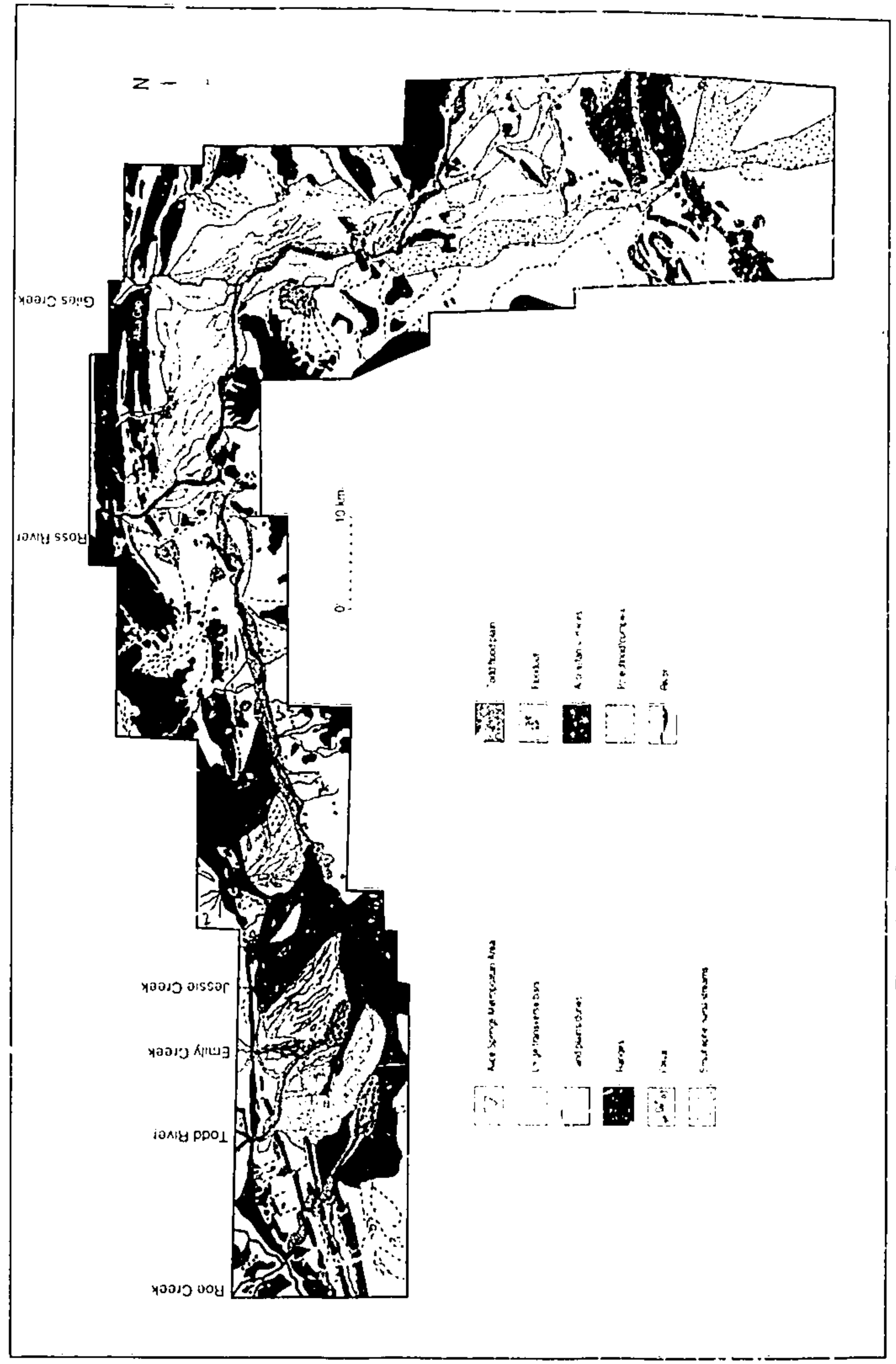


Figure 5.1a. Geomorphic map of the Todd River catchment.

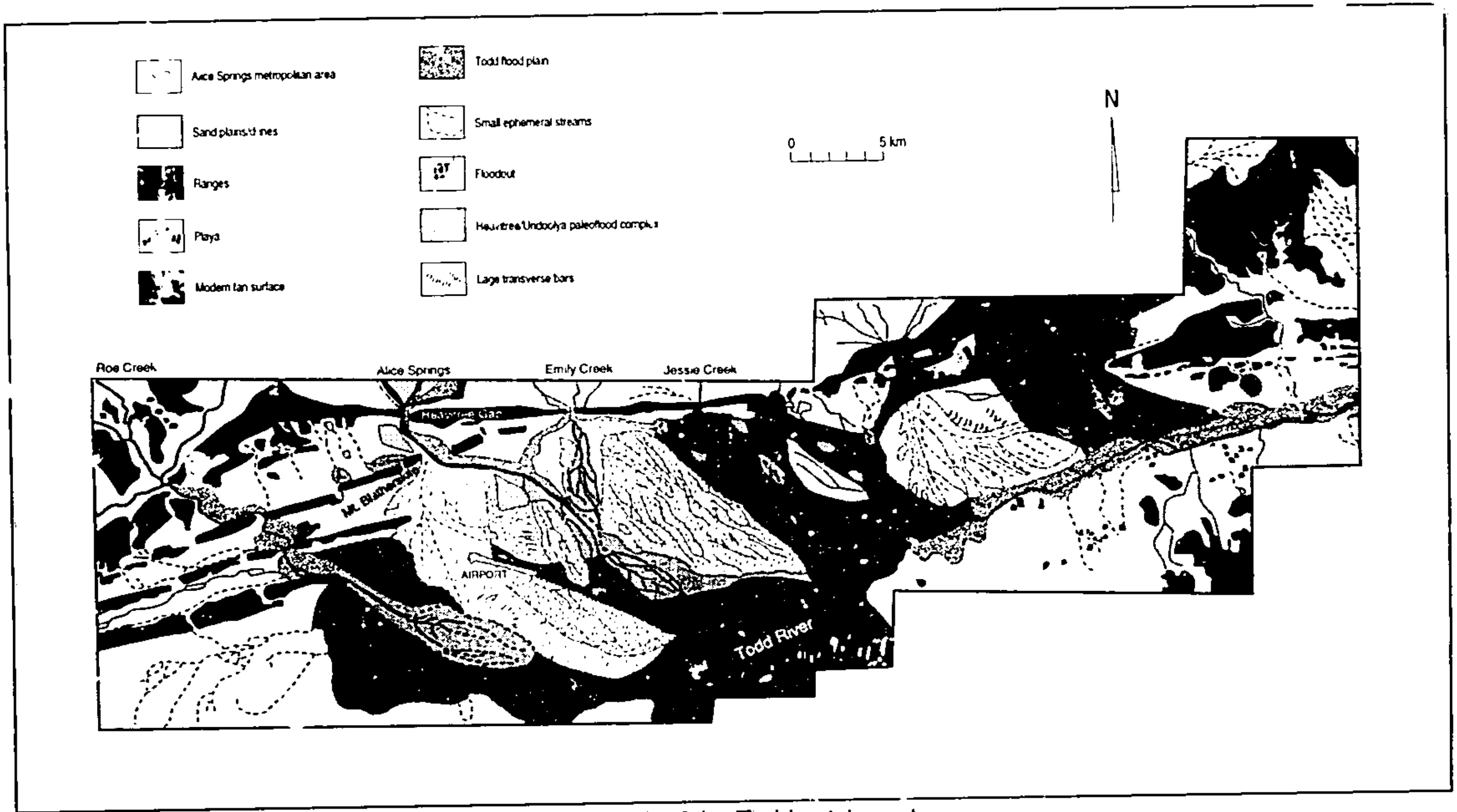


Figure 6.1b. Geomorphic map of the piedmont reach of the Todd catchment.

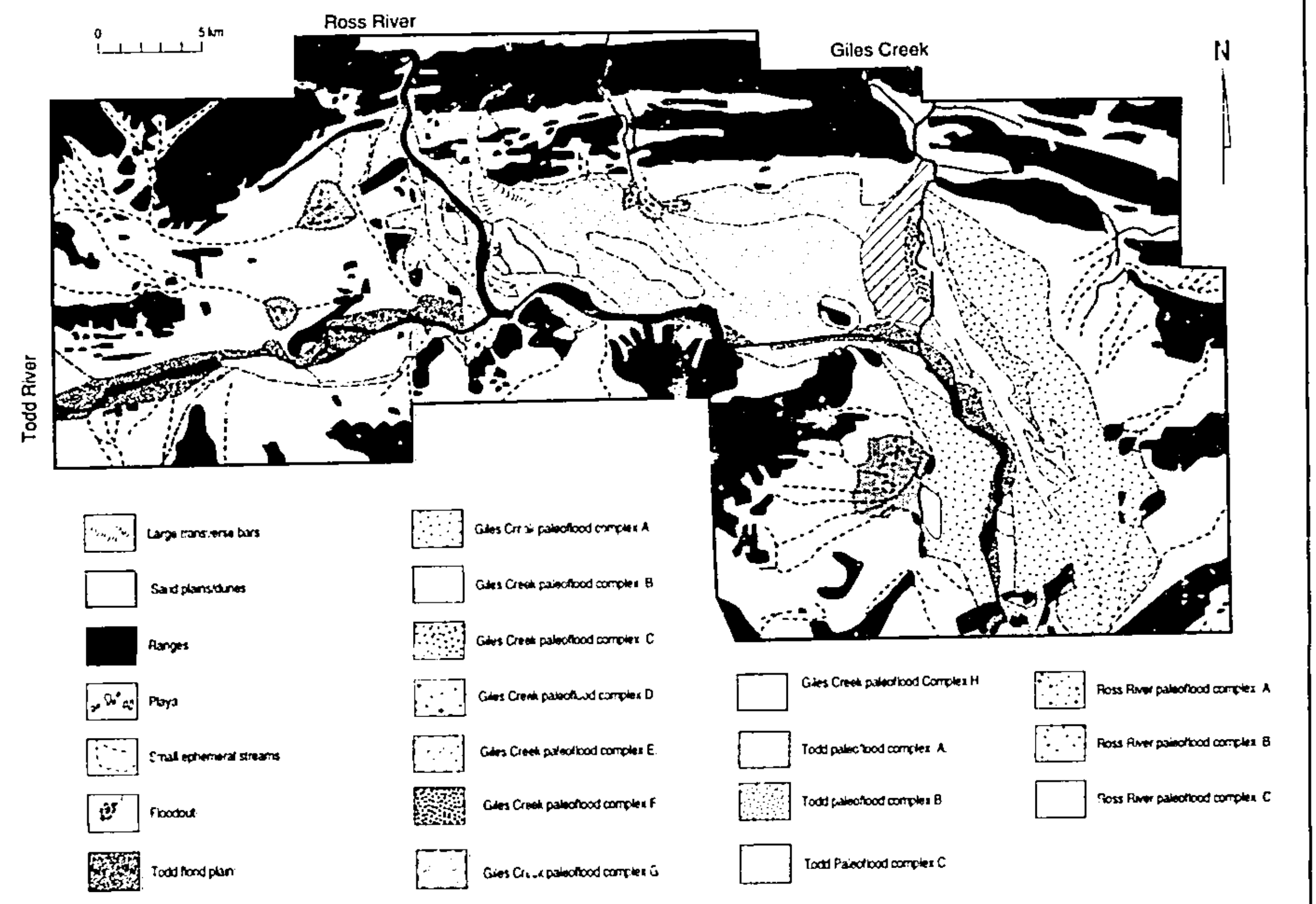


Figure 6.1c. Geomorphic map of the Ross and Giles piedmont reach.

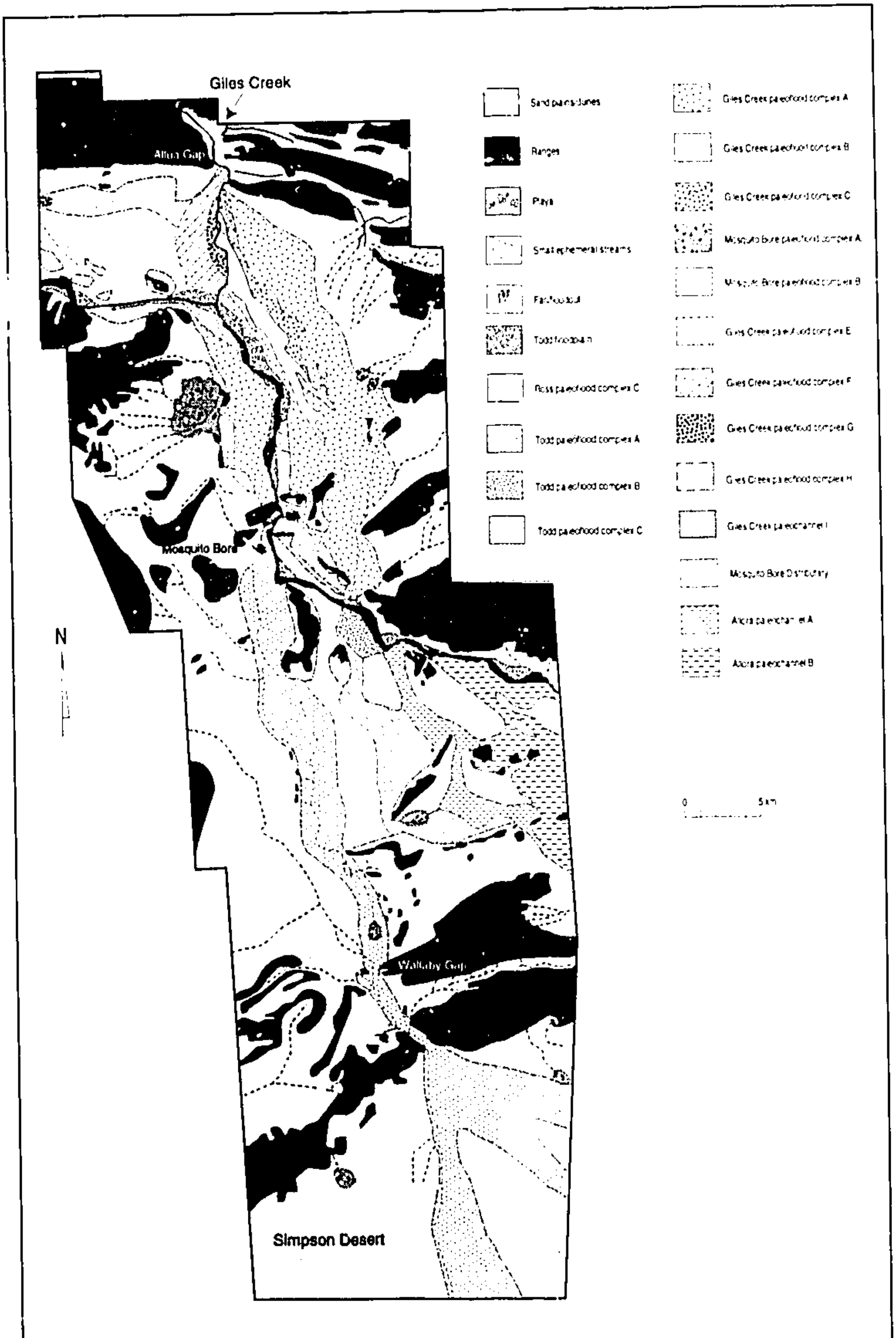
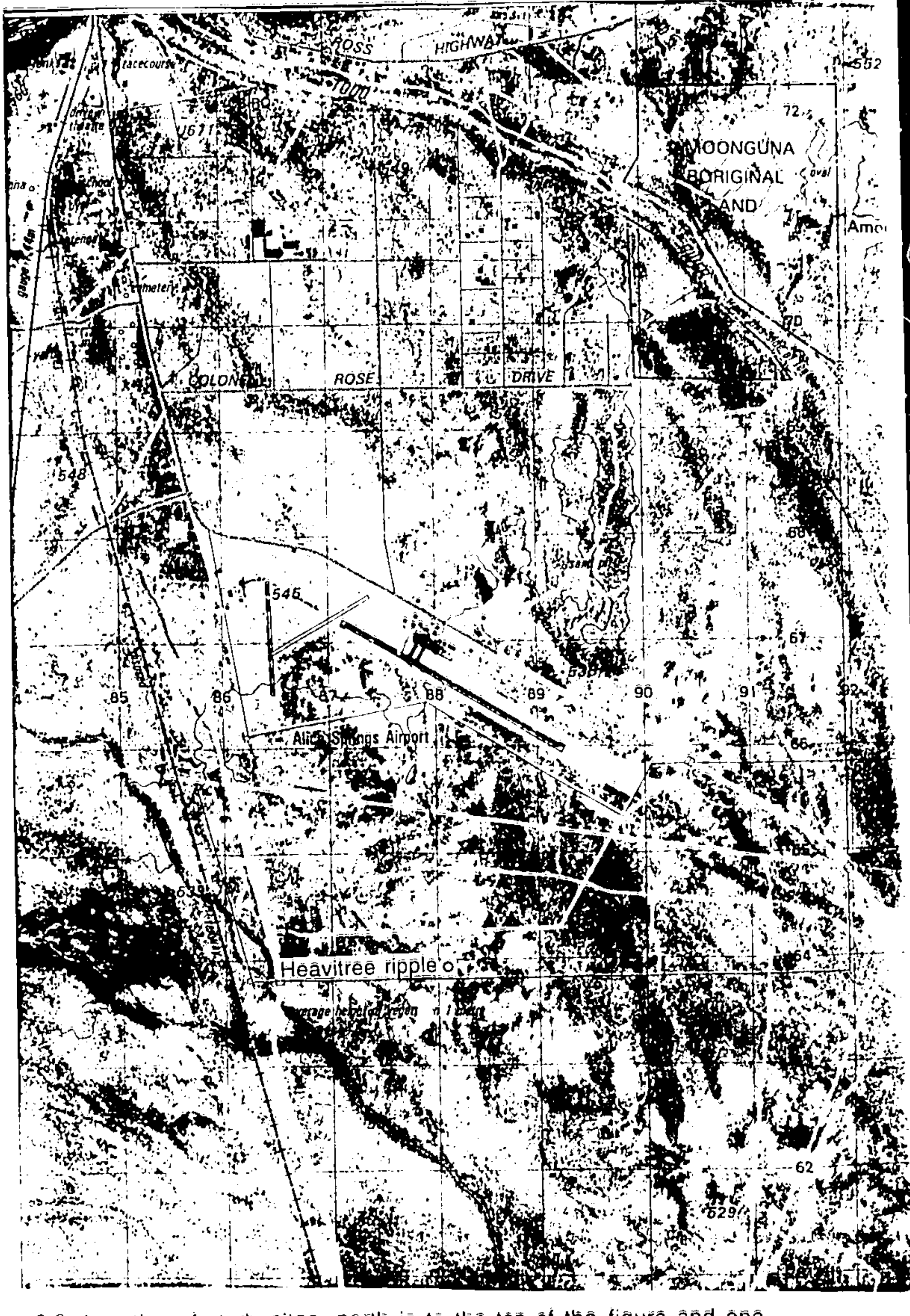
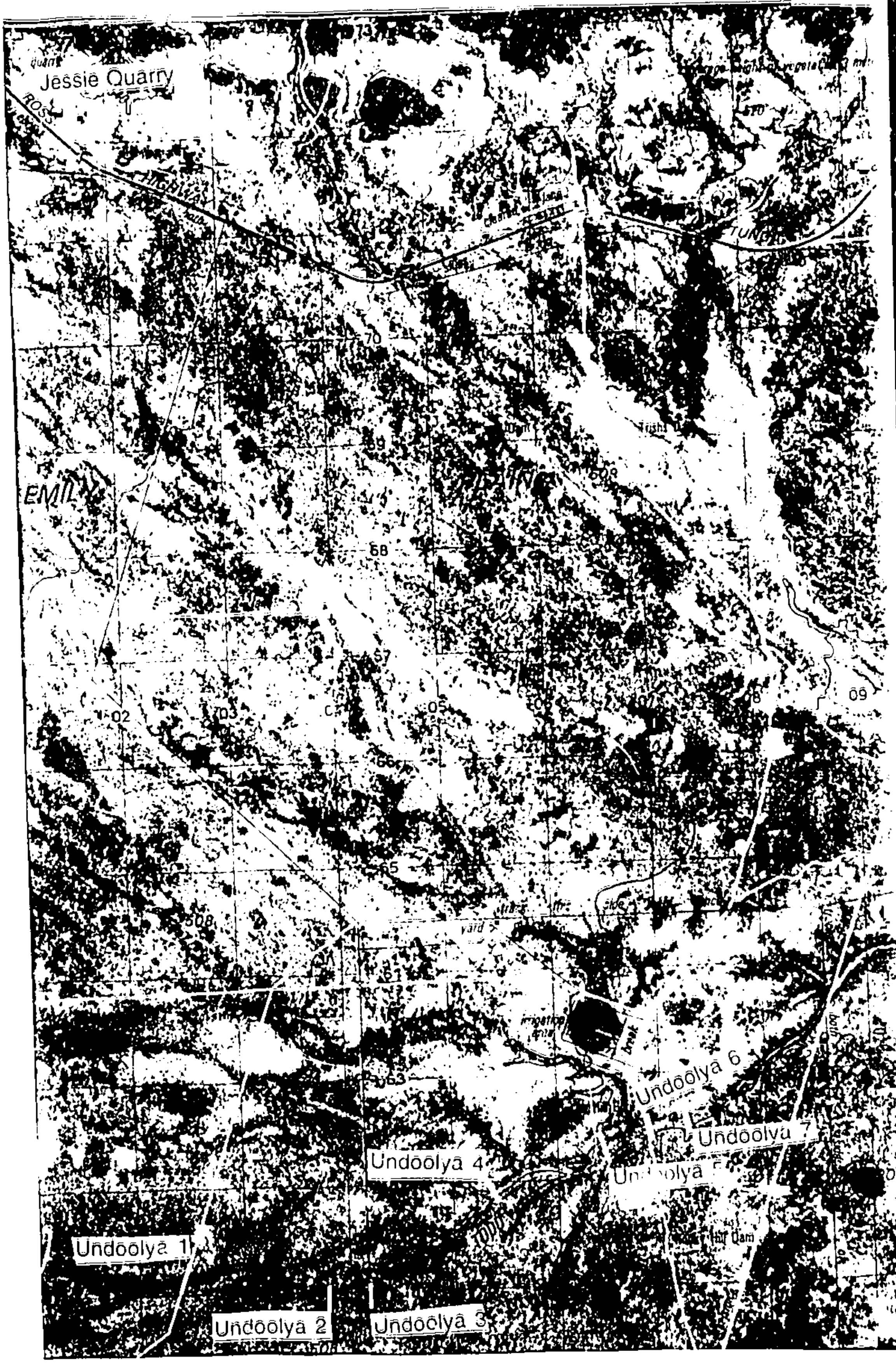


Figure 6.1d. Geomorphology map of the eastern Todd catchment.

6.2. Location of study sites, north is to the top of the figure and one square is 1 km.

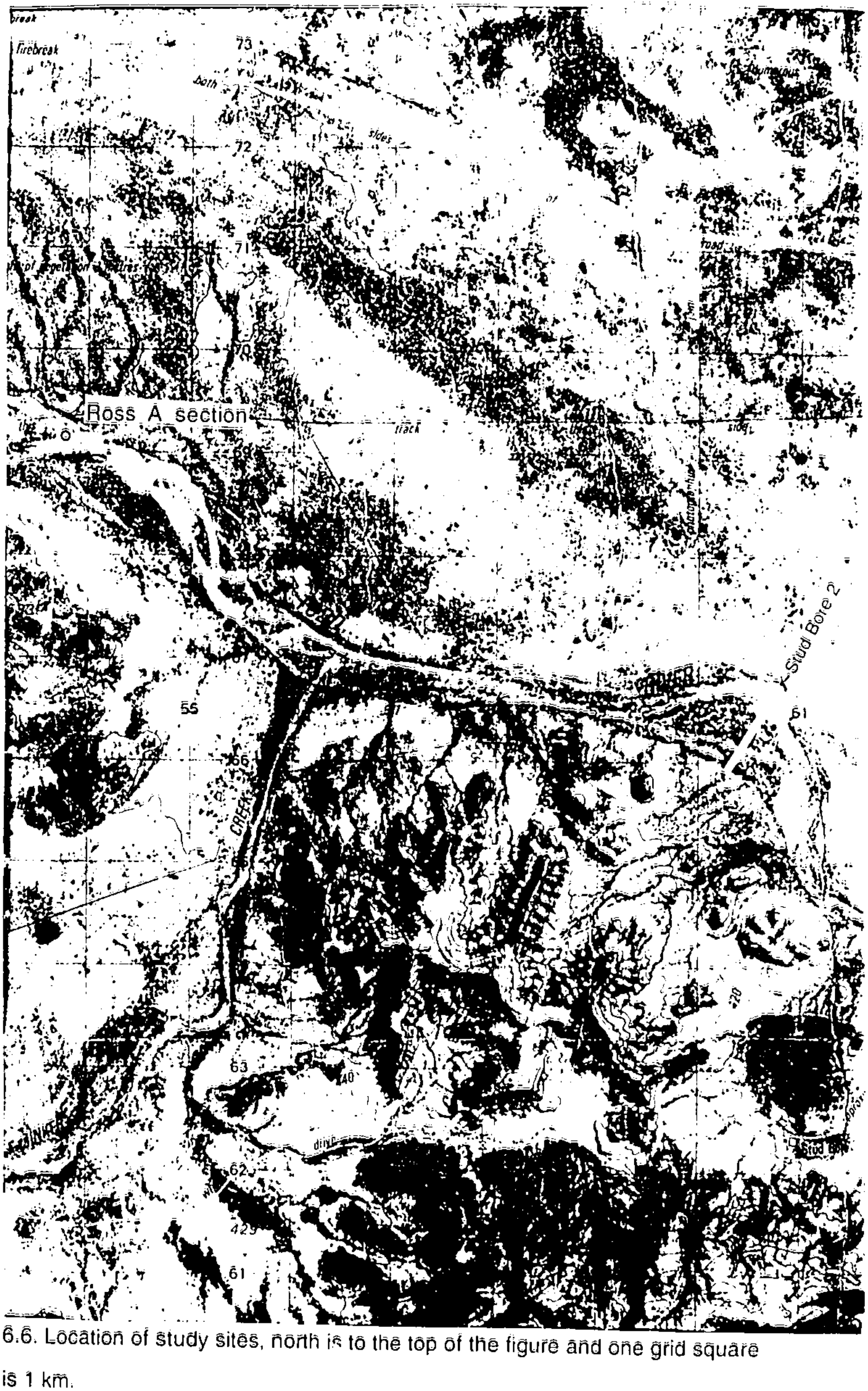


6.3. Location of study sites, north is to the top of the figure and one square is 1 km.

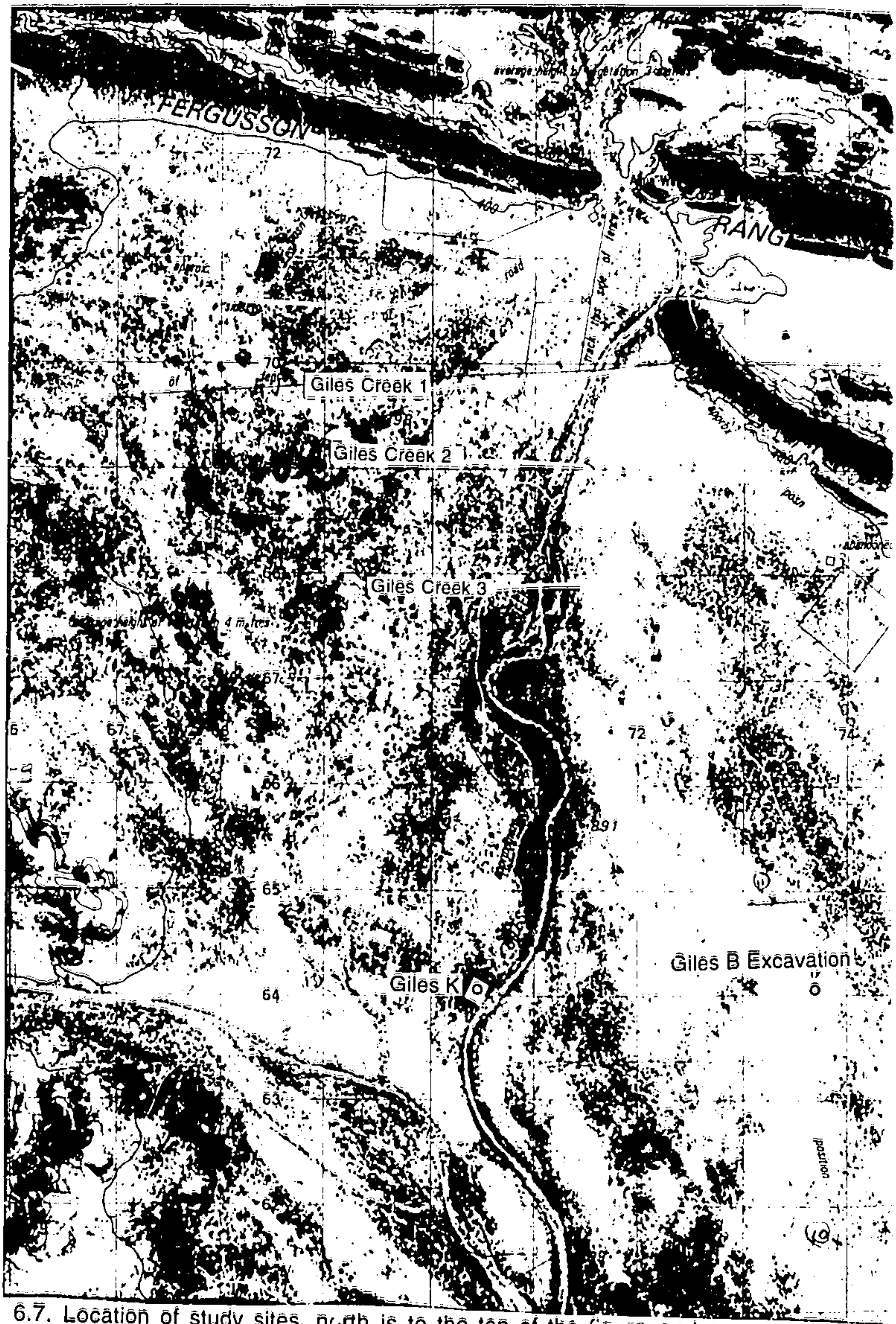


6.4. Location of study sites, north is to the top of the figure and one square is 1 km

6.5. Location of study sites, north is to the top of the figure and one square is 1 km.



6.6. Location of study sites, north is to the top of the figure and one grid square is 1 km.

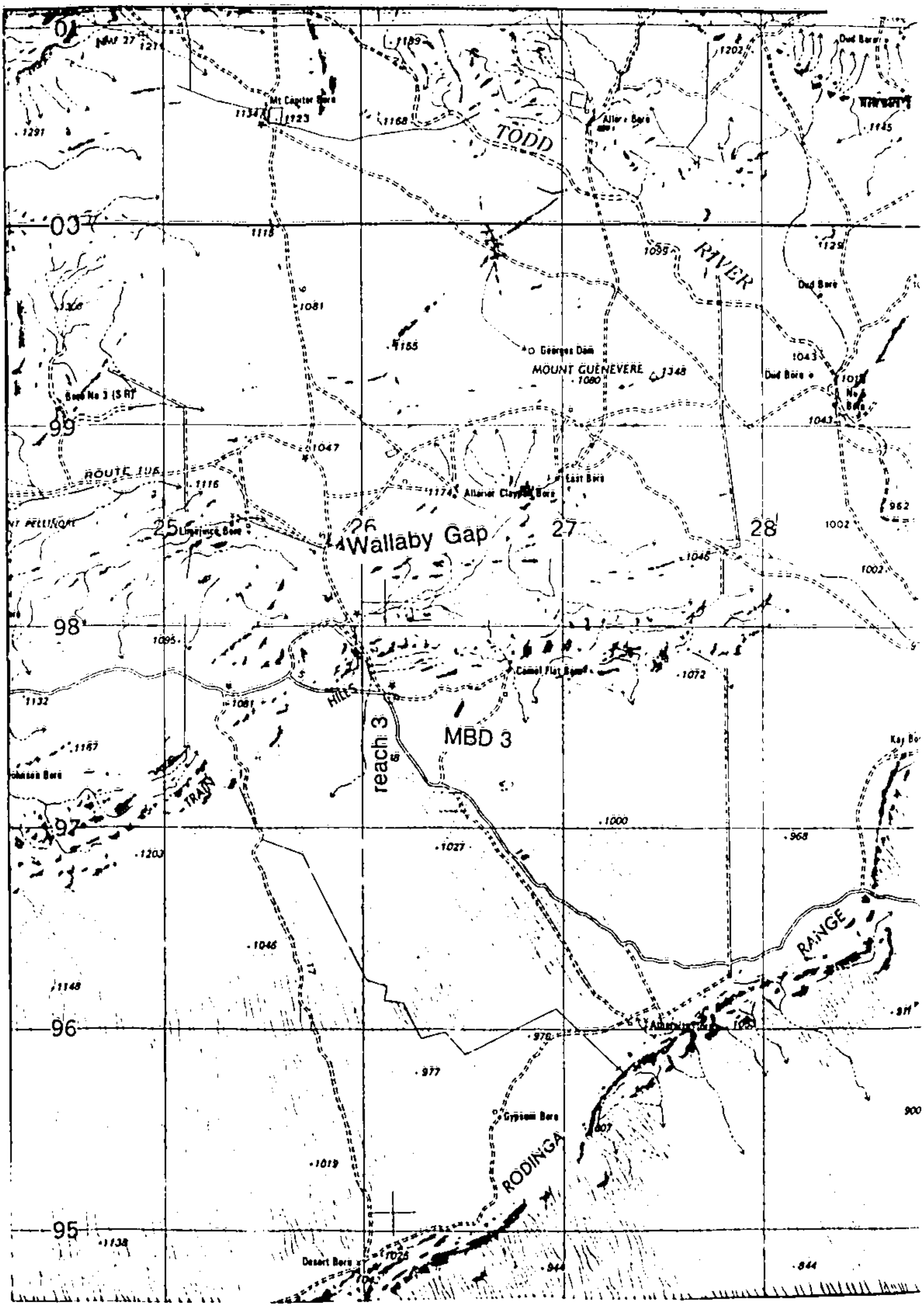


6.7. Location of study sites, north is to the top of the figure and one square is 1 km.

6.8. Location of study sites, north is to the top of the figure and one grid square is 1 km.

6.9. Location of study sites, north is to the top of the figure and one grid square is 1 km.

6.10. Location of study sites, north is to the top of the figure and one square is 1 km.



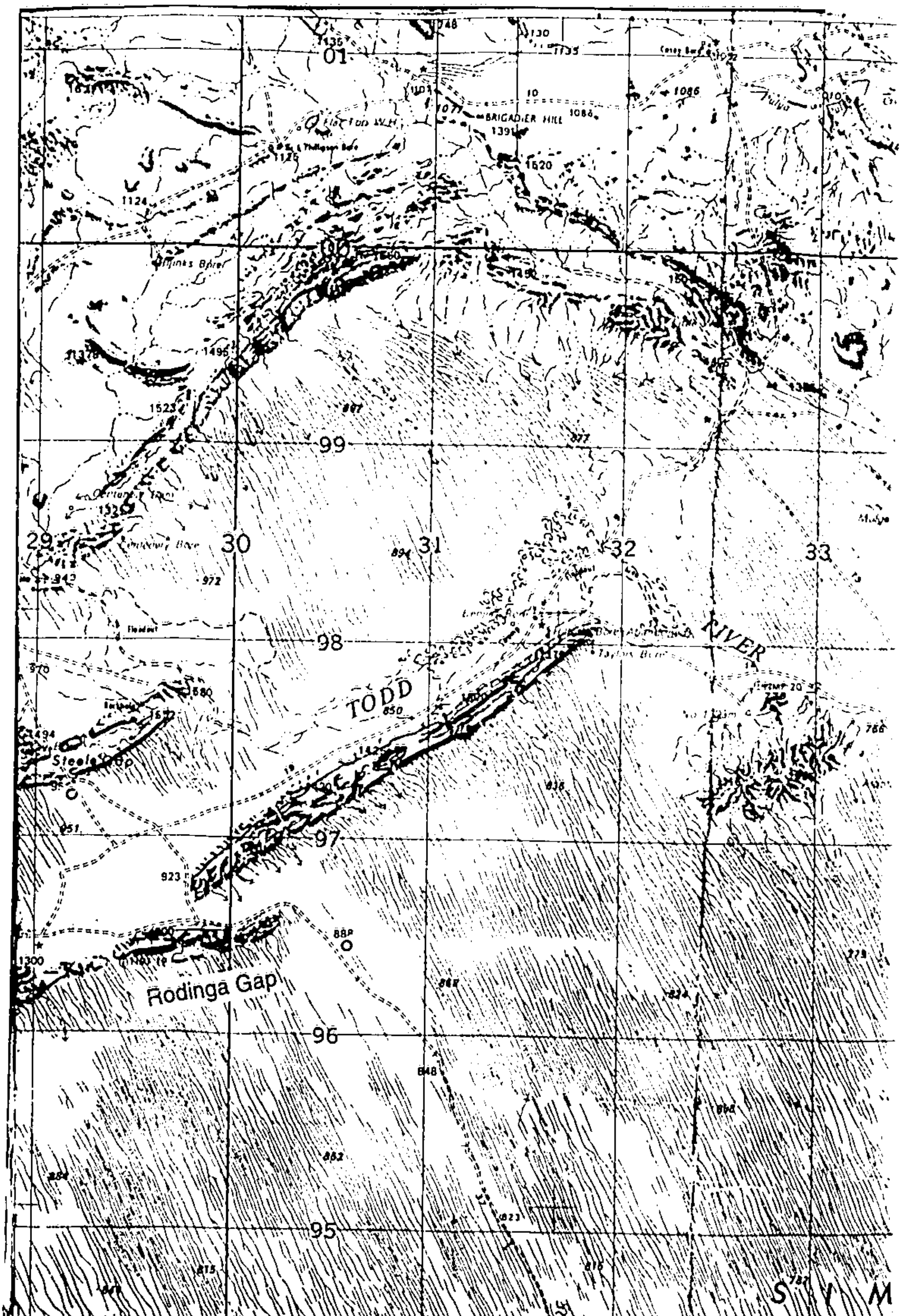
6.11. Location of study sites, north is to the top of the figure and one square is 9.2 km.

6.12. Location of study sites, n. rth is to the top of the figure and one grid square is 1 km.

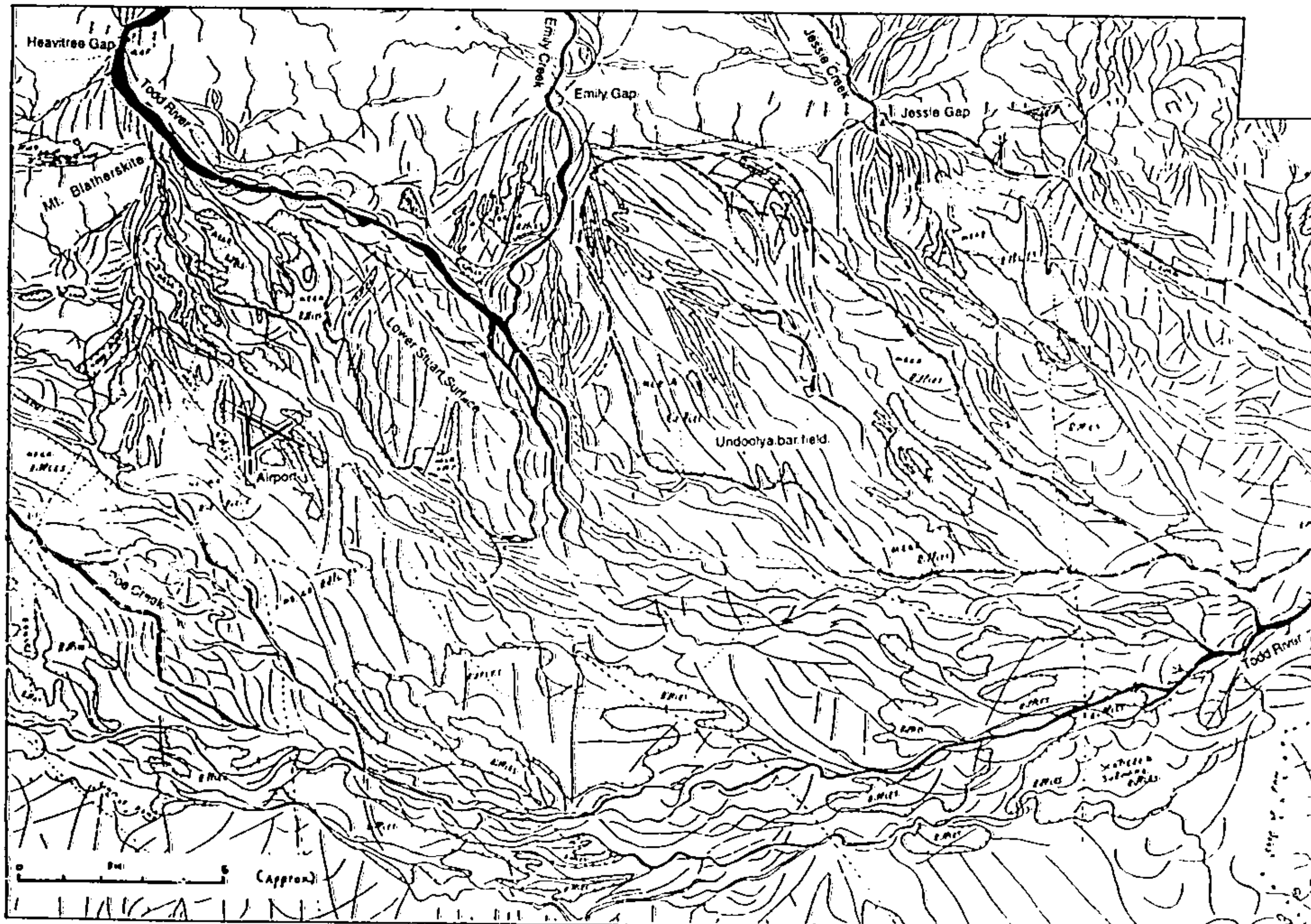
6.13. Location of study sites, north is to the top of the figure and one square is 1 km.

6.14. Locati n of study sites, north is to the top of the figure and one square is 1 km.

6.15 Location of study sites, north is to the top of the figure



6.16. Location of study sites, north is to the top of the figure and one square is 9.2 km



6.17. Map of drainage lines in the Todd catchment piedmont area drawn by K. Fitchett (unpublished).

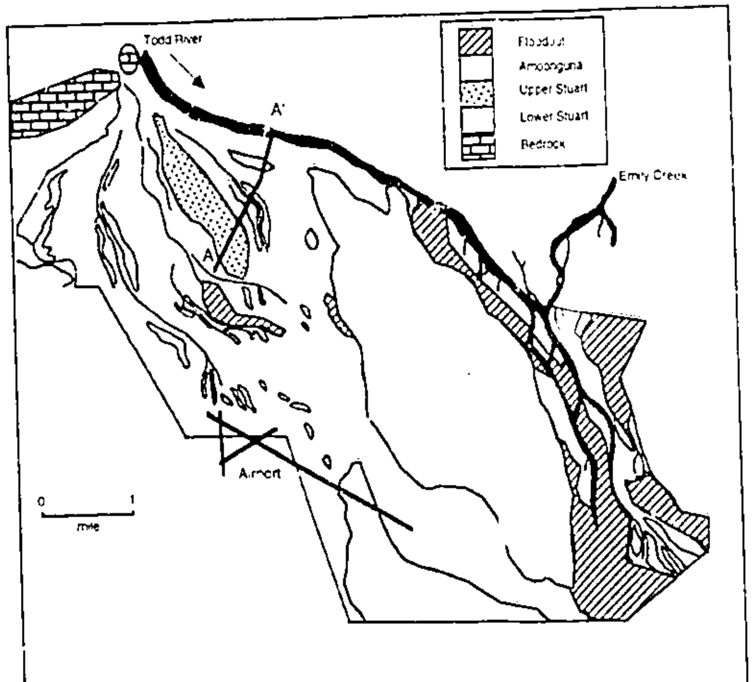


Figure 6.18a. Litchfield's (1969) alluvial surfaces on the Heavitree fan.

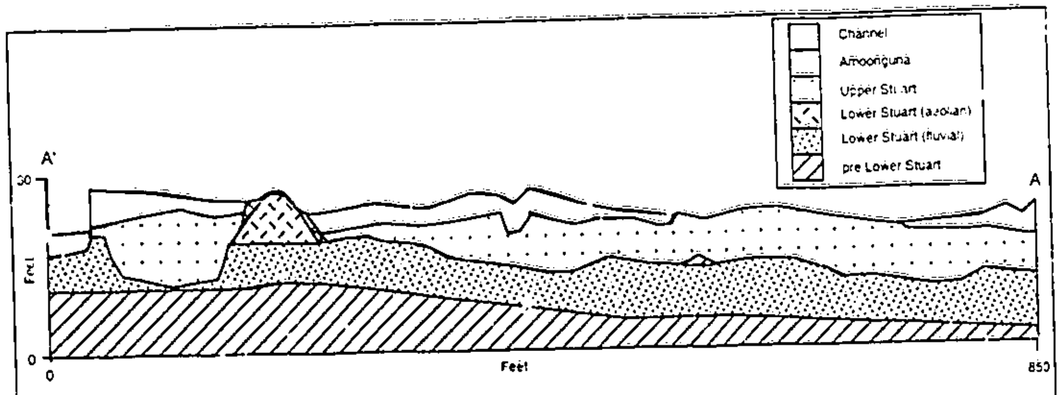


Figure 6.18b. Cross section of Heavitree fan from Litchfield (1969)

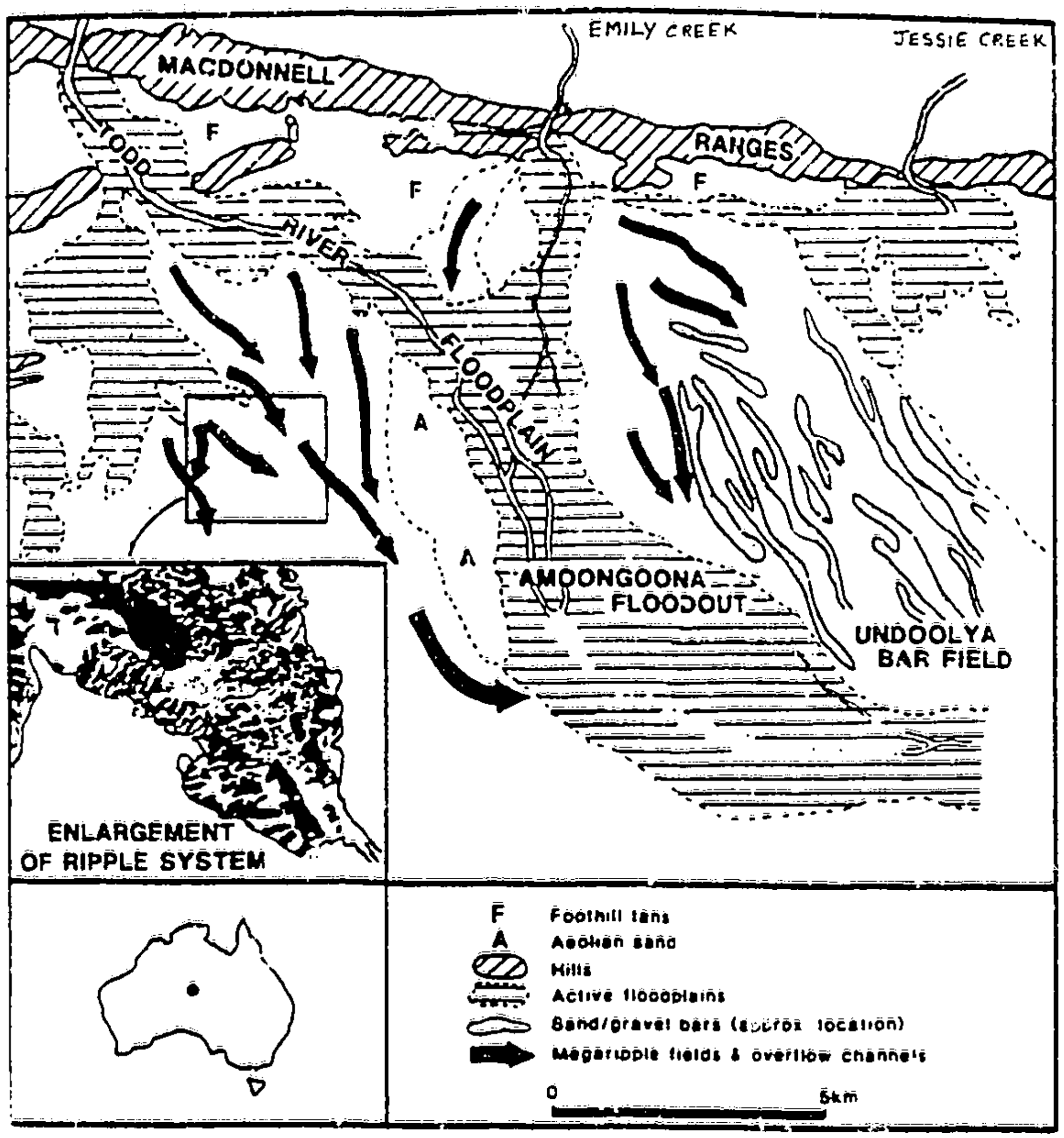


Figure 6.19. Undoolya bar field (Pickup, 1991).

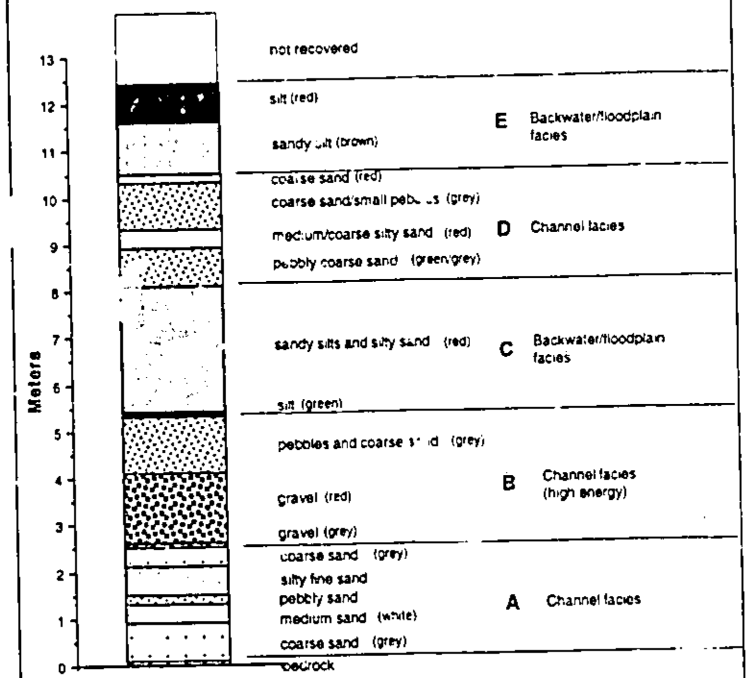


Figure 6.20. Profile of Heavitree Gap site 2 core, location in Figure 6.2.

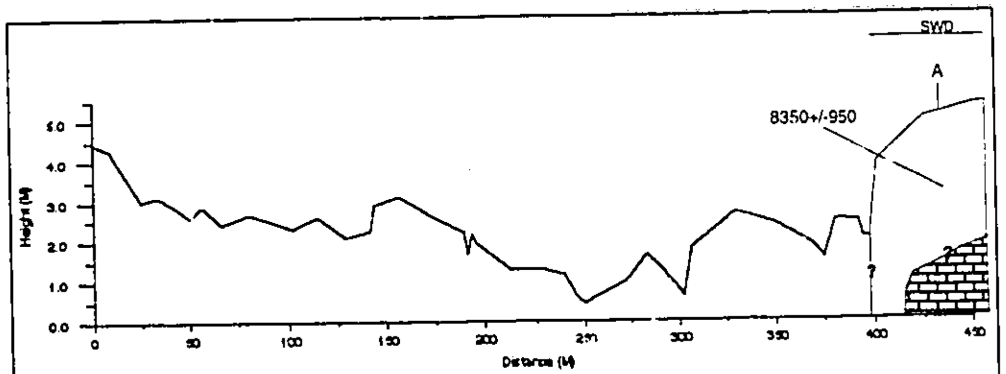


Figure 6.21. Morphostratigraphy of Shannon Bore slack water deposit, Location in Figure 6.5

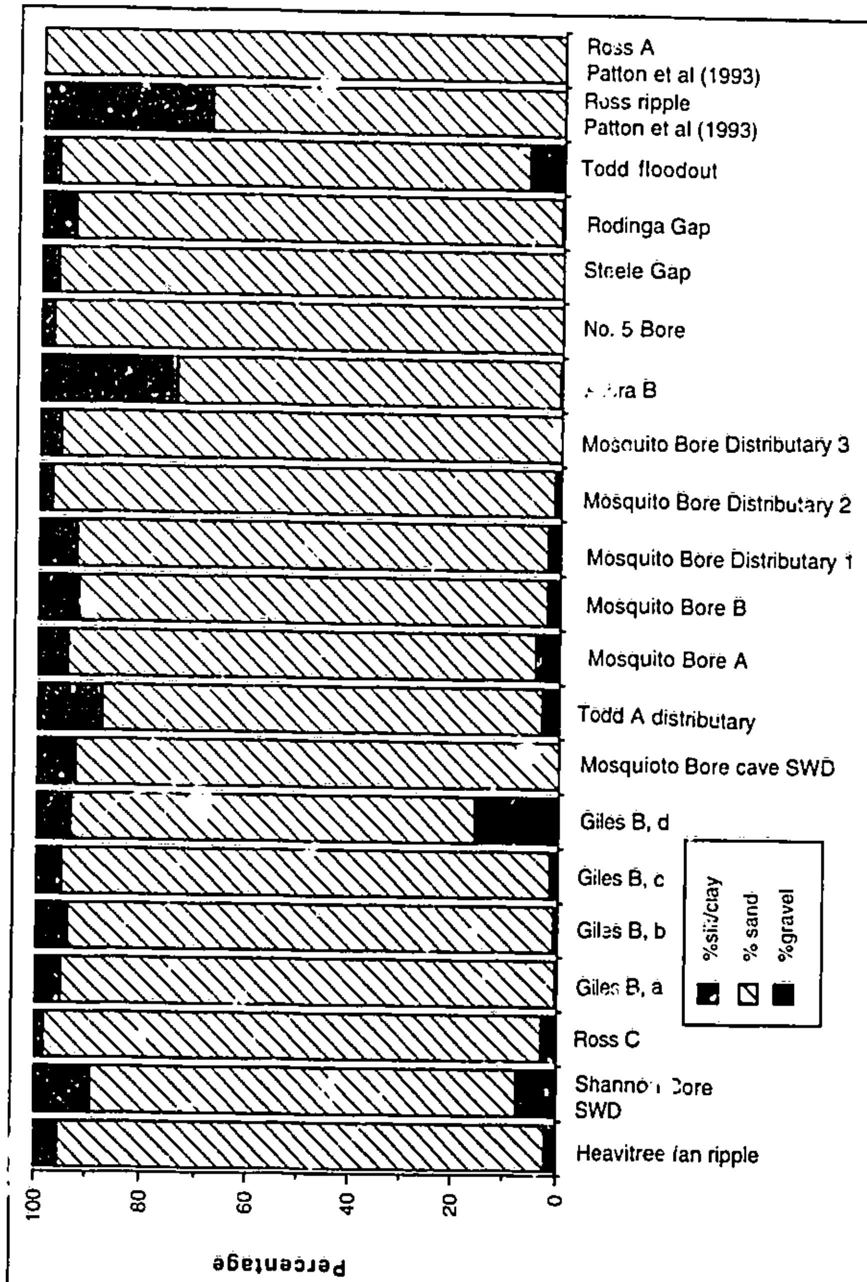


Figure 6.22. Stack bar graph indicating the percentage gravel, sand and silt/clay in paleoflood sediments.

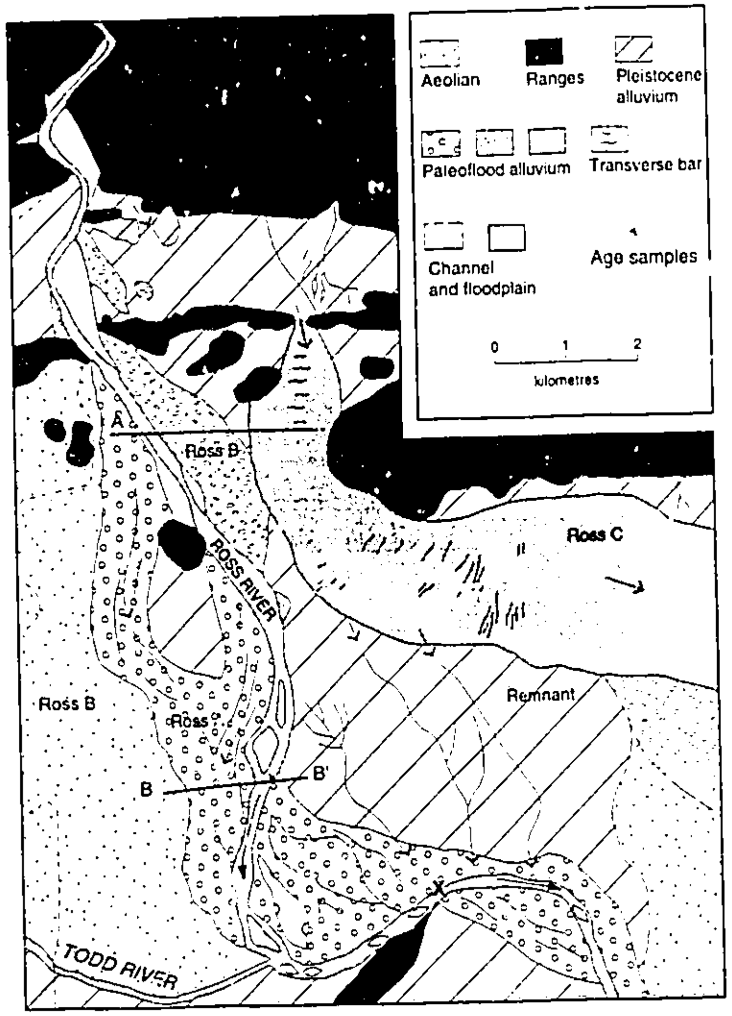


Figure 6.23. Ross paleoflood complex after Patton et al., (1993), annotated to show the Ross B surface (dotted line). X indicates location of Ross A excavation.

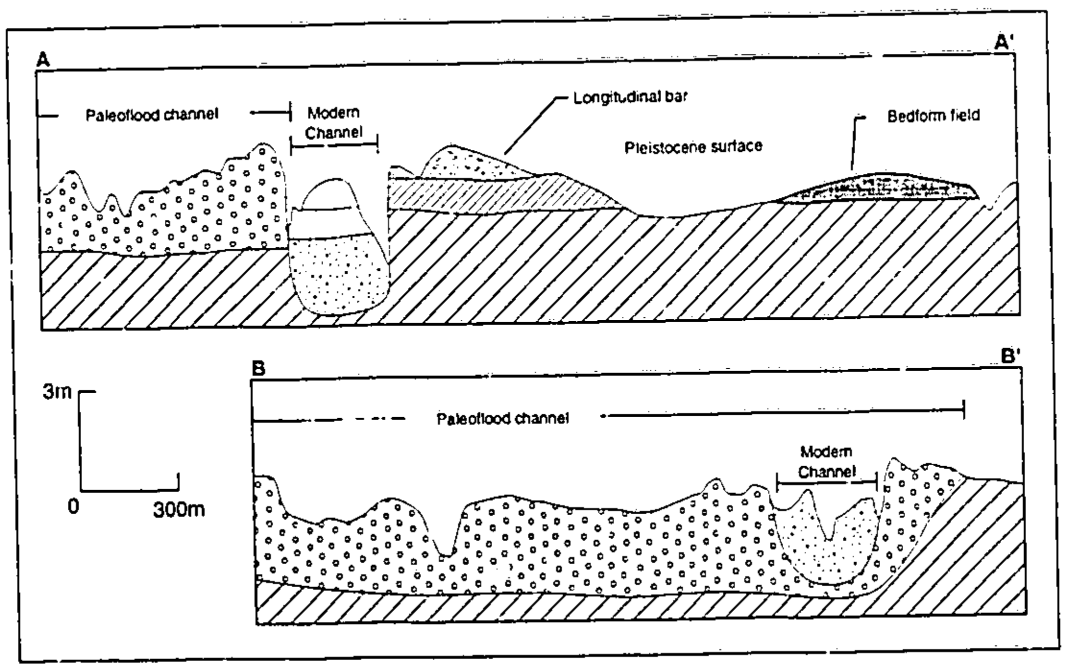


Figure 6.24. Ross A paleoflood complex cross sections after Patton et al (1993), location in Figure 6.23.

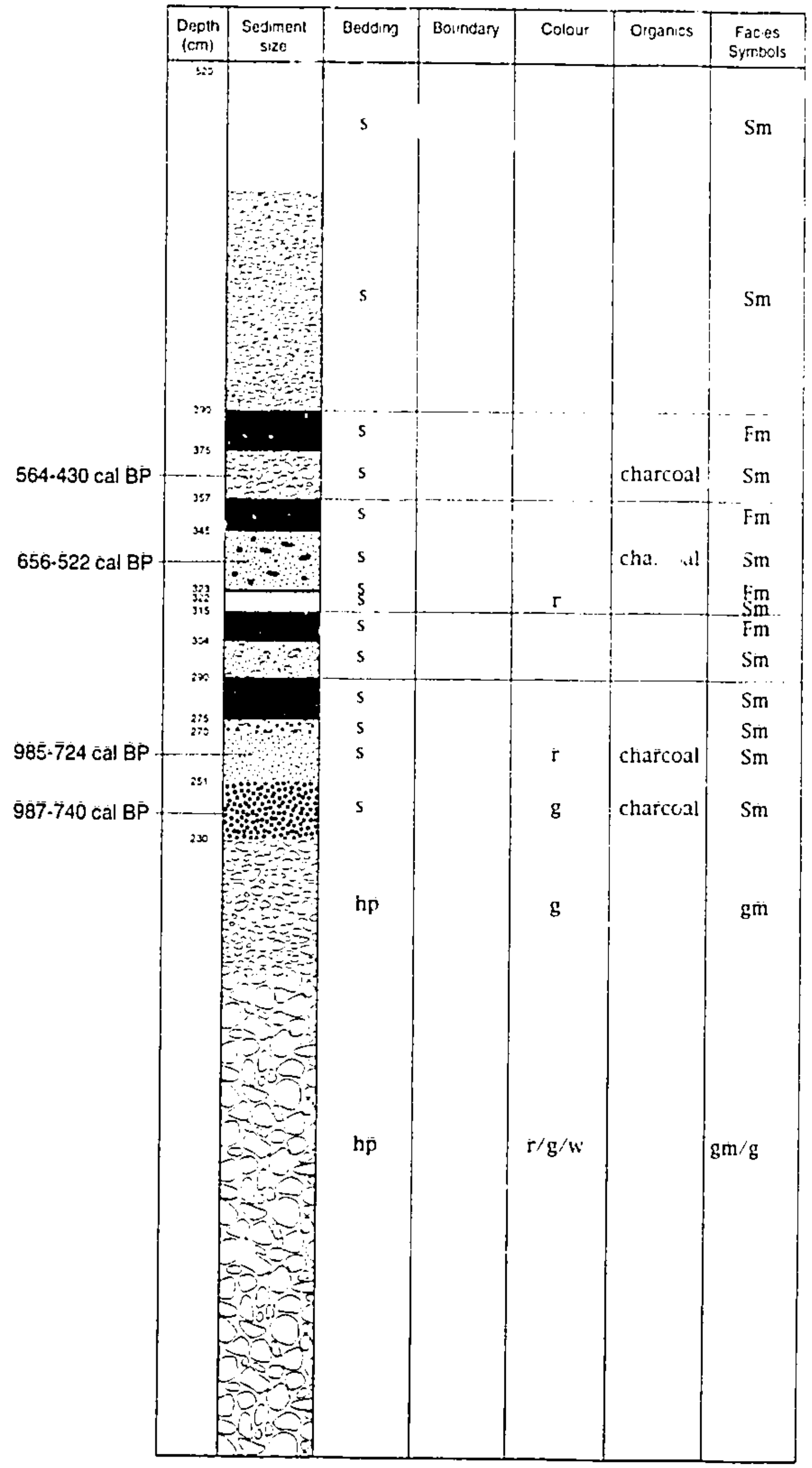


Figure 6.25. Profile A, Ross A paleoflood complex, location in Figure 6.23.

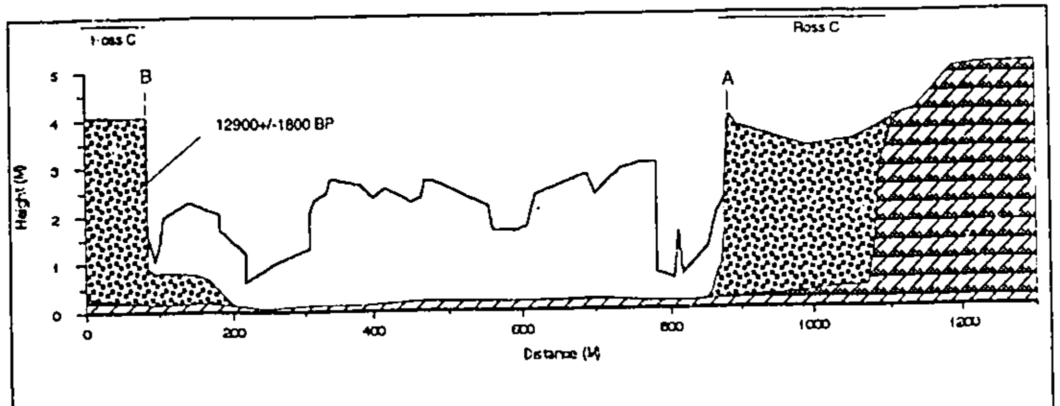


Figure 6.26. Morphostratigraphy of Ross C paleoflood complex at Stud Bore 2 site, location in Figure 6.6.

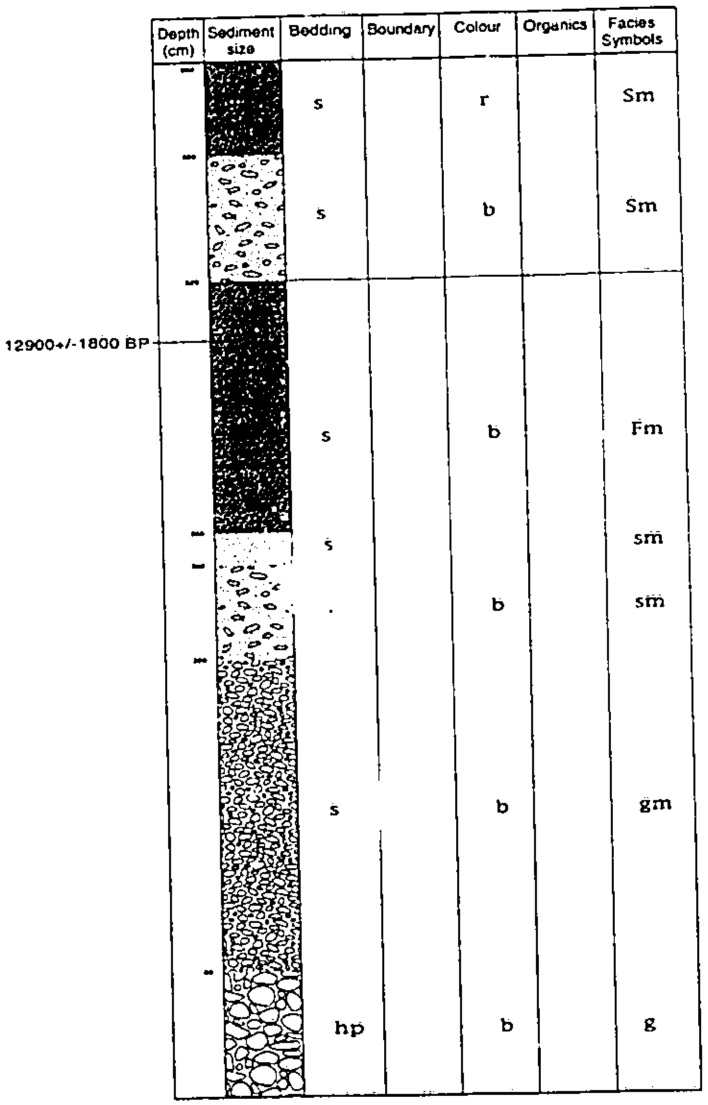


Figure 6.27. Profile B, Ross C paleoflood complex,S tud Bore 2 site, location in Fig. 6.26, see also Plate 6.5.

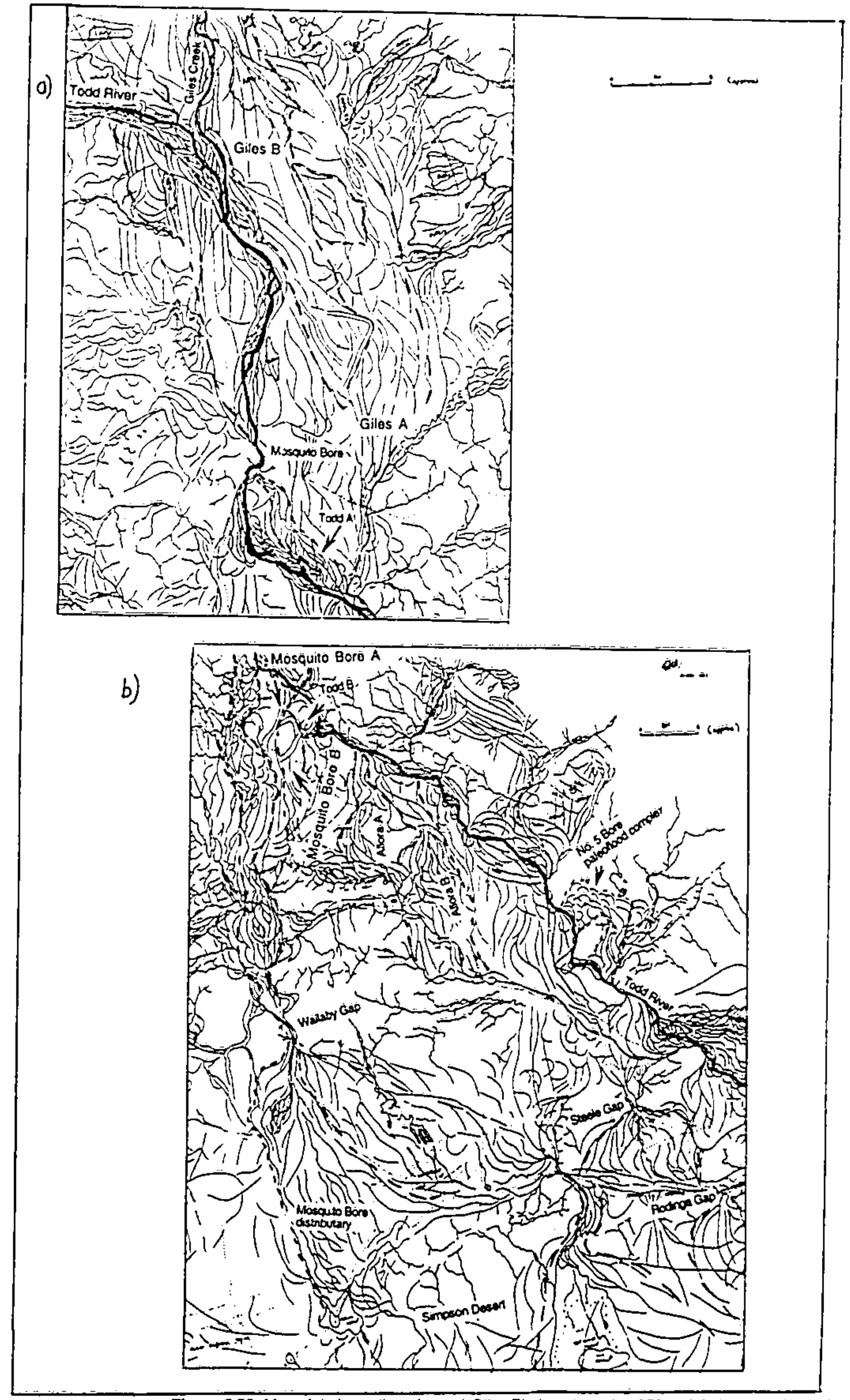


Figure 6.28. Map of drainage lines from a) Giles Piedmont reach, 1950 aerial photographs and b) Mosquito Bore Distributary, 1971 aerial photographs (Fitchett, unpublished). Relationships between paleoflood tracts and channels has to be confirmed in some areas

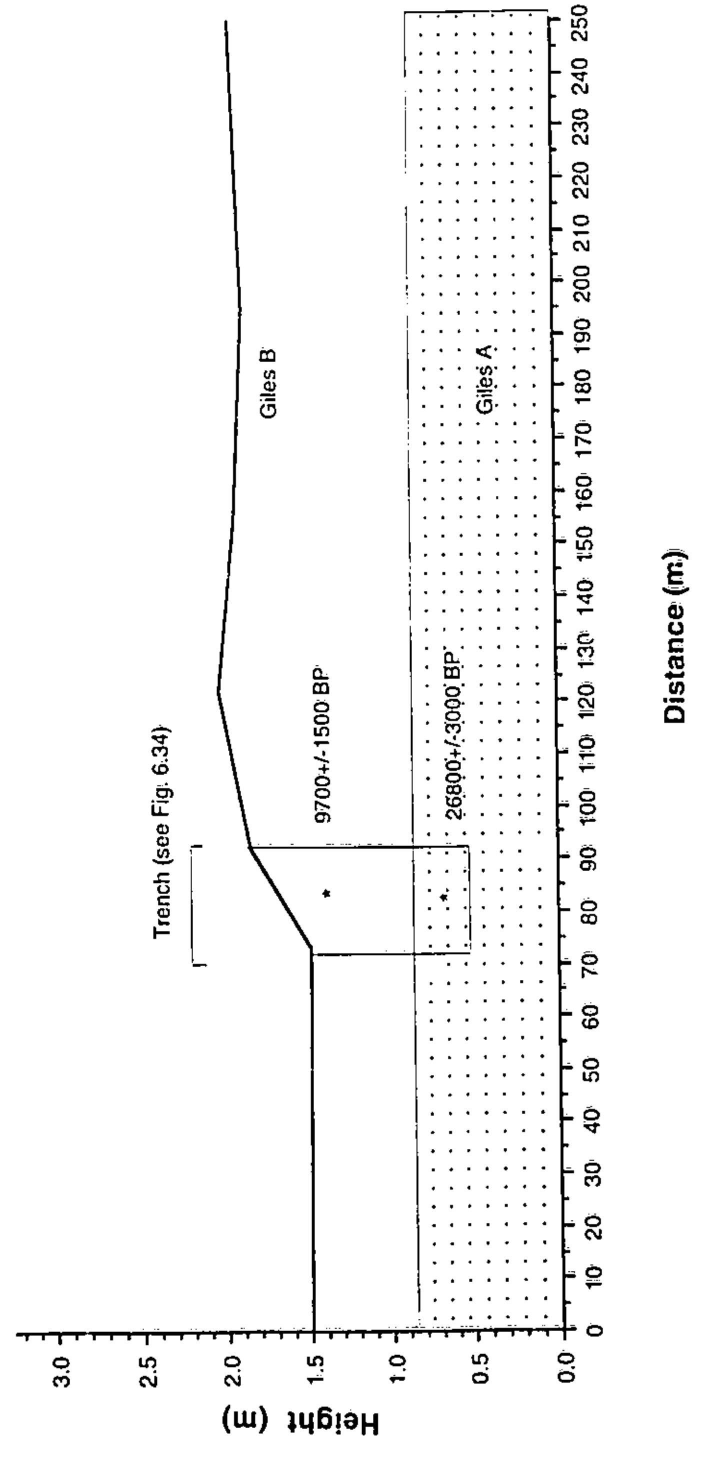


Figure 6.29. Cross profile of longitudinal bar excavation on Giles B, location in Fig. 6.7.

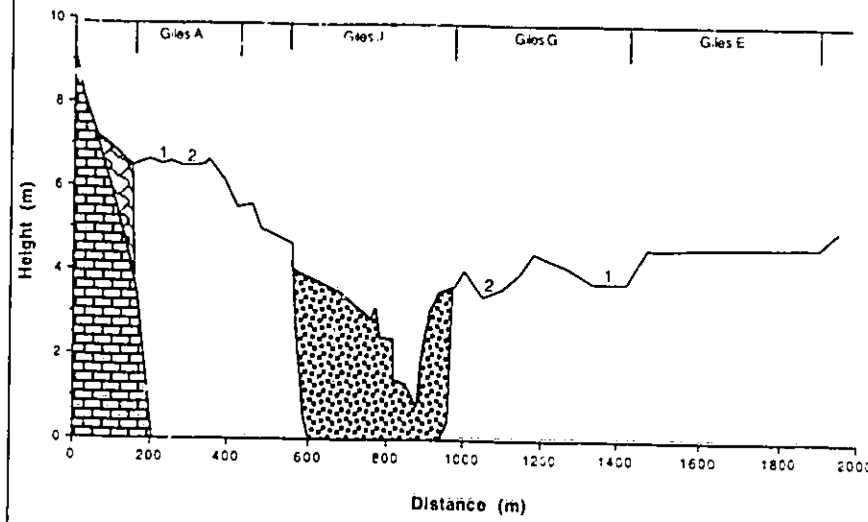


Figure 6.30. Morphostratigraphy of Giles J, at Giles Creek 1 site, location in Figure 6.7.

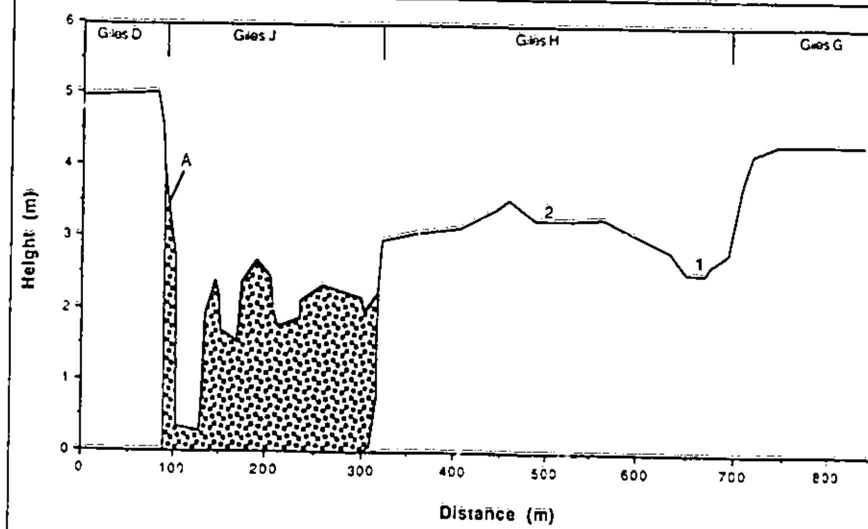
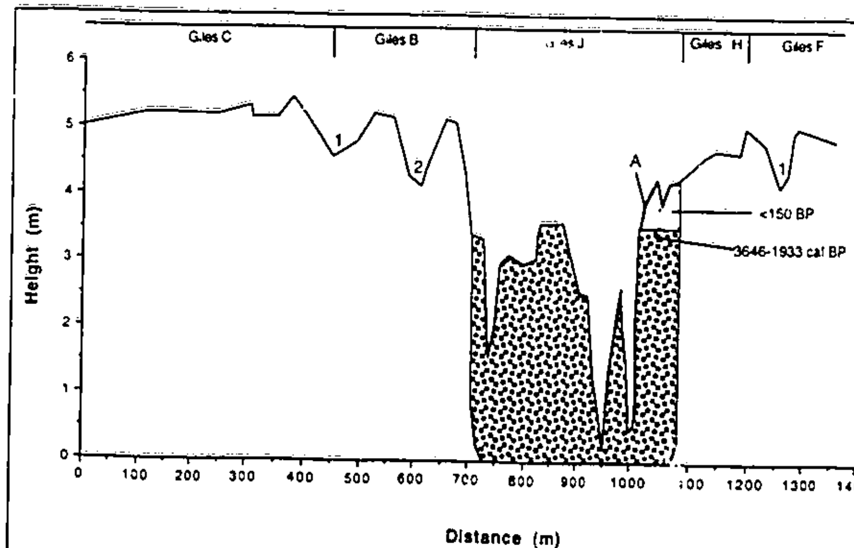


Figure 6.31. Morphostratigraphy of Giles J at Giles Creek 2 site, location in Figure 6.7.



gure 6.32. Morphostratigraphy of Giles J at Giles Creek 3 site, location in Figure 6.7.

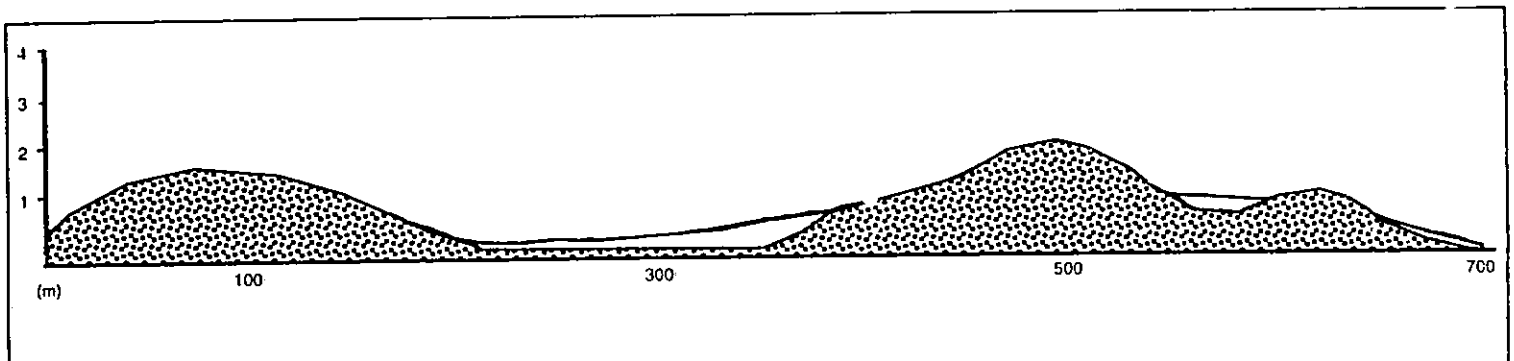


Figure 6.33. Cross profile of longitudinal bar complex on Giles B. Note the vertical axis is estimated.

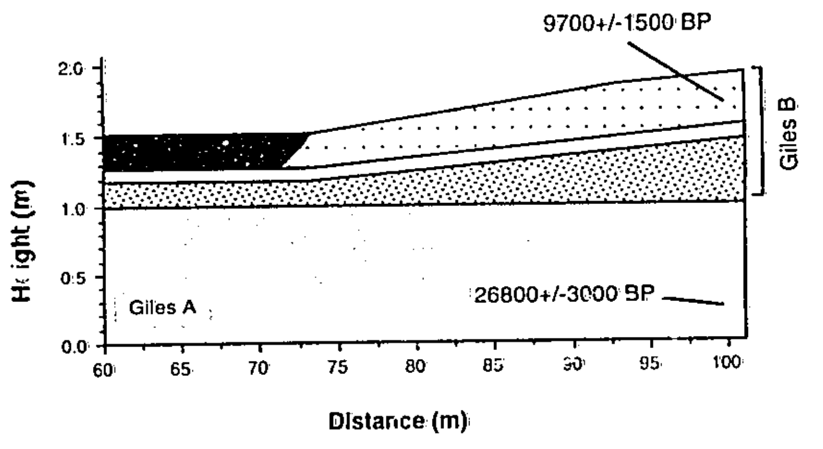


Figure 6.34. Excavation trench in Giles B longitudinal bar, location in Fig. 6.29, Plates 6.7a and b

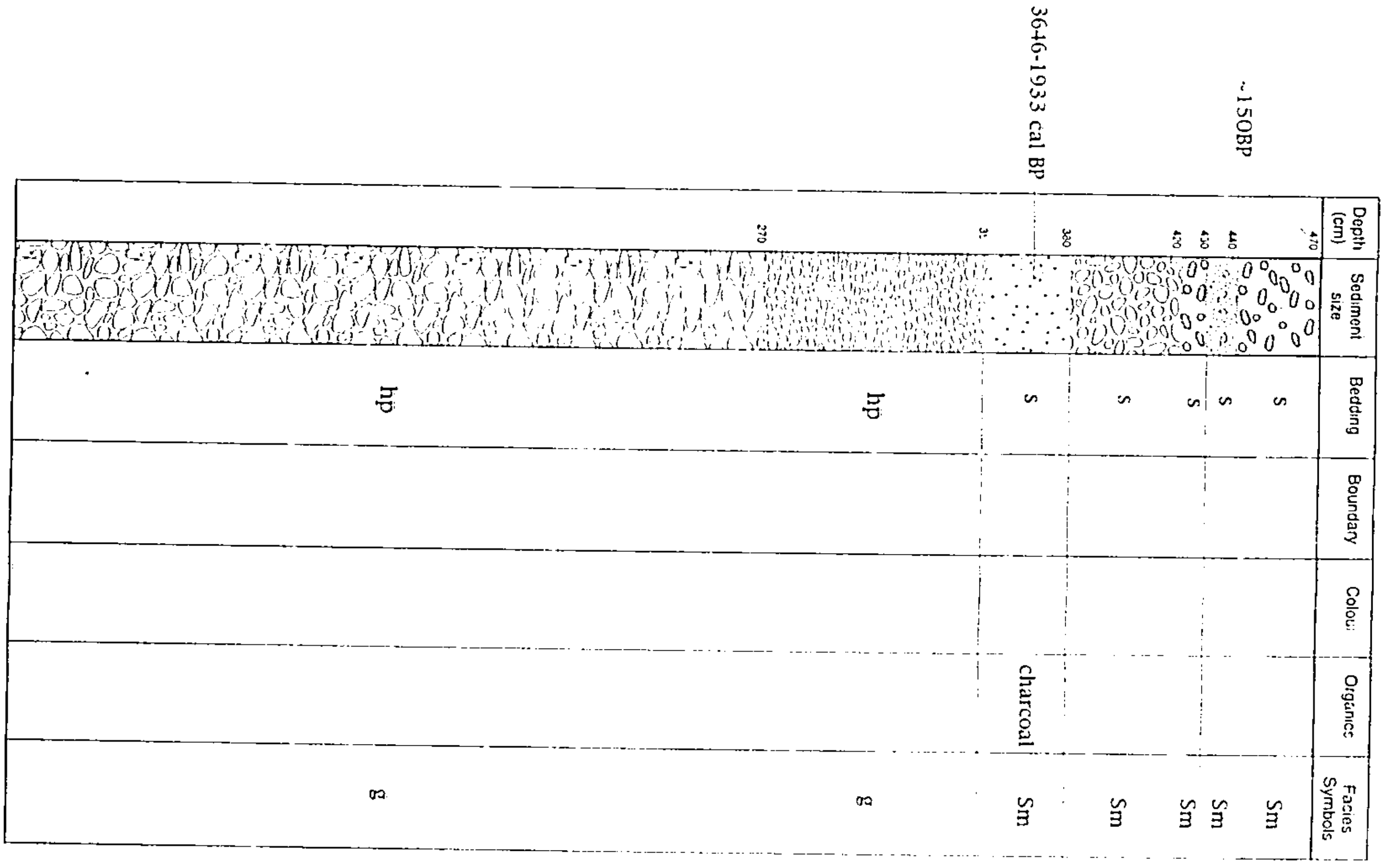


Figure 6.35. Profile A, Giles J paleoflood complex, location in Figure 6.32.

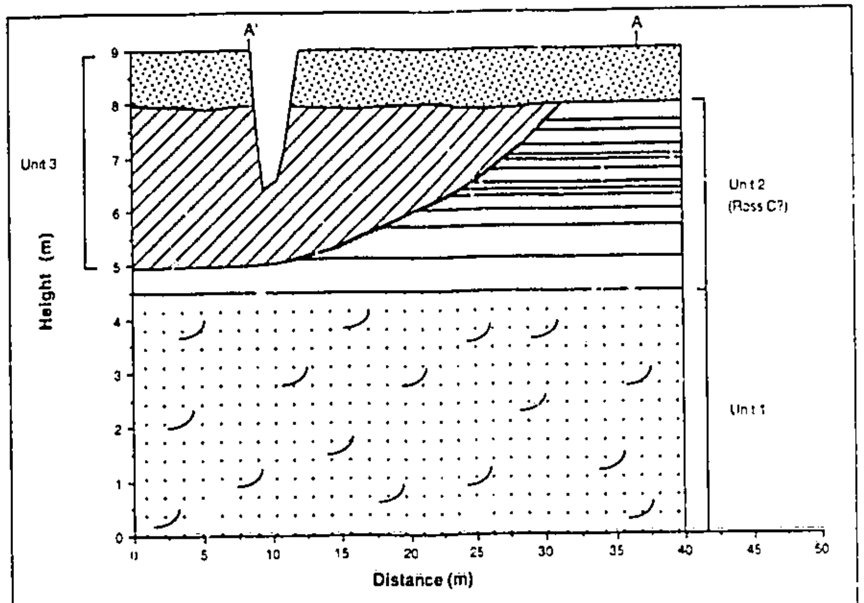


Figure 6.36. Profile of Giles K, location in Figure 6. 7, see text for description.

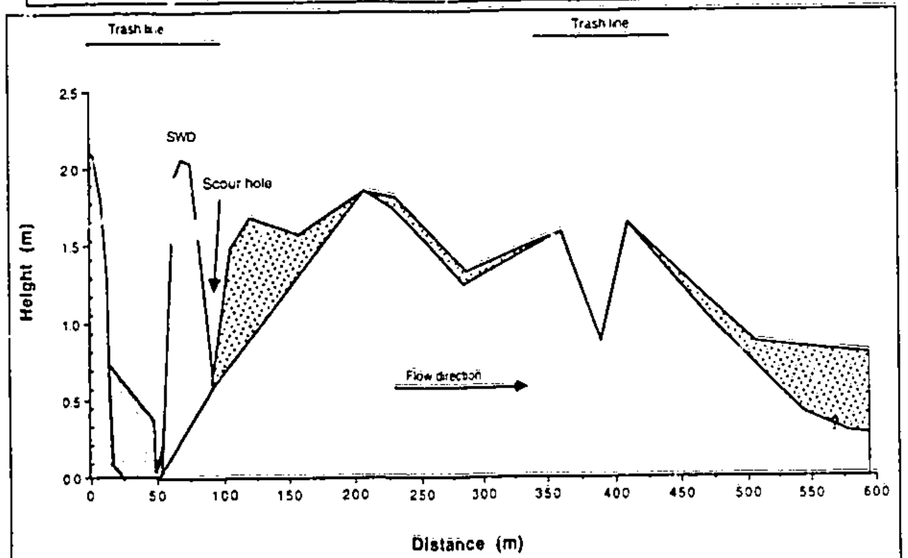


Figure 6.37. Long profile, Todd A distributary channel, location in Figure 6.8.

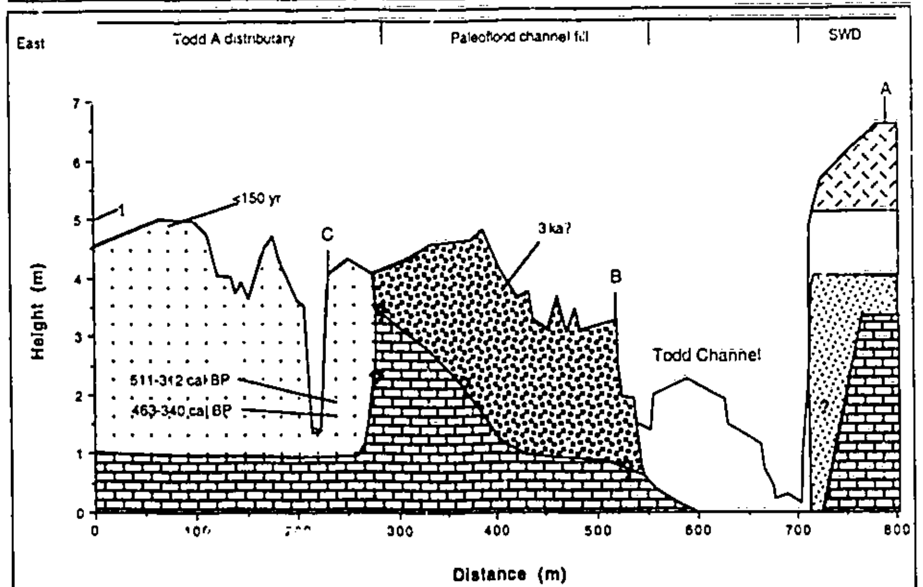


Figure 6.38. Morphostratigraphy of Todd A and associated paleoflood deposits at Mosquito Bore 2 site, location in Figure 6.8. 1 is the approximate elevation of the upper cave SWD

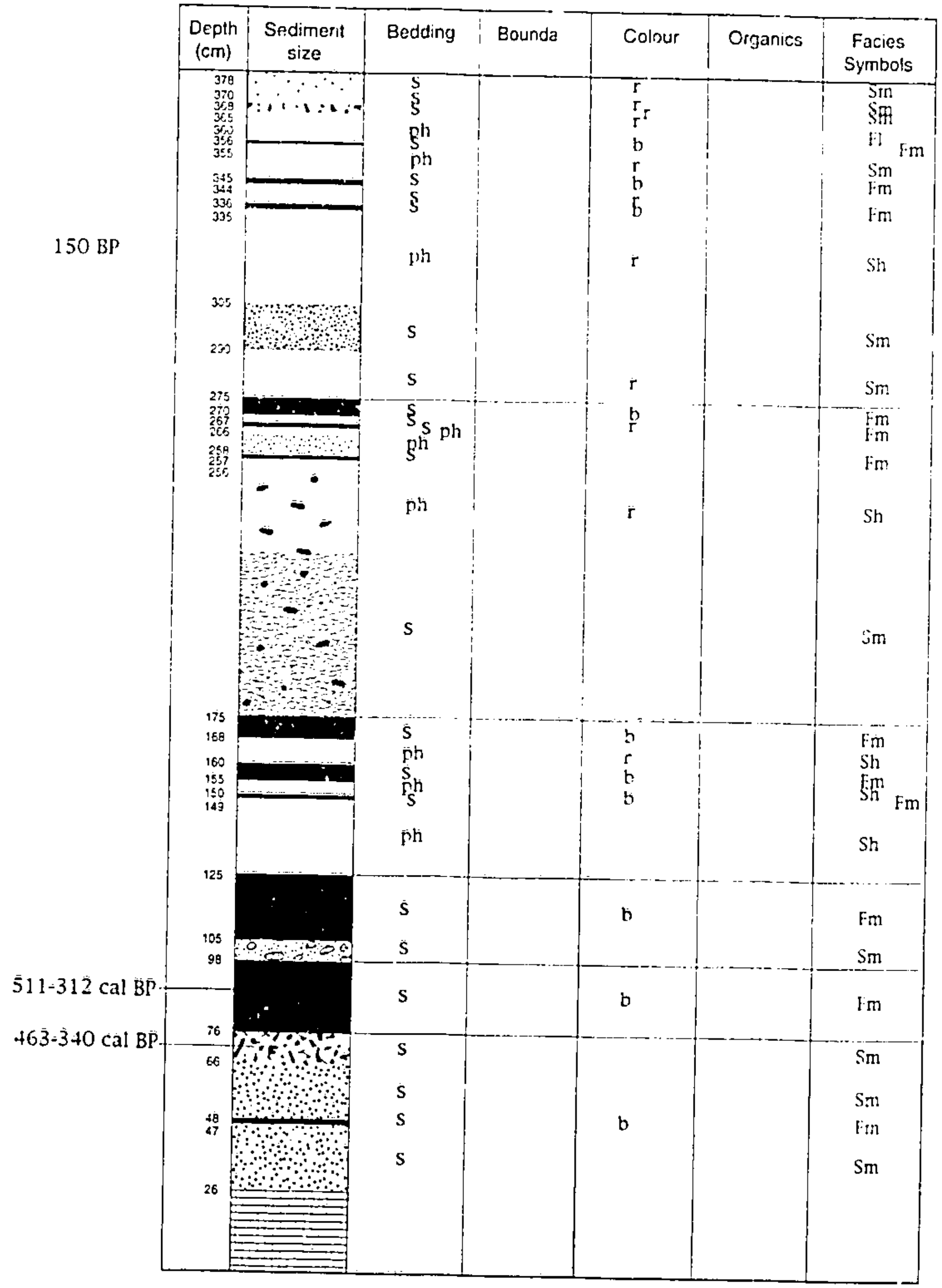


Figure 6.39. Profile C, Todd A paleoflood complex, location in Figure 6.38.

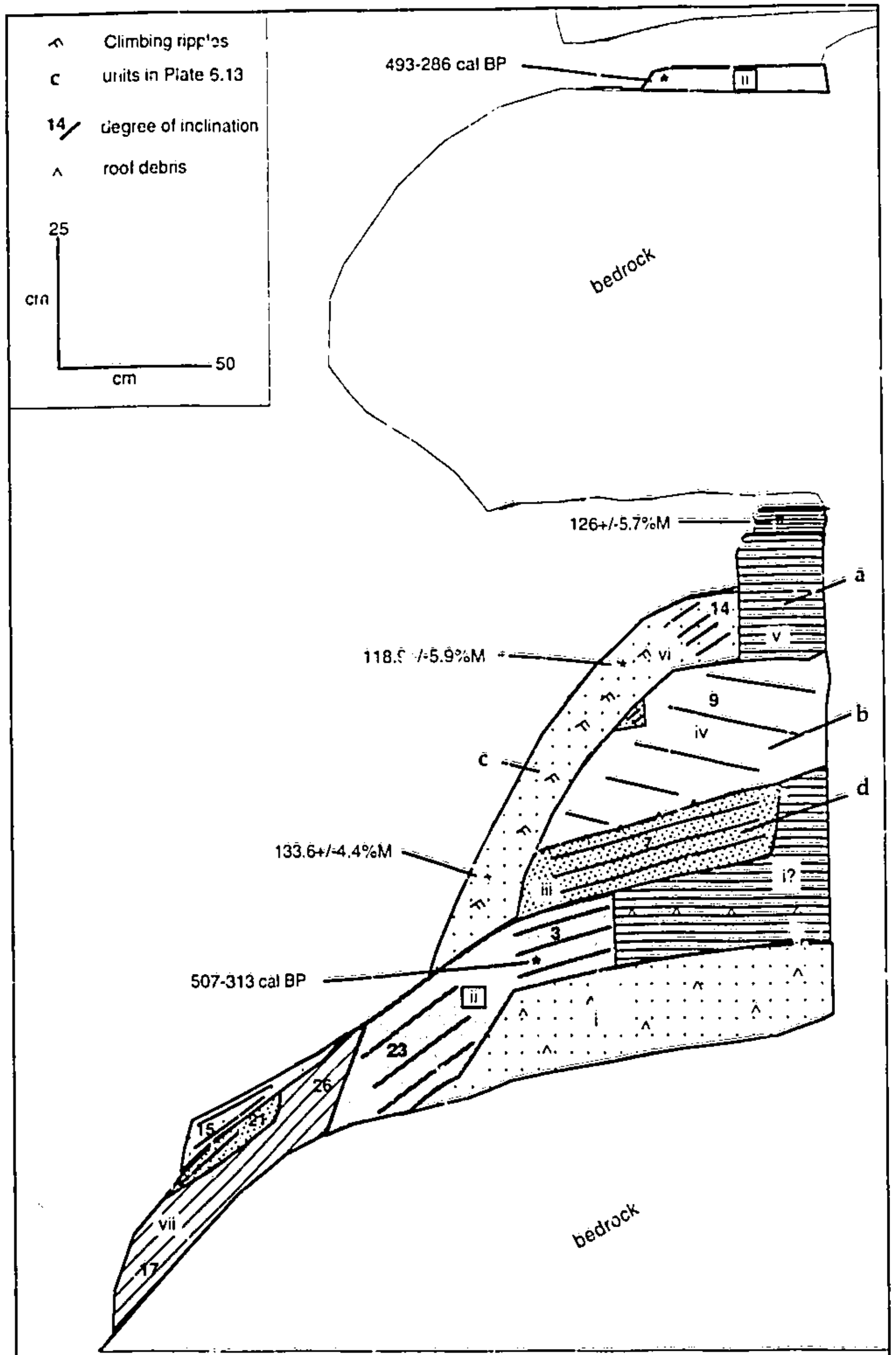


Figure 6.40. Schematic figure of Mosquito Bore cave slack water deposit. Shading indicates differentiation of units on texture, sedimentary structure and cut and fill relationships.

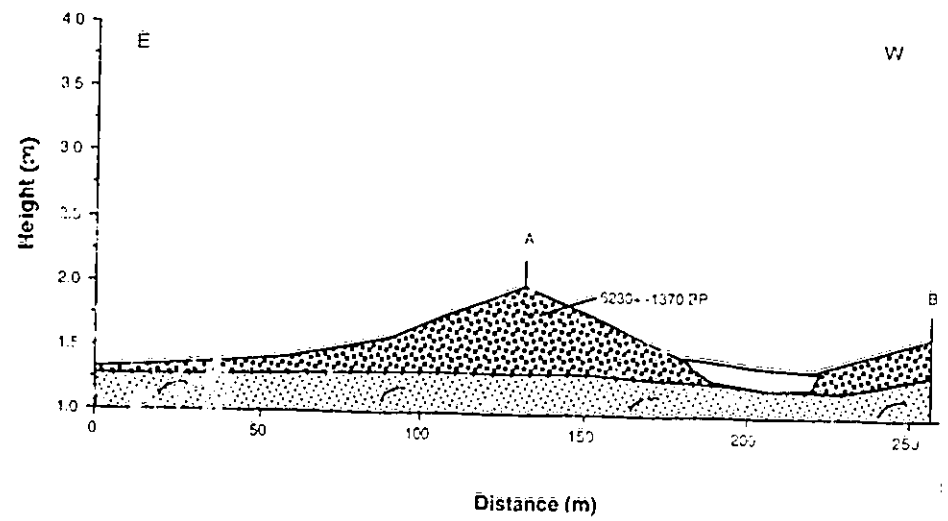


Figure 6.41. Morphostratigraphy of Mosquito Bore Distributary cross section 1, location in Figure 6.9.

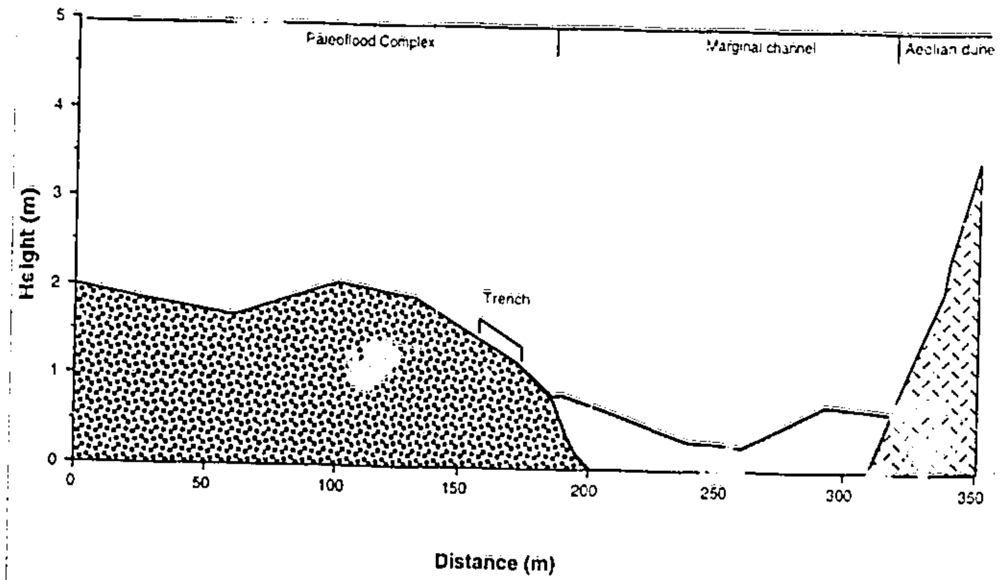


Figure 6.42. Morphostratigraphy of Mosquito Bore Distributary at cross section 2, location in Figure 6.9.

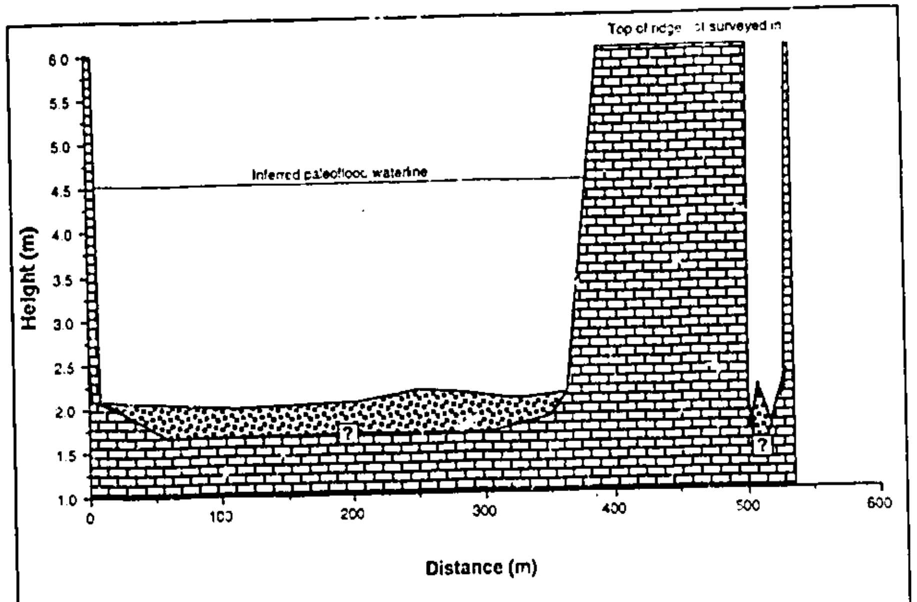


Figure 6.43. Wallaby Gap cross section, location in Figure 6.10.

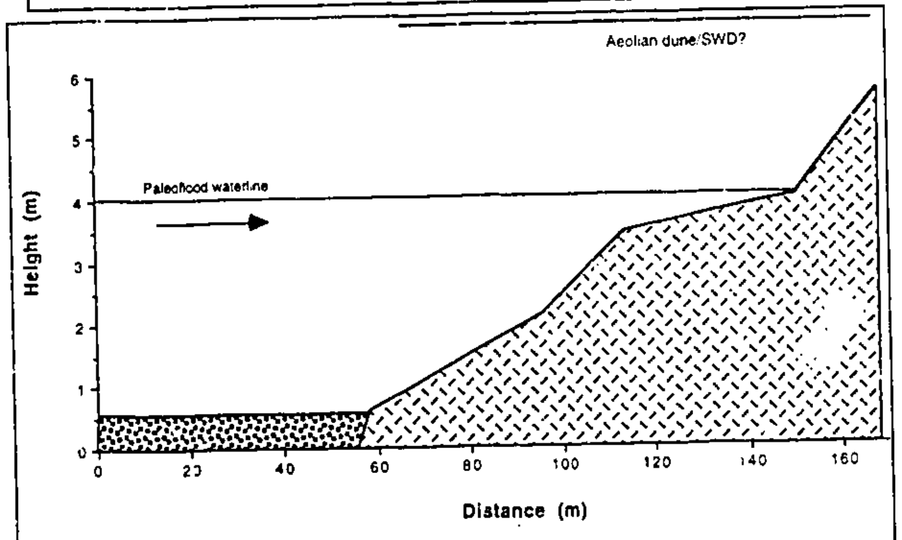


Figure 6.44. Long profile at Wallaby Gap, location in Figure 6.10.

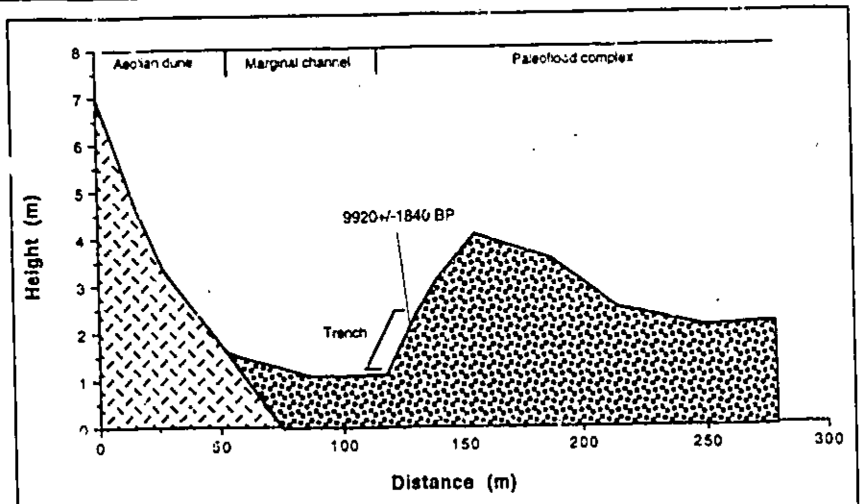


Figure 6.45. Morphostratigraphy of Mosquito Bore Distributary, cross section C, location in Figure 6.11.

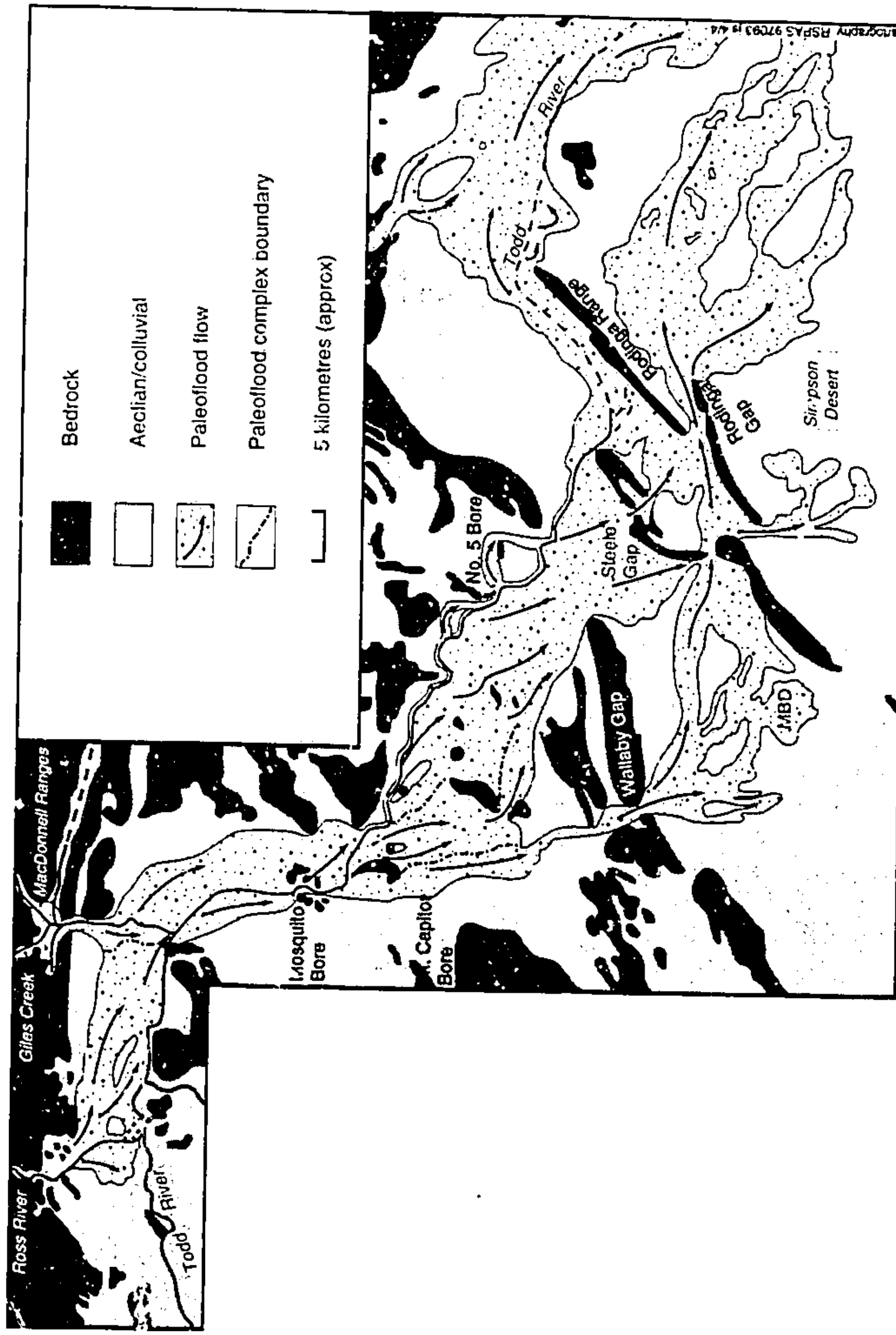


Figure 6.46. Paleoflood geomorphology of eastern Todd catchment.

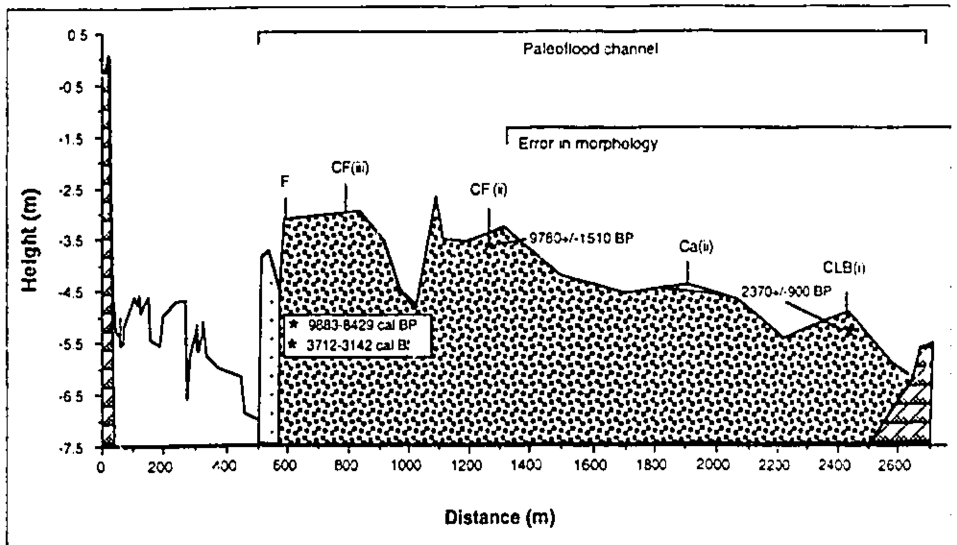


Figure 6.47a. Morphostratigraphy of Todd B paleoflood complex at Expansion Scour site, location in Figure 6.12.

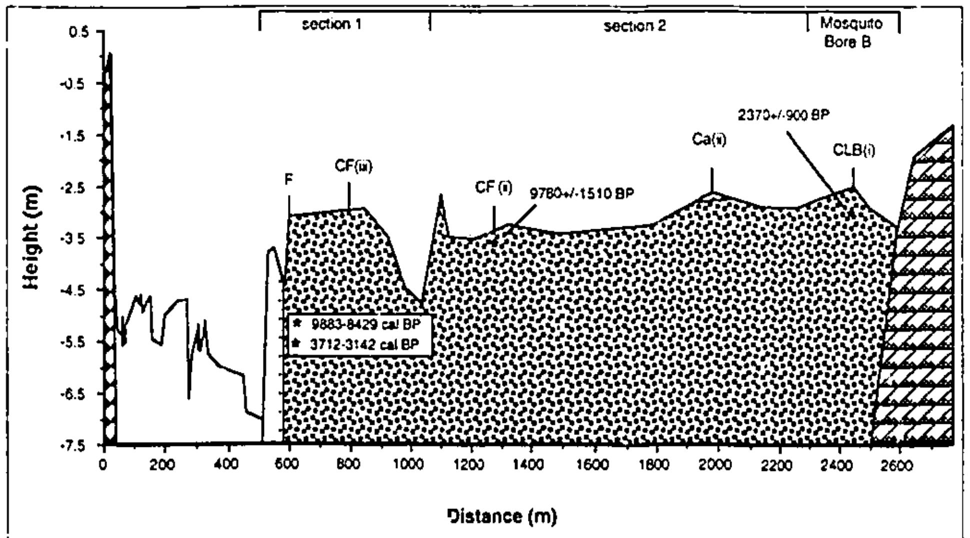


Figure 6.47b. Sketch oı morphostratigraphy of Todd B paleoflood complex at the Expansion Scour site, ıucation in Figure 6.12.

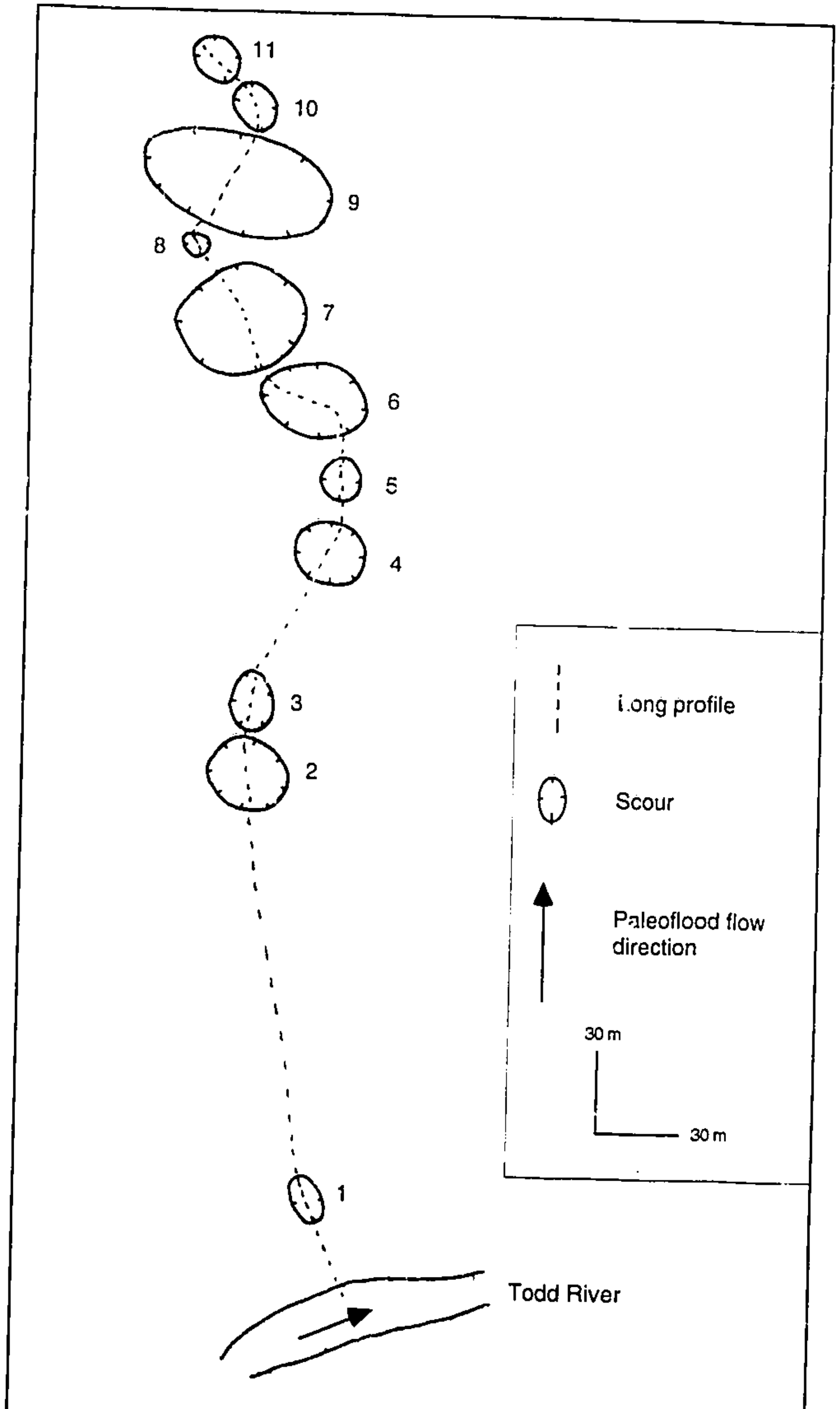


Figure 6.48. No. 5 Bore macroturbulent scour planform, location in Figure 6. 14.

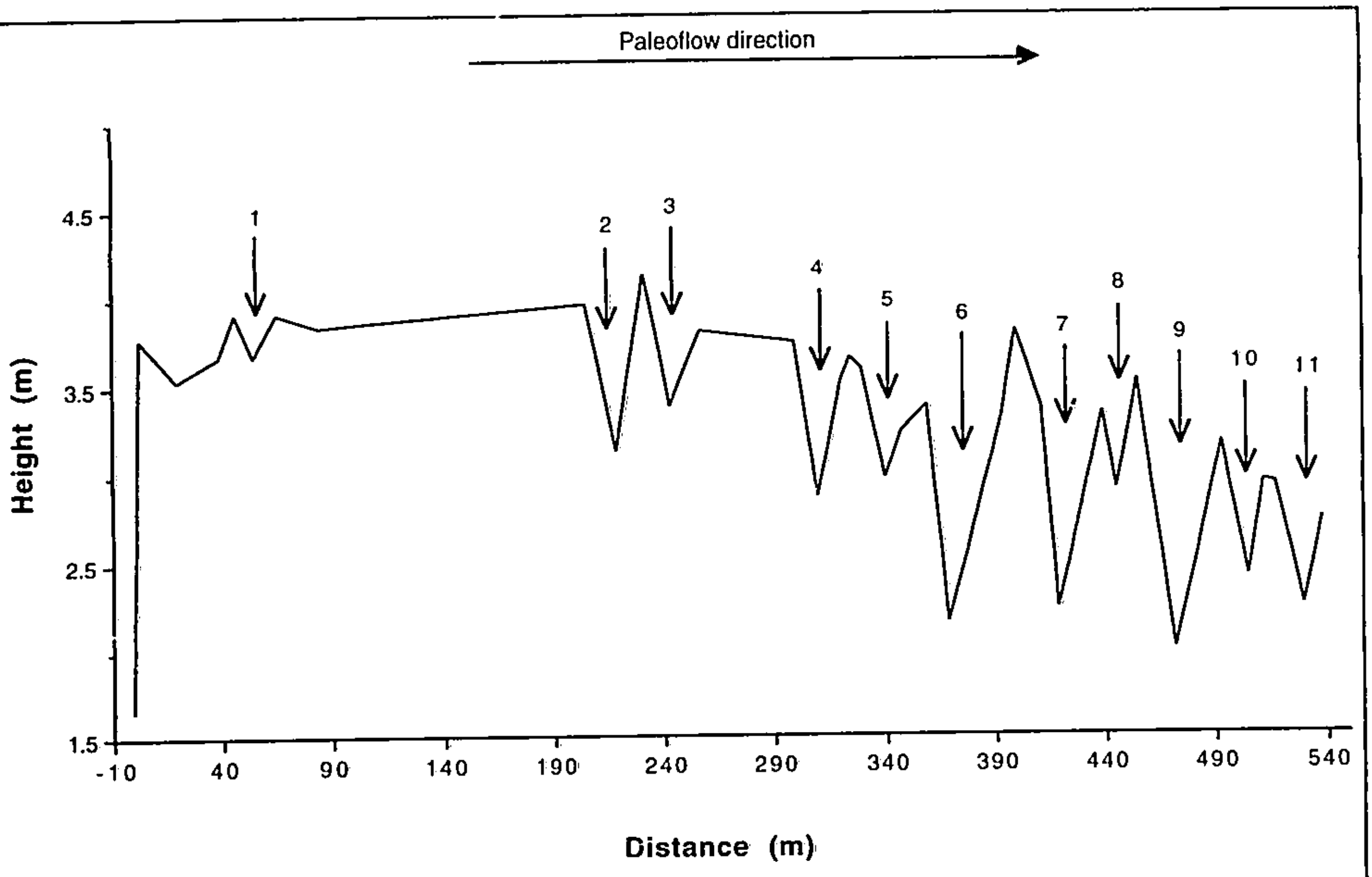


Figure 6.49. Long profile through No. 5 Bore paleoflood macroturbulent scours.

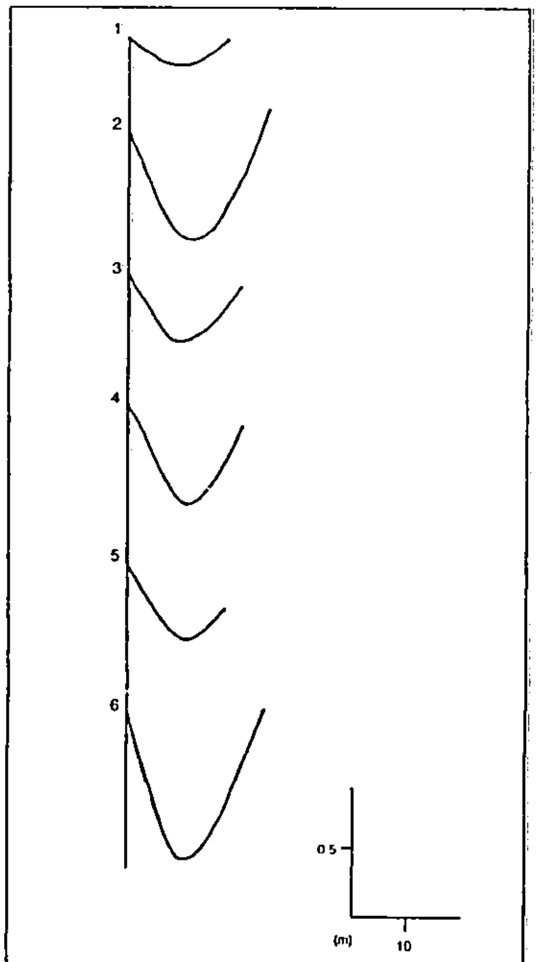


Figure 6.50a. Long profiles of No. 5 macroturbulent scours. Numbers refer to Fig. 6.49

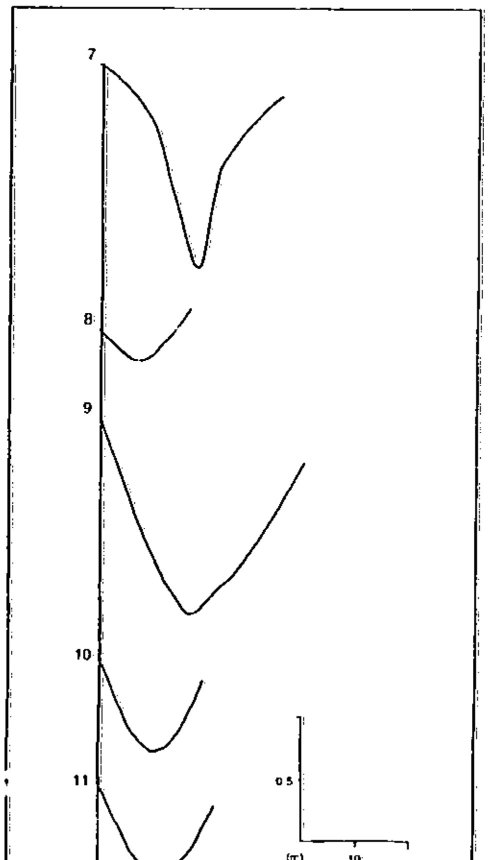


Figure 6.50b. Long profiles of No. 5 macroturbulent scours. Numbers refer to Fig. 6.49.

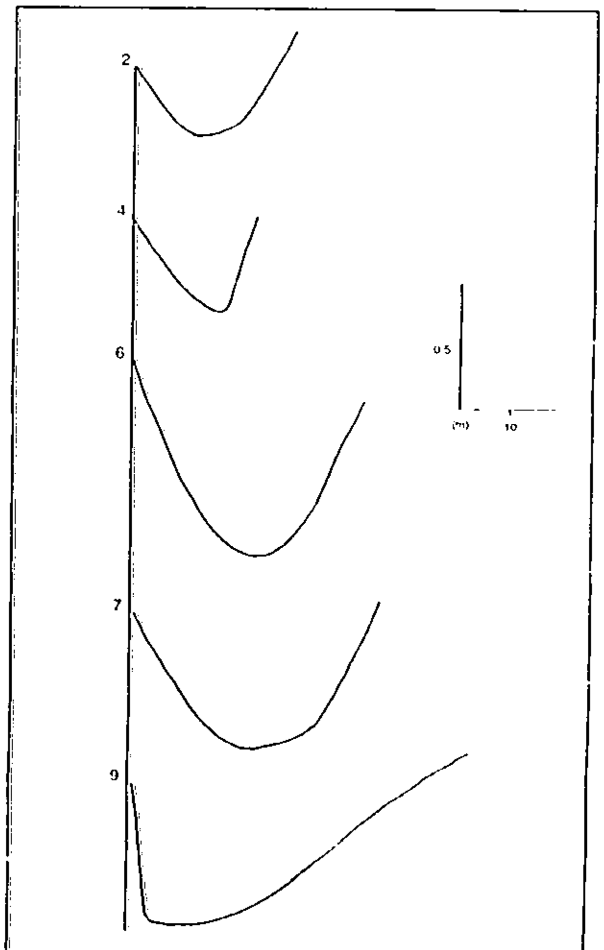


Figure 6.51. Cross section of No. 5 macroturbulent scours. Numbers refer to Fig. 6.49.

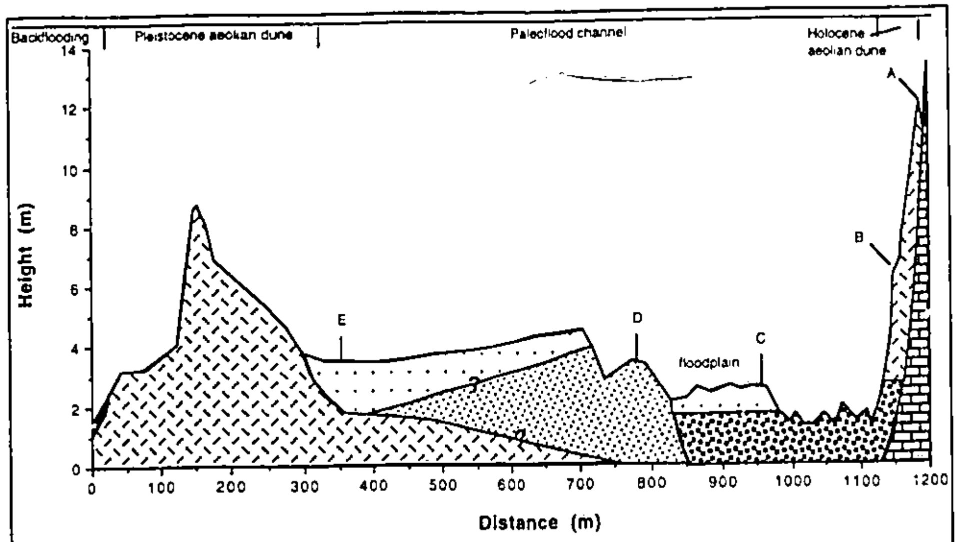


Figure 6.52. Morphostratigraphy of No. 5 Bore paleoflood complex at cross section 5.2, location in Figure 6.15.

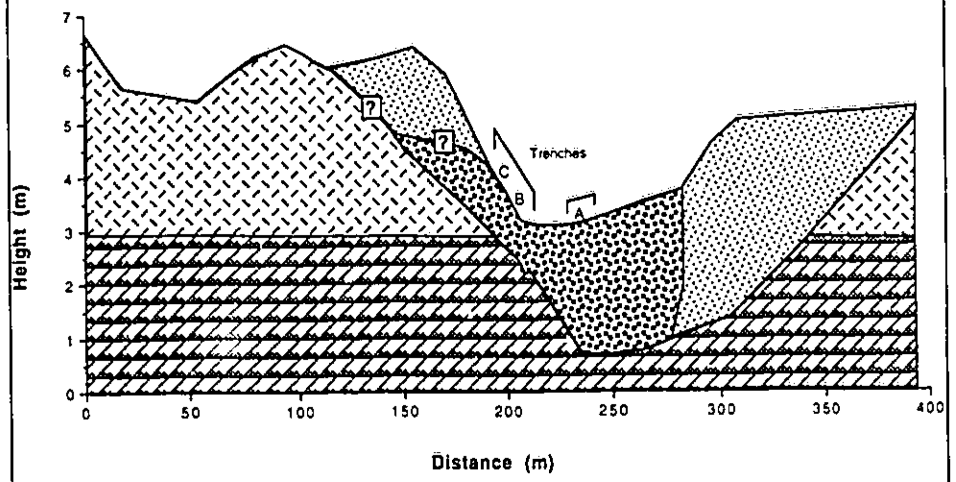


Figure 6.53. Morphostratigraphy of No. 5 Bore paleoflood complex at excavation cross section, location in Figure 6.15.

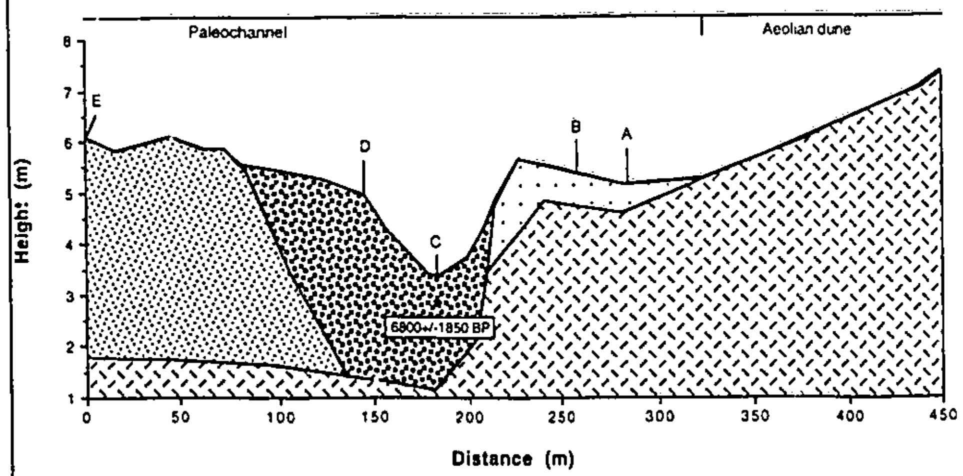


Figure 6.54. Morphostratigraphy of No. 5 Bore paleoflood complex at site 5.1, location in Figure 6.15

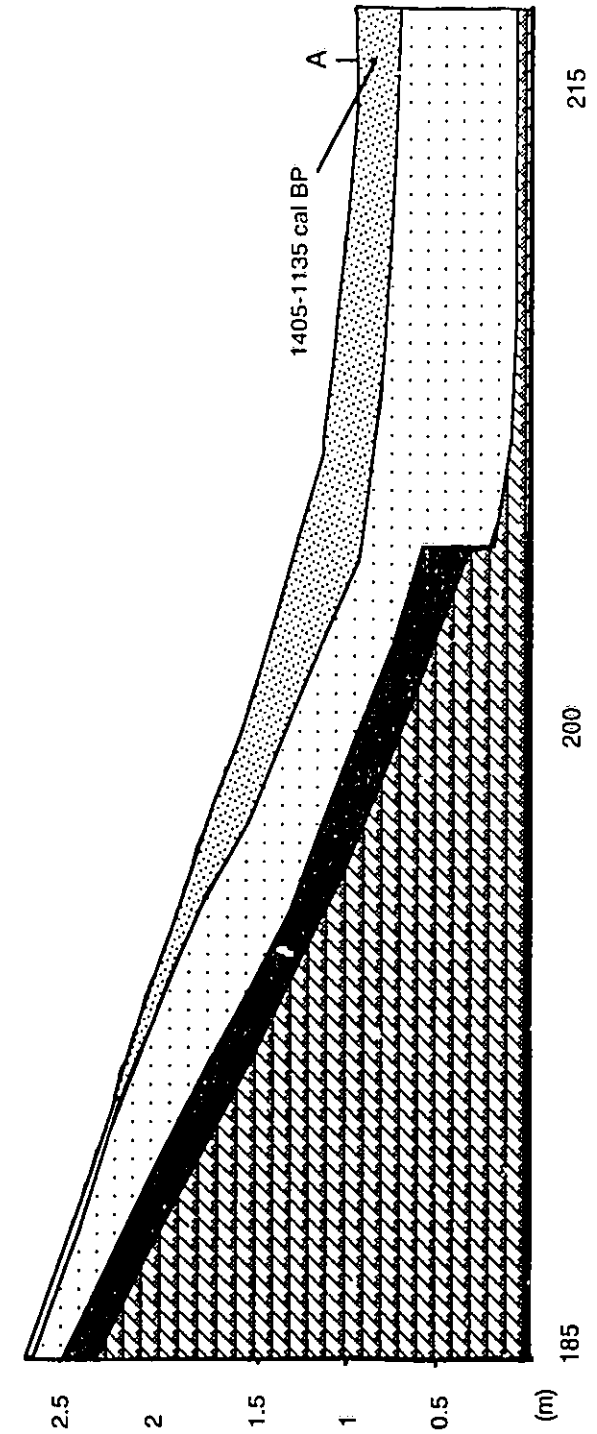


Figure 6.55. Excavation trench, No. 5 Bore paleochannel, location in Figure 6.15.

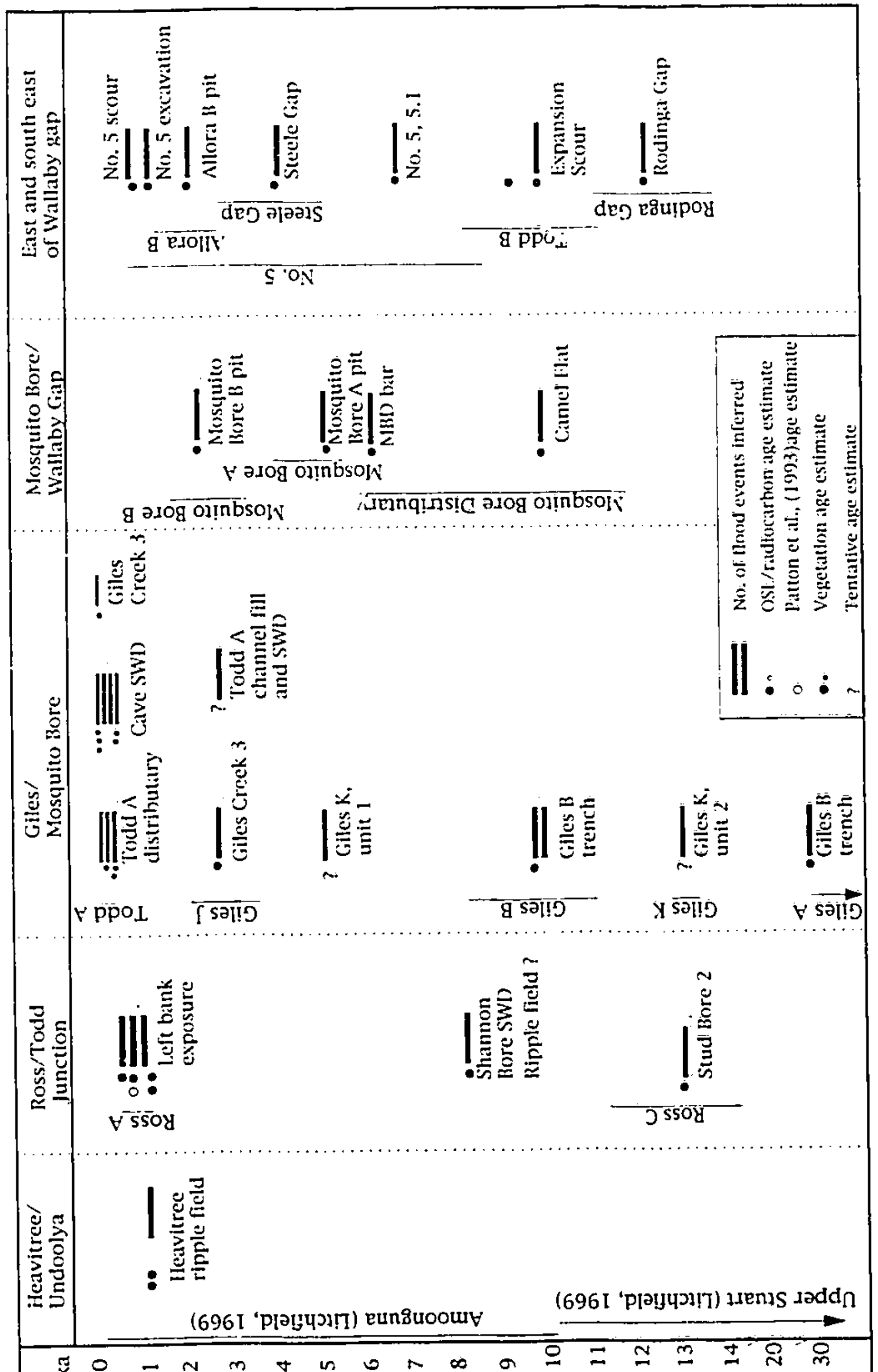


Figure 6.56. Age relationships of paleoflood deposits in Todd Catchment.

Stratigraphic Unit	Age (Pickup, 1991)	Surface (Litchfield, 1969)	Landforms	Sediment
Recent Fluvial Deposit	<10,000 yr with most deposits <2,000 yr	Amoonguna Surface	Overlies Red and Brown Earths, extends considerable distance from modern river. Active flood plains, sand threads and sand sheets shallow channels bars and ripples, erosion cell mosaics	Limited soil profile development
Brown Earth	>10,000 - <59,000	Upper Stuart Surface	Confined to shallow channels and embayments in the ancient wash plains. < 1m thick and form bare crusted surfaces erosion cell mosaics	Fine texture, solonetzic and sometimes contain carbonate nodules
Red Earth	>59,000	Lower Stuart Surface	Extensive, ancient wash plains which underlie flood plains and appear at the surface as low rises. erosion cell mosaics	Varies locally from clayey sand to gravelly sand

Table 6.1. Stratigraphic and surface units associated with central Australian flood plains from Pickup (1991).

Sample Name	Mean (phi)	Sorting	
Heavitree/Undoolya			
Bar 3@20	2.11	1.36	Poorly sorted, fine sand
Ross River			
RR SB SWD @200	2.08	1.87	Poorly sorted, fine sand
EX 5.1	2.01	1.19	Poorly sorted, fine sand
Giles Creek Piedmont			
EX 4.1a	2.41	1.24	Poorly sorted, fine sand
Ex 4.1b	2.7	0.99	Moderately sorted, fine sand
EX 4.1c	2.66	1.19	Poorly sorted, fine sand
EX 4.1d	1.36	2.06	Very poorly sorted, medium sand
Southern Plains			
Cave SWD Ca8	3.20	0.64	Moderately well sorted, very fine sand
MBCC 1.1	2.23	1.72	Poorly sorted, fine sand
EsCh 1.1	2.38	1.53	Poorly sorted, fine sand
EsCh 2.1	2.74	1.25	Poorly sorted fine sand
Mosquito Bore Distributary			
MBD 7	2.01	1.42	Poorly sorted, fine sand
Ex 3.1	1.84	1.32	Poorly sorted, medium sand
Ex 1.1	2.75	.86	Moderately sorted, fine sand
Eastern Systems			
AB 1.1	2.77	1.44	Poorly sorted, fine sand
DPc 1.3.1	2.4	0.88	Moderately sorted, fine sand
E3	2.48	1.01	Poorly sorted, fine sand
RG 1.2	2.67	.99	Moderately sorted, fine sand
E2.2@200	1.55	1.6	Poorly sorted, medium sand

Table 6.2. Statistical parameters of paleoflood sediments, Graphical Method (Folk and Ward, 1957)

Sample Name	ANU ID	Location	^{14}C Age BP	^{14}C Age cal BP	Deposit	Material Dated
Heavitree/Undoolya						
Bar 3 9-13	8833	Heavitree fan	1040±140	1196-691	Transverse bar	Charcoal
Bar 3 9-30	8834	Heavitree fan	125±174	339-0	Transverse bar	Charcoal
A(ii) @ 86	8835	Jessie Quarry	1430±270	1874-886	Aeolian (source bordering)	Charcoal
A2 75-88	8836	Jessie Quarry	2830±70	3087-2775	Hearth in Aeolian (Source bordering)	Charcoal
Ross						
XLi	10060	Ross A	480±60	564-430	Paleoflood	Charcoal
XM	10061	Ross A	600±70	656-522	Paleoflood	Charcoal
XNi	10062	Ross A	970±70	985-724	Paleoflood	Charcoal
XO	10063	Ross A	990±60	987-740	Paleoflood	Charcoal
XK	10059	Stud Bore	450±160	673-248	Paleoflood	Charcoal
Giles Creek Piedmont						
XD+XB	10068	Giles Creek T1	2710±380	3646-1933	Paleoflood	Charcoal
Mosquito Bore area						
C12	9285	mosquito slit cave	320±60	493-286	Paleoflood SWD	Charcoal
CJ	9657	mosquito cave	380±60	507-313	Paleoflood SWD	Charcoal
XI	10064	Todd A	390±60	511-312	Paleoflood	Charcoal
XJ	10065	Todd A	210±140	463-340 340-0	Paleoflood	Charcoal
CB	9650	Expansion Scour	8290±340	9883-8429	Paleoflood	Shell
CA	9649	Expansion Scour	3220±120	3712 3142	Paleoflood	Charcoal
Eastern Systems						
CH	9655	No. 5 Paleochannel	1020±90	1083-726	Paleoflood	Charcoal
XP+[illegible]	10058	No. 5 Paleochannel	1380±70	1405-1135	Paleoflood	Charcoal

Table 6.3. Radiocarbon age of paleoflood deposits in the Todd River.

Sample	ANU$_{OD}$ ID	Site Name	Complex	OSL Age	Error (±)	Deposit Type
Heavitree/Undoolya						
B3	170	Bar 3, Airport	Heavitree	1,030	420	Transverse bedform
Ross River						
42	171	Shannon Bore	Ross River Gorge	8,350	950	Tributary mouth mounded SWD
Ex 5.1	161	Ross	Ross C	12,900	1,800	Paleoflood Channel
Giles Creek Piedmont						
Ex 4.11	160a	New Yard	Giles B	9,700	1,500	Longitudinal Bar
Ex 4.13	160b	New Yard	Giles A	26,800	3,000	Longitudinal Bar
Southern Plains						
EsCh 1.1	164	Expansion Scour	Giles D	5,160	1,340	Paleoflood Channel
CLB (1)	165	Expansion scour	Giles E	2,370	900	Paleoflood Channel
Mosquito Bore Distributary						
MBD7	162	MBD	MBD	6,230	1,370	Paleoflood Channel
EX 1.1	168	Camel Flat	MBD	9,920	1,840	Paleoflood Channel
Eastern Systems						
CF (ii)	163	Expansion Scour	Todd B	9,780	1,510	Paleoflood Channel
AB 1.1	166	Allora Bore	Allora B	2,110	760	Paleoflood Channel
DPc 1.3.1	169	No. 5 Bore	No. 5	6,800	1,850	Paleoflood Channel
E3	167	Steele Gap (SG)	SG-RG	3,900	1,270	Paleoflood Channel
Rg 1.2	172	Rodinga Gap (RG)	SG-RG	12,310	1,400	paleoflood Channel

Table 6.4. OSL age of paleoflood deposits.

Site	Distance Downstream (km)	Width (m)	Depth (m)	Texture
Giles Creek 1	2.5	410	>3.5	Cobbles, coarse gravels
Giles Creek 2	3.4	230	>4	Cobbles, coarse gravels
Giles Creek 3	4.5	390	>3.6	Cobbles, coarse gravels

Table 6.5. Giles J paleoflood channel fill dimensions.

Macroturbulent Scour Number	Maximum Length (m)	Maximum Width (m)	Maximum Depth (m)
1	18.5	NA	0.24
2	26	30	0.99
3	21	NA	0.475
4	21	23	0.87
5	18	NA	0.6
6	26	43	1.22
7	35	47	1.57
8	16.5	NA	0.6
9	38	74	1.51
10	19	NA	0.76
11	21	NA	0.825
Average	23.6	43.4	0.88
Max	38	74	1.57
Min	16.5	30	0.24

Table 6.6. Macroturbulent scour dimensions

Site	Gradient (m/m)
Heavitree fan	.0024
Undoolya fan	.0048
Ross C paleoflood complex	.0013
Giles A paleoflood complex	.0022
Giles B paleoflood complex	.0002
MBD paleoflood complex reach 2	.0016
MBD paleoflood complex reach 3	.00097
Allora B	.0013
Steele Gap to Rodinga floodout	.00085
a) Steele Gap to Rodinga Gap	.00164
b) Rodinga Gap area	.00107
c) Mid to lower floodout	.0016[illegible]

Table 6.7. Pal[illegible]lood gradients.

Paleoflood system	Number of events	Age of events (Ka BP)	Landforms	Sediments
Heavitree/Undoolya	>3	~1 15-10 30-15	Transverse bar, distributaries, bar fields, avulsions	Coarse cobbles to silt
Ross River A (Patton *et al.*, 1993)	<5	~0.7 <1.5	Wide braided paleochannel, longitudinal bars.	Unweathered sand and gravel
Ross A	~5	~0.5-0.9	Wide braided paleocnannel, longitudinal bars.	Unweathered sand and gravel and cobbles.
Ross C	~2	~8-13	Slack water deposit large channel transverse bar field.	Weathered sand gravels and cobbles.
Giles A	1?	~ 27	Extensive area of inundation, dissected by shallow wide channels	Gravelly sand
Giles B	~3	~10	Large splay channel with longitudinal bars (0.5-2 m high) with transverse bars superimposed.	Gravelly sand
Giles J	1 2	~3 < 0.15	Channel fill 1 m of flood plain aggradation.	Cobbles and coarse gravel.
Todd A Distributary Channel	1 ~6	0.4 <0.4	Paleoflood distributary, 3 m aggradation	Small gravels, sand, mud.
Cave SWD	~2 5	0.4 <0.4	Slack water deposit in cave	Very fine sand
Mosquito Bore Distributary	1	6	Braided channel complex 2.5 km wide, longitudinal bars 70 cm high interspersed with shallow channels 50 m wide, eroded longitudinal sand dunes.	Gravelly sand, finer in channels.
Mosquito Bore Distributary at Camel Flat Bore	1	10	Splay complex, longitudinal bar, eroded longitudinal sand dunes.	Gravelly sand
No. 5 paleochannel scours No. 5 Excavation	1 1	1 1 1	Channel, splays, macroturbulent scours, avulsion. Channelr	Cobbles, gravel, coarse sand. Gravelly coarse sand
No. 5 paleochannel, 5.1		7	Channel	gravels in coarse sand
Mosquito Bore A	1	5	500-700 m wide channel	Coarse sand
Mosquito Bore B	1 1	2 >2	500-700 m wide channel	Coarse sand
Todd B Expansion scour	1	9	Wide channel	Gravelly sand
Allua B	1 1	2 >2	Erosion of dunes	Fine sand and silts
Steele Gap	1	4	Erosion of dunes, splay channel	Gravelly sand
Rodinga gap	1	12	Erosion of dunes, splay channel.	Gravelly sand

Table 6.8. Age and number of flood events in Todd Catchment

Plate 6.1a. Core from Heavitree Site 2, Units A and B at base of core.

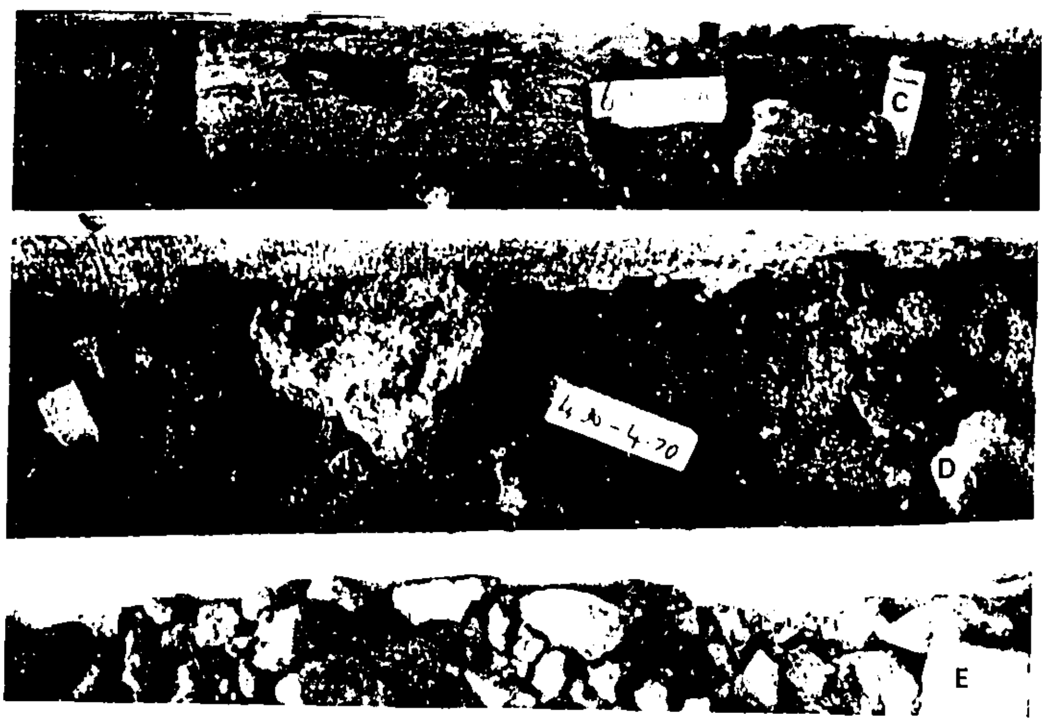


Plate 6.1 b. Core from Heavitree Site 2, Units C, D and E towards top of core.

Plate 6.2. Oblique view across Fmily and Jessie piedmont fans. (a) Jessie Creek, (b) Emily Creek (c) climbing dune (Jessie Quarry).

Plate 6.3. An oblique view across the Ross paleoflood complex. (a) is an aeolian sand sheet with low hummocky dunes, (b) is the Ross B paleoflood complex, (c) is the Ross A paleoflood complex and (d) is a streamlined remnant.

Plate 6.4. An oblique view of the Ross River Gorge slack water deposit (a), located at the mouth of a strike valley tributary. Location in Fig. 6.5.

Plate 6.5a. The top 3.5 m of Ross C deposits at Stud Bore 2. Note the fine texture and weathering of the profile.

Plate 6.5b. The lower portion of section through the Ross C deposits at Stud Bore 2. Note the coarse cobble framework gravels at the lower right. (a) indicates a flood plain inset on an eroded bench in the paleoflood sediments.

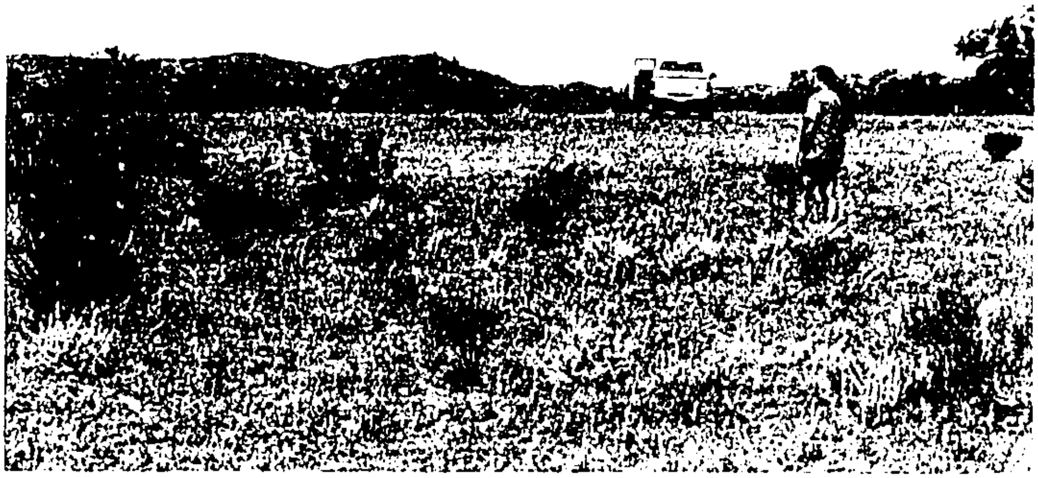


Plate 6.6a. View looking north east across a 1 m high longitudinal bar on Giles B.

Plate 6.6b The coarse texture of sediment in longitudinal bars on Giles B.

Plate 6.7a. Trench excavated in left flank on a longitudinal bar on Giles B.

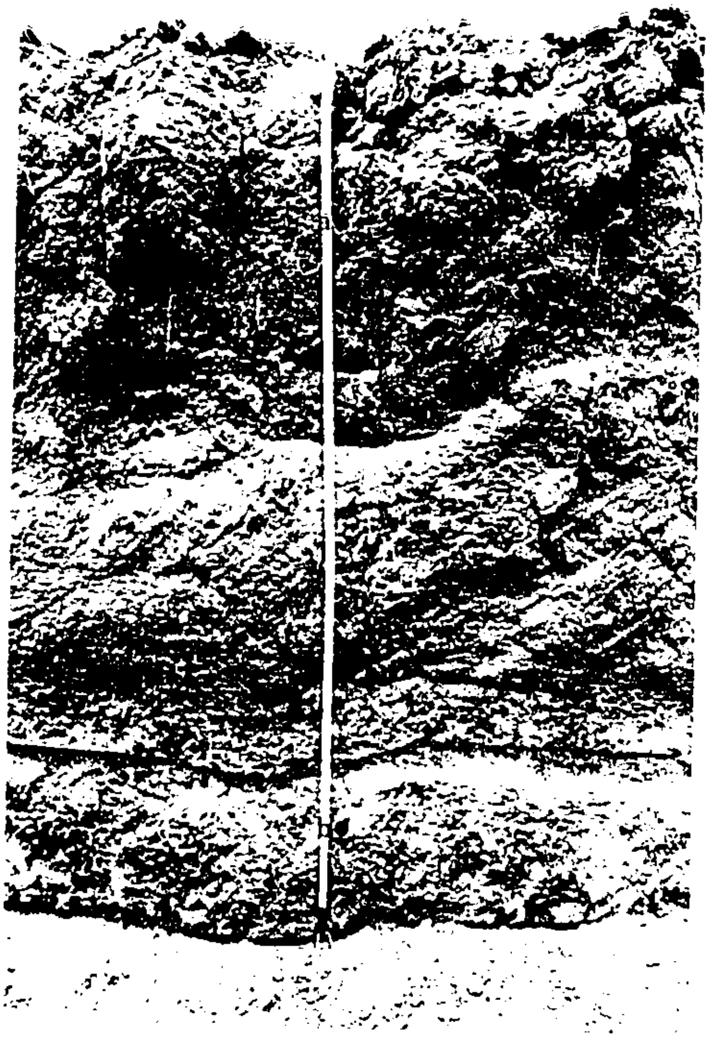


Plate 6.7b. Longitudinal bar flank sediments. a and b are the locations of OSL samples.

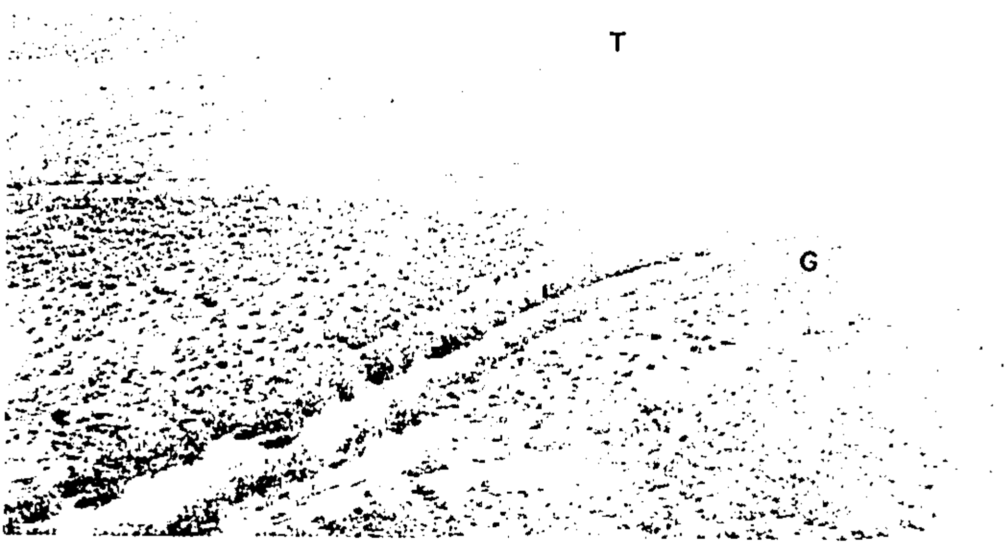


Plate 6.8. Oblique aerial view looking upstream along the Todd (T), across Giles Creek (G). Flow is to the left of the photo and Giles K is to the immediate right of the plate. It is inferred that the confluence was located in the centre of the photo in the past.

Plate 6.9. Giles K site. The exposed roots of the River Red Gum indicate the amount and rate of channel incision. a) Unit 1, b) Unit 2, c) Unit 3.

Plate 6.10. An oblique aerial view looking north east at Mosquito Bore (a). The Todd A paleoflood distributary (b) drains towards the lower right of the plate and flow of the Todd channel is to the bottom right of the plate.

Plate 6.11. View looking east across the Todd A distributary 2.4 km from intake. Note the subdued morphology, fine sediment texture and the buried Coolabah trees. Distributary flow is to the right.

Plate 6.12. The Mosquito Bore 1 slack water cave deposit is located in the shade behind the young River Red Gum. Note the fresh sediment deposited outside the cave after the January 1995 event. Flow is to the left.

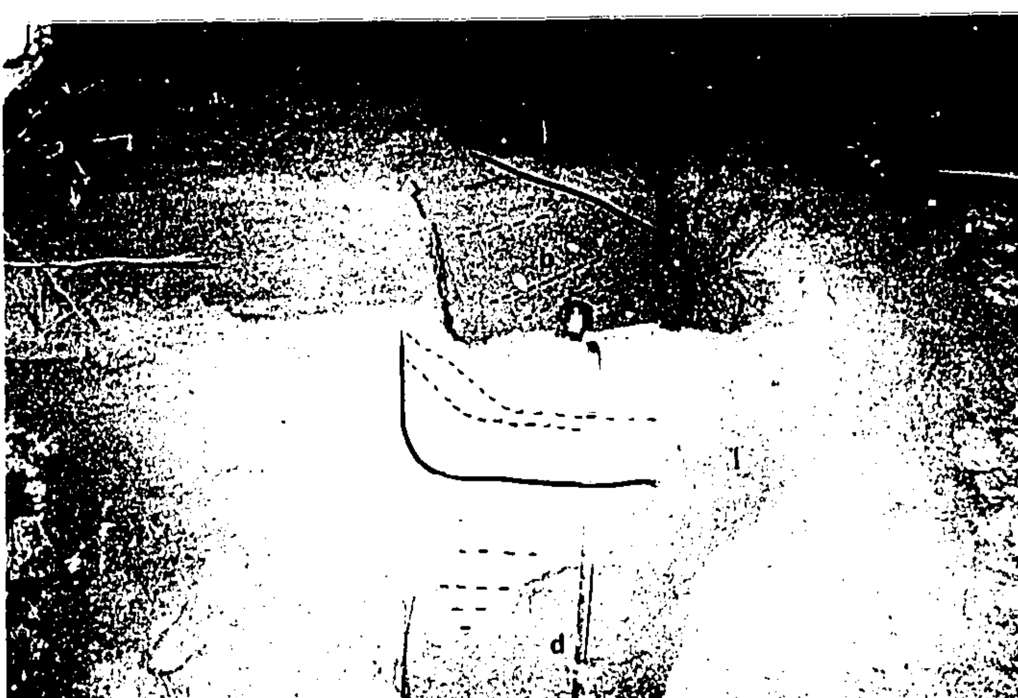


Plate 6.13. The slack water sediments in the Mosquito Bore cave. Note the roof fall boulder to the right of the photo. a-d are units identified in Figure 6.39, the dotted line indicates layer dip and the solid line the boundary between deposits.

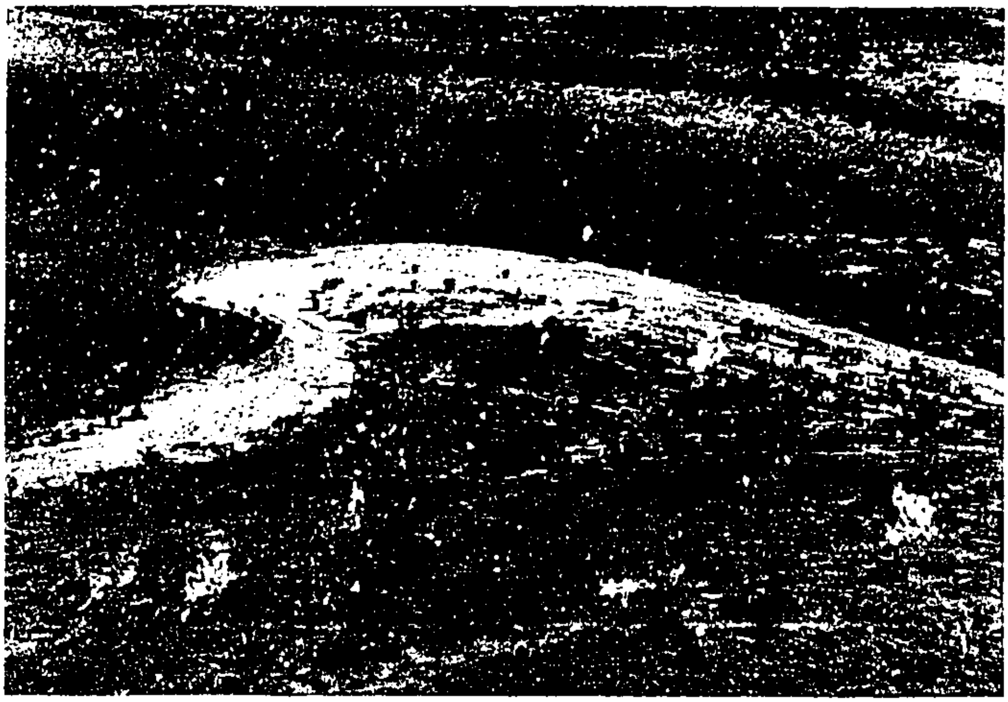


Plate 6.14. Aerial view looking south west at the sharp meander bend on the Todd River south of Mosquito Bore. Vegetation picks out the Mosquito Bore Distributary paleoflood complex.

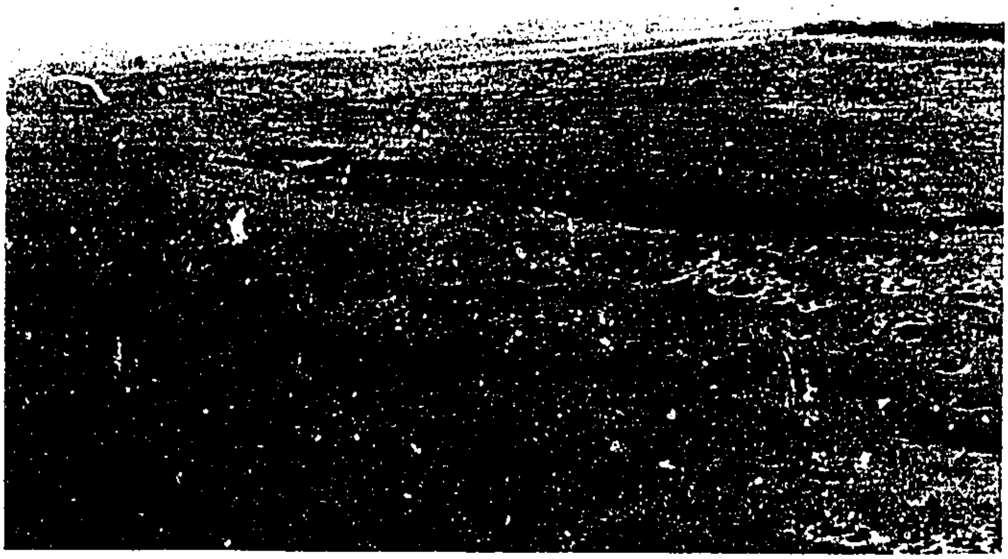


Plate 6.15. Aerial view looking south west across the Mosquito Bore Distributary. The green vegetation in the marginal channel (a) picks out the boundary between the aeolian dune field (b) and the paleoflood complex (c).

Plate 6.16. No. 5 paleoflood channel macroturbulent scour (a).

Plate 6.17. No. 5 paleochannel. Looking upstream along the climbing dune nourished from the paleochannel and reworked by local runoff from the ridge.

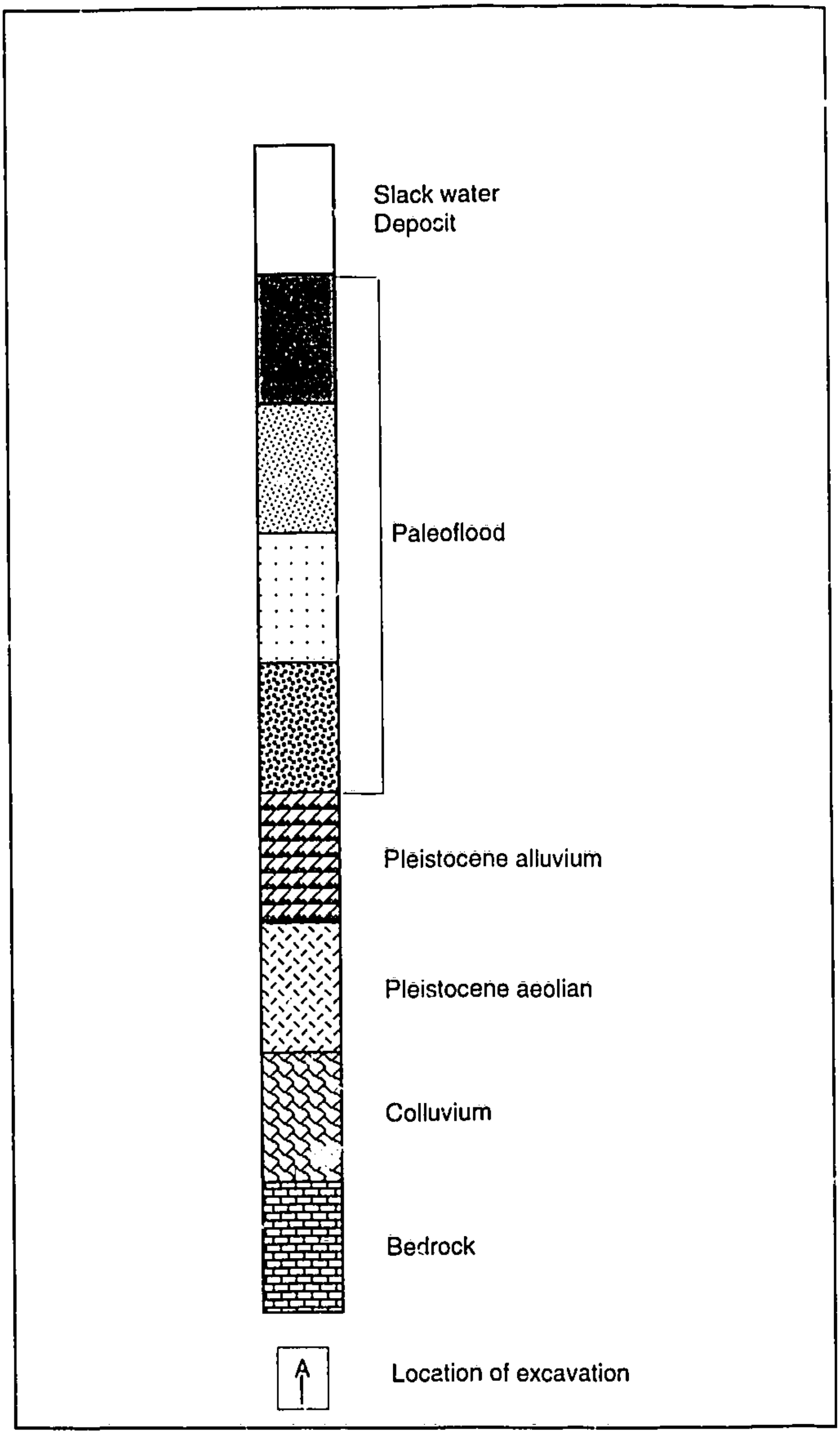
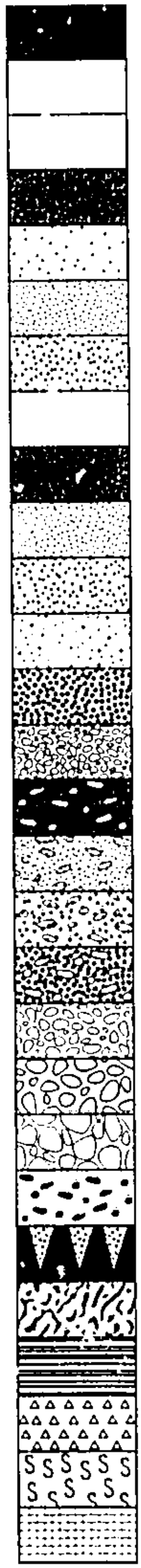


Figure 6.57. Legend for paleoflood morphostratigraphic figures.



Mud
Very fine sand
Fine sand
Silty fine sand
Silty medium sand
Medium sand
Silty medium and coarse sand
Medium and fine sand
Silty coarse sand
Medium and coarse sand
Coarse sand
Fine and coarse sand
Very coarse sand
Fine gravel
Fine gravel in silty sand
Gravel in medium sand
Gravel in coarse sand
Gravel in very coarse sand
Medium gravel
Coarse gravel
Cobbles
Mud balls in fine sand
Dessication cracks
Soil/organics
Sand with mud laminations
Colluvium
Pleistocene sediment
Bedrock

Bedding		Colour		Boundary	
d	discontinuous	r	red	E	erosional
h	horizontal	w	white		
i	inclined	b	brown		
s	structureless	g	grey		
p	parallel				
r	ripple				

Horizontal line indicates inferred break in deposition

Figure 6.58. Legend of symbols used in flood plain and paleoflood profiles.

CHAPTER 7

Paleoflood Geomorphology and Chronology

7.1. Introduction

The aim of this chapter is to interpret the data presented in Chapter 6. The first section synthesises the paleoflood depositional and erosional landforms, compares them to similar forms described elsewhere, and infers paleoflood processes and magnitude. The second section interprets the chronological data set and compares it with results from other Australian and tropical studies.

7.2. Review of Australian Paleoflood Studies

Awareness of flood hazards in central Australia extends beyond the instrumental record. There are Dreamtime stories of large raging floods, and there is anecdotal evidence of two Aboriginal men washed from a cave 3 m above Emily Creek (J. Hayse, *pers. comm.*). Paleofloods are recognised in the following description of the site of Alice Springs in 1901:

> On the first survey flood marks were found which showed that the whole country to the north of the range must have been converted into a huge lake 15 feet deep, and the blacks tell that before the white men came they had once to take refuge from the flood on the hill. (White, 1909, p.34)

Paleoflood research in Australia has sampled a wide geographical area from the tropical river systems in Queensland (Burdekin River, Wohl, 1992a; Herbert River, Wohl, 1992b), northern Australia (East Alligator River, Wohl *et al.*, 1994a; Waterfall Creek and Wangi Falls, Nott and Price, 1994; Nott *et al.*, 1996; Katherine River, Baker and Pickup, 1987), and the Kimberley Ranges (Lennard River, Gillieson *et al.*, 1991; Fitzroy, Brooking, and Margaret Rivers Wohl *et al.*, 1994a; and Piccaninny Creek, Wohl, 1993), to the arid zone (Finke River, Pickup *et al.*, 1988; Ross River, Patton *et al.*, 1993; Todd River, Pickup, 1991, Bourke and Pickup, in press; Bourke, in press; and Sandy Creek, Jansen *et al.*, 1996), and in south eastern Australia, (Nepean River, Saynor and Erskine, 1993). Much of this work has concentrated on flood paleohydrology and chronology (Baker *et al.*, 1983; Baker *et al.*, 1985; Pickup *et al.*, 1988; Wohl *et al.*, 1994) and therefore the focus has principally been on the gorge reaches of river systems where evidence of flood stage, such as slack water deposits and high stage gravel deposits are preserved. This has led to the recognition of the importance of high magnitude floods in determining gorge channel morphology and sedimentology (Baker and Pickup, 1987; Pickup *et al.*, 1988; Wohl, 1992a, 1992b).

Very large floods generate sufficient stream power to erode bedrock channel boundaries. One spectacular example is scabland development on upland bedrock surfaces described in Katherine Gorge (Baker *et al.*, 1987). Wohl (1992a, 1992b) has attributed erosional features in the Burdekin and Herbert gorges in Queensland, such as spectacular longitudinal grooves, potholes, inner channels, bedrock benches and irregular longitudinal profiles to extreme flows. Baker and Pickup (1987) observed that pool scour is located at extremely sharp bends and hypothesised that pools are drilled at joint intersections by intense hydraulic action of strong and persistent vortices. Large-scale depositional features are also controlled by the hydraulics of extreme floods. Wohl (1992 a, b) found boulder bars located at sites of stream power minima such as at channel bends and locations of shallow water.

Pickup (1991) looked beyond the confined gorge reaches for evidence of large floods on the Todd River in central Australia and briefly described a suite of landforms, including sand sheets, sand threads, ripple fields and ov[illegible] channels, inferred to be deposited by extreme floods. He suggested that deposition during the last 10,000 years was dominated by a few gigantic floods. Patton *et al.*, (1993) described the morphology and sedimentology of the Ross River floodout plain as one which is composed of a series of large-scale paleo-braid channels, levee deposits and broad, low relief bars. These landforms are associated with a series of high magnitude floods which occurred between approximately 1500 and 700 BP. The geomorphic effects were thought to be limited to close to the ranges and Pickup (1991) says:

> Large-scale fluvial landforms are best developed very close to the mountain ranges where runoff is at its most intense. Further downstream they are more difficult to detect...

The data in Chapter 6 indicate that the effects of the paleofloods are detected over 100 km downstream. Similarly Tooth (in press) noted the presence of transverse bedform fields on the Sandover River, 240 km from the MacDonnell Ranges.

7.3. Paleoflood Geomorphology and Processes: First Order Effects

The central Australian landscape provides one of only a few examples so far described, of large-scale, arid zone fluvial landforms resulting from extreme rainfall events. This landscape thus presents a rare opportunity to study the characteristics and processes of high magnitude events through an investigation of their morphology and stratigraphy. The following section synthesises the geomorphology of the Todd River paleofloods.

First order effects (Fig. 7.1) are the immediate geomorphic effects of the floods. Second order effects incorporate the longer-term impacts and recovery of the catchment from the paleoflood events (Chapter 8).

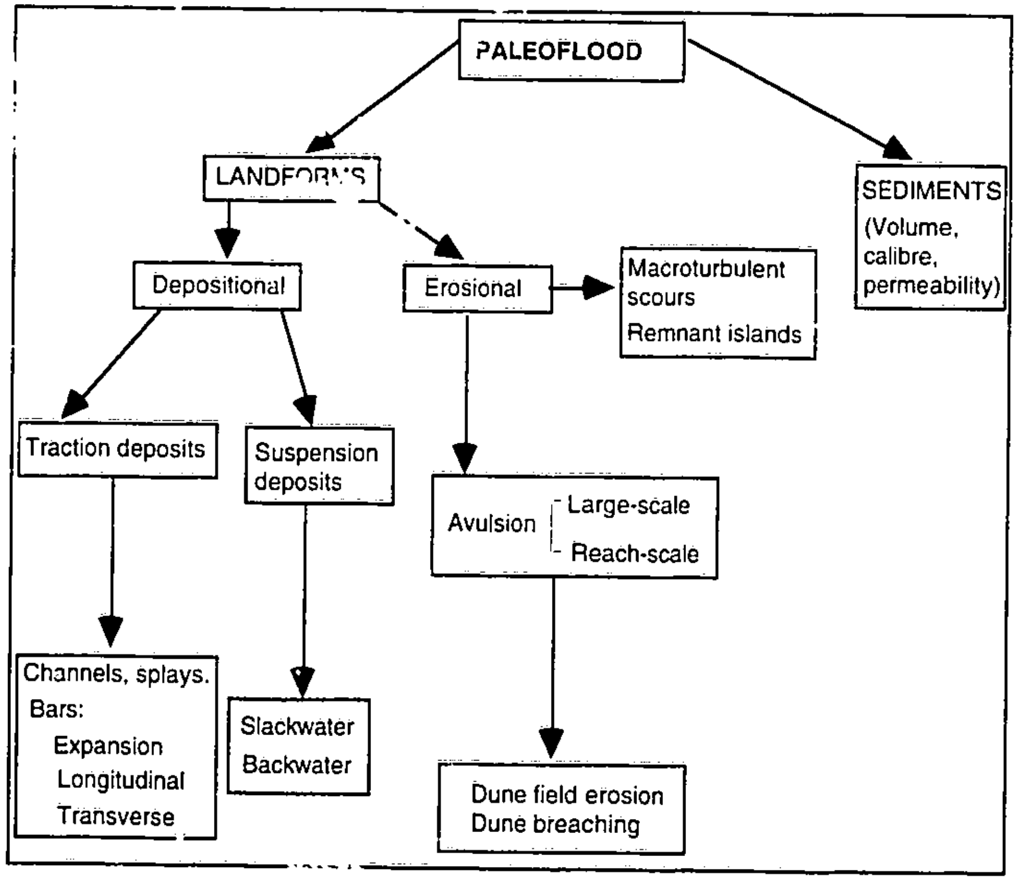


Figure 7.1. First order geomorphic effects of paleofloods

7.3.1. Depositional landforms

7.3.1.1. Channels

Paleofloods flowed from confined gorges and gaps in the MacDonnell Ranges onto the unconfined piedmont fans. Ancient sheet flows are identified from remnant transverse bars: *e.g.* the unnamed piedmont system (Appendix 4, Image 3); on the Ross River floodout, previously described by Patton *et al.*, (1993; Appendix 4, Image 5a); and at the terminus of flood distributaries *e.g.* Heavitree fan complex (Appendix 4, Image 2a).

Except for sheet flows immediately below gorge reaches, most flows tended to concentrate in channels of various magnitude and type. The larger-scale complexes were of a sufficient magnitude to extend as continuous channels from the ranges to the northern fringe of the Simpson Desert in relatively straight paths (*e.g.* the Mosquito Bore Distributary, Fig. 6.1d), but were diverted around outlying bedrock ridges and remnants of older and higher Quaternary deposits (*e.g.* Ross River floodout plain, Fig. 6.23) and through pre-existing gaps in high bedrock ridges (*e.g.* Wallaby Gap, Fig. 6.1d). Where channels are diverted around bedrock highs (*e.g.* Mosquito Bore B, Fig. 6.1d), a concave channel develops due to the intensified turbulence initiated by flow separation and diversion. O'Connor (1993) described a similar 'erosional moat' around pendant bars from the Bonneville floods. The larger paleoflood complexes include braided flood channels on convex bodies of sediment, which may measure 4 km kilometres in total width. The flood channels are separated by low relief bars which extend up to 4 km kilometres downstream. Examples of braiding channels are seen on the expansion bar complexes at reach 1 of the Mosquito Bore Distributary (Appendix 4, Image 8a) and on the Ross A paleochannel (Appendix 4, Image 5a). Smaller scale splay channels (~12 km long) extend away from a channel margin (*e.g.* Giles B and Giles C Fig. 6.1d) and are best developed along Giles Creek where they are superimposed by longitudinal and transverse bars. Sediment size decreases rapidly from cobble and gravel deposits in the confined gorges to gravelly sand in the unconfined reaches.

The pre-existing catchment topography may have an important role in dictating the location and type of paleoflood channel. South of Mosquito Bore, the late Pleistocene longitudinal dunes influence the channel configuration of smaller paleoflood channels as some of the flows are confined within swales or diverted around higher aeolian remnants. This contrasts with the

piedmont reaches of the catchment where the lower amplitude dunes and sand sheets (Patton *et al.*, 1993) may have been more conducive to splay, distributary channel and transverse bar deposits.

The following section describes tractive deposits associated with the paleoflood channels, followed by a description of suspended load deposits.

7.3.1.2. Tractive deposits

Three types of tractive deposits are recorded in the Todd catchment: expansion bars, longitudinal bars and transverse bars. Other paleoflood studies have described three additional forms not found in the Todd; point bars, pendant bars, and eddy bars (Baker, 1973; O'Connor, 1993).

<u>Expansion and longitudinal bars</u>

Expansion bars and longitudinal bars generally occupy wider portions of the flood route and are absent from the more confined reaches such as Wallaby Gap but reappear immediately downstream. These are similar to the observations of expansion bar location in the Bonneville and Missoula floods (O'Connor, 1993; Baker, 1978).

Expansion bars are wide gravel and sand deposits located where flow expansion creates a low velocity zone and deposition occurs (Baker, 1978). In the Todd catchment they are found along the Mosquito Bore Distributary in reachs 2 and 3 where they measure 4 km and 8 km wide. Although further drilling is required, the estimated depth of the expansion bar complex in reach 2 (surface area of 42 km^2) is ~2 m thick and in reach 3 is ~3 m thick, similar to the deposit thicknesses and surface area of the Missoula deposits (~3 m thick and 50 km^2, Baker, 1978). Expansion bars terminate at abrupt constrictions (*e.g.* Wallaby Gap) or where the energy rapidly dissipates (*e.g.* in dune fields).

O'Connor (1993) found that expansion bars deposited by the Bonneville floods increased in elevation downstream. This was not measured for the Todd but rough estimates from 1:50,000 maps indicate a decrease in elevation downstream. The two expansion bars along the Mosquito Bore Distributary have well-developed marginal channels which occur at the aeolian dune field boundary (Figs. 6.42, 6.45). A smaller braided channel network on the surface has left a morphology of multiple lenticular bars.

Longitudinal bars (1-~3 km long and <1 m high) were deposited on splays and in distributary channels parallel to the flow. Morphologically they occur as part of a larger longitudinal bar complex (Fig. 7.2a). The stratigraphy of the longitudinal bars in the Todd is poorly preserved and difficult to interpret but it was established that they are deposited on a horizontal surface (Fig. 7.2a) and it is likely that the bar core is a prograding tractive deposit (7.2b). O'Connor (1993) and Baker (1973) found well developed gravel foreset bedding overlain by armoured, coarse, horizontally bedded sediments in the traction deposits of the scabland floods. The fine textured upper deposits of Todd River longitudinal bars drape the traction core deposit and are more likely to have been deposited from suspension (Fig. 7.2b). The bar swales are often incised by a channel, and a larger marginal channel drains the area between longitudinal bar complexes.

Transverse bars

Transverse bars are found in piedmont areas and at the distal reaches of distributaries (*e.g.* Heavitree fan, Appendix 4, Image 2a). They are difficult to recognise in the field as they have low morphologies (~20 cm) and long wavelengths (~150 m) and are severely degraded by cattle near watering points. Transverse bars have been described elsewhere in central Australia (Patton *et al.*, 1993; Tooth, in press) and in south Australia (Rust and Gostin, 1981) and are inferred to be the deposits of very large sheet floods (Pickup, 1991; Patton *et al.*, 1993; Tooth, in press). The following data is taken from other central Australian research. The bars occur in clusters or 'fields' and up to twenty ripples can occur within each train (Tooth, in press). Individually they are low mounds (Ross: 0.22 m; Sandover: 0.3-0.4 m) composed of silty very fine sand in the Ross River and granules in silt in the Sandover River (Patton *et al.*, 1993; Tooth, in press). The stoss slopes tend to be vegetated and the crests and intervening swales are devoid of vegetation appearing as scalds (Patton *et al.*, 1993; Tooth, in press). Wavelengths are typically 100 to 200 m and individual, sometimes sinuous, crest line lengths are between 350 and 700 m, the internal stratigraphy often obliterated by bioturbation (Tooth, in press).

Wells and Dohrenwend (1985) described similar deposits (meso-and macro-bars) in the Mojave Desert which are steeper (.02-.06 m/m), have shorter wavelengths (tens of meters) and consist of coarser sediment (median grain sizes between 2 and 8 mm (meso) and coarse gravel-sized caliche rubble

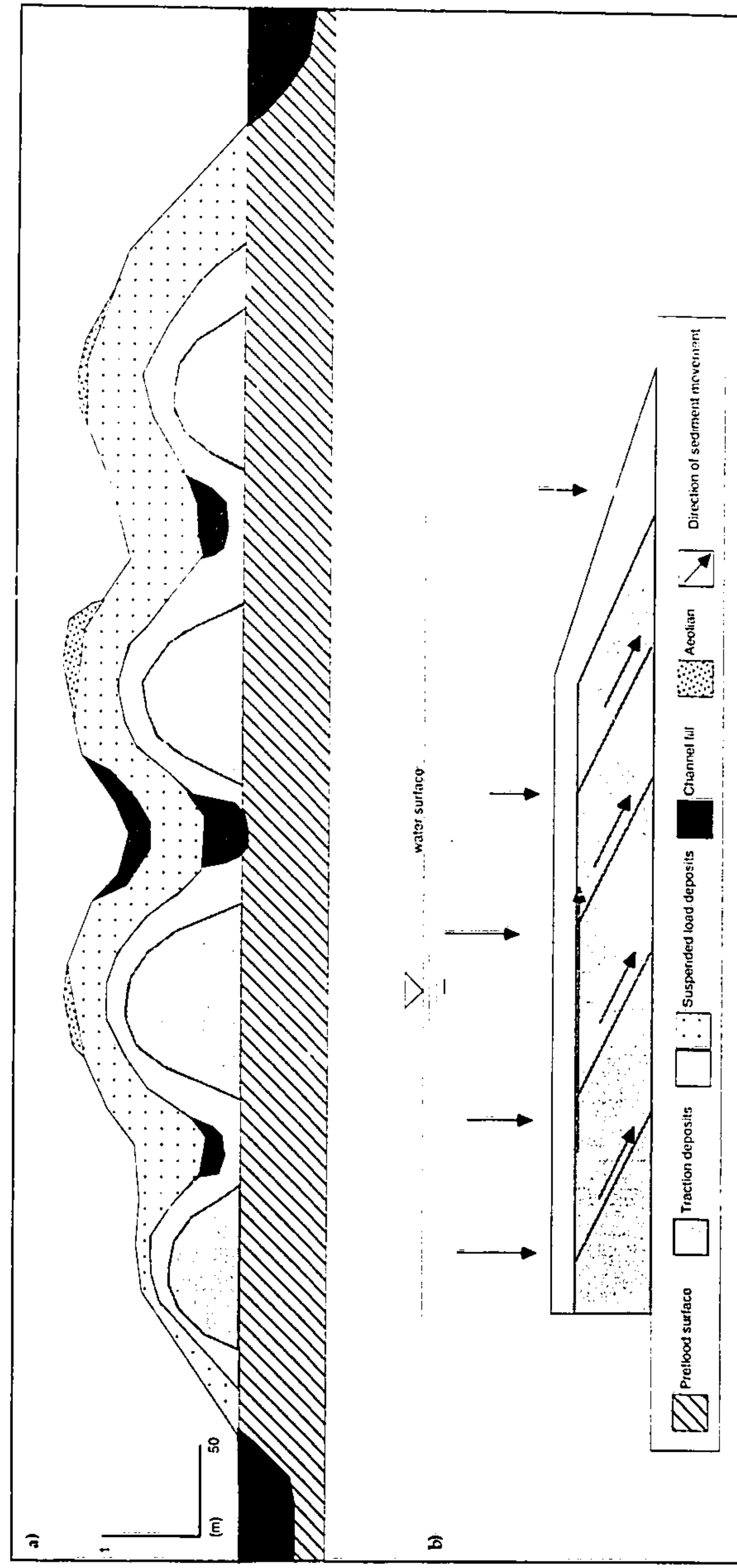


Figure 7.2. Schematic figure of longitudinal bar complexes.
a) Cross section shows the traction core and suspended load drape.
b) The traction core progrades downstream by forset avalanche aggradation and suspended load drape over core.

(macro)), than the central Australian forms. Similar to the central Australian features, they lack prominent internal structure but show some weak imbrication. The mean wavelength of the macro bedform is 31 m, mean bedform width is 17 m and mean height is .36 m (Wells and Dohrenwend, 1985).

7.3.1.3. Suspended load deposits

Typically slack water sediment is well sorted fine or very fine sand, and where exposed, is subjected to erosion and deposition by wind and local runoff. The slack water deposits in the caves at Mosquito Bore preserve a record of a large number of events but were subjected to more frequent erosion relative to the mounded slack water deposit.

Slack water deposits in less confined reaches were found in dune swales adjacent to flood channels. The juxtaposition of fluvial mud and red aeolian dunes produces a clear flood stratigraphy, as during quiescent flood periods well sorted fine red sand is deposited in thin layers across the mud. Although subject to long-term deflation, the aeolian sand is suitable for dating by luminescence and the clay pans and eroded dunes are good indicators of flood inundation.

7.3.2. Erosional landforms

The paleofloods inundated large areas of older alluvial, colluvial and aeolian landforms. In many locations the contact between the surfaces is erosional and truncated remnant landforms lie adjacent to the paleoflood complexes. These include truncated coarse-textured colluvial fans, the eroded downstream extensions of older flood channels and individual longitudinal dunes and large sections of dune fields.

7.3.2.1. The erosion of longitudinal dunes

Collectively, the Todd River paleoflood complexes have caused the pattern of landform assemblage to shift and reform throughout the landscape. The laterally eroded dune field along the western margin of reach 3 on the Mosquito Bore Distributary is inferred to be part of a wider dune field which today survives as aeolian remnants among the eastern paleoflood systems (Fig. 6.1d, Appendix 4, Image 8a). In other locations flows were less catastrophic and individual dunes were breached and survive today as short longitudinal dunes or longer breached remnants with the base of the breach infilled with alluvium.

Three mechanisms of dune erosion are inferred. The first is convex bank erosion by meandering channels as described in section 5.3.2. The second is direct shear by high energy flows inferred to be the dominant mechanism where the erosional flood boundary is transverse to the longitudinal dune orientation. Downstream from Wallaby Gap the south easterly direction of flow towards Rodinga Gap is across the north westerly dune orientation and dunes close to the gap are preserved only in the flow shadow of the bedrock ridge (Appendix 4, Image 9).

The third mechanism is the saturation of the permeable aeolian sand by high water levels. Breached dunes facilitate the passage of floodwaters into swales and thereby increase flow stage on both sides of dunes so that sediment in the aeolian 'islands' quickly saturates, slumps, and is transported downstream. There is evidence to suggest that water depths in the dune fields were sufficient to enable this type of erosion. South of Wallaby Gap, the dunes are 6 m high and a cross section from the dune crest, north east of the expansion bar (Fig. 6.42) indicates at least 3 m of paleoflood fill and the survey from the dune on the western margin of reach 3 of the Mosquito Bore Distributary (Figure 6.45) indicates 2 m of paleoflood fill. Assuming a flow depth of 2 m above these deposits, the dune swales were inundated by ~5 m deep waters placing them just 1 m below the dune crest.

7.3.2.2. Sediment dynamics at gap constrictions

Of particular interest are the hydraulic processes operating at the numerous gaps through which the Todd River and its tributaries pass. The pools which characterise these gaps are intermittently infilled with sediment and may have large boulder bars extending from them. This sediment dynamic is illustrated by the recent event history in Redbank Gorge in the Western MacDonnell Ranges. A large flood in 1983 infilled the bedrock water hole with sediment which was re-scoured during the 1988 flood, demonstrating that many of the short gaps in the ranges currently infilled with sediment (*e.g.* Jessie and Emily Gaps), may in the past have held long-term surface water bodies.

The scour and fill sediment dynamics in gaps are yet to be adequately studied. Jansen *et al.* (1996) found that in Sand Creek, NSW, pool reaches were

preferentially scoured during floods sufficiently large to override the velocity shifts along alternating riffle-pool reaches, while low flows prompted pool aggradation. Laboratory studies by Jopling and Richardson (1966) and modelling of the passage of turbidity currents from a submarine canyon to a submarine plain (Komar, 1971) indicate that the transition at the canyon mouth involves an abrupt loss of mechanical energy through the generation of intense turbulence so that within the hydraulic jump, erosion or non-deposition takes place. Further downstream is a site where only the coarsest fraction will be deposited because the downstream flow will remain competent. Baker (1978) showed that flow through several scabland constrictions was supercritical and proposed that a hydraulic jump may have occurred at the mouth of constrictions where maximum velocities and intense turbulence occur. The morphological and sedimentological features he associated with a hydraulic jump include the deep rock-cut basin such as Soap Lake followed downstream by the large gravel fan of the northern Quincy Basin. Although at a different scale from the scabland features, the large boulder bars which extend downstream from Ormiston Gorge and Simpson Gap in the MacDonnell Ranges are probably related to similar constriction-mouth hydraulics.

7.3.2.3. Macroturbulent scours

Macroturbulence is a three dimensional flow phenomenon which describes the development of secondary circulation, flow separation, and the birth and death of vorticity around obstacles and along irregular boundaries (Matthes, 1947). It has been invoked to explain the formation of swirl holes in bedrock fluvial systems (*e.g.* Baker, 1978; Jennings, 1983; O'Connor, 1993) and also has applications in alluvial settings (Collins and Schalk, 1937).

The most erosive form of macroturbulence is the kolk (Baker, 1978). These intermittent, upwardly spiralling vortices are powerful erosive agents which remove sediment by suction and cavitation (Matthes, 1947). Kolks are reported to be generated under the following conditions: 1. a steep energy gradient, 2. a low ratio of actual sediment transported to potential sediment transport, and 3. an irregular, rough boundary capable of generating flow separation (Matthes, 1947). These preconditions for macroturbulence were probably met during the No. 5 paleoflood flows, the transverse longitudinal aeolian dunes creating an irregular boundary. Further downstream, where the flood channel runs parallel to the dunes, scours are not found. Their asymmetric long profile is similar to that reported by Baker (1978) for the scabland potholes and is consistent with experimental flume studies of vortex scour (Baker, 1978).

7.3.2.4. Remnant Pleistocene islands

Eroded remnants of the pre-paleoflood topography stand as higher streamlined surfaces above the wide alluvial paleoflood plains. The remnants are composed of sediments laid down by older aeolian, alluvial and colluvial systems. Some of these remnants have been fully or partially inundated, and others which have not have the pre-flood topography preserved on the surface. This is similar to Baker's (1978) findings in the scablands. In the Todd catchment the majority of remnants were not been inundated during the paleofloods. One exception to this is the remnant on the left bank of the Ross River dissected by channels (Fig. 7.3) (Appendix 4, Images 5a). A reconnaissance of these features indicates broad, concave gravelly sand channels. Aerial photography and satellite image analysis shows that traces of these channels extend beneath the Ross C ripple field alluvium and are interpreted to represent older alluvial fills which have been stripped and streamlined by paleofloods. It is likely that inundation of the remnant by paleofloods has re-excavated the channel fills producing the channel network on the remnant surface.

Remnants survive in the flow shadows of bedrock ridges, on the margins of paleoflood complexes and on local topographic highs. Downstream from the high bedrock ridges remnant morphology is influenced by ridge size, and orientation relative to the flood flow Figure 7.4 illustrates the streamlined morphology of remnants in the Mosquito Bore area. Remnants which extend beyond the flow shadow tend to be cemented or contain sediment too coarse to be entrained.

Remnants of dune fields not protected by bedrock ridges may survive because they were located on the margin of two paleoflood complexes, *e.g.* the remnant between Mosquito Bore B and Allua A (Fig. 7.4). Alternatively, remnants may be preserved because they are located on topographic highs above the flood waters. Prior to the construction of the longitudinal dune fields, large river systems drained the region as evidenced by the Pleistocene alluvial sediments exposed in the dune swales. Survey data for the Lake Eyre region to the south shows that the dune fields were deposited on the topographically irregular alluvial surface (J. Magee, *pers. comm.*). Although

no data are available for the Todd catchment it is proposed that the survival of aeolian remnants may be due to their location on topographic highs.

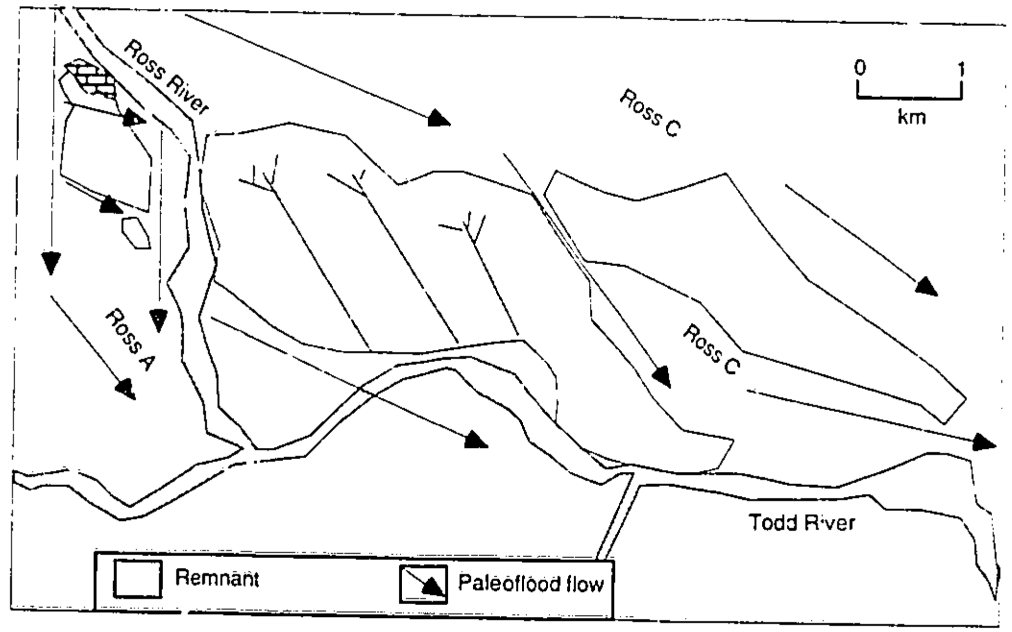


Figure 7.3. Remnant islands associated with the Ross River paleoflood complexes.

7.3.2.5. Avulsion

Channel avulsion has been an important process in the geomorphic evolution of the catchment. Three effects are of note: firstly, the resultant landscape is a mosaic of abandoned channels; secondly, the floods have collectively inundated a larger area than if the flow paths had remained fixed over time; and thirdly, channel entrenchment by nick point erosion has been widespread. Channel avulsion has resulted in the abandonment of <5 km reach lengths but has also changed the depositional end point of the drainage basin. These two scales of avulsion will now be discussed.

Reach-scale avulsion

Two study sites have been affected by the reach-scale avulsion dynamics; they are No. 5 Bore and the Giles-Todd confluence.

No. 5 Bore

The evidence for avulsion at No. 5 bore lies in the preservation of the eastern paleochannel which runs sub-parallel to the western modern channel (Fig. 7.5). Already described in Chapter 6, the stratigraphy records fluvial activity from at least 7 ka to 1 ka. The preserva... macroturbulent scours at the upstream end of the paleochannel indicates the high magnitude of the last flow through the channel and is inferred to be the avulsion event. After the avulsion a new channel was incised to the west and cr ·ently lies ~2.5 m below the bed of the paleochannel (Fig 6.49). The alluvial stores along the western limb are inset into Pleistocene sediment and appear young, displaying little pedogenesis, burial of trees and pr· : vation of sedimentary structures. In addition the radiocarbon ages (380±100 cal BP and three post 1950) indicate that the western limb at No. 5 bore has been active only since ~1,000 BP. Since that time, flows across the right bank flood plain removed part of the dune field close to the channel and breached dunes 800 m away (Fig. 7.5).

Giles Creek

The stratigraphy of Giles K which was described at a site 4 km upstream from the Giles-Todd confluence (section 6.4.4) is interpreted in Figure 7.6 to reflect the dynamic channel changes resulting from the avulsion of Giles Creek. The incision of Giles Creek is inferred to have followed the deposition of Units 1 to 3 (Fig. 7.6a) and to have resulted from the avulsion of the Giles/Todd junction to the present confluence 4 km downstream. The relatively lower local bas· level :t the new confluence would have caused incision of the channel upstream (Fig. 7.6b). This incised channel was subsequently infilled by coarse textured paleoflood sediment (~3 ka, Giles J. Fig. 7.6c) and subsequent incision of the paleoflood channel fill was probably initiated downstream of the point of paleoflood fill aggradation (7.6 d). Evidence for a rapid rate of charnel degradation is indicated by the exposure of the root system of a large River Red Gum (200-300 years old) (Plate 6.9). These Red Gums are not known to grow well on the carbonate-·:h Pleistocene sediments and this tree is inferred to have grown on the

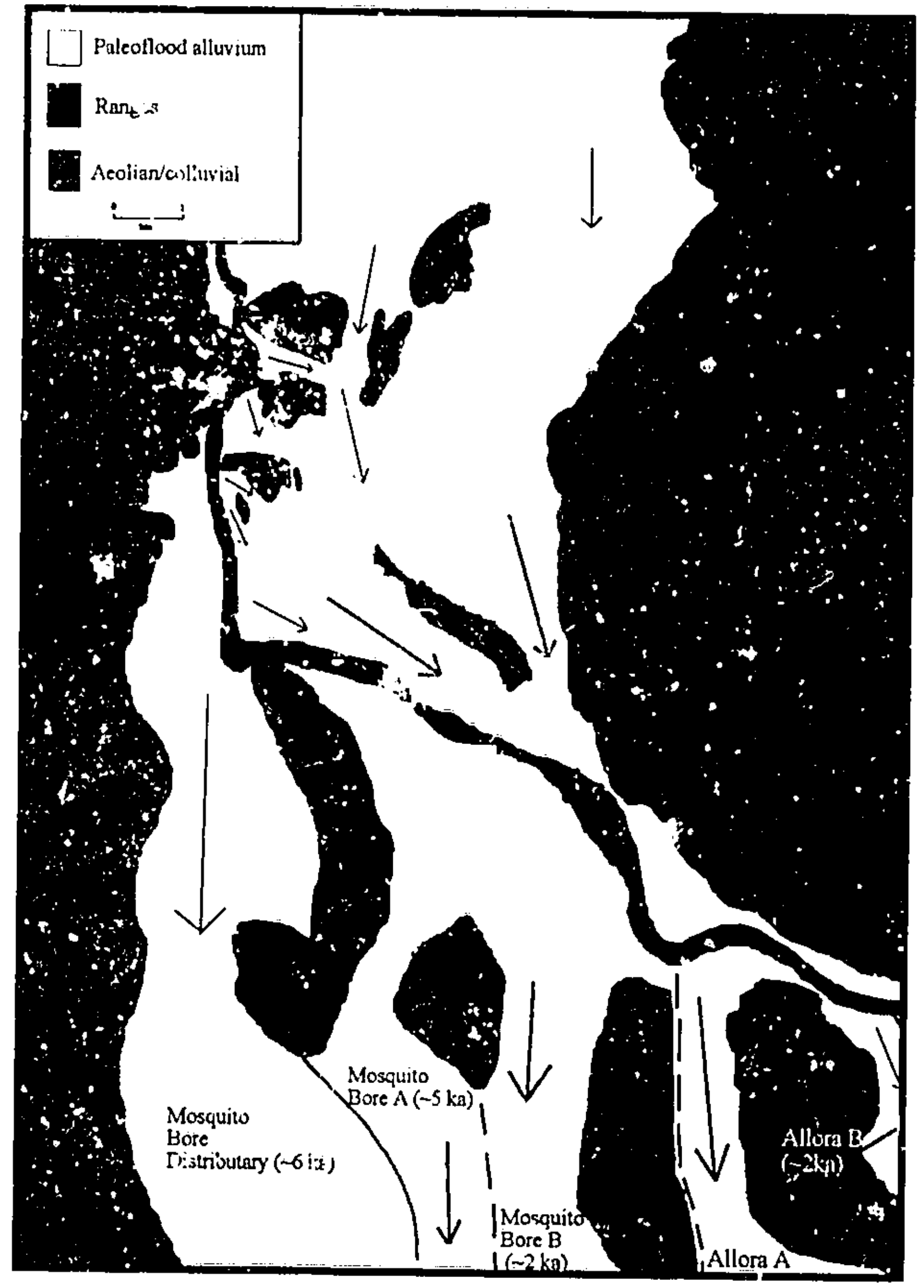


Figure 7.4. Remnant Pleistocene 'islands'.

Figure 7.5. Avulsion at No. 5 Bore.

Giles J channel fill. The root system is now exposed by ~5 m *i.e.* an incision rate of 2 m/100 yr or a few geomorphically effective floods.

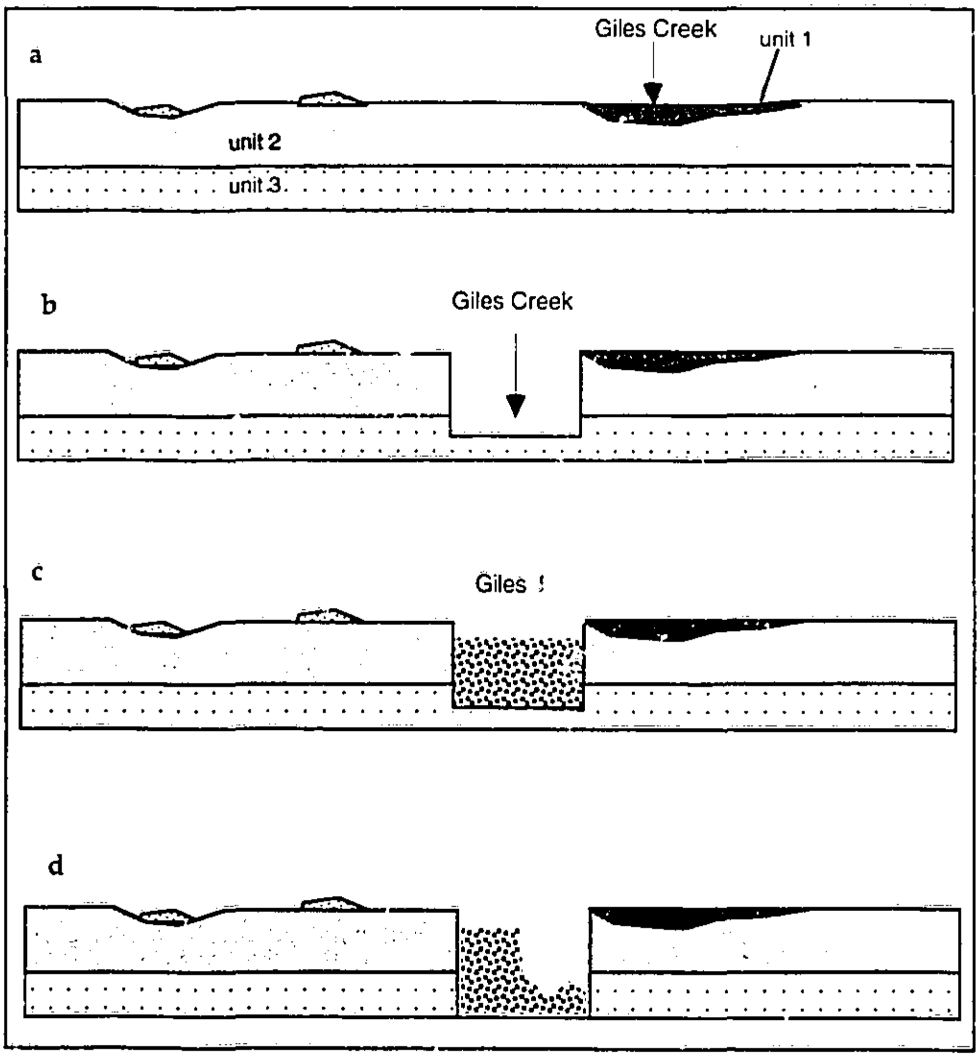


Figure 7.6. Schematic figure of channel change on Giles Creek.

a) The piedmont fan aggrades by distributary and splay accretion

b) Avulsion of Giles Creek, moving confluence 4 km south, channel incision.

c) Aggradation of incised channel by deposition of Giles J paleoflood deposits.

d) Incision of paleoflood deposits

Large scale avulsion

The terminal floodout has been positioned in three different locations during the Late Pleistocene. Moving from west to east they are the Mosquito Bore Distributary, the Rodinga Gap Distributary and the modern floodout (Fig. 7.7). The elevation of these sites and their age at the time of abandonment (Table 7.1) indicate a possible sequence of position of the terminal floodout.

The oldest age so far determined for a floodout is 12310±1400 sampled from the Rodinga Gap floodout. This sample is from a depth of 40 cm below the surface and indicates the last period of fluvial activity in this part of the floodout. The flow path was most probably across the eastern plains through Steele Gap and there may have been additional flow from the Mosquito Bore Distributary.

The Mosquito Bore Distributary floodout is at the highest elevation and is estimated by OSL to be approximately 10 ka. This phase of flooding is also recorded at Giles B and the Expansion Scour site (Fig. 6.56), indicating a significant phase of flooding which is absent from sites further east, and suggests that the channel avulsed from the Rodinga Gap floodout to the Mosquito Bore Distributary between 12 ka and 10 ka to flow directly south. South of Wallaby Gap the Mosquito Bore Distributary traverses the longitudinal dunes in a south easterly direction abutting against the Rodinga Range, and was diverted east and west forming a small floodout downstream of the gap in the Rodinga Range. Some of the flood waters from this system may have drained into the Rodinga Gap terminal floodout. Dated samples from south of Steele Gap (3900±1270 BP) and from the alluvium in the No. 5 bore paleochannel (6800±1850 BP) indicate that flow in the Todd had switched back to the easterly systems by at least 7 ka.

The geomorphology of the modern floodout differs from that of the early Holocene floodouts in sediment texture and surface morphology and the Todd River may only have occupied this position since the mid to late Holocene, as indicated by a radiocarbon age of 5936-5561 cal BP from a flood plain in the desert reach. The modern terminal floodout is a flat clay pan with little vegetation. It is at a lower elevation (195 m) than the late Pleistocene/early Holocene floodouts which are composed of coarse sand and gravel and have a have a higher relief composed of longitudinal bars and multiple channels on an expansion bar. It is inferred that the prior topography of the Late

Figure 7.7. Late Pleistocene terminal floodouts of the Todd catchment

Pleistocene river systems underlying the dune fields influenced the location of the paleoflood complexes and terminal floodouts.

The Todd River paleofloods therefore had the tendency to inundate new, additional areas during successive events by switching the location of the main flood path. For this reason, the total area affected is much larger than if a consistent flood path was maintained. In addition to the geomorphic effects, this tendency has had ecological and archaeological impacts which are discussed in Chapter 8.

Terminal Floodout (from west to east)	OSL Age (BP)	Elevation (m)
Mosquito Bore Distributary (MBD)	9,920±1,840	248.5
Between MBD and Rodinga floodout	not available	228.5
Steele Gap	3,900±1,270	234.5
Rodinga floodout	12,310±1,400	215
Modern floodout	Modern	195

Table 7.1. Elevation and age of terminal floodouts in the Todd catchment (Fig. 7.7). Elevation data is extrapolated from topographic maps.

7.3.3. Sources of sediment

Paleofloods deposited coarse and medium gravel and gravelly sand in the piedmont fans, gravelly sand in the middle reaches and sand with some gravel in the northern fringe of the Simpson Desert with silts and clays in the terminal and backwater areas. The modern flood plain sediments are similar as they are sourced from the paleoflood sediments. The absence of boulder, cobble and coarse gravel deposits is due to two factors. Firstly, the source of coarse clasts is limited, beyond the high energy reaches of the ranges and secondly, flow competence was greatly reduced in the unconfined flow conditions beyond the gorges and incised channels. There are three main sources of sediment for the Todd River paleofloods. These include the slope deposits in the headwaters reach, the Pleistocene alluvial, aeolian and colluvial sediments and the older paleoflood deposits.

In the ranges coarse boulder and gravel sized sediment is supplied to the channel through rockfall, steep tributaries and fans. This sediment is entrained during the higher magnitude flows and stored in high, lateral and central channel bar systems in the gorge. The upstream reach of the Ross River gorge has an extensive, incised gravel fill. Coarse sediments are not abundant beyond the gorge but are found in isolated pockets such as the cobbles in the Ross A section downstream from the Ross-Todd confluence (Fig. 6.25). Giles Creek differs somewhat in that the channel bed is considerably coarser than the Ross and the incised Giles J channel fill contains gravel and cobble sized material ~10 km from the ranges. Where flow is predominantly unconfined however gravel sized sediment may be present but is not abundant.

As the flood waters drained through the aeolian dune fields they eroded entire sequences of parallel longitudinal dunes and breached dunes on the periphery of the channel complex. The flood waters entrained this fine and medium sand which was redeposited in expansion bars downstream along with isolated granule material. The loosely packed well sorted dune sand would have been easily entrained, but the cemented cores provided more resistance and were transported as angular cemented aeolian clasts now preserved as red sand balls in a grey/brown paleoflood sand matrix.

The third source of sediment for the paleofloods is the pre-existing paleoflood deposits. As successive paleoflood tracts were repeatedly inundated or younger paleoflood routes abutted against older routes the contact was erosional and the loosely ...ed sediment easily entrained.

7.3.4. Summary of geomorphology of paleofloods

Large paleoflood complexes containing a network of braided channels, longitudinal and expansion bars can be traced in relatively straight paths from the MacDonnell Ranges to the Northern Simpson Desert, 100 km downstream where avulsion has facilitated the erosion of wide tracts of a...n dunes and the paleofloods have laid down extensive mud and sand deposits with some gravel (*e.g.* south of Wallaby Gap). A 300 m long swirl pit train along the No. 5 paleochannel indicates the turbulent nature of the flow and where the flow path was deflected by bedrock ridges, lee side remnants of the pre-flood topography are preserved. Winnowing of the paleoflood deposits nourished adjacent climbing dunes. These floods caused the pattern of landform assemblages to shift and reform throughout the landscape.

In addition to the immediate geomorphic effects of extreme floods, the degree to which paleofloods generated the modern drainage systems is of relevance as the channel planform and gradient are still recovering from the geomorphic impacts of events that occurred over one thousand years ago.

Chapter 8 ...plores some of the flood-forced changes and adaptations of the modern river.

7.4. Paleoflood Hydrology

There are two accepted approaches to reconstructing the paleohydrology of large floods. One is to estimate flow velocity from sediment transport (Maizels; 1983, Williams, 1983; Costa, 1983) and then to apply this over the paleoflood cross section. The results are conservative. The second approach u...s paleochannel width in bedrock passages, where flow depth is identified from paleostage indicators (PSI). The PSI can be biological, including vegetation damage and regrowth patterns and/or sedimentological including erosional and depositional flood features. While the sampling of flood deposits in unconfined reaches of the catchment improves the chronological record, it is difficult to estimate accurately the paleohydrology of the flows. Few slack water deposits were located in confined areas and cross sectional morphology was too variable.

7.4.1. Discharge and velocity

The slope area method was applied to five sites to calculate peak discharge (Q). The method involves the estimation of peak flood levels in the field. Criteria to be fulfilled for the accurate application of the method include, a straight channel, uniform cross section and slop... no channel obstructions, no backwater effects, no erosion or deposition during the flood and no overbank flow. While it is difficult to fulfil these criteria in the Todd River, this method was applied in order to get a best estimate of the peak discharge. The parameters to be determined include the cross sectional area (A), the hydraulic radius (R [A/P where P is the wetted perimeter]), Manning's coefficient of roughness (n) and the slope of the water surface (S) at the cross section. These are input to the following equation:

$$Q = A\,R^{2/3}\,S^{1/2} / n$$

There were not many locations where there were well defined flow boundaries and reasonable peak flow levels preserved. At the Todd A flood distributary, channel slope and cross section data were available (Fig. 6.37 and 6.38) and as some vegetation currently grows in the channel a Manning's n value of .035 was assigned (Goudie, 1981). The second location was a channel cross section on Giles Creek J (Fig. 6.32) where a high energy paleoflood infilled an incised channel approximately 3000 BP with coarse cobbles and gravel. The dimensions of the entrenched channel were estimated from the remnants of the fill, which was about 4 m (Table 6.5) and a Manning's roughness coefficient for mountain streams was used. The modern channel slope was used as an approximation of the water surface slope. Peak stage was assumed to be bankfull height at the Giles J site and the Todd B site but could have been higher.

For the cross section at Wallaby Gap (Fig. 6.43), the channel slope was measured from orthophoto maps for the reach immediately upstream between Mt. Capitor Bore and Wallaby Gap (reach 2, Table 6.7) and was assumed to represent the water surface slope. The channel cross sectional area was surveyed in at the gap and an erosional step in a sand deposit downstream from the constriction was used to estimate the peak stage (Fig. 6.44). A Manning's roughness coefficient for a clean winding channel with some pools and shoals was used (Goudie, 1981).

At Stud Bore 1 the cross sectional area of paleofloods was reconstructed using the morphostratigraphic remnants of former channel widening events (Fig. 5.6) At Mosquito Bore 1 flow width was estimated in the same manner and flow depth was estimated from the height of the cave slack water deposit. A peak velocity of 2.5 ms^{-1} was assumed for both sites.

It is assumed that the resultant discharge values (Table 7.2) greatly underestimate peak discharge at the first three sites (Todd A, Giles J and Wallaby Gap) as flows were overbank upstream and downstream of the reconstructed cross sections. In addition, the measured discharges for the largest flows in the gauged record at the Wills Terrace gauge in Alice Springs (catchment area of 450 km^2, 1190 m^3s^{-1} in 1988 and 900 m^3s^{-1} in 1983) are known to have had a very limited impact on the paleoflood forms on the Heavitree Fan. Additionally, the scale of the recent flood forms does not match those of the paleoflood forms. It is concluded therefore that the application of these methods is unsuitable in this instance.

Site	Cross sectional area (m^2)	$R^{2/3}$	$S^{1/2}$	n	Q ($M^3 s^{-1}$)	V (Ms^{-1})	(U_c) (ms^{-1}) [IA cm]
Todd A	60 [25x2.4]	2.23	.05	.035	1.91	3.19	0.20 [0.1]
Giles J	1560 [390x4]	2.61	.07	.04	7125	2.61	1.[illegible] [9]
Wallaby Gap	1225 [350x3.5]	2.13	.04	.04	2610	2.13	0.27 [0.2]
Stud Bore 1 event B (Fig. 5.6)	825 [250x3.3]				2062	2.5 (e)	
Stud Bore 1 event A (Fig. 5.6)	1650 [375x4.4]				4125	2.[illegible] (e)	
Mosquito Bore 1	1350 [270x5]				3375	2.5 (e)	

Table 7.2 Discharge and velocity estimates for paleoflood flows. Flow velocities of 2.5 m^3s^{-1} are assumed for Stud Bore 1 and Mosquito Bore. R: hydraulic radius, S: slope, n: Manning's roughness coefficient, Q: Manning's peak discharge, V: Manning's peak velocity, U_C: mean velocity (Costa, 1983), IA: intermediate axis, (e) estimated.

Two methods were used to calculate the mean velocity of the paleofloods. The Manning's uniform flow equation requires the calculation of R, S and n, which are input into the following equation.

$$V = R^{2/3} S^{1/2} / n$$

The second method is based on the calculation of the velocity threshold for the entrainment of sediment. The largest particle moved at each site was measured and input into the following equation (Costa, 1983).

$$U_C = 57D^{0.455}$$

where U_C is the mean flow velocity (m s^{-1}) and D is the intermediate axis. Threshold velocities at each site are less than velocities given by Manning's equation probably because the largest size sediment capable of being transported was not available in the downstream reaches.

The conservative peak discharges (*e.g.* at Giles J, 7125 m^3s^{-1} and Stud Bore 1, 4125 m^3s^{-1}) are still large for the Todd catchment (9,300 km^2) and are the equivalent of the eight year flood on the Daly River (49,000 km^2) in monsoonal northern Australia (Chappell and Bardsley, 1985).

7.4.2. Flow depths

The size of sediment in transverse bars on the Ross River piedmont plain (Ross C) as used to estimate the depth of sheet flow. Wells and Dohrenwend (1985) have attempted to calculate the mean flow depth (D) from transverse rib sedimentology (b_{max}) and paleoslope (S) using the equation developed by Koster (1978):

$$D = 0.15\, (b_{max})^{0.67} / S$$

where b_{max} is the intermediate axis (cm) of the ten largest clasts. The mean depth of flow over the Ross River C plain using a paleoslope of .0034 m/m and a particle size of .1 cm (Patton *et al.*, 1993) is 0.093 m. This is clearly an underestimate as the height of the transverse bars average at 0.22 m.

7.4.3. Criteria for estimating flood magnitude rank

The high magnitude of paleofloods is testified by the erosion of pre-existing landforms, the scale and spatial extent of the flood forms, the coarse texture of the deposits, the distance the flood effects extend from the ranges, the avulsive nature of the flows and the fact that no floods since European settlement of the region have inundated the higher paleoflood surfaces.

Paleoflood discharge estimates based on channel sections above are very likely to be too low. In order to assess the flow magnitudes four criteria were chosen (Table 7.3). Firstly if the paleoflood channel has been inundated by historical floods (A) it is assumed to be formed by a low magnitude paleoflood (category 1) as the historical floods are reported to have had little effect on the paleoflood forms (Pickup, 1991). Secondly, if a paleoflood is traced >50 km from the ranges (B), particularly if flow was unconfined, it is assumed to have been a high magnitude flow (category 3). Thirdly, if the flood deposits cover a large area (C) it is assumed to have been a high magnitude flood (category 2 or 3). Fourthly, is the scale of the flood forms (D), as large bars (>2 km long) and wide channels indicate a significant flow magnitude (category 2 or 3). Floods were assigned a flood magnitude weighting between 1 and 3, 1 indicating the smallest events.

Category 1 and 3 events were easier to distinguish than category 2 and 3. A combination of the magnitude criteria improves the flood magnitude ranking. For example, unconfined flows which extended over 50 km from the ranges were ranked as category 3 events if they were associated with traction deposits and if the flow was unconfined for the 50 km, but if the flow had avulsed from confined to unconfined in the latter part of the flow path it was given category 2 status. Thus, category 1 = A; category 2 = C+D; category 3 = B+C+D. The size of sediment was not used as there tended to be little difference between flood deposits on that basis (Table 6.2).

Flood Magnitude Criteria	Problems
A) Inundated by historical floods	May not be inundated due to avulsion or channel incision.
B) Unconfined flood channel >50 km from Ranges.	Transmission losses into the channel bed may increase between floods.
C) Large areal coverage of event	Erosion of evidence by successive floods.
D) Large scale of flood features, *e.g.* 4 km wide complexes, 2 km long bars.	May be dependent on sediment supply or degree of flow confinement.

Table 7.3. Criteria for Flood Magnitude ranking: category 1 = lowest magnitude. category 1 = A; category 2 = C+D; category 3 = B+C+D.

7.4.3.1. Category 1 paleofloods

These are the lowest magnitude paleofloods, identified by their proximity to the main channel and mapped in Figure 6.1 as the Todd A and Todd C complexes. Although formed by high magnitude floods they have been occupied during historical times, as shown by the lack of pedogenesis, modern radiocarbon ages and the presence of young River Red Gums and Coolabahs. Flood distributaries extend from the channel and channel dimensions decrease rapidly.

The Todd A paleoflood complex is a typical category 1 example (Fig. 6.1d). The flood record at this site extends from the underlying bedrock and the chronology begins close to that boundary (Fig. 6.38). Preservation of sediments and morphology since 400 BP suggest that no events of magnitude category 2 or 3 have occurred in the last 400 years in this reach of the basin.

7.4.3.2. Category 2 paleofloods

Category 2 events are greater in magnitude than category 1 events but are smaller than category 3. At the higher end of this category are the floods in the Mosquito Bore A, and Mosquito Bore B channels (Table 7.4). These paleoflood complexes diverge strongly from the channel and are interpreted as part of a wider flow path which is no longer visible, draining across the modern channel. They inundated wide areas, and deposited gravelly sand. They terminate in sump basins downstream controlled by the location of outcropping ridges (Fig. 6.1d).

Others flows in this category were somewhat smaller than Mosquito Bore A and B, as they deposited principally suspended loads (Allua B). The traction deposits are mainly limited to confined reaches (Ross A, No. 5 bore, Giles J) and evidence suggests rapid dissipation of flood effects downstream (*e.g.* Heavitree transverse bedform field).

7.4.3.3. Category 3 paleofloods

Category 3 floods are identified on the basis of three criteria. (i) they travelled the maximum distance downstream, (ii) their flow path was essentially unconfined, except at bedrock passages and (iii) they deposited traction deposits the entire length of the flood complex (Table 7.4). Two paleoflood complexes stand apart from the rest in this regard: the Mosquito Bore Distributary (Fig. 6.1d) and the Rodinga Gap paleoflood complex (6.46). Both maintained unconfined flow paths depositing traction deposits including gravel over 50 km downstream from the ranges and eroded surrounding aeolian landforms.

When the ages of paleofloods are plotted against their respective flood magnitude category (Fig. 7.8) it appears that the highest magnitude flows occurred during the late Pleistocene and early to mid Holocene. Category 2 floods occur from the mid Holocene and the lowest magnitude floods occur during the last 400 years. The youngest events in category 3 are the smallest events in that group and the oldest events in category 2 are the largest in

that group. The plot suggests that the mid Holocene events may require a separate category.

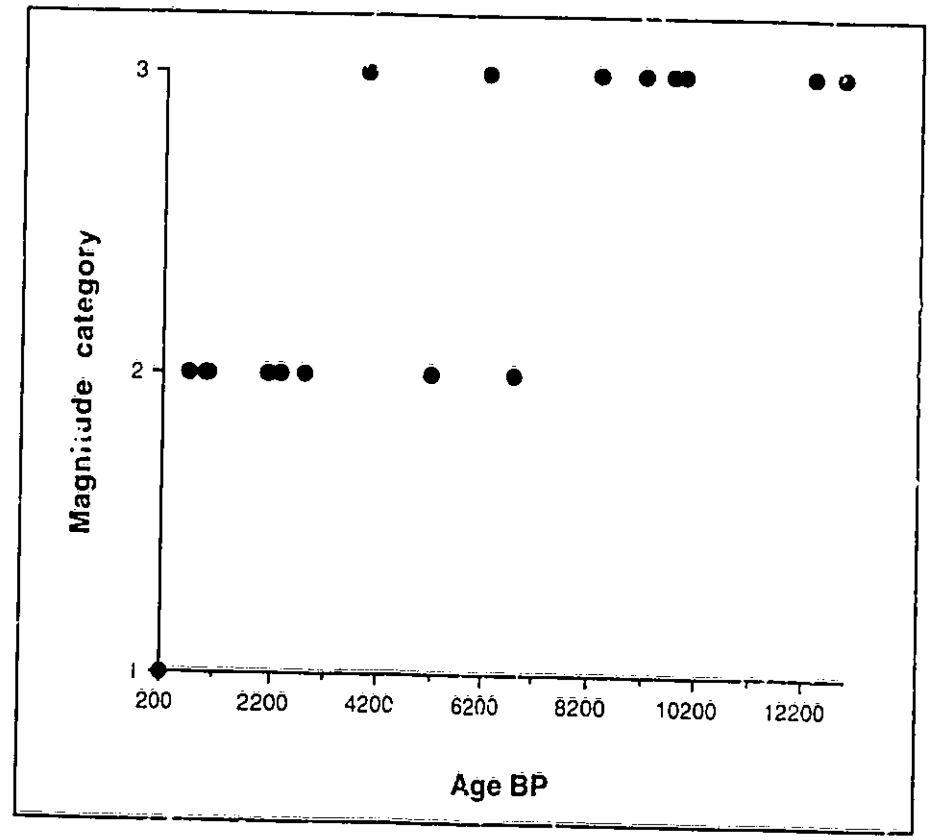


Figure 7.8. Scatter plot of flood magnitude category and paleoflood chronology.

This decrease in flood magnitude over time agrees with the findings of Nott *et al.* (1994, 1996) for the north of Australia. At Waterfall Creek flood discharges during the early to mid Holocene were 2-3 times greater than those occurring over the last 5 ka. The extreme floods recorded at Wangi Creek represent the likely maximum possible discharge achievable for a catchment of that size (55 km²). Other paleoflood work in tropical Australia has found the Late Holocene floods to be relatively low magnitude events. Although the paleoflood records from Katherine Gorge and Finke Gorge only extend back 2 ka (Baker and Pickup, 1987; Baker *et al.*, 1983) the estimated maximum discharges for this period were only one fifth that of the possible maximum discharges for their respective catchment sizes (Nott *et al.*, 1996). Similarly, the work of Wohl (1992a and b) in tropical Queensland indicates that the floods recorded during the last 1200 yr show maximum peak discharges less than half that of the likely possible maximum discharge (Nott *et al.*, 1996). It is only in the Kimberley region of Western Australia (Gillieson *et al.*, 1991; Wohl, 1994a), that discharges close to maximum possible discharges been recorded since the mid Holocene (Nott *et al.*, 1996).

7.5. Paleoflood Chronology

7.5.1. Introduction

Radiocarbon and OSL were used to estimate the absolute age of the paleofloods. The methods and data are discussed in Appendix 1. The following section will present the paleoflood ages and discuss them in the context of other Australian and tropical paleoflood chronologies.

7.5.2. Results and interpretation of dating program

Fourteen OSL and thirteen radiocarbon age estimates are used in the analysis of the chronology of the paleofloods. Given the large error bars of the OSL ages it is not possible to separate individual floods accurately based on chronological data alone, but separation was achieved by combining the stratigraphic and chronological data (Table 7.5). Figure 7.9 is a scatter plot of the OSL and radiocarbon ages for each of the paleoflood complexes.

The dated flood deposits fall within the last 13000 years, with one outlier at 26800±3000 BP (Fig. 7.9). The resolution of the dating techniques and the use of two methods makes it difficult to identify whether events were more common in some periods than others, but it was of interest to see if phases of floods could be identified and this was undertaken qualitatively (Fig. 7.10), firstly and conservatively by including the error bars associated with the ages and secondly by ignoring the error bars.

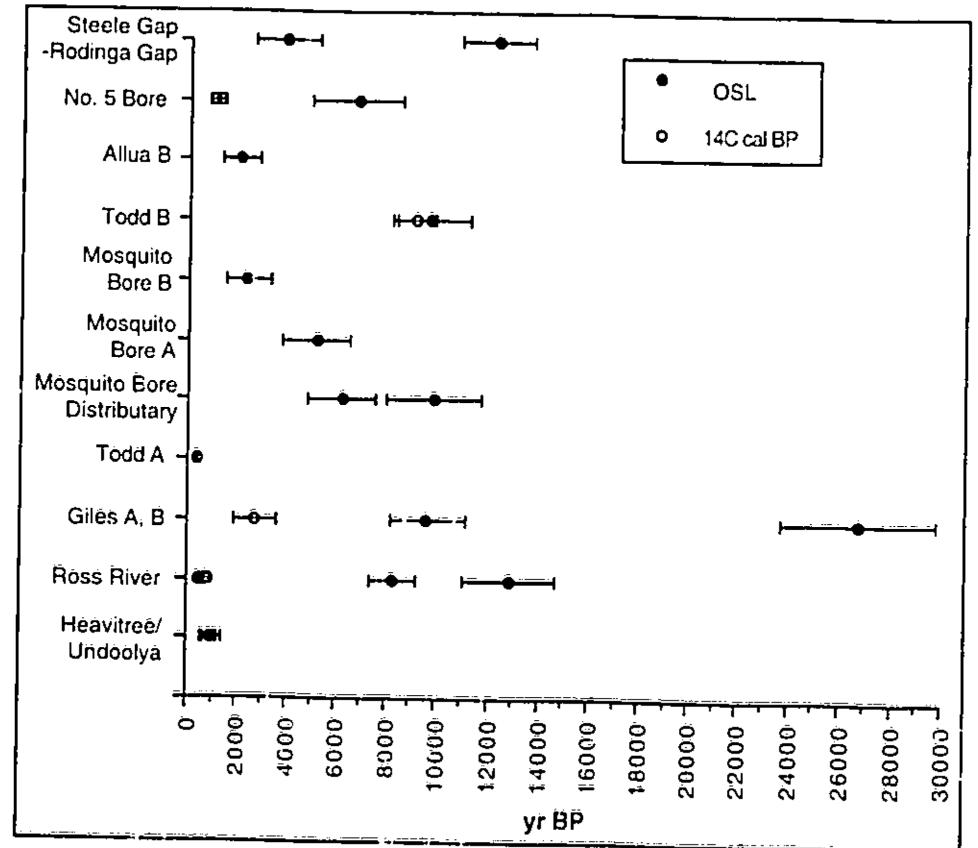


Figure 7.9. Scatter plot of ^{14}C and OSL paleoflood ages.

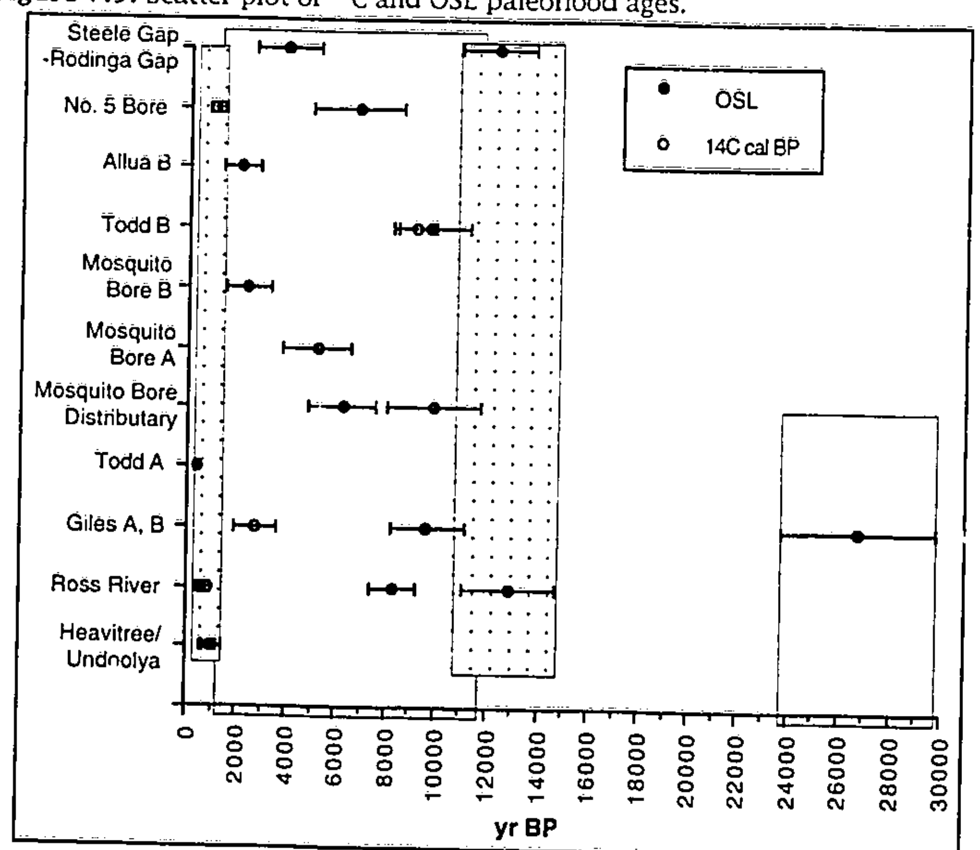


Figure 7.10. Todd River paleoflood phases.

A phase of activity was identified by grouping similar age data incorporating the values of each of the error bars. In principle, each phase should be separated from the next on the basis that the mean age of the sample does not fall within the error margin of a sample in the next phase. It is important to note that because the error bars are larger for the OSL ages than for the radiocarbon ages, this extends the range of flood phases for the older events which were principally dated using OSL (see Figure 7.10). Four phases of flooding are identified although the youngest three are not clearly separable statistically. The oldest phase (24000 - >30000 BP, Table 7.5) is easily differentiated from the remaining data but comprises only one data point. The second phase has two data points and occurs between 11000 and 15000 BP, the third extends from 11000 to 1500 BP and could not be subdivided on the basis of the criteria defined above, and the fourth extends from 1500 to 300 BP (Table 7.5).

The second approach which does not include the error terms obviously is problematic but highlights some possible trends that are masked by the more conservative approach above. The data are plotted as frequency, indicating the number of ages in 2000 year intervals (Fig 7.11a) and 500 year intervals (Fig. 7.11b). In constructing these figures, where a stratigraphic layer was sampled twice only one age estimate was included.

Results for the 2000 year interval (Fig. 7.11a) suggest three phases of flooding: >26 ka, 12 - 14 ka and <10 ka, with gaps that may indicate periods of quiescence between 26-14 ka and 12-10 ka. The same data set grouped 500 year intervals indicates some possible patterns in the Holocene data (Fig. 7.11b). Six periods with no dated flood deposits and possibly no floods are identified at around 1500, 3000, 4000, 5500, 7000, 9000. The time series suggests that floods in central Australia are randomly spaced.

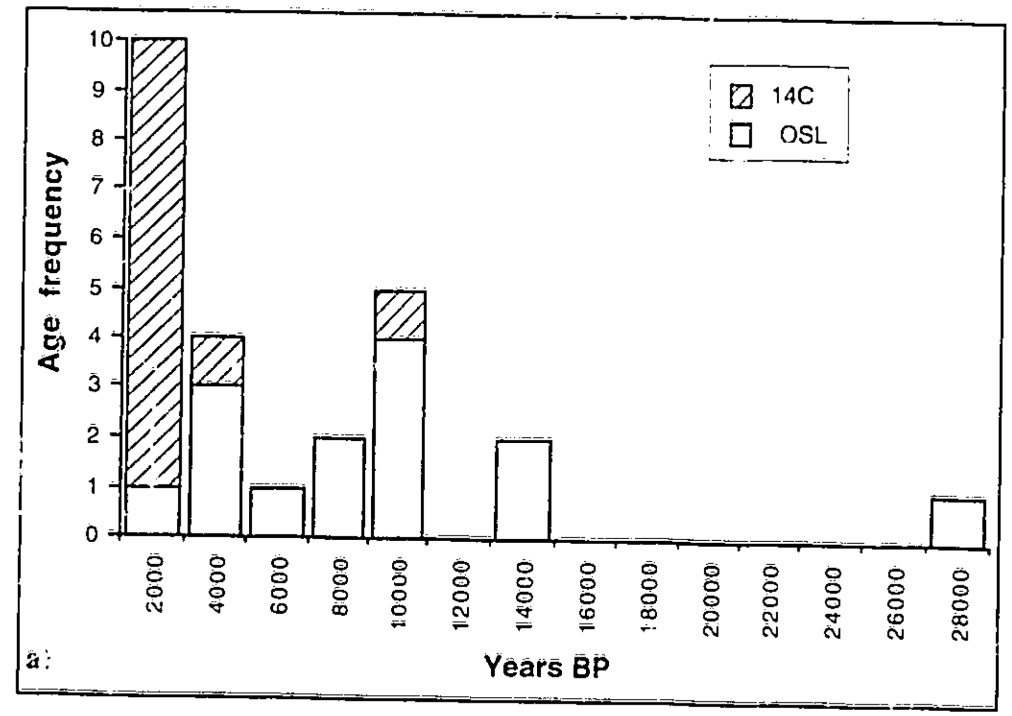
Drawing any solid conclusions from the data set is limited by the size of the error bars. Therefore it is concluded that extreme floods in the Todd catchment have occurred repeatedly since 15000 BP with little clear evidence at present to suggest a clustering of events.

Paleoflood system	A) Inundated by historical floods (evidence) [problems]	B) Unconfined channel >50 km from Ranges.	C) Large areal coverage of event	D) Large scale flood forms, traction deposit	Flood Magnitude Category (Age ka BP)
Heavitree/Undoolya	No [Channel avulsion, 1880]	No (10 km)	extensive (4)	yes/yes	2 (.7-1.5)
Ross River A	No [Channel avulsion/incision]	No (10 km)	extensive (4)	yes/yes	2 (.7-1.5)
Ross C Channel	No [Channel avulsion/incision]	No (10 km), but possibly linked to Rodinga Gap	extensive (2)	yes/yes	3 (11-15)
Ross C SWD	No	No (0 km)	extensive (5)	yes/no	3 (8-11)
Giles B	No [Channel avulsion/incision]	No (12 km), but linked to Camel Flat	extensive (5)	yes/yes	3 (8-11)
Giles J	Yes (incised channel)	Confined	moderate (3)	yes/yes	2 (1.5-3.5)
Todd A	Yes (Coolabah and young Red Gum growth)	No, local distributary	no	no	1 (< 3 and .3-.7)
Cave SWD	Yes (radiocarbon chronology)	Confined	no suitable evidence	no suitable evidence	1 (<.3 and .3-.7)
Mosquito Bore Distributary	No [Channel avulsion/incision]	No (30 km)	extensive (4)	yes/yes	3 (4-8)
Mosquito Bore Distributary at Camel Flat Bore	No [Channel avulsion/incision]	Yes (75 km)	extensive (2)	yes/yes	3 (8-11)
No. 5 paleochannel	No [Channel avulsion/incision]	Confined	extensive (4)	yes/yes	2 (4-8)
No. 5 paleochannel (scours)	No [Channel avulsion/incision]	Confined	extensive (4)	yes/yes	2 (.7-1.5)
Mosquito Bore A	No [Channel avulsion/incision]	No (45 km)	extensive (4)	yes/yes	2 (4-8)
Mosquito Bore B	No [Channel avulsion/incision]	No (45 km)	moderate (3)	yes/yes	2 (1.5-3)
Todd B Expansion scour	No [Channel incision]	No (35 km)	extensive (5)	yes/yes	3 (8-11)
Allua B	No [Channel avulsion/incision]	Yes (>50 km)	extensive (3)	yes/no	2 (1.5-3)
Steele Gap	No [Channel avulsion/incision]	Yes (80 km)	extensive (4)	yes/yes	3 (4-8)
Rodinga Gap	No [Channel avulsion/incision]	Yes (100 km)	extensive (2)	yes/yes	3 (11-15)

Table 7.4. Inferred Todd River paleoflood flow magnitudes.

Flood period	Date ranges	No. of flood events	Sites (no. of floods)	Flood deposit type
0-300	< modern	~5	Cave SWD (5)	*Suspended* Slackwater deposit
	<150 years		Giles I (2)	Floodplain
300-700	660-590-520	~6	Ross A (4)	*Traction* Channel
	570-500-430		Ross A	Channel
	490-390-290		Cave SWD (2)	*Suspended* Slackwater deposit
	510-410-310			
	510-410-310		Todd A(6)	Distributary
	460-400-340			
700-1500	980-860-740	~1	Ross A (1)	*Traction* Channel
	985-855-725			
	1450-1030-610		Heavitree (1)	Transverse bars
	1190-940-690			
	1405-1270-1135		No.5 paleochannel (1)	Channel
	1085-905-725		No. 5 paleochannel (2)	Channel
1500-3500	2870-2110-1350	~3	Allua B (2)	*Suspended :* Floodout deposit
	3270-2370-1470		Mosquito Bore B (2)	*Traction* Flood distributary
	3640-2790-1940		Giles I (1)	Channel
4000-8000	5170-3900-2630	~4	Steele Gap (1)	*Traction* Expansion bar
	6500-5160-3820		Mosquito Bore A (1)	Flood distributary
	7600-6230-4860		Mosquito Bore Distributary (1)	Distributary channel.
	8650-6800-4950		No. 5 paleochannel (1?)	Channel
8000-11000	9300-8350-7400	~5	Ross River C (1)	*Suspended* Slack water deposit
	9890-9160-8430		Todd B	*Traction* Channel
	11200-9700-8200		Giles B (3)	Longitudinal bar
	11290-9780-8270		Todd B (1)	Channel
	11760-9920-8080		Mosquito Bore Distributary (1)	Channel, expansion bar
11000-15000	13710-12310-10910	~2	Rodinga Gap (1)	*Traction* Flood distributary, expansion bar
	14700-12900-11100		Ross C (1)	*Suspended* Channel
24000-30000	29800-26800-2	~1	Giles A	*Traction* Flood distributary

Table 7.5. Flood period and number of flood events.



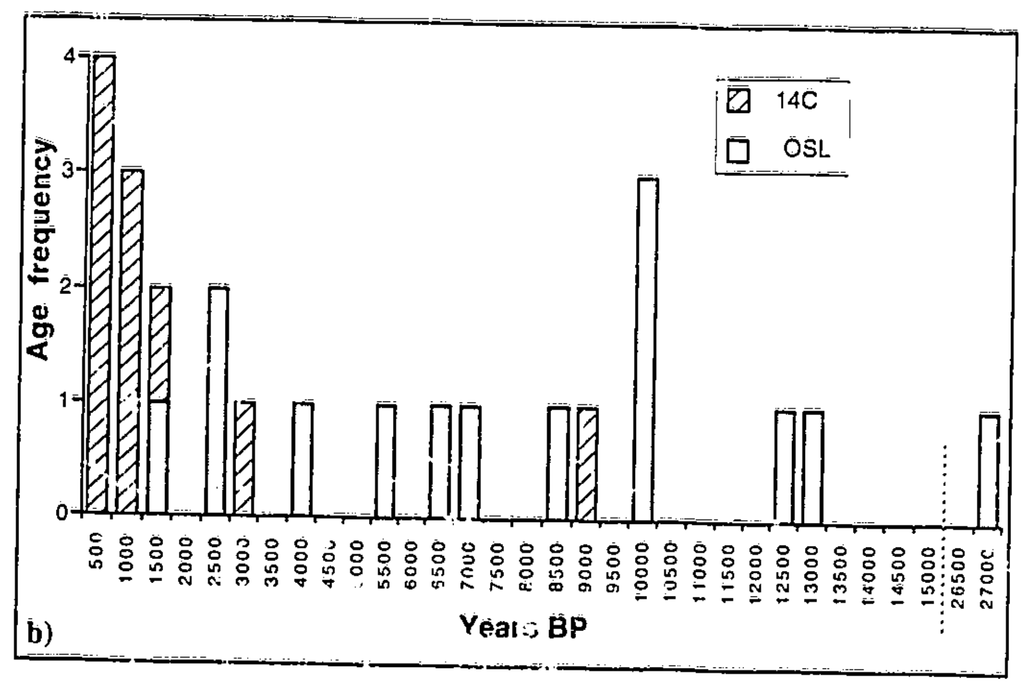


Figure 7.11. Frequency of paleoflood age.
a) At 2000 year intervals.
b) At 500 year intervals.

7.5.3. Australian paleoflood chronologies

The age estimates of the Todd paleofloods were plotted against data from other paleoflood studies in Australia (Fig. 7.12). While the research covers a relatively good spread of river systems in north west, north, north east and central Australia, few records extend past 4000 yr BP and the majority of flood ages are younger than 2000 yr BP. The more extensive flood records are from the Todd River and Waterfall Creek and indicate that the other flood records are incomplete. Possible explanations for why records are limited are explored in Appendix 1.

As the monsoon-driven weather systems which cause flooding in central Australia originate in the north west, north and north east it is possible that the flood record from these regions may be similar to that of the Todd River. Figure 7.13 was plotted to ascertain if some of the large floods were coeval in central Australia and across the north of the continent.

A flood phase centred on 400 BP is recorded in all four regions, and a phase at 600 BP is recorded in central Australia and the Burdekin River in Queensland. Evidence of a phase at 1000 BP is also found in all four regions and the phase at 2100 is found in the centre and the north west. Phases with evidence reported from all four regions may indicate a stronger monsoon, with extreme rainfall events occurring more widely or more frequently than at present across the top half of the continent. Dating of floods in the Finke River (Pickup *et al.*, 1988) and the Todd River indicate that some events which appear to have extended throughout the Todd (Fig. 6.56) may also have been coeval in the Finke River. This would suggest that the high rainfall weather systems were regional and future investigations should find evidence of these floods in other central Australian catchments.

Over a longer period Figure 7.14 shows that the two longest paleoflood records from Australia are similar and suggest enhanced fluvial activity beyond the 1.2 ka episode (as noted from Fig. 7.13) occurred at 5,000, 6700, 8300 and 12500 BP. However the Waterfall Creek record (Nott *et al.*, 1996) does not appear to record the strong 10 ka signal found in the Todd.

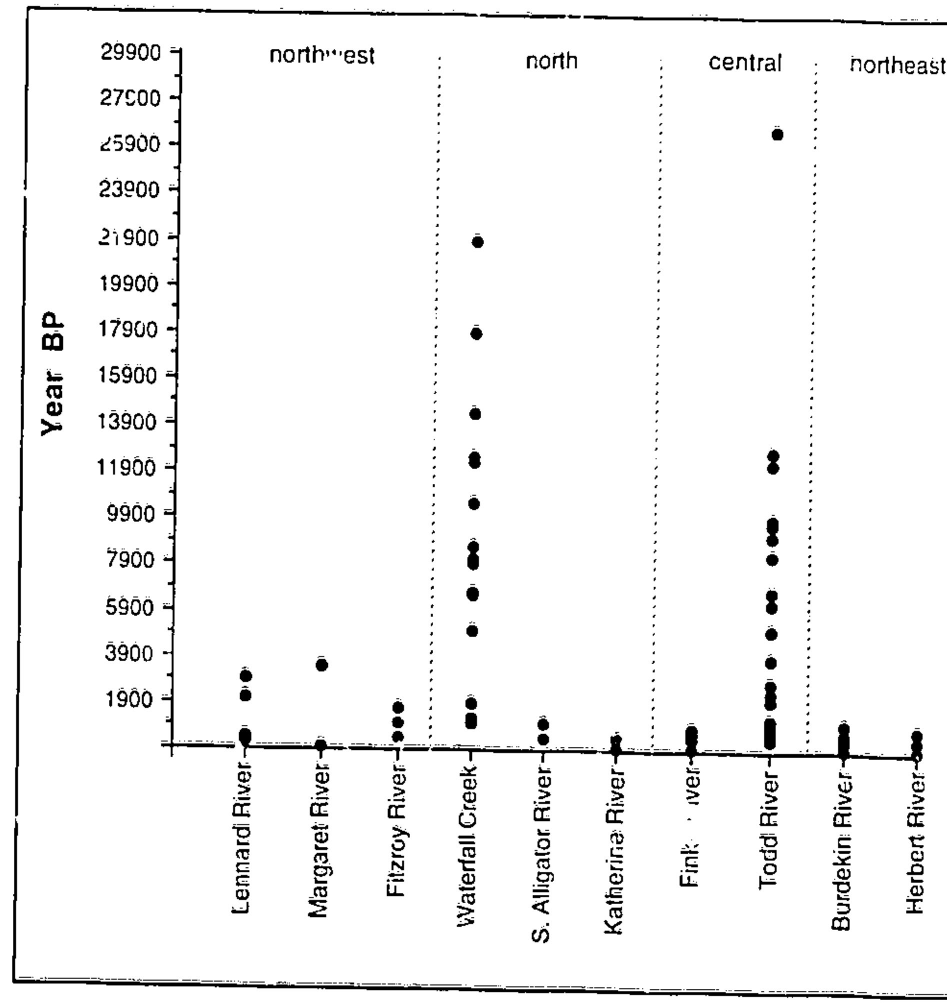


Figure 7.12. Age of floods north of 25°S in Australia. Data is from the following sources: Lennard River (Gillieson *et al.*, 1991); Margaret and Fitzroy Rivers (Wohl *et al.*, 1994 a); Waterfall Creek (Nott and Price, 1994; Nott *et al.*, 1996); South Alligator River (Wohl *et al.*, 1994 c); Katherine River (Baker and Pickup, 1987); Finke River (Pickup *et al.*, 1988); Todd River (Patton *et al.*, 1993, and this study); Burdekin River (Wohl, 1992a); Herbert River (Wohl, 1992b). The Waterfall Creek data includes all luminescence age estimates as individual floods were not identified by Nott *et al.*, (1996). Data from Wangi Creek was not included as there are age reversals.

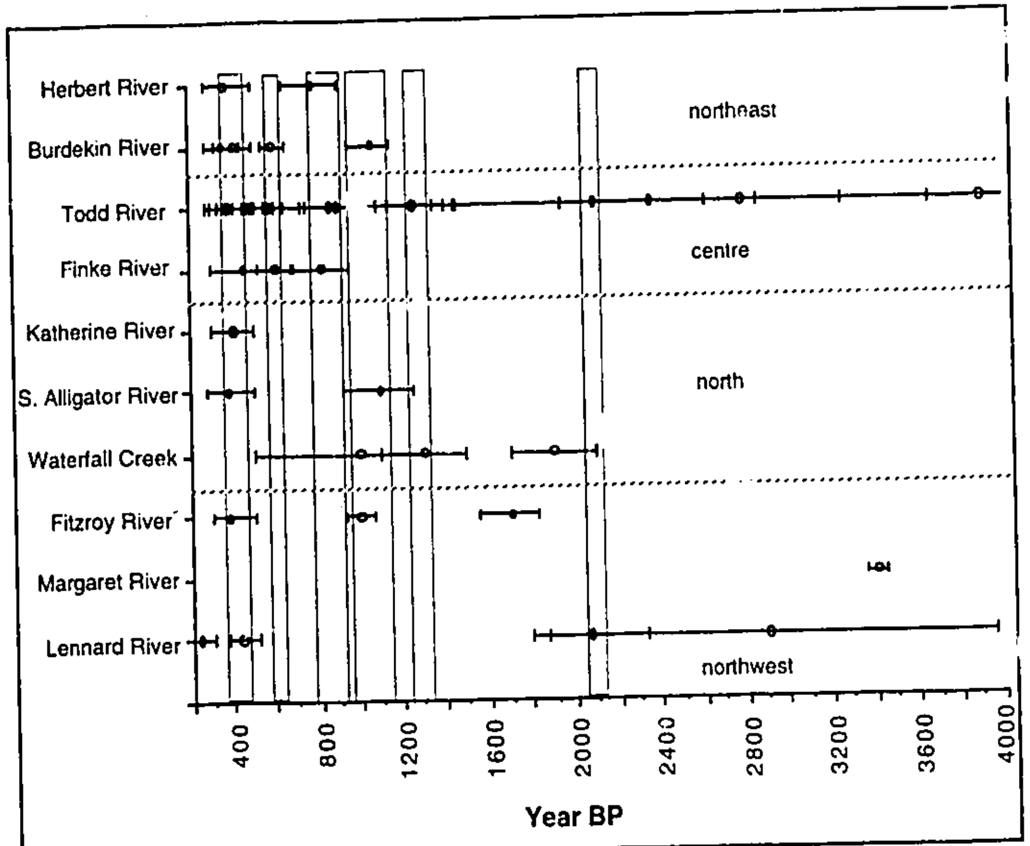


Figure 7.13. Coeval paleoflood phases in central and northern Australia. For source of data see Figure 7.12.

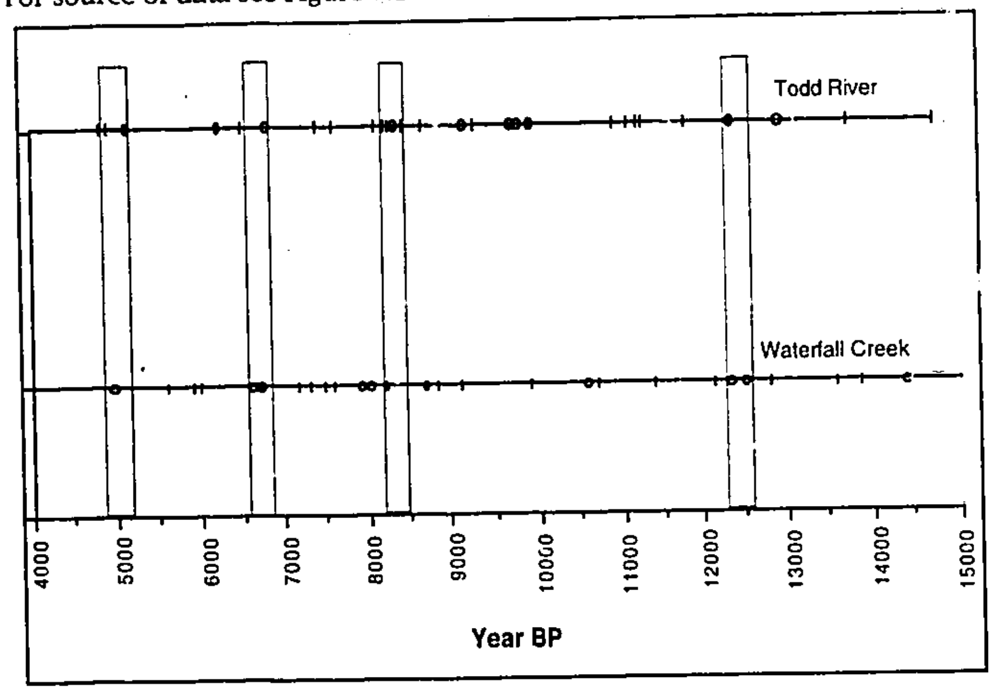


Figure 7.14. Coeval paleoflood phases 4 ka - 15 ka in central and northern Australia. North Australia data from Nott and Price (1994).

There is growing consensus that in the tropics and in extra tropical regions impacted by tropical flood-generating storms, the largest floods have occurred in the last century (Baker, 1995). This hypothesis is proposed on the basis of data gathered in many places including sites in the US (Ely *et al.*, 1993), India (Kale *et al.*, 1994; Ely *et al.*, 1996), and northern and central Australia (Pickup *et al.*, 1988; Wohl *et al.*, 1994c).

In monsoonal Australia, Wohl *et al.*, (1994c) reviewed all dated paleofloods that occurred during the last 2 ka BP but they ignored the age error terms in the analysis and could only tentatively suggest that there has been a higher frequency of floods during the last 150 years. Pickup *et al.* (1988) considered that three of the four largest floods in the Finke Gorge occurred in the last 100 years (1921, 1974, 1988) and Gale and Bainbridge (1994) suggest that the 1990 floods of inland Australia which inundated more than 22000 km² of southern Queensland and northern New South Wales were possibly the largest since European settlement of the region 200 years ago. This does not match the flood data of the Todd catchment as the 10 ka flood that is represented by the Giles B splay and longitudinal bar deposit, the expansion bar deposit at the end of the Mosquito Bore Distributary and the Expansion Scour deposit which is estimated to have deposited $8.6^{-6}m^3$ of sediment, far exceeds any historical flood. Indeed no category 3 paleofloods have occurred historically.

Three reasons have been suggested to explain the apparent clustering of exceptionally large floods in the last century (Pickup *et al.*, 1988; Wohl *et al.*, 1994c. Firstly, a climatic shift that increased the frequency of large flood generating storms is a popular assumption but one which has been impossible to prove. Such a climatic shift might be natural or the result of atmospheric warming induced by increases in the atmospheric concentrations of radiatively active gasses such as carbon dioxide. As these flood patterns have been linked to global hydro-climatic phenomena associated with tropical storms, clearly the delineation and understanding of the causes of change are important. Secondly, in the Finke, a recent combination of drought and overgrazing has enhanced runoff production in the drainage system. While it is likely that land use, both pastoral and traditional aboriginal fire-stick-farming, will increase sediment yield, high magnitude floods in large catchments such as the Finke are far less influenced by drainage basin properties than the more frequent lower magnitude events (Pickup *et al.*, 1988). The third factor may be that the small sample size increases the possibility that only the most recent events are recorded.

While an analysis of the flood data from the shorter flood records (<2 ka) does seem to indicate that the recent floods are the largest floods in many of the catchments it is not as yet a consistent record across the tropical and sub tropical region. Certainly, the high magnitude of the 1970 floods in the Finke were not as large across the eastern MacDonnell Ranges and Nanson and Tooth (in press) report that there has been little change to the Sandover River since 1950. The lack of synchronicity in terms of the magnitude of events within and between regions indicates that more field research needs to be undertaken in locations where a longer temporal record can be assured before we can invoke atmospheric scale climatic change due to the greenhouse effect or any other mechanism.

CHAPTER 8

Second Order Geomorphic Response to Paleofloods

8.1. Introduction

This chapter describes the second order response of the catchment to the paleofloods (Fig. 8.1). This is the first study to assess the recovery and long-term geomorphic effects of a sequence of high magnitude floods that occurred beyond the historical record in an arid zone catchment. The first section of the chapter describes the adjustment of small-scale catchment drainage patterns, the formation of aeolian dunes, and the change in vegetation. This is followed by a discussion of the modern channel adjustments to and recovery from changes in local base level and sediment supply. The chapter finishes by examining the impact of the paleofloods on the Aboriginal occupants of central Australia and the implications for the archaeological record.

During recovery from a flood, all parts of the catchment do not exhibit the same trends at the same time nor will any part of the stream necessarily exhibit the same tendency at different times. This spatial and temporal variability in behaviour has been labelled by Schumm (1977) as a complex response. While the length of the interval between catastrophic events will influence the degree to which channels are changed (Gupta and Fox, 1974, Wolman and Gerson, 1978), in humid regions moderate events can rapidly return a channel to its pre-flood condition (Costa, 1974), whereas recovery in arid and semi-arid environments generally takes much longer (Wolman and Gerson, 1978; Harvey, 1984). However, the ability of the river to recover is dependent on the magnitude of change wrought by the flood. For example, if the floods have eroded adjacent longitudinal dunes which were built during extended arid phases this landscape will not be repaired by intervening floods; similarly if the channel avulsed during the flow, repositioning of the channel may never be achieved. Sequences of large floods may be an important factor in major channel changes and geomorphic effectiveness (Burkham, 1972) as the catchment may have been unable to recover recover in the short period between two large floods.

8.2. Small-scale drainage systems

Small (<20 km^2) systems drain the piedmont ranges and outlying ridges south of the MacDonnell Ranges. These channels drain steep bedrock gullies and flow onto the piedmont area, aeolian dune fields (Appendix 4, Image 9) and trunk-stream alluvium (Appendix 4, Image 5) forming small piedmont fans. The drainage eventually disintegrates in a system of distributary channels where channel dimensions decrease downstream to a terminal floodout. Observations in the field and from satellite photographs indicate that where these small channels intersect with paleoflood alluvium the drainage pattern may disintegrate or it may be diverted. The following examples are selected to illustrate the paleoflood-induced adjustment of small drainage patterns.

Drainage disintegration is a common effect in the Todd catchment where the small channels intersect paleoflood alluvium. This can be seen in the channel at Camel Flat which drains from the high bedrock ridge east of Wallaby Gap (Appendix 4, Image 9; Fig. 8.2a). The channel flows into the swales of a longitudinal dune field and maintains a single thread channel pattern until it meets the Mosquito Bore Distributary deposit. At the point of change from the inter-dune swale to paleoflood sediments the drainage system rapidly disintegrates (Fig. 8.2a) to form a densely vegetated terminal floodout. The disintegration of channel form is related to the rapid increase in transmission loss into the channel bed once it flows over the more permeable paleoflood sediments.

Where small-scale channels intersect larger paleoflood channels they may occupy topographic lows in the paleoflood deposit. An example of this is seen along the Mosquito Bore Distributary where a small channel which drains from the west south of Mosquito Bore (Appendix 4, Image 8a) flows along the marginal paleoflood channel of reach 2 (Fig. 8.2b, Plate 6.19). Another example can be seen upstream of the Wallaby Gap constriction (Appendix 4, Image 9a) where a small westerly-draining channel is abruptly diverted south along a paleoflood braid channel. In this way the paleoflood braid network diverts the smaller channel network and dictates the location of the channel and the direction of flow.

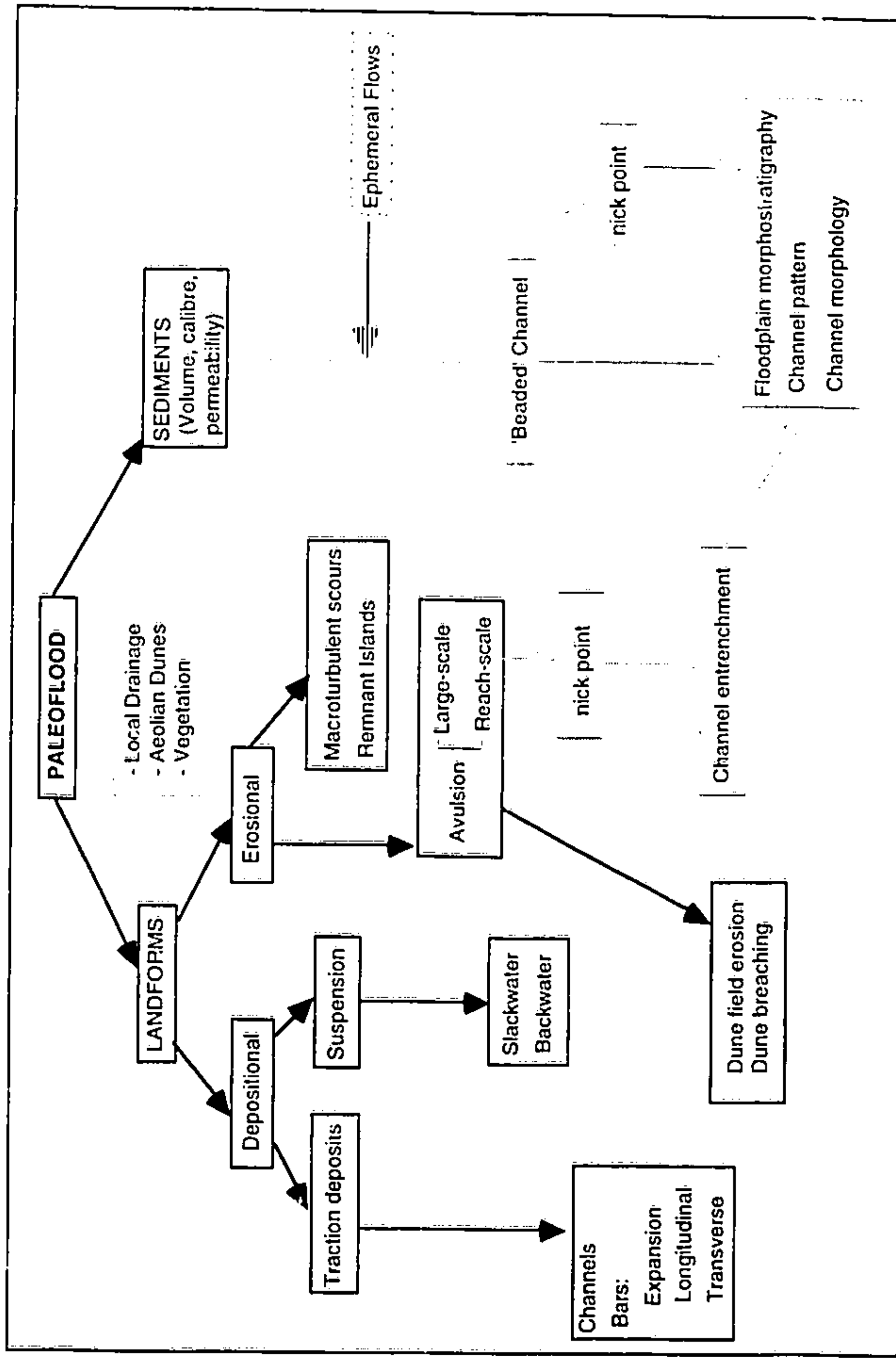


Figure 8.1. First order (black line) and second order (grey line) effects of the Todd River paleofloods.

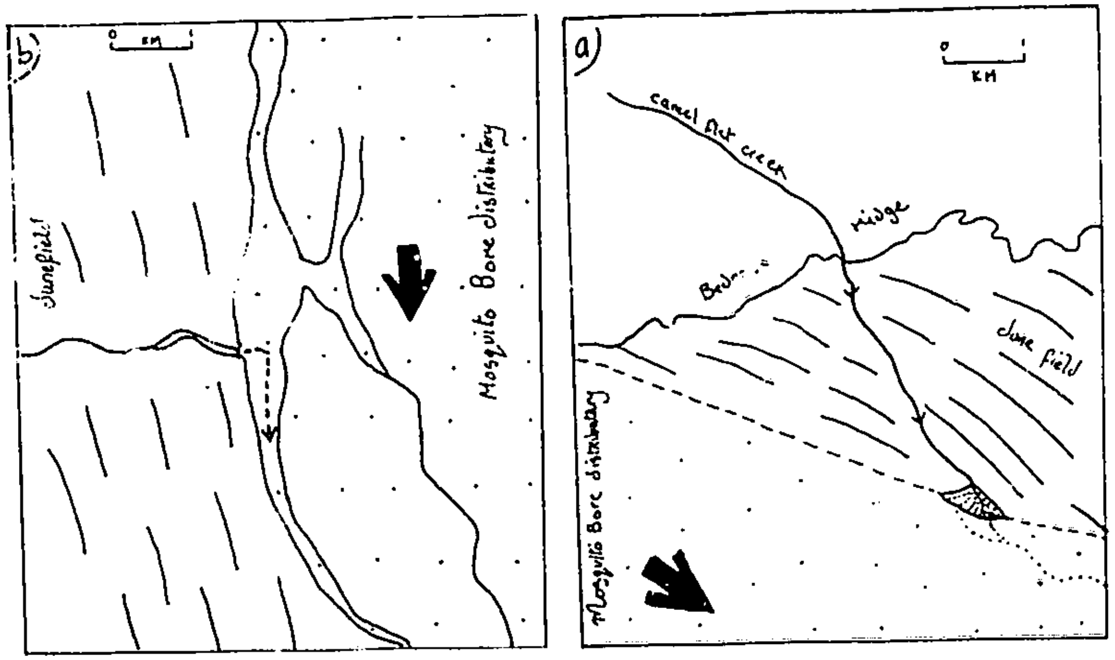


Figure 8.2. Second order effects on small catchments.

a) Drainage disintegration on Camel Flat Creek, south of Wallaby Gap.

b) Drainage diversion on the Mosquito Bore Distributary 10 km south of Mosquito Bore.

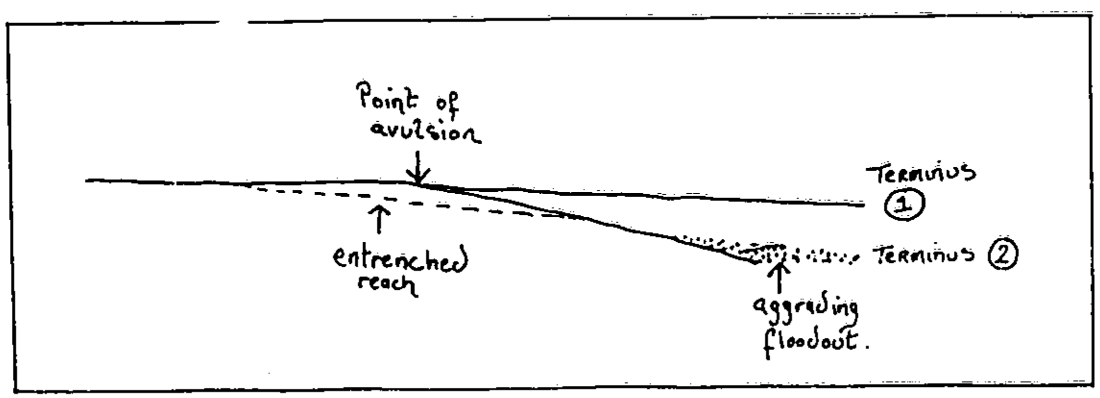


Figure 8.3. Avulsion and channel entrenchment induced by paleofloods.

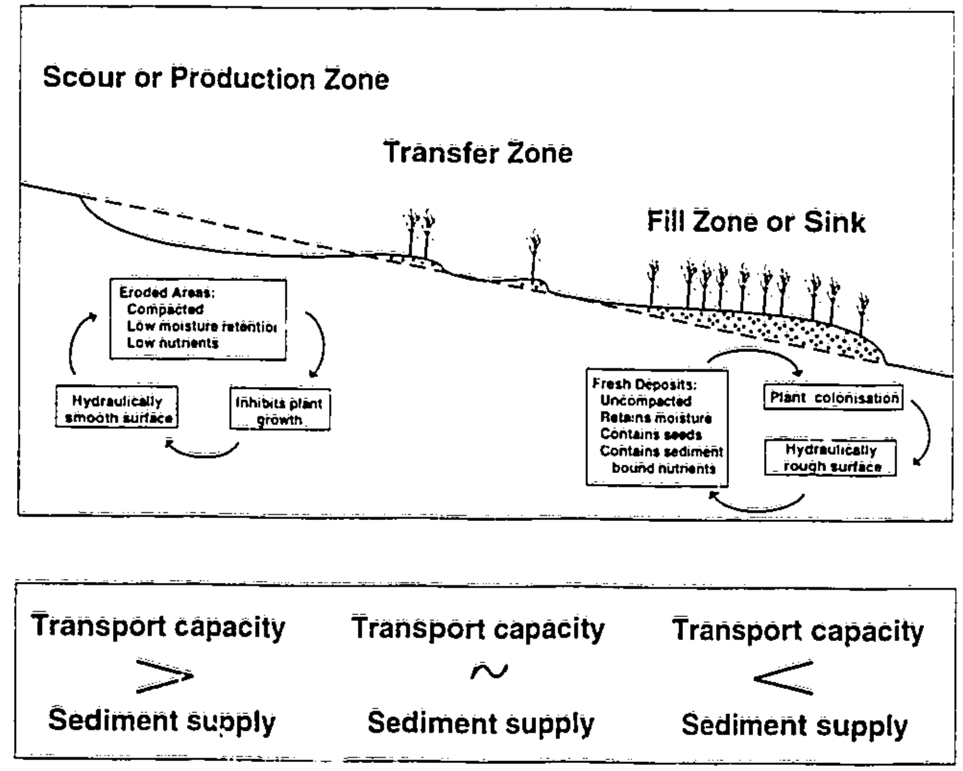


Figure 8.4. Schematic model of an erosion cell (Pickup, 1985). The scour zone is an area of erosion and actively sheds both water and sediment. The transport zone occurs down slope and is occupied by discontinuous patches of sediment in transit or in temporary storage. The fill zone lies downslope and is an area of sediment deposition.

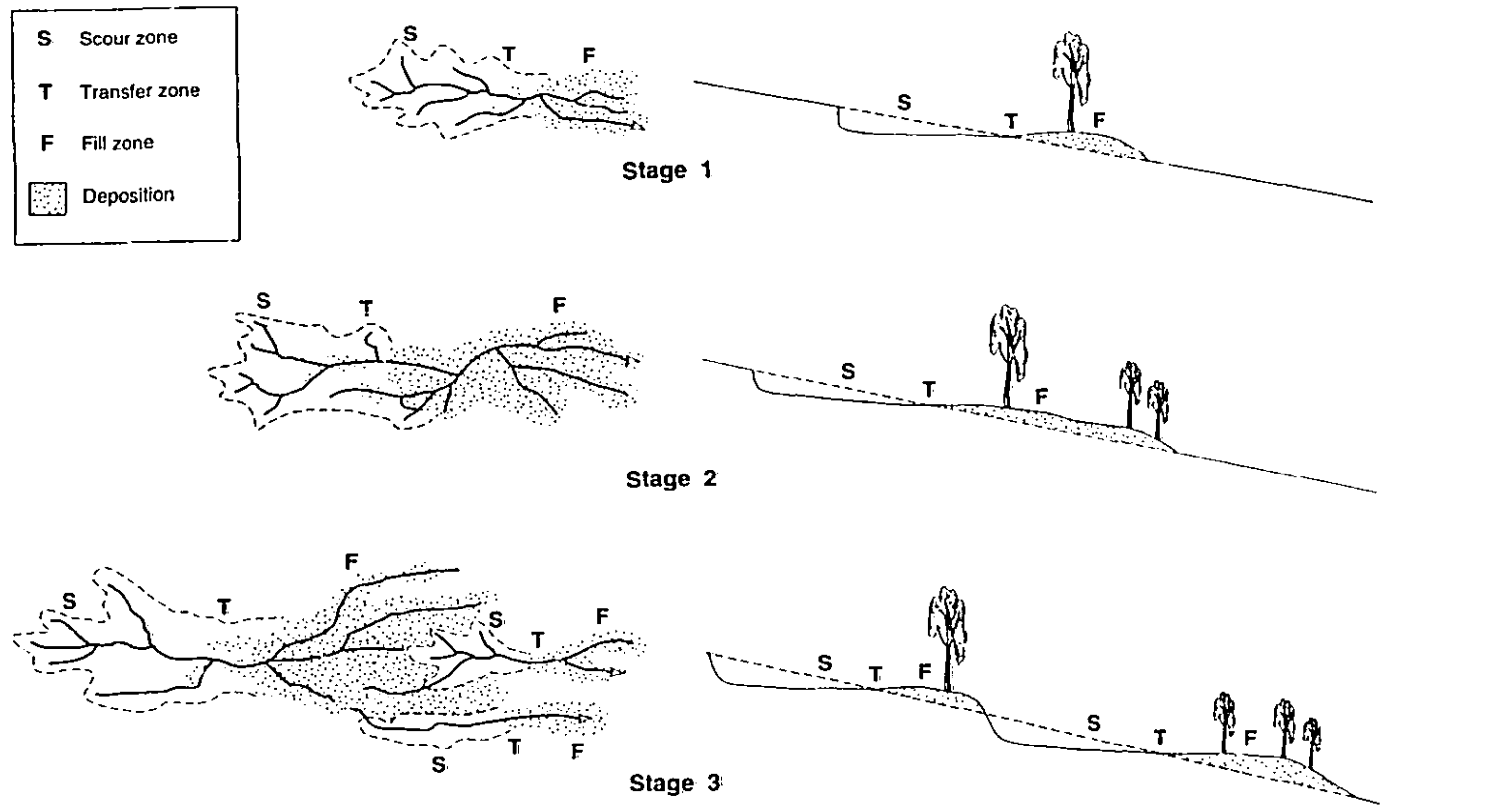


Figure 8.5. The temporally variable landform assemblage in an erosion cell mosaic in a flat area such as an alluvial fan or flood plain (Pickup, 1985). The shifting pattern is determined by the evolution of the erosion cell over time and the complex interaction of large and small scale erosion cells. Stage 1: Erosion cell established as small p.oduction, transfer and sink zones. All three zones begin to extend upslope. Stage 2: Upslope extension continues. The sink zone is elongated and no longer aggrades at the downstream end, but receives sediment deficient flow. Gully initiation in distal reaches. Stage 3: Development of new erosion cell. Primary sink may become inactive as new erosion cells capture diverted runoff.

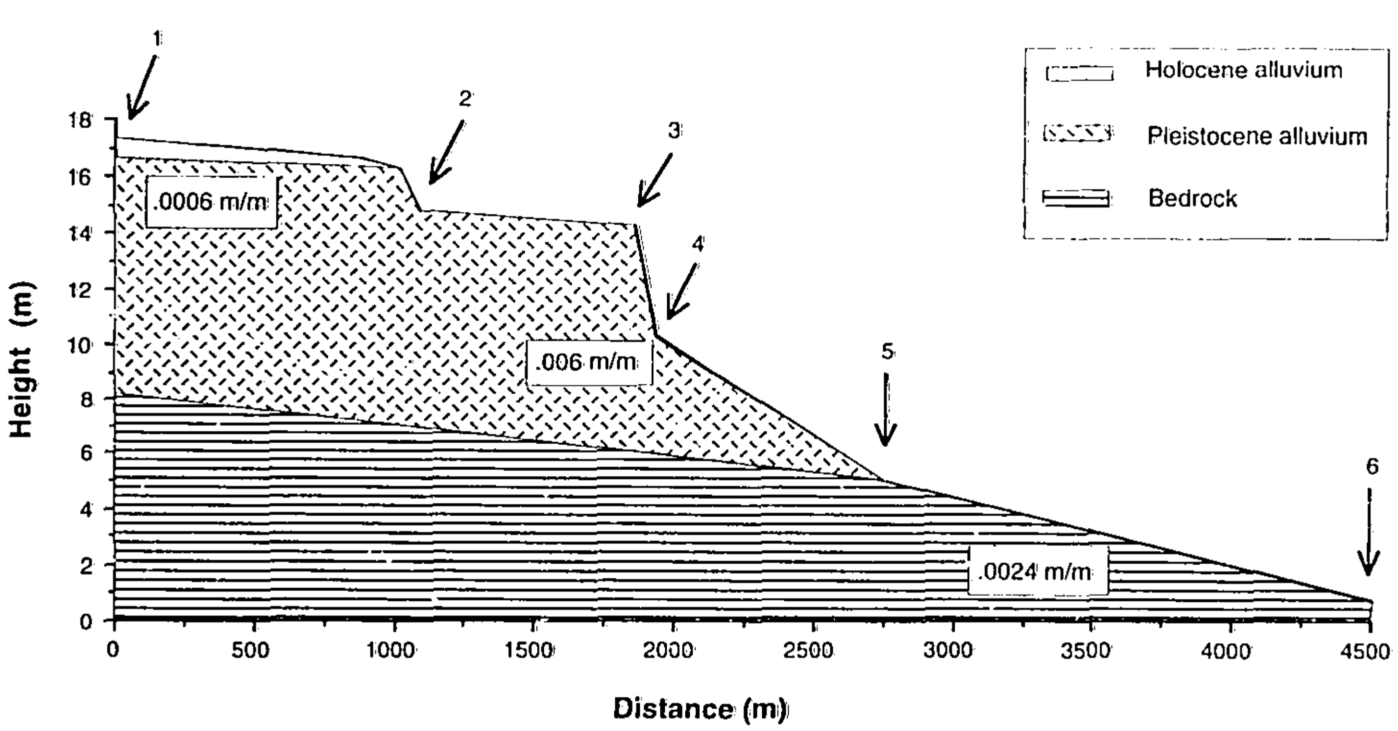


Figure 8.6. Long profile fr m Heavitree floodout downstream to bedrock channel. Arrows indicate the location of cross sections in Figure 8.7.

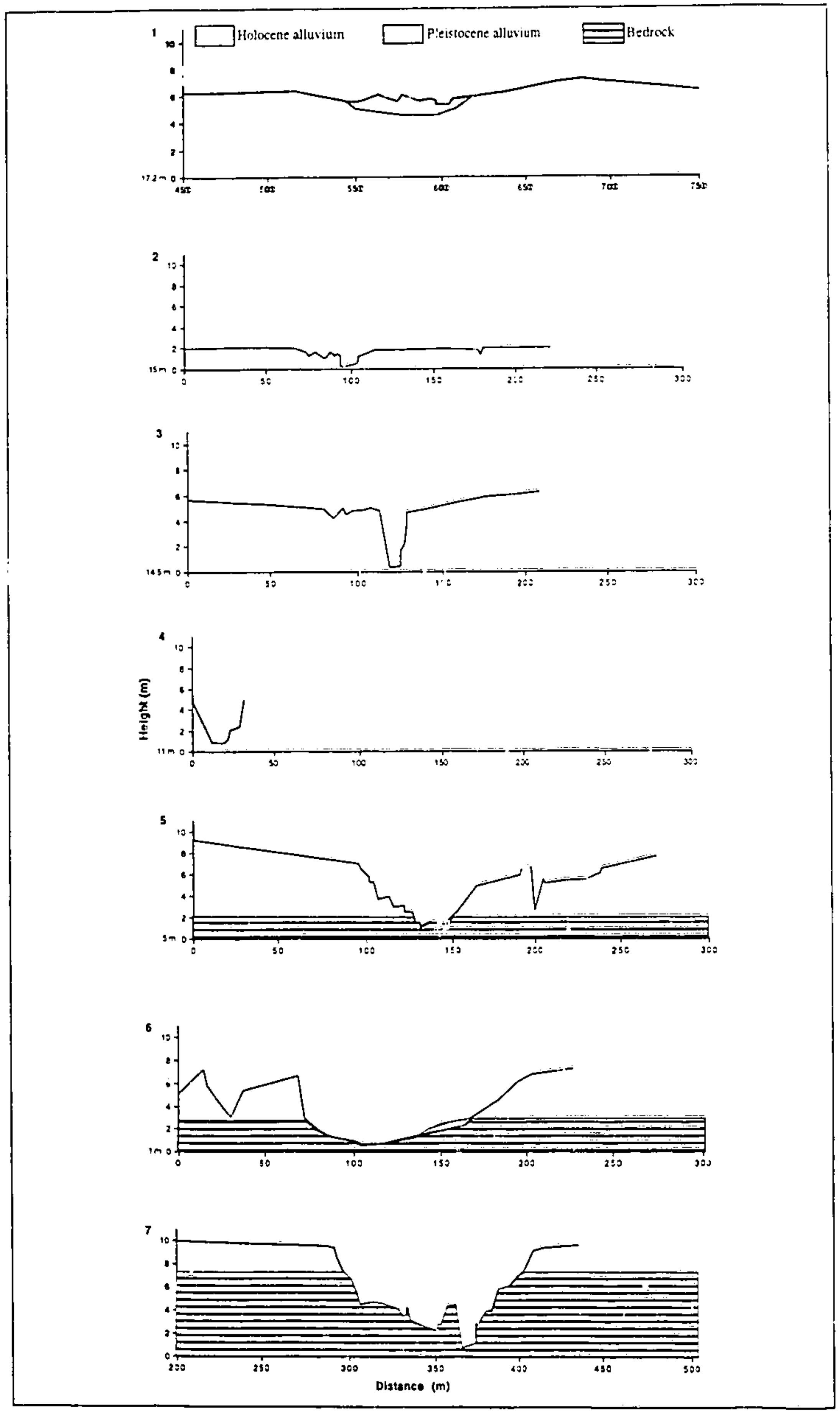


Figure 8.7. Channel cross sections from the Heavitree floodout to the bedrock channel.

8.3. Aeolian dunes

The surfaces of all paleoflood deposits have been reworked to some degree by aeolian processes. This is indicated by winnowed surfaces, small aeolian mounds trapped around vegetation and low dunes on paleoflood surfaces. Most impressive are source-bordering climbing dunes located close to paleoflood channels where adjacent high bedrock ridges trap the mobile sand fraction.

Large source-bordering dunes were investigated adjacent to Jessie Gap, (see section 6.2.4, Plate 6.2) and No. 5 Bore paleoflood channel, (see section 6.6.5 2., Plate 6.27). The dunes at No. 5 Bore are believed to postdate deposition of sediment in the adjacent paleoflood channel (Fig. 6.61) and are related to fluvial events which introduce sediment rather than signalling a period of increased aridity or windiness. Radiocarbon dates from the Jessie Quarry dune which has its source in Jessie Creek, indicate that it was forming at 3087-2775 cal BP and 1874-886 cal BP. Source-bordering dune sediment varies between fine and coarse moderately and poorly sorted sand (Table 8.1). The sand in the Holocene dune close to Jessie Gap is coarser than that of the underlying Pleistocene dune owing to the proximity of the Holocene dune to its sediment source. The Pleistocene dune is part of the larger Simpson Desert dune field, nourished from a more regional source. The Holocene dunes are reworked by small runoff gullies which transport the higher elevation aeolian deposits into small fan platforms on the lower climbing dune flank, enhancing its complex slope morphology.

Sample Name	Mean (phi)	Sorting	
DPc 1.5 (i) *No. 5 Bore* *Holocene Dune*	2.82	.83	Moderately sorted fine sand
Quarry B (ii) *Jessie Gap* *Holocene Dune*	0.81	0.75	Moderately sorted coarse sand
Quarry Ci *Jessie Gap* *Pleistocene Dune*	1.38	1.21	Poorly sorted medium sand

Table 8.1. Sediment size in aeolian dunes

On the ridge to the east of reach 1 on the Mosquito Bore Distributary is a series of steep fans (Appendix 4, Image 8). The sediment in these fans is derived from two locations: the angular coarse gravel fraction from the weathered rubble scattered across the ridge and a fine to medium aeolian sand matrix from the adjacent paleoflood alluvium which OSL age estimates show was deposited between 10 and 6 ka BP. Runoff has reworked the aeolian sediment into a series of piedmont fans which now abut the paleoflood channel. ther descriptions of alluvial f. ns nourished by aeolian processes were not found in the literature.

8.4. Vegetation

The relatively nutrient- and seed- rich, permeable paleoflood sediments (Bourke and Pickup, in press) support vegetation which differs from that of the surrounding older landforms. For example the older Pleistocene alluvial surfaces support *Hakea eyreana* and some *Acacia estrophiolata* and the aeolian surfaces *Micromyrtus flaviflora, Duboisia hopwoodii, Plectrachne schinzii.* The paleoflood surfaces support *Acacia murrayana, Acacia victoriae, Hakea eyreana, Acacia estrophiolata, Eucalyptus camaldulensis* and *Eucalyptus microtheca.* The effect on Aboriginal settlement of the arid interior of the change in vegetation distribution in the last 15,000 years, owing to an increase in flood frequency and magnitude, is discussed in section 8.6.

8.5. Manifestations of channel disequilibrium

Downstream changes of channel morphology, planform and sediments are step-like in the Todd River. The channel bed is downcutting where it flows over bedrock or cemented Pleistocene alluvium, while elsewhere it is locally aggrading, with metre-scale thick sequences of sandy tabular bars, and channel width tends to increase and decrease abruptly. Factors affecting this variability include sediment transport, local sediment supply, bedrock outcrops and entrenchment.

The following sections focus on three aspects of channel disequilibrium in the Todd and discusses the way in which the channel changes between high magnitude events. The first topic is channel entrenchment, second is the response to sediment injected by paleofloods and the third concerns the complex response at tributary confluences.

8.5.1. Channel entrenchment.

Channel entrenchment can be associated with an increase in flood frequency. For example Warner (1987) inferred that recent channel incision in some New South Wales coastal valleys reflects the influence of large floods since the 1940's. However, channel response in semi-arid regions may be different: for example, Pickup (1991) inferred from sedimentary evidence in dams in central Australia that entrenchment occurs upstream and deposits downstream in dry phases. Sediment trapped in the dams includes a lower layer of compacted mud heavily trampled and mixed by cattle, which accumulated during a relatively dry interval prior to 1967. This is overlain by stratified sand. Pickup inferred that during the drought prior to 1967, there was less vegetation and grazing enhanced erosion so that channels became choked with sediment and only silt and cla were transported. During the 1970's, Alice Springs experienced two to three times its average annual rainfall for several consecutive years and the landscape became revegetated. Erosion was reduced, sediment delivery to channels decreased and channel :our commenced releasing sand which was transported to the dam. Similarly, but at a larger scale, the first order effects of the Todd River paleofloods were aggradation of flood plains and channels rather than incision, which developed later.

Entrenched reaches are identified where the channel and recent flood plains are inset within high (>5 m), usually vertically walled, terraces composed of cemented Pleistocene or paleoflood sediments. Three mechanisms of entrenchment and nick point retreat are identified: avulsion-induced entrenchment, arroyo-style entrenchment and the incision of inherited steeper Pleistocene surfaces. Channel entrenchment releases large volumes of sediment to the system, which when transported downstream has a profound effect on channel form.

8.5.1.1. Avulsion induced channel changes

One of the major geomor. ic effects of paleofloods was avulsion. Reach-scale avulsion has already been discussed for the Giles Creek paleoflood channel fill and the No. 5 bore section (section 7.3.3.1.1). Large scale avulsion changed the flow path position from event to event. In addition to relocating the channel, avulsion sometimes changed the depositional end point of the system to a point of lower elevation (see Chapter 7). It is inferred that this process would have triggered widespread channel entrenchment in a manner similar to that shown in Figure 8.3 but specific examples are difficult to identify in the field.

8.5.1.2. Arroyo-style entrenchment

Working in eastern Wyoming and New Mexico, Schumm and Hadley (1957) observed the coexistence of both aggrading and entrenched channels. They proposed that the formation of a discontinuous gully was a semicyclic phenomenon, whereby an initial gully migrates headwards so that tributaries become entrenched and the rate of sediment removal is increased. When the supply of sediment from retreating headcuts exceeds the capacity of flows to remove it, deposition occurs in the lower reaches of the gully. The alluviating valley floor develops a local steep gradient at the downstream end which subsequently erodes, forming a trench which may eventually unite with the upper channel to form a continuous channel dissecting the channel fill.

Aggradation and entrenching are related to sediment supp which in arid fluvial systems is episodic and spatially variable, and is strongly linked to patterns of vegetation cover. Pickup (1985, 1988) proposed that the principles derived from arroyos and discontinuous gully systems by Schumm and Hadley (1957) apply in the larger, lower gradient fluvial systems of central Australia. Pickup's model, shown in Figure 8.4, uses the three ›nes identified by Schumm (1977), a sediment production zone where scour or stripping injects sediment, a sediment transfer zone and a sediment fill zone or sink. Newly deposited sediment in the fill zone is uncompacted and therefore can hold more moisture than older deposits; it may also contain seeds and a higher proportion of sediment-bound nutrients, and provides a favourable environment for vegetation which tends to stabilise the surface. The vegetated surface is also hydraulically rough which disperses subsequent flows and promotes further deposition (Fig. 8.4).

Eroded areas (scour zones) provide unstable surfaces for plants, are often hydraulically smooth and have limited capacity to capture and retain low density plant matter such as seeds. Taken together, the tendency for sediment to be transported and deposited in slugs, and the effect of stabilisation by plants, lead to alternating patterns of erosion and deposition of varying persistence in the landscape. These patterns may develop at a

variety of scales with the larger elements in the pattern becoming more pronounced as erosion intensifies.

The operation of scour-transport-fill (STF) sequences in arid zone rivers differs from perennial rivers in both the scale of operation and the complexity of spatial patterning. In perennial systems, the STF sequence is generally manifest as chains of ponds within unchanneled swampy meadows (Eyles, 1977), and by within-channel but vegetated ephemeral sediment islands, whereas in arid systems it operates across a number of scales, from the gully network to the large-scale floodout system. These scales are not exclusive and may have complex spatial relationships, for example, the distal zone of one floodout may be the production zone for another erosion cell sequence or smaller cells may be embedded within larger ones (Fig. 8.5). Erosion cells may not be fully developed or may represent latent features in the landscape (Pickup, 1988). The transition between linked cells tends to be morphologically abrupt. This patterning of arid zone sediment transport and deposition produces a shifting mosaic of ground surface sediments known as an erosion cell mosaic (Pickup, 1985, Fig. 8.5).

This model is useful in understanding floodout formation. Intermediate floodouts are located in the piedmont zone of the MacDonnell Ranges. The channel sequence through the headwaters - piedmont fan - floodout, close to the MacDonnell Ranges represents a physical expression of Pickup's (1988) STF model (see Fig. 6.1b). The bare granitic rock surfaces in the steep gradient headwaters of the Todd River upstream of Alice Springs, comprise the scour zone and supply coarse weathered sediment to the channel through a series of small-scale fans feeding into a network of gullies. This, in itself, is a small-scale version of the STF sequence. The main channel of the Todd River transports this sediment, along with some locally reworked alluvium contained in temporary storage in the transport zone. Once the channel emerges from the ranges it crosses the broad piedmont zone over a deep alluvial fill. Transmission losses are significant and deposition occurs in the fill zone. The intermediate floodouts of the Todd catchment are noted for their dense distribution of Coolabah vegetation (*Eucalyptus microtheca*), decreasing channel capacities, distributary channel patterns and extensive aggrading silty sand deposits.

The fc ɔwing is an example of the geomorphic impact of channel entrenchment downstream of a zone of aggradation in the Todd River. The Anabranching site is located in the entrenched channel formed as a result of the reach scale avulsion at the No. 5 paleochannel approximately one thousand years ago (section 7.3.1.1.). The channel width increases abruptly in this reach and contains a series of large islands separated by smaller channels (Plate 8.1). The cross section indicates a narrow channel incised into the underlying Pleistocene sediment (Fig. 4.22). This cross section identifies the way in which channel beads act as stores for sediments that are transported in pulses downstream and the importance of entrenchment in remobilising those stores. There is a nick point developed not only in the channel thalweg but also on the higher flood plain surfaces that are activated during the larger magnitude events (Plate 8.2). The planform of the bead suggests that channel widening is along the left bank as a result of aggradation and flow deflection most probably during subsequent lower magnitude events. The exploitation of the surface channels is effective in remobilising the sedimen ore and leaving an assemblage of remnant islands. The rate of nick point incision appears rapid (Plate 8.2) and if not reworked, these islands may remain as perched terrace deposits or perhaps be infilled by finer textured sediments until the next pulse of sediment is moved downstream to be stored preferentially in this location.

8.5.1.3. The incision of inherited Pleistocene surfaces

The third mechanism of channel entrenchment is evident in the entrenched reach downstream from the Heavitree floodout (Fig. 6.4). Figure 8.6 shows the current location of a series of nick points close to the catchment headwaters. Figure 8.7 indicates the changing channel morphology as the nick point moves upstream. Cross section 1 is through the Heavitree floodout and indicate a multiple shallow channel network. Cross sections 2 to 4 indicates arroyo-style channel entrenchment in the indurated Pleistocene sediments. Cross sections 5 to 7 are incised in bedrock. Two well-developed nick points are visible in the long profile (Fig. 8.6); the upper one is at the junction between the Holocene floodout sediments and the Pleistocene alluvium and the lower one is incised in Pleistocene alluvium. A third nick point, not measured in the long profile, is located 500 m downstream of cross section 6 in the Merinee sandstone. Cross section 7 is not shown in Figure 8.6. The entrenchment of the channel along this 5 km reach is probably related to the difference in gradient between the steeper underlying Pleistocene surface fan surface and the overlying Holocene intermediate floodout terminus (Fig. 8.6).

8.5.2. Paleoflood sediment texture, supply and reworking

The second example of channel disequilibrium is the geomorphic response to the injection of large volumes of paleoflood sediment.

8.5.2.1. The Influence of paleoflood sediments on channel morphology

The weathering history and textural composition of the sediments through which the channel passes profoundly influence the channel morphology, planform and pattern. Where the boundary is composed of resistant sediment such as cemented Pleistocene alluvium, and at some sites coarse paleoflood sediments and bedrock, the channel is relatively narrow and the banks are high. Coarse paleoflood deposits provide an abundant source of sediments, but have variable boundary resistance. Because lesser flows have insufficient energy to transport coarse sediments easily, paleoflood deposits are a major influence on modern channel and flood plain morpho-stratigraphy. Two sites illustrate the effect of the coarse textured sediment: Giles Creek 3 (Fig. 8.8a) and Mosquito Bore 2 (Fig. 8.8b) where the present channel has incised paleoflood fill emplaced approximately 3 thousand years ago. The paleoflood sediments at both sites, composed of cobbles and coarse gravel, reflect their proximity to the ranges. Large remnant islands composed of *in situ* paleoflood cobbles and gravel remain 3 m above the thalweg (Fig. 8.8a (i)) and deflect the modern channel flow into an anastomosing pattern (Fig. 8.a (ii)). In these locations the present flood plain is often a thin smear (Fig. 8.8a (iii) and b (i)) over a laterally more extensive paleoflood core (Fig 8.8a (iv) and b (ii)). The morphology of exposed paleoflood sediment is often step-like (Fig. 8.8a (v) and b (iii)) due to the relative resistance of individual sedimentary layers. In these reaches active channel bars tend to be coarser in textural composition than further downstream due to the local source of coarse paleoflood sediments (Fig. 8.8b (iv, Plate 8.3).

Further downstream, the channel incises finer textured paleoflood sediments which are easier to erode and transport. Many of these locations are associated with channel aggradation and channel widening which are discussed further below. In these reaches the channel typically includes a series of braid bars or stable islands separated by smaller channels. Channel banks may show a series of elliptical notches resulting from flow deflection at the downstream end of these bars. Plate 8.1 shows an aerial view of a narrow channel entrenched in Pleistocene sediments; Plate 8.7 and 8.8 show wide reaches set in paleoflood sediments.

8.5.2.2. Channel width variability

One of the notable features of the Todd River is the variation of channel width, particularly downstream of the Ross River junction. This variability is linked to the distribution of paleoflood sediments. Channel width was measured from the headwaters upstream of Alice Springs to 170 km downstream at 500 m intervals (Fig 8.9) from 1:50,000 orthophoto maps. The following section explores the relationship between channel width variation, sediment characteristics and channel boundary resistance, particularly at locations where width increases dramatically. Factors which influence channel narrowing include resistant boundary sediments such as bedrock, cemented Pleistocene sediments and coarse compacted paleoflood sediments, and the development of incised channels due to reach avulsion or local steepening of channel gradient due to aggradation upstream. Factors which influence channel widening are the location of erodable channel banks and the temporary storage of sediment waves.

Channel beads

Of particular interest in the Todd River is the occurrence of reaches of abrupt widening and narrowing (Plate 8.4, Fig. 8.10b), here referred to as 'beads', a term adopted from the fluvioglacial literature which describes similar sudden expansions in subglacial channel widths as beaded eskers. Plates 8.4-8.8 illustrate the planform of some Todd River channel beads. By way of example the Allora Bore bead increases from a width of 120 m to 400 m and maintains this width for 1,000 m downstream where it abruptly narrows again to 120 m (Plate 8.4). Typical dimension relationships are as follows: $W_n/W_b = 0.25$ and $W_b/L_b = 0.4$, (W_n = width of narrow reach, W_b = width of bead L_b = length of bead; see Fig. 8.10a). These reaches are associated with alluviation in the form of bars and islands, braided channel patterns (Plate 8.1, 8.6) and a decrease in channel slope. Similar sharp increases in width were observed in Spanish rivers by Thornes (1974, 1976a, b, 1977, 1980) who noted an association with high sediment discharges, transmission losses and high magnitude floods, all of which are found in the Todd River. The wide beads in the Todd River typically show evidence of lateral erosion, channel deflection and channel aggradation. Low magnitude channel widening

Plate 8.1. An oblique view of the Anabranching site. A beaded channel located upstream from a bedrock constriction. Dashed line indicates the approximate location of the cross-section. Flow is towards the top of the plate. Arrow indicates location of Plate 8.2.

Plate 8.2. View looking upstream at a nick point incision into gravelly Pleistocene alluvium at the Anabranching site. The streamlined remnant downstream from the young (~5-10 yr) eucalyptus tree indicates the rapid rate at which the nick point incision is occurring in this reach.

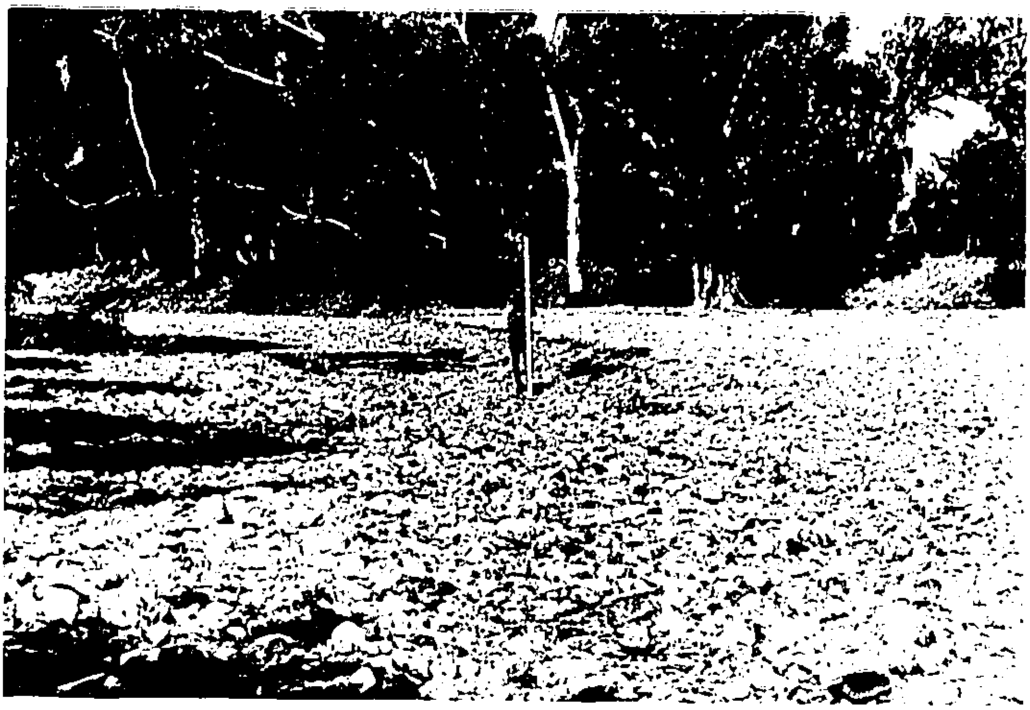


Plate 8.3. A headcut in coarse cobble and gravel paleoflood sediment Giles Creek. Flow is towards the left of the photo.

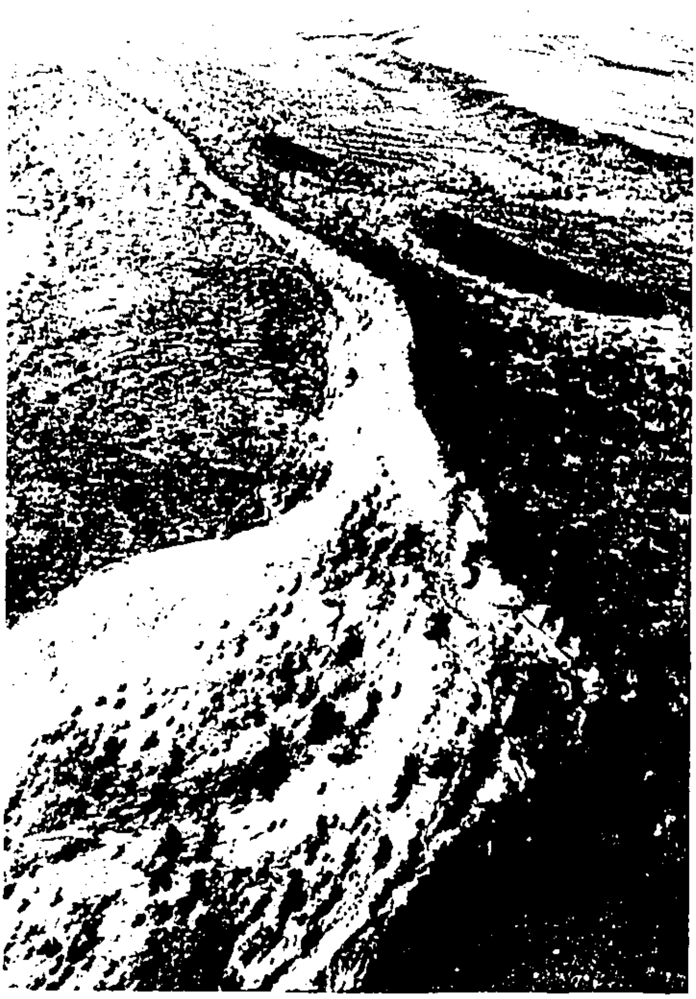


Plate 8.4. A oblique aerial photo of the Allora Bore channel bead. Note the abrupt nature of channel widening. Flow is towards the bottom left of the plate.

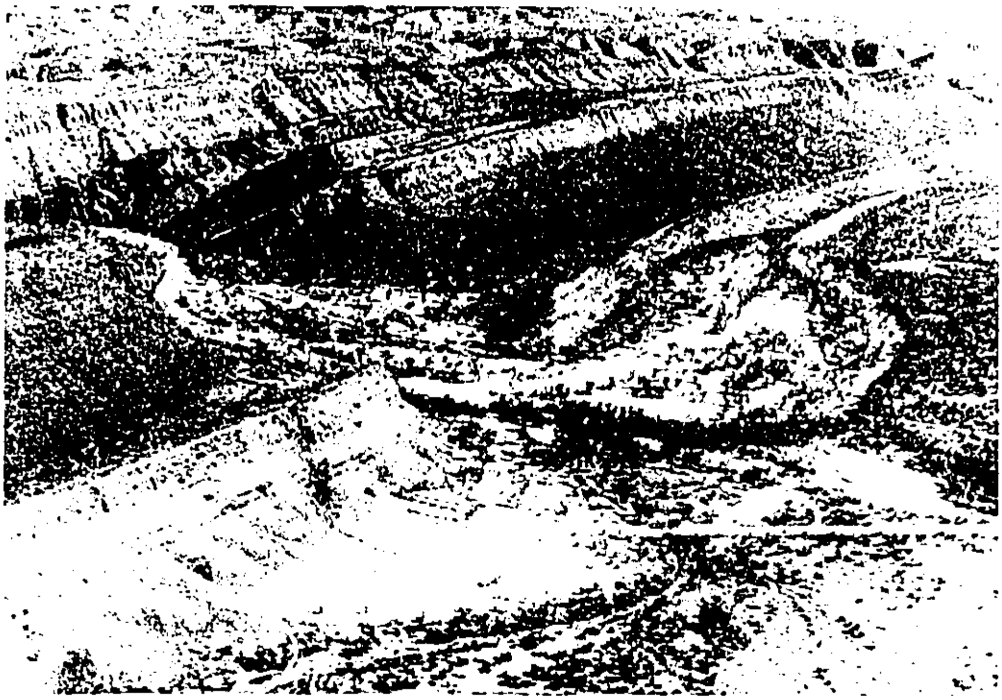


Plate 8.5. Oblique aerial view of the Ross River at entrance to Ross River Gorge. Flow is towards the left of the plate.

Plate 8.6. Oblique aerial view of channel 'bead' at entrance to Ross River Gorge. Flow is towards the right of the plate.

Plate 8.7. Channel bead located downstream from Giles Creek confluence indicating the impact of flow deflection on channel width. Flow is towards the bottom right of the plate.

Plate 8.8. Looking upstream along the Ross at the Ross/Todd confluence. Note the beaded channel planform associated with channel aggradation and bar formation. Flow is towards the bottom right of the plate.

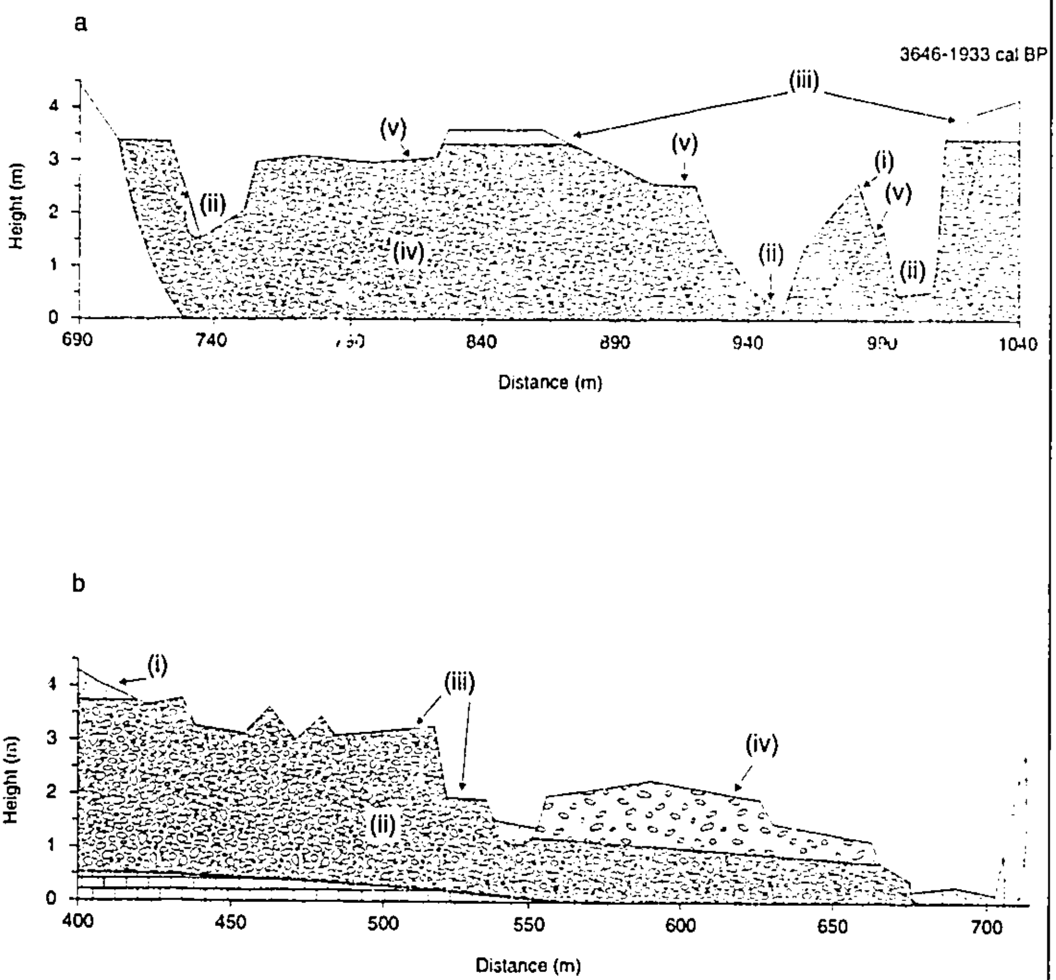


Figure 8.8. Morphostratigraphy indicating the impact of resistant paleoflood sediments on contemporary channel morphology and morphostratigraphy.

a) Giles Creek 3. (i) Island composed of remnant paleoflood sediments. (ii) locally anastomosing channel pattern. (iii) Modern flood plain smear. (iv) Flood plain core composed of remnant paleoflood sediments. (v) Stripped surfaces with step morphology (surface flood plain remnant).

b) Mosquito Bore 2. (i) Modern flood plain smear. (ii) Flood plain core composed of remnant paleoflood sediments. (iii) Stripped surfaces with step morphology, (surface flood plain remnant). (iv) Remobilised paleoflood gravels forming large channel bar.

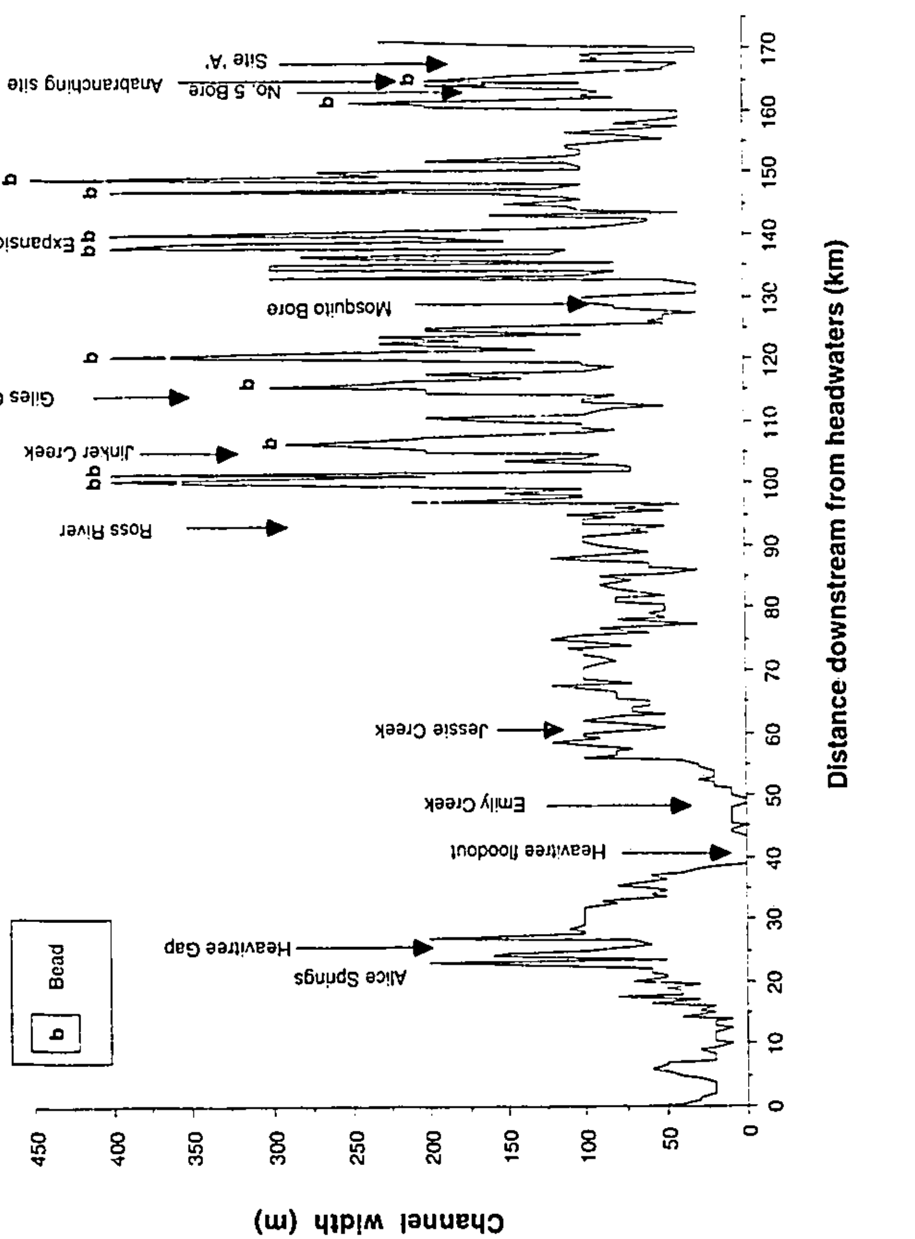


Figure 8.9. Channel width in the Todd catchment.

processes which include channel diversion, notching and undercutting also contribute to the development of beads.

The location of reaches of channel beads in the Todd is controlled by the following three factors: bedrock constrictions, sediment supply from tributaries and the inferred loss of flow energy (Fig. 8.11).

Upstream of the constriction at the head of the Ross River Gorge (Plates 8.5 and 8.6) channel width increases from 60 m to a maximum of 300 m presumably because the bedrock constriction has ponded the river and caused local channel aggradation. Beschta (1983a, b) reported a similar effect on the Kowai River in New Zealand where sites above constrictions such as bridges and gorges were sites of major aggradation and channel widening. In the Ross River, bedrock constrains the right side of the bead and the thalweg channel is deflected along the left bank (Plate 8.5) where it cuts into Holocene alluvium. Channel bars and a tendency to braiding can be seen in Plate 8.6.

Downstream of major tributaries, the Todd River is prone to aggradation and the development of braided beads with large bars or islands followed downstream by an abrupt narrowing of the channel. This can be seen on the satellite images where aggradation is evident downstream from the Ross River (Appendix 4, Image 5a) Jinker Creek (Appendix 4, Image 5a) and Giles Creek (Appendix 4, Image 7a). Similarly, Beschta (1983a, b) found sediment input from tributaries of the Kowai River caused downstream channel widening of the trunk stream and Mabbutt (1977) found the decline in channel width on the Finke River was reversed at major river junctions.

Channel beads not associated with constrictions or tributaries are inferred to form as a result of a decrease in transport energy downstream, in particular where aggradation coincides with erodable channel boundaries (Plate 8.7). Channel bar formation and the lateral diversion of flow important in widening the Todd channel where the banks are composed of erodable paleoflood sediments (Plate 8.5). Because an increase in sediment derived from local bank erosion augments the supply from upstream, braiding occurs and longitudinal bars develop, which enhance the lateral migration and widening of the channel. Such mid channel bars may continue to enlarge either by supply of sediment from upstream or locally from the eroded channel banks, thereby enhancing the lateral erosion. Coalescing bars may develop resulting in asymmetrical expansion of a portion of the channel, as is found on the Ross River channel bar (Plate 8.8). Similarly, Beschta (1983a, b) noted the positive feedback whereby in locations of channel aggradation additional sediment became available as the river cut laterally into formerly inaccessible sediments. In locations of channel aggradation where flow is diverted around large channel bars low magnitude flows diverted into side channels can effect significant erosion (Plate 8.7). Pickup (1988) described the process of 'buffering' for flat arid and semi-arid landscapes, which can be applied to reaches of extreme channel width increase or 'beads' on the Todd River. Sediment pulses tend to travel a short distance downstream and, in order to compensate for the increase in local sediment supply, the channel adjusts its pattern from a dominantly single thread to one of extended braiding sections; sediments are placed in stores such as islands, flood plains and large central bar systems (Pickup, 1985). Buffering occurs once depositional or fill areas become vegetated as they act as a bottle neck which eventually restricts the sediment volume and transport distance. It is common to find deposits building up so much that flow is diverted around them (Pickup, 1985, 1988) eroding the channel boundary and forming abrupt increases in channel width described as channel 'beads'.

8.5.2.3. Residence time of sediment in stores

In arid regions transmission losses tend to cause lobes of sieve deposits to form in the main channel. Their position and permanency relate to the frequency and magnitude of flows (Thornes, 1977; Finley and Gustavon, 1983). Similarly, mid-channel bars, sediment splays and aggraded channel beads are ephemeral sediment stores in the Todd channel. Paleoflood deposits comprise large sediment stores, which are reactivated by lateral channel erosion, channel avulsions and/or flood plain stripping events.

Channel aggradation deposits are initially reworked by the succeeding low magnitude flows which incise a narrow channel in the former channel bed thus indicating a low residence time at this scale. Once colonised by vegetation or incorporated into a more stable form such as an island, residence time increases. A radiocarbon age of ~300 BP was estimated for a sample from the channel fill at the base of an island in the Anabranching bead (sample '95 C1, Table 4.4b) indicating at least 3 m of incision since that time and an inferred residence time of ~300 years in these bead stores. It also suggests that this pulse of sediment may have been mobilised by the series of high magnitude floods which occurred approximately 400 BP. The tendency

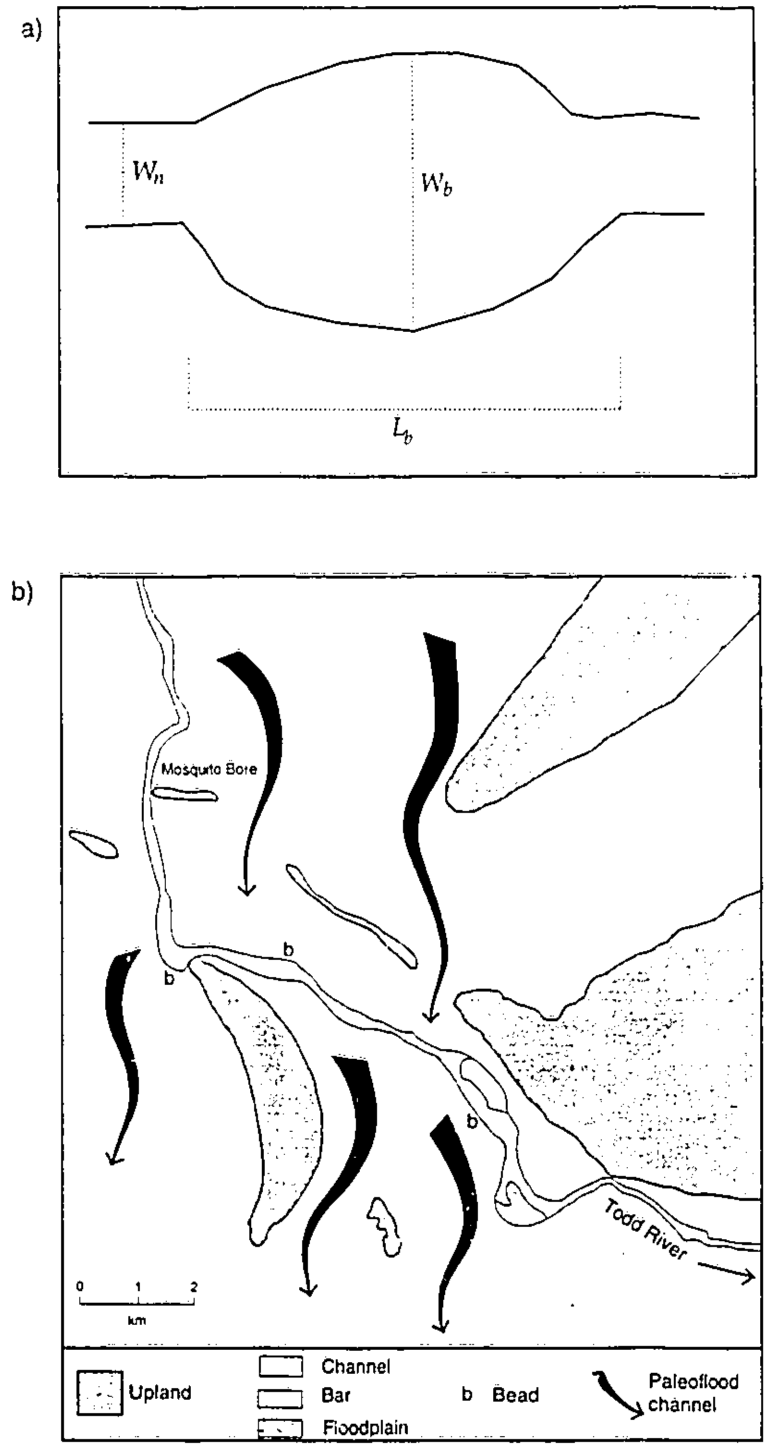


Figure 8.10. Channel beads.

a) Bead planform: Width of narrow reach(W_n), width of bead (W_b), length of bead (L_b)

b) Channel beads downstream from Mosquito Bore.

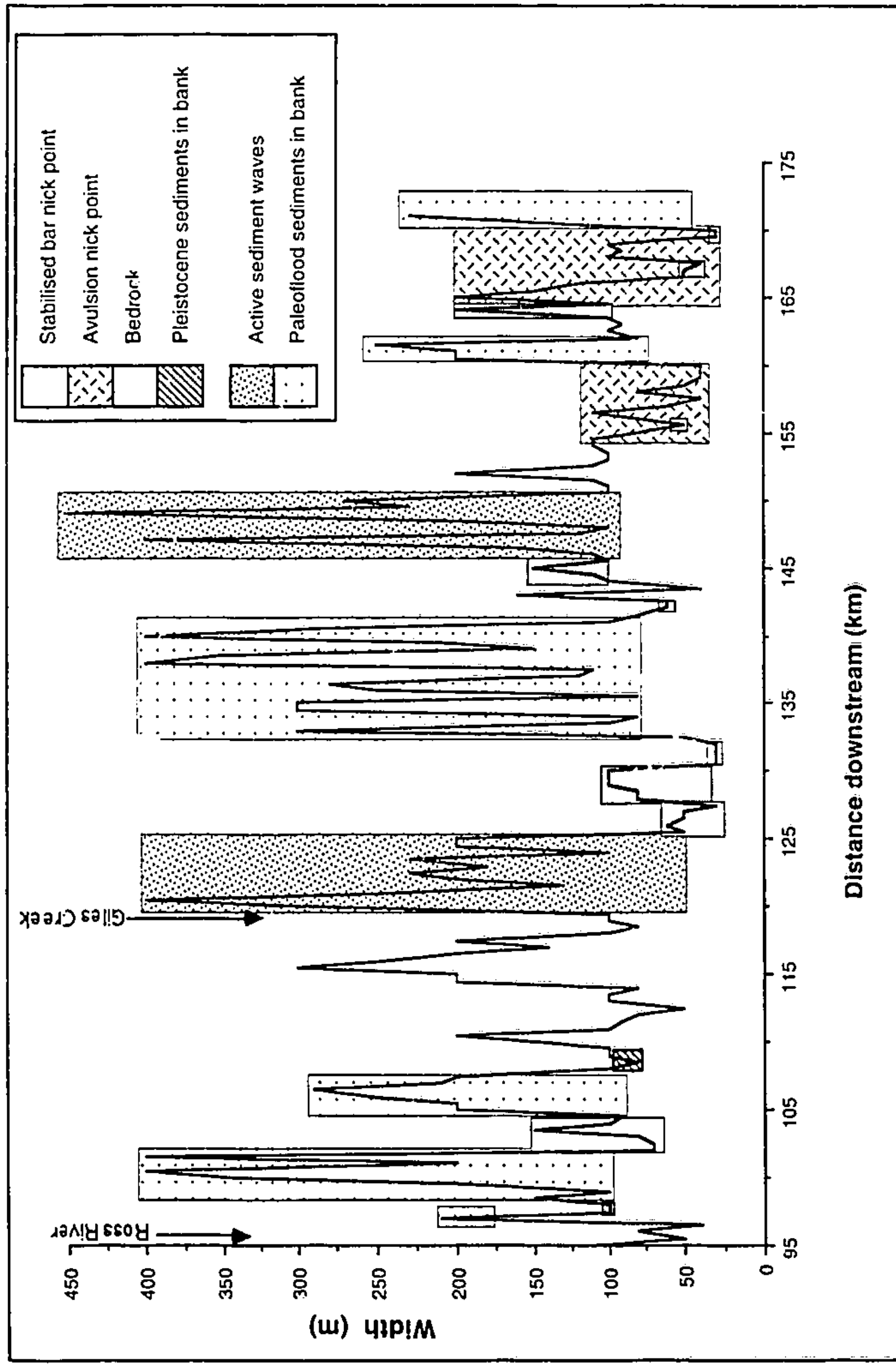


Figure 8.11. Factors influencing channel width variability in the Todd River.

for channel avulsion leads to long residence time for paleoflood sediments. Where inset by the modern channel the paleoflood stores are reworked principally by channel entrenchment and deflection. At the three Giles Creek sites the modern channel has reworked 8%, 30% and 5% respectively of the 3,000 year old Giles Creek paleoflood fill (Giles J), indicating a long residence time.

In the Todd River, reaches that are alternately constrained and unconstrained by channel banks may operate as series of reservoirs of intermittently stored sediment. In a study of the Kowai River in New Zealand, Beschta (1983a, b) considered that the ability of each reservoir to store sediment depends on the primary dimensions and morphology of potential storage locations, the amount of sediment in storage, the rate of sediment influx, the stream flow regime and the characteristics of the constricted channel/valley.

Summary

The southerly incursion of monsoonal depressions has generated large floods in the MacDonnell Ranges. These floods inundated vast areas of the landscape eroding large tracts of the dune fields and spreading floodwaters and sediments across the catchment. In addition to the immediate geomorphic effects the channel and floodplain system has continued to adjust to pulses of floods and sediment supply (Fig. 8.1). Large dunes aggraded adjacent to some paleoflood channels, smaller drainages have adjusted their planforms to the more permeable paleoflood sediments and preexisting channel networks. Vegetation patterns have been altered as relatively fertile seed-rich alluvium was emplaced in the desert margins. Once adjustments are initiated the system complexity ensures that they will be propagated throughout the drainage network, but because fluvial processes are intermittent in arid regions the responses are also discontinuous and the system may adjust only partially (Graf, 1988a).

The modern channel continues to adjust to the most recent pulse in sediment supply as it transports the paleoflood sediment in waves down the catchment forming channel beads and entrenchment, mobilising the sediment stores and enhancing the width variability. Channel planform is locally controlled by the nature of the paleoflood boundary sediments which are resistant to erosion or easily eroded. As the recovery time from paleofloods most probably exceeds their recurrence interval the Todd River remains in disequilibrium.

Given the sum of the first order geomorphic effects in combination with the second order effects and adjustments, it is proposed that it is catastrophic events that have dictated the late Quaternary geomorphology in this arid environment.

8.6. *Effects of the paleofloods on the central Australian Aboriginal occupants and implications for the archaeological record*

In addition to the geomorphic effects it is worth noting the impact of these events on the Aboriginal people of the central desert region and the implications for the archaeological record. The high magnitude floods increased the resource area in the Todd catchment by transforming the desert margin composed of longitudinal dunes to alluvial plains dotted with small ephemeral lakes. Two effects will be discussed here: the formation of new ephemeral surface water bodies and the extension of resource paths into previously relatively uninhabitable terrain.

The floods avulsed and changed the path of the flow during events thereby spreading flood water and sediment over a new area during each flood. This coupled with the distance downstream that these floods travelled extended resources relatively far into the harsh dune system of the Simpson Desert. This immediately brought ephemeral water resources into remote areas and deposited fertile soils which has had long-term beneficial effects.

As the flood channels carved fresh paths through dune fields, they emplaced tongues of coarse gravelly sand over the eroded aeolian landscape. Marginal to these flood channels, the swales between longitudinal dunes were back filled with flood waters and numerous ephemeral small lakes were formed in topographic lows (Fig 6.20). Today these lake beds are still apparent as small (<300 m^2) dry clay pans. The stratigraphy of one of these paleolakes has been described in Chapter 6 (section 6.6.3.). There is evidence to suggest that these resources were utilised by Aborigines, as stone artefacts are located on the crest of a flood-truncated dune adjacent to one of these small lakes.

Therefore, the paleofloods constructed a series of surface water bodies which were repeatedly occupied by flood waters and provided a valuable but short-

lived utilised resource. These flows transported river gravel, sand and mud into the swales of the longitudinal dune systems, depositing nutrient- and seed-rich alluvium in a relatively nutrient-poor environment. Soils in arid Australia are highly weathered and poor in nutrients with less than half the mean levels of phosphorus and nitrogen observed in other similarly arid regions (Stafford-Smith and Morton, 1990). Within Australia there is a marked differentiation in soil nutrients between sub-environments such as aeolian dune fields and alluvial soils (Table 8.2). This has resource significance as soil nutrients which are more abundant in the upper 10 cm of soil (Charley and Cowling, 1968; West 1981) are susceptible to wind and water transport (Noy-Meir, 1981), so that alluvial deposits and nutrients tend to move together (Pickup, 1985). This results in abrupt ecological boundaries (Stafford-Smith and Morton, 1990) which match the different geomorphic environments.

Environment (Vegetation)	Nitrogen 0 - 1 cm	Phosphorus 0 - 1 cm	Nitrogen 5 - 10 cm	Phosphorus 5 - 10 cm
Sand dunes (spinifex grasslands)	15.0	9.5	6.8	2.8
River floodout (mixed open woodland)	67.4	44.4	13.4	18.8

Table 8.2. Nitrogen and phosporus levels in Australian arid soils. From Table 2, Stafford-Smith and Morton (1990).

The relationship between vegetation zonation and areas of inundation is well established (Hughes, 1994). Riparian and flood plain vegetation are influenced by the magnitude, frequency and duration of stream flows. Hupp and Osterkamp (1996) state two reasons for a given species growing on a particular landform. First,that the site is suitable for germination and establishment and secondly, that the ambient environmental conditions are suitable for growth until reproductive age. In arid and semi-arid fluvial systems which are at the severe end of the stress-equilibrium gradient, vegetation distribution may be driven largely by the tolerance of species to specific geomorphic processes (Hupp and Osterkamp, 1996).

Therefore the injection of tongues of relatively nutrient-rich alluvial soils into aeolian dune fields has had a major ecological effect by increasing the local biodiversity. A singular important effect of large floods is the changing vegetation patterns which emerge and are sustained over long time periods, perhaps for several hundred years or more owing to the nutrient-rich alluvium.

These event-based changes in plant diversity result in the emergence of productive pockets of food resources. Stafford-Smith and Morton (1990) suggest that the dearth of nutrients has cascading effects throughout the food-webs. They predict that there is an increasing diversity of birds and leaf-eating insects in a transect from spinifex grasslands to riverine flood plains. In addition, as very few desert animals are dependent on free water, these pockets would have sustained this diversity after the replenished and new surface water bodies had disappeared. The longer term impact of these pulsed changes will now be discussed.

8.6.1. The Duration of Enhanced Resources

By examining the vegetation introduced into the dune fields along the older flood paths we can estimate the survival time of colonising species such as the Coolabah (*Eucalyptus microtheca)*. Flood tracks older than 4 ka have undergone minor aeolian reworking which is reflected in the sparse, low, shrubby vegetation. These reworked paleochannels are located tens of kilometres from the modern channel and have not been subjected to inundation for several thousand years. Plant life in these locations is supported by periodic rainfalls. In other locations there is evidence to suggest that even on abandoned flood channels, species survival times are long. Evidence from recent floods indicates that Coolabah trees colonise distributary flood paths throughout the catchment. This species is commonly found on heavy soils and floodouts and on seasonally flooded areas around swamps and lagoons, and less commonly along watercourses. They are used as food, the bark is used medicinally and water is obtained from the roots. In addition the hollow trunks are favoured sites for native bee colonies and birds' nests (Latz, 1995; Urban, 1993). Once established, the fire and drought tolerant Coolabah is almost indestructible in its natural habitat (Latz, 1995). Their presence today along the No. 5 Bore paleoflood channel suggests that Coolabah communities have survived since at least the last major inundation which, on the basis of geomorphic and chronological, data occurred approximately 1,000 years BP. The paleochannel vegetation contrasts markedly with the vegetation growing on the surrounding Pleistocene aeolian dunes. As it is unlikely that the trees are 1000 years old, it is

suggested that Coolabah community has been maintained by the seed stock, relatively nutrient-rich soils, periodic rains, local runoff and the ability of perennials to tap rapidly into ground water resources. The soil moisture extremes are crucial in defining the diversity of plant life histories on different soils. While germination and establishment are maintained by periods of high soil moisture, persistence is controlled by adaptation to moisture deficiency (Stafford-Smith and Morton, 1990). River channels in particular have a relatively continuous supply of subterranean water which is tapped by perennial plants. These plants are not dependent on short-term rainfall for their moisture and sustain growth for long periods after ground water recharge; they are important in stabilising production for some types of consumers and, despite the variable climate, many animals and plants do not vary in a linear way with rainfall (Stafford-Smith and Morton, 1990).

At present, there is no evidence of sapling growth around the base of these Coolabah trees along the No. 5 paleochannel, probably owing to grazing. This is unlike other species in the area such as Ironwood (*Acacia estrophiolata)* and Gidgee (*Acacia georginae*), both of which are unpalatable to browsing cattle and responded to the rains of the 1980's. It is assumed, therefore, that unless land use changes, this stand of Coolabah will stop propagating. It is proposed that in favourable conditions, without introduced domesticates, the Coolabah stands may survive naturally for >1000 years.

8.6.2. Implications for the archaeological record

Others have commented on the relationship between geomorphic processes and preservation of archaeological sites. Head (1983) described the effect of sea level stabilisation in the Discovery Bay region, Victoria, up to 3,000 BP as initiating a period of major sediment movement and shell midden erosion. Ross (1989) noted that the nature of surface camps, *i.e.* having little or no stratigraphic context, made them sensitive to destruction by changes in sedimentological regimes.

The preservation of sites for central Australian archaeology is affected by the destructive flood processes. During the very arid period in central Australia 15,000-25,000 years ago, people in the central desert regions retreated to the range country where springs and permanent water holes and associated resources would have remained accessible (Smith, 1989). Many of these archaeological sites are in gorge reaches, and are subjected to erosion during large floods. The large floods were equally destructive beyond the confined gorge reaches. They carved fresh paths through dune fields and formed distributary channels across existing flood deposits. During these events evidence of all sites that existed on the fluvial plains in the path of the flood would have been destroyed. The sequence of large floods which have affected the Todd catchment would have systematically destroyed evidence of occupation that may have accumulated prior to and between floods during the late Pleistocene and Holocene

Although the destructive effects of smaller floods rapidly diminish with distance from the channel (Bourke and Pickup, *in press*), there are implications for occupation sites located adjacent to confined channels. Bourke (1994) has modelled the effects of medium magnitude floods on flood plain processes in the Todd catchment and found them to accrete and erode rapidly. There are two implications for the preservation of occupation sites. Firstly, young sites along the riverine corridors which are subject to this rapid accreting process are now likely to be buried under several meters of sediment, as was found at the Mosquito Bore site (Bourke, 1997). Secondly, the length of preservation of these sites is a function of the spacing between large floods, as a relatively high magnitude event may completely remove the aggrading sediment down to the resistant underlying sediment, generally bedrock or cemented Pleistocene sediments (Bourke, 1994), removing evidence of both the buried and surface occupation sites.

Ross *et al* (1992) proposed that invisible occupation may account for the dearth of postglacial sites; I agree that the absence of evidence of occupation is not evidence for the absence of people and suggest that an additional hypothesis is the destructive nature of natural climatic events which are known to have occurred during the late Pleistocene and Holocene in central Australia on the Todd (Bourke, *in press* a), Finke (Pickup *et al.*, 1988), Dashwent, Derwent (Pickup, *pers. comm.*) and Sandover (Tooth, *in press*) River systems.

8.6.3. Summary

The occurrence of high magnitude rainfalls and floods during the late Pleistocene and Holocene increased the distribution of water, plant and animal foods. These events extended resource pathways into the dune fields of the northern Simpson Desert. The survival time of many of the enhanced natural resources was close to, if not longer than, the recurrence interval of the floods and therefore presented quasi-dependable resources. The flood

record for central Australia illustrates that large floods were more frequent than has previously been recognised and would have assisted the survival of central Australian occupants. Certainly, the effect was to dampen the variability of resource availability.

The destructive nature of floods which rework large tracts of the channel and flood plain has implications for the preservation of Aboriginal riverine occupation sites. The understanding of the broad scale implications of climatic variability and occurrence of floods may have implications in other areas of Australia. Indeed, the occurrence of large rainfall events in central Australia, sourced from North Australia, would undoubtedly have had sister events in the Western Australian region.

CHAPTER 9

Conclusion

9.1. Introduction

This thesis investigates the fluvial geomorphology of the arid Todd River in central Australia and contributes to the discipline by describing the structure and behaviour of arid zone flood plains, which is a neglected area of fluvial geomorphology. It is also one of few studies to describe the geomorphic effects of large scale paleofloods beyond confined gorge reaches.

9.2. Todd River Arid Zone Flood Plains

Aim 1. To describe the morphology, morphostratigraphy, sedimentary characteristics and morphodynamics of the Todd River in arid central Australia.

Chapter 4 describes two types of flood plains recognised in the Todd River system: confined and relatively unconfined. Confined flood plains occur where lateral movement of the channel is impeded by resistant channel boundaries, so that a single thread, generally straight channel is flanked by a narrow stepped flood plain which may have well developed benches within the channel. Relatively unconfined flood plains are associated with channels which are laterally unstable and often occur downstream of confluences and at areas of increased channel width; multiple channels are separated by high, well vegetated, stable islands and braid bars. Both confined and relatively unconfined flood plains are stepped and the surface reworked by back channels, flood channels and swirl pits.

Flood plain morphostratigraphic units include small and large scale channel fills, channel insets (bench and oblique) and flood plain insets, buried paleoflood deposits, flood plain remnants, veneers and swirl pit fills. Two morphostratigraphic models are [illegible]ed which summarise the main characteristics of the confined and [illegible]ned Todd River flood plains. The sedimentary characteristics of confined and relatively unconfined flood plain profiles are similar however and are summarised in a single profile model. Composed principally of vertically accreted layers emplaced during floods, boundaries between layers often are erosional as indicated by truncated bedding. Sediment textures vary from coarse gravel through to mud, with sand and gravel beds occurring in any order vertically through the profile. Flood couplets are fairly common and mud balls are preserved in sandy deposits. Bioturbation often disrupts layer boundaries, and laminations and horizontal bedding are prominent. Dipping parallel and trough cross bedding occur in fill deposits and veneers, although structureless sediments also are common.

The age of flood plain deposits in both confined and unconfined reaches, was determined by radiocarbon analysis and ^{137}Cs ^{210}Pb concentrations, and is predominantly modern but ranges to 400 years. The mean residence time of sediment before remobilisation in the active flood plain thus is estimated as 100-300 years.

Morphodynamics are described in Chapter 5. Flood plain destruction is by channel widening, flood plain stripping, swirl pit scour and flood channel scour. Channel widening occurs during floods, as do lateral channel migration, removal of stable islands and bars, exploitation of prior surface channels and notching. Remnants of laterally eroded flood plains appear as steps on the flood plain after its partial reconstruction. Multiple flood plain remnants indicate repeated channel widening and/or flood plain stripping which operates from the micro (<20 cm) to the meso (<5 m) to the mega scale (valley bottom sequences). Alluvial morphostratigraphy shows that the valley bottom alluvial fill has been stripped and rebuilt repeatedly. Swirl pit scours from macroturbulent vortices contribute to flood plain erosion. Flood channels scour the flood plain to depths of 1 m but confined back channels are more effective agents of both lateral and vertical erosion.

Mechanisms of flood plain construction include the deposition of insets, flood plain veneer, channel and swirl pit fill, and bar and island formation. Channel insets are remnants of lateral bars whereas flood plain insets are deposited on partly or completely stripped flood plain and terraces. A flood plain may be composed of several flood plain and channel insets and veneer deposits. Surface channel and swirl pit fills often are blanketed by more extensive sand sheets. The filling of abandoned channels is important in the construction of relatively unconfined flood plains.

Two models of flood plain formation are proposed in Chapter 5. Both involve catastrophic erosion, followed by the formation of new channels and flood plain reconstruction, but channel switching, and expansion and wider-reaching fill processes characterise unconfined reaches. The complex

morphostratigraphy of flood plains is related to the timing, magnitude and downstream changes by transmission losses and tributary inflows.

9.3. Todd River Paleoflood Geomorphology

Aim 2: To describe the morphology, morphostratigraphy, sedimentary characteristics and morphodynamics of the Todd River paleofloods and their deposits.

The central Australian landscape provides one of only a few examples so far described of large-scale assemblages of fluvial landforms of the arid zone resulting from extreme rainfall events. Depositional landforms include large aprons associated with braiding or anastomosing channels which extend from the ranges to the northern fringe of the Simpson Desert in relatively straight paths. Flow was diverted around outlying bedrock ridges and older higher Quaternary deposits. Smaller scale splay channels (~6~12 km long) are superimposed by longitudinal bars. Unconfined ~4 km wide braided aprons include expansion bars downstream of confined passages, longitudinal bars and transverse bar fields. Marginal channels occur at paleoflood complex boundaries. Well sorted fine sand slack water deposits are located in caves and within the mouths of small tributaries; silty backwater deposits occur in dune swales adjacent to floodout channels.

Eroded remnants of landforms adjacent to paleoflood channels include truncated coarse textured colluvial fans and older flood channels and longitudinal dunes and dune fields. Dune swales in the northern Simpson Desert were inundated by ~5 m deep floodwaters which flowed as high as 1 m below the dune crest. However, remnant Pleistocene 'islands' standing above the paleoflood plains, generally were not inundated by later paleofloods and survive downstream of bedrock ridges and on topographic highs throughout the catchment.

Paleoflood channel avulsion has had three effects; firstly, the paleoflood plains of Todd carry a mosaic of abandoned distributary channels; secondly the floods inundated a larger area than if the flow paths had remained fixed; and thirdly, channel entrenchment has been widespread. Avulsion occurred at two scales; reach-scale and large scale, and the latter resulted in three different locations of the terminal floodout during the Late Quaternary.

The sources of paleoflood sediment were slope deposits in the headwaters and Pleistocene alluvial, aeolian and paleoflood deposits. Paleoflood sediment texture is coarse gravel and cobble in the headwaters and parts of the piedmont zone and medium and small gravel in a coarse to fine sand matrix further downstream. However comparatively little sediment reached the Simpson Desert.

9.4. Central Australian Flood Record

Aim 3: To determine the age of the Todd River paleofloods in order to infer patterns of Late Pleistocene and Holocene extreme rainfall variability from the Todd River paleoflood record.

Fourteen OSI and thirteen radiocarbon age estimates were used to date the paleofloods. The paleoflood record for the Todd River now extends back to 26800±3000 BP although no events between 24,000 and 15,000 BP were recognised. Extreme floods have occurred repeatedly since 15000 BP with little clear evidence at present to suggest a clustering of events.

The highest magnitude flows occurred between 14,000 and 4,000 BP and paleofloods since then have been smaller. Some paleofloods may be coeval with events in northern Australia but dating is not sufficiently precise to prove this.

9.5. Catchment Recovery

Aim 4: To describe the geomorphic responses and adjustments of an event-driven arid zone river to impacts of high magnitude floods.

Where small-scale drainage systems intersect with the paleoflood alluvium the drainage pattern may disintegrate or it may be diverted. The surface of all paleoflood deposits have been reworked to some degree by aeolian processes and most impressive are the source bordering, climbing dunes nourished by paleoflood sediments.

Channel width is highly variable in the Todd catchment. Factors which have influenced the narrow channel reaches include resistant boundaries. Factors which have influenced channel widening are the presence of erodable channel banks (generally paleoflood sediments) and the temporary storage of sediment waves. Reaches where there is a sudden and large increase in channel width (~200 to 450 m), are termed channel beads and an

increase in the frequency of beads coincides with the input of high sediment loads from the Ross River and Giles Creek tributaries which rework paleoflood sediments. Channel beads form by lateral erosion during extreme floods and during lower magnitude flows where channel bed aggradation causes channel deflection. The location of reaches of channel aggradation is controlled by the loss of energy to transport sediment downstream, the ability of constrictions to promote aggradation upstream and sediment supply from tributaries.

Large-scale Late Pleistocene and Holocene floods have dominated the fluvial processes of the Todd catchment since the Late Pleistocene arid phase. These events transformed large sections of an aeolian landscape to a fluvial landscape. The adjustment and recovery of the Todd River is ongoing and as the recovery time most probably exceeds the recurrence interval of these high magnitude floods, the channel is in a constant state of disequilibrium.

Given the sum of the first order geomorphic effects in combination with the second order effects and adjustments, it is proposed that it is the catastrophic event that has dictated the Late Pleistocene geomorphology in this arid environment.

9.6. The Hierarchy Of Landforms In Variable Discharge Regimes: A Multi-scale Approach To Arid Zone Fluvial Landforms.

Fluvial landforms assembled under variable flow regimes are best described using a multi-scale approach (Gupta, 1988; Pickup, 1991; Gupta, 1995). In these environments, landform assemblages are associated with discrete flow magnitudes. Examples include the compound channel systems described by Graf (1988b), the large scale boulder bar sequences in the Burdekin gorge, north east Australia (Wohl, 1992), and the fluvial forms of the Auranga River, India (Gupta, 1995). Under the present climatic regime, the Todd River is subject to flows at extremes of the magnitude/frequency spectrum, *i.e.*, there are periods with no flow in the channel and periods when the catchment is inundated by paleofloods. Resultant fluvial landforms evolve episodically rather than continuously and occur at relatively discrete spatial scales.

The fluvial landforms are subdivided for convenience into three categories based primarily on landform scale and inferred flow magnitude: landforms generated by within-channel flows (small-scale); flows which extend across flood plains (medium-scale); and paleoflood flows (large-scale). Form dimensions across this three-scale hierarchy of flows and landforms vary by an order of magnitude. For example, longitudinal bars measure 10 m, 100 m, and up to 2,000 m for the respective categories. The resultant landforms are identified as small, medium and large-scale fluvial landforms. It is the geomorphic impact of these different scale flows that results in the complex assemblage and variable scales of fluvial landforms in the Todd River catchment.

In central Australia, and other arid and semi-arid regions, the extreme event is the most geomorphically important event (Schick, 1974; Thornes, 1976; Wolman and Gerson, 1978; Patton and Baker, 1977; Harvey, 1984; Pickup, 1991; Patton *et al.*, 1993; Schick, 1995). The largest-scale fluvial landforms in the Todd River are formed by extreme flows. These events occur rarely and inundate vast tracts of the landscape including aeolian dune fields. These large scale landforms evolve episodically, remain inactive for long periods of time (Baker, 1977; Pickup, 1991) and may appear to be in disequilibrium with the modern channel.

The assemblage of small, medium and large scale fluvial forms in the Todd catchment displays a complex range of landforms. In this nested landform hierarchy, high magnitude events influence lower magnitude flows and forms. The persistent erosion and depositional patterns generated by paleofloods modulate the landscape response to subsequent and smaller events. This is especially the case where the active channel is flowing through or adjacent to paleoflood deposits. The higher paleoflood surfaces direct the path of medium scale flows and store a large volume of gravelly sand available for transportation by medium and small scale flows. This sediment supply affects channel pattern, planform and flood plain morphostratigraphy in the modern channel.

In addition to the geomorphological imprint of a variable discharge regime it is important to note that the scale of flows affects the frequency of reworking of alluvial deposits. The headwaters and piedmont reaches of the system are more frequently reworked by small and medium scaled flows than downstream reaches. This is related to the decrease in flow magnitudes downstream due to transmission losses. In this way the distance decay of flow magnitude in ephemeral streams influences the spatial and temporal variability of landforms in the catchment.

APPENDIX 1

Chronology

Field Sampling Strategies

The sedimentary paleoflood record is incomplete because high magnitude floods often remove the alluvial record of previous events. Nanson and Tooth (in press) found that the number of samples decreases exponentially with age within the flood plain deposits in confined reaches in river systems north of Alice Springs. However they also found that within unconfined distal reaches the tendency for the channels to change course during the Holocene resulted in the extensive preservation of older Holocene alluvium.

Of the ten studies portrayed in Figure 7.12, the records from eight do not extend past 4 ka. In paleoflood studies the incompleteness of records may lead to erroneous conclusions regarding the magnitude and frequency of floods over time. In particular there are problems associated with the limited preservation of flood deposits in confined gorge reaches where most of the paleoflood research has been undertaken. Erosion of prior sediments by floods makes it most unlikely that complete records can be obtained. The research in northern Australia at Waterfall Creek (Nott and Price, 1994) and in the Todd River in central Australia (this study) have produced the longest paleoflood records for Australian rivers and indeed globally.

The longer records were achieved by sampling in relatively unconfined reaches, being plunge pool levees near the base of waterfalls in the Northern Territory (Nott and Price, 1994) and in unconfined flood distributaries in the Todd River. These are locations where the sediment record has a higher probability of being preserved than in the gorge reaches, where slack water records in Australia so far have been proven to be more sparse than has been found elsewhere (Wohl *et al.*, 1994).

Hence, this study is based on samples from paleoflood deposits in downstream, unconfined sites which are spatially extensive and preserve the deposits of alternative channels followed by different floods. However morphostratigraphic continuity is sometimes difficult to demonstrate, as described in Chapter 6.

In contrast with many previous studies, two dating methods, optically stimulated luminescence (OSL) and radiocarbon, were utilised systematically.

Luminescence Dating

Luminescence dating is a radiogenic dating method which measures cumulative non isotopic effects of radioactive decay in electron energy traps and produces numerical ages (Colman *et al.*, 1987). Luminescence is an 'umbrella' term covering the thermoluminescence (TL) dating technique and the various methods of optical dating. When applied to sediments, these methods give the age of the last time that the mineral grains were exposed to sunlight. Optical dating was introduced by Huntley *et al.* (1985) as a means of dating unburnt sediment and sedimentary quartz is well suited for OSL dating (Spooner *et al.*, in review). The Todd River OSL data is presented in Tables A1.1 and 1.2)

All luminescence dating methods rely upon measuring the time-dependent accumulation of charged electrons and holes trapped at defects within mineral grains, such as quartz and feldspar. The trapped charges are rapidly reset by exposure to sunlight, and slowly regrow after burial, with exposure to ionising radiation emitted by the decay of the radioactive elements in soil and rock, plus a small contribution from cosmic rays. When the mineral grains are exposed to heat or light, the electrons are released from their traps, then subsequently recombine with trapped holes, producing photons. In the natural environment exposure to sunlight zeroes the luminescence signal which then begins to accumulate once the mineral grains are buried. Samples must be collected in the field without exposing them to light and processed in the laboratory under safe light conditions.

Luminescence dating involves measuring the trapped charge accumulated since the resetting event (the 'natural' luminescence) by detecting the light emitted by electron and hole pairs as they recombine. The absolute number of trapped electrons cannot be measured directly; instead the dose of radiation received by a sample is measured by exposing separate aliquots of the sample to laboratory-administered radiation doses either incrementally added to the natural luminescence (additive dose method), or by rebuilding the luminescence after the natural luminescence has been measured and the sample reset to zero (regeneration method) (Aitken, 1985; Prescott *et al.*, 1993). These methods provide a measurement of the total radiation dose absorbed during burial, termed the 'paleodose' (P). Age calculation requires

the measurement of P, and also the measurement of the environmental radiation dose received by the sample which is determined by field and laboratory assays of radioisotope concentrations (^{238}U, ^{232}Th, ^{40}K, ^{87}Rb). The water content affects the effective environmental dose rate and must be measured; the depth of burial affects the dose from cosmic rays (Aitken, 1983).

In choosing a luminescence dating technique, OSL is more suited to the dating of fluvial sediments than TL. It was discovered that the 325°C TL peak in quartz is very readily bleached by both visible and ultra-violet wavelengths of light and corresponds to the signal used in conventional 'green light' OSL dating (Spooner *et al.*, 1988; Spooner 1994). Thus reliable optical dates can be calculated for sediments that were exposed only briefly to sunlight prior to burial in antiquity; such conditions prevail during flood flows. On the other hand, TL dating relies on a major peak at 375°C which is much less sensitive to light, and is less reliable for dating water-lain sediments. Furthermore, from a technical standpoint, the signal-to-noise ratio is more favourable for OSL than TL and optical dates can be obtained with more precision, and for younger sediments, than is possible using TL.

TL was applied with only limited success to the dating of young paleoflood slack water sediments in northern Australia (Murray *et al.*, 1992) and central Australia (G. Pickup, *pers. comm.*) where age errors were as much as 24-46% owing to uncertainty about the degree of initial resetting. Similar uncertainties were reported by Gillieson *et al.*, (1991) in dating slack water sediments in Windjana Gorge in the Kimberley, and Patton *et al.*, (1993) in attempting to date recent flood plain sediments in the Ross River found insufficient accumulated TL dose to allow for differentiation from the modern surface sediments. These limitations affect OSL to a much lesser degree than TL. Thus OSL is more suitable for sediments transported and deposited in turbid floodwater under low ultraviolet conditions.

Luminescence Methods

- Field sampling: The OSL samples were taken in stainless steel tubes of 50 mm diameter by 160 mm long, from vertical faces of exposed pits. In each case there was at least 30 cm separation of the sample from the sedimentary layer boundary, or any isolated stones except for sample 42, which was augured. The OSL samples were sealed in light proof black plastic, labelled and stored in light proof bins. Samples for dosimetry (1 kg) and moisture samples (200 gm) were taken adjacent to the core. Modern analogue samples (2 tubes of 25 mm diameter, 200 mm long) were taken from a recently active fluvial deposit located close by at 5 cm depth.

- Laboratory methods: The field samples were opened in safe light conditions and 20 mm at each end of the sample cylinder was removed. The sample was weighed and oven dried at 70°C for 24 hours and the moisture content was calculated.

  Chemical preparation and loading of sample aliquot on stainless steel discs was done under the supervision of Dr. N. Spooner and guidance of Mr. N. Hill. All steps in the sample pretreatment and luminescence dating were carried out in the dark with red safe light illumination.

The following sample pretreatment procedure was applied to the 14 OSL samples and the 7 modern analogue samples:

- Remove carbonates with 10-20% HCl, soak for 24-48 hours.
- Remove acid with 5 distilled water washings in an ultrasonic bath.
- Wash in alcohol and acetone twice in the ultra-sonic bath.
- Disaggregate clays with NaOH in the ultra-sonic bath.
- Remove alkaline with 5 distilled water washings in the ultra-sonic bath.
- Wash in alcohol and acetone twice in the ultra-sonic bath.
- Leave to dry (24-48 hours).
- Sieve to isolate the 90-125μm and 180-212μm fraction. Keep all fractions.

The following procedures were undertaken by Mr. N. Hill:

- Heavy liquid separation (sodium poly-tungstate) to remove heavy minerals.
- HF etching to remove a thin surface layer affected by external alpha particles.
- Wash in HCl to remove fluorides.
- Wash in alcohol and acetone.

The clean quartz grains of 90 - 125 microns from each sample are divided into 64 aliquots and loaded onto [illegible] mm diameter disks which have been sprayed with adhesive and placed in the tray for loading into the irradiator.

- The OSL methods used in the ANU luminescence dating laboratory are described by Spooner *et al.*, (in review) as follows: The OSL is detected when blue/green light is shone onto the quartz. An Elsec Type 9010 automated reader provides visible light stimulation for OSL measurements using a filtered halogen lamp light source (500±40 nm), with detection of UV emissions from quartz by an EMI 9235QA photomultiplier shielded with Hoya U-340 and Schott UG 11 colour glass filters. Laboratory irradiations were administered from an Elsec type 9022 irradiator with a 100 mCi ceramic $^{90}Sr/^{90}Y$ (beta) and ^{241}Am (alpha) source. Preheats of 300 s at 220°C were given prior to OSL measurement.
- OSL growth curves were produced by Dr. N. Spooner using the multiple disc additive dose method. The data were fitted using single saturating exponentials for young samples and combined with data collected by the regeneration method for older samples using the 'Australian slide' method introduced by Prescott *et al.* (1993).
- Environmental radioisotope concentrations were assessed by laboratory assays at CSIRO Division of Water Resources, Canberra, using Ge-gamma spectrometry. The samples were prepared and cast under the supervision of Ms. J. Olley and doses measured by Dr. P. Wallbrink.
- The radioisotope data were combined with measured soil water content and calculated cosmic ray dose by Ms. D. Questiaux in order to calculate the environmental dose rate.

Radiocarbon Dating

Radiocarbon is an isotopic dating method which measures changes in isotopic composition due to radioactive decay and produces numerical ages (Colman *et al.*, 1987). All ^{14}C ages were obtained as conventional radiocarbon ages using the 5568±30 year half-life (Libby *et al.*, 1949; Stuiver and Polach, 1977). The dating of the radiocarbon samples was carried out by laboratory staff in the Quaternary Dating Research Ce[illegible] (QDRC), ANU.

Methods

All samples were wood, charcoal, or fine organic particles except for one shell sample. Organic samples were pretreated in the QDRC using standard methods. Inorganic carbonate was removed using hot HCl, and soluble humic acids were leached from the samples using hot NaOH. Activity of ^{14}C was determined for seventeen samples using liquid scintillation counters. δ^{13} for age normalisation was measured mass spectrometrically in the QDRC.

Calibration of conventional radiocarbon ages greater than 300 yr b.p. was done using the CALIB 3.0 program (Stuiver and Reimer, 1993). Radiocarbon ages between 0 and 300 years are not commonly calibrated because of industrial injection of CO_2 into the atmosphere which is not uniform globally. Ages younger than AD 1950 are represented by anomalously high ^{14}C activity owing to above-ground nuclear testing but are not resolved other than being stated as >modern. In this thesis conventional radiocarbon age errors are presented at the 2 sigma level.

Problematic Samples:

There are three factors which may result in an erroneous radiocarbon age. The first is that the movement of small charcoal pieces downwards through the sedimentary deposit may result in a sample which is younger than the sediment and downwards transport by ants and termites is highly likely in central Australia. Secondly, samples may be contaminated by younger carbon from rootlets or humic acids (Williams *et al.*, 1993). Thirdly, the sample may be reworked from an older deposit which is particularly likely in fluvial environments (Blong and Gillespie, 1978).

^{137}Cs and ^{210} Pb Dating

Caesium-137 (^{137}Cs) and lead-210 (^{210}Pb) have half-lives of 30 and 22 years respectively (Pennington *et al.*, 1973, 1976) and are useful for dating recent sediments where they tend to be adsorbed onto clay particles (Wise, 1980).

^{137}Cs is an artificially generated isotope distributed by atmospheric thermonuclear weapon tests since 1954 (Wise, 1980); ^{210}Pb is a natural radioisotope within the ^{238}U decay series (Krishnaswamy *et al.*, 1971), and is distributed through diffusion of ^{222}Rn gas, a predecessor in the ^{238}U decay series. Both ^{210}Pb and ^{137}Cs are washed out of the atmosphere by precipitation (Koide *et al.*, 1973) and are adsorbed by clays and organic colloids within the soil where they are found largely within the top 30 cm of the profile (Davis, 1963; McHenry *et al.*, 1973).

Silt and clay were separated with a 63 micron sieve. The fine fraction (<63 microns) was sieved from nine flood plain samples in order to measure levels of ^{210}Pb, ^{137}Cs and R; the latter indicates the presence of ^{210}Pb derived not from ^{222}Rn but ^{238}U within the sample. The samples were processed in CSIRO under the supervision of Dr. Wallbrink. The presence of significant

concentrations of radionuclides that would indicate post 1950 AD ages was ascertained if radiometric concentrate was three times greater than the instrumental background (Wallbrink, *pers. comm.*)

The Paleoflood Dating Program

Both OSL ages and calibrated radiocarbon ages are combined in the chronology data set (Figure 7.9). Although the ^{14}C ages appear more precise they may be less accurate as the charcoal may be reworked from older deposits (Blong and Gillespie, 1978). However, in central Australia small pieces of flood-transported wood probably date from within a few years before the flood because termites, which are pervasive in central Australia, rapidly consume detrital wood (Pickup, *pers. comm.*).

Three radiocarbon ages from paleoflood deposits were discarded in the final analysis. They are samples ANU 10059, 8834 and 9649.

- ANU 10059 (673-248 Cal BP) conflicts with the OSL date (12900±1800, $ANU_{OD}161$) from the same sedimentary layer and is considered to be false. The soil profile of the deposit is red, rich in soil carbonate and is similar to other profiles with assigned Late Pleistocene ages.
- ANU 8834 returned a modern age (339-0 cal BP) which conflicts with two other samples from the same stratigraphic layer; 1196-691 cal BP (ANU 8833) and 1030±420 BP ($ANU_{OD}170$) and is considered to be false.
- ANU 9649 (3712-3142 cal BP) conflicts with an overlying sample of 9883-8429 cal BP (ANU 9650) and an OSL sample from the same alluvial unit (9780±1510 BP, $ANU_{OD}163$) 600 m away and is considered to be false.

Testing the Dating Methods

Where possible, both OSL and radiocarbon samples were taken from the same stratigraphic layer. However charcoal, wood and organic material were rarely found in the older deposit, and in many instances the radiocarbon laboratory was unable to date the radiocarbon sample taken adjacent to the OSL sample because the quantity of carbon was too small. Hence, paired OSL and radiocarbon ages were obtained from only two sites.

- Paired samples were obtained from a transverse bar on the Heavitree Piedmont fan (Fig. 6.3). The OSL sample returned an age of 1030±420 BP ($ANU_{OD}170$) and the radiocarbon sample returned an age of 1196-691 cal BP (ANU 8833): thus the results are in good agreement.

- Paired samples from the Todd B paleoflood complex (Fig. 6.47), an OSL and a radiocarbon sample, collected from the same stratigraphic unit 600 m apart, returned similar ages: 9883-8429 cal BP (ANU 9650) and 9780±1510 (ANU_{OD}163).

<u>Testing the Age Estimates</u>

In order to test the individual age estimates multiple samples from a single profile at four sites were dated.

- At the site on the Ross A paleoflood complex (Figs. 6.6, 6.25) the four radiocarbon age estimates were in the correct age sequence.
- At the site on the Todd B flood distributary, charcoal in a hearth provided a radiocarbon age of 463-340 cal BP (ANU 10065) which statistically overlaps the age of an overlying flood plain sand sheet with an age of 511-312 cal BP (ANU 10064).
- The ages of the 4 radiocarbon samples in the Mosquito Bore cave SWD (Fig. 6.40) are in stratigraphic order.
- The two OSL samples taken from a stratigraphic profile in the Giles B and A paleoflood complex (Fig. 6.34) returned stratigraphically consistent ages *i.e.*, 9700±1500 BP (ANU_{OD}160a) overlies 26800±3000 BP (ANU_{OD}160b).

- Two radiocarbon samples taken from paleochannel deposits at different locations downstream along No. 5 paleoflood complex (Fig. 6. 14, 6.15) returned similar ages of 108[illegible] cal BP (ANU 9655) and 1405-1135 cal BP (ANU 10058).

Sample	ANU_{OD}	Figure	Latitude	Longitude	Altitude (m)	Present Depth (m)
Ex 4.11	160a	6.29	23°50'40"	134°44'30"	390	0.30
Ex 5.1	161	6.27	23°48'30"	134°37'	410	2.00
MBD7	162	6.41	24°01'	134°42'56"	360	0.30
CF (ii)	163	6.47b	24°01'40"	134°45'10"	350	0.37
EsCh 1.1	164	6.12	24°03'20"	134°43'55"	350	0.35
CLB (1)	165	6.47b	24°02'	134°45'	350	0.30
AB 1.1	166	6.13	24°08'40"	134°50'30'	330	0.30
E3	167	6.16	24°20'	135°02'	285	0.25
Ex 1.1	168	6.45	24°19'35"	134°47'45"	316	0.80
Dpc 1.3	169	6.54	23°11'30"	135°00'30"	300	0.80
B3	170	6.3	23°50'	133°53'	535	0.20
42	171	6.21	23°41'55"	134°29'35"	460	2.00
Rg 1.2	172	6.16	24°25'	135°09'	265	0.40
Ex 4.13	160b	[illegible]	23°50'40"	134°44'30"	390	1.10

Table A1.1. OSL sample data

Sample Name	Palaeodose (P) (Gy)	Error (±) (Gy)	Modern Analogue Paleodose (MAP)	Error (±)	Dose Rate (d) (Gy/ka)	Age BP (P-MAP/d)	Error (±)
Heavitree/Undoolya							
B3	4.7	1.65	200	540	4.38	1,030	420
Ross River							
42	41.4	2.3	5,500	1,700	4.30	8,350	950
Ex 5.1	57.6	3.5	5,500	1,700	4.05	12,900	1,800
Giles Creek Piedmont							
Ex 4.11	31.8	1.4	820	-300 +1,560	3.21	9,700	1,500
Ex 4.13	77.2	4.4	820	930	2.85	26,800	3,000
Southern Plains							
ES Ch 1.1	20.1	2.5	2,260	850	3.45	5,160	1,340
C LB (1)	10.8	1.87	2,260	850	3.62	2,370	900
Mosquito Bore Distributary							
MBD7	25.1	1.4	2,390	2,530	3.65	6,230	1,370
EX 1.1	28.4	2.9	100	200	2.85	9,920	1,840
Eastern Systems							
CF (ii)	34.5	1.75	2,260	850	3.29	9,780	1,510
AB 1.1	10.0	1.3	2,260	850	3.67	2,110	760
DPc 1.3.1	25.8	5.4	2,260	850	3.46	6,800	1,850
E3	15.0	2.5	2 260	850	3.26	3,900	1,270
Rg 1.2	46.0	3.3	2,260	850	3.55	12,310	1,400
Modern Analogues							
46			5,500	1,700			
95(1)			2,260	800			
95(2)			200	540			
CFMA			2,260	850			
GCRMA			820	930			
MBDMA			2,390	2,530			
Ex 1.1MA			100	200			

Table A1.2. OSL age estimates

Sample Name	ANU ID	Figure	Material Dated	^{14}C Age BP	95% probability distribution window	Relative area under 95% window
Heavitree/Undoolya, Paleoflood Complex						
Bar 3 9-13	8833	6.3	Charcoal	1040±140	1196-691	.97
Bar 3 9-30	8834	6.3	Charcoal	125±174	339-0	.87
A(ii) @ 86	8835	6.4	Charcoal	1430±270	1874-886	.98
A2 75-88	8836	6.4	Charcoal	2830±70	3087-2775	.98
Ross River, Paleoflood Complex						
XLi	10060	6.25	Charcoal	480±60	639-598 564-430 381-323	.06 .86 .08
XM	10061	6.25	Charcoal	600±70	656-522	1
XNi	10062	6.25	Charcoal	970±70	985-724	1
XO	10063	6.25	Charcoal	990±60	987-740	1
XK	10059	6.6	Charcoal	450±160	673-248	.91
Giles Creek Piedmont, Paleoflood Complex						
XD+XB	10068	6.35	Charcoal	2710±380	3646-1933	1
Mosquito Bore Distributary, Paleoflood Complex						
C12	9285	6.40	Charcoal	320±60	493-286	1
CJ	9657	6.40	Charcoal	380±60	507-313	1
XI	10064	6.39	Charcoal	390±60	511-312	1
XJ	10065	6.39	Charcoal	210±140	463-340 340-0	.21 .79
CB	9650	6.47b	Shell	8290±340	9883-8429	1
CA	9649	6.47b	Charcoal	3220±120	3712-3142	.99
Eastern Systems, Paleoflood Complex						
CH	9655	6.14	Charcoal	1020±90	1083-726	1
XP+XQ	10058	6.55	Charcoal	1380±70	1405-1135	1

Table A1.3. Calibrated radiocarbon age of paleoflood samples.

APPENDIX 2

Site Descriptions

The following is a description of the confined and relatively unconfined flood plain sites described in Chapter 4. They are presented in order of distance downstream along the Todd, then the Ross River sites and finally Giles Creek sites.

Todd River

Todd/Ross Confluence

This site lies at the confluence of the Todd and Ross Rivers (Fig. 4.2). The flood plains are confined on the right bank by cemented Pleistocene alluvium and on the left bank by paleoflood sediments and a cemented Pleistocene terrace. The Ross River is the hydrologic, geomorphic and sedimentologic dominant system in this reach. Many of the flows from Alice Springs fail to reach this junction. Channel pattern is essentially straight with local braiding. The site is located on a large meander bend of the Ross River imposed by the ridge of fossiliferous siltstone and minor dolomite of the Arumba Sandstone formation. Channel gradient downstream of the confluence is .0032 m/m and the flood plain to channel ratio is 1.7:1.

Stud Bore 2

The channel boundary at this site is composed of loosely packed gravelly sand which facilitates the local lateral movement of the channel in addition to providing a large supply of sediment. This relatively unconfined site has a total channel width of 177 m and a flood plain to channel ratio of 3.4:1, the total flood plain is 607 m wide (Fig. 4.2).

Stud Bore 1

Located 3.6 km downstream of Stud Bore 2 (Fig. 4.2), the channel pattern in this confined flood plain reach is relatively straight. Channel width measures 70 m and the flood plain to channel ratio is 4.75:1. Gradient is steep measuring .0023 m/m. The total active width of the channel and flood plain surfaces is 400 m.

Mosquito Bore 1

The channel at this confined flood plain site flows through a gap in an outcrop of the Proterozoic Bitter Springs formation composed of dolomite, limestone, siltstone and sandstone (Fig. 4.4). The channel pattern is essentially straight with changes in channel direction dominated by the location of outcropping bedrock ridges. Channel width is 102.2 m with a flood plain to channel ratio of 2.07:1. The local channel gradient measured along a channel distance of 279 m is .00063 m/m.

Mosquito Bore 2

The right bank of this confined flood plain site abuts a high stage slackwater deposit overlying bedrock (Fig. 4.4). The left bank is confined by coarse textured paleoflood deposit. Channel pattern continues to be dominated by outcropping ridges. Channel width is 160 m and splits around a large longitudinal bar. Flood plain to channel ratio is 0.6:1.

Expansion Scour

This relatively unconfined flood plain site is inset into extensive sandy/gravel deposits of the Todd B paleoflood complex. It is located at the beginning of a reach where there is an increase in channel width from 50 m to a maximum of 450 m (Fig. 4.5). Channel width is narrow measuring 50 m and the flood plain to channel ratio measures 8.4:1. Channel gradient is low measuring .00087 m/m.

No. 5 Bore

The channel is confined in an eroded gap in the Proterozoic Limbla formation, which is composed of sandstone, calcarenite and siltstone. The channel pattern is predominantly straight (Fig. 4.6) with changes in channel direction dominated by the location of outcropping ridges. Channel width at this site is 84 m, and the limited flood plain development is indicated by a flood plain to channel ratio of .06:1.

Anabranching Site

This relatively unconfined flood plain site is an area of localised channel widening and the channel splits around large stable islands (Fig. 4.6). Total

active channel and flood plain width measures 1537.6 m. Channel width measures 153.6 m and flood plain to channel ratio is 9:1.

Site 'A'

The left bank confined flood plain is inset against colluvial deposits from the outcropping ridge of the Cambrian Giles Creek Dolomite (Fig. 4.6). Further downstream it flows between this formation and an outcropping ridge of Arumba Sandstone. It is a straight, narrow (62 m wide) section of channel located downstream from a wide braiding reach. The local channel gradient is .0016 and flood plain to channel ratio is 4:1.

The Simpson Desert

Once the Todd River passes through the Rodinga floodout and turns south towards the Simpson Desert, channelled flow ceases. The multiple channel morphology of the floodout is replaced by a broad flat featureless plain. As the system drains south it abuts against an outcrop of the Brewers conglomerate (Fig. 4.7) and channelled flow is reformed. Further downstream, the channel occupies the inter-dune corridors of the longitudinal north-west trending dunes of the northern Simpson Desert. It maintains a well defined channel then bifurcates and ponds in the terminal floodout of the dune swales. The spatial pattern of the longitudinal dunes controls the prevailing trellised channel pattern in this reach. Incipient flood plains form spatially discontinuous, bench-like insets composed of fine sandy material, lying unconformably on the underlying aeolian unit.

Atoota Dam 2

This confined flood plain site is a natural scour hole located downstream of a bedrock constriction in the Brewers conglomerate. This conglomerate provides a source of coarse gravels and cobbles for entrainment by the higher energy flows. Earlier this century a retaining earthen wall was built at the downstream end of the waterhole (Fig. 4.7) to improve water retention at this remote site for cattle. However the 1974 flood restored the channel to its former dimensions (*pers. comm.*, K. Pick).

Atoota Dam 1

In this confined flood plain reach (Fig. 4.7) the channel flows between Pleistocene longitudinal dunes. Channel width is 28 m, the flood plain to channel ratio is 9.4:1. Channel gradient along this reach measures .0005 m/m.

Desert Reach

Here the channel is confined in the 300 m wide swale between two longitudinal aeolian dunes (Fig. 4.7). Flood plain to channel width is approximately 1:1. The channel is actively incising the underlying cemented aeolian/alluvium sediment. There is little human impact in this reach as no palatable water sources exist.

The Ross River

In the headwaters, the Ross River and its tributaries flow through narrow and sometimes meandering gorges. Further south, folding of the Pre Cambrian quartzite, sandstone and accessory carbonate rocks has formed low, sharp crested, east-west trending ranges separated by narrow plains. Trellised rivers drain these parallel strike ridges and tributary gorges, passing through short gorges superimposed on the strike ridges and alluvial or sand plain reaches in the strike valleys. Two sites were examined in locations where the gorge width was less confined.

Ross River Striped site

The Ross River meander pattern is controlled by bedrock outcrops (Fig. 4.1). Gorge width at this site is 1.2 km. Channel gradient is steep (.0023 m/m) and total channel width measures 112 m and the flood plain to channel ratio is 4.3:1. Flood plains support a relatively dense stand of partially buried River Red Gums absent from the main channel.

Shannon Bore

Located 5.3 km downstream from the Striped Site (Fig. 4.1). Immediately upstream of the site, the channel passes through a narrow gorge reach and at the site locally widens at a small strike valley further downstream the channel is once again confined and eventually flows into the piedmont

reach. Channel width is 114 m and flood plain width is 260 m. The resultant flood plain to channel ratio is 2.3:1. The flood plain supports a stand of mature River Red Gums and the vegetation changes to Mulga on the higher terraces.

Giles Creek

Many of the streams flowing from the ranges onto the piedmont plains drain through paleoflood deposits. In Giles Creek these paleoflood deposits are composed of very coarse gravels and cobbles and provide a resistant boundary to the channel. The morphology of three sites was examined along Giles Creek over a distance of 2 km (Fig. 4.3). Channel gradient increases downstream along this reach (Table 4.1). A long profile was surveyed a distance of 2.8 km downstream from cross section 1. The channel is confined by coarse paleoflood deposits.

Giles Creek 1

This site is located 2.5 km downstream from Allua gap in the ranges (Fig. 4.3). The single thread channel transports medium/coarse gravel and coarse sand. Channel width is 62 m and flood plain to channel width ratio is 0.7:1. Channel gradient in this reach is steep, measuring .00122 m/m. The flood plain surfaces have thick stands of young *Eucalyptus camaldulensis* probably seeded during the floods of the 1970's and 1980's.

Giles Creek 2

This site is located 900 m downstream of Giles Creek 1. At this site the channel width has narrowed to 30.5 m with no flood plain development. Along this reach coarse textured paleoflood sediment is exposed in channel banks and reworked in the channel bed. Channel gradient steepens to .00295 m/m. This site is located immediately downstream of a prominent nick point in the channel long profile.

Giles Creek 3

This site is located 1.12 km downstream from Giles Creek 1. Channel width measures 54 m and the flood plain to channel ratio is 2.6:1. This site is located upstream of a nick point where Pleistocene sediment is exposed in the channel banks. The reach gradient is steeper than the previous two reaches (.0048 m/m).

APPENDIX 3

Statistical Parameters of sediment size

Graphical Method (Folk and Ward, 1957).

ϕn is the particle size (phi-units) at which n% by mass of the distribution is coarser.

1. Graphic Mean: (M_Z)

$$M_Z = (\phi 16 + \phi 50 + \phi 84)/3$$

2. Inclusive Graphic Standard Deviation (σ_I)

$$\sigma_I = [(\phi 84 - \phi 16/4] + [(\phi 95 - \phi 5) / 6.6]$$

APPENDIX 4

Satellite Images

Enhanced satellite images of the Todd Catchment from February and June 1988. The largest flow on the gauged record at Alice Springs occurred in March 1988 and the red areas indicate vegetation growth following the March 1988 rains. The images were donated by Dr. G. Pickup and enhanced by Mr. G. Pearse, CSIRO, Alice Springs.

Image 1. [illegible]e Springs, enhanced TM satellite image, February 1988.

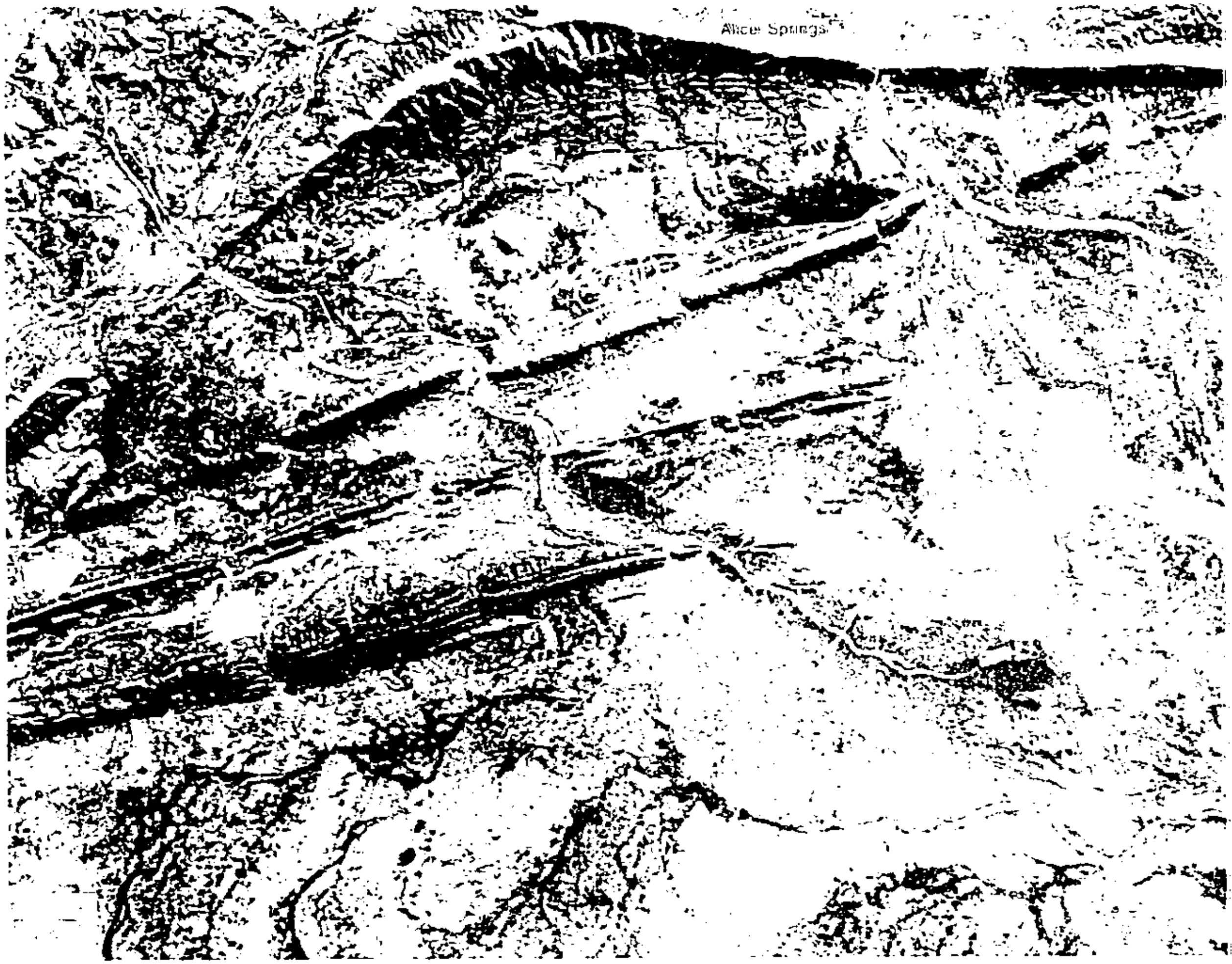


Image 1a. Alice Springs, enhanced IRS satellite image, June 1988.

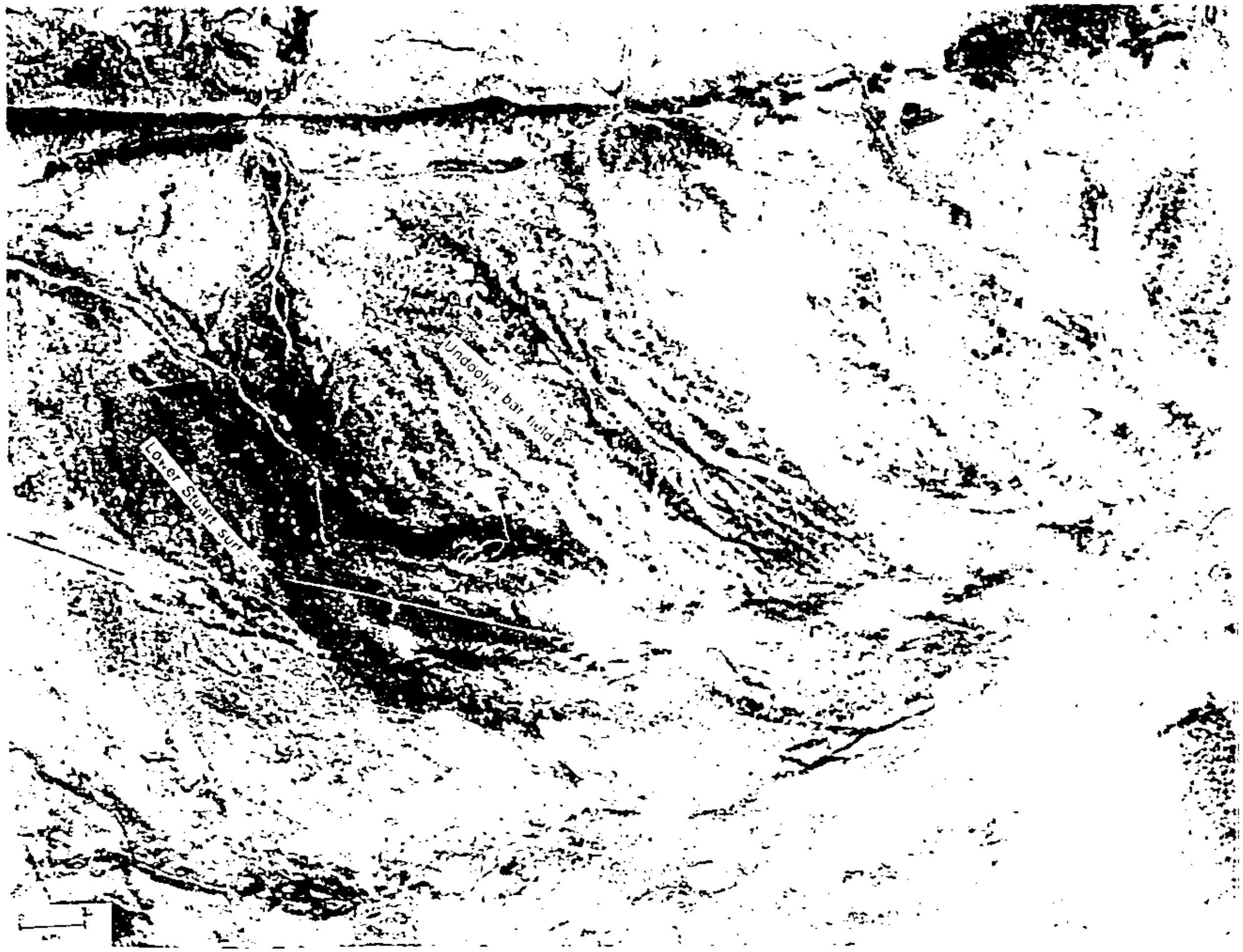


Image 2. Emily and Jessie Creeks, enhanced TM satellite image, February 1988

Image 2a. Emily and Jessie Creeks, enhanced TM satellite image, June 1988.

Image 3. Undoolya Creek, enhanced TM satellite image, February 1988.

Image 3a. [illegible], enhanced TM satellite image, June 1988

Image 4. Williams Creek, enhanced TM satellite image, February 1988.

Image 4a. Williams Creek, enhanced TM satellite image, June 1988.

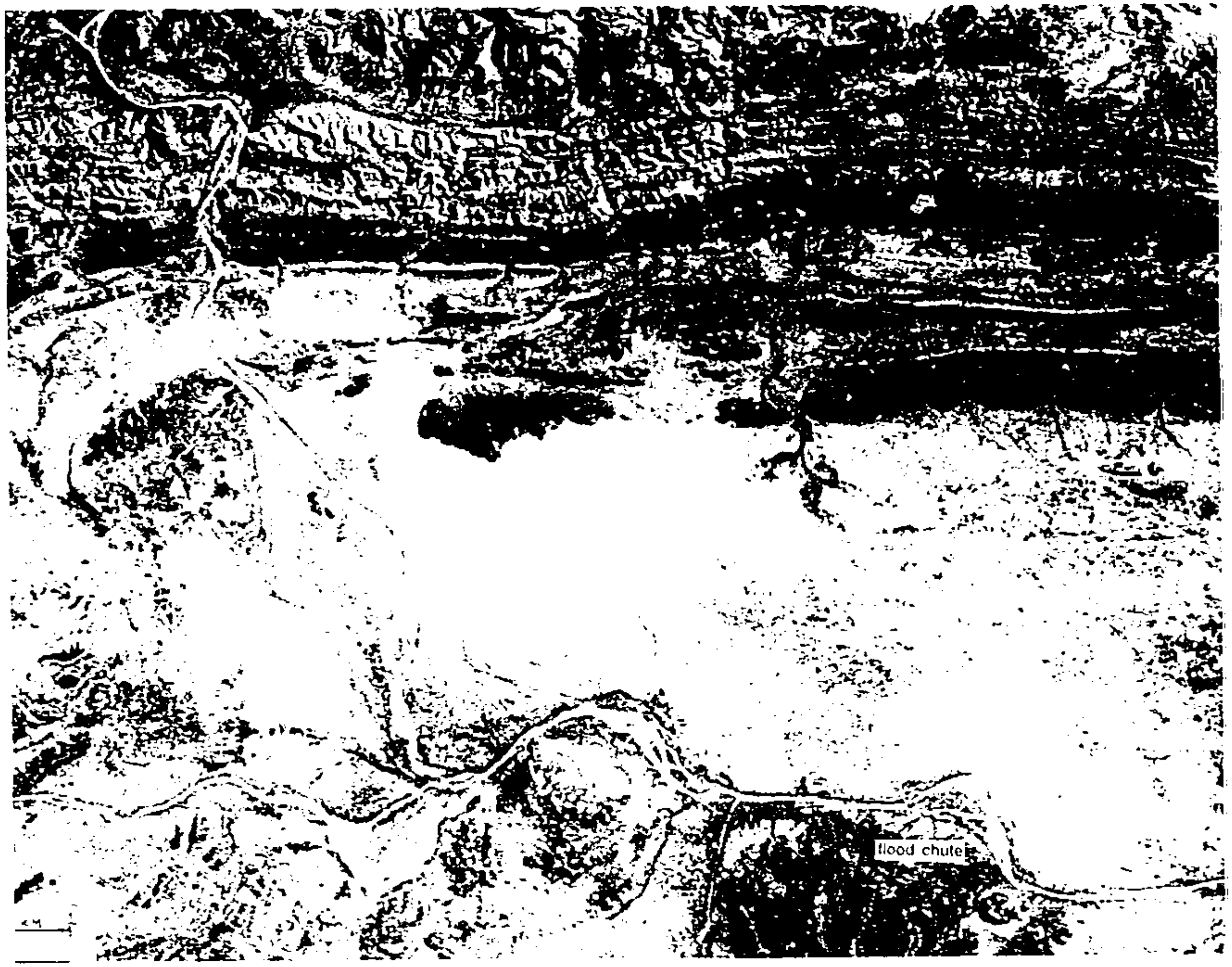


Image 5. Ross River, enhanced TM satellite image, February 1988.

Image 5a. Ross River, enhanced TM satellite image, June 1988.

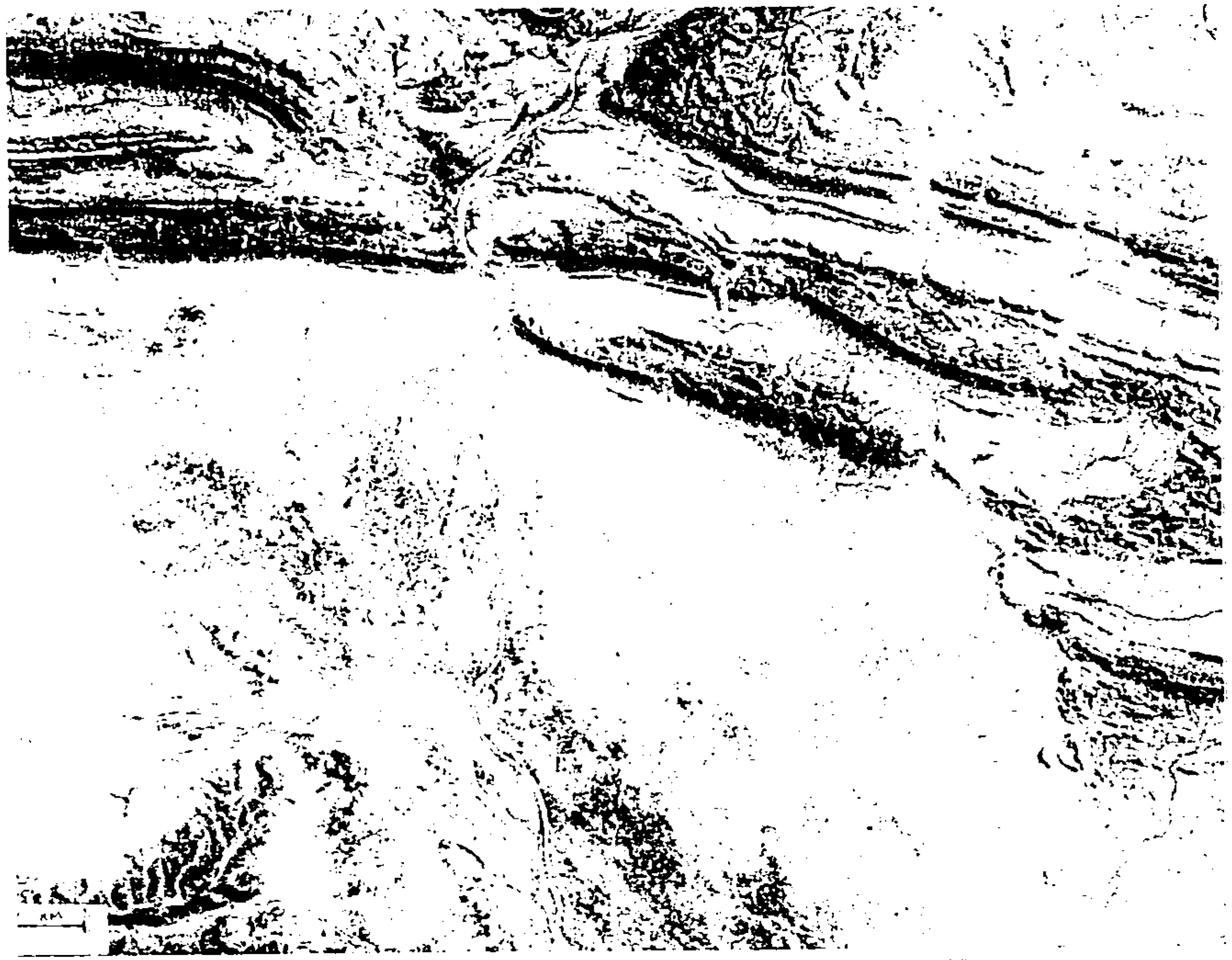


Image 6. Giles Creek, enhanced TM satellite image, February 1988.

Image 6a. Giles Creek, enhanced TM satellite image, June 1988.

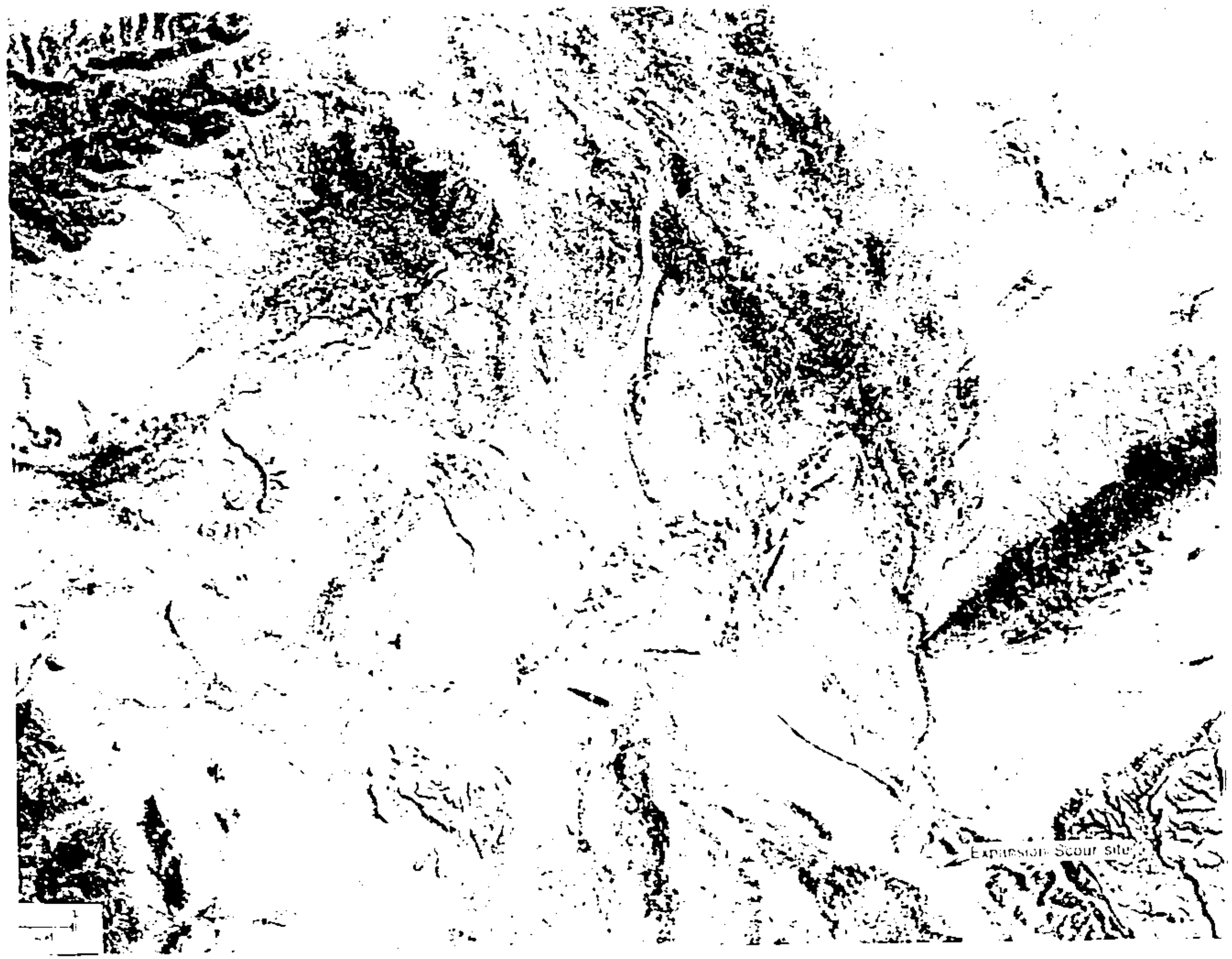


Image 7. Mosquito Bend, enhanced [illegible] satellite image, February 1988.

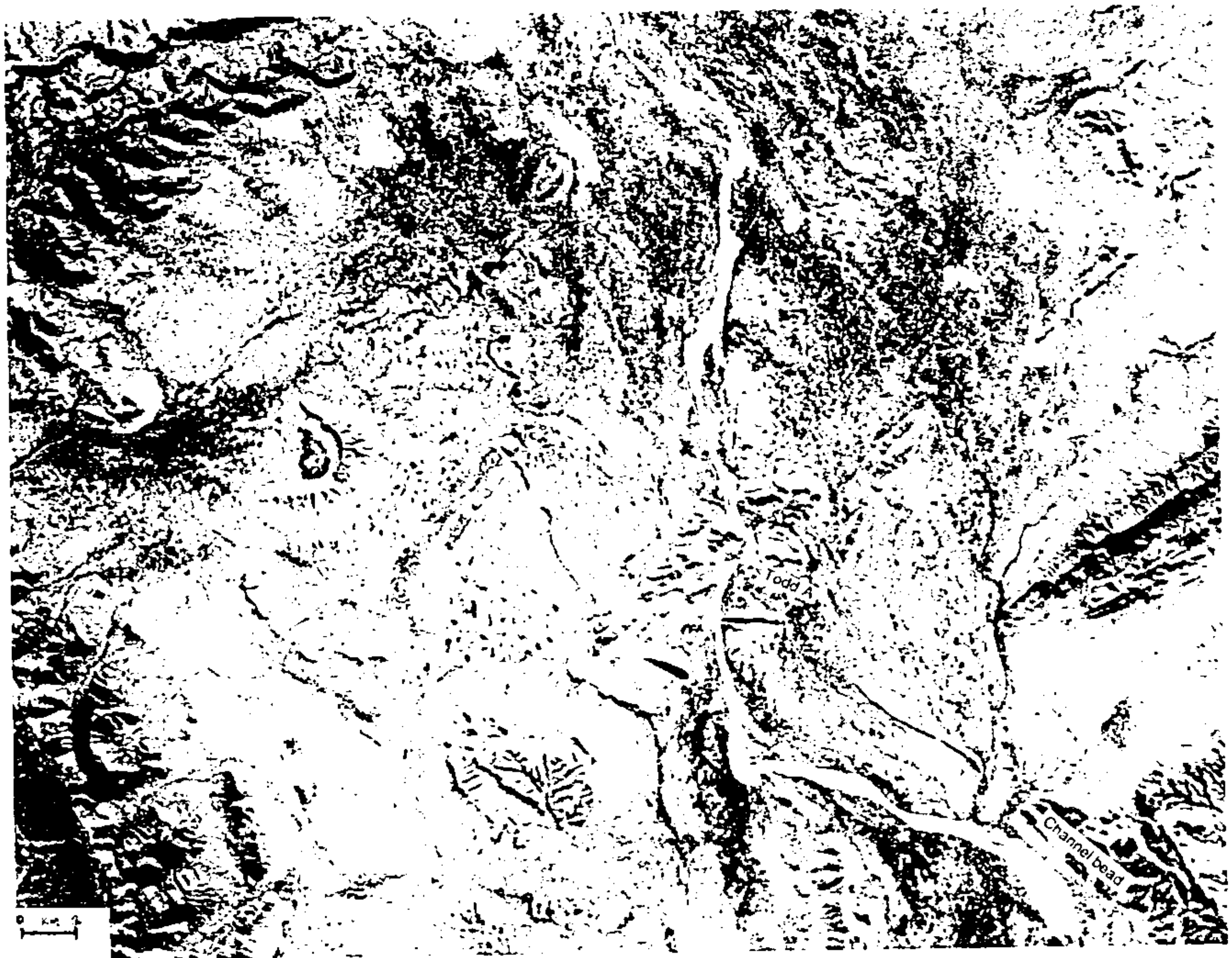


Image 7a. Mosquito Bore, enhanced TM satellite image, June 1988.

Image 8. Mosquito Bore Distributary, enhanced TM satellite image, February 1988.

Image 8a. Mosquito Bore Distributary, enhanced TM satellite image, June 1988.

Image 9. Wallaby Gap, enhanced TM satellite image, February 1988.

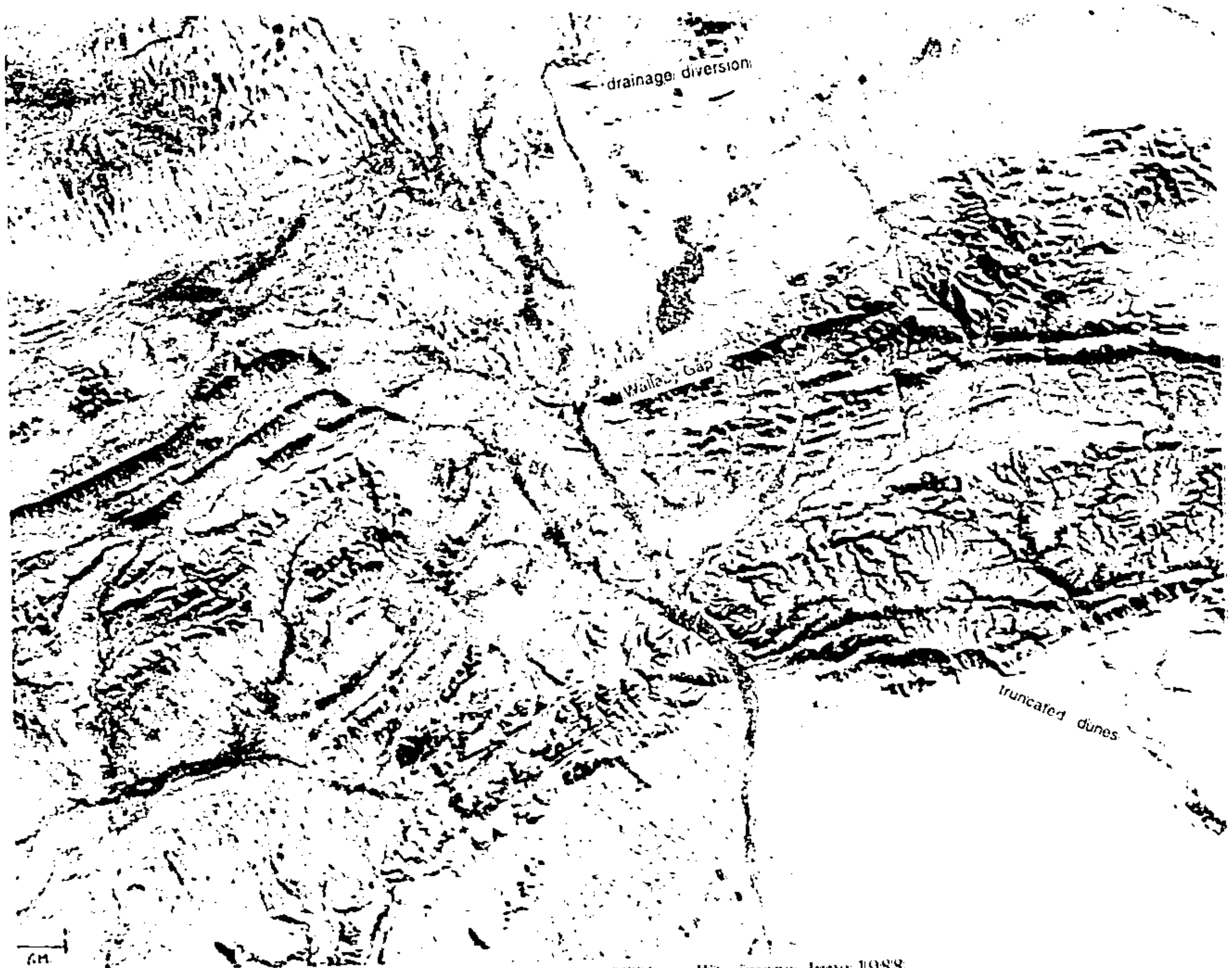


Image 9a. Wallaby Gap, enhanced TM satellite image, June 1988.

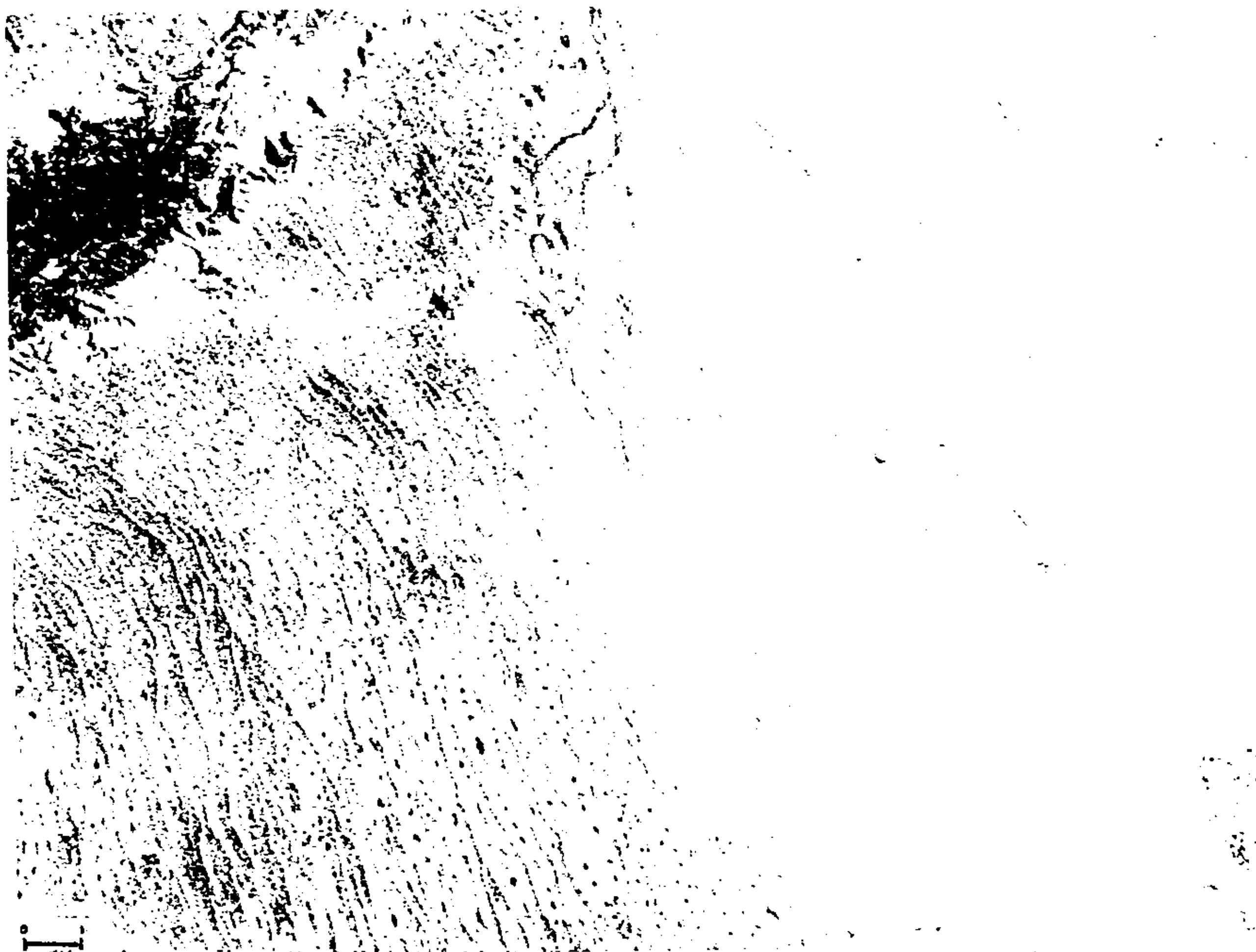


Image 10. Wallaby Gap Distributary, enhanced TM satellite image, February 1988.

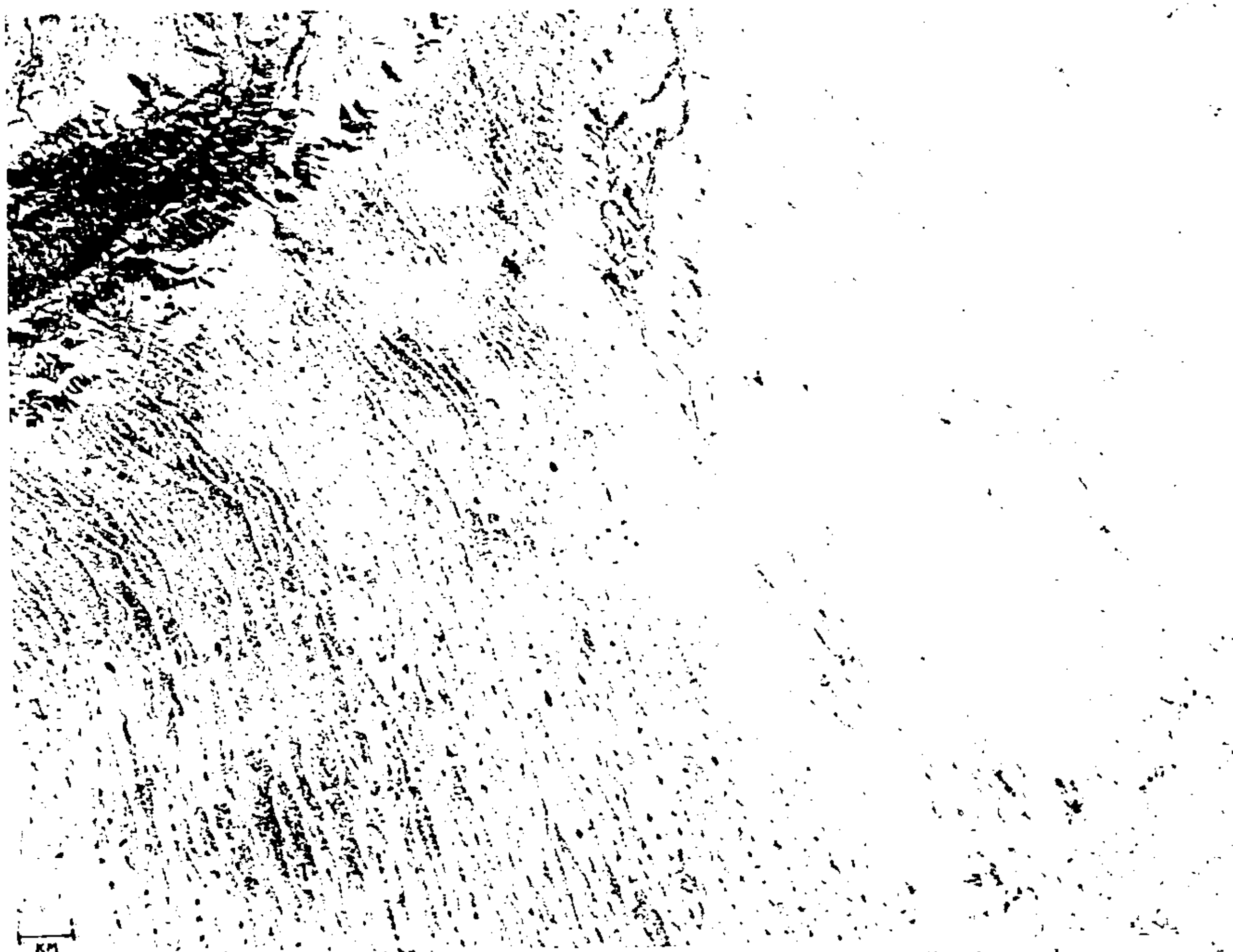


Image 10a. Wallaby Gap Distributary, enhanced TM satellite image, June 1988.

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