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Channel and floodplain characteristics of Cooper Creek, Central Australia

Simon D. Fagan
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CHANNEL AND FLOODPLAIN CHARACTERISTICS OF COOPER CREEK, CENTRAL AUSTRALIA

A thesis submitted in (partial) fulfillment of the requirements for the award of the degree

DOCTOR OF PHILOSOPHY

from

UNIVERSITY OF WOLLONGONG

by

SIMON D. FAGAN
BA (Hons.)(Oxon.)

SCHOOL OF GEOSCIENCES

2001
I, Simon D. Fagan, declare that this thesis, submitted in partial fulfilment of the requirements for the award of Doctor of Philosophy, in the School of Geosciences, University of Wollongong, is wholly my own work unless otherwise referenced or acknowledged. The document has not been submitted for qualifications at any other academic institution.

Simon D. Fagan
Simon Fagan

Channel and floodplain characteristics of Cooper Creek, central Australia

Abstract:

This thesis is an investigation of the formation and distribution of the different channel patterns that co-exist on the broad, low-gradient floodplain surface of the Cooper Creek, an ephemeral, arid-zone river in south west Queensland, Australia. Although previous research described some aspects of the channel as being inherited from wetter Pleistocene periods, all the alluvial surface features were found to be consistent with a contemporary origin broadly in equilibrium with present environmental conditions.

The anastomosing channel system is comprised of interconnected narrow and deep channels with resistant, cohesive boundaries. Levees are formed whose size is related to the size of the proximal channel, and their textural variations are more complex than the usual simple distal fining trends. Levee-like landforms are also formed by the deposition of sediment by flow converging on the channel from the floodplain surface. These forms were termed ‘eevels’ here to indicate that they form in the opposite way to levees, and there are textural differences between these forms and levees. The planform of anastomosing anabranches was analysed by both the direct measurement of channel planform and the fractal method and is consistent with that of freely meandering channels. Results suggest that the disruption of the channel boundary materials by wetting and drying cycles enables and determines bank erosion and channel migration, which seems to be greater in smaller channels due to greater fracturing of the boundaries of these channels. Smaller channels can be much more sinuous than larger channels and frequently form cutoffs.

Elements of the extensive floodplain channel-system formed here interact with and influence the evolution of the larger, inset anastomosing channels. The angles of bifurcation junctions between anastomosing channels tend to be much larger than confluence angles, and a comparison with the angular geometry of junctions of braid-form or floodway channels and anastomosing channels suggests that the latter develop by the enlargement of floodways in a gradual avulsion process. Like the floodplain-surface channels here, the anastomosing channels form in response to high flows and their formation has the effect of increasing the efficiency of overbank flow transport by the floodplain. Reticulate and braid-form patterns cover more than 80% of the floodplain surface, and their distribution and occurrence at both the small and large scale is shown to be determined by overbank flow patterns. Flow patterns also influence the morphology of reticulate networks which have both transitive and space-filling aspects. Braid-form channels co-occur with anastomosing channels where the depths, velocities and inundation
frequencies of overbank flow are relatively high. Reticulate patterns occur where overbank flow is weaker and very infrequently inundated areas are unchanneled. The variation in floodplain-surface pattern occurrence is probably due to the mechanical effect of overbank flow power on the formation of prominent gilgai with which the reticulate pattern co-occurs.
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# Table of contents

1  **INTRODUCTION AND BACKGROUND** ............................................................... 1

1.1 Introduction ........................................................................................................ 1

1.1.1 Aims .................................................................................................................. 3

1.1.2 Selection of the study area ............................................................................... 3

1.1.3 The structure of the study ............................................................................... 4

1.2 Literature assessment ....................................................................................... 5

1.2.1 Definitions of anastomosing rivers .................................................................. 5

1.3 Characteristics of anastomosing channels ....................................................... 9

1.3.1 Channel boundary material and cross sectional form ..................................... 9

1.3.2 Anabranch planform ....................................................................................... 12

1.3.3 Floodplain morphology ................................................................................... 15

1.3.4 Anastomosing channel island planform ......................................................... 18

1.3.5 Flow regime ..................................................................................................... 20

1.4 Anastomosing channel dynamics .................................................................... 20

1.4.1 Avulsion events ............................................................................................... 20

1.4.2 Avulsion and anastomosis – autogenic and exogenic .................................. 29

1.5 The diversity of anastomosis ......................................................................... 30

1.6 Higher-order explanations of anastomosing channel form ......................... 32

1.7 Summary ........................................................................................................... 34

2  **METHODS** .................................................................................................... 36

2.1 Statistics ............................................................................................................. 36

2.2 Sediment size analysis ..................................................................................... 37

2.3 Dating techniques ............................................................................................. 37

2.4 Channel morphology ....................................................................................... 42

2.4.1 Introduction ...................................................................................................... 42

2.4.2 Cross section characteristics .......................................................................... 43

2.4.3 Planform characteristics ................................................................................ 47

2.4.4 Fractal geometry ............................................................................................ 50

3  **REGIONAL SETTING, QUATERNARY HISTORY AND RIVER CHARACTERISTICS** ............................................................... 63

3.1 Introduction ...................................................................................................... 63
3.2 Regional setting ............................................................................................................. 63
  3.2.1 Lake Eyre Basin morphology and climate ................................................................. 63
  3.2.2 Geology .................................................................................................................... 67
3.3 Quaternary history ........................................................................................................ 73
3.4 River characteristics .................................................................................................... 77
  3.4.1 Hydrology ................................................................................................................ 77
  3.4.2 Floodplain landforms ............................................................................................... 82
  3.4.3 Anastomosing channels .......................................................................................... 84
  3.4.4 Waterholes ............................................................................................................. 88
  3.4.5 Floodplain-surface channels ................................................................................... 94
4 ANASTOMOSING CHANNEL PLANFORM ANALYSIS .......................................................... 106
  4.1 Introduction ................................................................................................................ 106
  4.2 Processes of channel bank erosion ............................................................................... 108
  4.3 Observed regularities in single channel planform properties ....................................... 112
    4.3.1 Sinuosity ............................................................................................................... 112
    4.3.2 Regularities in meandering channel form .............................................................. 113
  4.4 Characteristics of the data set ..................................................................................... 117
  4.5 Analysis results .......................................................................................................... 121
    4.5.1 Stream power and slope-discharge predictions ..................................................... 121
    4.5.2 Channel sinuosity ................................................................................................ 121
    4.5.3 Morphological indicators of dynamism: cutoffs and scroll bar occurrence ............. 124
    4.5.4 Meander geometry – results from the direct measurement of channel planform ...... 124
    4.5.5 Fractal Analysis .................................................................................................... 132
  4.6 Discussion .................................................................................................................. 136
  4.7 Summary .................................................................................................................... 141
5 ANASTOMOSING CHANNEL JUNCTIONS ................................................................. 144
  5.1 Introduction ................................................................................................................ 144
    5.1.1 Confluence literature review ................................................................................ 145
    5.1.2 Bifurcation literature review ................................................................................ 150
  5.2 Methodology .............................................................................................................. 151
  5.3 The morphology of channel junctions ....................................................................... 153
    5.3.1 Channel confluences ............................................................................................. 153
  5.4 Junction angular geometry ....................................................................................... 161
5.5 Braid-form channel - anastomosing channel junctions

5.6 Discussion

6 CHANNEL-FLOODPLAIN BOUNDARY FORM AND PROCESS

6.1 Introduction

6.2 Anastomosing channel levee morphology and texture

6.2.1 Methodology

6.2.2 Levee accretion rates and contemporaneity

6.2.3 Levee morphology

6.2.4 Summary of levee morphology analysis results

6.3 Levee texture

6.3.2 Levee texture results

6.3.3 Scroll bar texture

6.3.4 Summary and discussion of levee texture analysis results

6.4 The form and texture of expanding and contracting channel boundaries

6.4.1 Boundary morphology

6.4.2 Texture variations

6.4.3 Summary

6.4.4 Comparison of floodplain sediment-transport implications, anastomosing channel and waterhole levees

7 FLOODPLAIN-SURFACE CHANNELS

7.1 Types and controls on distribution

7.1.1 Introduction

7.1.2 Methodology

7.1.3 Soil properties

7.1.4 Elevation and distribution

7.1.5 Floodplain width and transmission losses

7.1.6 Controls on gilgai formation and floodplain-surface channel pattern

7.1.7 Transitions between floodplain-surface channel pattern types

7.2 Reticulate network morphology

7.2.1 Introduction

7.2.2 Reticulate network topology

7.2.3 Polygon form

7.2.4 Reticulate network morphology: polygon distribution and orientation

7.2.5 The formation and adjustment of reticulate networks

7.3 The braid-form pattern
## Index of figures

Figure 1.1: The north-eastern Lake Eyre Basin showing the study area.  

Figure 1.2: Nanson and Knighton's (1996, p.236) classification of channel patterns. Type 1 channels are low energy, cohesive sediment anastomosing rivers, Type 2 sand dominated, island forming rivers, Type 3 mixed load meandering rivers, Type 4 sand dominated ridge-forming rivers, Type 5 wandering gravel bed rivers, and Type 6 gravel dominated, laterally stable channels.  

Figure 1.3: Floodplain profiles from (a) the anastomosing Alexandra River in Canada showing the prominent development of floodbasins (from Makaske, 1998, p.26), and (b) Red Creek in Wyoming showing marked floodplain convexity (from Schumann, 1989, p.283).  

Figure 1.4: Scatterplot of braid bar length and width measurements, from Rachocki (1981, p.67).  

Figure 1.5: Scatterplot of anastomosing river channel island length and width measurements, Upper Irrawaddy River, from Yonechi and Maung (1986, p.109).  

Figure 2.1: Comparison of surveyed and aerial photograph channel width measurements.  

Figure 2.2: Definition of meander characteristics  

Figure 2.3: The construction of a Koch Curve.  

Figure 2.4: Sierpinski's Sieve.  

Figure 2.5: The measured length-scale of measurement relationship for the South African coast, the west coast of Britain, and a circle (after Mandelbrot, 1982, p.33).  

Figure 3.1: Map of the study area, after Knighton and Nanson (1994a, p.138).  

Figure 3.2: (i) Precipitation and (ii) evaporation distributions over the study area (after Bureau of Meteorology, 2000). Evaporation measured with a Class A Pan.  

Figure 3.3: The geology of the study area (from Queensland Geology 1:2500000 Sheet MD34-75-8) and a schematic cross section of the Cooper Creek valley fill deposits showing the typical thicknesses of each layer.  

Figure 3.4: Comparison of typical floodplain surface morphologies from a heavily gilgaied reticulate area and an unchanneled area, and illustrations of the surface morphology typical of each area.  

Figure 3.5: Soil micrographs showing the increasing compaction of the mud layer with depth.  

Figure 3.6: Selected relative flow magnitude (Q/Qpeak) curves from the flow records of the Curareeva and Nappa Merrie gauging stations (from Knighton and Nanson, 1994a, p.145).  

Figure 3.7: Landsat TM false colour composite image (21/1/95) of the Wilson River (WR) in flood.  

Figure 3.8: Schematic illustration of the floodplain landforms and coexisting channel patterns of the Cooper Creek (from Tooth and Nanson, 2000a, 5).  

Figure 3.9: The anastomosing channel network of the Cooper Creek.  

Figure 3.10: The anabranch channel width-mean depth relationship.  

Figure 3.11: (a) Tooley Wooley (T) and Little Tooley Wooley (LT) floodplain-surface waterholes, (b) Didhelginna Waterhole and (c) Bogala Waterhole, flow from north to south.  

Figure 3.12: Meringhina Waterhole.
Figure 3.13: Examples of reticulate networks showing the intricate, space-filling nature of the patterns.

Figure 3.14: Examples of reticulate channel cross sections.

Figure 3.15: A schematic diagram showing the location of reticulate channels in depressions between gilgai mounds. The bankfull level of the reticulate channel approximates the level of the gilgai depression, and reticulate channels do not occupy every gilgai depression.

Figure 3.16: An example of the greater intensity of cracking of reticulate channel beds compared to the surrounding, higher areas of floodplain surface. Higher intensities of cracking and boundary disruption caused by exposure to higher frequencies of wetting and drying cycles probably facilitate channel formation.

Figure 3.17: Examples of the braid-form floodplain surface pattern.

Figure 3.18: Typical braid-form channel cross sections and an illustration of a braid-form channel in the field.

Figure 3.19: Differences in the relationships between braid-form channels (grey) and anastomosing channels (black) recognised by (a) Rundle (1977) and Rust (1981) who asserted that braid-form channels generally run continuously across anastomosing channels with no disruption to their planform, and (b) Mabbutt (1967) who stated that braid-form channels tend to originate from and terminate at anastomosing channels.

Figure 3.20: Gullied convex channel bank top.

Figure 4.1: The variation of cohesive sediment erodibility with sediment compaction, from Postma (1967, p.158).

Figure 4.2: A heavily fractured, dry bed of a small anastomosing channel showing the effects of drying on boundary sediments.

Figure 4.3: Channel width and reach length characteristics of planform analysis reaches.

Figure 4.4: Examples of meandering channel form from the study reach showing the planform regularity shown by many reaches, cutoff development, and consistency of channel form with Brice's (1984) sinuous canaliform class.

Figure 4.5: Slope-discharge and stream power plots for the 67 anabranch sections in Figure 3.9 (discharge calculated by Mannings n). The slope-discharge plot also shows the channel pattern thresholds of Leopold and Wolman (1957) and Ackers and Charlton (1970).

Figure 4.6: Relationship between channel size and sinuosity of Cooper Creek anabranches.

Figure 4.7: Reach wavelength distribution skewness values.

Figure 4.8: Skewness of reach radius of curvature distributions.

Figure 4.9: Distribution of $\lambda_m/W$ relationships.

Figure 4.10: Distribution of reach median $r_m/W$ values and scatterplot of $r_m$ against $W$ showing the marked variability present.

Figure 4.11: Watawarra Channel morphological map showing reach subdivisions and surveyed cross sections illustrating the downstream contraction of the channel.

Figure 4.12: Selected fractal analysis results.

Figure 4.13: Channel shifting trajectories over time expected as a result of (i) lateral migration, (ii) vertical accretion and (iii) the accretion release process.

Figure 5.1: Mississippi River Basin confluence angle distribution (Mosley, 1976) $n=68$.

Figure 5.2: Braid bar distal angles/ braided channel anabranch confluence angles (Rachocki, 1981)$n=130$. 

vii
| Figure 5.3: Confluence angle distributions, anastomosing Poronai and Irrawaddy Rivers (Yonechi and Maung, 1986). | 146 |
| Figure 5.4: Zones of confluence flow, from Best (1987, p.28). | 148 |
| Figure 5.5: Definitions of measured junction characteristics. | 152 |
| Figure 5.6: Surveyed morphology of the accordant junction at the head of Goonbabina Waterhole and the location of this and the three other junctions subsequently described. | 154 |
| Figure 5.7: Strongly discordant junction north of Goonbabina Waterhole. | 155 |
| Figure 5.8: Bogala Waterhole confluence junction showing bed discordance and, unusually, post-junction bar formation and active junction adjustment. | 155 |
| Figure 5.9: A tributary fan formed at a strongly discordant junction on a feeder channel to Bogala Waterhole. | 157 |
| Figure 5.10: Separation zone formation at confluentes of channel of differing depth, from Best and Roy (1991, p.413). | 160 |
| Figure 5.11: Comparison of Cooper Creek confluence and bifurcation angle distributions. | 162 |
| Figure 5.12: An example of a sequence of ‘scorpion’ bifurcations. Flow from top left, channels densely tree lined. | 162 |
| Figure 5.13: Variation of junction channel combinations with quartile of junction angle: (i) bifurcations, and (ii) conflueses. (Quartile 4=upper quartile). | 165 |
| Figure 5.14: Comparison of confluence angle distributions, Cooper Creek and (i) Mississippi Basin conflueses (Mosley, 1976), (ii) braided channels (Rachocki, 1981), (iii) Poronai and (iv) Irrawaddy Rivers (Yonechi and Maung, 1986). | 166 |
| Figure 5.15: Examples of the interactions between braid-form and anastomosing channels. | 168 |
| Figure 5.16: A comparison of BF-AC and AC-AC junction angle distributions. | 169 |
| Figure 5.17: The postulated development of BF-AC bifurcations into AC-AC bifurcations showing the increase in junction angle with channel enlargement. | 169 |
| Figure 6.1: Location of sampled levees and scroll bars. | 179 |
| Figure 6.2: Definition of levee properties. | 180 |
| Figure 6.3: Dated levee profiles, Levee A. | 183 |
| Figure 6.4: Relationships between levee width WL and the width of the proximal channel. | 185 |
| Figure 6.5: Examples of normalised levee morphology to illustrate the variability in levee form found. | 187 |
| Figure 6.6: Levee morphology and mean grain size trends. | 194 |
| Figure 6.7: Levee morphology and sand content (%) trends. | 195 |
| Figure 6.8: Trends in the sediment texture of scroll bar topography, Meringhina Waterhole. | 198 |
| Figure 6.9: Planform of Didhelginna Waterhole and location of sampled transects. | 203 |
| Figure 6.10: The morphological variation of Didhelginna Waterhole from survey data, calculated discharge by Manning’s n_r (Yen, 1992). | 204 |
| Figure 6.11: Morphology and texture trends of the channel-floodplain boundary around Didhelginna Waterhole. | 205-206 |
| Figure 7.1: Sediment texture plots of surface samples from different floodplain environments, reticulate, unchanneled and braid-form areas. | 216 |
| Figure 7.2: Location and topographic profile of a reticulated, high area of floodplain south of Lake Yamma Yamma. | 218 |
| Figure 7.3: Floodplain topographic profile showing the occurrence of reticulate and braid-form patterns in both high and low areas of the floodplain. | 219 |
Figure 7.4: Location, topographic profile and satellite image (SPOT false colour composite) of an unchanneled high area of floodplain in the south of the study reach.

Figure 7.5: SPOT false colour composite image of the prominent unchanneled areas of floodplain just south of Lake Yamma Yamma. Images of the areas were studied by Robinove (1978) and Baker (1986).

Figure 7.6: The relationship between variations in floodplain and floodplain-surface pattern extent over the study reach.

Figure 7.7: The continuum model of floodplain-surface pattern distribution.

Figure 7.8: The transition between braid-form and reticulate floodplain-surface patterns near the confluence with the Wilson River.

Figure 7.9: The fretting of braid-form channels around a braid-form island and the beginning of reticulate pattern formation at the transition between braid-form and reticulate pattern areas, South Galway area north of Lake Yamma Yamma.

Figure 7.10: Transition from braid-form to reticulate pattern, MVS area north of Lake Yamma Yamma.

Figure 7.11: Transition between braid-form and unchanneled floodplain areas.

Figure 7.12: Definition of reticulate network link types. LIL - looped interior links, TIL - terminal interior links, EL - exterior links.

Figure 7.13: Oblique aerial photographs of reticulate networks showing clear polygon development and the hierarchy of network links defined in Figure 7.12.

Figure 7.14: Definition of reticulate polygon morphology characteristics measured and examples of polygon form from the Naccowlah area.

Figure 7.15: Location of M2 study area.

Figure 7.16: Reticulate polygons and polygon orientations, M2 area.

Figure 7.17: Surface plots of polygon descriptor variables, M2 area.

Figure 7.18: Location of and polygon orientations in the Barrolka reticulate study area.

Figure 7.19: Map of polygon orientations in the MVS study area.

Figure 7.20: Surface plots of polygon descriptor variables, MVS area.

Figure 7.21: Location of and polygon orientation in the Naccowlah study area.

Figure 7.22: Surface plots of polygon descriptor variables, Naccowlah area.

Figure 7.23: (a) Location of the braid-form pattern study areas in the north of the study reach, and (b) an aerial photograph showing the pattern developed over a part of the smaller of the two study areas shown in (a).

Figure 7.24: Examples of braid-form island planform.

Figure 7.25: Scatterplot of braid-form island length and width dimensions.

Figure 7.26: The floodplain accretion transect showing the position of samples and variations in elevation and floodplain-surface channel pattern occurrence over the area.

Figure 7.27: Floodplain accretion dated profile from the Wilson's Swamp area.
Index of tables

Table 1.1: Summary of published anastomosing river characteristics data. 10-11
Table 1.2: Causes of avulsion events. 23
Table 2.1: A comparison of TL and OSL dating results. 41
Table 3.1: Climate averages for Hughenden Post Office Meteorological Station (Bureau of Meteorology, 2000). 66
Table 3.2: Climate averages for Windorah Meteorological Station (Bureau of Meteorology, 2000). 66
Table 3.3: Climate averages for Moomba Meteorological Station (Bureau of Meteorology, 2000). 66
Table 3.4: Selected hydrological characteristics of the Cooper Creek. 77
Table 3.5: Range of reticulate channel morphological characteristics found. 95
Table 3.6: Range of braid-form channel morphological characteristics found. 101
Table 4.1: Reported \(\lambda\)-W relationships for freely meandering alluvial channels. 114
Table 4.2: Empirical relationships between meander and channel properties. 128
Table 4.3: Watawarra Channel planform, cross section and calculated hydraulic characteristics (reach values averaged from 3 cross sections per reach). 132
Table 4.4: Characteristics of reaches and results of fractal analysis. 133
Table 4.5: The variation in calculated fractal dimension with measurement scale, Reach W3. 135
Table 4.6: Sinuosity development of a meandering channel with differing lateral shifts at bend apices. 138
Table 4.7: Accretion depths needed to enable an initially straight channel of \(\lambda=10W\), \(W=10d_{\text{max}}\) and 45° outbanks to attain various sinuosities by the accretion release process. 140
Table 5.1: Angles of stream divergence and their variation with stream order on the Sutlej-Yamuna Plain terminal fans, India (after Mukerji, 1976, p.195). 150
Table 5.2: Capacities and calculated discharges of waterhole confluences. 161
Table 5.3: Summary of junction angle results. 163
Table 5.4: Braid-form channel-anastomosing channel junction angular geometry results. 170
Table 6.1: Morphological characteristics of anastomosing channel levees, for locations see Figure 4.1. 182
Table 6.2: Results of levee shape analysis. 186
Table 6.3: Characteristics of texture analysis levees. 192
Table 6.4: Composition of sediment samples (n=57). 192
Table 6.5: Correlations between sediment size variables (n=57). 192
Table 6.6: Comparison of sand composition, proximal and distal levee boundaries. 196
Table 6.7: Channel boundary morphology, Didhelginna Waterhole. 207
Table 7.1: Results of reticulate pattern occurrence regression analyses (n=57). *Denotes not significant at \(\alpha=0.05\). 224
Table 7.2: Results of braid-form pattern occurrence regression analyses (n=57). *Denotes not significant at \(\alpha=0.05\). 224
Table 7.3: Results of unchanneled area occurrence regression analyses (n=57). *Denotes not significant at \(\alpha=0.05\). 225
Table 7.4: The variability of reticulate polygon form (n=1122). 238
Table 7.5: Results of correlation analysis between polygon descriptor variables (n=1122) (* indicates \(\rho\) not significant at \(\alpha=0.05\)). 238
Table 7.6: The variation of braid-form island morphology (n=62). 257
Table 7.7: Results of braid-form island morphology analysis (n=62). * Denotes not significant at \(\alpha=0.05\). 260
Table 7.8: Braid-form island angle variation (n=62). 260
List of symbols and abbreviations

α - statistical significance level
λ - meander wavelength
ρ - Pearson’s Correlation Coefficient
%_{BRA} - percentage of floodplain width occupied by the braid-form channel pattern
%_{RET} - percentage of floodplain width occupied by the reticulate channel pattern
%_{UNCH} - percentage of floodplain width unchanneled
A - channel bankfull cross sectional area
A_{B} - braid-form island area
A_{p} - reticulate polygon area
AC - anastomosing channel, used to refer to a type of waterhole
AC-AC - refers to junctions (confluences or bifurcations) between two anastomosing channels
BF-AC - refers to junctions (confluences or bifurcations) between a braid-form channel and an
anastomosing channel
CV - coefficient of variation
d_{ave} - average depth
d_{max} - maximum depth
D - fractal dimension
Dd - distance downreach
\hat{D} - braid-form island distal angle
E_{B} - braid-form island elongation (L_{B}/W_{B})
E_{p} - reticulate polygon elongation (L_{p}/W_{p})
EL - exterior link
F - channel width/depth ratio
FS - floodplain slope
g - acceleration due to gravity (9.81 ms^{-2})
G - measurement scale (fractal analysis)
H - levee height
h(x) - levee height at distance x from the channel edge
L_{B} - braid-form island length
L_{p} - reticulate polygon length
LIL - looped interior link
LSLR - least squares linear regression
M - percentage of silt and clay in the channel boundary
N - number of lengths of scale G required to measure an object by the divider method (fractal
analysis)
n_{n} - roughness coefficient for the dimensionally balanced Manning equation (Yen, 1992)
OSL - optically stimulated luminescence
P - sinuosity
\hat{P} - braid-form island proximal angle
PMH - point of maximum levee height, in metres from the channel edge
PMH/W_{L} - distance of levee high point from the channel edge as a proportion of total levee
width W_{L}
Q_{av} - average discharge
Q_{B} - bankfull discharge
Q_{mean} - mean discharge
Q_{max} - maximum discharge
Q_{peak} - peak discharge
R - hydraulic radius
r_{n} - radius of curvature
S - slope
TIL - terminal interior link
TL - thermoluminescence
v - flow velocity
$W$ – bankfull channel width
$W_B$ – braid-form island width
$W_{BRA}$ – width of floodplain occupied by braided channels
$W_F$ – floodplain width
$W_L$ – levee width
$W_p$ – reticulate polygon width
$W_{RET}$ – width of floodplain occupied by reticulate channels
$W_{UNCH}$ – width of floodplain unchanneled
$x$ – distance across levee
1 INTRODUCTION AND BACKGROUND

1.1 Introduction

While not terrae incognitae, inland Australia and its fluvial geomorphology are much less well known than the rivers of the more populated coastal regions (Tooth and Nanson, 1995). This is true of desert rivers generally, although it must be acknowledged that a great deal of work has been done on them in recent years (for reviews see Graf, 1988; Cooke et al., 1993; Tooth, 2000). Reasons for the comparative neglect of Australian desert streams are the relative inaccessibility of the inland regions and also the greater practical imperative to understand the behaviour of rivers in densely populated areas. Although deserts are defined by low rainfall, fluvial processes may still be very significant in determining landscape form there as is evidenced by the common occurrence of landforms like wadis, alluvial fans and playas. Also, in many areas like the study area desert rivers are exoreic (Cooke et al., 1993) and rise in wetter regions before flowing into or through the desert, leading fluvial processes to be more important than local precipitation would suggest.

Cooper Creek is part of a vast area of the 1.14million km² Lake Eyre basin in central Australia known as the Channel Country and characterised by intricate networks of channels crossing broad valley floors. The Lake Eyre basin was formed by a downwarping event during the late Cainozoic (Wopfner and Twidale, 1967) and is of very low relief. The climate of the basin is arid to semi-arid and the tributaries of the river rise in the seasonally wet north of the catchment from where flow is fed to the more arid areas downstream. The Cooper Creek in name begins at the confluence of the Thompson and Barcoo Rivers - one of the few instances of two rivers combining to form a creek - and terminates some 1500km from its source at Lake Eyre 15m below sea level. This thesis studies the reach of the Cooper Creek from Windorah to its contraction into a single channel within the Innamincka Dome, which despite the singular channel suggested by the name Cooper Creek is the only part of the study reach where the river does flow as a single channel (Figure 1.1). The gradient of the river in the study reach is very low and approximates 0.00015. The valley ranges up to approximately 60km in width and is covered by an array of channel types and scattered aeolian dunes. The development of oil and gas extraction industries in the area since the 1960’s has made it relatively accessible and as a consequence a number of geomorphic studies have been undertaken over the last 25 years (e.g. Rundle, 1977; Veever and Rundle, 1979; Rust, 1981; Nanson et al., 1986; Rust and Nanson, 1989; Maroulis, 1992; Maroulis, 2000).
Figure 1.1: The north-eastern Lake Eyre Basin. The Cooper Creek extends from the confluence of the Thompson and Barcoo Rivers to Lake Eyre. The boundary of the study reach is shown by the dashed line.
1.1.1 Aims

The broad aim of this study is to investigate the geomorphology and related dynamics of a large, inland, low-gradient anastomosing river system. Only two other arid-zone anastomosing rivers have been investigated in any detail; Red Creek in Wyoming, U. S. A. (Schumann, 1989), and the inland Niger Delta (Makaske, 1998). Dryland anastomosing rivers are still considered unusual to such an extent that Miall (1996, p.52) described the Cooper Creek as "anomalous", implying that it is somehow not a 'normal' anastomosing river, whatever that might be. What an anastomosing river is and what characteristics qualify a river to be described as anastomosing are significant questions that will be addressed later in this chapter. The more detailed aims of this study are:

1) to investigate anabranch dynamics and behaviour through the study of planform properties using both the direct measurement of channel planform and the fractal method;
2) to investigate the morphology and angular geometry of anastomosing anabranch junctions, and, particularly, to compare the angular geometry of confluence and bifurcation junctions which has, it would appear, not previously been undertaken in multiple channel systems;
3) to investigate levee formation, occurrence and the variability of levee morphology and textural variations because levees are significant determinants and signatures of the process of anastomosing formation;
4) to describe and explain the occurrence and morphology of the various floodplain-surface channel pattern types which are formed extensively here because the patterns have not previously been described or studied in any detail, here or elsewhere;
5) to formulate a model of anastomosing channel formation and dynamics, and to consider the applicability of previous theories of anastomosing channel formation to the system.

1.1.2 Selection of the study area

The study area on Cooper Creek was selected for a number of reasons. Firstly, there is now detailed knowledge of the Quaternary history of this river (Rundle, 1977; Rust, 1981; Rust and Legun, 1983; Rust and Nanson, 1986; Nanson et al., 1986; Nanson et al., 1988; Nanson and Tooth, 1999; Maroulis, 2000) which has made it possible to evaluate the possible influence of past conditions on contemporary fluvial geomorphology. Secondly, the hydrology of the system (Knighton and Nanson, 1994a, 2000b) and the large, expanded sections of channel termed 'waterholes' (Knighton and Nanson, 1994b; Knighton and Nanson, 2000a) have been studied in detail and provide a useful basis from which to investigate those aspects that have
not been so studied, including the extensive floodplain-surface channel patterns, the angular geometry of confluence and bifurcation junctions, and levee occurrence and characteristics. Thirdly, the river is very different in character to the majority of dryland rivers that have been studied which have generally been small rivers with coarse sediment loads flowing in relatively steep basins (e.g. Leopold and Miller, 1956; Thornes, 1977; Rendell and Alexander, 1979; Reid and Frostick, 1987; Walters, 1989; Hughes and Sami, 1992; Laronne et al., 1994). For this reason the findings of this study should be a useful addition to the present knowledge of dryland rivers.

1.1.3 The structure of the study

Chapters 1 reviews the previous work on anastomosing channel morphology and dynamics. Chapter 2 outlines the analytical methods used including a brief description of the statistics, the dating techniques and the methods of channel form analysis. The theoretical basis of the fractal method which will be used to study channel planform is outlined. To complete the introduction Chapter 3 describes the regional setting and characteristics of the study area including physiography, climate, geology and Quaternary history. The aeolian and fluvial floodplain landforms of the study reach are described and the characteristics of the range of channel pattern types found and their processes of formation briefly considered. The findings of previous studies of the area are reviewed and some deficiencies in present knowledge highlighted. Chapter 4, following a review of the findings of studies of single channel planform regularities, describes the results of the planform analyses of the anastomosing channels and considers their implications both for channel behaviour and methodologically. In Chapter 5 another aspect of the anastomosing channel system, the junctions between anastomosing channels, is investigated. The morphology of confluence junctions is considered in some detail, and the angular geometries of confluence and bifurcation junctions are studied and compared. Junctions between anastomosing channels and floodplain-surface channels are also described, and the implications of the results of junction analysis for anastomosing channel dynamics considered. In Chapter 6 the analysis of the morphology and texture of the channel-floodplain boundary of anastomosing channels, both constant in width and changing markedly in size, is reported. In Chapter 7 the distribution of floodplain-surface channel patterns, their character and morphological variations, and controls on their formation are described. Chapter 8 presents a summary of the major findings and their implications, and a model of anastomosing channel behaviour is outlined. Some philosophical issues prompted by the study are raised and suggestions for future research considered.
1.2 Literature assessment

1.2.1 Definitions of anastomosing rivers

There has recently been great interest in the morphology, formation and stratigraphy of anastomosing rivers (Smith, 1973; Smith and Putnam, 1980; Smith and Smith, 1980; Rust, 1981; Smith, 1983; Nanson et al., 1986; Rust and Nanson, 1986; Nanson et al., 1988; Schumann, 1989; Smith et al., 1989; Miller, 1991; Kirschbaum and McCabe, 1992; Harwood and Brown, 1993; Knighton and Nanson, 1993; Törnqvist et al., 1993; Knighton and Nanson, 1994b; Smith and Pérez-Arlucea, 1994; Nanson and Knighton, 1996; Schumm et al., 1996; Smith N. D. et al., 1997; Makaske, 1998; Nanson and Huang, 1999). This has been because of both an increasing recognition of contemporary anastomosing river occurrence, and the realisation that anastomosed fluvial deposits are much more widespread in the stratigraphic record than was formerly appreciated. The first recorded description of a river as anastomosing was in 1834 when the term was viewed as being synonymous to the term 'braided' and distinctions between the two terms were not recognised until relatively recently (Smith and Putnam, 1980). One of the first uses of the term anastomosis which distinguished its meaning from the term braided was by Bretz (1923) in his work on the Channeled Scabland in Washington State, U.S.A. Bretz applied the term to the erosionally developed network of scabland channels separated by bedrock islands which he stated were often relict topographic remnants of a more continuous surface, however Bretz did not explicitly define the term and the precise meaning and scope he intended is unclear. There have been many different definitions of anastomosing rivers, and while there is still no universally accepted definition there is a deal of agreement on the typical morphology and characteristics of anastomosing rivers. In the first widely cited definition of an anastomosing river Schumm (1968), using examples of Australian rivers, defined an anastomosing river as a laterally inactive, low gradient, predominantly suspended load multiple channel stream in which channels are divided by stable vegetated islands. Later, Schumm (1977) noted that individual anabranches of anastomosing streams may exhibit their own channel patterns and be straight, meandering or braided. Anastomosing rivers as defined here thus differ greatly from braided rivers which are single channel bedload rivers divided at below bankfull flows by island bars composed of either sediment or vegetation (Schumm, 1977). Braided rivers tend to be high energy, highly dynamic channels transporting large quantities of bedload whose bar-forms and channels are in constant flux (e.g. Leopold and Wolman, 1957; Fahnestock, 1963; Bridge, 1993; Ferguson, 1993; Murray and Paola, 1994).
In a more inclusive definition than that of Schumm (1968), Smith (1983) defined anastomosis on the basis of both channel and floodplain characteristics stating that “anastomosing channels are rapidly aggrading low-energy channel and wetland complexes” (p155). Knighton and Nanson (1993) consider the morphological relationship between anastomosing anabranches and floodplain islands as distinctive of the anastomosing pattern, and so define an anastomosing river as one which “consists of multiple channels separated by islands which are usually excised from the continuous floodplain and which are large relative to the size of the channels” (p.613). The inclusion of some aspects of floodplain morphology in the above definitions of anastomosing rivers is a recognition that some channel patterns are best described by a more inclusive and general classification than considering within-channel processes alone allows (Richards et al., 1993). The following definition will be adopted here as a working definition of an anastomosing river. An anastomosing river is a low energy multiple channel system with resistant banks in which stable or slowly migrating anabranches are separated by irregular islands of approximately floodplain height that are large with respect to the width of the channels. This definition shares some characteristics with definitions discussed earlier, and the justification for it will become clear in our subsequent detailed consideration of anastomosing river characteristics.

The restriction of the term anastomosing to low energy channels is generally accepted (Smith and Putnam, 1980; Schumm, 1981; Knighton and Nanson, 1993; Nanson and Knighton, 1996) although not universal (e.g. Miller, 1991). Anastomosing and braided rivers are not the only types of multiple channel system recognised and a number of authors have proposed more inclusive classifications of multiple channel systems. Nanson and Knighton (1996) consider anastomosing rivers to be a low energy subclass of a larger multiple channel class which they define as ‘anabranching’. The term ‘anabranching’ was originally introduced in Australia in the early nineteenth century as an abbreviation of ‘anastomosing branch’ (David and Browne, 1950), and an anabranching river is defined by Nanson and Knighton (1996, p.218) as being “a system of multiple channels characterised by vegetated or otherwise stable alluvial islands that divide flows at discharges up to nearly bankfull.” This definition will be adopted here. The classification of Nanson and Knighton (1996) is shown in Figure 1.2. Nanson and Knighton (1996) only consider low energy anabranching channels with relatively laterally stable anabranches to be anastomosing and also recognise high energy anabranching channels such as wandering gravel bed rivers like the Bella Coola and Squamish Rivers in British Columbia, Canada.
Figure 1.2: Nanson and Knighton's (1996, p.236) classification of channel patterns. Type 1 channels are low energy, cohesive sediment anastomosing rivers, Type 2 sand dominated, island forming rivers, Type 3 mixed load meandering rivers, Type 4 sand dominated ridge-forming rivers, Type 5 wandering gravel bed rivers, and Type 6 gravel dominated, laterally stable channels.
Richards et al. (1993) make a different distinction between classes of non-braided multiple channel systems and propose that there is a continuum between anastomosing channels and what they define as avulsive channel systems which parallels the meandering-braiding continuum. The importance of avulsion processes is said to vary between the two classes of channels with avulsion being widespread in anastomosing river dynamics but “the dominant long term influence” on the sedimentary history and channel evolution of avulsive channel systems (Richards et al., 1993, p.201, emphasis in original). Richards et al. (1993) give as an example of an avulsive channel system the River Rapti flowing on the northern Gangetic Plain. While acknowledging that possible differences need to be assessed more fully with data from more river systems, Richards et al. state that differences between the two classes seem to be that there are: “(i) larger-scale tracts of land between channels in avulsive systems in unconfined sedimentary basin settings compared to confined valley settings; (ii) relatively more stable channels in anastomosing systems...compared to the behaviour of dominant channels in avulsive systems; (iii) high sediment loads and rapid aggradation in avulsive systems, generating topographic differences relatively rapidly; (iv) large discharge ranges with sustained high flows in avulsive systems, encouraging large-scale channel diversion; and (v) a higher ratio of channel to overbank aggradation in avulsing systems, and rapid bed aggradation which results in channels perching above surrounding floodplain areas” (p. 201). Richards et al. state that the differences between classes of river reflect different scales and rates of various depositional processes, however while these differences undoubtedly occur between some rivers they do not seem to occur in all cases. For example, a number of anastomosing channel systems have been shown to be aggrading extremely rapidly and the anastomosing upper Columbia River in Canada is aggrading at 0.6m/100yrs (Smith, 1983), anastomosing rivers sometimes show high ratios of channel to overbank aggradation (e.g. Törnqvist et al., 1993; Miall, 1996; Makaske, 1998), and anastomosing rivers typically experience sustained high flows and high discharge ranges (e.g. Schumann, 1989; Makaske, 1998). Also, it will be argued later in this chapter that avulsion is the dominant influence on the sedimentary history and channel evolution of anastomosing channels, and it may be more important in anastomosing systems than avulsive channel systems because as Richards et al. note channels in anastomosing systems tend to be more laterally stable than those in avulsive channel systems. While the distinction made by Richards et al. may be descriptively useful, the differences between the channel types noted seem to be primarily differences in setting as opposed to fundamental differences in river type.
The distinction between types of multiple channel system made by Nanson and Knighton (1996) is useful and will be used here. There are clearly marked differences in morphology and characteristics between classes of anabranching channel, although it is not clear to what degree these reflect differences in the formative processes and essential characteristics of types of anabranching channel pattern. Different types of anabranching channel patterns may have the same cause, and as with the continuum between anastomosing and avulsive channel systems outlined by Richards et al (1993), differences between most anabranching pattern types recognised by Knighton and Nanson (1996) seem to be mainly of degree rather than kind, with the exception of systems like ridge forming channels which generate multiple channels by the within-channel deposition of sediment ridges rather than by the avulsion processes which characteristically produce multiple channels in the other anabranching patterns.

1.3 Characteristics of anastomosing channels

A summary of published data on anastomosing rivers is presented in Table 1.1

1.3.1 Channel boundary material and cross sectional form

Strong bank materials which are highly resistant to erosion are a common feature of anastomosing channels (Smith, 1973; Smith and Smith, 1980; Knighton and Nanson, 1993). In humid environments vegetation is often important in causing banks to be resistant and on anastomosing rivers in humid regions of Canada provides a 'rip-rap like' protection to banks from erosion (Smith, 1973). Smith (1976) found that the root mats of vegetation on channel banks of the Alexandra River in Canada, are 20,000 times more resistant to erosion than comparable root free sediment. In more arid areas, bank resistance is generally provided by high proportions of cohesive sediments as is the case for Red Creek in Wyoming (Schumann, 1989) and the Cooper Creek. Commonly, anastomosing channels have a combination of cohesive bank sediments and protective bank vegetation and the only river in Table 1.1 with largely non-cohesive, sandy banks is Magela Creek in the humid Northern Territory of Australia which has resistant banks due to the protection of a dense mat of tree roots (Nanson et al., 1993).

The bed materials of anastomosing channels are not as uniform as bank materials tend to be and vary from predominantly cohesive silt and clay for the Cooper Creek and Red Creek
Table 1.1: Summary of published anastomosing river characteristics data.

<table>
<thead>
<tr>
<th>River</th>
<th>Source/Reference</th>
<th>Climate</th>
<th>Flow regime</th>
<th>Discharge (cumecs)</th>
<th>Slope (m/m)</th>
<th>Lateral activity</th>
<th>Anabranch Sinuosity</th>
<th>Anabranch Width-Depth ratio</th>
<th>Sedimentation rate (m/100yr)</th>
<th>Sediment load</th>
<th>Bed material</th>
<th>Bank material</th>
</tr>
</thead>
<tbody>
<tr>
<td>Gearagh Harrow (1993)</td>
<td>Harwood and Brown</td>
<td>Humid temperate</td>
<td>Perennial</td>
<td>1.5-2 year flood = 61</td>
<td>-</td>
<td>Stable</td>
<td>Low</td>
<td>4.5-15.7 mean = 11.3</td>
<td>-</td>
<td>gravel bedload, fine suspended sediment</td>
<td>gravel</td>
<td>fine sediments, root masses</td>
</tr>
<tr>
<td>Red Creek, Wyoming</td>
<td>Schurmann (1989)</td>
<td>Arid</td>
<td>Seasonal snowmelt dominated</td>
<td>Qb = 13</td>
<td>0.0011</td>
<td>Slow, transient migration</td>
<td>1.1-1.9</td>
<td>4-12 (main channel)</td>
<td>-</td>
<td>-</td>
<td>Cohesive silt-clay</td>
<td>Cohesive silt-clay</td>
</tr>
<tr>
<td>Magela Creek</td>
<td>Nanson et al. (1993)</td>
<td>Tropical monsoonal</td>
<td>Seasonal snowmelt and glacier dominated</td>
<td>Qb = 40 Qmax = 3,000-4,000</td>
<td>0.0005</td>
<td>Stable</td>
<td>1.1</td>
<td>10 (mean)</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>Medium sand</td>
</tr>
<tr>
<td>Upper Columbia</td>
<td>Smith (1983)</td>
<td>Cold temperate</td>
<td>Seasonal snowmelt and glacier dominated</td>
<td>Qb = 275</td>
<td>0.000096</td>
<td>Very slow lateral accretion</td>
<td>1.16</td>
<td>16</td>
<td>Aggrading 0.15</td>
<td>-</td>
<td>Medium sand</td>
<td>Fine sand/silt</td>
</tr>
<tr>
<td>Lower Saskatchewan n</td>
<td>Smith (1983)</td>
<td>Subhumid continental</td>
<td>Seasonal snowmelt and glacier dominated</td>
<td>Qav = 500 Qb = 1,400</td>
<td>0.000022</td>
<td>Stable</td>
<td>1.4</td>
<td>10</td>
<td>Aggrading 0.15</td>
<td>-</td>
<td>Medium sand</td>
<td>Fine sand/silt</td>
</tr>
<tr>
<td>Alexandra River</td>
<td>Smith and Smith (1980)</td>
<td>Cold temperate</td>
<td>Seasonal snowmelt and glacier dominated</td>
<td>Qb = 66</td>
<td>0.0006</td>
<td>Stable</td>
<td>up to 2.5</td>
<td>8-23 mean = 13</td>
<td>Aggrading 0.18</td>
<td>-</td>
<td>Gravel/coarse sand</td>
<td>Silt dominated</td>
</tr>
<tr>
<td>North Saskatchewan n</td>
<td>Smith and Smith (1980)</td>
<td>Cold temperate</td>
<td>Seasonal snowmelt and glacier dominated</td>
<td>Qb = 164</td>
<td>0.0010</td>
<td>Stable</td>
<td>Variable</td>
<td>9.35 mean = 16</td>
<td>Aggrading 0.18</td>
<td>-</td>
<td>Gravel/coarse sand</td>
<td>Silt dominated</td>
</tr>
<tr>
<td>Mistaya</td>
<td>Smith and Smith (1980)</td>
<td>Cold temperate</td>
<td>Seasonal snowmelt and glacier dominated</td>
<td>Qb = 34</td>
<td>0.0039</td>
<td>Stable</td>
<td>Variable</td>
<td>8-41 mean = 15</td>
<td>Aggrading 0.06</td>
<td>Coarse</td>
<td>Gravel/sand</td>
<td>Silt dominated</td>
</tr>
<tr>
<td>Lower Attawapiskat River</td>
<td>King and Martini (1984)</td>
<td>Humid subarctic</td>
<td>Seasonal snowmelt dominated</td>
<td>Qmean = 508 Qmax = 3,115</td>
<td>0.00052</td>
<td>Stable</td>
<td>&lt;1.5</td>
<td>4-12 main channel</td>
<td>Degrading</td>
<td>Solution and suspended load dominant</td>
<td>Silts/sands</td>
<td>Silty clay</td>
</tr>
</tbody>
</table>
Table 1.1 (continued): Summary of published anastomosing river characteristics data.

<table>
<thead>
<tr>
<th>River</th>
<th>Source</th>
<th>Climate</th>
<th>Flow regime</th>
<th>Discharge (cumees)</th>
<th>Slope (m/m)</th>
<th>Lateral activity</th>
<th>Sinuosity Ratio</th>
<th>Sedimentation rate (m/100yr)</th>
<th>Sediment load</th>
<th>Bed material</th>
<th>Bank material</th>
</tr>
</thead>
<tbody>
<tr>
<td>Magdalen River, Colombia</td>
<td>Smith (1986)</td>
<td>Tropical savanna</td>
<td>Two high flow periods</td>
<td>Qb = 8,800</td>
<td>0.000095</td>
<td>Stable</td>
<td>Variable</td>
<td>Aggrading 0.38</td>
<td>10% bedload</td>
<td>Medium sand</td>
<td>Mud/fine sand</td>
</tr>
<tr>
<td>Cooper Creek</td>
<td>Nanson and Knighton (1996); This study</td>
<td>Arid/semi-arid</td>
<td>Seasonal, monsoonal</td>
<td>Qmean = 40-100</td>
<td>0.00015</td>
<td>Slow lateral migration</td>
<td>Variable 1.1 to 3</td>
<td>Aggrading 0.01max 0.004mean</td>
<td>Suspended load</td>
<td>Silt/clay</td>
<td>Silt/clay</td>
</tr>
<tr>
<td>Yangtze River</td>
<td>Baker (1986)</td>
<td>Humid temperate</td>
<td>Summer rainy season</td>
<td>Qmax &gt; 110,000</td>
<td>-</td>
<td>Slow lateral migration</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>Solimões River</td>
<td>Baker (1978)</td>
<td>Humid tropical</td>
<td>Constant but seasonal</td>
<td>3,500 (measured)</td>
<td>0.0001 to 0.00003</td>
<td>-</td>
<td>13</td>
<td>40-60</td>
<td>Solid load 30% of total load, bedload 5% of solid load</td>
<td>Gravel/coarse sand</td>
<td>Silt</td>
</tr>
<tr>
<td>Ovens River</td>
<td>Schumm et al. (1996)</td>
<td>Humid temperate</td>
<td>Winter flooding dominant</td>
<td>Qb = 70 - 280</td>
<td>0.00051 to 0.00088</td>
<td>Slow, transient migration</td>
<td>Variable 1.31-2.2</td>
<td>Slowly aggrading</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>Bani River</td>
<td>Makaske (1998)</td>
<td>Tropical semi-arid</td>
<td>Seasonal, monsoonal</td>
<td>Qmax = 5,300</td>
<td>0.0000345</td>
<td>Stable, point bars small or absent</td>
<td>1.0-1.2</td>
<td>40-60</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>Japura River</td>
<td>Baker (1978)</td>
<td>Humid tropical</td>
<td>Constant but seasonal</td>
<td>-</td>
<td>0.0001 to 0.00003</td>
<td>-</td>
<td>1.1</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
</tbody>
</table>
channels to gravel and coarse sand in rivers such as the North Saskatchewan and Alberta Rivers (Table 1.1). Data on sediment load is very patchy, however as one would expect bedload transport is more important in channels with coarse bed material. Red Creek and Cooper Creek which both have cohesive bed materials are suspended load dominated, while the gravel bedded North Saskatchewan River has a gravel dominated sediment load (Smith and Smith, 1980). The lower Attawapiskat River which has bed material of silt and sand has a sediment load dominated by solution and suspension (King and Martini, 1984). Channel cross-section form is influenced by sediment load and channels transporting predominantly suspended loads tend to be narrower and deeper than those with a bedload dominated sediment load (e.g. Schumm, 1960a, 1960b, 1961; Pickup, 1976). As well as sediment load, bank strength exerts a significant influence on channel geometry (Andrews, 1984; Knighton, 1987; Huang and Nanson, 1997; Huang and Nanson, 1998). For example, Huang and Nanson (1998) found that variations in bank strength along a channel may produce a three fold change in channel width, a two fold change in channel depth, and a 1.6 fold change in channel capacity. The data for anastomosing rivers, although limited, suggests that for these channels the ability of the channel to erode bank materials may be more important than the nature of sediment load in determining channel section form as channels with a dominant suspended load do not have consistently smaller width-depth ratios than bedload dominated channels. The suspended and solution load dominated lower Attawapiskat River has a width depth ratio varying between 30 and 140 and experiences frequent bank erosion by slumping and ice-flow abrasion (King and Martini, 1984), while the bedload dominated Alexandra and North Saskatchewan Rivers have mean ratios of 13 and 16 respectively and have a rip-rap like protection of their banks by dense root mats and show little signs of bank erosion (Smith and Smith, 1980).

1.3.2 Anabranch planform

Characteristic of many anastomosing rivers is the relative lateral stability and low sinuosity of individual anabranches (Smith, 1973; Smith and Smith, 1980; Smith, 1983; Nanson and Knighton, 1996). Many anastomosing anabranch planforms are said to be ‘straight’ (e.g. Smith and Smith, 1980; Smith, 1983), while others are noted as having low sinuosities (e.g. Baker, 1978; King and Martini, 1984). The observed active meandering of anabranches is said to be uncommon (the term meandering as it will be used here implies that active lateral migration is occurring), as is the occurrence of floodplain morphology indicative of migration (Smith, 1983), although a number of anastomosing channels do have laterally migrating anabranches.
such as the Yangtze River (Baker, 1986), the Solimões River (Baker, 1978), Red Creek (Schumann, 1989) and the Ovens and King Rivers (Schumm et al., 1996) (see Table 1.1).

Channels in nature have been said to fall along a continuum of variations in flow energy and boundary resistance. Leopold and Wolman (1957) considered the influence of flow properties as controlled by discharge and slope on the occurrence of channel patterns. Plots of channel pattern occurrence with these variables enabled Leopold and Wolman to define a line of discrimination between meandering and braided channels of the form

$$S = 0.06Q^{-0.44}$$

(3.1)

Where $S$ is slope and $Q$ is discharge. Channels above the line tended to be braided while those below the line tended to meander, and at the same slope, meandering channels tend to have lower discharges than braided channels (Leopold and Wolman, 1957). Subsequent work has extended this analysis to explicitly include the effects of variations in boundary material on erodibility and channel form (e.g. Henderson, 1963; Osterkamp, 1978; Carson, 1984, Ferguson, 1987). In non-cohesive materials, an increase in boundary sediment size tends to cause an increase in the slope - discharge threshold between patterns (Osterkamp, 1978). Treatment of the erodibility of cohesive and vegetated materials is more problematic, and the consideration of the influence of these factors on channel pattern has been largely qualitative.

The tendency toward straightness is uncommon in nature as is noted by Leopold and Wolman (1957) and Wilson (1973). Definitions of straight channels vary in their restrictiveness. Leopold and Wolman (1957) define a stream as straight if it has a sinuosity less than 1.5, as do Rust (1978), Begin (1981), and Chang (1988) while Moody-Stuart (1966) is more restrictive defining only streams with sinuosities lower than 1.3 as straight. Interestingly, Chang (1988) terms channels with a sinuosity less than 1.5 ‘straight or sinuous’ as opposed to merely ‘straight’, thereby removing the implication that planform is geometrically straight. Most commonly, channels are said to be straight because they lack the energy necessary for them to meander (e.g. Schumm and Khan, 1972). The justification for this is the fact that flow thalwegs tend to wander even when channel planform is not sinuous, and so channels which are straight are seen as being trapped inside restrictive bank materials they are incapable of eroding, their natural tendency to meander suppressed. In a flume experiment, Schumm and Khan (1972) found that with discharge held constant, a geometrically straight channel cut into the flume
remained straight only below a threshold slope. In a study of British rivers Ferguson (1981) found that laterally inactive straight or sinuous channels tended to have lower bankfull stream powers than active channels with median values of $15\text{Wm}^{-2}$ compared to $30\text{Wm}^{-2}$. Another explanation of 'straight' channel planform is dynamic rather than static, involving planform actively changing to become 'straight'. Vanoni et al. (1972) and Begin (1981a) state that 'straight' streams may develop higher shear stresses than streams which meander. These streams are intermediate in flow energy between meandering and braiding, and in this case 'straightness' results from the conservation of flow momentum which discourages turning (Begin, 1981a, 1981b). These channels develop large bend radii of curvatures to reduce momentum losses, and while 'straight' by definition in that they have sinuosities below 1.5 they are in fact actively meandering and show planforms with systematic, cyclic downstream variations.

The present distinction between meandering and straight channels by sinuosity alone is unacceptable and unrealistic. It implies a genetic difference between the two classes of channels - one actively meandering and laterally eroding and the other not - which does not necessarily exist. As an illustration of this Leopold and Wolman (1960) in their paper on river meandering included channels with sinuosities as low as 1.1. Makaske (1998) proposed a genetic definition of straight streams which defines all those which are laterally stable and not actively meandering as straight. Thus as Makaske (1998) states 'straight streams' may have high sinuosities and convoluted planforms as long as this is not due to an active meandering process. This is a logical non sequitur. The term straight has a particular geometric meaning and a stream should not logically be defined as straight when its high sinuosity may make it far from being so. A preferable classification was proposed by Nanson and Knighton (1996) who used the term meandering as a process based descriptive term and sinuous as a passive term for form with no implication that any sinuosity present is the produce of an active process. For descriptive purposes, stable channels may be divided according to sinuosity and defined as either 'laterally stable low-sinuosity' or 'laterally stable high-sinuosity' channels using the generally accepted division of 1.5 between 'straight' and 'meandering' channels. The term straight may also be retained and used in a very strict, geometric sense consistent with its meaning. Clarification of the existing terminology is certainly warranted because of its misleading implications and inadequacies.

The characteristic channel geometry, relative lateral stability and low sinuosity of anastomosing channels are related to channel properties. Slopes of anastomosing rivers are
generally very low. The greatest slope of any of the rivers for which there is data is 0.0011 m/m and slopes are generally much less. Low slopes cause the channel flow to have low energy, and data collated on anastomosing rivers by Nanson and Knighton (1996) showed that all rivers for which there was data had specific stream powers of 10 Wm$^{-2}$ or less. The low energy and resistant banks of the channels resist lateral erosion leading the planform of individual anabranches to be relatively stable. The anabranches of laterally dynamic anastomosing rivers tend to migrate slowly. Lateral migration of the anastomosed Yangtze River may be inferred from the occurrence of scroll bars (Baker, 1986), and scroll bars may be clearly seen on the anastomosing Solimões River in Brazil although radiocarbon dating of archaeological material found on one channel island showed the island to be at least 2050 years old and indicates that channel migration there is not rapid (Baker, 1986). The meandering of Red Creek in Wyoming (Schummann, 1989) and the Ovens and King Rivers in Australia (Schumm et al., 1996) was inferred from planform and channel geometry variations because the slow rates of channel change precluded changes from being directly observed or recorded in the historical record. While many definitions of anastomosing rivers allow individual anabranches to be braided, the occurrence of braided anabranches on anastomosing channels is very rare because of the low energy of the channels and Knighton and Nanson (1993) found that anastomosing channels tended to plot below both meandering and braiding channels on slope-discharge plots.

1.3.3 Floodplain morphology

A number of distinctive anastomosing floodplain environments have been recognised, and Smith’s (1983) inclusive channel and floodplain based definition of anastomosing was given earlier (Section 1.2.1). From work on rapidly aggrading anastomosed rivers in humid temperate areas of Canada Smith (1983) recognised six common sedimentary environments and facies; channel deposits of sand or fine gravel, crevasse splay deposits of sand or fine gravel, levee deposits of sandy silt, lacustrine deposits of mud, marsh deposits of mud and organic material, and bog deposits of peat. Coarse, aggrading bedload deposits are confined to channels and crevasse splays and levees border lower areas more distal to channels where floodwater tends to accumulate and pond, leading to suspended sediment deposition and the formation of lacustrine, marsh and bog deposits. The anastomosing islands formed on the Alexandra River have a saucer-like morphology that has been recognised on many anastomosing river floodplains (Baker, 1978; Smith and Putnam, 1980; Smith and Smith, 1980; Smith, 1983). The low central areas have been termed floodbasins (Makaske, 1998), and are sometimes connected to major anastomosing channels by channels at their downstream ends (Smith and Smith,
1980). Like Smith (1983), Makaske (1998) considers floodplain environments so characteristic of anastomosing rivers that they are central to his definition of an anastomosing river. Makaske (1998, p.27) defines an anastomosing river as one “composed of several interconnected channels, which enclose floodbasins.” He states that “...floodbasins are the unifying element between the multiple channels [of anastomosing rivers]. They play a crucial role in their genesis, being the place where crevassing and avulsion take place.” (Makaske, 1998, p.47)

The occurrence and prominence of floodbasins varies greatly between anastomosing rivers as Figure 1.3 shows. Floodbasins or floodponds are prominent on the Alexandra River, and generally not present on Red Creek where Schumann (1989, 283) states that “vertical accretion floodplains tend to slope away from the channels, giving the floodplain a convex shape when viewed in cross section”. On the Cooper Creek even though large anastomosing channels are generally bordered by levées (this has been contested as will be described in Chapter 3, but is confirmed by the results presented in Chapter 6) floodbasins are not a prominent feature of floodplain morphology. The term ‘floodbasin’ implies that the floodplain surface depression is three dimensional and not just apparent on a two dimensional floodplain cross section. Even though levees generally occur along anastomosing channels on Cooper Creek they are frequently dissected by floodplain-surface channels joining or splitting from the anastomosing channels, and an examination of air photographs following flood events shows none of the ponding of floodwaters that one would expect if such floodbasins were present. Also, as will be described in Chapters 3 and 7 the topography of the areas between anastomosing channels is not usually that of simple concave basins but instead the areas are undulating and covered by braid-like channels and islands. Given the patchy occurrence and distinctiveness of floodbasins and floodplain environments in anastomosing streams, one can conclude that floodbasins are not a necessary component of the anastomosing pattern and so should not be central to any definition of anastomosis.

Work on anastomosing rivers in arid and semi-arid zones has shown that some of the sedimentary conditions characteristic of humid anastomosing river floodplains are absent from more arid areas. Work on the Cooper Creek (Rust, 1981; Nanson et al., 1986) and Red Creek in Wyoming (Schumann, 1989) has shown that marsh, bog, and lacustrine deposits are generally not found in these environments due to their greater aridity. The difference in floodplain environments between anastomosing rivers led Nanson and Croke (1992) in their classification of floodplains to recognise two distinct types of anastomosing river floodplains, organic-rich and inorganic floodplains. Organic-rich floodplains occur in humid environments and a significant proportion of floodplain sediments consist of lacustrine deposits and organic
Figure 1.3: Floodplain profiles from (a) the anastomosing Alexandra River in Alberta, Canada, showing the prominent development of floodbasins (from Makaske, 1998, p.26), and (b) Red Creek in Wyoming showing marked floodplain convexity (from Schumann, 1989, p.283).
material often as peat (Nanson and Croke, 1992). In contrast, inorganic floodplains occur in more arid areas where accretion rates tend to be lower and lacustrine and organic deposits tend to be absent (Nanson and Croke, 1992). This is primarily a climatic distinction, and differences do not necessarily reflect a difference in the dynamics or genesis of the anastomosing pattern.

1.3.4 Anastomosing channel island planform

The plan morphology of anastomosing channel islands is one of the distinctive features of the anastomosing channel pattern, and anastomosing channel islands are distinctly different to bars in braided channels in a number of ways (e.g. Schumm, 1977; Yonechi and Maung, 1986; Nanson and Knighton, 1996; Makaske, 1998). Firstly, anastomosing channel islands are much more stable than islands in braided channels and exhibit lower rates of change. The stability of island form is directly related to the low lateral dynamism of the adjoining channels and is often expressed in the development of mature vegetation on islands. Bars in braided channels tend to be smaller in width than the adjacent channels (Trevena and Picard, 1978), and the channels defining braid bars tend to be curved segments of relatively simple planform in which flow patterns are determined by flow in adjacent anabranches (Bridge, 1993). In contrast anastomosing islands tend to be much wider than the channels which define them (Yonechi and Maung, 1986). As Schumm (1977) notes, individual anabranches of anastomosing rivers may have their own unique planforms and be braided, meandering or straight. This is due to the fact that flow patterns in anastomosed anabranches are independent of flow patterns in adjacent anabranches (Bridge, 1993). This enables anabranches to develop their own patterns and island form to be complex, as Figure 1.2 recognises. There have been relatively few detailed studies of braid-bar geometry. Rust (1972) reported that braid-bars 150m-450m in length on the braided Donjek River in Canada showed an average length/width ratio of 3.9, were elongated parallel to the flow direction, deviated from an elongate rhombic form pointed at both ends, and were commonly asymmetrical. A more detailed study of bar form was carried out by Rachocki (1981) on braid-bars between approximately 10cm and 6m from both the Skeidararsandur in Iceland and experimental fans. Rachocki found that the bars showed consistent length/width ratios of approximately 2.4 indicating that they tend to approximate scale invariant characteristic shapes (Figure 1.4). The tendency to adopt characteristic shapes has been found in a number of streamlined fluvial bar and island types and by analogy with geometric streamlined forms such as the lemniscate loop is thought to represent a tendency to minimise flow resistance (Baker and Milton, 1974; Baker and Kochel, 1978; Baker, 1979; Komar, 1983, 1984). Hills and bars in the Channeled Scabland have been naturally streamlined by catastrophic flood flows and $r^2$ values for relationships between length, width and area
Figure 1.4: Scatterplot of braid bar length and width measurements, from Rachocki (1981, p.67).

Figure 1.5: Scatterplot of anastomosing channel island length and width measurements, Upper Irrawaddy River, from Yonechi and Maung (1986, p.109).
measurements of the forms range between 0.87 and 0.94 (Baker, 1979). Komar (1984) found that islands in the Mississippi, Missouri, and Columbia Rivers also showed strong length/width relationships with an $r^2$ value of 0.96. In contrast to braid bars and natural streamlined forms, anastomosing channel islands do not show consistent length/width ratios and do not tend to approximate characteristic shapes, as Figure 1.5 of island form in the upper Irrawaddy River illustrates (Yonechi and Maung, 1986). This is due to greater irregularities in the planform of the channels defining the islands and the generally larger size of anastomosing islands. A related difference in the distal angles of anastomosing islands and braid bars was also noted by Yonechi and Maung (1986) who found that the distal angles of anastomosing islands on the Poronai and Irrawaddy Rivers were generally more variable than the distal angles of braid bars reported by Rachocki (1981). Interestingly there has been no study of island proximal angles, akin to the bifurcation angles, of the bordering channels, nor has there been any comparison between confluence and bifurcation angles on anastomosing rivers.

1.3.5 Flow regime

An examination of Table 1.1 shows that many anastomosing rivers are characterised by seasonal or highly irregular flow regimes due commonly to either monsoonal rainfall or snowmelt. Similarly, Makaske (1998) notes that humid climate anastomosing streams experience regular prolonged periods of annual flooding. The occurrence of large flood events forces large volumes of flow overbank and promotes the formation of new anabranches.

1.4 Anastomosing channel dynamics

1.4.1 Avulsion events

The anastomosing planform is dynamic despite the frequent lateral stability of the planform of individual anabranches, and the constancy of anastomosing channel position was overstated in early facies models of Smith and Putnam (1980) and Smith and Smith (1980) which proposed that ‘walls’ of channel deposits accumulated beneath a laterally stable, fixed network of vertically accreting channels (e.g. Figure 6, Smith and Smith [1980]). Further study of ancient anastomosed deposits has shown that channel persistence for long periods of aggradation does not occur and that the positions of the channels change (Smith, 1983; Törnqvist et al., 1993; Miall, 1996). The dynamism of anastomosing channel planform is due to the occurrence of avulsion events which are characteristic of anastomosis (e.g. Rust, 1981; Harwood and Brown, 1993; Schumann, 1989; McCarthy et al., 1992; Knighton and Nanson, 1993; Schumm et al.,
Some kind of avulsion process is necessary, in the absence of bar deposition by braiding processes or the within channel deposition of sediment ridges which occurs in some anabranching channels (e.g. Wende and Nanson, 1998), to account for the multiple channels and islands of the anastomosing pattern.

1.4.1.1 Definitions of avulsion

The Oxford English Dictionary (1989) defines avulsion in a number of ways: “1. The action of pulling off, plucking out, or tearing away; forcible separation...2. A part torn off, a detached portion...3. Law The sudden removal of land, by change in a river’s course or by the action of flood, to another persons estate...” (p.826). Common to all these usages is the connotations of violence (“tearing away”, “forcible separation”) or swiftness they carry. Geomorphic definitions of avulsion also stress the rapidity of the process, which is typically described as an ‘event’. Miller (1991, p.228) quotes the 1972 definition of the American Geological Institute, which defines an avulsion as “a sudden cutting off or separation of land by a flood or by an abrupt change in the course of a stream...whereby the stream channel diverts its old channel for a new course” (emphasis added). Slingerland and Smith (1998, p.435) state that an avulsion occurs when “a channel belt is abruptly abandoned for a new course at a lower elevation” (emphasis added), and the rapidity of avulsion is also central to the definitions of Allen (1965), Törnqvist (1994), and Makaske (1998).

While the definitions of avulsion above adequately captures the nature of some avulsion events, they oversimplify the process of channel change in many others. Rapidity and violence of change are not universal features of avulsions which in some cases may be very slow, gradual events with new channels developing incrementally. As Smith et al. (1997) note, our understanding of the mechanisms governing the location and timing of avulsions is imperfect and many models of channel avulsions treat them as random events. The factor enabling an avulsion is generally the existence of a gradient advantage for flow between channel and floodplain. During flood events, this gradient advantage causes a diversion of flow from the main channel onto the floodplain to have a greater slope than the established channel and so to capture its flow (e.g. Bridge and Leeder, 1979; Mackey and Bridge, 1995). Avulsion events can be divided into three stages:

- a pre-avulsion stage in which a new anabranch may be developing;
- an initial avulsion stage in which the new anabranch draws discharge from the main channel at a flow stage equal to bankfull flow or below;
- and a post-avulsion adjustment stage, in which the channel system adjusts to the incorporation of the new anabranch.
The development of the new anabranch may have been occurring for some time in the pre-avulsion stage before the avulsion occurs (e.g. Schumann, 1989; Schumm et al., 1996). The pre-avulsion stage, where it occurs, may vary in duration from no more than a couple of events to much longer periods of time, and during this phase the incipient anabranch is being developed during flood events which exceed bankfull discharge, however does not capture a significant proportion of in-channel flow below bankfull stage. The development of a new anabranch may also begin in the initial avulsion stage as in the case of the rapidly aggrading Yellow River (van Gelder et al., 1994), and avulsions of this type commonly occur in rapidly aggrading systems during particularly large flood events or are caused by channel blockages. When this occurs, the new channel is formed and channel flow captured during a single event. Following the avulsion there is a post-avulsion phase in which the multiple channels adjust following the development of the new anabranch (Smith et al., 1989; Törnqvist, 1994; Mackey and Bridge, 1995). There are three possibilities: the newly formed channel may capture all the discharge of the older channel almost instantaneously, channels may coexist; or the new anabranch may be abandoned. Adjustment of channels following avulsion may take a long period of time. Based on the dating of channel fill deposits from anastomosed Rhine-Meuse delta channels, Törnqvist (1994) found evidence of both ‘instantaneous’ abandonment (where channels coexisted for a short time below the resolving power of $^{14}$C dating) and gradual adjustment following avulsion. Whether or not channels coexist, adjustments to channel morphology are likely to persist for a significant length of time following an initial avulsion (e.g. Smith et al., 1989; Brizga and Finlayson, 1990; Smith and Pérez-Arlucea, 1994).

It will be argued in this thesis that the anastomosing channels of the Cooper change very slowly and that small floodplain-surface channels enlarge to become anastomosing channels over long periods spanning numerous events. The discordance of this process with the suddenness and violence implied by the term ‘avulsion’ necessitates the qualification of the term with the adjective gradual. ‘Gradual avulsion’ is proposed here to describe the extremely slow, incremental development of a new channel course and formation of an island of excised floodplain.

1.4.1.2 Causes of avulsion, direct and indirect

Avulsions may be said to have direct and indirect causes. Specific direct and indirect causes of avulsions which have been recognised are summarised in Table 1.2. Direct causes are those which trigger the formation or inception of a new anabranch and examples of direct causes from Table 1.2 are flood events and log jams. The occurrence of a direct cause does not always lead to an avulsion, for example floods do not always cause avulsions and log jams are often
washed out by high flows. Indirect causes are those which create the circumstances favourable for a trigger event to cause avulsion such as aggradation of the channel belt causing the channel to be perched high above the surrounding floodplain or factors favouring aggradation such as eustatic sea level rise (Törnqvist, 1994) or tectonic downwarping (Smith, 1986).

Table 1.2: Causes of avulsion events

<table>
<thead>
<tr>
<th>Cause</th>
<th>Source</th>
<th>Type of cause, direct or indirect</th>
<th>Situation of occurrence</th>
</tr>
</thead>
<tbody>
<tr>
<td>Flood event</td>
<td>Speight (1965b)</td>
<td>Direct</td>
<td>Meandering channel, Alluvial fan, Subaqueous turbidity fan channels</td>
</tr>
<tr>
<td></td>
<td>Gole and Chitale (1966)</td>
<td>Direct</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Weimer (1991)</td>
<td>Direct</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Van Gelder et al. (1994)</td>
<td>Direct</td>
<td></td>
</tr>
<tr>
<td>Vegetation blockages (log jams and within channel vegetation growth)</td>
<td>Taylor and Woodyer (1978)</td>
<td>Direct</td>
<td>Alluvial fan, Anastomosing channel</td>
</tr>
<tr>
<td></td>
<td>McCarthy et al. (1992)</td>
<td>Direct</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Harwood and Brown (1993)</td>
<td>Direct</td>
<td></td>
</tr>
<tr>
<td>Lateral migration into palaeochannel</td>
<td>Fisk (1952)</td>
<td>Direct</td>
<td>Meandering channel, Wandering gravel bed river</td>
</tr>
<tr>
<td></td>
<td>Ferguson and Werrity (1983)</td>
<td>Direct</td>
<td></td>
</tr>
<tr>
<td>Ice jams</td>
<td>King and Martini (1984)</td>
<td>Direct</td>
<td>Anastomosing channels</td>
</tr>
<tr>
<td></td>
<td>Smith et al. (1989)</td>
<td>Direct</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Smith and Pérez-Arlucea (1994)</td>
<td>Direct</td>
<td></td>
</tr>
<tr>
<td>Gradual tectonic activity</td>
<td>Smith (1986)</td>
<td>Indirect</td>
<td>Anastomosing river, Drainage network, Alluvial fan</td>
</tr>
<tr>
<td></td>
<td>McDougall (1989)</td>
<td>Indirect</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Smith et al. (1997)</td>
<td>Indirect</td>
<td></td>
</tr>
<tr>
<td>Hippopotamus trails</td>
<td>McCarthy (1992)</td>
<td>Direct</td>
<td>Swamp</td>
</tr>
<tr>
<td>Mass movement levee or bank failure</td>
<td>Weimer (1991)</td>
<td>Direct</td>
<td>Subaqueous turbidity fan channels</td>
</tr>
<tr>
<td>Earthquake induced bank failure</td>
<td>Alexander and Leeder (1987)</td>
<td>Direct</td>
<td></td>
</tr>
<tr>
<td>Eustatic sea level rise</td>
<td>Törnqvist et al. (1993)</td>
<td>Indirect</td>
<td>Delta</td>
</tr>
<tr>
<td>Channel belt aggradation</td>
<td>Speight (1965)</td>
<td>Indirect</td>
<td>Meandering channels, Braided channels</td>
</tr>
<tr>
<td></td>
<td>Coleman (1969)</td>
<td>Indirect</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Brizga and Finlayson (1990)</td>
<td>Indirect</td>
<td></td>
</tr>
<tr>
<td></td>
<td>van Gelder et al. (1994)</td>
<td>Indirect</td>
<td></td>
</tr>
<tr>
<td>Floodwater scouring</td>
<td>Popov (1962)</td>
<td>Direct</td>
<td>Meandering River, Distributary fan</td>
</tr>
<tr>
<td></td>
<td>Riley (1975)</td>
<td>Direct</td>
<td></td>
</tr>
<tr>
<td>Within channel aggradation</td>
<td>Taylor and Woodyer (1978)</td>
<td>Indirect</td>
<td>Anastomosing rivers, Alluvial fan</td>
</tr>
<tr>
<td></td>
<td>Schumann (1989)</td>
<td>Indirect</td>
<td></td>
</tr>
<tr>
<td></td>
<td>McCarthy et al. (1992)</td>
<td>Indirect</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Schumann et al. (1996)</td>
<td>Indirect</td>
<td></td>
</tr>
<tr>
<td>Increasing channel sinuosity</td>
<td>Schumann (1989)</td>
<td>Indirect</td>
<td>Anastomosing channels</td>
</tr>
<tr>
<td></td>
<td>Schumann et al. (1996)</td>
<td>Indirect</td>
<td></td>
</tr>
</tbody>
</table>

Table 1.2 is not exhaustive, but does capture some of the great variance in causes of avulsion events. It is notable that some avulsions have a number of causes and that the distinction between direct and indirect causes is not always precise or simple. For example, the direct cause of the 1954 avulsion of the meandering Anabunga River was flood events, while the indirect cause was aggradation of the channel belt which enabled a crevasse channel breaching the levee to form (Speight, 1965b). Abandonment of the channel here was relatively rapid with
anabranch coexistence lasting for only between 3 and 8 years before the old channel was completely abandoned (Speight, 1965b). Avulsion mechanisms in anastomosing channels

A number of different avulsion mechanisms have been found to operate in anastomosing channels. These are described here.

1.4.1.3 Crevasse avulsion

By far the most studied mechanism of channel avulsion is undoubtedly the crevasse avulsion (Lewin and Hughes, 1980; Smith et al., 1989; van Gelder et al., 1994; Slingerland and Smith, 1998). Crevasse channels, so named because of their form, are cut into levees by overbank flows down the relatively steep slopes often found perpendicular to the channel. The steeper the slope the more energy overbank flow has and so the greater the likelihood a crevasse channel will develop. For this reason, crevassing is generally restricted to channels with prominent levees. A good example of a river which experiences regular crevasse avulsion events is the lower Yellow River near its delta (van Gelder et al., 1994). The aggradation rate of the lower Yellow River is remarkable. Suspended sediment load may reach 222kg/cumec, overbank deposits from individual flood events may be metres thick, and the channel itself aggrades so rapidly that its bed may be perched several metres above the surrounding floodplain (van Gelder et al., 1994). These conditions are ideal for crevasse avulsions which occur in anastomosing rivers that develop prominent levee and floodbasin topography (Makaske, 1998). These tend to occur where there are high aggradation rates, such as on the Columbia, Mistaya, North Saskatchewan, Alexandra, and lower Saskatchewan Rivers (Smith and Smith, 1980; Smith, 1983). Although high aggradation rates are not common to all anastomosing rivers some authors have claimed that crevassing is the characteristic avulsion mechanism of all anastomosing channels (Miall, 1996; Makaske, 1998).

The evolution of a crevasse channel will depend on the volume of sediment supplied to the crevasse channel relative to its sediment transporting capacity, and the crevasse may heal up and become blocked, remain the same, or enlarge (Slingerland and Smith, 1998). Initially when a breach or crevasse is small the water flowing through it is drawn from high in the main channel flow and will contain low concentrations of suspended sediment and the sediment transport capacity of the flow will often exceed supply causing the crevasse to enlarge (Slingerland and Smith, 1998). As the crevasse becomes deeper and captures more flow from the main channel it receives flow extending deeper down the water column which contains a greater concentration of suspended sediment increasing the volume of sediment supplied to the
crevasse channel. Enlargement of the crevasse will continue until either the sediment flux from
the main channel comes to equal the sediment transport capacity of the crevasse channel, or the
crevasse capacity is reduced by decreasing water surface slope (Slingerland and Smith, 1998).
Smith and Putnam (1980) note that the formation of a crevasse channel does not necessarily
mean an avulsion will occur. They state that on low gradient floodplains avulsion channel
development will only occur if flow can re-enter the established channel because only then will
the flow have sufficient energy to scour a new continuous channel. Where this is not the case
floodwaters will tend to pond, reducing the water surface slope of the crevasse and causing
aggradation to block the crevasse channel (Smith and Putnam, 1980). After an avulsion, a
channel which is abandoned will begin to fill with sediment and contract in size. Such a
channel will tend to be laterally inactive due to lack of flow energy, and so may develop an
underfit stream morphology (Brizga and Finlayson, 1990; Richards et al., 1993).

In a crevasse avulsion studied by Smith et al. (1989) which occurred on the Saskatchewan
River in the Cumberland Marshes in 1873, the formation of an anastomosed pattern was a
transitional stage of adjustment following avulsion. The reach where the avulsion occurred is
an unstable meandering river and an ice jam caused the river to breach its levee bank and flow
into an adjacent low area of floodplain (Smith et al., 1989). Four stages of channel evolution
following the avulsion event were recognised: (i) an ‘initial avulsion stage’ in which new
channels and crevasse splays on the invaded floodplain increased in number rapidly, (ii) an
‘anastomosed stage’ in which new channels began to form an integrated network and crevasse
splay development was accompanied almost equally by crevasse splay abandonment, (iii) a
’reversion stage’ which began when deposition had extended across the entire floodplain and
saw the concentration of flow into a smaller number of channels due to an increased rate of
channel abandonment, and (iv) the eventual construction of a single channel (Smith et al.,
1989). This mechanism encapsulates the characterisation of anastomosing as a transitional
channel pattern dependent upon high rates of aggradation which is based on the assumption
that extremely high rates of aggradation cannot persist for sustained periods of geologic time.
However, Table 1.1 shows that not all anastomosing systems are rapidly aggrading and while
in some cases anastomosing channels may be transient features there appears to be no reason to
regard the pattern generally as being transient. As Knighton and Nanson (1993) point out, some
anastomosing channels systems exhibit great longevity and the Cooper Creek seems to have
persisted as an anastomosed system for at least 50,000 years.

1.4.1.4 Floodplain scour

Popov (1962) and Riley (1975) both proposed that anabranch development can take place
without aggradation and crevassing. Popov (1962) states that during high floods anabranches may be scoured into lines of weakness on the floodplain surface, or where overbank flow depths and velocities are high. Riley (1975) proposed a similar mechanism for the Namoi-Gwydir distributary system, which is similar to an anastomosing channel system in slope, channel sectional form, sediment load and flow energy, the principal difference being that, because it is a large alluvial fan, channel recombination is rare. Riley (1975) noted the absence of levees in the Namoi-Gwydir system, and inferred that levee crevassing was not important in producing anabranches there. Like many semi-arid or arid zone streams, the Namoi-Gwydir system is characterised by an highly irregular precipitation and flow regime, and experiences very large flood events. As Riley (1975) pointed out, by this mechanism avulsion events may occur even if the existing channels are in equilibrium with imposed sediment transport conditions. In the low gradient, low energy areas characteristic of anastomosing streams the scour of a new anabranch of sufficient size and continuity to capture significant amounts of below bankfull flows would most likely take a great deal of time and a large number of flood events, and so in the terminology proposed here is likely to be a gradual avulsion event.

1.4.1.5 Upstream headcut migration

Schumann (1989) in his study of Red Creek, an arid-zone stream in Wyoming, recognised a gradual avulsion mechanism involving a headcutting process. Red Creek is a small, low gradient, suspended load stream with a seasonal flow regime dominated by spring snowmelt runoff (Schumann, 1989). From space-time substitution based on observations of anabranches of different inferred ages and study of channel sediments, Schumann proposed an avulsion mechanism which is initiated by flow from the main channel being diverted overbank during a flood event, generally on the outside of meander bends due to the super-elevation of the water surface that occurs there (Schumann, 1989). Flow energy is dissipated as flow invades the floodplain and spreads out, and so no channel is immediately formed, however after some distance, channelisation occurs as a very wide, shallow channel and when this channel intersects with the main channel the steep drop in gradient causes a headcut to form at its downstream end. This headcut migrates back up the wide shallow channel enlarging it and progressing back to intersect the main channel causing an avulsion event and creating a new anabranch. Schumann found that the meandering and increase in anabranch sinuosity over time caused a lowering of channel slope and flow energy and an increase in flow resistance which triggered the deposition of suspended sediment within the anabranches decreasing channel capacity and driving increasing amounts of floodwater overbank (Schumann, 1989). The importance of increased sinuosity in causing avulsion and anastomosis has also been noted by Smith (1983) and Schumm et al. (1996). Schumann states that anabranch enlargement occurs
concurrently with the reduction in the capacity of other anabranches, and notes that larger anabranches of Red Creek tend to be more sinuous than smaller anabranches due to having more energy for channel migration and meander enlargement.

1.4.1.6 *Upstream and downstream headcut migration*

Schumm et al. (1996) proposed a model similar in a number of respects to that of Schumann (1989) for the Ovens and Kings Rivers in Australia, low gradient, slowly laterally migrating anastomosing rivers transporting a small sediment load with a coarse sand and gravel bedload (Schumm et al., 1996). They found that overbank flow concentrates into relatively straight floodplain depressions into which channels are scoured and that, as on Red Creek, a headcut incised by floodwaters proceeded up the depression from its downstream intersection with the main channel. However Schumm et al. also found that an incision also proceeded down the depression from the place where flood flows went overbank and that it was the meeting of these incisions which caused the avulsion event and the formation of a new anabranch. The gradient advantage between the old and new channels is created by anabranch sinuosity increasing with age which also causes the deposition of sediment within the channel and a reduction in channel capacity. However, unlike Red Creek much of the deposited sediment is material the river is transporting as bedload. Aggradation of the channel belt was not cited as a contributory factor to avulsion occurrence here. Once anabranches have achieved high sinuosities they do not have the energy to continue eroding and form cutoffs which would shorten the channel, lower sinuosity and increase channel slope and flow energy. Rather, the loss in energy makes them incompetent forcing increasing proportions of floodwaters overbank and leading to headcut formation (Schumm et al., 1996). The infilling anabranches may adopt underfit stream morphologies.

1.4.1.7 *Vegetation obstruction*

Both McCarthy et al. (1992) and Harwood and Brown (1993) studied anastomosed channels whose dynamics are heavily influenced by vegetation. McCarthy et al. (1992) studied a distributary channel of the Okavango River in Botswana which flows into a very extensive swamp. Channel boundaries here are highly permeable and composed of vegetation, and the channels transport a sand bedload. Avulsions here are stimulated by aggradation of the channel bed which causes a rise in channel water level and increases the rate of water loss through the channel banks. This further accelerates channel aggradation causing the channels to become shallower, and as flow velocity declines and bedload transport decreases the channel bed is colonised by bottom growing plants which impede flow and trap sediment and floating plant.
debris causing vegetation blockages (McCarthy et al., 1992). The blockages force water into the surrounding swamp areas and cause new channels to develop along pre-existing lines of weakness, such as hippopotamus trails, which tend to accumulate more water than the surrounding swamp. After the avulsion has occurred, the swamp around the older channel is deprived of water and is ultimately destroyed by fire leaving the elevated surface caused by within-channel accretion (McCarthy et al., 1992).

Harwood and Brown (1993) studied the anastomosed Gearagh in Ireland. This river flows through a heavily forested area in which a number of vegetation influenced fluvial processes operate. Avulsions are generated by debris dams, which form due to the dense vegetation present. Root masses cause a concentration of overbank flow and scour, and tree throw creates easily erodible, unvegetated depressions which are scoured by flood flow. All of these factors promote the development by avulsion of new channels and enable the maintenance of an anastomosed channel system (Harwood and Brown, 1993). Channel changes here seem to be relatively slow and probably occur by gradual avulsion rather than avulsion.

1.4.1.8 Cooper Creek

The anastomosing channels of the Cooper Creek have been studied by Rundle (1977), Rust (1981), Nanson et al. (1986), Rust and Nanson (1986), Knighton and Nanson (1994a), and Knighton and Nanson (1994b). Rust (1981) noted that headcuts, knickpoints and crevasse splays are generally not found on the Cooper Creek and suggested that avulsion occurs here without the formation of crevasse splays. Rust interpreted poorly defined anastomosed channels as being in the process of being abandoned and excluded the possibility of their being in the process of formation. This would seem to rule out a gradual process of channel formation, and implies that avulsion on the Cooper is a catastrophic event which takes place only in extreme flood events, however this suggestion is problematic because large flood events over the period covered by aerial photograph records have produced no apparent change to the anastomosing channels. The largest flood on record for the Cooper Creek was the 1974 flood which was by all accounts a huge event (e.g. Baker, 1986; Kotwicki, 1986), and a comparison of channel form pre- and post-flood from aerial photographs shows that no discernible change in channel form was caused by the event This suggests that an alternative "avulsion" or channel dynamic mechanism must operate in these channels and that a gradual avulsion process may occur here. The formative processes of the Cooper channels will be considered in more detail in subsequent chapters and findings summarised in Chapter 8.
1.4.2 Avulsion and anastomosis – autogenic and exogenic

The occurrence of avulsion or gradual avulsion is central to all the mechanisms of anastomosing channel formation outlined above. The excision of floodplain islands and formation of new channels is the characteristic dynamic of anastomosing rivers and the occurrence of avulsion or gradual avulsion is a prerequisite for anastomosing channel occurrence. As one would expect, the characteristics of anastomosing channels are generally those which promote or at least facilitate the occurrence of avulsions. There are two classes of avulsion or gradual avulsion which promote anastomosing and which will just be referred to as avulsions for simplicity here; *autogenic avulsions* in which avulsion may be caused by the flow and sediment dynamics of the channel itself, and *exogenic avulsions* which are caused by channel blockages due to factors other than the water and sediment regime of the channel.

The occurrence of autogenic avulsions is promoted by factors which are common to many anastomosing rivers; resistant channel banks and relatively low flow energy, high aggradation rates, and a flood dominated flow regime. High aggradation rates tend to create a strong gradient advantage between the channel and floodplain as occurs on the Yellow River, a flood dominated flow regime ensures large volumes of flow are displaced overbank which enhances the ability of the river to form a new anabranch, and resistant channel banks and low flow energy mean that the river cannot respond to high discharges by easily increasing channel size which again ensures that large amounts of flow are displaced overbank. A number of autogenic avulsion mechanisms occurring in anastomosing channels have been described earlier - crevasse avulsion, headcut migration à la Schumann (1989) and Schumm et al. (1996) and floodplain scour.

Factors which have been shown to cause exogenic avulsions are channel blockages caused by ice jams (King and Martini, 1984) and vegetation (Harwood and Brown, 1993). Channel blockages of this nature can be considered as disturbances. This type of anastomosis may be defined as ‘exogenic anastomosis’ because it is caused by the occurrence of exogenic avulsions, and may be seen as a transitional channel pattern which will persist while the disturbances causing avulsion events persist. Above the anastomosing reach of the lower Attawapiskat River the river exhibits a meandering pattern (King and Martini, 1984), and in the absence of the ice jams which occur there the lower Attawapiskat would most likely revert to a meandering pattern. Harwood and Brown (1993) cite the continuing occurrence of vegetation channel jams as necessary for the maintenance of the anastomosed pattern of the Gearagh.
The persistence of multiple channels produced by avulsion events is dependent upon both avulsion frequency and the persistence of anabranches. From a study of the stratigraphic facies of the Rhine-Meuse delta, Törnqvist et al. (1993) found that avulsion frequency controlled channel pattern here during the Holocene. Anastomosing periods were associated with a rate of groundwater level rise exceeding 1.5mm/yr caused by eustatic sea level rise and the anastomosing facies were accompanied by numerous crevasse splay deposits indicating a high frequency of avulsion (Törnqvist et al., 1993). Less rapid rates of sea level rise (1.0 - 1.5mm/yr) caused less frequent avulsions, fewer crevasse splay deposits and facies representative of a pattern transitional between anastomosing and meandering (Törnqvist et al., 1993). This suggests that frequent avulsions are needed to maintain an anastomosing pattern as has been suggested by Makaske (1998), however this is not necessarily the case. ‘Frequent’ is an imprecise and relative term, but however it is defined, if anabranches produced by infrequent avulsions persist for long periods of time then the avulsion frequency needed to maintain a multiple channel system may be very low. Further, it would seem that this is likely to be the case in the anastomosing systems in which the avulsion process is incremental and more like gradual avulsion such as Red Creek (Schumann, 1989), the Ovens and King Rivers (Schumm et al., 1996) and perhaps the Cooper Creek. Channels are likely to persist for long periods for the same reasons that it takes them many flood events to form; low rates of aggradation and high levels of inertia.

As will become clear later in this thesis, the concept of gradual avulsion is of particular importance to the operation of the Cooper Creek where to describe changes merely as ‘avulsions’ would be extremely misleading.

1.5 The diversity of anastomosis

It is clear from the discussion above that anastomosing rivers are a diverse group and not everywhere the same. While some authors have attempted to remove the need to explain differences by declaring some anastomosing channels as “anomalous” as Miall (1996, p.52) states regarding arid-region anastomosing streams like the Cooper Creek, other authors have proposed different classifications of anastomosing rivers which recognise that different types of anastomosing river exist (Nanson and Knighton, 1996; Schumm et al., 1996).

Schumm et al. (1996) distinguish between ‘Canadian’ and ‘Australian’ types of anabranching channel although any differences between anastomosing channels are certain not to be due to
nationality alone! Schumm et al. (1996), based on work by Smith and Smith (1980) state that the ‘Canadian’ type anastomosing river is relatively straight and stable, is characterised by rapid vertical deposition in channels and on floodplains, and does not experience much channel avulsion. In fact as stated previously it is now recognised that avulsions are much more common on the low gradient anastomosing streams in Canadian montane valleys than was first thought (Törnqvist et al., 1993; Miall, 1996). Aggradation of fine sediment on the floodplain does not keep pace with the aggradation of coarse material in the channels and so avulsions occur. Schumm et al. describe ‘Australian’ anastomosing channels as being of high sinuosity and laterally active with slowly accreting floodplains and characterised by channel avulsions (Schumm et al., 1996), however this characterisation is problematic because Australian anastomosing streams are not as uniform as Schumm et al. suggest. Schumm et al. based their characterisation on the Ovens and King Rivers which are quite unlike other Australian anastomosing rivers such as Magela Creek which is aggrading relatively rapidly and the anabranches of which do not laterally migrate and have low sinuosities of approximately 1.1 (Nanson et al. 1993; Table 1.1). Also, other anastomosing rivers such as the Cooper Creek and the other anastomosing Channel Country rivers cover an area of the continent dwarfing that represented by the Ovens and King Rivers and seem to be unlike the Ovens and King Rivers in a number of ways including bed material characteristics and aggradation rate which is much lower on the Cooper Creek. Because of the diversity of anastomosing rivers within countries it is not particularly useful to divide them on the basis of nationality.

Nanson and Knighton (1996) recognise three classes of anastomosing river, Type 1a organic systems such as the Okavango Delta, Type 1b organo-clastic systems producing stratigraphic deposits composed of both clastic and organic sediments like the Upper Columbia River, the Mistaya River, the Alexandra River, and Type 1c mud dominated systems like Red Creek and Cooper Creek. Nanson and Knighton (1996) classify the Solimões River as a Type 3 mixed load, laterally active anabranching system which they acknowledge are usually similar to Type 1b anastomosing channels but distinguished by their greater lateral dynamism. This is consistent with the restriction of anastomosing to low-energy channels, although there does not seem to be any necessary differences in either formative process or morphology between Type 1 and Type 3 rivers and precisely how rapidly a much a river needs to migrate before it is no longer an anastomosing channel is a moot point. Because of this no distinction between Type 1 and Type 3 channels is made here and both are considered to be anastomosing. The rate and occurrence of lateral migration are nonetheless important characteristics of anastomosing channels and three types of behaviour may be recognised. Anastomosing channels like Magela
Creek and the lower Saskatchewan, Alexandra and Mistaya Rivers are laterally stable, on the Ovens and King Rivers and Red Creek migration is transient and occurs only at certain stages of anabranch evolution when channels are low in sinuosity, and lateral migration occurs continuously on the Yangtze and Solimões Rivers.

The previous analysis of avulsion mechanisms divided anastomosis into autogenic and exogenic anastomosis based on the type of avulsion mechanism causing anastomosis. This distinction between autogenic and exogenic anastomosis is a useful addition to Nanson and Knighton’s classification of anastomosing channels in Type 1a, b and c systems based on morphology because it recognises an important genetic difference in formative mechanism between anastomosing channels not recognised by previous classifications but central to the explanation of why anastomosing channels form. Also, recognising that exogenic anastomosing channels are truly anastomosing systems acknowledges the significance of what have previously often been dismissed as exceptional cases.

1.6 Higher-order explanations of anastomosing channel form

There have been two orders of cause proposed to explain the development of the anastomosing pattern; first order explanations which consider the direct physical cause of anastomosis development, and second order explanations which attempt to address the function of anastomosing channels and explain anastomosis as an equilibrium or quasi-equilibrium channel pattern. Avulsion or gradual avulsion processes are the general first order cause of channel anastomosis and, while most treatments of anastomosis have not attempted to go beyond an analysis of first order cause, both sediment transport efficiency (Baker, 1986; Nanson and Knighton, 1996; Nanson and Huang, 1999) and sediment storage (Schumann, 1989; Makaske, 1998) have been proposed as second order explanations of anastomosis development.

Baker (1986) states that the anastomosed patterns of rivers such as the Yangtze River may be explained by the fact that very large rivers with high suspended loads must often split their channels to maintain the efficient transport of water and sediment. This is because large discharges of water and suspended sediment are best accommodated in narrow, deep channels, and although width can easily increase with river size, depth is often limited by the depth of the alluvial valley fill and cannot (Baker, 1986). To compensate for this, rather than forming a single large channel of optimal geometry the river must split into a number of separate channels which each individually approach optimal geometries (Baker, 1986).
A similar but more inclusive argument has been advanced by Nanson and Knighton (1996) to explain anabranching generally. Nanson and Knighton (1996) state that the “fundamental advantage of an anabranching river is that, by constructing a semi-permanent system of multiple channels, it can concentrate stream flow and maximise bed-sediment transport (work per unit area of the bed) under conditions where there is little or no opportunity to increase gradient.” (p.217) This argument is developed further by Nanson and Huang (1999) who from an analysis of basic hydraulic relationships for alluvial channels demonstrate that conversion of a wide, shallow single channel to a number of narrower and deeper multiple channels will, with constant total discharge, reduce total width and increase average flow depth, hydraulic radius, and flow velocity causing a potential increase in bedload transport. Nanson and Huang (1999) state that anabranching rivers seem to be “closer to exhibiting the most efficient sections for the conveyance of both water and sediment than are equivalent wide, shallow channels at the same slope” (p.477), and that the varying first order causes identified earlier seem to be “subsidiary to the primary requirement to maintain water and sediment throughput without recourse to increasing gradient but under conditions where channel width/depth ratios can be reduced due to the formation of resistant banks.” (p.489) A number of aspects of Nanson and Huang’s (1999) analysis bear further examination. Firstly, Nanson and Huang consider the change from a wide single channel to a number of narrower, deeper channels with equivalent discharge. However, the conversion to a number of smaller channels is not necessarily an optimal outcome. While it is more efficient than the initially wide, shallow channel a more efficient single channel can be constructed by making channel depth equal to the depth of any anabranch and width sufficient to contain channel discharge. This single channel will have a smaller wetted perimeter and larger hydraulic radius than the aggregate for the multiple channels and be more efficient. However, there may be good physical reasons why such a channel cannot be formed such as that bank strength may be inadequate to support a very deep channel, or as in the case of the Yangtze River and other large rivers the depth of valley fill may be insufficient to allow a single, large channel to form. Secondly, maximising bedload transport would seem to be desirable for some but not all autogenic anastomosing systems, and autogenic anastomosing rivers would seem to be divisible into bedload and suspended load systems. For example, the Upper Columbia, lower Saskatchewan, Alexandra and Mistaya Rivers all have coarse bed material and transport significant amounts of bedload (Smith, 1983; Smith and Smith, 1980), while in contrast bedload is not a significant component of sediment load in Red Creek which is suspended load dominated (Schumann, 1989).

Makaske (1998) studied an anastomosing reach of the upper Columbia River and found that
Despite the presence of multiple channels, most channels were shallow and inefficient and did not contribute greatly to the total discharge of water and sediment by the system. Indeed, Makaske stated that the upper Columbia River could practically be considered a single channel as far as water and sediment discharge are concerned and concluded from this that anastomosis is not an attempt by the system to maximise water and sediment throughput as Nanson and Huang (1999) suggest. Rather, Makaske suggests that anastomosis in a result of bedload transport inefficiency in a system which cannot laterally migrate and store coarse bedload sediments in point bar deposits. Anabranches act as stores for coarse bedload sediments which cannot be transported and must be stored in the channels. This causes a decrease in anabranch capacity and promotes the occurrence of avulsion events and anastomosis. Makaske (1998) notes that this model is similar to that of Schumm et al. (1996) described earlier in that channel capacity reduction drives avulsion occurrence and anastomosis development. As a general explanation of anastomosis, this model is subject to one of the criticisms applied to the model of Nanson and Huang (1999) in that not all anastomosing rivers transport significant amounts of bedload. Also anabranches of some are laterally active and so can store bedload in lateral accretion deposits. Schumann (1989) states that anabranches of Red Creek are sources of floodplain sediments deposited during flood events which is reflected in the convex profile of the floodplain surface. Schumann briefly notes that the dispersion of anabranches across the floodplain enables a widespread distribution of sediment over the floodplain as each anabranch builds its own quasi-independent vertical accretion floodplain.

It is clear from the earlier discussion of avulsion mechanisms that there are myriad first order causes of anastomosis, and it may well be that there is no general second order cause of anastomosis. While difficult to test, the application of the above hypotheses to the Cooper Creek anastomosing network will be considered.

1.7 Summary

In conclusion, anastomosing rivers are very diverse in morphology and genesis, more so than many authors have recognised. Despite this diversity, for reasons outlined earlier anastomosing channels tend to share some similar characteristics and to be characterised by resistant banks, low slopes, low flow energies, and generally low anabranch width-depth ratios. The definition of anastomosing rivers proposed by Makaske (1998) is too prescriptive and precise and as a consequence is too proscriptive excluding from the anastomosing class a number of rivers best classified as anastomosing. A more liberal definition is proposed here by which an
anastomosing river is defined as a low energy multiple channel system with resistant banks in which stable or slowly migrating anabranches are separated by irregular islands of approximately floodplain height that are large with respect to the width of the channels.
2 METHODS

This chapter will review the methodological aspects of this study that warrant description. Firstly the statistical methods used will be described followed by the sediment size analysis method and dating techniques used. The measures of channel cross section and channel plan morphology utilised are outlined, with particular attention paid to the fractal method that is used here to analyse single channel planform.

2.1 Statistics

A variety of statistical measures were used in this study, most of which are widely known and need no explanation. Analysis of variance (ANOVA) was carried out to determine whether differences between data sets were significant. ANOVA analysis is based on the assumption that the samples are independent, normally distributed, and have exactly equal variances (Lindman, 1992). While real data never satisfy these assumptions there are safe limits within which test results are valid and data sets were examined to ensure they were within safe margins. The assumption of sample independence is most important, and if undermined will compromise the validity of results (Lindman, 1992). The Tukey-Kramer HSD (honestly significant difference) test was used in a number of cases to explore the significance of differences between individual groups. This is a means comparison test which compares the actual difference between group means with the difference that would be statistically significant and if more than two groups are compared indicates precisely which sample means are statistically different (SAS Institute, 1994). It is more rigorous than many other tests such as the students t test because it protects the significance tests of all combinations of within sample pairs, and least significant differences measured by this test are larger than those calculated by the student's t (SAS Institute, 1994). Correlations reported used Pearson's Correlation Coefficient which measures the linear association between variables.

Descriptive measures of sample distributions were used to evaluate both the spread and form of distributions. Most measures used such as the mean, median, and skewness are straightforward. Also considered here was the kurtosis of the distribution which is a useful measure of distribution form that is often ignored (e.g. Hopkins and Weeks, 1990). Kurtosis is measured relative to the shape of a normal distribution, and is often considered to be a measure of the strength of the central peak of a distribution, akin here to the degree of dominance of a characteristic value or range of values. The kurtosis value of a distribution is influenced by the
size of the tails as well as the size of the central peak and positive kurtosis values indicate that, compared to a normal distribution, there is an excess of values near the mean as well as far away from it, and a depletion of values on the flanks of the distribution (Kotz and Johnson, 1982). Distributions with positive kurtosis values are known as "peaked" or "leptokurtic" distributions. Negative kurtosis values indicate distributions with less prominent peaks and heavier tails and are known as "platykurtic". An extreme example of a platykurtic distribution is a uniform distribution which has no peak. The standard deviation is the most commonly used indicator of the spread of data, however it is not easily comparable between data sets with differing magnitudes of values. For this reason the coefficient of variation (CV) which expresses the standard deviation as a percentage of the mean was used here as a measure of data spread. Small CV values indicate that the data is closely grouped around the mean value.

2.2 Sediment size analysis

This was carried out using a Malvern laser sediment sizer with a measurement range up to particles of 1000µm diameter, and pre-analysis sieving determined that no particles larger than this size were present. The particle size analyser calculates the volume of a particle and represents the size of the particle as the diameter of a sphere of equivalent volume. Sediment was ultrasonically dispersed before analysis and the 'ultimate' particle size distribution was measured rather than the 'effective' particle size distribution (Walling and Kane, 1984). Sediment aggregation may have a significant impact on sediment behaviour and Nanson et al. (1986), Rust and Nanson (1989), Maroulos (1992) and Maroulos and Nanson (1996) have all considered the process implications of sediment aggregation on the Cooper Creek floodplains. While the effective particle size distribution may be the more geomorphically relevant measure, the ultimate size distribution is the most practical to obtain and most comparable (Reid and Frostick, 1987).

2.3 Dating techniques

Two luminescence dating techniques, thermoluminescence (TL) and optically stimulated luminescence (OSL) were used to provide information on floodplain dynamics and channel behaviour. Particular aims were to investigate differences between floodplain accretion rates in areas of different channel types, and to assess the contemporaneity of levee deposits. Luminescence dating was the most suitable method of dating the floodplain sediments here, as both the rarity of dateable organic carbon in floodplain sediments, especially at depth, and the
possibility that samples would exceed the 30-40ka age limit of radiocarbon dating (Callen et al., 1983) made that technique unsuitable. Luminescence dating was developed to provide a method of dating archaeological artifacts like pottery and burnt flints (Aitken, 1985) and has since been applied to date a range of sedimentary features including aeolian dunes and loess deposits (e.g. Janotta et al., 1997), lunettes (e.g. Nanson et al., 1992; Dutkiewicz and Prescott, 1997), tsunami deposits (Bryant et al., 1990), and fluvial floodplain and channel deposits (e.g. Nanson et al., 1986; Nanson et al., 1988; Fuller et al., 1998). A number of multi-dating studies have demonstrated that luminescence dating can yield results consistent with those of other dating methods (e.g. Nanson and Young, 1987; Nanson et al., 1988; Bryant et al. 1990; Abeyratne, M. et al., 1997; Smith et al., 1997).

The theoretical basis of luminescence dating is as follows. Sedimentary deposits are continually exposed to a natural radiation flux derived primarily from the radioactive decay of uranium, thorium and potassium isotopes in or around the sediment (Aitken, 1985). More minor doses of radiation are derived from cosmic rays, and there may be a usually insignificant contribution from rubidium decay (Aitken, 1985). In certain crystalline minerals such as quartz and feldspar, the excitation caused by this exposure to radiation detaches electrons from atoms which may become trapped in flaws or lattice-charge disequilibrium states in the crystal lattice called 'traps'. The electrons will remain trapped until released by vibrations of the crystal lattice which are intensified by increases in temperature and exposure to light. When released, electrons mostly return back into the crystal lattice structure, however some are transmitted to luminescence centres, particular types of defect in the crystal lattice usually caused by the presence of silver or manganese atoms, causing photon emission (Aitken, 1985).

When electrons are released and traps emptied the material is said to be ‘bleached’. The buildup of electrons acts as a ‘clock’ recording the last time the sediment was exposed to light or heated to a high temperature with older sediments accumulating more trapped electrons than younger sediments. The time since the sediment was last exposed to light can be obtained by taking into account the strength of the radiation flux around the material and is calculated as Luminescence age = ED / AD, where ED, the palaeodose or equivalent dose, is the total amount of energy absorbed since last exposure as quantified by the energy stored in electron traps and AD is the annual dose (Aitken, 1985). Traps differ in how well they hold electrons, with ‘deep’ traps holding electrons more securely than ‘shallow’ traps and needing more energy to be applied before electrons are released. Thermoluminescence measurements are usually restricted to traps from which electrons are released only above 200°C. To measure the
intensity of thermoluminescence a sample is progressively heated to a temperature of usually around 500°C and the light emitted by the sample during heating measured. This produces what is called a 'glow curve', in which a number of different luminescence peaks may be observed at different temperatures. These peaks correspond to different types of trap, with higher temperature peaks being produced by emissions from 'deeper' traps than lower energy peaks (Aitken, 1985). In contrast, in OSL analysis a sample is exposed to light within a given wavelength range, usually centred on 514nm (Rees-Jones et al., 1997; Aitken, 1998), and the luminescence emitted at this wavelength of stimulation measured. The luminescence emitted on OSL stimulation represents only one of the peaks seen in TL analysis, the 325°C peak that is easiest to bleach (Aitken, 1998).

Important issues in the luminescence dating of sedimentary deposits are the need to establish whether the annual radiation dose has been constant over the period since deposition - sites where moisture levels have changed significantly are likely to have also experienced non-constant radiation dose rates (e.g. Prescott and Hutton, 1995) as water acts to attenuate the radiation dose and so variations in moisture levels must be taken account of, and whether the sediments were effectively bleached prior to deposition. If materials are not effectively bleached then the strength of their luminescence signal does not reflect the time since last deposition but some undeterminable longer period. TL measurements commonly include signals from peaks which are not easily bleached, and bleaching of the 375 degree peak commonly used for TL analysis requires 'prolonged' exposure to sunlight (Price, 1994). This is of particular concern in the study of fluvial deposits as turbid water acts as a barrier to light penetration and may prevent effective bleaching (e.g. Spooner et al., 1988; Berger, 1990; Gemmell, 1997), and has led to the suitability of the TL technique for the dating of fluvial sediments to be questioned. To address this concern, a number of studies have been conducted to determine in which fluvial environments inadequate bleaching of the TL signal may occur. Nanson et al. (1991) examined the TL signal from contemporary sediments in three different depositional environments on the Gilbert River Delta in northern Australia and found that floodplain sediments and sediments from shallow channels were thoroughly bleached while sandy bedload from deep channels was relatively poorly bleached. As well as the environment of deposition, the duration of transport, latitude through its influence on the strength of sunlight, and contingent weather conditions may also influence the effectiveness of bleaching. Gemmell (1997) found that the bleaching of the suspended sediment load of an alpine glacial meltwater stream was more effective on clear days and at low sediment concentrations. In stormy conditions when the intensity of daylight was reduced by thick cloud cover, suspended
sediment in the stream was less well bleached and this effect was compounded by the higher sediment loads generally carried by the increased runoff during stormy periods.

In contrast to the TL signal the OSL signal is produced only by easily bleachable traps and any exposure to sunlight or daylight is sufficient to bleach the traps utilised by OSL analysis (Aitken, 1998; Stokes, 1999). This makes the use of OSL preferable in situations where inadequate bleaching is a possibility, and a number of experimental studies have demonstrated the greater ease with which the OSL signal is bleached in comparison to TL signals (e.g. Godfrey-Smith et al., 1988; Rendell et al., 1994). Most samples in this study are dated using the TL technique, however both OSL and TL analysis are applied to a number of samples in order to confirm the validity of the TL technique by comparison with the more sensitive OSL method. Thermoluminescence (TL) dating in this study utilised the regenerative / additive coarse-grain technique (Readhead, 1988) which has been widely used (e.g. Nanson and Young, 1987, 1988; Short et al., 1989; Shepherd and Price, 1990; Page et al., 1991, 1996; Price, 1994; Maroulis, 2000). Dating was carried out in the TL laboratory at the University of Wollongong by David Price. Optically stimulated luminescence (OSL) samples were measured using a Risø TL/OSL reader fitted with a broad-band, blue green stimulation source (420-550nm) and three U-340 detection filters (290nm-370nm), using the 63μ-150μ size fraction. OSL dating was carried out at the University of Wollongong laboratory by Tim Pietsch under the supervision of Dr. Debabrata Bannerjee.

Experience in the study area suggests that TL dating of the sediments here is reliable as young TL ages have been obtained for a number of near-surface sediment samples from the Cooper Creek floodplain which suggests that overbank mud deposits are well bleached (D. M. Price and G. C. Nanson, per. comm., 2000; Maroulis, 2000; this study). Also, comparative Uranium/Thorium radiometric dating of older sand deposits has produced results consistent with TL dates to within error margins (Nanson et al., 1988). The reliability of TL dating in the study area was confirmed by the dating results here which showed a good correspondence between TL and OSL ages (Table 2.1). In 2/6 cases the error margins of the dates overlap (W2839 and W2846) while in the other 4 cases the error margins of the dates are almost touching. There is no systematic difference between TL and OSL ages, and in 3/6 cases the OSL date obtained is older than the TL date while in the other 3 cases the opposite is true. Full details of sample dating results are given in Appendix A. Two TL dates showed abnormal glow curves indicating that the dates may be unreliable. Sample W2843 showed a stepped plateau on the temperature glow curve which may indicate either that the sample was been insufficiently bleached prior to deposition or that it may have been partially re-exposed following exposure.
In this case the age given represents the time of the last solar exposure. This date is taken from a profile in which four samples are dated, and is consistent with the other three dates obtained suggesting that it is reliable. Sample W2847 returned a very young age of $0.7\pm0.1\text{ka}$ and again showed an abnormal glow curve with a sloping temperature plateau indicating that the date may be unreliable. This date was the shallowest of three taken from a profile through a heavily gilgai area showing prominent reticulate channel development, and the abnormal result is thought to be due to contemporary surface sediments being mixed with older, shallowly buried sediments by self-mulching of the cracking soil. The two dates from lower down the profile returned normal glow curves and the comparative OSL dating of the lowest sample (W2849) produced a similar age suggesting the lower dates are accurate.

<table>
<thead>
<tr>
<th>TL Lab. no./ Sample no.</th>
<th>TL Palaeodose (Gy)</th>
<th>TL Age (ka)</th>
<th>OSL Palaeodose (Gy)</th>
<th>OSL Age (ka)</th>
</tr>
</thead>
<tbody>
<tr>
<td>W2839/SF1</td>
<td>40.1$\pm$2.8</td>
<td>20.4$\pm$1.6</td>
<td>49$\pm$0.2</td>
<td>21.1$\pm$1.4</td>
</tr>
<tr>
<td>W2844/SF6</td>
<td>11.9$\pm$1.0</td>
<td>5.3$\pm$0.5</td>
<td>11$\pm$0.3</td>
<td>4.2$\pm$0.3</td>
</tr>
<tr>
<td>W2845/SF7</td>
<td>14.1$\pm$1.1</td>
<td>7.3$\pm$0.6</td>
<td>14$\pm$0.3</td>
<td>6.1$\pm$0.6</td>
</tr>
<tr>
<td>W2846/SF8</td>
<td>15.2$\pm$1.2</td>
<td>8.8$\pm$0.7</td>
<td>15$\pm$0.5</td>
<td>7.5$\pm$0.5</td>
</tr>
<tr>
<td>W2849/SF11</td>
<td>15.0$\pm$0.3</td>
<td>6.1$\pm$0.5</td>
<td>23$\pm$0.2</td>
<td>7.3$\pm$0.5</td>
</tr>
<tr>
<td>W2850/SF12</td>
<td>13.4$\pm$1.3</td>
<td>5.7$\pm$0.6</td>
<td>21$\pm$0.5</td>
<td>7.2$\pm$0.5</td>
</tr>
</tbody>
</table>

Samples for dating were collected both by the removal of soil blocks from excavated trenches and by augering. Block samples were collected during daytime, and the outer, light exposed portion of sediment removed before analysis leaving the unexposed centre of the block to be analysed. Auger sampling produces loose fractured material and was undertaken at night to ensure that no part of the sample was exposed to daylight. Only the centre portion of the auger tube was sampled to minimise the risk of sampling sediment that had fallen down the hole from above. All sediment samples were stored in light proof containers which preserved the moisture content of the sediment at the time of sampling. Care was taken to ensure that sediment was taken from a homogeneous unit with no dissimilar material within a radius of 0.3m, as dissimilar material such as pebbles, pedogenic nodules, or different sedimentary materials can expose the sample to an annual radiation dose whose magnitude is not determinable from the sample itself. For block samples it was possible to ensure that no dissimilar material was present around the sample in the vertical plane while for augered samples it was only possible to determine that no dissimilar material was present within 0.3m of the samples vertically.
2.4 Channel morphology

2.4.1 Introduction

Morphometric measures are filters through which we can view a world changed to facilitate our understanding of it. Just as when we look through a visual filter the appearance of the world is dependent upon the nature of the filter as well as the nature of the world, so the information we receive from morphometric measures is no different and depends upon the construction of the measure itself as well as the object measured. The calculation of a morphometric measure necessitates a loss of information with the morphology of a complex object being represented by a more simple index that generally only captures some aspect or aspects of an object’s geometry. It is precisely this simplification that makes the measures useful but which also necessitates their careful construction to ensure that those aspects of morphology relevant and interesting to us are accurately represented. Results from morphometric measures are not always directly comparable as they do not necessarily represent the same aspect of an object’s form, and even when they do aim to represent the same aspect of geometry differences in the construction of the measures may lead to differing results. For example, Brice (1974a) found that results from the manual definition and measurement of meander bend geometry of a reach of the White River differed greatly from those obtained by spectral analysis of the same reach by Chang and Toebes (1970). The mean wavelength of meander bends measured by Brice was consistently half to a quarter of the values found by Change and Toebes, a difference probably attributable to the fact that spectral analysis measures wavelengths around bends rather than across them as direct measurement does (Ferguson, 1981). The usefulness of a morphometric measure is not dependent only upon its efficacy in characterising relevant aspects of geometry but also on the geomorphic inferences possible from the results. For this reason, the popularity of a method enhances its utility by providing a more extensive database against which results can be compared and enables inferences about the geomorphic significance of results to be more confidently drawn.

Both cross-sectional and planform measures were used in this study and are reviewed here. Most cross-sectional measures are based on the reliable identification of channel bankfull which will be discussed first. Following this, the calculation of channel width-depth ratio and flow discharge are briefly reviewed. A number of different morphometric measures of channel planform have been applied to characterise the form of both single (e.g. Speight, 1965a; Hickin, 1977; Nikora, 1991; Andrie, 1996a) and multiple channel systems (e.g. Rust, 1978; Hong and Davis, 1979; Richards, 1982; Friend and Sinha, 1993). Methods for measuring and characterising single channel planform are considered here and those applicable to multiple
channel systems will be discussed later in reference to particular applications. Williams (1986) states that there are two fundamental approaches to analysing the planform of meandering channels, "the traditional approach, which assumes and emphasises an underlying regularity or meander geometry...and the series approach, which seeks to accounts for varying degrees of irregularity or quasi-randomness using a detailed analysis of the meander trace" (p.147). Both approaches are used here and two methods of planform characterisation are applied. The direct measurement of aspects of channel planform based around the identification of characteristic regularities and sequences of planform change has undoubtedly been the most popular method used to study single channel planform and has provided the bulk of the results from which empirical relations describing morphology have been formulated. As well as this method, fractal analysis which seeks to characterise the constancy of planform irregularity variations with scale is also used (e.g. Tarboton et al., 1988; La Barbera and Rosso, 1989; Nikora, 1991; Montgomery, 1996).

2.4.2 Cross section characteristics

2.4.2.1 Bankfull definition

The definition of bankfull width and depth is central to the calculation of both the width/depth ratio and hydraulic radius and to evaluating the adjustment of channel planform to channel size. Definitions of bankfull have been based on various sedimentary, vegetative and morphometric criteria and Williams (1978) records eleven different definitions of bankfull. Common definitions are the level at which the channel width/depth ratio is a minimum (Wolman, 1955), the height of the lower limit of perennial vegetation (Speight, 1965b) and the elevation of various bench levels in more complex channels (e.g. Schumm, 1960; Woodyer, 1968; Riley, 1972, 1975). In this study the bankfull level of surveyed cross-sections was subjectively defined in a way consistent with the identification of bankfull as the level with the minimum width/depth ratio after Wolman (1955). Riley (1972) has demonstrated that this method is not appropriate for complex channels with a number of benches and that it is best applied to channels with simple cross sections and steep banks, however this limitation is not problematic here because of the almost exclusively simple nature of channel cross-sections. With one exception all anastomosing channels surveyed were simple and canal-like in cross section and the within-channel benches that commonly characterise Australian rivers (e.g. Woodyer, 1968; Riley, 1975; Erskine and Livingston, 1999) were absent. Although more subdued features than anastomosing channels, the floodplain-surface channels surveyed also possessed bankfull levels identifiable easily from survey data and in the field.
For the study of the relationship of planform properties to channel size in anastomosing channels and the influence of channel size on junction form, channel bankfull widths were measured from aerial photographs. The effectiveness of this technique has been demonstrated by Riley (1972). In order to maximise precision, widths were measured using a graduated microscope lens from the most detailed photographs available, which varied between 1:30,000 and 1:90,000 scale depending upon location. Government land surveys have provided spatially complete coverage of the study area with overlapping vertical aerial photographs suitable for stereoscopic analysis and generally between 1:65,000 and 1:90,000 scale. From these photographs, rectified, 1:100,000 scale orthophoto maps have been produced. Systematic coverage began in 1948 and the study area has been re-photographed on average once every 20 years since. Recent photographs were selected from periods of bankfull flow when the contrast between channel and floodplain was clear and stark.

In order to evaluate the accuracy of the aerial photograph width measurement method, measurements were made using the method of a number of channel sections also surveyed. The results presented in Figure 2.1 clearly show that measurements from the different methods are consistent to within an acceptable degree of accuracy for the range of channel sizes shown. The maximum difference between measurements shown is 4m, however differences are generally smaller and averaged 1-2m or 3% of channel width. Such close agreement is due to the availability of photographs from periods of bankfull flow and the relatively steep banks and simple nature of channel cross sections. The method was unsuitable for measuring the width of particularly narrow channels, and for channels below 10m in width produced results with significant errors of up to 30%. This occurred primarily because the error of 1-2m noted above as being typical of all measurements represents a much larger proportion of the total width of small channels. A further problem is caused by the form of narrower channels. Many narrow channels are shallow, subdued features which operate only infrequently. It is difficult to obtain photographs of these channels operating at bankfull, and their subdued nature means that vegetative signatures of the transition between channel and floodplain are often not distinct, making determining the bankfull width of empty channels difficult. When narrow channels are deep and full, significant measurement error still occurs because the canopy of the large trees which occur adjacent to all deep channels can obscure a significant proportion of the channel on aerial photographs.

2.4.2.2 Width/depth ratio (F)

Width/depth ratio is one of the most popular indices of channel form (Pickup, 1976) and there has been extensive investigation of the relationship between sediment properties and channel
Figure 2.1: Comparison of surveyed and aerial photograph channel width measurements
width/depth ratio (e.g. Schumm, 1960a, 1960b, 1961). Width/depth ratios are calculated here using average channel depth $d_{av}$ rather than maximum channel depth $d_{max}$ except where otherwise stated. Width/depth ratio is not a unique measure of channel form and sections of differing shape and size may have identical width/depth ratios. Pickup (1976) noted a number of limitations to the width/depth ratio measure, the first of which is that it is generally calculated at bankfull and so subject to all the difficulties associated with the definition of bankfull. Secondly, the measure is insensitive to some types of shape difference and it may not effectively capture channel bed-width variations (Pickup, 1976). This may be problematic as bed-width variations are often what reflect differences in the composition of the channel boundary which is what many previous studies have used the measure as a proxy index of (Pickup, 1976). Despite these limitations, width/depth ratio is a useful measure of channel form when used to describe the geometry of simple channels with a dominant boundary texture like those studied here.

2.4 2 3 Hydraulic calculations

Discharge calculations in this study were made using the dimensionally balanced form of the Manning’s ‘$n$’ equation given by Yen (1992). Manning’s ‘$n$’ is recognised as one of the most useful ways of expressing flow resistance to steady, fully developed flow and calculating sectional discharge (ASCE, 1963; Yen, 1992). The equation is

$$v = \sqrt{g/n_y \cdot R^{2/3} \cdot S^{1/2}}$$

(2.1)

where $v$ is velocity, $g$ is acceleration due to gravity, $n_y$ is the dimensionally balanced Manning roughness value, $R$ is hydraulic radius and $S$ is slope. The slope used in the calculation was the energy slope rather than bed slope. Graf (1988) notes that in many dryland rivers roughness changes dramatically during a flood event complicating the use of the Manning equation, however this is not problematic here as roughness and channel form do not change appreciably over a single event. Channels were assigned a roughness value of 0.072, the suggested value for winding and sluggish excavated earth channels without vegetation (Yen, 1992) which all the sections closely resemble. The velocities calculated using Manning’s $n_y$ were comparable to those measured during the 1990 flood event (Knighton and Nanson, 2000a) suggesting that both the equation and roughness value selected are appropriate for discharge calculations in these channels.

In one instance above a channel junction discharge was calculated for a composite channel section to account for discharge from a significantly lower, densely vegetated area between the
channels as well as the channels themselves. This section was assigned a roughness value of 0.313 which corresponds to the normal $n_y$ for medium to dense brush (Yen, 1992). The channel was compositied by the summing of individually calculated values for the feeder channels and the vegetated central section as suggested by Chang (1988). This approach does not consider transfers of momentum between sections and assumes flow velocities in each section are uniform and independent. While this tends not to be the case, the vegetation here is strong and robust, is not flattened by flood flows, and because it is never completely inundated mitigates against the transfer of flow momentum across the boundary between sections. For these reasons the level of inaccuracy introduced by not considering momentum transfers should not be too great, and in fact inclusion of the central vegetated section only affected total calculated discharge very slightly because of the high roughness of the section.

2.4.3 Planform characteristics

2.4.3.1 Sinuosity

The most simple direct measure of channel planform is channel sinuosity ($P$). A number of different sinuosity measures have been formulated (Richards, 1982). The most simple measure of sinuosity is used here which is calculated as the distance measured along a channel between two points divided by the straight line distance between the points. Like width/depth ratio sinuosity is not a unique measure of channel form and an infinite number of different planforms may have precisely the same sinuosity (Hey, 1976). The magnitude of channel sinuosity may be quite deceptive, and channels with sinuosities approaching one may appear markedly sinuous.

2.4.3.2 Direct measurement of meander geometry

More detailed direct measurements of single channel planform properties are centred on the definition and measurement of regularities in channel planform associated with meandering (e.g. Leopold and Wolman, 1957; Leopold and Wolman, 1960; Williams, 1986). Single channels which do not meander may be identified by an inability to detect and reliably define planform regularities associated with meandering, or by the results of analysis showing that the measured regularities differ from those typical of actively meandering channels. Planform properties are measured directly from a trace of the channel or channel centreline. The inflection points which define the alternate bends characteristic of meandering channels are marked dissecting the trace into discrete bends and aspects of the geometry of each bend are measured as shown in Figure 2.2. Most of the properties defined in Figure 2.2 may be measured directly, the exceptions being arc angle and radius of curvature which are defined
Figure 2.2: Definitions of meander characteristics
using a circular arc as a geometric analogy to meander bend form. Circular arcs have been proposed as geometric models or analogies of meander bends by both Chitale, (1970) and Brice (1974b), however, as Brice (1984) notes no specific geometric form is satisfactory as a model of meander geometry. As well as the complexity inherent in natural meander trains, individual bends exhibit an inherent downstream skew or asymmetry due to the mechanics of flow in them (Kinoshita, 1961; Parker et al., 1983; Carson and Lapointe, 1983). This makes all models based on bend symmetry such as the circular arc model or the sine generated curve model of Leopold and Langbein (1960) unsatisfactory and means that the definition of arc angle and radius of curvature inevitably involves some subjectivity and abstraction. Two different measures of bend radius of curvature \( r_m \) of a bend have been used; the first describes the form of the bend as a whole, while the second measure used here describes the maximum curvature of the bend.

The most significant consideration regarding the manual measurement of planform properties is the subjectivity inherent in the process. Even though most of the attributes being measured can be measured directly the identification and definition of those attributes may vary greatly between individuals as meandering planforms are often highly complex (e.g. Bruce, 1974a; Ferguson, 1975; Parker et al., 1983). Williams (1986) found that definitions of bend lengths and meander wavelengths for a reach with well defined meanders and no intervening straight sections varied by up to 15% between individuals, and discrepancies between the measurement of the radius of curvature of a bend were higher at up to 25% reflecting the greater subjectivity in the application of this measure. Discrepancies between measurements of more complex reaches will be higher. Concerns over subjectivity have led a number of authors to formulate automated and objective techniques for measuring single channel planform (Hickin, 1977; O'Neill and Abrahams, 1986; Howard and Hemberger, 1991). These approaches use computer analysis of the digitised channel trace to identify reflection points and measure various aspects of meander geometry. The automated definition of meander bends does not suffer from subjectivity but from precisely the opposite problem, a lack of discretion in the definition of bends. Very small directions changes in digitised channel traces may represent real irregularities present in channel planform due to the early stages of meander bend growth or inhomogeneities in channel boundary materials, however they may also represent errors introduced through the digitising process, especially as channels are generally digitised at very small point spacings in an attempt to represent planform as accurately as possible. Approaches to dealing with small scale variations in channel planform have varied, with O'Neill and Abrahams (1986) adopting a subjectively defined threshold of minimum channel curvature before a section of channel is recognised as a bend, and Howard and Hemberger (1991) taking
every inflection point as defining a new bend and relying on averaging repeated digitisings to exclude digitising errors. While automated methods of analysis may be useful, manual measurement of meander planform is adopted in this study. The manual method is simpler to implement, and replicated manual definitions of reach meander properties showed that meander properties here could be reliably defined using the method. Also, some of the bends present in channel reaches here are of low amplitude due to the reaches being low in sinuosity. This means that while there are consistent trends in channel direction change along bends, the direction change between individual digitised points may be very small and in this situation the possibility that digitising errors will affect the results of an automated definition of channel properties is significant.

2.4.4 Fractal geometry

It has been claimed by a number of authors that fractals are of great practical and theoretical importance in the characterisation, understanding and representation of geographic forms (e.g. Mandelbrot, 1982; Feder, 1988; Goodchild and Mark, 1987; Klinkenberg, 1992; Tarboton et al., 1988; La Barbera and Rosso, 1989), and the method has been used to study the geometry of many natural forms such as river channels (e.g. Tarboton et al., 1988; Nikora, 1991; Beavavis and Montgomery, 1996; Montgomery, 1996; Sapozhnikov and Foufoula-Georgiou, 1996), coastlines (e.g. Mandelbrot, 1982), sediment particles (Orford and Whalley, 1983), landscapes (e.g. Mark and Aronson, 1984), soil patterns (Burrough, 1985) and lunar craters (Woronow, 1981). While a number of different kinds of fractal measures and methods have been defined and applied (e.g. Feder, 1988; Klinkenberg and Goodchild, 1992), only the method and type most used in geomorphology, the divider method and statistically self-similar fractals, are considered here. The method will be applied to the anastomosing anabranches of the Cooper Creek and as part of the investigation of planform properties and behaviour (Chapter 4). The finding of river channel fractality — implying that rivers express no characteristic scale of meandering — would seem to be inconsistent with the findings of many previous studies that have detected characteristic scales of meandering determined by channel size in natural channels (e.g. Inglis, 1949; Leopold and Wolman, 1957, 1960; Dury, 1964c; Carlston, 1965; Schumm, 1968; Tinkler, 1971; Schumm and Khan, 1971, 1972; Davis and Sutherland, 1980; Williams, 1986). As well as using the fractal method and the direct measurement of channel planform to investigate channel behaviour here, the findings of the methods and their implications for channel behaviour will be compared to determine which method is more useful for studying channel planform.
2.4.4.1 Fractal objects

It is related of the Socratic philosopher Aristippus that, being shipwrecked and cast upon the shore of the Rhodians, he observed geometrical figures drawn thereon, and cried out to his companions: ‘Let us be of good cheer, for I see the traces of man.’ With that he made for the city of Rhodes and went straight to the gymnasium. (Vitruvius, 1960, p.167)

As the shipwrecked Aristippus noted, there are often clear differences between the geometry of natural forms and those of Euclidean geometry that enable us to clearly distinguish between the two classes. The simple figures of Euclid’s geometry which characterise many man-made forms are generally smooth and possess a characteristic scale of pattern, while many natural forms often appear to possess some degree of irregularity at whatever scale they are viewed. The recognition that Euclidean geometry is inadequate to describe some aspects of the geometry of many forms led Mandelbrot (1967, 1975, 1982) to formulate fractal geometry to take account of variations in geometry over scale. The usage of the term ‘fractal’ has been rather imprecise. Here it will be taken to mean a figure exhibiting self-similarity in any form (e.g. self-similarity, statistical self-similarity, self-affinity, statistical self-affinity) and more precise terms will be defined and used where appropriate.

The purest type of fractal is a ‘self-similar’ fractal which is a “geometrical figure that consists of an identical motif repeating itself on an ever-reduced scale” (Lauwerier, 1987, p.1). Every part of a self-similar fractal however small is identical to the whole, and so every self-similar fractal looks the same at any particular scale as it does at any other. Although always seen as static figures, fractals are without exception produced by dynamic generative processes and the presence of self-similarity is the signature of a scale-free process. Self-similar fractals were originally discovered as the products of certain mathematical functions which were found to describe figures with unusual, counterintuitive geometries such as lines of infinite length and bounded spaces with infinitely small areas (Mandelbrot, 1982; Lauwerier, 1987). An example of a scale-free self-similar fractal is Koch’s curve. The curve is constructed by transforming an initially straight line into a complex, more convoluted figure (Figure 2.3). The middle third of a straight segment of line is replaced by two joined segments each equal in length to the segment of line initially removed. In the next iteration, the same transformation is applied to all straight segments produced by the previous iteration and so on ad infinitum. The motif of the curve is repeated at scales below the initial scale of the object because the generative process cascades down scales with each application transforming smaller and smaller segments of curve with each iteration, the result being an infinitely long curve between any two points however
Figure 2.3 The construction of a Koch Curve. The middle 1/3 of the straight line in (a) is replaced by two joined lines each equal in length to the 1/3 removed (b), and the process repeated to the straight lines in (b) resulting in the figure in (c) The infinite repetition of the process will result in an infinitely long line between two points however arbitrarily close.

Figure 2.4 Sierpinski's Sieve
arbitrarily close they may be. Another example of a self-similar fractal produced by a similar type of iterative process is Sierpinski's Sieve (Figure 2.4) which is generated by subdividing an equilateral triangle into four smaller, equally sized equilateral triangles the central of which is then removed and the process repeated on each of the triangles left and so on. This produces a bounded figure which may have a large extent but has an infinitely small area.

2.4.4.2 The recognition of natural fractals

The precision and infinite scaling required for a figure to be a self-similar fractal means it is likely that only abstract mathematical constructions and computer simulations can ever come close to satisfying the definition. Natural forms do not exist as coherent entities over an infinite range of scales, and are never constructed by perfectly homogeneous processes operating on perfectly homogeneous materials. While they can never show the infinite scaling and precision of self-similar fractals, natural forms may satisfy a looser definition of a fractal object and the recognition of this arose from attempts to meaningfully determine the length of features such as coastlines (Mandelbrot, 1967). Measuring the length of any natural boundary such as a coastline, contour line or river bank is complicated because the result of measurement is itself dependent upon the scale at which we measure the line (Richardson, 1961; Steinhaus, 1969). Viewing a natural object at a larger cartographic scale generally reveals more detail of the object's form and because this new-found detail will be included in our length measurements at this scale the length of the feature will apparently increase as we measure it in more detail. Because of this it does not make any sense to talk of a natural feature such as a coastline having a length, because what particular length should one choose?

Richardson (1961) investigated the scale dependence of coastline length measurements by measuring the lengths of coastlines on the same map with different lengths of ruler as a surrogate for measuring the same feature on maps of varying scale, a standard cartographic technique known as the ruler or divider method (Mandelbrot, 1982; Andrie, 1992). The dividers of length G are walked along the line counting the number of steps N from one intersection of the dividers with the line to the next, and the length of the line is calculated as the product of the number of steps taken and the length of the dividers (N×G). The results of Richardson's analysis of the South African and British coasts and the length of a Euclidean form, a circle, are shown in Figure 2.5. The measured lengths of coastlines plotted against various ruler length scales on a log-log scale fall approximately on an apparently straight line of negative slope. The negative slope is a consequence of the measured length of the coastline increasing as we measure it in more detail because of the inclusion of previously unmeasured
Figure 2.5: The measured length-scale of measurement relationship for the South African Coast, the west coast of Britain, and a circle (after Mandelbrot, 1982, p.33)
irregularities. Similar to the natural forms, as the scale of measurement of the circumference length of the circle is decreased the apparent length also initially becomes greater, however in contrast to the coastline measurements circumference length rapidly approaches a constant because after a scale of measurement sufficiently detailed to adequately measure the smooth form is reached decreasing the scale of measurement fails to capture more of the circle's geometry as no previously unnoticed irregularities are revealed. In contrast, the measurements of coastline length do not approach a constant because the coastlines are irregular below the scale of the most detailed measurements (approximately 10km for the west coast of Great Britain and 100km for the South African coast). If the apparent constancy of slope of coastline plots is real this is an indication that some aspect of coastline geometry is constant over the range of measurement scales and that the object exhibits a type of self-similarity. However, if so, it is not the precise self-similarity of a self-similar fractal but rather a statistical self-similarity. Unlike self-similar fractals like the Koch Curve, each smaller part of the coastline is not exactly the same shape as the coastline as a whole, but rather some statistical characteristic of the pattern, in this case the irregularity of the form, is the same at all scales.

Fractal measures of coastline geometry are derived from the apparent constancy of the coastline scale of measurement-length relationship. The number of lengths of scale G needed to approximate the length of the coastline is

\[ N = aG^{1-D} \]

where D is the fractal dimension of the coastline and a is a constant (Mandelbrot, 1982). Self-similarity is a necessary but not sufficient characteristic of self-similar fractality as many figures that are self-similar but not fractal, for example while a straight line and empty space look the same at whatever scale they are viewed (Roach and Fowler, 1993) they are not fractal objects as a necessary characteristic of a fractal object is that it has a non-integer fractal dimension. In Euclidean geometry a point has a dimension of 0 reflecting the fact that ideally it is infinitely small. A line has a dimension of one as it has no width but only length, and an area has a dimension of 2 as it has both length and width but no thickness. D summarises something about the lines that conventional geometry does not grasp (Mandelbrot, 1982). The higher the value of D between 1 and 2 the greater the space-filling tendency of the line, and a line of high fractal dimension is presumed to appear significantly more irregular than a line of low fractal dimension. In Euclidean terms fractal lines are in between a straight line and an area.
2.4.4.3 The confirmation of statistical self-similarity

To suggest that a natural object is a type of fractal is to suggest something that in purely probabilistic terms is very unlikely. It is trivial to note that one tends to detect more detail in the morphology of natural landforms the closer one looks, however claims of fractality go much further than this. For an object to be statistically self-similar not only does the form of the object have to appear more detailed the closer one looks, but the complexity or irregularity of the object has to be constant over scale. Two kinds of argument have been advanced to support the ‘fractal’ nature of the natural world, the ‘argument from resemblance’ and statistical testing.

The argument from resemblance This argument is based on the visual similarity between the appearance of fractal models of natural features, mostly landscapes, and the appearance of the features themselves. As stated by Mandelbrot (1975, p.3826), the argument claims that the fractality of natural forms “need no longer be tested solely through the quality of fit between predicted and observed values of a few exponents. It is my belief (perhaps a controversial one) that the degree of resemblance between massive simulations and actual maps or aerial views must be treated as evidence.” It is unquestionable that fractal models can produce forms which appear very similar to natural forms. Xu et al. (1993, p.254) state that they “are unable to deny the resemblance of fractal model surfaces to some natural landscape surfaces” and argue that because “the human eye (and brain) has the unsurpassed ability to observe and interpret complex phenomena...it is easy to infer that landscapes possess self-similar properties.” Xu et al. do go on to say that such properties should not be defined visually, and quote Piech and Piech (1990) who note that a suggestive visual appearance does not constitute proof. As Nabokov has stated in a literary context, “Resemblances are the shadows of differences. Different people see different similarities and similar differences” (Nabokov, 1962, p.894). Objects may resemble each other and yet still be fundamentally different. To say that things are similar is a much weaker claim than to say they are identical. This should not be ignored and claims of identicality cannot be justified from merely resemblances, especially because the fractal model is based upon a precise constancy of the complexity-scale relationship that in purely probabilistic terms is highly unlikely.

The statistical confirmation of statistical self-similarity In objects measured using the divider method this is based on testing the linearity of the log-log plots derived. Initially it is assumed that the fractal dimension of the object being studied is constant for the range of scales over which the object was measured and a single fractal dimension for the object is produced (Xu et al., 1993). The scales at which an object is sampled are only a few of the infinite number of
scales falling within the range of scale over which the object is measured, and it is assumed that
these are representative of the object's form over the range. The reality of the calculated fractal
dimension is only determinable through further analysis, as a straight line of best fit can be
forced through even the most non-linear collection of points and so the method will produce a
fractal dimension for an object even if the object possesses no fractal characteristics. This
makes the statistics used to verify the actual existence of a linear relationship between the
plotted variables crucial to the reliability of the method, especially because, as is well known
by practicing scientists who often use it to their discursive advantage, log-log plots are not very
sensitive to deviations from linearity (e.g. Andrle and Abrahams, 1989, 1990). Francis (1970,
p.186) notes that “[d]ouble log plotting is the worlds finest and simplest methods of hiding
errors”.

In the earliest studies of the statistical self-similarity of natural forms most evaluations of fit of
a line to data were made visually, however recent advances in computing technology and data
processing techniques have made a number of more rigorous automated techniques available.
The most popular of these is least squares linear regression (LSLR) which has been used in
most fractal-based geographic studies (Andrle, 1997). LSLR was originally developed for
predicting the value of one variable from that of another, however in fractal analysis the
technique is instead used for estimating the slope coefficient of the line of best fit and so the
fractal dimension of the object. The r² value is used as an indicator of linearity, with high
values taken as indicating statistical self-similarity r² values are decreased by both systematic
deviation from linearity and by random non-systematic ‘noise’, and significant non-linearity
may be present in relationships with high r² values. For example, Andrle (1997) reports
significant curvature in relationships with r² values greater than 0.999 which is possible due to
the insensitivity of log-log plots to deviations from linearity. For this reason high r² values are
not sufficient to confirm linearity of a relationship and further tests need to be carried out to
ensure that any apparent linearity is real (Andrle, 1996b, 1997; Andrle and Abrahams, 1990).

To confirm the linearity of the relationship between variables one should also examine the
residuals from the regression line and test for serial positive autocorrelation amongst the
variables using a test such as the Durbin-Watson test (Andrle, 1997). If the relationship
between variables is truly linear then the residuals of the regression, the difference between the
observed values of the data and those predicted from the regression equation, should be
randomly distributed around the line. Systematic trends and positive autocorrelation in
residuals indicates that variations in the data are more complex that the line of best fit.
Given that neither visual assessment of the linearity of log-log plots or high $r^2$ values are sufficient indicators of statistical self-similarity, many of the objects previously described as statistically self-similar on the basis of results from these tests may in fact turn out to have been wrongly classified. Goodchild and Mark (1987, p.265) have claimed that “a wide variety of spatial phenomena have been shown to be statistically self-similar over many scales, suggesting the importance of scale-independence as a geographic norm”, however it is questionable whether the evidence supporting this claim is as strong as Goodchild and Mark suggest. An examination of 22 journal articles by Andrle (1997) showed that none presented sufficient evidence to confirm the linearity of relationships the claimed to exist. The west coast of Britain was considered the paradigmatic example of a naturally occurring statistically self-similar object (e.g. Mandelbrot, 1967, 1982) and was of great discursive importance in ensuring the acceptance of the fractal concept by geographers. The coastline was described by Mandelbrot as a random self-similar figure (Mandelbrot, 1967, p.638), the evidence for which was the alleged linearity of the log(NeG)-logG plot (Figure 2.5). However, even a cursory visual examination reveals significant deviations from linearity in the plot that were in fact noted by Richardson (1961) on whose data Mandelbrot’s analysis is based, and a re-examination of geometry of the coastline by Andrle (1996c) found significant non-linearity in the complexity-scale relationship and determined using a non-fractal method of analysis called the ‘angle measure’ that the coastline possessed two distinct peaks in complexity with lower complexity at the scales in between. The larger scale peak in complexity corresponded to the influence of large-scale tectonic features including granite batholiths and parallel grabens on coastline form, while the smaller-scale peak in complexity was found to correspond to small bays representing partially submerged glacial and fluvial valleys (Andrle, 1996b).

2.4.4.4 Partially fractal objects: fractal elements and their detection

Statistical self-similarity in natural forms cannot extend over infinite ranges of scales because, as noted earlier, natural forms generally do not exist as coherent entities over an infinite range of scales. Scheidegger (1970, p.9) noted that the apparent linearity of Richardson’s plots only extended over a limited range of scales and stated that as well as not being justified by the evidence it seemed “physically absurd” to postulate the existence of a constant complexity-scale relationship over all scales of measurement. Forms at different temporal and spatial scales are commonly operated on by different sets of processes (e.g. Schumm and Lichty, 1965; Harvey, 1968) which are a priori unlikely to produce the same characteristic geometry. For this reason claims of statistical self-similarity are commonly restricted to small, limited ranges of scale called fractal elements (e.g. Mark and Aronson, 1984; Klinkenberg and
Goodchild, 1992; Avnir et al., 1998). Contrary to the infinite scaling implied by the term ‘fractal’, many identified fractal elements extend over very limited ranges of scale (Avnir et al., 1998). In a study of the influence of valley type on the complexity-scale relationship of river channels, Beauvais and Montgomery (1996) recognised fractal elements extending over as little as 0.2 of an order of magnitude (Beauvais and Montgomery, 1996, the South Fork Hoh River, Figure 5 page 1445), and Montgomery (1996) defined fractal elements commonly extending over 1 order of magnitude range of scale. The common practice of identifying such small ranges of fractal behaviour led Avnir et al. (1998) to question whether such objects should really be described as fractal, and the question is of more than semantic importance because the smaller the range over which self-similarity is defined the less confidence one may have in its reality (Xu et al., 1993; Andrle, 1996b). This reason for this is that the statistical problems in testing for the linearity of log-log plots discussed earlier are compounded by a number of additional problems. Firstly, the scales chosen to sample the object may be crucial to determining whether small fractal elements are detected, and as one cannot know a priori where fractal elements are one cannot space sampling scales to ensure their detection. Obviously, the smaller the fractal element or the coarser the scales of our sampling the greater the likelihood that the element will be missed. More importantly as Xu et al. (1993, p.255) point out, “[s]ometimes the object from which one extracted the log-log plot is non-fractal and the log-log plot is actually an arc and the so-called “linear segment” is just a segment of the arc.” Geomorphic data are inherently noisy as processes are never applied uniformly to uniform materials and because landforms are often not in equilibrium with processes currently operative but may also contain elements related to many past periods of activity (e.g. Caldarelli et al., 1997). Landscapes have been described as being akin to palimpsests. This noise becomes problematic to the detection of fractal elements because as the range of scale under consideration is reduced the degree of scatter of points about the best-fit line increases with respect to the length of the line which may confound tests such as the Durbin-Watson test, mask any non-linearity present and lead us to conclude that the object in question is statistically self-similar even if it is not (Andrle, 1996b; Andrle, 1997).

The fractal model is only one possible form of the complexity-scale relationship and, as noted earlier, one that at least in purely probabilistic terms is highly unlikely. In another form of the relationship complexity continually varies with scale peaking at scales dependent upon the processes operating on the object in question (e.g. Burrough, 1985; Andrle, 1996b). The fractal dimension of non-fractal objects will vary with the range of scales that is used to compute it
(Burrough, 1985; Andre, 1996a, 1996b), and it thus makes as much sense of talk about the fractal dimension of a non-fractal object as it does to talk about the length of a coastline.

2.4.4.5 Fractal studies of river channels

The fractality of river channel geometry and behaviour has studied by a number of authors (e.g. Tarboton et al., 1988; La Barbera and Rosso, 1989; Beauvais and Montgomery, 1996; Montgomery, 1996; Nikora et al., 1996; Sapoznikov and Foufoula-Georgiou, 1996, 1998). Results from a number of theoretical simulation models of meandering channel dynamics suggest that fractal dynamics may be inherent in meandering channel behaviour (Furbish, 1991; Montgomery, 1993; Stolum, 1996), however it is the fractal geometry of channel planform which is of interest here and which will be investigated on the Cooper anastomosing anabranches. Most fractal-based investigations of channel planform have used the divider or ruler method described earlier (e.g. Nikora, 1991; Beauvais and Montgomery, 1996; Montgomery, 1996). There has been a marked conflict between the findings of many studies, with channels being found to be different types of fractal object by different authors (Nikora et al., 1996).

Nikora (1991) used the divider method and logN-logG plots to study the complexity-scale relationship in 46 Moldovan rivers, finding them all to be fractal. Nikora found that statistically self-similar fractal behaviour was limited to a range of scales between the width of the channel and the width of the river valley, however did not report any tests of the linearity of the log-log plots beyond visual inspection which, as stated earlier, is not sufficient to confirm the reality of any apparent fractality. Nikora reported fractal dimensions ranging up to 1.33.

Montgomery (1996) and Beauvais and Montgomery (1996) both used the divider method to examine the scaling relationships of rivers flowing in different situations. Montgomery (1996) compared log(N×G)-logG plots of alluvial channels, incised channels, and channels developed in glacial spillways. Montgomery examined the central portion of the log(N×G)-logG plots over which N×G varies and outside which it is constant, and applied a lack-of-fit test which compares the slope of different areas of the log-log plot to assess the linearity of slope of this central portion. This test is described by Andre (1992) and is based on the fact that if the relationship is truly linear then regression lines fitted to different portions of the data should have the same formula and the same slope. If they do not then this indicates that the relationship is in fact nonlinear. Montgomery (1996) found that alluvial channels showed evidence of statistical self-similarity described by a single fractal dimension, while incised
bedrock channels were not fractal and glacial channels appeared to be 'bi-fractal' with two contiguous ranges of different but constant fractal dimension. Montgomery also compared channel sinuosity and computed fractal dimensions which were as high as 1.5 for some channels, finding a good positive relationship between channel sinuosity and fractal dimension with more sinuous channels having higher fractal dimensions. Montgomery speculated that the difference between the results for alluvial and incised channels suggested that the added constraints on channel form imposed by the bedrock channel boundaries acted to simplify channel form which would otherwise tend to be more complex and fractal.

Beauvais and Montgomery (1996) used essentially the same methodology as Montgomery (1996) except that testing of the linearity of a relationship was based on the examination of regression residuals. Beauvais and Montgomery examined the alluvial, incised, and glacial valley channels and restricted the range of scales over which they described fractality to that between channel width and the scale of the largest single meander bend. This limitation combined with the distribution of sampling scales chosen meant that log-log plots were constructed from as few as 6 data points. Beauvais and Montgomery found fractal dimensions ranging up to 1.176, and in contrast to Montgomery (1996) found that incised channels were statistically self-similar fractals described by a single fractal dimension and that alluvial channels were bi-fractal statistically self-similar objects possessing two distinct but not necessarily contiguous ranges of scale with different fractal dimensions. Like Montgomery (1996) Beauvais and Montgomery (1996) found that channels in inherited glacial valleys were bi-fractal objects.

2.4.4.6 Methodology

In the implementation of the fractal method here, LogG-logN plots were constructed using the divider method which was implemented with the Benoit© fractal analysis program. The program measures the length of a digitised line represented as a pixellated bitmap. Channel centrelines were digitised at point spacings of less than one channel width, and a vectorised line interpolated between the points. This line was then converted to a bitmap with the channel centreline being represented by a single pixel and care was taken to ensure that each pixel represented a distance no greater than a third of channel width to ensure that pixellation did not influence results. The program adds both the number of full ruler lengths used to measure the line to the fractional number of partial ruler lengths used to measure the residual not measured by unit ruler length/s. This is considered to introduce the least error of the various methods.
used to deal with the residual from unitary ruler lengths (Andrle, 1992). Log-log plots of the number of steps against the scale of measurement were constructed from 200 scales ranging in length from the channel width to approaching the length of the reach. The use of channel width as the lower bound for investigation is consistent with other studies (e.g. Nikora, 1991; Beauvais and Montgomery, 1996), however previous studies have defined a range of upper scale limits for statistically self-similar behaviour including the valley width (Nikora, 1991) and the scale of the largest bend or point at which scatter in the residuals of the regression analysis abruptly increases (Beauvais and Montgomery, 1996). Here all scales of measurement up to those approaching the straight line length of the study reach were initially included, and the position of the upper scale limit then reconsidered based on the results of the initial analysis. LSLR was applied to the data points and a fractal dimension computed from the line of best fit. Residual plots were then examined and the Durbin-Watson test for the autocorrelation of residuals applied to investigate the presence of any systematic structure in the residuals.
3 REGIONAL SETTING, QUATERNARY HISTORY AND RIVER CHARACTERISTICS

3.1 Introduction

This chapter begins by reviewing the morphology, geology and climate of the Lake Eyre Basin and study area. The Quaternary history of the area is reviewed and the landforms of the floodplain region are briefly described. The morphology and characteristics of the various types of channel found are reviewed in some detail, the findings of previous work described, and gaps and inconsistencies in current knowledge highlighted.

3.2 Regional setting

3.2.1 Lake Eyre Basin morphology and climate

The Cooper Creek is part of the Lake Eyre Basin (Figure 1.1) which at approximately 1.14 million km² is one of the largest internally drained catchments in the world (Kotwicki, 1986; Nanson et al., 1988). The elevation and relief of the basin are generally very low and the basin is composed mostly of gently rolling hills and expansive plains. Although both elevation and relief increase approaching basin margins, 90% of the land surface is below 500m in elevation and 70% below 250m with the low point basin being the bed of Lake Eyre towards the south western margin of the basin at -15m. The study reach is shown in more detail in Figure 3.1. The Lake Eyre Basin is subject to an arid to semi arid climate classified as BSh to BWh under the Koppen system. Monthly average climate statistics from three stations in the basin (see Figure 1.1 for locations) are given in Tables 3.1-3.3, and precipitation and evaporation variability over the study area shown in Figure 3.2. Long, hot summers alternate with short, cool winters, and the area experiences the large diurnal temperature variations typical of desert climates. At Moomba approximately 100km west of the downstream end of the study area mean daily temperature variations approximate 15 degrees throughout the year (Bureau of Meteorology, 2000).

As one would expect of such a large area, there are marked variations in climate over the basin. Precipitation and evaporation levels are roughly inversely proportional to one another, with precipitation highest in the north east of the basin and evaporation greatest in the south west (Figure 3.2). Precipitation is strongly seasonal with summer (November to March) monsoonal
Figure 3.1: Map of the study area, after Knighton and Nanson (1994a, p.138)
Figure 3.2: (i) Precipitation and (ii) evaporation distributions over the study area (after Bureau of Meteorology, 2000). Evaporation measured with a Class A Pan.
Table 3.1: Climate averages for Hughenden Post Office Meteorological Station. Elevation 324m, length of record 126 years (Bureau of Meteorology, 2000).

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<th>Mean maximum temperature (°C)</th>
<th>Mean minimum temperature (°C)</th>
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Table 3.2: Climate averages for Windorah Meteorological Station. Elevation 126m, length of record 113 years (Bureau of Meteorology, 2000).

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Table 3.3: Climate averages for Moomba Meteorological Station. Elevation 39m, length of record 18 years (Bureau of Meteorology, 2000).

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rain making up approximately 70% of the yearly total and a significant secondary winter rainfall maximum of 15-20% of the yearly total falling in May and June. Precipitation in arid areas tends to be highly unreliable and irregular (Goudie, 1987; Cooke et al., 1993; Tooth, 2000) and the Lake Eyre Basin is no exception with precipitation becoming more irregular as the Basin becomes more arid. The northern region of the basin lies on the southern margin of monsoonal influence, and most precipitation is derived from incursions of moist tropical air masses, monsoonal depressions and thunderstorms (Allen, 1988, 1990). The dominance of monsoon rains compounds the precipitation irregularity typical of arid areas, and the influence of the monsoon is not confined to northern regions and may affect the entire basin and penetrate far inland. In the 1984 filling of Lake Eyre much of the water fell near the lake from a tropical depression that had originated off the north-west of Australia and subsequently tracked south-east (Kotwicki, 1986). Tropical cyclones are also an extremely significant although even more unpredictable and irregular source of precipitation during the warmer months and have been the cause of many major floods. The smaller, winter rainfall maximum is caused by precipitation from southern maritime and continental, generally easterly tracking air masses (Kraus, 1955).

3.2.2 Geology

The Lake Eyre Basin was formed by the Lake Eyre depression, a downwarping event which faulting near the lake itself indicates occurred in the late Cainozoic (Wopfner and Twidale, 1967) and is composed of a number of sub-basins of varying degrees of independence. The Cooper Creek in the study area overlies the Cooper Basin, a sub-unit of the larger 1.7 million km² Eromanga Basin which underlies most of the Lake Eyre Basin. Structurally, the Eromanga Basin is a north-easterly trending syncline which contains about 3000m of material. Accumulation in the Eromanga Basin began due to major downwarping, and the resulting depositional sequence is composed mainly of Jurassic and Cretaceous terrestrial units the texture of which varies reflecting alterations between fluvial and lacustrine depositional environments (John and Almond, 1987). The Cooper Basin is elliptical in form and extends from the beginning of the Cooper Creek in name at Windorah to Lake Eyre (Watts, 1987). A geological map of the study area is given in Figure 3.3. The depo-centre of the Cooper Basin is also the depo-centre of the Eromanga Basin (John and Almond, 1987), however the Cooper Basin is distinguished from the Eromanga Basin by the presence there of younger Permian and Triassic deposits not laid down extensively enough to extend into the Eromanga Basin (Senior et al., 1968). The uplift of the eastern margin of the Eromanga Basin which forms the eastern margin of the Lake Eyre Basin occurred in the Tertiary, when the north-west to north-easterly
Figure 3.3: The geology of the study area (from Queensland Geology 1:2500000 Sheet MD34-75-8) and a schematic cross section of the Cooper Creek valley fill deposits showing the typical thicknesses of each layer.
trending fold structures which dominate the internal structure of the Eromanga and Cooper Basins also formed (Wopfner, 1960; Gregory et al., 1967; Senior et al., 1968; Wecker, 1989). These fold structures control the location and planform of the Cooper Creek valley in the study reach (Figure 3.3). Above the confluence with the Barcoo River, the Thompson River which joins the Barcoo at Windorah to form the Cooper Creek follows the Thompson Syncline (Gregory et al., 1967). Between Windorah and Lake Yamma Yamma, a 30km wide playa lake occupying a synclinal depression, the Cooper valley occupies a structural low between confining dome and anticline structures (the location of Lake Yamma Yamma and of other landforms specifically referred to subsequently in this chapter is given in Figure 3.1 unless stated otherwise). Downstream of Lake Yamma Yamma the valley axis trends north to south following a southward plunging syncline extending for approximately 210km, and then swings west along the E-W trending Wilson Syncline at the confluence with the Wilson River (Gregory et al., 1967; Senior et al., 1968). The river then flows approximately 100km west to the Innamincka Dome which marks the downstream extent of the study reach.

The surfaces surrounding the river valley display the reddish hues characteristic of Australia’s ‘red centre’ and show a range of landforms. Gibber surfaces of broken siliceous material dominate the land surface with the underlying bedrock outcropping in places. The gibber material is mostly derived from the weathering of the Tertiary silcrete duricrust which previously formed a mostly continuous and flat land surface over the study area (Mabbutt, 1967). This surface is preserved intact in some places, while in others it is dissected into spectacular mesas and buttes up to two hundred metres high. The oldest outcropping rocks are the interbedded sandstones and mudstones of the Late Cretaceous Winton Formation (Gregory et al., 1967). This Formation outcrops only at structural highs and is overlain by the more widely outcropping Tertiary Glendower Formation parts of which are strongly silicified and make up some of the Gibber material (Gregory et al., 1967; Senior et al., 1968). The red of the surrounding surfaces contrasts with the dark alluvium of the Cooper Creek valley which seen from the air cuts a dark swath across the burnished landscape.

Whitehouse (1944, 1948) was the first to describe the geomorphology of the area in detail. He considered the contemporary floodplains to be small remnants of the floors of a number of lakes each equal in size of Lake Eyre today which he claimed had formed due to ponding of the channels “in the past geological age when rainfall was tremendously heavy” (Whitehouse, 1948, p11). Whitehouse stated that the subsequently drier climate which still exists today had caused aeolian sand to encroach over the plains limiting their extent, however the actual pattern of Late Tertiary and Quaternary sedimentation has been very different. While the period has
seen phases of both erosion and deposition, the general trend has been for the accumulation of sediment in the low points of the land surface around structural highs (Gregory et al., 1967; Senior et al., 1968). The Cooper valley in the study area varies in width from just over 100m upstream of the Innamincka Dome to over 59km wide north of Lake Yamma Yamma. Above the weathered Tertiary basement rocks, the valley fill is composed of up to 100m of alluvium which is mostly Quaternary in age but may be Late Tertiary in deeper areas of the valley (Senior et al., 1968). The valley fill may be divided into two units; a lower, predominantly sand unit up to 90m in thickness, and an overlying fine-grained mud unit (Rundle, 1976; Veevers and Rundle, 1979; Rust, 1981; Rust and Nanson, 1986; Nanson et al., 1986; Nanson et al., 1988; Maroulis, 2000). A schematic section of the valley fill deposits is given in Figure 3.3, and the units will be described in turn.

3.2.2.1  The Katipiri Formation

The bulk of the upper sand unit is made up of the Katipiri Formation. Sediments from this Formation have been dated using thermoluminescence dating (Rust and Nanson, 1986; Nanson et al., 1986; Nanson et al., 1988; Maroulis, 2000) and Uranium series ages have been obtained for post-depositional pedogenic minerals (Nanson et al., 1988). The Katipiri Formation ranges in age from greater than 740ka BP (the age of the oldest dated TL sample from Maroulis, 2000) to approximately 35ka BP (Maroulis, 2000). Although a number of previous authors considered the Katipiri Formation to have been produced by large, braided channel systems (e.g. Rundle, 1977; Veevers and Rundle, 1979; Rust, 1981; Rust and Legun, 1983), the unit is characterised by upward fining sequences in medium to coarse sands with large scale tabular bedforms which indicate deposition by laterally active, wide and probably shallow meandering channels (Nanson et al., 1986; Rust and Nanson, 1986; Nanson et al., 1988; Nanson and Tooth, 1999; Maroulis, 2000). In the upper part of the Formation near the boundary with the younger overlying mud unit, excavations have revealed the existence of scroll bars and palaeochannels from these channels which are significantly larger than any existing today (Nanson and Tooth, 1999; Maroulis, 2000). Stratigraphic studies by Rust (1981) and Rust and Nanson (1986) suggested that the stratigraphy showed a cyclic pattern of stacked mud dominated and sand dominated units, however subsequent study has shown that this is not the case (Maroulis, 2000). Some small deposits of fine-grained muds from overbank deposition are interbedded with the sands here, however there are no large, continuous fine-grained units and overall there is little fine-grained material present (Maroulis, 2000).
3.2.2.1 The mud unit

The overlying "mud" unit varies in thickness between 1.5m and 10m and averages 2-3m (Maroulis, 2000). This unit is the contemporary floodplain and the unit in which the present channels are formed. While some channels extend down to the Katipiri Formation, they generally do not penetrate into it because the surface of the sand sheet is heavily indurated and resistant to erosion (Nanson et al., 1988). This unit ranges in age from 118ka in some places at its base to contemporary at the surface. The characteristics and behaviour of the mud deposits have been described by Nanson et al. (1986), Nanson et al. (1988), Rust and Nanson (1989), Maroulis (1992), Maroulis and Nanson (1996), Gibling et al. (1998) and Maroulis (2000).

Analysis conducted in this study shows that disaggregated sediment from the mud unit is generally composed of 35-60% clay, 40-60% silt and 4-9% sand and has a mean grain size of 25-30μm. Smectite, an expandable clay, is a dominant clay mineral occurring in the soil and causes the soils to shrink and swell with moisture changes (Maroulis, 1992; Maroulis and Nanson, 1996). The mud unit can be divided into upper and lower layers by depth on the basis of pedologic characteristics and these layers will be described in turn beginning with the upper layer.

The upper layer

This layer averages 0.4-0.5m in thickness and its soil structure is dominated by cracking and fracturing of the soil mass. An examination of soil micrographs shows pedogenic aggregates to be a prominent component of the soil (Nanson et al., 1986; Rust and Nanson, 1989). Soil aggregates may form by a number of processes including erosion from weathered mudstones (Nanson et al., 1988), as faecal pellets (Rust and Nanson, 1989), by the adhesion of particles by organic matter (McCave, 1984) and salt efflorescence (Bowler, 1973). Aggregate formation here is a pedogenic process occurring in response to the wetting and drying of expandable sediment resistant to dispersion in low salinity conditions (Nanson et al., 1988; Rust and Nanson, 1989; Maroulis, 1992; Maroulis and Nanson, 1996). Aggregate size is sensitive to the nature and rate of wetting and an experimental study showed aggregates formed by simulated rainfall were significantly larger (0.70-0.75mm) than those formed by immersion (0.15mm) (Maroulis, 1992; Maroulis and Nanson, 1996). Immersion is analogous to wetting by overland flow, and is probably the most important mechanism of aggregate formation on the floodplain (Maroulis, 1992; Maroulis and Nanson, 1996). Maroulis (1992) found that aggregates formed after disaggregated sediment experienced 2-3 wetting and drying cycles, and flume studies (Nanson et al., 1986; Maroulis, 1992; Maroulis and Nanson, 1996) have shown that the aggregates are durable and persist for significant periods of time in transport. When transported as bedload in sufficient quantities the aggregates can form ripple, dune and antidune bedforms, and calculations using the Shield's entrainment equation by Nanson et al.
(1986) indicate that the low density aggregates (2.34 g/cm$^3$ cf. 2.65 g/cm$^3$ for quartz sand) could be entrained at flow depths of approximately 0.2 m on the Cooper floodplain. The geomorphic significance of the aggregates will be considered in more detail in subsequent sections.

Although Sturt (1965) reports observing cracks up to 2.4 m deep in the soils of the region, none so deep have been observed in the study area. The upper soil layer is self-mulching, and when the soil cracks and fractures, loose, dry chunks of sediment fall down the cracks and on wetting are absorbed into lower soil layers. Over a number of wetting and drying cycles this churns the soil layer and destroys any fine scale sedimentary structures present (Maroulis, 2000). An arid region form of patterned ground, gilgai, forms across all areas of the floodplain but are largest in areas with reticulate channels. In fine-grained soils like those here gilgai formation is caused by uneven wetting and drying of cracking, swelling sediments (Hallsworth et al., 1955; Verger, 1964; Cooke et al., 1993). Cracks in expansive soils are an expression of uneven drying patterns and in dry conditions represent points of extreme dryness with soil moisture content increasing as one moves away from cracks into the centre of intact soil blocks (Hallsworth et al., 1955). Soil around cracks is wetted quicker than soil which must be wetted by infiltration into intact blocks which causes the soil near cracks to expand fastest, increased pressure near cracks, and the deformation of the soil surface to equalise pressures which results in either the uplifting of blocks of soil between cracks as slippage occurs (Hallsworth et al., 1955) or the deformation of soil blocks between cracks into dome shapes (Verger, 1964). Both methods cause the formation of surface undulations and alternate hummocks and hollows which often given an approximately sinusoidal pattern to surface elevation variations. The peaks in surface height are called gilgai mounds or "puffs" and the troughs are called depressions (Harris, 1959). On flat surfaces, gilgai are generally circular or polygonal in form with a central mound surrounded on all sides by depressions and possess no general orientation although observations suggest that some may be aligned with regional wind directions (Goudie et al., 1990). On slopes gilgai generally possess a strong orientation and tend to be linear features aligned with slope direction (Hallsworth et al., 1955; Elberson, 1983). Linear mounds extend downslope with linear troughs on either side giving the sloping surface the appearance of a ploughed field. The scale of gilgai developed in a soil is dependent upon a number of factors which can be divided into internal and external factors. Internal factors are soil properties which determine the swelling potential of the soil such as the swelling potential of the clay type present, and gilgai will tend to be larger in soils with greater swelling potentials (Hallsworth et al., 1955). Important external factors controlling gilgai formation are the frequency of wetting and drying cycles which drive the cracking and swelling of the soil, and the rate of erosion which may prevent the organisation of the soil surface and gilgai formation (Elberson, 1983;
Paton et al., 1995) This erosion may be the result of slope processes (Elberson, 1983; Paton et al., 1995), or in the case of the generally flat Cooper floodplain, fluvial or aeolian processes.

Detailed soil surface profiles from both a heavily gilgaied reticulate area and an unchanneled area are given in Figure 3.4 and illustrate the variability of soil surface form. Heavily gilgaied areas show much greater magnitudes and larger scales of surface elevation variations than do weakly gilgaied areas. The largest gilgai that have been observed in the study area are those shown in the surface-profile given in Figure 3.4 and, even where prominent gilgai are developed, gilgai sizes are usually much lower than that of those shown.

The lower layer In contrast to the upper layer, the lower soil layer is more compact, massive in structure and is not cracked. Aggregates are not a prominent component of soil structure in this layer as compaction which increases with depth makes the soil here denser and soil structure more homogeneous (Rust and Nanson, 1989) (Figure 3.5). The mud layer as a unit is largely impermeable to moisture. While the cracks in the soil surface allow water to penetrate into the upper layers of the soil mass, the absorption of water by the soil and consequent swelling causes cracks to close and wetted soil surfaces quickly seal. The sealing is so effective that a trench excavated three metres below the water level of the adjacent Goombabina Waterhole experienced no seepage over a number of days even though the waterhole bank was only 1m away (J. C. Maroulis and G. C. Nanson, personal communication, 1997).

3.3 Quaternary history

Possible causes for the change of the Cooper Creek and other Channel Country streams from the sand dominated system which deposited the Katipiri Formation to the present mud dominated system are environmental change, tectonic activity, a change in the character of sediment supplied to the rivers or any combination of these factors. Fortunately a brief consideration of these factors can exclude all but environmental change and the mud unit is thought to be the product of a change to a drier climate causing a reduction in the energy of the channels (Nanson et al., 1988; Gibling et al., 1998; Nanson and Tooth, 1999; Maroulis, 2000). While there is local tectonic activity in parts of the Channel Country and some evidence for the recent uplift of the Innamincka Dome, there is no evidence of the widespread, regional-scale tectonism which would be needed to produce the widespread, regional changes in sedimentation observed (Rust, 1981). The sediment supplied to the Cooper floodplain is derived ultimately from the valley sides of the basin and supplied to the floodplain primarily by
Figure 3.4: Comparison of typical floodplain surface morphologies from (a) a heavily gilgaied reticulate area [R] and (b) an unchanneled area [U]. Profile locations given in (c). (d) shows an unchanneled area similar to [U] with subdued surface relief, and (e) small scale gilgai showing the typically undulating nature of the surface pattern (carcass for scale is approximately 60cm across).
Figure 3.5: Soil micrographs showing the variation of mud unit compaction with depth. Horizontal field of view 3.5mm in all cases. (a) Shows loose surface material with aggregates the dominant component of soil structure, (b) soil at 1m depth with aggregates less visible and void spaces much reduced, and (c) more dense soil at 2m depth with void spaces absent.
tributary channels, and it is conceivable that the change in the sediment regime of the Cooper reflects a change in the nature of sediment supply to the floodplain which would most likely be due to environmental change. However, an examination of tributaries joining the Cooper Creek in the study reach shows that they are generally sand transporting channels implying that the change of the Cooper to a mud-dominated system has not been triggered by a shift of tributary channels to a mud-dominated sediment regimes, although it is likely that the rate of sediment supply to the floodplain has varied greatly as a result of climatic change. Environmental change would seem to have influenced the character of sedimentation primarily through its influence on flow regime, and so the sedimentary record of the Cooper Creek provides a detailed history of Quaternary environmental change (Nanson et al., 1988; Nanson and Tooth, 1999; Maroulis, 2000). As stated previously, the valley fill may be Late Tertiary in age however the record studied to date extends only up to 750ka BP and covers the last two glacial cycles in some detail (Nanson et al., 1988; Nanson and Tooth, 1999; Maroulis, 2000). Dates have been obtained from both the Katipiri Formation and the overlying mud units with interest being focused on timescales of glacial cycles. Results show that the previous two interglacial periods (244-190ka BP and 130-74ka BP) have been characterised by pluvial episodes which caused extensive fluvial activity and the deposition of the bulk of the deposits of the upper Katipiri Formation (Nanson and Tooth, 1999; Maroulis, 2000). During these periods channels were much larger and the flow regime more perennial than that during the intervening arid glacial periods (Nanson and Tooth, 1999; Maroulis, 2000). During glacial arid phases, it is thought that sediments like those of the contemporary mud dominated floodplain were deposited, of which only small vestiges remain because of stripping and reworking in the subsequent (interglacial) periods of enhanced fluvial activity (Gibling et al., 1998; Maroulis, 2000). Gibling et al. (1998) speculate that these units would have been preserved had accretion rates been significantly higher. The current interglacial period is unusual in that it has not (so far) been characterised by enhanced fluvial activity, however, records from past interglacials suggest that the onset of enhanced fluvial activity may lag the climatic indicators like sea level and temperature by which interglacials are defined by up to 15ka (Nanson et al., 1992), and so the present lack of large, dynamic meandering channels and enhanced (relative to the present) fluvial activity is not (yet) inconsistent with the Quaternary record. Maroulis (2000) records evidence of some fluvial activity in the interstadial at ~160-170ka BP, and there is also evidence of fluvial activity in Oxygen Isotope Stage 3 (60-32ka BP). The record from the Cooper Creek is generally consistent with records from other areas of Australia and the palaeohydrology of Lake Eyre (Nanson et al., 1988; Nanson et al., 1992; Kershaw and Nanson, 1993; Magee et al., 1995; Nanson et al., 1995; Nanson et al., 1998; Croke et al., 1999; Nanson and Tooth, 1999; Maroulis, 2000).
3.4 River characteristics

3.4.1 Hydrology

The Cooper Creek is an ephemeral stream which flows only intermittently, and like other Channel Country streams such as the Diamantina and Georgina Rivers, drains an exceptionally large basin with a very low gradient and transports a principally fine, suspended sediment dominated load. Arid, inland central Australia is poorly gauged with an average of one gauging station per 350,000 km\(^2\) (McMahon, 1979). There are four gauges in the study reach, two of which, Currareeva at the upstream end of the reach and Nappa Merrie at the downstream end, have reliable records from 1967 onwards while the others have shorter and less complete records (Knighton and Nanson, 1994a). The hydrelogy of the system has been studied by Knighton and Nanson (1994a, 2000). Selected flow characteristics from the Currareeva and Nappa Merrie gauging stations are given in Table 3.4, and examples of hydrographs from each of the stations are given in Figure 3.6. The distance between the two stations is ~420 km.

Table 3.4: Selected hydrological characteristics of the Cooper Creek

<table>
<thead>
<tr>
<th>Hydrological characteristic</th>
<th>Currareeva</th>
<th>Nappa Merrie</th>
</tr>
</thead>
<tbody>
<tr>
<td>Mean annual discharge (10(^6) m(^3))</td>
<td>3,050</td>
<td>1,257</td>
</tr>
<tr>
<td>Standard deviation, mean annual discharge (10(^6) m(^3))</td>
<td>5,020</td>
<td>2,830</td>
</tr>
<tr>
<td>Drainage area (km(^2))</td>
<td>150,220</td>
<td>236,985</td>
</tr>
<tr>
<td>Mean annual flood (cumecs)</td>
<td>96.7</td>
<td>40</td>
</tr>
</tbody>
</table>

Like precipitation, runoff in the Cooper Creek and the Lake Eyre Basin is very variable. The variability of runoff is generally higher in Australia than it is in the rest of the world (Finlayson and McMahon, 1988), and the variability of runoff in the more arid areas of the Lake Eyre Basin equals the highest in Australia (DNDAWRC, 1978). At both Currareeva and Nappa Merrie the standard deviations of mean annual discharge (Table 3.4) are significantly higher than mean annual discharge itself, and the hydrographs in Figure 3.6 show sometimes long periods without flow. This variability is characteristic of the basin as a whole, and modeling results suggest that Lake Eyre has received flow only once every two years over the last century (Kotwicki, 1986). The largest flood of the Cooper Creek on record occurred in 1974, when major flooding occurred throughout Queensland and most of New South Wales (Douglas, 1979). The headwaters of the Channel Country streams received up to 1 m of rainfall due to influence of an exceptionally strong monsoon which penetrated much further southwards into the interior of the continent than normal (Baker, 1986). In the first three months of 1974 the cumulative discharge of this flood at Innamincka was 8.6 times the mean annual discharge.
Figure 3.6: Selected relative flow magnitude ($Q/Q_{peak}$) curves from the flow records of the Currareeva and Nappa Merrie gauging stations (Knighton and Nanson, 1994a, p.145).
there, and maximum daily averaged discharges at the flood peak reached 24,974 cumecs at Currareeva and 5,812 cumecs at Nappa Merrie. The flood was spectacular in both its spatial extent and magnitude, and satellite imagery of the flood has been studied by both Robinove (1978) and Baker (1986).

As the distribution of precipitation over the basin would suggest, flow in the Cooper Creek is mostly generated in the wetter north of the catchment where runoff is much higher (DNDAWRC, 1978). Inputs of water from the Thompson and Barcoo Rivers are undoubtedly responsible for supplying most water to the Cooper, especially the larger flows that reach Lake Eyre, and Knighton and Nanson (1994a) have shown that discharges at Currareeva and Nappa Merrie are very highly correlated. The Cooper is not just a conduit for flow from the wetter north, however, and also receives flow from tributaries along the study reach, from local rainfall on the floodplain and from sheet-flooding of the valley side. Local rainfall onto the floodplain itself may cause localised inundation but is of relatively little significance to river flows because of the size of the floodplain and basin, and to greatly influence channel flows precipitation events must extend over wider areas than just the floodplain. Sheet-flooding of the valley side does not provide significant amounts of water to the channel although like local rainfall it may cause localised inundation. Tributary inputs along the study reach may provide large volumes of water to the Cooper. The Wilson River is the largest tributary to the Cooper in the study reach with a catchment area of approximately 25,000 km². LANDSAT imagery of the Wilson River flood event of January 95 suggests that inputs of water from the Wilson River and some other large tributaries may strongly influence the flow of the Cooper Creek (Figure 3.7).

Like periods of no flow, periods of continuous flow may be relatively long and the hydrograph from 1969 in Figure 3.6 shows continuous flow at Nappa Merrie from mid April to late August. Many flow periods are caused by single precipitation events, however the periods of largest, most sustained runoff are due to the coincidence and clustering of a number of events (Knighton and Nanson, 2000b). For example, in 1950 following an extremely wet year in 1949, heavy summer rains were followed in March by the incursion of three tropical cyclones in succession and then four months of average rains (Wopfner and Twidale, 1967). This concatenation of events caused sustained periods of high flows in all the Channel Country rivers and Lake Eyre North to fill to the second highest level on record (Kotwicki, 1986).
Figure 3.7: Landsat TM false colour composite image (21/1/95) of the Wilson River (WR) in flood. Floodwaters appear electric blue and the floodwave from the Wilson has reached the confluence with the Cooper Creek. Some floodwaters are seen on the eastern side of the Cooper Creek floodplain upstream from the flooding of Windula Creek (WC) and Cunnawulla Creek (CC).
3.4.1.1 Transmission losses

Apparent from Table 3.4 is the large difference between both mean annual discharge and the mean annual flood at Currarreeva and Nappa Merrie due to transmission losses between the two stations (Knighton and Nanson, 1994a). Transmission losses often exert a significant influence on the hydrology of dryland rivers (e.g. Walters, 1990; Hughes and Sami, 1992), and are generally caused by four factors: evapotranspiration; evaporation which is greatest at times of shallow overbank flow and during summertime; infiltration into the channel boundary and floodplain surface; and the storage of water in pond-like sump areas (Knighton and Nanson, 1994a). All factors except evapotranspiration significantly contribute to transmission losses on the Cooper Creek. When the entire floodplain is inundated flow expands to 60km wide and a huge surface area is exposed to evaporation. Evaporative losses are augmented by the fact that the peak flow period of the Cooper Creek is the extremely hot summertime when temperatures exceeding 45° are the norm. The cracked channel boundary and floodplain surface soils absorb water during the initial inundation phase before the soils seal and become impermeable. Although some larger channels extend down to the contact between the mud unit and the sand Katipiri Formation there is little or no percolation of water into the underlying sand sheet because the surface of the sand layer is quickly sealed by the deposition of the fine particles in the interstices between the sand grains (Knighton and Nanson, 1994a). Storage also contributes to transmission losses over the reach. Lake Yamma Yamma covers an area of approximately 900km² and satellite imagery shows that the lake fills during large flood events and may remain partially filled for months after flooding has subsided. While not particularly deep, even filled to an average depth of only 1m it is capable of storing 900,000,000m³ of water which is approximately 70% of the total annual discharge of the Cooper at Nappa Merrie. Knighton and Nanson (1994a) found that transmission losses varied non-linearly with stage, and that above a threshold flow of 25% duration transmission losses exceeded 75% of flow. Transmission losses are relatively low at low flow stage when flow is confined to the major anabranching channels, and rise sharply as smaller channels become active and the flow invades the floodplain surface.

3.4.1.2 Travel times

Travel times of flood peaks through the study reach vary markedly between approximately 400 and 1200hours (Knighton and Nanson, 2000b). This corresponds to (straight line) average speeds of between \( \approx 0.3 \text{ms}^{-1} \) and \( \approx 0.1 \text{ms}^{-1} \) or \( \approx 1.06 \text{kmh}^{-1} \) and \( \approx 0.36 \text{kmh}^{-1} \). Travel times appear to be related to flood magnitude (Knighton and Nanson, 2000b), and the lowest travel time recorded over the study reach was that of the largest flood on record which occurred in 1974. Stage height data from the Bureau of Meteorology (1996) show that the flood peaked at
Curareeva at 9am on the 2nd of February and at Nappa Merrie 420km downstream at 8pm on the 16th of February, traversing the study reach with an average speed of 1.06kmh⁻¹.

3.4.2 Floodplain landforms

The floodplain displays a complex array of landforms whose intricacy and complexity has attracted a good deal of comment and interest (e.g. Whitehouse, 1944, 1948; Mabbutt, 1967, 1977; Rundle, 1977; Rust, 1978; Veevers and Rundle, 1979; Rust, 1981; Rust and Legun, 1983; Baker, 1986; Rust and Nanson, 1986; Nanson et al., 1986; Nanson et al., 1988; Knighton and Nanson, 1994b; Knighton and Nanson, 2000a; Tooth and Nanson, 2000a). As mentioned previously, the floodplain of the Cooper Creek in the study area varies between 100m and 59km in width over its ~400km length, not including the cul-de-sac offshoot of Lake Yamma Yamma. Floodplain slope varies between 0.0001 and 0.0002 and averages 0.00015 over most of the reach. Floodplain relief, with the exception of isolated dunes, is very low and generally does not vary by more than a few metres across kilometres of floodplain. The floodplain is composed primarily of fluvial depositional units and landforms with some aeolian components (e.g. Rundle, 1977; Rust, 1981; Nanson et al., 1988; Gibling et al., 1998; Maroulis, 2000). Prominent aeolian sand dunes up to approximately 10m high are found in a number of locations on the floodplain surface. Analysis in this study revealed that dunes tend to occur in greater numbers in wider areas of floodplain, and study of floodplain dunes away from the valley side in the middle and lower areas of the study reach by Maroulis (2000) showed that many are source-bordering dunes which formed next to prior sand transporting channels and are not currently active. Luminescence dating suggests the dunes are being both degraded by erosion and buried by accretion of the floodplain around them (Maroulis, 2000). Nearer the valley sides the presence of dunes on the floodplain may be due to the supply of sand from the valley side or the expansion of the floodplain into adjacent dunefields. As stated previously, the pattern of Quaternary sedimentation here has been for infilling of very shallow, saucer-like structural lows and valley widths may expand markedly in response to relatively small depths of accretion.

Alabyan and Chalov (1998) recognise three levels of fluvial relief: the floodplain surface, floodplain-surface channels, and low water channels. The Cooper Creek exhibits a range of coexisting channel patterns and displays all three levels of fluvial relief with a network of anastomosing low water channels, well developed floodplain-surface channels and significant areas of unchanneled floodplain (Figure 3.8). Common to all three levels is an extremely slow rate of change. A study of the 40 year long aerial photograph record reveals no discernible
Figure 3.8: Schematic illustration of the floodplain landforms and coexisting channel patterns of the Cooper Creek (from Tooth and Nanson, 2000a, p.5).

Figure 3.9: The anastomosing channel network of the Cooper Creek. Large, narrow and deep anastomosing channels here co-occur with the braid-form floodplain-surface channel pattern, described later in this Chapter and in Chapter 7, as they do generally. The figure also illustrates some of the complexity and discontinuity of the anastomosing channel network.
change in any channels despite the occurrence of a number of large floods events in the period, and vertical accretions rates on the floodplain are known to be very low (e.g. Nanson et al., 1988; Maroulis, 2000). The characteristics of the various channel patterns will be briefly described and relevant literature reviewed.

3.4.3 Anastomosing channels

3.4.3.1 Continuity

A network of relatively narrow and deep anastomosing channels of various sizes is inset into the floodplain surface which is considered by most authors to be a contemporary pattern formed under the present flow regime (Whitehouse, 1948; Rust, 1981; Rust and Legun, 1983; Nanson et al., 1986; Rust and Nanson, 1986; Knighton and Nanson, 1994b; Tooth and Nanson, 2000a) (Figure 3.9). The largest and deepest channels receive flow at all discharges, while the smallest channels operate only near bankfull which is attained here by the two year flood event. Channels divide and rejoin at high angles, and at first glance there appears to be no direction to the pattern with it being difficult to determine flow direction from aerial photographs. At the small scale the dominant impression of the anastomosing network is one of discontinuity. However, the connectivity of the system varies and the anastomosing network does form a sufficiently continuous system to enable lower bankfull and sub-bankfull flows to be transported down the study reach solely by the anastomosing channels. That said, not only individual channels but extensive sections of anastomosing network may be isolated from one another. For example, the extensive anastomosing network feeding and draining Bogala and Naccowlah Waterholes on the eastern side of the floodplain north of the confluence of the Cooper and Wilson Rivers covers an area exceeding 500km² and is operative only during overbank flow events. The network here is isolated from the anastomosing network on the western side of the floodplain which is dominant and feeds flow through Meringhina, Didhelginna and Goonbabina Waterholes.

As well as changing in size at channel junctions, anabranches often vary in size along their length, in places dying out and in others growing in prominence. Knighton and Nanson (2000a) traced a number of anastomosing anabranches through reaches of changing channel size and found that variations in channel prominence are related to variations in floodplain topography. Anabranches grow in size and number where surveys of the floodplain surface suggested floodplain flow tends to concentrate, and they decrease in size where flow diverges. Size discontinuity is not dominant however and many anabranches remain relatively constant in size over lengths many hundreds of times their width.
3.4.3.2 Cross sectional morphology, channel boundary texture and sediment load

In cross section anabranches are narrow, deep, and simple and canal-like in form. The geometry of 67 surveyed anabranch and waterhole channel cross sections ranging from 11m to 106m in width was analysed. The study of aerial photographs revealed that the largest anabranch found in the study area is 123m wide (Eulberti Waterhole), and so this range of widths covers most of the range found in the study area. Maximum channel depths observed range up to 7.5m in the alluvial reaches of the river, and while the depth of the single, bedrock confined channel at the Innamincka Dome which marks the downstream limit of the study reach is considerably greater at about 27m (G. C. Nanson, per. comm., 2000) this section is atypical of the reach as a whole (Baker, 1986). The relationship between anabranch width and average depth are shown in Figure 3.10. Although a nonlinear relationship shows no statistically significant improvement on the linear relationship, the distribution of values in Figure 3.10 suggests that the rate of increase in channel depth with width tends to decrease at the upper range of channel widths found. This is possibly because the thickness of the mud layer generally does not exceed 6m and this sets a limit on channel depth because the upper surface of the underlying sand sheet is highly indurated and resists erosion (Nanson et al., 1988).

Anastomosing channel banks generally vary in maximum slope from 20° to 70°, with maximum bank slopes up to 90° in a small number of cases. No overhanging channel banks were observed, and no evidence of large-scale mass movement failure was seen on any of the channels observed or surveyed in the field. The highest channel banks occur along the large waterholes where bank heights may reach up to 7.5m. Channel banks and channel boundaries generally are composed of cohesive sediments which sediment size analysis showed do not differ significantly from floodplain sediments which are usually composed of 35-60% clay, 40-60% silt and 4-9% sand. The channels transport a predominantly fine-grained sediment load consistent with their fine-grained channel boundaries. The largest sediment size fraction transported by the anastomosing channels is sand and observations of dry channel beds indicate that a small sand load is transported by most anabranches. Sand only occurs here as a thin drape over the underlying cohesive boundary materials and no thick channel fill deposits of coarse sediment like those in the anastomosing reaches of the Alexandra, North Saskatchewan and Columbia Rivers (e.g. Smith and Putnam, 1980; Smith and Smith, 1980; Makaske, 1998) are found here. The study of dry channel beds also reveals that, at least at low flows, the sand is organised into well sorted patches or ribbons which contain negligible amounts of finer sediment and which overlie the cohesive channel boundary. The sand patches generally only
Figure 3.10: The anabranch channel width-mean depth relationship. The dotted lines delimit the range of data.

$r^2=0.58, p<0.0001, n=67$.
Mean $W/d_m=2.8$
cover a relatively small percentage of the channel bed, and their occurrence is consistent with Rust and Nanson's (1986) statement that the sand load is transported as bedload in traction. Where sand is present small-scale bedforms are formed and sand patches observed following a small flushing flow in the main channel showed small-scale ripple development with ripple heights of approximately 20mm and mean spacing of approximately 70mm. Where sand is absent hydraulic bedforms composed of mud aggregates may be formed and have been observed following flood events (Rust and Nanson, 1989; Maroulis and Nanson, 1996). As the channel bed dries the mud bedforms are destroyed by the cracking of the bed. Measurements of suspended sediment load during the February-March 1997 flood event indicate that concentrations of suspended sediment are very low. Sediment concentrations were inversely proportional to flow stage and varied between 427mg/l at about half bankfull to about 132mg/l at bankfull.

3.4.3.3 Planform and the characteristics of the channel-floodplain boundary

Pools and ripples are absent from the channels and within-channel benches are generally absent. In Brice's (1984) terminology the anabranches are 'sinuous canaliform' channels and remain constant in width even around bends in the channel. Channel cross sections are asymmetrical at bends and resemble meandering channel cross sections with the thalweg displaced to the steeper outer bank of the bend and a more gently sloping inner bank. Sedimentologically however, the channels do not form recognisable point bars and excavations of channel sections has revealed no sedimentary evidence of lateral migration (Rust, 1981; Nanson et al., 1988). This may reflect a real lack of lateral migration or may be because the predominantly cohesive load of the channels means that points bars are formed but are indistinguishable in texture from overbank deposits. As mentioned in Section 2.4.2.1, within-channel benches are generally absent from the channels. Preliminary results reported by Beaman (1995, unpublished) suggest that the size of anabranches is related to their sinuosity with smaller anabranches being more sinuous than larger anabranches. More detailed analyses of anastomosing channel planform were carried out here and are reported in Chapter 4.

The anabranches are foci for vegetation growth as many channels in arid regions tend to be. Vegetation tends to be more dense around larger anabranches, and large anastomosing channels are lined by mature coolibah trees (Eucalyptus microtheca) up to 8m high which concentrate there because soil moisture is important for seed germination, establishment and longevity (Williams, 1979). Lignum (Muehlenbeckia cunninghamii) and flowering lignum (Eremophila polyclada) bushes 1-2m high also cluster there for similar reasons. Eucalypts are
schlerophyllous and adapt to drought stresses not by economising on water use but by developing hard tissue which enables them to endure severe and permanent wilting without lasting damage (Williams, 1979). Eucalypts have wide ranging, extensive (Cannon, 1911) root systems with coarse ultimate rootlets occupying a large soil volume at low densities (Pryor, 1976; Williams, 1979). The within-channel vegetation commonly found in many Australian dryland rivers (e.g. gracme and Dunkerley, 1993; Wende and Nanson, 1998; Tooth and Nanson, 2000b) is absent here, and channel beds and banks are invariably devoid of both ephemeral or perennial vegetation.

There has been some debate over whether or not the anastomosing channels are flanked by levees. The occurrence of levees along the channels has been commented on by a number of authors (e.g. Mabbutt, 1977; Rundle, 1977; Nanson et al., 1986; Rust and Nanson, 1986; Nanson et al., 1988; Gibling et al., 1998), however they have not been studied in great detail and in a recent paper on waterhole formation Knighton and Nanson (2000a) stated that “the anastomosing channels rarely have levees” (p.107). Some levees are clearly apparent during overbank flood events when they emerge as ribbon-like islands above the surrounding inundated floodplain. As the review in Chapter 1 showed, the presence or otherwise of levees may be a significant factor in determining the process of anastomosis formation, and the generality of levee occurrence will be examined in more detail later.

3.4.4 Waterholes

A notable feature on many of the channels of the Cooper Creek are the larger, expanded channel sections known as waterholes (Figure 3.11). Waterholes occur on all the Channel Country rivers and are sections of channel which hold water at low flow and during periods when there is no flow at all. Historical evidence suggests that some waterholes in the study reach have held water continuously since European settlement of the area over 100 years ago despite sometimes long periods without flow (Knighton and Nanson, 2000a). Waterhole capacities are generally many times that of any channels feeding or draining them, and their origin has been the subject of some debate (Whitebuuse, 1948; Rundle, 1977; Knighton and Nanson, 1994b; Knighton and Nanson, 2000a). Knighton and Nanson (1994b) divided waterholes into 3 classes; those formed between dunes and the valley side, those formed between dune constrictions on the floodplain, and those formed in neither situation within the anastomosing channel network. Waterholes can be more simply divided into two types with very different morphologies: ‘floodplain-surface’ (FS) waterholes which are formed at overbank flow constrictions on the floodplain and which are not well integrated with the
Figure 3.11: (a) Tooley Wooley (T) and Little Tooley Wooley (LT) floodplain-surface Waterholes, flow from the north-east to south-west. Note the radial dispersion of flow from the downstream end of Tooley Wooley Waterhole. (b) Didhelginna Waterhole and (c) Bogala Waterhole, flow from north to south.
anastomosing network, and ‘anastomosing channel’ (AC) waterholes which are essentially expanded sections of channel integrated into the main anastomosing channel network. This distinction corresponds to waterholes but is less ambiguous than that made by Rundle (1977) between ‘short’ (floodplain-surface) and ‘long’ (anastomosing channel) waterholes.

3.4.4.1  Floodplain surface (FS) waterholes
The constrictions causing FS waterholes to form are mostly aeolian dunes and occasionally the valley side (Knighton and Nanson, 1994b). These waterholes are very wide with high width-depth ratios, are short relative to their widths hence Rundle (1977) calling them ‘short’ waterholes, and are of very low sinuosity or geometrically straight. Tooley Wooley Waterhole (Figure 3.11a) is typical of FS waterholes although larger than most. It is formed between a dune constriction on the floodplain surface, is 3km long, 160m wide at its widest point, and has a width-maximum depth ratio of approximately 50 (Knighton and Nanson, 2000a). levees are absent from FS waterholes, and so is the relatively large and dense woody vegetation found along major anastomosing channels with only much smaller, less woody and possibly more ephemeral vegetation found bordering floodplain scour waterholes. This suggests that the floodplain surrounding FS waterholes and FS waterholes themselves receive flow much less frequently and reliably than the areas surrounding anastomosing channels.

FS waterholes form when overbank flow during flood stage concentrates and causes a shallow, wide channel to be incised into the floodplain surface (Rundle, 1977; Knighton and Nanson, 1994b). Rundle (1977) stated that floodplain surface waterhole formation occurs when the protective cover of surface vegetation is breached by scour from flood flows, however, this is not correct as away from the main anastomosing channels the floodplain surface is only sparsely vegetated, most of the soil surface is exposed, and vegetation does not bind the surface of the floodplain together in the way Rundle implies. They tend to occur on the wider areas of floodplain only because the narrower floodplain areas tend to be better connected to the anastomosed network meaning that any waterhole forming there is likely to be well integrated with the anastomosing channel network, and because there are significantly fewer dune constrictions present on narrower areas of floodplain due to the greater tendency for the dunes to have been eroded and reworked by either the present anastomosing or former sand-bed meandering channels there. It is clear that the morphology of Tooley Wooley and Little Tooley Wooley Waterholes reflects the pattern of floodwater distribution around them. As flow approaches the line of dunes across its path it beings to converge and becomes deeper and faster, and small channels are incised into the floodplain surface which become progressively
larger toward the constriction reflecting the increased focusing of floodwaters on the channels. Both this focusing of flow on FS waterholes upstream and the dispersion of flow from them downstream are reflected in the orientation of channels near the waterholes which tend to focus in on the upstream end of the waterholes and to radiate away from their downstream end (e.g. Figure 3.11a).

3.4.4.2 Anastomosing channel (AC) waterholes

The degree of integration of AC waterholes varies with some more densely connected to the anastomosing network than others, however, all are fed and drained by at least one anastomosing anabranch which persists for some kilometres beyond the waterhole itself. In cross sectional geometry AC waterholes are narrow and deep like anastomosing channels (Figure 3.9) and are generally flanked by what seem to be levees. The length of AC waterholes varies greatly. The longest in the study reach is Eulberti Waterhole at 17.3km while there is no well defined minimum length since they are simply expanded sections of anastomosed channel. A notable characteristic of AC waterholes is the abruptness with which they tend to begin and end. AC waterholes generally begin at confluences in the anastomosing channel network (Knighton and Nanson, 1994b) (e.g. Didhelginna, Bogala, Meringhina and Goonbabina Waterholes), and only rarely does an AC waterhole slowly peter out; more usually they terminate at some sort of abrupt planform discontinuity such as a particularly sharp bend (e.g. Bogala) or a channel bifurcation (e.g. Didhelginna, Meringhina, Eulberti and Bogaller). As Richards (1980) notes, most variation in channel dimensions usually occurs at nodes in the channel network, however, the width and capacity of AC waterholes often varies manyfold without significant tributary inputs or distributary outputs and over distances too short for evaporation or infiltration losses to be significant and the capacity of Didhelginna Waterhole (Figure 3.11b) varies from 70m² at its inception to 353m² at a point exactly half its 2.5km length downstream. Variations in AC waterhole form appear to be systematic, and in contrast to how they have been characterised elsewhere (c.f. Rundle, 1977) AC waterholes do not become gradually smaller downstream but instead typically expand in size from their point of inception until between about half to two-thirds of their length and then begin to contract at a greater rate than they expanded (e.g. Figure 3.11b).

AC waterholes were thought by Rundle (1977) to be relict features which, due to their capacities being much larger than those of their feeder channels, behaved in a quasi-lacustrine manner during flood events, however more recent work has concluded that like channel patterns here generally they are contemporary features (Knighton and Nanson, 1994b; Knighton and Nanson, 2000a). The contemporary explanation proposed by Knighton and
Nanson (2000a) regards AC waterholes as in equilibrium with the discharge supplied to them at flood stage by both the anastomosing feeder channels and overbank flow from the floodplain surface. Observations of flow patterns in an around expanding and contracting AC waterholes have proved difficult to obtain because of the irregularity of flood events and the tendency for flooding to be a region-wide event which often not only cuts the roads across the Cooper Creek floodplain but also isolates the surrounding areas. An attempt was made to collect such data during the flood event of March 2000, however flooding in the surrounding region cut roads to the area and meant that the salient parts of the floodplain were unreachable when the flood-wave passed through. In the absence of relevant flow data three main strands of evidence support the contemporary explanation of waterhole origin: 1) they fill to bankfull about once every 2-3 years; 2) they occur at points of convergence in the anastomosing channel network; and 3) and they exhibit contemporary sediment splays at their downstream ends. The tendency of waterholes to occur at points of flow convergence supports the contemporary hypothesis because it suggests that waterhole position is determined by the current configuration of the floodplain surface (Knighton and Nanson, 1994b). The occurrence of sediment splays at the downstream ends of waterholes is perhaps the strongest sedimentary evidence for a contemporary origin of waterholes as it indicates contemporary scour or at least the contemporary transport of bedload out of the waterholes (Knighton and Nanson, 1994b, 2000a). Splays here are much more subdued features than those formed by large rivers like the Yellow River (e.g. Van Gelder et al., 1994) or many anastomosing channels such as the lower Saskatchewan River (Smith and Pérez-Arlucea, 1994) and Columbia Rivers (Smith, 1983), and are often undetectable on the ground due to their low relief. Nanson et al. (1988) reported that a prominent splay at the downstream end of Pritchilla Waterhole was composed of sandy-mud and stated that the high sand component present was excavated from the waterhole during flood events. The relative coarseness of the splay deposits makes them clearly visible on aerial photographs.

Meringhina Waterhole (Figure 3.12) is unusual in that it is the focal point of almost all the anastomosing channels across the floodplain which implies that almost the entire cross sectional bankfull flow of the Cooper passes through Meringhina Waterhole (Knighton and Nanson, 2000a). Meringhina Waterhole is also unusual in that it is the only channel in the study reach along which scroll bars occur, indicating that the channel is laterally migrating. The median annual flood at Meringhina is estimated to be 330 cumecs, and the average bankfull cross sectional area of Meringhina Waterhole is 355m² (Knighton and Nanson, 2000a). The mean capacity of the waterhole is within 15 % of the channel capacity predicted for a channel with cohesive bed and banks and a discharge equal to 300cumecs from regime equations which
Figure 3.12: Meringhina Waterhole
Knighton and Nanson (2000a) state is sufficiently close to support a particular form of the contemporary hypothesis which regards waterholes as discontinuous sections of an equilibrium single channel which cannot be maintained.

Knighton and Nanson state that AC waterholes are “self maintaining scour features” (2000a, p.101) and that they resemble the ‘chain of ponds’ morphology described by Eyles (1977). Eyles (1977) found that Birchams Creek in the Great Dividing Range (Australia) incised irregularly sized and spaced ponds up to 2.15m deep and 150m² in surface area into the valley floor. The ponds tended to be elliptical in form, elongated with the slope of the valley floor, and were not connected to each other by channels. Eyles stated that pond formation seemed to be related to the structure of the soil profile, and that ponds have a minimum depth of 0.3m which represents the depth of the resistant topsoil which must be breached before a pond can form. After breaching has taken place ponds scour rapidly into the easily erodible subsoils and expand laterally as undercutting of the resistant topsoil leads to bank collapse. Chains of ponds represent a discontinuous channel form that is nonetheless adjusted to present conditions and the similarly discontinuous nature of AC waterholes led Knighton and Nanson (2000a) to suggest that AC waterholes on the Cooper Creek may represent an analogous system also formed by and adjusted to contemporary conditions, with individual waterholes expanding upstream by a headcutting process and terminating because, for some reason, they are unable to continue transporting the sediment they erode.

The occurrence of what are apparently levees around channels concentrating flow from the floodplain is perhaps surprising as levees are generally formed by the transfer of sediment overbank from the channel to the floodplain and so typically reflect the divergence of flow from channels (e.g. Kesel et al., 1974; Pizzuto, 1987; Marriott, 1992; Guccione, 1993; Asselman and Middelkoop, 1995; Brierley et al., 1997; Cazanacli and Smith, 1998). The morphology and texture of the channel-floodplain boundary around expanding and contracting channels will be studied in detail in Chapter 6.

### 3.4.5 Floodplain-surface channels

The floodplain-surface channels of the Cooper can be divided into reticulate and braid-form pattern types.
3.4.5.1 Reticulate networks

Reticulate networks are an unusual type of channel pattern which occurs over approximately 40% of the Cooper Creek floodplain (Figure 3.13). The reticulate pattern has an extremely high drainage density and is defined by a prevalence of approximately right-angled confluences and bifurcations between channels which produce a complex interlaced network of jagged, dissected polygon islands. The range of reticulate channels found is given in Table 3.5 and sample cross sections in Figure 3.14.

Table 3.5: Range of reticulate channel morphological characteristics found (n=16).

<table>
<thead>
<tr>
<th>Characteristic</th>
<th>Range</th>
<th>Mean</th>
</tr>
</thead>
<tbody>
<tr>
<td>Width W (m)</td>
<td>1.8-8.5</td>
<td>5.3</td>
</tr>
<tr>
<td>Maximum depth d_m (m)</td>
<td>0.18-0.63</td>
<td>0.42</td>
</tr>
<tr>
<td>Average depth d_av (m)</td>
<td>0.12-0.31</td>
<td>0.20</td>
</tr>
<tr>
<td>W/d_av</td>
<td>14.6-58.3</td>
<td>26.2</td>
</tr>
<tr>
<td>Cross sectional area (A) (m²)</td>
<td>0.2-2.6</td>
<td>1.1</td>
</tr>
</tbody>
</table>

Reticulate channels occur in areas with prominent gilgai development and are inset into depressions between gilgai mounds with the bankfull level of the channels being at the level of the base of the gilgai depression as shown schematically in Figure 3.15. The planform of individual reticulate channels tends to be very irregular and to exhibit little smoothness, and the reticulate pattern differs from the anastomosing pattern in its much smaller scale, the greater intensity of the divisions between channels, and the common occurrence of terminal distributary channels. Reticulate patterns occur both proximal to and distant from the valley side, and both in areas where floodplain flow is confined by dunes and areas where flow is unconstricted.

Previous study of the reticulate networks has considered their morphology and formation only very briefly. Most attention has been focussed on the co-occurrence of the reticulate pattern and prominent gilgai development and the apparent similarity between the two patterns has led to the assertion that the channels are only partly fluvial in origin and a product of the interaction between fluvial processes and gilgai formation on the floodplain (Mabbutt, 1967; Rundle, 1977). Rundle (1977, p.57) states that it “seems that the gilgai and the drainage must have developed simultaneously, the swelling mounds controlling the distribution of floodwaters.” The fact that the bankfull levels of reticulate channels are at the level of the base of gilgai depressions implies that the channels are incised into the depression which is probably facilitated by the greater cracking and disruption of the soil that occurs there (Figure 3.16). The complex, undulating floodplain topography caused by the presence of gilgai promotes channel
Figure 3.13: Examples of reticulate network patterns showing the intricate, spacefilling nature of the patterns.
Figure 3.14: Examples of reticulate channel cross sections. Note that the channels are inset into gilgai depressions. Dashed lines indicate bankfull levels.

Figure 3.15: Schematic diagram showing the location of reticulate channels in depressions between gilgai mounds. The bankfull level of the reticulate channel approximates the level of the gilgai depression and reticulate channels do not occupy every gilgai depression.

Figure 3.16: An example of the greater intensity of cracking of reticulate channel beds (C) compared to the surrounding, slightly higher areas of floodplain surface (S). Higher intensities of cracking and boundary disruption caused by exposure to higher frequencies of wetting and drying cycles probably facilitate channel formation.
division and recombination leading to interconnected networks of channels being formed. A similar link between pedogenic processes and reticulate patterns to that proposed by Mabbutt (1967, 1977) and Rundle (1977) has been proposed by White and Law (1969). They studied alluvial clay depressions in Iraq and Sudan which form a gilgai type surface morphology. As they develop, White and Law state that shallow surface depressions are occupied by what would be termed here reticulate channels although they termed them interlocking or 'tabra' channels after the soil type on which they occur. On the Cooper floodplain the undulating soil surface is not due to the differential soil swelling in gilgai formation but to the opposite process, differential soil compaction which takes place due to soil weathering and causes the removal of void spaces and the re-orientation of clay minerals (White and Law, 1969).

The reticulate pattern was thought by previous authors to be confined to lower, sump areas of the floodplain where water ponds (Whitehouse, 1948; Rundle, 1977). This was not based on survey data, but rather on the assumption that gilgai development would be greater in these areas because they tend to be wetter. However, as we shall see later the situation is more complex than this and reticulate networks are not confined to low areas of the floodplain which implies in turn that, since reticulate channels and prominent gilgai patterns co-occur, neither are prominent gilgai patterns. Chapter 7 will examine the distribution of reticulate channel patterns over the floodplain surface and the controls on their occurrence.

3.4.5.2 Braid-form patterns

The braid-form pattern is the most extensive floodplain-surface pattern type found here and occupies 44% of floodplain in the study reach (Figure 3.17). The braid-form pattern appears somewhat similar in appearance to the pattern of a typical braided channel and has been described as ‘braided’ by a number of authors (e.g. Rundle, 1977; Rust, 1978; Rust, 1981; Nanson et al., 1986; Maroulis and Nanson, 1996) the term ‘braid-form’ is used here instead because it is unclear that the form and process of the pattern is consistent with that recognised as characteristic of braided channels (e.g. Bridge, 1993; Ferguson, 1993) as the use of the term ‘braided’ would imply. This question will be addressed in Chapter 7. The braid-form pattern is composed of elongate floodplain islands between which small, wide and shallow channels flow (Table 3.6 and Figure 3.18). As Figure 3.18 illustrates, braid-form channels do not possess levees, bankfull levels are generally not local floodplain surface high points, and the channels are inset slightly into the floodplain surface. The sinuosity of braid-form channels is generally very low and the channels are generally straight or moderately sinuous. There is little small scale variability of channel planform, and where significant sinuosity is developed it is mostly
Figure 3.17: Oblique aerial photographs of the braid-form floodplain-surface pattern showing the subdued nature of the braid-form islands, the wide shallow dividing channels, and apparent similarities with the pattern of typical braided channels.
Figure 3.18: Typical braid-form channel cross sections and an illustration of a typical braid-form channel in the field. Note that the channels appear to be depressed into the surrounding floodplain. Dashed lines indicate bankfull levels.
due to large scale, irregular wanderings of the channels. A comparison of aerial photographs from 1948 and 1988 showed no discernible change in channel position or pattern geometry, implying that rates of morphological change of the braid-form pattern must be extremely slow in contrast to the dynamism typical of braided channels (e.g. Bridge, 1993; Ferguson, 1993; Murray and Paola, 1994). Braid-form islands range between 400-3500m in length and 90-630m in width. The relief of the pattern is very subdued, and island summits are generally no more than 1m above the beds of the braid-form channels bounding them.

Table 3.6: Range of braid-form channel morphological characteristics found (n = 19).

<table>
<thead>
<tr>
<th>Characteristic</th>
<th>Range</th>
<th>Mean</th>
</tr>
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<tbody>
<tr>
<td>W (m)</td>
<td>3.5 - 32m</td>
<td>15.16m</td>
</tr>
<tr>
<td>d_m (m)</td>
<td>0.13 - 0.50m</td>
<td>0.39m</td>
</tr>
<tr>
<td>d_av (m)</td>
<td>0.07 - 0.44m</td>
<td>0.19m</td>
</tr>
<tr>
<td>W/d_av</td>
<td>28.57 - 198.80</td>
<td>86.39</td>
</tr>
<tr>
<td>A (m^2)</td>
<td>0.43 - 14.08m^2</td>
<td>3.36m^2</td>
</tr>
</tbody>
</table>

The origin of braid-form pattern has been the subject of a deal of debate centred around a contemporary vs. relict argument. One of the earliest contributions was made by Mabbutt (1967) who termed the channels ‘floodways’ and noting that many originate and terminate at junctions with anastomosing channels concluded that they are contemporary features formed by overbank flow during flood events. Mabbutt compared the channels to the erosional swales or floodplain scour routes described by Thornbury (1968). Subsequent to Mabbutt’s work, relict hypotheses explaining the origin of the pattern gained acceptance. Rundle (1977), Rust (1981) and Rust and Legun (1983) considered the present form of the Cooper Floodplain to reflect what Rust (1981) termed ‘disequilibrium’ and proposed that the braid-form pattern is the surface expression of a relict, inactive sand-bed braided channel system over which the mud deposits of the contemporary floodplain had been draped. The braided channel system was said to have formed during a prior, wetter climate when the river had a larger, more persistent flow regime than exists at present. Rust (1981) stated that over time continued floodplain accretion would form a flat, featureless floodplain traversed only by anastomosing channels. Rundle’s (1977) hypothesis is broadly similar to that of Rust (1981) but recognises three distinct braided phases compared to Rust’s one. Rundle distinguished between two braid-form patterns on the basis of differences he identified in the scale of the pattern and was able to deduce three braiding phases from these as he considered that one of the phases did not need to be represented by a distinct pattern as it occurred in the “remote past” (Rundle, 1977, p.86). The smaller scale pattern identified was ascribed to the action of smaller flows than those...
which formed the larger scale forms, and Rundle proposed a chronology of pattern development based on morphological relationships between channel types on the floodplain surface. In contrast to Mabbutt (1967), Rundle states that small anastomosing channels are 'cut into' the smaller of the two braid-form patterns he detected and that where braid-form channels intersect anastomosing channels they run continuously across the anastomosing channels with no planform discontinuities. Rundle stated that small anastomosing channels also often occur as 'underfit' streams in the depressions around braid-form islands which he interpreted as infilled formerly larger braided anabranches (Rundle, 1977, p.82). Rust (1981) concurred with this interpretation, and both authors concluded that these morphological relationships implied that the braid-form pattern formed prior to the anastomosing pattern. The difference between Mabbutt's (1967) and Rundle and Rust's (Rundle, 1977; Rust, 1981) characterisations of the intersections between anastomosing and braid form channels is shown in Figure 3.19 and will be studied in Chapter 5. Rundle stated that the smaller braid-form pattern 'cuts across' the larger braid-form pattern which he took to imply that the smaller braid-form pattern formed later than the larger braid-form pattern. An attempt was made to replicate Rundle's classification of braid-form patterns, however even with reference to his original photographs and diagrams all attempted divisions of the braid-form pattern into different classes by scale were unsuccessful.

Like Mabbutt (1967), Nanson et al. (1986) and Rust and Nanson (1986) proposed that the braid-form pattern was a contemporary feature formed under the current, ephemeral flow regime. Augering of the floodplain surface by Rust and Nanson demonstrated that the positions of the braid-form channels is unrelated to the surface-form of the underlying sand sheet which led Nanson et al. (1986) and Rust and Nanson (1986) to conclude that the braid-form pattern was a surficial feature formed in equilibrium with present climatic conditions. Nanson et al. (1986), Rust and Nanson (1986), Maroulis and Nanson (1996) and Gibling et al. (1998) considered the braid-form pattern to be braided in form and process. The occurrence of braiding in low energy channels with cohesive boundary sediment like those found on the Cooper floodplain is unusual as braiding generally occurs in relatively high energy, coarse, non-cohesive load streams (e.g. Ferguson, 1993) and a recent theoretical model has suggested that an easily erodible channel boundary composed of non-cohesive sediments is a pre-requisite for braiding (Murray and Paola, 1994, 1997). Rust and Nanson (1989, p.294) state that the "apparent anomaly of a braided channel system composed of clay-rich mud is explained by the fact that the mud is mainly present in the form of sand-sized aggregates." Because these aggregates have a lower density than sand they can be more easily transported as traction load over low gradients and thus, although their load is composed of mud, Rust and Nanson state that
Figure 3.19: Differences in the relationships between braid-form channels (grey) and anastomosing channels (black) recognised by (a) Rundle (1977) and Rust (1981) who asserted that braid-form channels generally run continuously across anastomosing channels with no disruption to their planform, and (b) Mabbutt (1967) who stated that braid-form channels tend to originate from and terminate at anastomosing channels.
the rivers are in effect sandy braided systems. The Shield's criterion indicates that they can be transported by flow depths greater than 0.2m (Nanson et al., 1986) which often occur during flood events when flow depths over the floodplain surface may exceed 1m. The veracity of the description and interpretation of the braid-form channels as truly braided in form and process will be examined in Chapter 7.

3.4.5.3 Floodplain drainage channels

The networks of floodplain-surface channels are formed by flood flows from upstream but also tend to concentrate runoff from intense local rainfall and act as floodplain drainage networks. However, there are often large areas of undrained floodplain between these channels and large areas of unchanneled floodplain in the study reach overall. Gullying of the floodplain surface in response to runoff from local rainfall is rare because the extremely low slope of the floodplain means that the shallow surface flows from local rainfall events do not possess much erosive energy. Gullying does occur on some steeper surfaces where a threshold slope is exceeded and small gully networks are commonly observed on steeply sloping, convex, unvegetated bank tops (Figure 3.20).
Figure 3.20: Gullied convex channel bank top. Ruokaspek for scale is 45 cm tall.
4 ANASTOMOSING CHANNEL PLANFORM ANALYSIS

4.1 Introduction

Although in multiple channel systems the development of any individual anabranch is ultimately dependent upon its relationship to other anabranches (Richards et al., 1993), anabranches in anastomosing channel systems may develop their own width-scale flow patterns not directly dependent upon the flow pattern in other anabranches (Bridge, 1993) and their own planform patterns (Schumm, 1977). The planform of single channel anabranches is an important indicator of channel behaviour and, in the absence of braiding development, whether the channels are meandering or laterally stable. Freely meandering channels have been shown to develop characteristic relationships between channel size and meander form variables as a result of the meandering process itself (e.g. Leopold and Wolman, 1957, 1960; Dury, 1966, 1967; Schumm and Khan, 1972; Hey, 1976; Ferguson, 1981; Richards, 1982; Williams, 1986; Hickin, 1977) and tend to show planform elements characteristic of the meandering process like cut-offs, point bars and scroll-bars. Anabranches of anastomosing channels are generally referred to as being straight or sinuous and laterally stable (e.g. Knighton and Nanson, 1993, 1996), however, lateral migration of anastomosing anabranches has been recorded on the anastomosing Solimões, Japura (Baker, 1978) and Yangtze Rivers (Baker, 1986). It is also inherent in the anastomosing model of Schumm et al. (1996) formulated to explain channel dynamics on the Ovens and King Rivers in south-east Australia. Also in Australia Riley (1975) and T. Pietsch (pers. comm., 2000) have noted the occurrence of cut-offs, point bars and scroll bars indicating channel meandering on anastomosing distributary channels of the semi-arid Namoi-Gwydir distributary system, a low gradient, mud-dominated anastomosing channel system in New South Wales not unlike the Cooper Creek in a number of ways.

The anastomosing-channel model of Schumm et al. (1996) has been described in some detail earlier (see Section 1.4.1.6) and proposes that systematic changes in channel planform drive the avulsion process and the formation of anastomosis. The model states that following avulsion a small channel is developed which enlarges over time. Because this channel is low in sinuosity it takes a relatively direct path downslope on the floodplain giving it a relatively steep gradient and sufficient energy to laterally migrate. With time, meandering of the channel causes its sinuosity to increase and the gradient and energy of the channel to decrease until eventually the channel no longer has sufficient energy to meander or to transport the sediment that is supplied to it. This causes channel size to contract as sediment is deposited within the channel and channel planform to become static and underfit. The contraction of the channel forces
increasing amounts of flow overbank which triggers the incision of a new anabranch and an avulsion event. This model has clear predictions for the change in channel planform over time. As channels become older, sinuosity should increase. Small channels may represent both old channels being filled in and dying away, in which case they should be either high in sinuosity and underfit or young, newly incised channels in which case they should be low in sinuosity and, provided there has been sufficient time for meanders to develop, have meander morphologies typical of freely meandering channels. Larger channels represent channels of intermediate age and should therefore have intermediate to high sinuosities depending upon how close they are to the threshold slope at which they become incompetent and static, and also possess meander geometries typical of freely meandering channels. This is examined here and the channel dynamics of individual anabranches investigated.

In some respects anastomosing systems are simple systems on which to study planform because a number of variables which have been shown to influence channel planform properties are constant for all anabranches and so do not need to be considered as possible influences on planform variability, although channel age may vary between anastomosing anabranches and is a complicating factor in the study of planform variability. The slope of the floodplain, which has been shown to influence both the type of pattern developed and the characteristics of that pattern (e.g. Friedkin, 1945; Carlston, 1965; Schumm and Khan, 1971, 1972), varies only between 0.0001 to 0.0002 along the length of the floodplain and is mostly an approximately constant 0.00015. The texture of channel boundary sediments, another factor of importance in determining planform properties (Henderson, 1963; Schumm, 1967, 1968; Osterkamp, 1978; Carson, 1984; Ferguson, 1987) is also essentially constant for all anabranches here, although as will be outlined the sedimentary properties of channel boundaries does vary with anabranch size as the compaction of floodplain sediments varies with depth as described in Chapter 3. However, while simple in some respects, the influence of age on the planform properties of anastomosing anabranches provides an additional complicating factor which is much less easily quantified than the factors mentioned above which are constant between anabranches.

Firstly, processes and determinants of channel bank erosion are reviewed. The magnitude and distribution of bank erosion rates are crucial determinants of single channel planform which cannot be properly understood without first understanding the means by which channel adjustments are made. Following this observed regularities in single channel planform properties and their possible causes are reviewed, and then the characteristics of the reaches studied are described and results from analysis of their geometry presented. As described in
Chapter 2, two different methods of planform analysis are applied to the reaches, in part as a comparison of the methods as well as for the results each provides. The relative efficacy and utility of the methods is compared and the implications of the results discussed.

4.2 Processes of channel bank erosion

Discounting the dissolution of bank materials which is rarely significant in alluvial channels, bank erosion occurs both by direct entrainment of bank material by channel flow and by mass movement processes (Thorne and Lewin, 1979; ASCE TC, 1998). Channel bank erosion rates are generally determined by the balance between the various forces exerted on the channel bank by gravity and channel flow and the resistance of the channel bank to those forces. In a study of the distribution of bank erosion rates on channels in Devon, England, Hooke (1980) found that river size was the most important factor determining rates of bank erosion with larger rivers eroding their banks at greater rates than smaller ones. The silt-clay content of the channel banks of the Devon rivers, a proxy measure of bank resistance, was of secondary importance (Hooke, 1980). Factors influencing the frictional erosion of channel banks, especially cohesive banks like those here, will be reviewed first followed by the influences on mass movement processes. Finally the channel banks along the Cooper Creek anastomosing anabranches will be described and possible factors influencing their erosion considered.

The resistance of bank materials to frictional fluid erosion is determined by the characteristics of the boundary sediments themselves and by additional factors such as vegetation which may bind bank sediments together (e.g. Smith, 1976; Andrews, 1984; Huang and Nanson, 1997, 1998). As noted in Section 1.3.1, Smith (1976) found that on the Alexandra River in Canada channel banks with dense root mats had approximately 20,000 times more resistance to erosion than bank sediments without a vegetation component. In the case of cohesive sediments like those composing anabranch banks here, one cannot predict sediment erodibility from texture characteristics alone because the electrostatic attractions between particles which bind the material together are influenced by a vary array of additional factors as well as sediment texture such as water content, mineralogy, organic content, sediment compaction and salinity conditions (e.g. Sundborg, 1956; Postma, 1967; Migniot, 1968; Partheniades and Passwell, 1970; Graf, 1971; McCave, 1984; Dade et al., 1992) and the erodibility of cohesive sediments has also been shown to be partially time-dependent (Partheniades and Passwell, 1970). The sediment shear strengths and erosional resistance of cohesive sediments increase with compaction as both water content and the volume of void spaces decrease (e.g. Sundborg,
Sediment erodibility is also affected by the influence of antecedent conditions on bank properties, particularly the water balance of the channel bank and exposure of the bank to water during the preceding period (e.g. Wolman, 1959; Hooke, 1979; Thorne and Lewin, 1978; ASCE TC, 1998; Simon et al., 2000). Negative pore pressures in bank sediments draw water from the channel into the floodplain sediments increase the frictional strength of bank materials by acting as a suction force between bank sediment particles (Simon et al., 2000). Positive pore pressures occur particularly during rapid draw-down of water due to falling channel flow (ASCE TC, 1998), and cause a flux of water from the floodplain to the channel. The loss of negative pore pressures removes the suction force between sediment particles and results in a reduction in the frictional strength of the bank material (Simon et al., 2000). This reduction in frictional strength will both reduce the resistance of the channel bank to frictional erosion and make the occurrence of mass movement failures more likely. Another erosion process caused by changes in stage is slaking which may occur if rising channel flow inundates a dry channel bank when positive pore pressures in bank sediments due to the compression of trapped air dislodge sediment (ASCE TC, 1998). The disruption of bank materials increases their susceptibility to frictional erosion and frost action has been found to markedly accelerate rates of bank erosion (Wolman, 1959; Lawler, 1986; Stott, 1997) and Lawler (1986) states that the influence of frost action on bank erosion is particularly important in cohesive materials. Not dissimilar to frost action, wetting and drying may significantly influence bank erosion by disrupting the structure of bank sediments and fracturing homogenous soil units (ASCE TC, 1998). The cracking that often results from wetting and drying can significantly increase erodibility by reducing shear strengths and is particularly damaging when desiccation cracks form (ASCE TC, 1998).

Figure 4.1 The variation of cohesive sediment erodibility with sediment compaction, from Postma (1967, p. 158). Postma notes that actual erodibility values will differ between particular cohesive sediments and that the values given in the figure should be treated conceptually rather than literally.

Figure 4.2 A heavily fractured, dry bed of a small anastomosing channel showing the effects of drying on boundary sediments. The surfaces between the prominent cracks visible are fractured and covered by a melange of soil peds that generally range up to approximately 20 mm in size.
4.2.1.1 Cooper Creek anastomosing channel boundary morphology and process

As described in Chapter 3, anastomosing channel banks generally vary in maximum slope from 20° to 70° with vertical banks occurring only rarely and overhanging channel banks absent. The highest channel banks observed occur along the large waterholes where bank heights may reach up to 7.5m. Channel boundary sediments are predominantly composed of cohesive materials with textures the same as floodplain sediments generally which are usually composed of 35-60% clay, 40-60% silt and 4-9% sand. That bank erosion rates are very low is evidenced by the lack of significant planform change recorded by sequential air photographs over a period of 40 years. The largest magnitudes of bank erosion observed in the field occurred on cattle tracks leading into and out of the channels and used by the cattle to get to water at low flow stages. Bank slopes here had often been greatly reduced compared to the slopes of the banks on either side of the tracks. Field observations enable a number of conclusions about operative erosion processes to be drawn. Firstly, vegetation seemed not to be a significant influence on bank erosion as while the eucalypt trees lining the banks of anastomosing channels possess extensive root systems, roots are not sufficiently dense to form mats which would retard erosion. Only very low densities of fine roots and isolated larger roots protrude from channel banks, and the large roots in some places enhance rather than retard erosion by increasing local turbulence and causing small scour depressions to form. Mass movement failures of the channel banks appear to be very rare, and only a handful of slump debris deposits were seen on the many miles of channel banks observed or surveyed in the field. Variations in pore pressure would seem to have little importance on bank erosion as the channel boundary sediments seal effectively on wetting preventing migration of water through the channel boundary (see Section 3.2.2.2).

Small scale disruption of the integrity of the channel boundary materials by cracking was caused by the exposure of material to wetting and drying cycles. The expandable cohesive sediments fracture extensively on drying as illustrated by Figure 4.2, and the melange of small (<20mm) soil peds which form a 10-20mm thick surface layer on dry channel boundaries were observed to be transported by subaerial processes. Peds were observed to be moved downslope by raindrop impacts and wind action and are probably also transported by slopewash from local rainfall although this was not observed. All these processes are likely to be more significant where channel banks are steeper. It is likely that the removal of soil blocks by rising flow is an important erosion process on these channels. Because of the increase in the compaction of floodplain sediments with depth (Rust and Nanson, 1989; Maroulis, 2000, see Section 3.2.2.2) it should be more difficult for channels to erode deeper material in the floodplain than it is for
them to erode the shallower, surficial floodplain sediments. Which processes of bank erosion are important here will become clearer following the analysis of channel planform characteristics and is addressed in Section 4.6.

4.3 Observed regularities in single channel planform properties

4.3.1 Sinuosity

The sinuosity of meandering channels has been found to be systematically related to hydraulic characteristics. Schumm and Khan (1971, 1972) in their experimental study of channel pattern formation found that, with discharge kept constant, as valley slope and so stream power increased the sinuosity of both channel and thalweg increased over most of the range of slopes over which meandering occurred. Just before the onset of braiding, meandering channel sinuosity decreased. Edgar (1984) found this relationship was replicated in both a repeat study of experimental channels and 45 alluvial channels from Illinois and Indiana. Richards (1982, Figure 7.1(d)) reports a general increase in the sinuosity of single channels with increasing stream power, although his data are not sufficiently detailed to detect the existence of any decrease in sinuosity immediately prior to the onset of braiding like that found by Schumm and Khan (1971, 1972) and Edgar (1984). The explanation of the increase in the sinuosity of meandering channels with valley slope and stream power is the greater energy of the flow to erode the channel boundary. Ferguson (1973b) studied the sinuosity of glacial streams flowing on ice and found that sinuosity was determined by initial stream power and that streams with larger initial stream powers evolved to become more sinuous than those with lower initial stream powers due to higher shear stresses at the outer bank of bends raising temperatures there and promoting bank melting. If it has sufficient energy, the channel will erode its boundary and form a meandering pattern, increasing its sinuosity which has the effect of both increasing the magnitude of skin resistance effects due to increasing channel boundary area and so increasing energy dissipation, and at the same time decreasing slope and reducing stream power. This continues until the channel becomes incompetent to erode its boundary and stable (e.g. Chang, 1988). The decrease in channel sinuosity immediately prior to the onset of braiding is due to the high energy of the flow discouraging turning and leading to the formation of bends with particularly large curvature radii (Vanoni et al., 1972; Begin, 1981a, 1981b).

As mentioned previously, sedimentary properties are significant in channel pattern formation, and Schumm (1960, 1961a, 1961b, 1963) reported associations between channel boundary materials and both cross sectional and planform properties. Schumm found that channels with cohesive boundaries tended to have small width-depth ratios and to be highly sinuous, while
channels with little cohesive material in the boundary tended to have high width-depth ratios and low sinuosities as expressed by the relationships

\[ P=0.94M^{0.25} \]  
\[ P=3.5F^{-0.27} \]

where \( P \) is channel sinuosity, \( F \) channel width/depth ratio and \( M \) the percentage of silt and clay in the channel boundary. Schumm's calculation of \( M \) has been criticised by Melton (1961) for being intercorrelated with channel dimensions which it contains in its derivation. Schumm's (1963) plot of the percentage of silt and clay in the bank sediments against sinuosity avoids intercorrelation, and shows that channel sinuosity is being consistently low with non-cohesive bank materials but widely variable in channels with cohesive bank materials. Baker (1978) found that the morphology of a number of channels in the Amazon Basin did not fit with the predictions of Schumm's relationships, which he interpreted as indicating the importance of both flow and sediment characteristics in determining channel form and which implies that flow properties are not necessarily expressed by channel sediment characteristics. This is also noted by Ferguson (1987). Rust (1978) notes another departure from the predictions of Schumm's formula in the existence of silty, low sinuosity braided rivers such as the lower Yellow River and the Slims River. Despite being composed almost entirely of silt, the boundaries of these channels do not behave cohesively being very dynamic and experiencing rapid bank caving which Rust speculates may be due to an absence of clay sized material promoting cohesion.

4.3.2 Regularities in meandering channel form

A number of empirical regularities have been observed between aspects of channel form in both natural and model meandering channels (e.g. Inglis, 1949; Leopold and Wolman, 1957, 1960; Dury, 1964c; Carlston, 1965; Schumm, 1968; Tinkler, 1971; Schumm and Khan, 1971, 1972; Davis and Sutherland, 1980; Williams, 1986). Relationships have been found between various channel characteristics including width, mean depth, bankfull cross sectional area, bankfull discharge, the mean annual discharge and indices of meander form like channel sinuosity, meander wavelength, bend length, radius of curvature, arc angle, and meander belt width. This review will be limited to the relationships between channel size meander wavelength (\( \lambda \)) and bend maximum radius of curvature (\( r_m \)). The relationship between channel size and sinuosity which will also be examined here was reviewed previously in this chapter.
4.3.2.1  Wavelength ($\lambda$) – width ($W$) relationship

Leopold and Wolman (1957) found a relationship of meander wavelength to channel bankfull width of

$$\lambda = 6.5W^{0.11}$$  \hspace{1cm} (4.3)

where $\lambda$ is meander wavelength and $W$ is channel width. This relationship was much stronger than that between bankfull discharge and meander wavelength, leading Leopold and Wolman (1957) to conclude that meander wavelength is dependent directly upon bankfull width and only indirectly on discharge. A number of other relationships between meander wavelength and channel width have been reported for freely meandering channels some of which are given in Table 4.1.

Table 4.1: Reported $\lambda$-$W$ relationships for freely meandering alluvial channels.

<table>
<thead>
<tr>
<th>Author</th>
<th>Relationship</th>
</tr>
</thead>
<tbody>
<tr>
<td>Leopold and Wolman (1960)</td>
<td>$\lambda = 10.9W^{0.10}$</td>
</tr>
<tr>
<td>Schumm (1968)</td>
<td>$\lambda = 11.7W^{0.1}$</td>
</tr>
<tr>
<td>Dury (1976)</td>
<td>$\lambda = 11W$</td>
</tr>
<tr>
<td>Hey (1976)</td>
<td>$\lambda = 12.6W$</td>
</tr>
<tr>
<td>Hickin (1977)</td>
<td>$\lambda = 10W$</td>
</tr>
<tr>
<td>Baker (1978)</td>
<td>$\lambda = 8.1W$</td>
</tr>
<tr>
<td>Davis and Sutherland (1980)</td>
<td>$\lambda = 10W$</td>
</tr>
<tr>
<td>Ferguson (1981)</td>
<td>$\lambda = 10-14W$</td>
</tr>
<tr>
<td>Richards (1982)</td>
<td>$\lambda = 12.3W$</td>
</tr>
<tr>
<td>Williams (1986)</td>
<td>$\lambda = 7.5W^{0.12}$</td>
</tr>
</tbody>
</table>

The consensus, then, is that meander wavelengths tend to approximate 10-13 channel widths. The relationships are typically very strong and that found by Williams (1986) has a correlation coefficient of 0.96. Relationships have been found to be similar between meanders in different types of media such as alluvial channels, the Gulf Stream, streams on glacial ice, surface tension meanders, solution channels in limestone, and model density currents (Leopold and Wolman, 1960; Davis and Sutherland, 1980) from which Leopold and Wolman (1960) concluded that meander geometry is determined by flow patterns rather than sediment transport. Meanders are not everywhere the same however, and meanders of Lunar and Venusian sinuous rills both seem to have different geometries to terrestrial channels of any kind (Baker et al., 1992). The $\lambda/W$ ratio has been used by a number of authors to identify Osage type underfit streams or to identify underfit streams in meandering valleys (e.g. Johnson,
Underfit streams are characterised by high values of the $\lambda/W$ ratio and values as high as 40-50 have been observed in ingrown meanders where the present channel follows the former (Dury, 1970).

Like sinuosity, meander wavelength has also been found to depend on valley slope (Friedkin, 1945; Carlston, 1965) and channel boundary texture (Schumm, 1967, 1968). Data presented by Carlston (1965) of the variation in the meander wavelength of Mississippi River meanders shows clear trends in meander wavelength which Carlston ascribes to changes in slope. The slope of the Mississippi River decreases sharply at Natchez, and meander wavelength consistently increases as one progresses downstream to Natchez as the channel enlarges and then declines after Natchez as slope decreases. The influence of slope is generally not strong however, and Schumm (1967) states that simple correlations of stream gradient against meander wavelength tend not to be significant. Schumm (1967, 1968) found that meander wavelength was influenced by the type of sediment transported by the channel as reflected in the composition of the channel boundary. Channels with low percentages of silt and clay in the channel boundary, interpreted as bedload channels, tended to have larger meander wavelengths than suspended load transporting channels of equivalent size and Schumm (1967, p.1550) states that “differences in meander wavelengths between rivers or changes in meander wavelength along a river cannot be attributed to changes of water discharge alone.” Schumm states that the relationship between meander wavelength and channel width alone will however be very good because channel width is strongly influenced by both sediment load and discharge.

4.3.2.2 Radius of curvature ($r_m$) - $W$

As stated in Chapter 2 meander bends are systematically not circular arcs, however a circular arc may reasonably approximate bend form or if a bend is strongly asymmetrical can measure the zone of maximum bend curvature. By substitution from $\lambda/W$ and $\lambda$-radius of curvature ratios, Leopold and Wolman (1960) found the relationship

$$r_m=2.3W$$  \hspace{1cm} (4.4)\]

where $r_m$ is radius of curvature. In their sample of 50 rivers, Leopold and Wolman (1960) found that 2/3rds of values lay between 1.5 and 4 with a median $r_m/W$ ratio of 2.7 and a mean of 3.1. Other authors have found similar relationships: Hickin (1977) states that $r_m/W$ ratios of 2 are typical, Ritter (1978) ratios of 2-3, and Williams (1986) found a relationship of
There is generally a scatter in \( r_m/W \) values and Williams' (1986) data showed values ranging between 1 and 8. Williams states that his data showed a stronger central tendency than that of Leopold and Wolman (1960) which he attributed to Leopold and Wolman's inclusion of streams with sinuosities below his cutoff value of 1.2. The use of \( r_m \) as an index of meander form is not uncontroversial, and both Hey (1976, 1984) and Ferguson (1981) found no \( r_m/W \) relationship for the channels they examined. Hey (1976, 1984) suggests that the additional use of bend arc angle (the angle swept out around the bend by the radius of curvature) can rationalise the lack of a relationship between \( r_m \) and \( W \) and that the \( r_m/W \) relationship is a function of arc angle. However, as Ferguson (1973a) notes the magnitude of bend arc angle is associated with bend sinuosity, and Leopold and Wolman (1960) found a marked inconsistency in the relationship between bend amplitude and \( W \) from which they concluded that bend amplitude and so sinuosity and arc angle are determined more by the magnitude of erosion than by hydraulic principles. Despite these reservations, along with the \( \lambda - W \) ratio the \( r_m - W \) ratio is perhaps the most commonly used to evaluate meander form.

4.3.2.3 Causes of meandering channel regularities

The various theories attempting an explanation of meander development are reviewed by Rhoads and Welford (1992). Fundamentally, flow induced preferential bank erosion is necessary to induce meander inception. In flume experiments Friedkin (1945) and Schumm and Khan (1971, 1972) found that with a curved flume entry meandering was induced at lower slopes than with a straight flume entry but that the meanders formed decreased in size and died out downstream. The formation of these meanders is well explained by the theory of Yalin (1971) which proposes that meandering is produced by perturbations to the form of large, macroturbulent eddies which are propagated downstream with a constant wavelength of \( \lambda = 2\pi L \) where \( L \) is the length scale of the macroturbulent eddies. The velocity fluctuations in the macroturbulent eddies are propagated due to the autocorrelation of turbulent flow structures. Hey (1976) has shown that such eddies typically extend only over half the width of the channel and taking this into account suggests a \( \lambda \) of \( 4\pi W \) or \( 12.6W \) typical of meandering natural channels (Hey, 1976). However, as Rhoads and Welford (1992) note the correlation structure of the turbulent eddies will die out with increasing distance from the perturbation causing meander amplitude to decrease with distance and Yalin's theory does not explain the formation of meanders which persist downstream and do not die out. An alternative theory proposes the existence of a migrating, infinitesimal doubly harmonic perturbation of the channel boundary.

\[ r_m = 1.5W^{1.12} \]
which is expressed above a threshold flow energy and which grows in amplitude at a preferred length scale causing the initiation of an associated meandering flow pattern with a characteristic scale (e.g. Parker, 1976). By this mechanism, meanders may begin to develop simultaneously down a reach rather than sequentially as the deflection of flow by the bars causes the erosion of channel banks. While not terribly intuitive, this theory provides a sufficient, physically based explanation of meander initiation and characteristic scale (Rhoads and Welford, 1992).

The reason for the consistency observed in the $r_m/W$ relationships of natural channels is contested. Bagnold (1960) showed that below an $r_m/W$ of approximately 2 bend flow resistance increases dramatically as flow separation occurs and an eddy is formed on the inside of the tight bend, and a minimum resistance occurs just prior to the onset of flow separation. Chang (1988) notes that the onset of flow separation is a limit to the tightness of bend curvature because of the erosional and depositional mechanisms flow separation initiates. Erosion of the outer bank occurs because flow separation decreases the effective width of the channel, leading to flow acceleration and the deflection of the maximum velocity filament to the outer bank. Deposition occurs in the separation zone because of the slackwater there, and both these processes combine to reduce bend curvature. It has been claimed both that the widely observed $r_m/W$ ratios of 2-3 represents a state of maximal bend efficiency (Bagnold, 1960; Hickin, 1974; Hickin and Nanson, 1975; Chang, 1988). However, the flow resistance-bend curvature relationship is quite complex, and Davis and Sutherland (1980) note that the median $r_m/W$ of 2.7 observed by Leopold and Wolman (1960) is closer to a local maximum of bend resistance at $r_m/W$ of 2.5 than a minimum and state that this suggests channel form adjusts to maximise flow resistance. Nanson and Hickin (1986) found that $r_m/W$ ratios of 3 were typical. As Richards (1982) notes, since bends develop to dissipate energy it is perhaps more logical that typical $r_m/W$ ratios should approximate resistance maxima.

4.4 Characteristics of the data set

The plan geometry of 27 reaches which satisfied three criteria was analysed here. All were approximately constant in width over the length of the reach, there were no significant confluences or bifurcations occurring along the length of the reach, and they all exceeded 100 channel widths in length. These criteria were applied to ensure that there was a sufficient length of reach to enable the meaningful measurement of sinuosity and other planform characteristics, and that reaches could validly be treated as homogenous. All reaches in the study area
satisfying the criteria were included in the data set which does not represent a sample of such channels but rather the entire population. In one instance, the constant width criterion was relaxed to permit the inclusion of the Watawarra Channel which systematically declines in size along its length despite the absence of significant bifurcations. This channel offers the possibility of a unique insight into the effect of changing channel size on planform along the reach and will be described in more detail later. For purposes of analysis, the Watawarra Channel was divided into three distinct reaches each of which exceeded 100 channel widths and was treated as homogenous. While this is not the case the subdivision of the channel reduces the inhomogeneity of each reach to acceptable levels.

The length and width characteristics of the reaches analysed are shown in Figure 4.3 and expressed in channel widths reach lengths vary between 139W and 1029W with a mean of 435W and median of 346W. In terms of absolute length (metres) reach lengths of smaller and larger channels were no different, however in relative terms (channel widths) the reach lengths of smaller channels tended to be much longer than those of larger channels. The channel width of reaches varied between 12.5m and 125m and covers almost the entire range of anastomosing channel sizes found including the largest. More smaller channels than larger channel are present in the data set, with only 3 channels over 50m in width and 24 below 50m. This deficiency of large channels in the data set reflects a real absence of large channels satisfying the sample criteria in the study reach and the relative frequency of occurrence of large and small anastomosing channels. In all the channels studied trains of alternate channel bends could be clearly and replicably defined. The planform of selected reaches is shown in Figure 4.4 and illustrates the regularity of meander geometry often seen. Many of the channels are below the sinuosity of 1.3 or 1.5 commonly used as the lower bound in definitions of meandering channels as meandering, however as mentioned previously sinuosity can be a deceptive measure, and lines which appear very sinuous may only have low measured sinuosity values. As an example, Eulberti Waterhole (Reach 27) shown in Figure 4.4 has a sinuosity of only 1.12 despite showing a clear, regular meander planform. As discussed previously, the sinuosity threshold employed in many definitions of meandering is purely arbitrary, and the range of sinuosities of the channels here (1.12-2.96) is similar to those studied by Leopold and Wolman (1960) (1.10-2.76) in their paper on river meandering.
Figure 4.3: Reach widths and reach lengths expressed in channel widths (n=27).
Figure 4.4 Examples of meandering channel form from the study reach showing the planform regularity shown by many reaches. Note the neck cutoff development along the smaller channels in (A) and (D) which are approximately 15 m wide. The channel in (D) shows the cross-sectional asymmetry characteristic of the meandering reaches together with the approximately constant bankfull width of the channels consistent with Brice's (1984) sinuous canaliform class. (C) shows Eulberti Waterhole which at 123 m wide is the largest anabranch in the study reach and has a sinuosity of 1.12.
4.5 Analysis results

4.5.1 Stream power and slope-discharge predictions

Figure 4.5 shows a slope-discharge plot of the reaches whose cross-sections are described in Figure 3.9. Slope was estimated as valley slope and the sinuosity of individual channels was not taken into account. It is clear that the reaches plot well below both the meandering-braiding threshold defined by Leopold and Wolman (1957) and the straight-meandering threshold defined by Ackers and Charlton (1970) which suggests that one would expect the anabranches to be laterally stable, straight channels. All anabranches have unit stream powers below 10 typical of anastomosing channels (Knighton and Nanson, 1993). Such low stream powers are also consistent with the channels being laterally stable. As mentioned in Chapter 1, Ferguson (1981) found in a study of British rivers that laterally inactive channels as a set have a median unit stream power of 15Wm$^{-2}$.

4.5.2 Channel sinuosity

The channels studied spanned a wide range of sinuosities from 1.12 to 2.96 and the relationship of channel sinuosity to channel size found here is shown in Figure 4.6. In contrast to the findings of Beaman (unpublished data, 1995, cited in Tooth and Nanson, 2000a) who reported a strong linear relationship ($r^2=0.76$) with channel sinuosity decreasing as size increased, no linear trend was detectable in the data here. However, there are clear differences in the range of sinuosities shown by large and small channels. Small channels less than 50m wide showed a wide range of sinuosities from 1.14 to 2.96, while channels above 50m in width exhibit a much more restricted range of sinuosities from 1.07 to 1.25. Because of the relative deficiency of large channels in the data set additional large channels were included to help ensure that the small number of such channels in the data set adequately represent their general sinuosity and to enable a better evaluation of trends in large channel sinuosity suggested by the data set. Five large (>50m wide) channels were included which satisfied two of the three sampling criteria in that they are relatively constant in width and lack significant confluences and bifurcations along their length. However, they do not satisfy the length criterion being only between 50 and 100 channel widths long. These channels are clearly distinguished in Figure 4.6 and varied in sinuosity from 1.07 to 1.25 as was found to be typical of large channels and confirming the pattern suggested by the channels which satisfy all three sampling criteria. The occurrence of small channels significantly more sinuous than the larger channels is the opposite of the tendency reported by Schumm and Khan (1971, 1972), Ferguson (1973b), Richards (1982) and Edgar (1984) for channels with more energy to develop greater sinuosities. It is also contrary to
Figure 4.5: Slope-discharge and stream power plots for the 67 anabranch sections in Figure 3.10 (discharge calculated by Manning’s $n_l$). a) The slope discharge plot also shows the channel pattern thresholds of Leopold and Wolman (1957) and Ackers and Charlton (1970) which are both derived from data on natural channels and according to which the Cooper anabranches would be expected to be straight and laterally stable. b) The unit stream powers of all anabranches are very low and a comparison with the findings of Ferguson (1981) also suggests that the anabranches would be laterally stable. Although because of the use of discharge in calculating stream power this figure is essentially a plot of $Q (x)$ against $Q/W (y)$, it is nonetheless useful in illustrating the range of stream powers shown by the reaches.
Figure 4.6: Relationship between channel size and sinuosity of Cooper Creek anabranches. Dashed line indicates a sinuosity of 1.25 which is exceeded by no channels of W>50m but most with W<50m. * Points represent supplementary data as described in text.
the predictions of the Schumm et al.'s (1996) model in which anabranch sinuosity is a function of age and which implies that that large channels should at least be moderately sinuous and more sinuous than some small channels. Indeed, as Figure 4.5 indicates, it is contrary to all the available empirical evidence for these channels to be sinuous at all! The causes of the sinuosity variations found will be considered later after all the planform analysis results have been presented.

4.5.3 Morphological indicators of dynamism: cutoffs and scroll bar occurrence

Both cutoff and scroll bar formation were recognisable along channels from aerial photographs however point bars were not found consistent with the earlier description of the anabranches as, in Brice's (1984) terminology, 'sinuous canaliform' channels. It was apparent that, as well as sinuosity differences between channels of differing size, there are also differences in the frequency of cutoff and scroll bar formation with anabranch size. Scroll bars are only developed on a short reach of a single large channel, the central section of Meringhina Waterhole (Figure 3.12). Cutoffs do not occur on larger channels, however, numerous neck cutoffs were observed along smaller and more sinuous channels in the study reach (Figure 4.4) and, as would be expected, cutoff frequency increases with channel sinuosity and is most intense along the most sinuous channels. This is contrary to the tendency reported by a number of authors for larger, higher energy channels to develop greater sinuosities than smaller channels (Schumm and Khan, 1971, 1972; Ferguson, 1973b; Richards, 1982; Edgar, 1984). It is also inconsistent with the anastomosing model of Schumm et al. (1996) which suggests that channel dynamism ceases when channels attain high sinuosities and that the channels then contract to become underfit streams. Rather, channel dynamism appears to be maintained even when small channels attain high sinuosities.

4.5.4 Meander geometry – results from the direct measurement of channel planform

Results are summarised here and the tabulated results for each reach are presented in Appendix B. Statistics describing meander form for a reach were calculated from measurements of all bends and meanders in the reach. Measurements of meander wavelength were computed from a minimum of 6, maximum of 59, and mean of 26 wavelength measurements per reach while radius of curvature statistics were computed from double the number of bends per reach. These sample sizes compare more than favourably to those used in previous studies. Williams (1986) reports that Leopold and Wolman (1960) occasionally only measured one subjectively selected representative meander from a reach, and states that in his study of Swedish channels the
number of wavelengths measured per channel ranged from 1 to 19 with an average of 8, and the number of radius of curvature measurements ranged between 1 and 56 with an average of 11. Williams reported reach averaged planform results, while Brice (1984) states more generally that bends or meanders of median size tend to conform to empirical meander relationships. As Williams (1986) notes, studies of meander form using the direct measurement approach generally assume an underlying characteristic regularity in channel planform and that some measure of the central tendency of the distribution, usually the mean, is representative of this characteristic regularity. This assumption is generally not tested and more detailed analyses of distribution form not reported. While it is difficult to quantify the strength of a tendency to a characteristic regularity, more detailed descriptions of distribution form are utilised here in an attempt to better evaluate whether measures of the location or centrality of a data set such as the mean or median represent a characteristic scale as this is not determinable from the measures themselves.

4.5.4.1 Meander wavelength

Meander wavelengths of the channels studied varied between a maximum of 2993m and a minimum of 51m. Wavelength distributions for all reaches were positively skewed and approximately log-normally distributed consistent with the findings of Brice (1984) that distributions of meander properties tend to be log-normally distributed or positively skewed (Figure 4.7). Skewness values ranged from +0.1 to +4.2 a mean skewness of +1.4 indicates that skewness is generally quite marked. Mean and median meander wavelengths were computed for each reach. The measures are very strongly correlated at above the 99% level ($\rho=99.3$, $\alpha<0.001$), and as would be expected given the positive skewness of the data sets median values are generally lower and an average of 90% of mean values. A further, more novel measure of location, the mid-mean, was computed for comparison purposes and is the mean of the values between the first and third quartiles which minimises the influence on outliers in skewed distributions on the mean value. Mid-mean meander wavelengths were very similar to median wavelengths and an average of only 0.8% greater than median values. Median values provide a more accurate measure of the dominant wavelength of meandering channels because of the skewness of the wavelength distributions and so results calculated using median values are reported here. The should not create any great difficulties in comparing these results to those of previous studies which have predominantly reported mean values and statistics derived using mean values (Williams, 1986) because meander loops of median size tend to conform to empirical meander laws (Brice, 1984).
Figure 4.7: Reach wavelength distribution skewness values (n=27).

Figure 4.8: Skewness of reach radius of curvature ($r_n$) distributions (n=27).

Figure 4.9: Distribution of $\lambda_n/W$ relationships (n=27).
Coefficient of variation (CV) values for reach meander wavelength distributions ranged from 25.8 to 97.9 with a mean of 53.1 and median of 46.3 (values for individual reaches are given in Appendix B). The mean kurtosis value for the reach meander wavelength distributions is +2.7 indicating marked leptokurtosis and a tendency towards peaked distributions consistent with a dominant characteristic scale of meandering. The maximum positive kurtosis value was +6.9, and only 4 of the 27 distributions were platykurtic. Assessing the implications of distribution descriptors is somewhat subjective, however both the strongly positive kurtosis values of the distributions and the relatively low CV values suggest that the channels here do tend to exhibit a characteristic wavelength of meandering which is more organised than one would expect if bend sizes were randomly determined, or if there was systematically no characteristic meandering scale.

4.5.4.2 Radius of curvature

Radii of curvature measured varied between 4m and 1714m, although both particularly low and high values were extremely uncommon. Extremely low values reflected rare sections where channel planform is almost angular and abrupt changes in planform direction occur. Radii of curvature distributions showed similarities with meander wavelength distributions, being positively skewed with skewness values ranging from +0.5 to +3.7 and a mean of +2.1 (Figure 4.8). Median radii of curvature values \( r_m \) were an average of 75.8% of mean values, and the greater relative difference between the mean and median radii of curvature values compared to the mean and median wavelength values is due to the generally greater positive skewness of radii of curvature distributions (+2.1 c.f. +1.4).

CV values were significantly higher than those for meander wavelength distributions, ranging from 26.3 to 155.8 with a mean of 84.0. Kurtosis values of the distributions were generally positive with only 1 platykurtic distribution and a mean kurtosis value of +5.56, rather higher than that of meander wavelength distributions. While kurtosis values suggest that reaches tend to possess a dominant radius or range of radii of curvature values, the high average CV values indicate that this characteristic curvature is apparently attained by fewer bends than attain the dominant meander wavelength and that more bends diverge markedly from the characteristic radius of curvature range.

4.5.4.3 The \( \lambda-W \) relationship

The distribution of the \( \lambda/W \) ratios for individual reaches in given in Figure 4.9. Values ranged between 4.24 and 16.9 and were approximately normally distributed (skewness=+0.16, kurtosis=-0.44). The mean \( \lambda/W \) value is 10.4, the median 10.1 and 48% of values fall between
9 and 14. These results would seem to be fairly consistent with values typically shown by freely meandering channels, especially when the mean or median is used to represent the data set. Relationships between $\lambda$ and $W$ were explored. As mentioned previously, larger channels are underrepresented in the data set due to a real absence of them in the study area. Because of this the data set was analysed in two ways; including all the data, and including only data in the range where data points were dense ($W<50m$) in order to examine the influence of the 'outlier' points representing the small number of large channels in the data set on the relationships. The relationships found are presented in Table 4.2. Correlation coefficients rather than $r^2$ values are quoted partly for consistency with statistics reported by previous studies, and also because in the case of the $r_m$ relationships it is unclear that $r_m$ is directly dependent upon channel width.

Table 4.2: Empirical relationships between meander and channel properties

<table>
<thead>
<tr>
<th>Equation no.</th>
<th>Equation</th>
<th>Range</th>
<th>$p$</th>
<th>$P(p)$</th>
<th>$n$</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>$\lambda = 6.1W^{1.14}$</td>
<td>All data ($W&lt;125m$)</td>
<td>0.90</td>
<td>0.000</td>
<td>27</td>
</tr>
<tr>
<td>2</td>
<td>$\lambda = 8.73W^{1.56}$</td>
<td>$W&lt;50m$</td>
<td>0.82</td>
<td>0.000</td>
<td>24</td>
</tr>
<tr>
<td>3</td>
<td>$r_m = 0.11W^{1.84}$</td>
<td>All data ($W&lt;125m$)</td>
<td>0.88</td>
<td>0.000</td>
<td>27</td>
</tr>
<tr>
<td>4</td>
<td>$r_m = 9.9W^{0.59}$</td>
<td>$W&lt;50m$</td>
<td>0.41</td>
<td>0.053</td>
<td>24</td>
</tr>
</tbody>
</table>

Although the relationship described by equation 1 is stronger than that described by Equation 2, Equation 2 is preferable because of the greater density of data points over the range covered by the equation and the correlation coefficient indicates that there is a strong relationship between $\lambda$ and $W$. The equation has a lower intercept than most other reported equations but a slightly higher coefficient. In this respect it is similar to Williams' (1986) equation $\lambda = 7.5W^{1.12}$ and Leopold and Wolman's (1957) $\lambda = 6.5W^{1.1}$.

4.5.4.4 The $r_m-W$ relationship

The distribution of reach median $r_m/W$ values is given in Figure 4.10. Values ranged from 1.08 to 6.95, with a median of 2.45 and a higher mean vale of 2.96 due to the positive skewness of the distribution (+1.03). 52% of values were between 1.5 and 3.5. Values computed using mean $r_m$ values ranged from 1.9 to 7.5, and ranges calculated using both the mean and median $r_m$ values are both very similar to those reported by Leopold and Wolman (1960) of 0.8 to 9.7 and by Williams (1986) of 1.0 to 7.0. While the empirical relationship between the variables calculated using the full data set was relatively strong, that for the truncated data set was quite
Figure 4.10: Distribution of reach median \( r_m/W \) values \((n=27)\) and scatterplot of \( r_m \) against \( W \) showing the marked variability present.
weak and not significant at $\alpha=0.05$. An examination of the scatterplots of both the entire data set and the truncated data set suggests that while $r_m$ values do tend to increase with an increase in channel size, the degree of scatter between the points also increases markedly. In summary, while the $r_m/W$ values of individual reaches found here are typical of those generally shown by freely meandering channels, there is no strong relationship between $r_m$ and $W$ shown by the data set although, unlike the findings of Hey (1976, 1984) and Ferguson (1981), some trends are detected.

4.5.4.5 Watararra Channel

This channel occurs in the wide area of largely unchanneled floodplain after the junction of the Cooper and Wilson Rivers, is fed by a larger anastomosing anabranch immediately upstream and runs for approximately 13km as the crow flies or 33km measured along the channel in a south/south-westerly direction (Figure 3.1 and 4.11). The channel changes markedly in size downstream, becoming progressively smaller before terminating in a plethora of small distributaries at a reticulate area, and has a composite cross section with a small, relatively narrow and deep channel inset into a wider, deeper channel (Figure 4.11). This inset channel is approximately constant in size over the length of the Watararra Channel contracting only slightly from 11.3m to 7.3m in average width from Reach W1 to Reach W3, and is probably a response to smaller flow events occurring more frequently than larger, bankfull events in the channel as a whole. The analysis here uses the width of the larger channel rather than the width of the inset channel because it is the larger channel which defines overall planform. The cause of the decrease in the size of the larger channel is the loss of discharge to small distributary channels when the channel is flowing near bankfull. Study of aerial photographs and satellite imagery of the area and field observations found that the channel only received flow at bankfull stage or above in the main anastomosing channel network and that the channel does not pond flow for significant periods of time following flood events at any point along its length. Planform variations between reaches are shown in Table 4.3 and cannot be due to the reaches being different ages because they are sub-reaches of a single channel thus enabling a possible cause of planform variability, which on the anabranches here cannot be easily assessed, is eliminated. The changing characteristics of the channel were measured from 3 surveyed cross sections for each reach, and reach channel and planform characteristics are summarised in Table 4.3.
Figure 4.11: Watawarra Channel morphological map showing reach subdivisions and surveyed cross sections illustrating the downstream contraction of the channel.
Table 4.3: Watawarra Channel planform, cross section and calculated hydraulic characteristics (reach values averaged from 3 cross sections per reach).

| Reach | W (m) | \(d_{\text{max}}\) (m) | \(d_w\) (m) | A (m²) | Sinuosity \(P\) | \(\lambda\) (m) | \(r_m\) (m) | \(\lambda/W\) | \(r_m/W\) |
|-------|-------|-----------------|---------|--------|---------------|----------------|-------------|------------|-----------|---------|
| W1    | 46.7  | 1.36            | 0.59    | 27.6   | 1.78          | 432            | 76          | 9.3       | 1.64     |
| W2    | 32.8  | 1.04            | 0.5     | 16.4   | 2.30          | 277            | 51          | 8.5       | 1.55     |
| W3    | 19.0  | 0.85            | 0.38    | 7.22   | 2.96          | 215            | 37          | 11.3      | 1.97     |

With channel slope = floodplain (0.00015)

<table>
<thead>
<tr>
<th>Reach</th>
<th>Q by Manning’s (n_r) (cumecs)</th>
<th>Velocity at Manning’s (n_r) ms⁻¹</th>
<th>Q by Manning’s (n_r) (cumecs)</th>
<th>Velocity at Manning’s (n_r) ms⁻¹</th>
</tr>
</thead>
<tbody>
<tr>
<td>W1</td>
<td>10.2</td>
<td>0.37</td>
<td>2.8</td>
<td>0.21</td>
</tr>
<tr>
<td>W2</td>
<td>5.4</td>
<td>0.33</td>
<td>1.4</td>
<td>0.22</td>
</tr>
<tr>
<td>W3</td>
<td>2.0</td>
<td>0.27</td>
<td>0.4</td>
<td>0.20</td>
</tr>
</tbody>
</table>

It is clear from Table 4.3 that the planform of the Watawarra Channel is adjusted to channel width, with meander wavelength and radius of curvature both decreasing with the decline in channel width and the \(\lambda/W\) and \(r_m/W\) ratios of the reaches remaining similar. \(r_m/W\) ratios are smaller than those typical of freely meandering channels and among the smallest observed on the Cooper channels, which is probably due to both the high tortuosity of the channels and the fact that \(r_m\) as used here measures maximum bend curvature. The sinuosity of the channel increases downstream with the decline in channel width, and the narrowest reach of the channel is also the most sinuous. This is consistent with the broad trend shown in Figure 4.6 for the sinuosity of smaller channels to be generally greater than that of larger channels.

### 4.5.5 Fractal Analysis

#### 4.5.5.1 Methodology

The presence of statistical self-similarity in the plan geometry of 9 reaches from the data set is reported here. The reaches covered the full range of channel sinuosity and width variations found in the study area (Table 4.4). Analysis of the remaining 18 reaches was undertaken and confirmed that results from the 9 reaches were representative of the data set generally.

The range of calculated single fractal dimension values found was consistent with those found by other studies of river channels which have reported values up to 1.5 (Montgomery, 1996). As found by Montgomery (1996), the calculated single fractal dimension of reaches showed a strong positive relationship \((p=+0.97, \alpha<0.001)\) to reach sinuosity with reaches of higher
Table 4.4: Characteristics of reaches and results of fractal analysis

<table>
<thead>
<tr>
<th>Reach</th>
<th>Channel width (m)</th>
<th>Reach length (W)</th>
<th>Sinuosity P</th>
<th>Calculated fractal dimension D</th>
<th>$r^2$</th>
<th>Sig. F Durbin Watson test</th>
</tr>
</thead>
<tbody>
<tr>
<td>8</td>
<td>26</td>
<td>363</td>
<td>1.14</td>
<td>1.013</td>
<td>0.99987</td>
<td>0.0000</td>
</tr>
<tr>
<td>22</td>
<td>20</td>
<td>628</td>
<td>1.41</td>
<td>1.043</td>
<td>0.99990</td>
<td>0.0000</td>
</tr>
<tr>
<td>27</td>
<td>123</td>
<td>141</td>
<td>1.12</td>
<td>1.018</td>
<td>0.99989</td>
<td>0.0000</td>
</tr>
<tr>
<td>35c</td>
<td>14</td>
<td>811</td>
<td>1.67</td>
<td>1.120</td>
<td>0.99886</td>
<td>0.0000</td>
</tr>
<tr>
<td>38</td>
<td>30</td>
<td>293</td>
<td>1.13</td>
<td>1.013</td>
<td>0.99987</td>
<td>0.0000</td>
</tr>
<tr>
<td>46</td>
<td>16</td>
<td>443</td>
<td>1.34</td>
<td>1.048</td>
<td>0.99978</td>
<td>0.0000</td>
</tr>
<tr>
<td>W1</td>
<td>47</td>
<td>268</td>
<td>1.78</td>
<td>1.120</td>
<td>0.99886</td>
<td>0.0000</td>
</tr>
<tr>
<td>W2</td>
<td>33</td>
<td>289</td>
<td>2.30</td>
<td>1.210</td>
<td>0.99621</td>
<td>0.0000</td>
</tr>
<tr>
<td>W3</td>
<td>19</td>
<td>583</td>
<td>2.96</td>
<td>1.239</td>
<td>0.99555</td>
<td>0.0000</td>
</tr>
</tbody>
</table>

Note: * Indicates significant at the 95% level.

Sinuosity having higher fractal dimensions. It is clear that there is significant nonlinearity present in all the logG-logN plots (Figure 4.12) despite the high $r^2$ values recorded for all the regressions. In some of the plots (e.g. Reach W2 and W3) this is apparent from the plots themselves while for other reaches (e.g. Reach 27 – Eulberti Waterhole) this is only apparent on examination of residual plots. The systematic structure apparent in the residual plots and confirmed by the Durbin-Watson test indicates that a single fractal dimension is not a valid description of the complexity-scale relationship shown by the reaches. An examination of the residual plots shows that they appear to be systematically curved and that there is a great deal of similarity between the pattern of residuals in the different reaches. Scatter in the residuals is generally greater in the middle of the plots and in most cases decreases outside this range both above and below the central region. This is most likely an artifact of the method. At scales of measurement approaching channel width, ruler lengths at nearby scales will all tend to fit closely to the channel trace which will diminish scatter over small ranges of measurement scale, while at scales approaching the same order as the straight-line length of the reach, the measure consistently misses most of the complexity of the line which again tends to cause little scatter. As the Benoit© program calculates and keeps fractional ruler lengths measurements of $N$ at large scales still change which explains the trends seen.

To illustrate the magnitude of the variability of the calculated fractal dimension due to the nonlinearity of the complexity-scale relationship, the fractal dimension of reach W3 was calculated over 5 different ranges of scale. The 200 scales over which the object was measured were
Figure 4.12: Selected fractal analysis results. Note the systematic structure and apparent curvature of the residual plots, and that the nonlinearity of the log-log plots is only clear on examination of the residuals. Only for reaches W2 and W3 can nonlinearity be confidently identified from the log-log plots.
divided into 5 ranges of scale each consisting of 40 successive scales and a fractal dimension calculated for each set by LSLR. The results are given in Table 4.5, and illustrate both that the fractal dimension of a non fractal object will vary depending on the range of scales over which it is measured, and that the magnitude of variation may be appreciable with fractal dimensions both above and below the value of 1.239 calculated over the entire range of scales.

Table 4.5: The variation in calculated fractal dimension with measurement scale, Reach W3.

<table>
<thead>
<tr>
<th>Range of scale (m)</th>
<th>Fractal dimension, D</th>
<th>( p )</th>
</tr>
</thead>
<tbody>
<tr>
<td>19-78</td>
<td>1.007</td>
<td>0.99791</td>
</tr>
<tr>
<td>78-207</td>
<td>1.344</td>
<td>0.98563</td>
</tr>
<tr>
<td>207-537</td>
<td>1.303</td>
<td>0.96595</td>
</tr>
<tr>
<td>537-1391</td>
<td>1.048</td>
<td>0.99381</td>
</tr>
<tr>
<td>1391-3666</td>
<td>1.015</td>
<td>0.99972</td>
</tr>
</tbody>
</table>

The reaches are not fractal over ranges of scale from the width of the channel to valley width as reported by Nikora (1991). The testing of fractal elements over small ranges of scale was considered to be inappropriate because of the curved appearance of the residual plots and the difficulties in the statistical confirmation of linearity over small ranges of scale outlined in Section 2.4.4.4. However, it was notable that the change points on the residual plots did not occur at consistent multiples of channel width and were not related to the scale of the largest meander bend which together with the curvature shown by residual plots strongly suggests that fractal elements do not exist here.

4.5.5.2 Summary of fractal analysis results

Fractality, where it is found in river channels, must be a reflection of the variability of \( \lambda \), meander amplitude, and channel direction changes (Montgomery, 1996). The reaches examined here do not seem to be fractal objects, which is consistent with the finding from the direct measurement of planform properties that dominant characteristic scales of planform variability are found. The fractal method is a very compressive measure which attempts to describe (aspects of) often highly complex forms with a single number. One of the limitations of the fractal method is that its construction makes the recognition of characteristic scales difficult, another is that as Murray and Paola (1996) note the method is incapable of recognising recurring sequences within patterns at particular scales. Since the essence of the meandering pattern is systematically recurring sequences of planform variations the fractal method is perhaps not the best choice for the analysis of river channel patterns and results here have borne this out. Given the well documented literature describing the characteristic scale of
meandering reviewed earlier in this chapter it is perhaps surprising that the description of meandering channels as fractal objects in contradiction to this literature has received little attention.

4.6 Discussion

Results from the direct measurement of channel planform indicate that the planform of the channels as a set is typical of that of freely meandering channels and that the channels develop meander-forms with a characteristic wavelength adjusted to channel width. The results from fractal analysis support this in as much as they indicate that the channels are not fractal objects, although for this reason the fractal measure was not particularly useful here as it is not a good descriptor of non-fractal objects. Adjustment of the radius of curvature is less clear, and although the distribution of \( r_m/W \) values from individual reaches found here is very similar to those reported by Leopold and Wolman (1960) and Williams (1986), the relationship between \( r_m \) and \( W \) is rather weak. One possible reason for this is the importance of channel sinuosity in determining \( r_m \). Channels with high sinuosities tend to develop tighter bends than straighter channels with very shallow bends, and given that there is no strong trend in channel sinuosity and size over much of the data set one would not expect a strong relationship between \( r_m \) and \( W \).

The sinuosity results are perhaps the most interesting. The finding that the sinuosity of small channels is generally greater than that of large channels indicates that sinuosity is not determined by flow strength alone. The fact that larger channels do not attain moderate sinuosities suggests that sinuosity is not determined by channel age alone as Schumm et al.'s (1996) model suggests, as if this was the case some larger channels should show at least moderate sinuosities due to being older than the smaller channels from which they developed. The fact that none of the anabranches appear to be significantly underfit and that channels of all sizes and sinuosities appear to be actively meandering is also inconsistent with Schumm et al.'s model which suggests that highly sinuous channels will atrophy and become underfit streams. The most sinuous reaches studied, the sub-reaches of the Watawarra Channel, all showed \( \lambda/W \) values typical of freely meandering channels, and all small, sinuous channels possessed numerous neck cutoffs indicating that channel migration and lateral dynamism is an ongoing process. The fact that there is no linear relationship between anabranch size and sinuosity is considered here to be due to differences in anabranch age with younger small channels being
much less sinuous than older small channels because of the shorter period over which erosional processes have had to operate. The range of sinuosities shown by small channels thus reflects the dynamism of the channel system. As larger channels would be expected to have developed from previously smaller channels the fact that they are not intermediate in sinuosity between smaller young channels and smaller old channels suggests that channels may enlarge faster than they meander, for if they meandered more rapidly than they enlarged then they would at least be moderately sinuous by the time they became large. The attainment of \( \lambda/W \) relationships typical of freely meandering channels by large anabranches indicates that these channels do tend to migrate even though they do not attain high sinuosities.

The fact that there is such a large disparity between the maximum sinuosities attained by large channels and those attained by small channels suggests that, as well as the more sinuous smaller channels probably being older, smaller channels are more capable of increasing their sinuosity than larger channels. The results from the analysis of the Watawarra Channel support this. As larger channels will generally have greater stream powers and flow velocities than smaller channels one would expect them to be more competent to erode their boundaries, and so the fact that they seem not to be suggests that the erosion of the wetted, coherent channel boundary sediments may be a relatively insignificant process of bank erosion here. Such a conclusion would be consistent with the predictions of slope-discharge plots and stream power considerations that all the channels here should be laterally stable. Two principal hypotheses are considered as explanations of the anabranchn planform variations observed:

1. because of the high resistance of boundary sediments to erosion, channel flow is incapable of frictionally eroding the channel boundary, therefore, channel sinuosity is developed not by a conventional lateral migration process, in which the channel shifts because sediment is deposited on the inner banks of bends and eroded from the outer banks of bends, but rather by an accretionary process termed here ‘accretion release’ in which the vertical accretion of the channel enables it to shift laterally without eroding its boundary;
2. the erosion of channel boundaries does not occur by the frictional erosion of coherent boundary sediments, but rather by the removal of loose, fractured sediment produced by the drying of channel boundaries.

These hypotheses will now be examined in detail.
4.6.1.1 Hypothesis 1: The accretion release hypothesis

This hypothesis is based on the premise that, in the absence of channel boundary erosion in an accreting system, the higher velocities developed on the outside of channel bends will mean that as the channel vertically accretes sediment will be preferentially deposited on the inner bank of the channel where flows are relatively weak, rather than the outer bank of channel bends where faster flow will ensure that deposition is suppressed. Because of this, as the channel accretes it also shifts position, with bends growing in amplitude and the channel becoming more sinuous. The shifting of the channel which would be expected by this process is shown in Figure 4.13. This process enables channels to develop a meandering planform, high sinuosity and neck cutoffs in the absence of any bank erosion whatsoever. In order to assess the likelihood of this mechanism being important here, a geometric model of channel shifting was developed based on the analysis of Chitale (1970). Meander bends were modeled as successive, opposing circular arcs of wavelength 10W which results in this chapter show is typical of the channels here. The evolution of a channel of width W from a sinuosity of 1 to various higher sinuosities was considered by calculating the effect on sinuosity of lateral shifts of the channel at bend apices in increments of one channel width. An initial shift of the channel at each bend apex of 1W makes meander amplitude equal to 3W, and subsequent lateral channel shifts of 1W increase meander amplitude by 2W each time. The sinuosities developed by the channel with various lateral shifts can be calculated from Figure 9 (p.208) in Chitale (1970) which specifies the form of the relationship between the \( \lambda \) /amplitude product and sinuosity. Sinuosities developed for various lateral shifts of nW are given in Table 4.6.

Table 4.6: Sinuosity development of a meandering channel with differing lateral shifts at bend apices.

<table>
<thead>
<tr>
<th>Lateral shift at bend apex (W)</th>
<th>Channel amplitude (W)</th>
<th>Sinuosity</th>
</tr>
</thead>
<tbody>
<tr>
<td>0</td>
<td>0</td>
<td>1.00</td>
</tr>
<tr>
<td>1</td>
<td>3</td>
<td>1.20</td>
</tr>
<tr>
<td>2</td>
<td>5</td>
<td>1.61</td>
</tr>
<tr>
<td>3</td>
<td>7</td>
<td>2.07</td>
</tr>
<tr>
<td>4</td>
<td>9</td>
<td>2.47</td>
</tr>
<tr>
<td>5</td>
<td>11</td>
<td>3.23</td>
</tr>
</tbody>
</table>

In the absence of bank erosion, the maximum lateral shift of the channel is limited by the slope of the cut bank. If a channel has vertical banks then it will not shift at all by this process, while the shallower the slope of the cutbank the more the channel can shift for any given increment of accretion. The amount of accretion needed to enable a channel of given dimensions to shift...
Figure 4.13: Channel shifting trajectories over time expected as a result of (i) lateral migration, (ii) vertical accretion and (iii) the accretion release process.
by one channel width can be calculated trigonometrically. An initially straight channel of
$W/d_{\text{max}}$ ratio of 10 with 45° cutbanks was used as the basis for calculations here as these
dimensions are fairly typical of anabranches in the study reach which have a median $W/d_{\text{max}}$ of
11.8 (n=67). A channel of these dimensions must accrete by 10d in order to shift laterally by
1W, and the accretion depths needed for this channel to attain the sinuosities given in Table 4.6
are given in Table 4.7.

Table 4.7: Accretion depths needed to enable an initially straight channel of $\lambda=10W$, $W=10d_{\text{max}}$
and 45° cutbanks to attain various sinuosities by the accretion release process.

<table>
<thead>
<tr>
<th>Sinuosity</th>
<th>Lateral shifts (W)</th>
<th>Necessary accretion depth ($d_{\text{max}}$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1.2</td>
<td>1</td>
<td>10</td>
</tr>
<tr>
<td>1.61</td>
<td>2</td>
<td>20</td>
</tr>
<tr>
<td>2.07</td>
<td>3</td>
<td>30</td>
</tr>
<tr>
<td>2.47</td>
<td>4</td>
<td>40</td>
</tr>
<tr>
<td>3.23</td>
<td>5</td>
<td>50</td>
</tr>
</tbody>
</table>

With an identical magnitude of accretion, smaller channels will be able develop a greater
sinuosity by this process than will larger channels because the magnitude of accretion will be a
greater multiple of the depth of small channels than of larger channels. Thus, this could
potentially explain the greater sinuosity developed by many smaller channels than by the larger
channels here. The depth of accretion needed for a channel of $d_{\text{max}}=1m$ and $W=10m$ to attain a
sinuosity of 1.2 by this process is therefore 10m, and for a channel of $d_{\text{max}}=2m$ the necessary
accretion depth is 20m. As stated previously, the depth of the mud unit in which the
anastomosing channels are found varies between 1.5m and 10m and averages 2-3m in thickness
(Maroulis, 2000). Because the accretion depths necessary to enable channels to attain high
sinuosities by this process so exceed the average depths of the mud unit in which the channels
are formed, the process cannot therefore be responsible for the high sinuosities developed by
many channels although it may have increased channel sinuosities very slightly. In general, this
process is most likely to operate on laterally stable channels with strong, cohesive banks. The
fact that such channels tend to be narrow and deep (Schumm, 1960a, 1960b) will tend to limit
the effects of this process by requiring extremely high accretion depths before meaningful
sinuosity increases are achieved.
4.6.1.2 **Hypothesis 2: The boundary fracturing hypothesis**

The importance of the frequency of events which disrupt the integrity of bank materials in determining erosion rates, especially in channels with cohesive boundaries, has been noted by Lawler (1986) and Stott (1997). It is clear from field observations of channels that the boundaries of smaller, shallower channels are significantly more intensely fractured that the boundaries of larger channels. The reason for this is unclear. One possible reason is that the boundaries of small channels are less continuously wetted than the boundaries of larger channels. Smaller channels hold ponded water for shorter periods than do larger channels because their size means that there is less volume of water to evaporate off. So while large channels often hold water during periods when there is no net flow [some particularly large channels in the study reach seem to have held water continuously since European settlement of the area (Knighton and Nanson, 2000a) in the late 1800's], small channels quickly dry out following the cessation of flow. Another possible reason consistent with the observed pattern of boundary fracturing is that the more compact, deeper floodplain sediments may fracture less easily and intensely. By this hypothesis the erosion of more continually wetted boundary sediments is relatively insignificant because the high resistance of such sediments to erosion means that even large channels cannot laterally migrate through them. This hypothesis is consistent with the observed pattern of large channels (W>50m) being only of low sinuosity and small channels (W<50m) being able to attain higher sinuosities. The hydraulic removal of fractured material will be most efficient where flow velocities are greatest at the outside of channel bends, and because these banks tend to be the steepest the downslope movement of soil blocks by wind action, slopewash and raindrop impact will also be greatest there. Although these last three processes would be expected to be relatively unimportant, it is difficult to properly assess their possible significance in a system changing so slowly. The results from the Watawarra Channel are consistent with this hypothesis.

4.7 **Summary**

Results suggest that the channels are not competent to laterally migrate through their wetted boundary sediments consistent with the predictions of slope-discharge plots and stream power considerations, and that erosion here is made possible by the exposure of channel boundaries to wetting and drying cycles. Wetting and drying causes disruption of the channel boundary material and erosion occurs when the material loosened by this process is removed on reweeting. Of secondary importance are the hydraulic forces generated by channel flow, and, if
boundaries were equally fractured, larger channels would probably migrate more rapidly as the faster flow which generally occurs in larger channels will be more efficient in removing fractured sediments. It is the disruption of the channel boundary by cracking and the subsequent removal of the fractured material which enables even small, highly sinuous channels to continue meandering and to form frequent neck cutoffs and which enables channel of all sizes to show planform geometries consistent with those of freely meandering channels. Meander wavelengths seem to adjust synchronously with any changes in channel size that occur. As noted in Chapter 3 these channels do show sedimentological evidence of meandering (Rust, 1981; Nanson et al., 1988) which in the light of the results here must be because the predominantly fine-grained sediment load of the channels means that lateral accretion deposits are indistinguishable from floodplain sediments and that sedimentary structures formed in the fine sediments deposited are destroyed by cracking.

The pattern of anabranch dynamism that can be inferred here is very different from that proposed in Schumm et al.'s (1996) model. Small, highly sinuous channels do not cease meandering, become underfit streams and atrophy, but instead continue meandering, continue to be dynamic, and continue to possess equilibrium $\lambda/W$ ratios. This suggests that it is not anabranch contraction which drives avulsion/gradual avulsion occurrence and anastomosis formation here. The anastomosing models of Schumann (1989) and Schumm et al. (1996) both state that it is the inability of anabranches to transport the sediment supplied to the channel from upstream that causes channel infilling and contraction. Bankfull capacities are very important in these models, and it is the inability of the channel to maintain sufficient bankfull capacities that drives flow overbank and triggers the formation of new anabranches. Systems operating this way fluctuate around equilibrium capacities which they cannot maintain. The anastomosing channel system here seems to operate rather differently and does not seem to tend toward forming an equilibrium number of channels or to maintain any particular capacity. In some sections of the study reach like that near Meringhina Waterhole only a couple of anastomosing channels are formed across the floodplain, while in other places up to a dozen significant anastomosing channels are found. The anastomosing network is, on the large scale, partly discontinuous, and some large complexes of anastomosing channels such as those feeding and draining Naccowlah Waterhole are only operative during overbank flows. The anastomosing system as a whole is discontinuous and while the study of channel planform in this chapter has shown that the planform of individual anastomosing channels is adjusted with bankfull width which is obviously a meaningful, important characteristic for these channels, the anastomosing network as a whole does not seem to be adjusted to any particular channel capacity or bankfull flow. As noted in Chapter 3, Knighton and Nanson (2000a) found that
anastomosing channels formed and enlarged where survey data suggested that flow from the floodplain surface would concentrate, and the formation of anastomosing channels seems to be a product of very gradual scour by high magnitude, overbank flood events. As Riley (1975) notes such scour can cause anabranch formation even if the existing channels are adjusted to some equilibrium sediment transport requirement. The development of anastomosing anabranches will be considered further in the next chapter, and the adjustment and efficiency of the channel system will be considered in more detail in Chapter 8. It is worth noting here that one consequence of the development of anastomosing channels will be an increase in the efficiency of the floodplain for carrying overbank flow, which Winker (1978) noted was also the case with the floodplain-surface channels formed on the floodplains of the Little River Drainage Basin in Texas.
5 ANASTOMOSING CHANNEL JUNCTIONS

5.1 Introduction

The variation of junction morphology between channel patterns and between confluences and bifurcations has not received a great deal of attention. There has previously been no comparison of confluence and bifurcation angular geometry in multiple channel systems and previous studies have concentrated on investigating only the distribution of confluence angles (e.g. Rachocki, 1981; Yonechi and Maung, 1986). The Cooper Creek with its plethora of junctions of both types offers an excellent opportunity to rectify this. Work on anastomosing channel junctions has suggested that confluence angles in these systems may be greater than in other systems and may be a diagnostic signature of the anastomosing pattern (Yonechi and Maung, 1986). This is investigated here. The morphology of channel junctions may also help reveal the process of anastomosing channel dynamics operating on the Cooper channels and for this reason the angular geometry of junctions between anastomosing channels and braid-form floodplain-surface channels is also studied and the nature of the interaction between braid-form and anastomosing channels described.

As well as angular geometry, the morphology of a number of confluence junctions was studied in order to examine the range of junction morphologies found. The anastomosing channel network contains junctions between channels of many differing sizes and this offers an opportunity to examine the influence of differing channel size combinations on junction form. Bifurcation morphology was not studied in detail because field observations indicated that bifurcations are relatively uniform and do not display the range of morphological variations that confluences do.

In this chapter, previous studies investigating the angular geometries and flow processes in confluences and bifurcations are reviewed. The morphologies of confluence junctions are described and their formative processes considered. The angular geometries of both confluences and bifurcations are then described and compared, and junctions between anastomosing channels and braid-form floodplain-surface channels are examined. This is followed by a detailed discussion of the findings and their implications for anastomosing channel-dynamics. Finally, the implications of the junction angle distributions found for models of junction form is briefly considered.
5.1.1 Confluence literature review

5.1.1.1 Confluence angular geometry

At the basin scale, confluences between streams have been studied as part of drainage network analysis (e.g. Horton, 1945; Lubowe, 1964; Howard, 1971; Abraham, 1984), however, in this work junction angles are generally defined as being the angle between either the average direction of the tributary channels, or valley directions. It is local channel planform at the junction that is of interest here and so this work is not particularly relevant. A number of studies have considered the local angular geometry of confluence junctions (Mosley, 1976; Rachocki, 1981; Yonechi and Maung, 1986; Best, 1985). Mosley (1976) studied confluences on the floodplains of the western Mississippi Basin and found that confluence angles, defined as the angle between the two tributaries above the confluence, vary between $<10^\circ$ and $90^\circ$ (Figure 5.1). Mosley found that a large number of small streams joined the Mississippi at right angles and stated that these junction angles are large because the small channels do not have sufficient energy to modify the junctions. Excluding these, confluence angles are approximately normally distributed which Mosley notes often reflects the operation of a number of independently operating processes which “could presumably be distinguished if our understanding was sufficiently advanced” (Mosley, 1976, p.560). Another study of the local angular geometry of confluence junctions in drainage networks is that of Best (1985) who found angles between $15^\circ$ and $105^\circ$ are most common.

Bridge (1993) states that common confluence angles between braided stream anabranches range from $15^\circ$ to $110^\circ$. The only detailed study of the angular geometry of braided channel confluences is that of Rachocki (1981) who measured the distal angles of 140 bars in braided channels on small alluvial fans. Bar distal angle is an approximation of confluence angle, assuming that the centrelines of the channels run parallel to the inner bank of the bar which will be the case if channel width is relatively constant above the junction. Rachocki found that bar distal angles ranged between $30^\circ$ and $90^\circ$ with a mean value of approximately $50^\circ$ (Figure 5.2). The distribution is positively skewed and distal angles of most braid bars are below $50^\circ$.

Yonechi and Maung (1986) found that the distal angles of channel islands on the anastomosing Poronai and Irrawaddy Rivers in south Asia varied between $20^\circ$ and $150^\circ$ on the Poronai, and $10^\circ$ and $140^\circ$ on the Irrawaddy (Figure 5.3). Both angle distributions are positively skewed, and on both rivers the overwhelming majority of island distal angles fall between $20^\circ$ and $70^\circ$. The range of confluence angles displayed by the Irrawaddy and Poronai Rivers is greater than those
Figure 5.1: Mississippi River basin river confluence angle distribution (Mosley, 1976) (n=68)

Figure 5.2: Braid bar distal angles / anabranch confluence angles (Racocki, 1981) (n=130).

Figure 5.3: Confluence junction angles, anastomosing Poronai and Irrawaddy Rivers (Yonechi and Maung, 1986)
reported for braided channels and the local angular geometry of confluences in dendritic drainage networks.

5.1.1.2 Confluence flow processes

Investigations of confluence form and process on both natural and experimental channels have yielded much detailed knowledge of the flow and sediment characteristics of confluences (e.g. Taylor, 1944; Mosley, 1976; Best and Reid, 1984; Petts and Thoms, 1987; Roy and Bergeron, 1990; Best and Roy, 1991; Ashmore et al., 1992; Biron et al., 1993; Kenworthy and Rhoads, 1995; Rhoads and Kenworthy, 1995; De Serres et al., 1999). Flow patterns in confluences are highly complex and, as would be expected in a zone of mixing, flow is not uniform (e.g. Best, 1985, 1987; Ashmore et al., 1992; De Serres et al., 1999). A number of zones of confluence junction flow can be delimited and flow patterns in a model confluence are shown in Figure 5.4. Firstly, there are areas of channel before and after the confluence where flow is directly unaffected by the mixing of flow at the junction. Prior to the confluence this is because backwater effects do not extend to these areas and after the confluence this is because the combined flow has adapted to the new single channel (Best, 1987). When flows enter the area affected by the confluence there is initially a zone of flow deflection, in which the flow paths of the tributary streams are deflected from their prior directions as the two flows meet (Best, 1987). Deflection will be greatest in the smaller of the tributaries, and often results in the formation of a zone of flow stagnation between the channels on the upstream side of the confluence as the zones of highest velocity are deflected downstream. If the channels differ significantly in size, this stagnation zone will be concentrated within the smaller tributary. Where flows converge a shear layer forms along which highly turbulent vortices form as the flows mix (Best, 1987). The increased turbulence here causes scour and a deepening of the channel bed, and is responsible for the production of the scour holes which often characterise channel confluences (Mosley, 1976; Kjerfve et al., 1979; Ashmore and Parker, 1983; Best, 1986; Bridge, 1994). Scour holes may be up to six times the depth of the tributary anabranches and their depth and morphology depend principally upon the configuration and size of the tributary channels (Ashmore and Parker, 1983). As the flows mix more completely the turbulence of the flow decreases and the scour hole gradually tapers out. Flow separation zones may form adjacent to the channel boundary on the downstream sides of the confluence if the geometry of the junction causes the main flow filament to be deflected away from the channel banks (Best and Reid, 1984; Best, 1987). Separation zones are most likely to form where confluence angles are large, are often foci for sediment accumulation and bar formation, and
Figure 5.4: Zones of confluence flow, from Best (1987, p.28).

Channel discharge ratio, $Q_r = \frac{Q_t}{Q_m}$

$\frac{b_1}{b_2} = b_3$
cause acceleration of the main flow filament by reducing the effective channel width after the confluence (Best, 1987). Net flow acceleration tends to occur through confluences and the capacity of streams after a confluence tends to be less than the total capacity of tributary channels (Roy and Roy, 1988). Flow acceleration may be due to the lateral slopes of the channel beds directing flow towards the centre of the channel, in narrow and deep channels or at near bankfull discharges to a reduction in boundary friction as Lyell suggested, or to a decrease in grain roughness at confluences in particularly coarse bed material (Roy et al., 1988).

5.1.1.3 Junction adjustment and predictive models of junction morphology

Howard (1971) states that the direction of flow deflection at the junction determines junction angle adjustment and that if tributary flow is deflected towards the main stream at the junction then flow will tend to erode the upstream side of the junction between the tributary channels and favour deposition on the downstream side of the junction causing the junction to migrate upstream and increasing the confluence angle. If flow is diverted towards the downstream side of the junction the pattern of erosion and deposition will be reversed and the junction will tend to shift downstream and become more acute (Howard, 1971). This process perhaps causes confluence angles generally to be acute and the downstream deflection of tributary channels at junctions and junction deferment is very common (Ahnert, 1998).

The non-uniform flow patterns in channel junctions are too complex for predictive models of confluence morphology to consider, and so all current predictive models of junction geometry are based on assumptions of simple, uniform flow (e.g. Howard, 1971; Mosley, 1976; Roy, 1983). The use of various kinds of extremal model which propose that junction form adjusts to minimise some function of energy or power expenditure at a junction have been popular (e.g. Horton, 1945; Howard, 1971; Mosley, 1976; Roy, 1983). Mosley (1976) used a momentum equation to predict junction angular geometry, and found that the equation “can predict the angle of exit of the lower channel from a confluence if the overall confluence angle and tributary flow momenta are known” but that “it is not possible to determine overall confluence angle in terms of hydraulic conditions, because it is an independent variable that itself controls flow processes in a confluence.” (Mosley, 1976, p.535). Like hydraulic geometry generally (Maddock, 1970; Phillips, 1990, 1991), Mosley (1976) found that junction form is indeterminate, as did Gippel (1985) for changes in channel hydraulic geometry at junctions. Roy (1983) proposed a model in which the adjustment of channel path lengths takes place to maximise the efficiency of junction form in a way analogous to the three point problem of industrial location theory. The path length of the most efficient channel is maximised and that
of the least efficient channel minimised, and if channels are relatively straight and their positions fixed at points above and below the junction, the adjustment of the length of each channel section determines the junction angle. Roy (1983) states that the predictions of his model produced a reasonable agreement with average junction angles in four dendritic drainage networks. Predictive models of junction form will not be applied here, however the results of junction angle analysis will be shown to be of relevance to the applicability of the models and the accuracy of the efficiency assumptions which underlie them.

5.1.2 Bifurcation literature review

5.1.2.1 Bifurcation angular geometry

Bifurcations have received much less attention than confluences largely due to their scarcity. A rare study of the angular geometry of channel bifurcations is that conducted by Mukerji (1976) on bifurcations of terminal fan-channels on the Sutlej-Yamuna Plain in India. Terminal fans are fluvial distributary systems in which all flow is dissipated internally in normal conditions and usually form in arid or semi-arid areas in response to high moisture deficits caused by high infiltration rates and high evaporation-precipitation ratios (Kelly and Olsen, 1993). Mukerji (1976) applied a stream-ordering system to the fan-distributaries based upon their sequence of divergence from the single channel at the fan. The channels resulting from the initial bifurcation were classed as 1st order distributaries, channels resulting from the bifurcation of 1st order distributaries as 2nd order distributaries, and so on. The results of Mukerji's analysis are shown in Table 5.1.

Table 5.1: Angles of stream divergence and their variation with stream order on Sutlej-Yamuna Plain terminal-fans, India (from Mukerji, 1976, p.195).

<table>
<thead>
<tr>
<th>Distributary order</th>
<th>Average bifurcation angle (°)</th>
<th>Distributary order</th>
<th>Average bifurcation angle (°)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>90</td>
<td>7</td>
<td>58</td>
</tr>
<tr>
<td>2</td>
<td>90</td>
<td>8</td>
<td>20</td>
</tr>
<tr>
<td>3</td>
<td>65</td>
<td>9</td>
<td>30</td>
</tr>
<tr>
<td>4</td>
<td>55</td>
<td>10</td>
<td>40</td>
</tr>
<tr>
<td>5</td>
<td>47</td>
<td>11</td>
<td>70</td>
</tr>
<tr>
<td>6</td>
<td>65</td>
<td>12</td>
<td>90</td>
</tr>
</tbody>
</table>

Bifurcation angles tend to be acute with an overall average of about 60°, although the occurrence of average bifurcation angles of 90° in three classes suggests that many are obtuse. With the exception of very high order distributaries, bifurcation angles tend to decrease with an increase in distributary order paralleling a decrease in fan slope and a decrease in fan...
convexity away from the apex (Mukerji, 1976). The decrease in fan slope and convexity with distance from the apex offers a partial explanation for the decrease in bifurcation angle with distributary order as incipient anabranches will tend to follow the slope of the fan surface and so diverge most near the fan apex (Mukerji, 1976).

5.1.2.2 Bifurcation flow processes

Most work which has considered the morphology and processes operating in channel divisions has been on braided rivers (Ashworth and Ferguson, 1986; Davoren and Mosley, 1986; Ashmore, 1991; Bridge and Gabel, 1992; Ferguson et al., 1992) or deltas (Russell, 1942; Amborg, 1948; Axelsson, 1967; Silvester and de la Cruz, 1970), and the subject has received some attention in the engineering literature because of its importance in the design of agricultural irrigation canals. Like confluences, bifurcation flow patterns have been studied using physical models (Bey, 1952; Blench, 1952; Bondurant, 1952; Lindner, 1952; Thomas, 1952; Law and Reynolds, 1966; Garde and Ranga Raju, 1985), however as Bridge (1993) notes flow in channel divergences is still very poorly known. Flow patterns in confluences and bifurcations are not symmetrical, and one notable difference is that whereas the central zone of confluence flow is characterised by intense turbulent mixing, the central zone of bifurcations is characterised by weakening, divergent flow as flow separates into the divergent channels.

5.2 Methodology

The geometry of 69 confluences and 75 bifurcations of anastomosing anabranches throughout the study reach was studied from aerial photographs (the 06/81 Windorah and 08/88 Durham Downs series). Junctions were selected at random, and both the angular geometry of the junction and the widths of the feeder and resultant channels immediately prior to adjustment at the junction were measured. The definitions of the properties measured is shown on Figure 5.5. Confluence angle is defined as the angle between the centralines of the two convergent channels above the junction. Where channels are curved leading into the junction, the direction of the channel centraline between the point at which channel width begins to adjust above the junction (marked on Figure 5.5 by the "W" sections) and the junction itself is taken to represent channel direction. As channel width only begins to adjust close to the junction, this property represents channel and so general flow direction at the junction. For the same reason, channel planform here tends to be approximately straight. The definition of bifurcation angle utilised the same method, with bifurcation angle being measured between the diverging channels below
Figure 5.5: Definitions of measured junction characteristics. Note channel width is measured before/after adjustment at the junction.
the junction. Detailed surveys and observations of a number of six channel junctions were made in order to enable the description of the morphology and diversity of junction form.

5.3 The morphology of channel junctions

5.3.1 Channel confluences

5.3.1.1 Accordant junctions

Confluences between large, similarly sized channels are most common at waterhole heads and the symmetric confluence shown in Figure 5.6 occurs at the head of Goombabina Waterhole. Like large anastomosing channels generally, the channels here are tree lined with large, mature eucalypts growing densely along the channel banks. Measurements of the density of trees within 4m of the channel bank showed that trees were spaced an average of 3.4m apart. Tree roots were not exposed along the banks which would seem to indicate some degree of stability in junction form and, as is the case for all junctions described here, no change in junction planform could be discerned from the 40 year long aerial photograph record. The channels here transport some sand although the bed material at the confluence is predominantly mud with a mean grain size of 39.1μm and a median grain size of 13μm. Silt and clay sized particles (<76μm) compose 83% of bed material. While junctions are often points of marked change in both the size and character of bed material (e.g. Lodina and Chalov, 1971; Knighton, 1980) bed material did not change here or at any other junctions surveyed. Like most junctions observed, the confluence of Goombabina Waterhole is simple in form with two canal-like channels combining to produce a third canal-like channel and as was the case for all confluences no scour hole is formed. The junction is accordant, with tributary channel bed levels at the confluence point being equal in accordance with Playfair's law (Kennedy, 1984) due to the equality in size of the feeder channels.

5.3.1.2 Discordant junctions

Kennedy (1984) states that discordance is the rule rather than the exception for channel junctions, and this was certainly found to be the case here with all junctions being strongly discordant except as described above where channels of equal size meet. Examples of discordant junctions are shown in Figures 5.7 and 5.8. The difference between channel bed levels at the Shire Confluence shown in Figure 5.7 exceeds 1.5 metres, and although the widths of the two channels are relatively similar at 24.3m for the main channel and 19.6m for the tributary, their capacities are very different at 46m² and 18m² respectively. Despite the marked
Figure 5.6: Surveyed morphology of the accordant junction at the head of Goombabina Waterhole (i). The location of this and the three other junctions subsequently described is given in (ii).
Figure 5.7: Strongly discordant junction north of Goonbabina Waterhole. (See Figure 5.6 for location)

Figure 5.8: Bogala Waterhole Confluence junction showing bed discordance and, unusually, post-junction bar formation and active junction adjustment. See Figure 5.6 for location
discontinuity of junction form headcuts were absent here as they were from all other channels and junctions observed.

The confluence at the head of Bogala Waterhole is also discordant and the bed of the shallower tributary (Tributary 1) at the mouth of the channel is 0.62m higher than that of Tributary 2 (Figure 5.8). Tributary 1 is significantly smaller than Tributary 2 with a capacity of 23m$^2$ compared to 43m$^2$. The Bogala Waterhole confluence is a rare example of a more morphologically complex junction in the study reach. A small post-confluence depositional bar of the kind described by many workers in symmetrical confluences like that here (e.g. Mosley, 1976; Best, 1986; Bristow et al., 1993) is found in the centre of the channel and extends 0.15m above the surrounding bed (Bar A). A second bar is erosional in origin and represents an excised portion of channel bank with a large number of exposed tree roots visible in the depression between the bar top and the channel bank (Bar B). Downstream of Bar A the channel attains a simple parabolic form which is maintained downstream.

5.3.1.3 Tributary fans
Discordant junctions with very prominent tributary mouth bars were observed in a number of locations in the study area, one example of which is shown in Figure 5.9. The tributary channel here has a capacity of 2.5m$^2$ compared to that of 9.5m$^2$ for the main channel above the junction, and has built out a fan-shaped tributary mouth bar (hereafter referred to as a 'tributary fan') into the main channel. The lobe extends across 65% of the width of the main channel and causes a 38% reduction in channel capacity which contracts from 9.54m$^2$ at Section H to 5.9m$^2$ at Section G. The top surface of the tributary fan is only slightly steeper than the tributary channel bed and terminates at a steep avalanche face similar in slope to the main channel banks. A small gully is incised into the tributary fan, however the size of this gully is exaggerated by the contour spacing in Figure 5.9 and it is incised no more than 50mm into the tributary fan surface. Two similar junctions were observed in other areas of the floodplain. At Goonbabina Waterhole a similar junction is formed where a small (4m wide, 0.5m deep) channel enters the ~60m wide waterhole. The junction here is strongly discordant with a difference in bed level of approximately 3.5m and the tributary fan protrudes 3m into the 60m wide channel. Another tributary fan was observed 500m upstream from Goonbabina Waterhole where a 12m wide and 1.7m deep tributary channel joins a larger 25m wide and 2.7m deep channel. The tributary fan here extends across 1/2 of the main channel, and again the junction is discordant and the bed of the tributary channel elevated 1m above the main channel bed. All the tributary fans observed were approximately symmetrical in plan and showed no apparent influence of main channel
Figure 5.9: A tributary fan formed at a strongly discordant junction on a feeder channel to Bogala Waterhole.
flow direction in their form and all were composed predominantly of fine silt and clay sized material.

Tributary fans have previously been described by Dolan et al. (1978), Kennedy (1984, 1999) and Reid et al. (1989). Kennedy (1999, p.156) states that “it is quite possible for a steep tributary with a coarse load to build a fan out into the main channel...” and a number of tributary fans of this type have been. Dolan et al. (1978) describe tributary fans in the Grand Canyon which are composed of very coarse materials and built out into the main channel during large magnitude tributary floods. The fans persist because the material they are composed of is too coarse to be transported downstream. Reid et al. (1989) studied the confluence of Widdale Beck and the River Ure and found that during an event in which flow discharges in the River Ure reached 23 times those of the smaller Widdale Beck, a prominent coarse gravel tributary fan was built out from the mouth of the River Ure across Widdale Beck (see Figure 8.6, Reid et al., 1989). The fan persisted for two years and changed the morphology of the junction by inducing erosion of the bank opposite the fan.

The occurrence of prominent tributary fans in low energy channels with fine-grained cohesive boundary materials such as those here has not been previously reported. The formation of tributary fans is clearly due to a preferential deposition of sediment and their persistence due to the inability of main channel flow to remove them. The relative flow stages of confluent channels may strongly influence the pattern of sediment erosion and deposition at a confluence (Kochel and Baker, 1988; Reid et al., 1989; Bristow et al., 1993; Kennedy, 1999), however the possible variability of flow stages in the channels here is constrained. Both the tributary and main channels here are anabranches of a locally interconnected anastomosing system and receive flow from the same source upstream. Because of this flow stages in the tributary anabranches must vary synchronously in contrast to confluences in dendritic drainage networks where flow stages in confluent channels may vary synchronously or asynchronously and where flow from either channel may dominate. Flow in the smaller tributary will commence only when flow stage in the main channel rises to exceed the bed level of the smaller channel, and when both channels are active flow stage will vary synchronously. Since the tributary fan can only form when the tributary is active, it must form at high flow stages when both the main channel and tributary are active. Observations made over the period 1998-2000 showed that the morphology of the fans had not changed despite the occurrence of a number of flood events during this period one of which (the March 2000 event) approximated the 10 year flood. This implies that the tributary-fans here are stable features and form gradually over many events in contrast to coarse grained tributary fans which typically form quickly over a single event like
the River Ure fan (Reid et al., 1989), or between bankfull events in the main channel like at the confluence of the Lovers' Lane stream and the Mahaveli River in Sri Lanka studied by Kennedy (1999).

All tributary-fans observed share a number of common characteristics: all occur where a relatively small tributary joins a much larger channel at a highly discordant junction and junctions are very asymmetrical with the smaller tributaries intersecting the main channel at approximately right angles. Clearly flow patterns strongly favour sediment deposition here. Flow patterns at strongly discordant junctions have been studied by Best and Roy (1991) and De Serres et al. (1999). Best and Roy describe how a separation zone, shown in Figure 5.10, forms in the lee of the mouth of the shallower channel, and De Serres et al. (1999) from work on a natural confluence describe the occurrence of recirculating eddies there. Sediment deposition in the slackwater of this separation zone provides a possible mechanism for tributary fan formation. Deposition of tributary sediment in the separation zone below the bed of the tributary channel will form a lobe projecting into the main channel that will influence the flow patterns around it. Taylor and Woodyer (1978) found that on the Barwon River, a low gradient suspended load anastomosing river in northern New South Wales, lobes of sediment from channel slumps which encroached into the channel acted as foci for sediment deposition. During periods when only the main channel is flowing it is likely that the sediment lobes deposited at tributary fan confluences provide similar foci triggering deposition in separation zones on both the upstream and downstream sides of the sediment lobe. This would explain the planform symmetry of tributary fans and the lack of apparent imprint of flow direction on the forms. The tributary fan will therefore grow by the deposition of sediment from both the main and tributary channels, and at flow stages when both main and tributary channels are active and when only the main channel is active. It is probable that the aggregation of fine sediments promotes the formation of tributary fans as aggregates may be deposited at considerably higher flow velocities than their component particles (McCave, 1984). It is likely that the tributary-fan will continue to grow until the cross sectional area of the main channel is so constricted that flow velocity there is increased to such a degree that sediment deposition ceases.

5.3.1.4 Confluence capacity adjustment

Channel capacities at the confluences of Goonbabina, Didhelgina and Bogala Waterhole were measured and channel discharges calculated using the dimensionally balanced Manning’s ‘n’ equation (Yen, 1992) as described in Chapter 2. The feeder channels to Goonbabina Waterhole were treated as a composite channel to account for discharge from the 18.6m wide densely vegetated, lower area between the channels that would receive water when the waterhole was at
Figure 5.10: Separation zone formation at the confluence of channels of differing depth, from Best and Roy (1991, p.413).
bankfull. As Table 5.2 shows, the capacities of the channels after adjustment below the confluence are all significantly lower than the combined capacities of the feeder channels as is generally the case (e.g. Roy and Roy, 1988).

Table 5.2: Capacities and calculated discharges of waterhole confluences

<table>
<thead>
<tr>
<th>Waterhole</th>
<th>Combined capacity of feeder channels ( (m^3) )</th>
<th>Capacity of resultant channel ( (m^3) )</th>
<th>Combined discharge of feeder channels ( (\text{cumecs}) )</th>
<th>Discharge of resultant channel ( (\text{cumecs}) )</th>
</tr>
</thead>
<tbody>
<tr>
<td>Didhelgima</td>
<td>89.6</td>
<td>69.9</td>
<td>54.6</td>
<td>54.8</td>
</tr>
<tr>
<td>Goonbabina</td>
<td>215.2</td>
<td>182.3</td>
<td>191.4</td>
<td>192.4</td>
</tr>
<tr>
<td>Bogala</td>
<td>66.4</td>
<td>58.9</td>
<td>41.6</td>
<td>39.9</td>
</tr>
</tbody>
</table>

Calculated discharges are conserved through the junction and again, as expected, calculated mean flow velocities increase through the junctions. Calculations indicate that at the Bogala Waterhole Confluence flow continuity is maintained through the junction despite the more complex morphological variation shown there. The channel capacity at Section E (Figure 5.8) at 72.5\( m^3 \) is marginally greater than the combined capacity of the feeder channels (66.4\( m^3 \)) and noticeably larger than the channel capacity at Section F (58.9\( m^3 \)), however the calculated bankfull discharge of Section E (43.7cumecs) is very close to both that of both Section F (39.9cumecs) and the feeder channels (41.6cumecs).

5.4 Junction angular geometry

The results of the angular geometry analysis are summarised in Table 5.3 and the frequency distributions of the junction angles given in Figure 5.11. It is clear from both Table 5.3 and Figure 5.11 that there are marked differences between the confluence and bifurcation angle distributions. Tukey-Kramer HSD test results confirmed that the apparent difference between the junction angle means of the groups was statistically significant at above the 99% significance level (\( \alpha=0.01 \)). Mean bifurcation angles are just under twice as large as mean confluence angles, and bifurcations display a much greater variability of junction angles. Both distributions are positively skewed with the confluence angle distribution being noticeably more so than the bifurcation angle distribution. The Shapiro-Wilk Test for normality showed that neither is normally distributed (at \( \alpha=0.05 \)).
Figure 5.11: Comparison of Cooper Creek confluence and bifurcation angle distributions.

Figure 5.12: An example of a sequence of ‘scorpion’ bifurcations. Flow from top left, channels densely tree lined.
Table 5.3: Summary of junction angle results

<table>
<thead>
<tr>
<th></th>
<th>Confluences (n=69)</th>
<th>Bifurcations (n=75)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Mean (°)</td>
<td>69.5</td>
<td>124.5</td>
</tr>
<tr>
<td>Median (°)</td>
<td>66.0</td>
<td>112.0</td>
</tr>
<tr>
<td>Maximum (°)</td>
<td>209.0</td>
<td>234.0</td>
</tr>
<tr>
<td>Minimum (°)</td>
<td>16.5</td>
<td>30.5</td>
</tr>
<tr>
<td>Range (°)</td>
<td>192.5</td>
<td>233.5</td>
</tr>
<tr>
<td>Inter-quartile range (°)</td>
<td>41.0</td>
<td>70.0</td>
</tr>
<tr>
<td>Skewness</td>
<td>+1.38</td>
<td>+0.53</td>
</tr>
</tbody>
</table>

Both confluences and bifurcations may have large, obtuse junction angles and one exceptional confluence has a junction angle of 209°. Large angled bifurcations are more common than large angled confluences and a number of bifurcations were observed where junction angles are greater than 180° and the channels below the bifurcation double back on themselves for a distance before again tending down-valley. These bifurcations are termed 'scorpion' bifurcations because of their form and examples of them are given in Figure 5.12. 19% of bifurcations studied were scorpion bifurcations. Scorpion bifurcations appear to be very unusual. None were observed on available images of other anastomosing rivers, and they were also absent from a number of deltaic distributary systems studied from images in the book 'Geomorphology from Space' (Short and Blair, 1986) which provides a library of remotely sensed images of deltas (Coleman et al., 1986). Possible reasons for the occurrence of scorpion bifurcations here will be discussed in Section 5.5.1.2.

5.4.1.1 Channel size and junction angle

The composition of the confluence and bifurcation data sets was analysed to ensure that the observed could not be due to a difference in the data sets in terms of the channel combinations at junctions. Both the absolute \[ \sum W(\text{smaller channel}), W(\text{larger channel}) \] and relative \[ \frac{W(\text{smaller channel})}{W(\text{larger channel})} \] widths of the tributary/distributary channels at each junction were calculated and results showed no significant difference between the combined widths of the tributary channels at confluences and the distributary channels at bifurcations which both had a mean combined width of 46m. (Tukey-Kramer HSD test, \( \alpha=0.05 \)). There was a small but significant difference between the mean relative sizes of tributary/distributary channels at confluences and bifurcations (HSD test, \( \alpha=0.05 \)) with the mean relative width of the smaller to larger tributary channel at bifurcations being 0.77 compared to 0.70 for confluences and indicating that distributary channels at the bifurcations studied tend to be slightly more equal in size than are the tributary channels at the confluences studied.
The data set was further examined to determine whether particular junction angles were associated with particular combinations of absolute or relative channel width at a junction. Multiple regression analysis was conducted on both the confluence and bifurcation data and results showed that relationships between junction angle and the combined absolute and relative sizes of the dual channels at a junction were extremely poor predictors of confluence and bifurcation angle with $r^2$ values of 0.026 for confluences and 0.12 for bifurcations. Neither relationship was significant at $\alpha=0.05$. Confluences and bifurcations were divided into quartiles by junction angle and the characteristics of the confluent or divergent channels at the junctions examined to see if differences between quartiles were apparent. Scatterplots of the data are shown in Figure 5.13, and both these plots and testing using the Tukey-Kramer HSD test ($\alpha=0.05$) indicate that there is no significant difference between junction angle quartiles in terms of channel size and combinations of channel size. As well as indicating that the slight difference in the composition of the confluence and bifurcation data sets noted above should not have caused the differences observed, most significantly the results indicate that variations in junction angle are not associated with particular combinations of channel size in contrast to confluences from the western Mississippi River Basin where high angled confluences are predominantly due to small streams joining much larger channels (Mosley, 1976).

5.4.1.2 Comparison of confluence angle distributions

A comparison was made between the confluence angle distribution found here and those reported in previous studies (Figure 5.14). Such a comparison was not possible for bifurcation angles due to a lack of comparable data. The Cooper Creek confluence angle distribution is markedly different from both that of the Mississippi Basin confluences (Mosley, 1976) and the distal angles of braid bars reported by Rachocki (1981). Many of the Cooper confluences have obtuse confluence angles while in contrast no confluences from the Mississippi Basin or Rachocki's braided channels have obtuse confluence angles. The distribution of confluence angles from the Mississippi is bimodal, but shares a common modal class of 40-50° with the unimodal braid bar distribution although the distributions do not appear particularly similar. The comparison between the Cooper confluence angle distribution and those of the anastomosing Poronai and Irrawaddy Rivers reveals more similarities. Like the Cooper, both the Poronai and Irrawaddy Rivers possess confluences with obtuse angles. In the case of the Irrawaddy River only one confluence angle is obtuse, however, as on the Cooper, a significant percentage of confluences on the Poronai River (over 13%) are obtuse and the two distributions appear very similar.
Figure 5.13: Variation of junction channel combinations with quartile of junction angle: (i) bifurcations, and (ii) confluences. (Quartile 4=upper quartile)
Figure 5.14(i): Comparison of confluence angle distributions, Cooper Creek and (i) Mississippi basin channels (Mosley, 1976), (ii) a braided channel (Rachocki, 1981), (iii) Poronai and (iv) Irrawaddy Rivers (Yonechti and Maung, 1986)
5.5 Braid-form channel - anastomosing channel junctions

5.5.1.1 *The nature of braid-form channel - anastomosing channel junctions*

Braid-form channels generally intersect anastomosing channels at strongly discordant junctions after passing through breaches in the flanking levees. Braid-form channels bifurcating from anastomosing channels are much more subtle features than crevasse channels, do not form recognisable sediment or crevasse splays, and usually persist for long distances until a junction with another braid-form or anastomosing channel. The two proposed spatial relationships between braid-form (BF) channels and anastomosing (AC) anabranches are shown in Figure 3.19, and instances of both proposed relationships can be found as Figure 5.15 shows. In order to determine which situation is generally the case 77 braid-form anastomosing channel (BF-AC) junctions were studied from aerial photographs of the areas shown in Figure 7.21. In 16 cases (21%) a braid-form channel appeared to cut directly across an anabranching channel with no apparent discontinuity in its planform, while in 61 cases (79%) the braid form channel either originated from or terminated at the anastomosing anabranch and did not cross it. This suggests that Mabbutt’s (1967) characterisation of the braid-form-anastomosing channel relationship is correct and that when braid-form channels do appear to cut continuously across anastomosing channels it is more by accident than design. This is consistent with the suggestion of Mabbutt (1967), Nanson et al. (1986) and Rust and Nanson (1986) that braid-form channels are formed by contemporary floodwaters rather than being partially buried relics of a formerly braided channel system as Rundle (1977), Rust (1981) and Rust and Legun (1983) have claimed.

5.5.1.2 *The angular geometry of braid-form - anastomosing channel (BF-AC) junctions*

As Table 5.4 below shows, the differences in the morphology of BF-AC junctions shadow those between anastomosing anabranches (AC-AC junctions) with bifurcation angles being larger and more variable than confluence angles. Results from the Tukey-Kramer HSD test indicate that the difference between the means of the two groups is significant ($\alpha=0.01$), however it is noticeable that the BF-AC confluence and bifurcation junction angle distributions are much more similar than are the AC-AC distributions.

The angular geometry of BF-AC and AC-AC confluences are very similar while bifurcation types show more limited similarities (Figure 5.16). Differences between the means of the distributions were testing using the Tukey-Kramer HSD test which revealed that mean junction angles of confluence types are indistinguishable (69.5° c.f. 71.4°), that the mean of BF-AC bifurcation angles (91.3°) is significantly higher than that of both confluence types, and that the
Figure 5.15: Examples of the interactions between braid-form and anastomosing channels. The braid-form channel in the foreground, A, can be seen to run apparently continuously across the larger, anastomosing channel, while in contrast channels B, C and D do not cross but seem to originate at the anastomosing channel.
Figure 5.16: A comparison of BF-AC and AC-AC junction angle distributions

Figure 5.17: The postulated development of BF-AC bifurcations into AC-AC bifurcations showing the increase in junction angle with channel enlargement [(i)-(iv)]. Note the shift of the downstream side of the bifurcation (a-d) and the consequent increase in bifurcation angle.
mean of AC bifurcation angles (124.5°) is significantly larger than that of all other junction types (α=0.05).

Table 5.4: Braid-form channel-anastomosing channel junction angular geometry results

<table>
<thead>
<tr>
<th></th>
<th>Confluences (n=58)</th>
<th>Bifurcations (n=66)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Mean (°)</td>
<td>71.4</td>
<td>91.3</td>
</tr>
<tr>
<td>Median (°)</td>
<td>70.5</td>
<td>90.0</td>
</tr>
<tr>
<td>Maximum (°)</td>
<td>141.0</td>
<td>218.0</td>
</tr>
<tr>
<td>Minimum (°)</td>
<td>29.0</td>
<td>34.0</td>
</tr>
<tr>
<td>Upper quartile (°)</td>
<td>90.0</td>
<td>104.3</td>
</tr>
<tr>
<td>90% value</td>
<td>110.3</td>
<td>132.3</td>
</tr>
<tr>
<td>Range (°)</td>
<td>112.0</td>
<td>184.0</td>
</tr>
<tr>
<td>Inter-quartile range (°)</td>
<td>39.5</td>
<td>34.8</td>
</tr>
<tr>
<td>Skewness</td>
<td>+0.36</td>
<td>+1.23</td>
</tr>
</tbody>
</table>

The similarity between braid-BF-AC and AC junction morphology is sufficient to suggest that the two types are genetically related, and that, as hypothesised earlier and as one might expect, the excision of anastomosing channel islands and the formation of multiple channels occurs by braid-form floodplain-surface channels evolving into anastomosing anabranches. The lack of change observed during even large flood events suggests that this occurs as a gradual avulsion rather than an avulsion process. The braid-form channels may become larger by incising into the floodplain surface or because accretion rates of the floodplain around the channels are greater than those of the channels themselves. The difference between BF-AC and AC-AC bifurcation angles suggests that bifurcation angles evolve to become larger as the braid-form channel enlarges into an anastomosing channel. As the bifurcation develops the junction point is likely to migrate downstream because most erosional energy will be directed to the proto-anabranch bank on the downstream side of the junction (Figure 5.17). Erosion of the channel bank on the downstream side and deposition in the separation zone on the upstream side of the junction will lead junction angle to increase as along the proto-channel further away from the junction channel flow will be directed more along the channel than at the channel bank as would be expected at the junction itself. Such adjustments do not occur at confluences because flow is not as directly focussed on the channel bank. The existence of scorpion bifurcations suggests a lack of a strong influence of floodplain slope on anabranch direction which is made possible by the extremely low slope of the floodplain.

5.6 Discussion

In terms of form, the morphology of channel junctions on anastomosing channels is unusual to various degrees. The confluences do not display scour holes commonly found at confluence
junctions although this is to be expected in such a low energy environment, and the occurrence of prominent, fine-grained tributary fans has not been previously reported elsewhere. In terms of angular geometry the results reveal a number of noteworthy observations. Firstly, at junctions between anastomosing channels, confluence angles of the Cooper Creek channels are markedly higher than those reported from dendritic drainage networks and braided streams, and show a good deal of similarity with the confluence angle distributions reported from other anastomosing channels. Secondly, there is a statistically significant difference between the angular geometry of anastomosing channel (AC-AC) confluence and bifurcation junctions with bifurcation angles tending to be much larger than confluence angles. Indeed, bifurcation angles tended to be particularly large and almost 75% of those examined were over 90° and 28% over 150°. Thirdly, the difference in the angular geometry of anastomosing channel confluences and bifurcations (AC-AC) is paralleled to a degree by differences in the geometry of confluences and bifurcations between anastomosing channels and braid-form floodplain-surface channels (BF-AC). Results here have shown that the braid-form floodplain-surface channels tend to originate and terminate at anastomosing channels, and the junction angle distributions of BF-AC and AC-AC confluences are statistically indistinguishable while those of AC-AC bifurcations tend to be larger than those of BF-AC bifurcations.

5.6.1.1 Differences in confluence and bifurcation angles

The difference between confluence angles in anastomosing rivers and other channel types is most likely due to the mode of junction production. Confluences in anastomosing channels are almost exclusively produced by avulsion (or gradual avulsion) processes (e.g. King and Martini, 1984; Schumann, 1989; McCarthy et al., 1992; Harwood and Brown, 1993; Schumm et al. 1996; Makaske, 1998) while those in dendritic drainage networks are predominantly produced by the long-term evolution of the drainage network and may persist in the same location for geological timespans. While some junctions in braided channels are produced by avulsion events, most occur due to bar-building processes or bar division (Leddy et al., 1993). In a scaled physical model of a braided channel, Leddy et al. (1993) found that channel divisions caused by bar building processes accounted for approximately 85% of total channel divisions, while what they termed ‘apex avulsions’ accounted for only 15% of cases. Apex avulsions occur preferentially on the outside of channel bends where flow is directed against the outer bank because either super-elevation of the water surface on the outside of the bend due to flow momentum causes fast moving, erosive flow to overtop the bank there during flood events (e.g. Schumann, 1989), or migration of the channel leads the anabranch to intersect with a palaeochannel or topographic low which then captures channel flows (e.g. Leddy et al., 1993). Flow leaving the channel will tend to follow the slope of the floodplain which, if the
channels develop levees, will be roughly perpendicular to and directed away from the channel, and new anabranches formed by apex avulsion or gradual avulsion will thus be likely to develop high bifurcation angles (Makaske, 1998). It was noticeable from study of aerial photographs that BF-AC bifurcations were preferentially situated on the outside of channel bends, and that BF-AC bifurcation angles were generally larger than confluence angles. It has also been suggested that the confluence angles of junctions developed by apex avulsions will be significantly higher than those of confluences developed by bar-building processes (Makaske, 1998). Confluence angles of both BF-AC and AC-AC junctions were generally larger than those of braided channel confluences which are generally, although not exclusively, formed by bar-building processes (Leddy et al., 1993), suggesting that this is indeed generally the case. The position of confluence junctions may be less determined and more random than that of bifurcation junctions with floodplain-surface flows re-entering anastomosing channels wherever levees are locally absent or small. The confluence angle would then initially be at whatever angle the floodplain-surface flows happened to re-join the anastomosing channel, with the similarity between BF-AC and AC confluence angles suggesting that confluence angles do not change subsequent to their initial formation. The random location of confluence junctions would explain why confluence angles tend not to be systematically so large as the angles of bifurcations which deterministically occur in positions which lead them to tend to develop junctions with large angles.

5.6.1.2 Determinants of junctions form: efficiency versus inheritance

Most previous studies of channel junctions have concentrated on junctions with easily deformable boundaries, however Mosley (1976) notes that confluence morphology may not be in equilibrium with current hydraulic conditions and that antecedent conditions may influence or determine confluence angles. The similarity between BF-AC and AC-AC confluence junction angles suggests that antecedent conditions determine the angular geometry of AC-AC confluences, while the differences in junction angle between BF-AC and AC-AC bifurcations suggests that for these junctions antecedent conditions are not so significant. Bifurcations would seem to have sufficient energy to modify their junctions while confluences do not. Differences in angular geometry between confluences and bifurcations are accentuated by the apparent enlargement of bifurcation angles as the bifurcation channel develops. The anastomosing channels here like many anastomosing channels are low energy channels with low slopes and resistant boundaries composed of fine, cohesive materials. Because of the combination of these factors the channels cannot easily erode their banks and this would seem to be the cause of the stasis in confluence junction angles seen. However, it is also clear that
the Cooper junctions are adjusted in some ways to the prevailing hydraulic conditions. Examination of AC-AC confluences has shown that channel geometry is adjusted to maintain flow continuity at bankfull stage and the junctions show the net decrease in channel capacity through the junction typical of confluences (Roy and Roy, 1988). This is the case even through channel junctions which appear to be actively adjusting like the confluence at the head of Bogala Waterhole.

The stasis of confluence angles suggests that these junctions do not adjust their angular geometries to maximal efficiency, at least not as efficiency is defined in current extremal models of junction form. Under uniform flow assumptions greater junction angles are associated with greater energy and momentum losses, and the confluence angles here (and on other anastomosing rivers) are generally larger than those of systems like braided channels which have more easily adjustable boundaries. Although the current extremal models of junction form were not formulated to apply to bifurcations, there is nothing in the structure of the models that implies they should not apply to bifurcations if the assumptions of efficiency maximisation and minimum energy loss contained in the models are general principles of junction adjustment. However, results here suggest that the models are not applicable to bifurcation junctions, as the inferred enlargement of bifurcation angles with the development of braid-form channels into anastomosing channels indicates that these junctions seem to evolve to become more inefficient and is thus problematic for extremal models based on momentum considerations (e.g. Mosley, 1976). The common occurrence of 'scorpion' bifurcations is also problematic for extremal models of junction angular geometry based on the adjustment of channel path lengths such as that of Roy (1983) because if channels double back on themselves, as channels do at scorpion bifurcations where junction angles exceed 180°, total path length is not minimised and so neither can efficiency be. Considering only linear momentum, as simple models do, this would also seem to maximise rather than minimise momentum losses.

5.6.1.3 Summary

In conclusion, the examination of junction form here indicates that Yonechi and Maung's (1986) claim that junctions between anastomosing channels differ from those between other channel types is correct and that junction angle may indeed be a diagnostic signature of the anastomosing pattern. Results also show there are large differences in the angular geometry of confluence and bifurcation junctions. A comparison of junctions between anastomosing
anabranches (AC-AC) and those between an anastomosing anabranch and a braid-form floodplain-surface channel (BF-AC) suggests that a gradual avulsion mechanism of anastomosis formation operates here, with braid-form channels slowly morphing into larger anastomosing channels. In the case of confluences this occurs without alteration to junction angular geometry, while bifurcations become systematically more obtuse. Both these situations are problematic for the application of extremal models of junction angular geometry, as they suggest that in the case of confluences junctions do not adjust their angular geometries to increase efficiency, while bifurcations seem to become systematically more inefficient. The gradual avulsion mechanism is one of incremental, longitudinal scour and headcuts or knickpoints have not been observed anywhere on the Cooper Creek floodplain.
6 CHANNEL-FLOODPLAIN BOUNDARY FORM AND PROCESS

6.1 Introduction

The morphology and presence of levees at the channel-floodplain boundary may be an important indicator of fluvial processes and channel dynamics (Brierley et al., 1997). As stated in Chapter 1, the formation of prominent levees is a central component of models of anastomosing channel development based on crevasse avulsions. In the absence of levees an alternative mechanism must operate as Riley (1915) notes, although this does not necessarily imply that where levees are found the anastomosing pattern must be generated by the occurrence of crevasse avulsions and some other mechanism may operate. A number of field-based and theoretical studies of levee form have been conducted, however both Brierley et al. (1997) and Cazanacli and Smith (1998) still consider our understanding of levee form and process to be limited and Brierley et al. (1997, p.2) state that "levees are neglected features of the fluvial landscape." In order to study levee form and process survey and texture data for a number of levees along anastomosing anabranches were collected. Anastomosing systems are good channel types on which to study levee form and process because the multiple channels of the pattern enable levees to be studied on channels of different sizes and ergodic methods to be applied, and it is no coincidence that the most detailed study of channel levees to date was carried out on the anastomosing lower Saskatchewan River in Canada (Cazanacli and Smith (1998).

The examination of levee form in this study serves a dual purpose; to study levee morphology and process for its own sake and to provide information on the behaviour and dynamics of the channels and waterholes here. Levee texture may provide useful information on the mode of transport of various sediment size-fractions. For example, if a large, deep channel transports coarse material large quantities of which are deposited near the channel edge this indicates that the material is transported in suspension. Following a brief review of levee form, texture, and evolution studies, the analyses and results of the study of anastomosing channel levees are presented and processes of levee formation considered. As stated in Chapter 3, the occurrence of apparent levees along channels concentrating rather than dispersing flow during flood events is perhaps surprising and the morphology and texture of these deposits will also be investigated here. For this purpose the morphology and texture of the boundary of Didheljimna Waterhole is studied and the findings compared with those from the study of anastomosing channel levees.
Didelginna Waterhole shows perhaps the most marked and systematic size changes without tributary inputs of any channel in the study reach (Figure 3.11).

6.1.1.1  **Levee formation**

Levee formation is the result of sediment accumulation rates adjacent to the channel tending to be higher than those in more distal areas of the floodplain (e.g. Allen, 1965; Kesel et al., 1974; Greteener and Stromqvist, 1987; Asselman and Middelkoop, 1995; Simm, 1995). Levees generally peak at or near the channel edge and decrease in height away from the channel until they grade into the floodplain surface (e.g. Brierley et al., 1997; Cazanacli and Smith, 1998). As well as being recognisable by their form, the texture of levee sediments is also said to be intermediate in size between channel and floodplain deposits, and to become finer with increasing distance from the channel source (Kesel et al., 1974; Pizzuto, 1987; Marriott, 1992; Guccione, 1993; Asselman and Middelkoop, 1995; Brierley et al., 1997; Cazanacli and Smith, 1998; Walling and He, 1998). During the 1973 flood of the Mississippi River in Louisiana a deposit averaging 53cm thick was left on the natural levee of which 68% was sand, while in the backswamp areas the thickness of the deposit fell to 1.1cm and 97% of the material was silt and clay (Kesel et al., 1974). Asselman and Middelkoop (1995) found a similar pattern of sediment deposition during a 3 day flood on the Rhine and Meuse Rivers in the Netherlands where deposition averaged 4kg per square metre on the levees and only 1.6kg in lower, more distal areas of the floodplain with the coarsest material being deposited close to the main channel.

6.1.1.2  **Floodplain inundation patterns**

Sediment is transferred from channel to floodplain by convection and diffusion (Pizzuto, 1987; Marriott, 1992). Convection is the transfer of flow with a dominant velocity direction, while diffusion occurs when there is flow movement and mixing due to turbulence but no dominant flow direction. Convection and diffusion may occur concurrently but tend to occur at different flow stages. Convection tends to dominate when flow initially goes overbank and invades the floodplain and diffusion when the entire floodplain is inundated. When flow initially goes overbank, the transfer of water and sediment to the floodplain may occur by bank breaching or bank spilling (Lewin and Hughes, 1980). During bank breaching floodwaters invade the floodplain at discrete points, while in bank spilling floodwaters overtop large sections of bank simultaneously (Lewin and Hughes, 1980). Bank spilling may occur at higher stages than bank breaching when flow overtops large sections of channel bank as well as passing through discrete breaches, or during initial overbank flow if no breaches are present. Both bank breaching and spilling create convection currents directed away from the main channel which may transport coarse bedload sediment away from the channels (Marriott, 1992; Simm, 1995).
Flow patterns in channels with inundated floodplains have been studied by Toebes and Sooky (1967), Rajaratnam and Ahmadi (1979), James (1985), Pizzuto (1987), Knight and Shiono (1996) and Sellin and Willetts (1996). When the floodplain is fully inundated and there is no lateral water surface slope, there tends to be no net flow between channel and floodplain. Flow on the floodplain surface is much slower than flow in the main channel because of the lower depth of flow and greater roughness of the floodplain surface which causes a momentum transfer between the zones of slow floodplain and fast channel flow that is manifest as intense turbulent eddies at the channel floodplain boundary (e.g. Knight and Shiono, 1996). Although there may be no net flow between channel and floodplain, the mixing ensures that there is a constant diffusion of water between the two zones. Sediment concentrations tend to be greater over the channel than the floodplain (James, 1985; Pizzuto, 1987) and, facilitated by the intense mixing at the channel-floodplain boundary, sediment is transported to the floodplain by diffusion (James, 1985; Pizzuto, 1987). As floodplain flows are generally less competent to transport sediment than channel flows due to their lower velocities much sediment may be deposited within a short distance of the channel. This process has been modeled by James (1985) and Pizzuto (1987), however models based solely on diffusion tend to underestimate the ability of floodwaters to carry coarse material away from the channels (Pizzuto, 1987; Marriott, 1992) due to the importance of convection as well as diffusion (Pizzuto, 1987).

6.1.1.3 Levee formation models

While many authors have considered the formation and morphology of levees in general terms, there have been few detailed, process-based models of levee development proposed. A notable exception is that of Cazanacli and Smith (1998) based on the study of the developing avulsion belt of the lower Saskatchewan River in Canada. They found that the levee deposits there tended to be wedge shaped with maximum levee elevations occurring at or near the channel edge, and that levee texture became finer with increasing distance from the channel. Cazanacli and Smith recognised considerable and systematic variability in levee form and texture determined by channel size and the stage of channel development. Narrower levees tended to be laterally steeper than larger levees and to border smaller, more recently developed channels. Cazanacli and Smith (1998) state that because newly formed, small channels have low banks that are easily breached by inundating flows much coarse sediment is transported out of these channels but is deposited within a short distance of the channel to form narrow and steep levees. As channels become larger, in part due to levee accretion raising channel banks and leading bankfull levels to be elevated above the level of the surrounding floodplain, the transport of coarse sediment out of the channels becomes more difficult as such material is
generally transported low in the water column and so the sediment transported out of the channel becomes finer. This material is more easily transportable across the floodplain surface and so a greater proportion of sediment is deposited further from the channel reducing the levee slope (Cazanacli and Smith, 1998). After a channel is abandoned sediment supply and deposition on higher parts of the levee cease, however sediment is still supplied to lower distal parts of the levee and the adjacent floodplain by the flooding of more distant, active channels which causes levee slope to become even more subdued (Cazanacli and Smith, 1998). The applicability of this model to levee development on the Cooper anabranches will be assessed here.

6.2 Anastomosing channel levee morphology and texture

6.2.1 Methodology

Survey data were collected and collated for 16 levees along anastomosing channels in the study reach which cover 9 different anabranches (Figure 6.1). Most (12) of the surveys were of levees on one side of the channel only because of difficulty of access, however on two channels levees on both sides of the channel were surveyed. Care was taken to ensure that the surveys were along reaches of constant channel size and not along channels markedly expanding or contracting in width. This is because, as shown later, levees along reaches that enlarge or reduce in size without tributary or distributary junctions may form differently and show different morphological and textural variability to levees forming along constant width channels. Samples of levee surface sediments were collected for 5 anastomosing channel levees and, for comparative purposes, two transects scroll bar topography from which levees were absent. The sample size here compares well to those that have been used in previous studies. The study of Cazanacli and Smith (1998) considered 15 levee sections from 8 channel transects while most studies have considered a few levees at most (e.g. Gretener and Strømquist, 1987; Guccione, 1993; Asselman and Middelkoop, 1995). One levee was dated in order to determine rates of levee accretion and to confirm that the levees are contemporary features.

Although sometimes subdued features, levees were clearly recognisable from survey data and by eye. Levees were measured from survey data and there were no significant vegetative or pedogenic changes associated with levee boundaries. Levees were defined as extending from the bankfull edge of the channel, easily recognisable on the anabranches studied, until a point at which either floodplain slope became and remained flat, or a low point between the levee and adjacent floodplain units (Figure 6.2). Where levees were bounded by a scour channel at
Figure 6.1: Location of sampled levees and scroll bars. Levee F (not shown) occurs along the western side of Tabbareah Waterhole whose location is given in Figure 3.1.
Figure 6.2: Definition of levee properties
their distal margins, as some of the levees surveyed were, the bankfull edge of the scour channel adjacent to the levee was considered the point of levee termination and not the bed of the scour channel. *Levee height* $H$ was defined as the elevation of the levee peak above the level of the adjacent floodplain. Channel bankfull levels were not always elevated above the level of the adjacent floodplain implying that levee formation here does not always have the effect of raising channel banks (see Figure 6.5). In a few cases bankfull levels were slightly below the level of the adjacent floodplain. As well as the direct measurement of levee geometry described above a number of indices of levee form were calculated; the ratio of levee height to width ($H/W$), the distance of the high point of the levee from the channel edge ($PMH$) expressed as a proportion of levee width ($PMH/W$), and the backslope of the levee calculated as the average slope of the levee between the high point and termination point. These indices together with the direct measurements of levee form capture most of the variation in levee morphology and enable the form of different levees to be compared. The morphological characteristics of the levees surveyed are given in Table 6.1.

**6.2.2 Levee accretion rates and contemporaneity**

The 1.65m high Levee A is formed adjacent to a 24.5m wide and 2.5m deep anastomosing channel approximately 2km north of Goonbabina Waterhole and occurs on the outside of a shallow channel bend (Table 6.1; Figure 6.1). The morphology of the levee with the dates obtained are shown in Figure 6.3. When error margins are taken into account there are no reversals in age with depth down the profiles, and the dates obtained for multi-dated (TL and OSL) samples are very consistent and in two cases (SF2 and SF8) the error margins of the dates overlap while in the two other cases (SF6 and SF7) the outer limits of the error margins fall within 0.2ka of each other. *Dates from the levee peak profile (LP)* show a relatively constant accretion rate since 8.15ka BP. Accretion rates (calculated using the mean of OSL and TL dates for multi-dated samples) are 0.27m/ka from 0-2.6kaBP (0-0.7m), 0.20m/ka from 2.6-6.7kaBP (0.7-1.53m), and 0.34m/ka between 6.7 and 8.15kaBP (1.53-2.02m). Although the sample from the top of the profile (W2843) returned an abnormal TL signature the consistency of the accretion rates down the profile suggests that the date is at least approximately accurate. The dates of the profile from the distal margin of the levee (LB) show that, relative to the general accretion rate of the Cooper floodplain which averages approximately 0.25m/ka, most of the sediment accreted very rapidly with 1.37m of sediment deposited over 2.1ka and most sediment (0.96m) accreting over a period of time too small to be differentiated by the dating technique. Since 18.4ka BP, the top date on the profile, 0.72m of sediment has accumulated at an apparent accretion rate of 0.04m/ka. The extremely high accretion rates shown by Profile
Table 6.1: Morphological characteristics of anastomosing channel levees, for locations see Figure 6.1.

<table>
<thead>
<tr>
<th>Levee</th>
<th>Levee Width $W_L$ (m)</th>
<th>Height max $H$ (m)</th>
<th>Point of maximum height $PMH$ (m)</th>
<th>PMH/$W_L$</th>
<th>$H/W_L$ Ratio</th>
<th>Backslope</th>
<th>Width of adjacent channel $W$ (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>A</td>
<td>57.5</td>
<td>1.65</td>
<td>15.0</td>
<td>0.26</td>
<td>0.028</td>
<td>0.038</td>
<td>24.5</td>
</tr>
<tr>
<td>B</td>
<td>33.5</td>
<td>0.53</td>
<td>15.0</td>
<td>0.45</td>
<td>0.016</td>
<td>0.028</td>
<td>7.5</td>
</tr>
<tr>
<td>C</td>
<td>150.0</td>
<td>1.89</td>
<td>40.0</td>
<td>0.27</td>
<td>0.013</td>
<td>0.017</td>
<td>9.0</td>
</tr>
<tr>
<td>D</td>
<td>65.0</td>
<td>0.75</td>
<td>15.0</td>
<td>0.23</td>
<td>0.012</td>
<td>0.016</td>
<td>9.0</td>
</tr>
<tr>
<td>E</td>
<td>188.0</td>
<td>0.49</td>
<td>34.0</td>
<td>0.18</td>
<td>0.0026</td>
<td>0.0031</td>
<td>59.0</td>
</tr>
<tr>
<td>F</td>
<td>63.9</td>
<td>0.88</td>
<td>2.5</td>
<td>0.04</td>
<td>0.014</td>
<td>0.014</td>
<td>36.0</td>
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<tr>
<td>G</td>
<td>170.0</td>
<td>1.07</td>
<td>10.0</td>
<td>0.06</td>
<td>0.0063</td>
<td>0.0067</td>
<td>68.0</td>
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<td>H</td>
<td>147.7</td>
<td>0.73</td>
<td>74.5</td>
<td>0.51</td>
<td>0.005</td>
<td>0.010</td>
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<tr>
<td>I</td>
<td>115.0</td>
<td>0.70</td>
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<td>0.17</td>
<td>0.006</td>
<td>0.0072</td>
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<tr>
<td>J</td>
<td>75.2</td>
<td>1.25</td>
<td>14.0</td>
<td>0.19</td>
<td>0.017</td>
<td>0.020</td>
<td>12.0</td>
</tr>
<tr>
<td>K</td>
<td>85.2</td>
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<td>26.0</td>
<td>0.31</td>
<td>0.014</td>
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<td>9.0</td>
</tr>
<tr>
<td>L</td>
<td>107.7</td>
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<td>28.5</td>
<td>0.17</td>
<td>0.0022</td>
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<td>M</td>
<td>81.2</td>
<td>1.28</td>
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<td>0.13</td>
<td>0.016</td>
<td>0.018</td>
<td>55.0</td>
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<tr>
<td>N</td>
<td>79.2</td>
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<td>10.0</td>
<td>0.13</td>
<td>0.0074</td>
<td>0.0085</td>
<td>35.0</td>
</tr>
<tr>
<td>O</td>
<td>63.0</td>
<td>0.25</td>
<td>13.0</td>
<td>0.21</td>
<td>0.004</td>
<td>0.0050</td>
<td>20.5</td>
</tr>
<tr>
<td>Q</td>
<td>45.7</td>
<td>0.47</td>
<td>20.5</td>
<td>0.45</td>
<td>0.01</td>
<td>0.018</td>
<td>20.5</td>
</tr>
</tbody>
</table>
Figure 6.3: Dated levee profiles, Levee A. LP - levee peak profile, LB - levee base profile. Dates obtained with TL only are given with error margins, and where samples were multidated with both TL and OSL the average of the two dates is given without error margins. Individual dates and error margins are given in Appendix A.
LB are best explained by migration of the channel bank during channel contraction occurring around the time of the Last Glacial Maximum with the levee accreting subsequently next to the contracted channel. The sequence is unlikely to represent a point bar sequence because as mentioned previously it occurs on the outside of a channel bend. These results support the contemporaneity of levee deposits.

6.2.3 Levee morphology

6.2.3.1 Levee occurrence

Both surveys and extensive field observations indicate that levees generally occur adjacent to medium to large anastomosing channels. Measurements made in the field showed that, while there is no precise lower limit of channel width below which levees were not formed, levees only form next to narrow channels if the channels are deep and have relatively large capacities, and levees were not found adjacent to channels less than both 15m wide and 1.5m deep. The levees are dissected by braid-form floodplain-surface channels joining, crossing or leaving the anastomosing channels (e.g. Figure 5.15), and are also absent where scroll bar topography occurs which in the entire study reach is only along the 4km long central section of Meringhina Waterhole (Figure 3.12). The data set was examined to investigate whether there was a relationship between channel width and the width of the adjacent levee deposit. A number of channel size-variables were considered including width, depth and cross-sectional area. Results using width to represent channel size are reported here because they showed the strongest and most consistent trends. Channels were divided into numerically equal classes by width and differences in levee size between the groups examined (Figure 6.4). Group 1 (n=8) comprised channels less than 24m in width and Group 2 channels greater than 24m in width (n=8). Figure 6.4 shows that, while there is a large of overlap between the groups, larger channels tended to be flanked by wider levees than smaller channels. The mean, maximum and minimum widths of levees in Group 2 are all higher than those of Group 1, however differences between the groups were not statistically significant (Tukey-Kramer HSD Test, α=0.05). The relationship between levee width and the width of the adjacent channel is significant but weak with $r^2=0.28$ (p=0.04). It is clear that there is significant scatter, and these results perhaps imply that while the size of the adjacent channel may be a significant determinant of levee width it is not the sole determinant. From Figure 6.4 there seems to be a minimum levee width for any given channel width, and that levees may be much wider in individual cases which is possibly representative of age differences between the channels. Levee height showed a less convincing pattern of variation with channel size. The mean height of levees bordering large (W>24m) channels is greater than that of levees bordering small channels (1.05m c.f. 0.88m), however, as a set, levees bordering smaller channels show a larger range of heights. Regression analysis
Figure 6.4: Relationships between levee width $W_l$ and the width of the adjacent channel $W$. 

Regression line: 

Channel width $W$: 

$W = 64.5551 + 1.05476 W_l$ 

$R^2 = 0.33$ 

$p = 0.02$
revealed that there is no significant relationship between channel width and levee height ($r^2=0.01, p=0.74$).

6.2.3.2 Levee height and width

The correlation between levee height and width was not significant ($p=0.07, p=0.80$) and divided into two equal groups by levee width there was little difference in the mean heights of the groups which were 0.95m for the narrower levees ($W_L<79m \ [n=8]$) and 0.98m for the wider levees ($W_L>79m \ [n=8]$). Because of the lack of a systematic increase in levee height with width, the ratio $H/W_L$ tends to decrease with increases in levee width. Mean $H/W_L$ ratios are 0.016 for the $W_L<79m$ group and 0.009 for the $W_L>79m$ group and the difference was statistically significant (Tukey-Kramer HSD Test, $\alpha=0.05$). These results indicate that levees are not self-similar and that wide levees are not simply larger but otherwise identical versions of narrower levees.

6.2.3.3 Levee shape

Levees here are not exclusively triangular, wedge shaped deposits whose point of maximum height (PMH) occurs at or near the channel edge. Only on 2 of the 16 levees surveyed is the PMH within 0.1 levee widths of the channel edge, on 6 it is between 0.1 and 0.2, and on a further 8 is greater than 0.2 reaching a maximum of 0.51. Non-dimensional levee form can be represented, after Cazanacli and Smith (1998), by plotting $\frac{x}{W_L}$ (x-axis) where $x$ is distance from the channel and $W_L$ is levee width against $\frac{h(x)}{W_L}$ (y-axis) where $h(x)$ is height at point $x$, and this enables the scale-free comparison of levee form. Figure 6.5 plotted using this method shows selected levee profiles to illustrate the variability of levee form found. Variations in levee shape with width were examined across the levee width categories of levee width defined in Section 6.2.3.2 and results tabulated in Table 6.2.

Table 6.2: Results of levee shape analysis

<table>
<thead>
<tr>
<th></th>
<th>Small levees ($W_L&lt;64m$) [n=6]</th>
<th>Medium levees ($64&gt;W_L&lt;100m$) [n=5]</th>
<th>Large levees ($W_L&gt;100m$) [n=5]</th>
</tr>
</thead>
<tbody>
<tr>
<td>Mean PMH</td>
<td>13.5m</td>
<td>14.5m</td>
<td>35.5m</td>
</tr>
<tr>
<td>Mean PMH/W_L</td>
<td>0.29</td>
<td>0.19</td>
<td>0.23</td>
</tr>
</tbody>
</table>

Table 6.2 indicates that in absolute terms the PMH of large levees tends to be further from the channel than the PMH of smaller levees, and there is a significant positive correlation ($p=0.52, p=0.04$) between $W_L$ and the distance of the high point from the channel edge. One
Figure 6.5: Examples of normalised levee morphology to illustrate the variability in form found.
possible reason for the PMH of larger levees being further in absolute terms from the channel is that as Section 6.2.3.1 showed larger levees tend to be associated with larger channels which will develop faster flow during flood events than smaller channels. The greater velocity differential between channel and floodplain flow velocities here will lead to more intense turbulent mixing and may cause deposition to be deferred and to begin further from the channel edge. The levees here may be particularly responsive to this because as we shall see subsequently in this chapter they are predominantly composed of fine sediments. Variations in the distance of the high point from the channel edge in proportional terms are less strong. The correlation between the PMH/\(W_L\) ratio and levee width is poor and not significant at the 95% level \((p=-0.18, p=0.51)\) and there is no real trend in variation of the PMH/\(W_L\) ratio with levee width which indicates that this aspect of levee shape does not significantly differ with levee size.

6.2.3.4 Backslope

Levee backslopes are gently sloping and vary between 0.039 and 0.003 with a mean of 0.016. As one would expect, levee backslope is closely related to \(H/W_L\) ratio and levees with high \(H/W_L\) ratios tend to have steep backslopes \((p=+0.96, p=0.000)\).

6.2.4 Summary of levee morphology analysis results

Levees here seem to be somewhat unusual in that they are not always triangular, wedge shaped deposits that peak at or near the channel edge, and that levee formation is often not accompanied by a raising of the channel banks. The relationship between levee width and the size of the adjacent channel, when combined with the results from the dating of Levee A, is strong evidence that levees are contemporary features formed under and adjusted to Holocene conditions. Wider levees tend to occur adjacent to wider channels than do narrower levees however results suggest that channel width is not the only influence on levee size and that age of the anabranch may also be important. Levee morphology is influenced by the accretion of the floodplain adjacent to the levee as well as supply of sediment from the adjacent channel. Cazanaeli and Smith (1998) note that levees may become more subdued if the floodplain next to the levee is accreting faster than the levee itself. It is likely that the varying accretion of floodplain units adjacent to the Cooper Creek levees causes some of the relatively large variability in the relationship between levee-size and the size of the adjacent channel. As mentioned previously, the distal boundary of a number of the levee deposits surveyed is a floodplain-surface channel. Areas occupied by floodplain-surface channels are probably more erosive environments and therefore are likely to accrete more slowly than unchanneled areas of
floodplain adjacent to levees. Floodplain-surface channels may even scour into the floodplain surface. Variability in the accretion rates of floodplain units adjacent to levees is likely to be responsible for some of the scatter in the morphological relationships noted above.

Compared to wider levees, narrower levees tend to:

1 - peak closer to the channel in absolute terms;
2 - have higher $H/W_L$ ratios;
3 - have steeper backslopes; and, however,
4 - they do not differ significantly in maximum height from wider levees.

These trends suggest, by the ergodic principle, a morphological model of levee growth with changes in channel size. Newly formed levees next to small channels are narrow features which peak close to the adjacent channel. As the levee grows the high point shifts further away from the channel in absolute terms but does not shift significantly relative to total levee width. Most levee growth is due to the extension by deposition of the distal flank of the levee which in absolute terms enlarges much more during levee growth than does the proximal flank. The lack of a significant change in levee height during growth means that $H/W_L$ ratios and backslopes are greatly reduced, making larger levees appear much more subdued features than smaller levees. These results are broadly consistent with the levee evolution model proposed by Cazanacli and Smith (1998). They found that the causative process of changes in levee morphology was a difference in the nature and rate of sediment supply which was reflected in the textural variations of levee deposits.

### 6.3 Levee texture

Following a brief review of the processes of sediment transport across the floodplain and the possible importance of the aeolian supply of sediment to the floodplain, the results of levee texture analysis are presented. After a summary of the results the relationships between levee texture and morphology will be considered as will the applicability of the levee evolution model of Cazanacli and Smith (1998) which found levee morphology to be determined by variations in sediment supply.
The cross-floodplain transport of fluvial sediments

The analysis of the variations in levee texture is complicated by the fact that fluvial sediments are often transported as aggregates which may behave quite differently to the individual particles which they are composed of (e.g. Walling and Kane, 1984; Nicholas and Walling, 1996). Walling and Kane (1984) define the in-transport sediment size distribution as the 'effective' size distribution and that of the same sediment when fully disaggregated as the 'ultimate' sediment size distribution. If aggregate formation is significant then the behaviour of the sediment in transport will not be reflected by the ultimate size distribution, and it has been suggested that aggregation may explain exceptions in the general tendency of sediment (and levee) deposits to become finer with increasing distance from the channel (Nicholas and Walling, 1996). As stated in Chapter 3 sand sized silt and clay aggregates are formed on the floodplains here and a Shields equation calculation suggests they may be transported by flow depths exceeding only 0.2m (Nanson et al., 1986). Such depths are often exceeded in flood events and the accuracy of the predictions of the Shields equation have been confirmed by the observation of aggregate transport in shallow overbank flows (G. C. Nanson, personal communication, 1997). The situation is complicated however by the fact that the aggregates are not homogenous. As well as silt and clay particles, some may contain sand and air is a prominent component of many aggregates until they are saturated (G. C. Nanson, personal communication, 2001). The presence of sand will tend to increase aggregate densities and so decrease buoyancy, while the presence of air will have the opposite effect. The results of texture analysis enable the relative importance of aggregated and non-aggregated transport of sand to be assessed, and this point will be considered further in the summary to this section. Further calculations with the Shields equation suggest that 200μm sand can be transported at flow depths of approximately 0.2m, however in this case the predictions of the Shields equation may not be as reliable as those for the aggregates as they are based on the assumption that the material transported is a component of a bed made of similar or identical material because the equation is constructed from empirical evidence corresponding to these conditions. Use of the Shields equation is problematic when the material of interest is atypical of the bed as a whole because the character, sorting and configuration of the bed can exert a significant influence on whether or not particles or types of particles are entrained (e.g. Brayshaw, 1984, 1985; Bluck, 1987; Iseya and Ikeda, 1987; Kirchner et al., 1990; Hattingh and Illenberger, 1995; Wathen et al., 1995; Sambrook Smith, 1996; Wilcock and McArdell, 1997). Levee texture variations may provide a guide to the accuracy of the predictions of the Shields equation prediction of sand transportability.
6.3.1.2  Aeolian sediments

In arid environments river channels are often not the only significant sources of floodplain sediments and Lewin (1992) states that in “dry conditions, wind-blown material, either derived from arid-zone deflation, glacial material, or from river deposits, may also be important components of floodplains” (p.146). Sediments may be supplied by the erosion of aeolian dunes on the floodplain and Goudie et al. (1990) described the fluvial sculpting of an aeolian dune on the floodplain of an arid zone river in the Napier Ranges of Western Australia. Erosion of floodplain dunes on the Cooper Creek seems to be mostly aeolian and Maroulis (2000) reports that thermoluminescence dating of the South Narberry Sand Dune approximately 20km north of Meringhina Waterhole suggests that there has been 3m of deflation since the last glacial maximum. The sand eroded will most likely be supplied to the proximal floodplain sediments. The deposition of aeolian sediments may reach significant magnitudes (e.g. McTainsh, 1980; McTainsh and Walker, 1982; Tsoar and Pye, 1987; McTainsh et al., 1997) and McTainsh (1980) found that the erosion and transport of sediment from the Lake Chad basin in west Africa by the Harmattan wind caused an annual deposition rate of approximately 100g/m² 1000km from the source of the sediments. The transported material was mostly fine silt and clay sized, however while theoretical models of aeolian transport suggest that the wind speeds necessary for the transport of sand sized particles are so great that the particles are generally not transported long distances (e.g. Tsoar and Pye, 1987), McTainsh and Walker (1982) found that sediments in the Chad Basin up to 250μm in size were transported distances exceeding 600km. It is possible that the supply of both small and relatively large aeolian sediment particles to the floodplain-surface is significant here both because floodplain accretion rates are generally so low and because areas of easily erodible sediments abound on the bare hillsides and dune fields surrounding the river valley.

6.3.2  Levee texture results

6.3.2.1  Sample characteristics

The textural variations of 5 levees were studied (Table 6.3). As stated in Chapter 3, waterholes are often characterised by marked size variations, however while two of the levees occur along sections of channel described as waterholes - Levee D along Goonbabina Waterhole and Levee G along Meringhina Waterhole - both occur where channel width is locally constant As well as being approximately constant in size, channels next to which the levees described here were located were morphologically and sedimentologically typical of anastomosing channels as described in Chapter 3, and an examination of channel beds indicated that all transported some sand.
Table 6.3: Characteristics of texture analysis levees.

<table>
<thead>
<tr>
<th>Channel</th>
<th>Width</th>
<th>Width samples (n)</th>
<th>Sediment samples (n)</th>
</tr>
</thead>
<tbody>
<tr>
<td>A</td>
<td>58</td>
<td>0.03</td>
<td>15</td>
</tr>
<tr>
<td>C_b</td>
<td>150</td>
<td>0.01</td>
<td>40</td>
</tr>
<tr>
<td>C_w</td>
<td>65</td>
<td>0.01</td>
<td>15</td>
</tr>
<tr>
<td>D</td>
<td>189</td>
<td>0.003</td>
<td>34</td>
</tr>
<tr>
<td>G</td>
<td>148</td>
<td>0.005</td>
<td>75</td>
</tr>
</tbody>
</table>

Five characteristics were calculated for each texture sample; mean grain size (\(\mu m\)), median grain size (\(\mu m\)), percentage clay (<5.69\(\mu m\)), percentage silt (5.69-76.32\(\mu m\)), and percentage sand (>76.32\(\mu m\)) which was the largest size fraction present. The samples from the 5 levees were combined in order to examine general relationships between sediment characteristics measured (Table 6.4 and 6.5).

Table 6.4: Composition of sediment samples (n=57)

<table>
<thead>
<tr>
<th></th>
<th>% Clay</th>
<th>% Silt</th>
<th>% Sand</th>
</tr>
</thead>
<tbody>
<tr>
<td>Mean</td>
<td>40.8</td>
<td>54.1</td>
<td>5.1</td>
</tr>
<tr>
<td>Maximum</td>
<td>59.9</td>
<td>68.5</td>
<td>17.7</td>
</tr>
<tr>
<td>Minimum</td>
<td>29.7</td>
<td>37.6</td>
<td>0.0</td>
</tr>
</tbody>
</table>

Table 6.5: Correlations between sediment size variables (n=57)

<table>
<thead>
<tr>
<th>Mean grain size</th>
<th>Median grain size</th>
<th>% Clay</th>
<th>% Silt</th>
<th>% Sand</th>
</tr>
</thead>
<tbody>
<tr>
<td>Mean grain size</td>
<td>+1.00</td>
<td>+0.65</td>
<td>-0.66</td>
<td>+0.31</td>
</tr>
<tr>
<td>Median grain size</td>
<td>+0.65</td>
<td>+1.00</td>
<td>-0.96</td>
<td>+0.87</td>
</tr>
<tr>
<td>% Clay</td>
<td>-0.66</td>
<td>-0.96</td>
<td>+1.00</td>
<td>-0.91</td>
</tr>
<tr>
<td>% Silt</td>
<td>+0.31</td>
<td>+0.87</td>
<td>-0.91</td>
<td>+1.00</td>
</tr>
<tr>
<td>% Sand</td>
<td>+0.91</td>
<td>+0.41</td>
<td>-0.43</td>
<td>+0.01*</td>
</tr>
</tbody>
</table>

* denotes not significant at the 95% level

Results indicate that mean grain size is primarily determined by the sand content of the samples which is explained by the great positive skewness of all sediment size distributions. As one would expect there is a negative relationship between mean grain size and percentage clay,
both because higher percentages of clay tend to depress the mean grain size of the sediments and because samples with high clay contents tend to have lower percentages of sand. Median grain size is much more strongly associated with clay content than sand content, however unlike mean grain size the median grain size is also strongly determined by percentage silt content ($\rho=-0.87$ c.f. $+0.31$ for mean grain size). This is because silt particles may comprise a comparable percentage of samples to clay particles while sand particles are generally a much less abundant.

6.3.2.2 Mean grain size

All the levees showed significant trends in mean grain size, however, none showed the simple distal fining generally characteristic of levee deposits (Figure 6.6). Although texture trends are usually displayed together, trends in individual sediment size characteristics are plotted separately here in order to display textural variations as clearly as possible. Fitted relationships were all very strong. Levees A, C, D and G show peaks in mean grain size within 10m of the channel edge after which grain size falls markedly before rising again, and on these levees the maximum mean grain size within 10m of the channel was an average of 1.45 times the mean grain size of the levee as a whole. As Figure 6.6 illustrates, on Levees D and G the rate of the decline to the minimum mean grain size value was relatively rapid and linear elements were fitted to the declining and rising grain size limbs separately, while on Levees A and C the rate of decline was lower, the points of minimum mean grain size further from the channel edge and variations were described by 2nd-degree polynomial curves. Levee C displays a different trend to the other levees and mean grain size consistently increases across the levee away from the channel.

6.3.2.3 % Sand

As was the case for mean grain size, all levees showed significant, and in this case very similar, relationships in sand content variations which fell from a maximum near the channel to a minimum further away before rising again towards the distal margins of the levee (Figure 6.7). The maximum sand contents of the levees within 10m of the channel was an average of twice the mean percentage sand contents of samples from the levee as a whole. Maximum sand contents near the channel vary markedly while sand contents at levee distal margins are much more uniform as Table 6.6 (overleaf) shows. Sand contents near the channel of Levees A, D and G which border large channels were much larger than those of Levees C and C which border the same smaller channel (Table 6.6). For the levees bordering large channels (Levees A, D and G) the maximum sand content near the channel was also the maximum for the levee as a whole in contrast to the levees bordering the smaller channel (C and C) where maximum
Figure 6: Levee morphologies and mean grain size (μm) trends. Distance from the channel edge. Only significant relationships shown. R values apply to solid lines while dashed lines indicate trends which did not reach significance.
Figure 6.7: Levee morphology and sand content (%) trends. Distances are from the channel edge. Only significant relationships (at $\alpha=0.05$) shown. $r^2$ values apply to solid lines which indicate fitted regression lines. Points excluded from fitted lines indicated by crosses. Dashed lines visually fitted.
sand contents occurred at the distal margins of the levees (Table 6.6).

The rate of decline to a minimum sand content varied, with sand contents of Levees $C_b$ and $G$ falling rapidly to a minimum over 10m while the decline in sand content of Levees A, $C_w$ and D was much more gradual, occurring over 30-40m. This is associated with levee form, and on levees where the levee peak is greater than 20m from the channel edge declines in sand content are much more rapid than on levees that peak closer to the channel.

Table 6.6: Comparison of sand composition, proximal and distal levee boundaries.

<table>
<thead>
<tr>
<th>Levee</th>
<th>Proximal Sand Content (%)</th>
<th>Distal Sand Content (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>A</td>
<td>17.8</td>
<td>8.9</td>
</tr>
<tr>
<td>$C_b$</td>
<td>5.0</td>
<td>6.5</td>
</tr>
<tr>
<td>$C_w$</td>
<td>6.7</td>
<td>7.5</td>
</tr>
<tr>
<td>D</td>
<td>8.7</td>
<td>6.2</td>
</tr>
<tr>
<td>G</td>
<td>10.3</td>
<td>5.9</td>
</tr>
<tr>
<td>Mean</td>
<td>9.7</td>
<td>7.0</td>
</tr>
<tr>
<td>Coefficient of variation (CV)</td>
<td>51.0</td>
<td>17.4</td>
</tr>
</tbody>
</table>

6.3.2.4 Median grain size

Only Levee $C_w$ showed a significant relationship between median grain size and distance from the channel. Median grain size increased away from the channel from 8.2μm to 16.9μm (These results are given in Appendix C).

6.3.2.5 % Clay

Again only Levee $C_w$ showed a significant relationship with clay content declining approximately asymptotically with increasing distance from the channel from 42.8% to 30.5% (Appendix C).

6.3.2.6 % Silt

Levees A, $C_w$ and G showed significant trends (Appendix C). In all cases a nonlinear relationship described by a 2nd degree polynomial function was found, with percentage silt rising to peak at an intermediate distance from the channel and then falling again. However, the significance of the relationship fitted to the Levee G data must be considered tenuous as the falling limb of the curve is defined by only a single point.

6.3.3 Scroll bar texture

The texture of two transects across scroll bar topography along the western side of Meringhina
Waterholes were analysed. The location of the transects is shown on Figure 6.1 and the scroll bar topography on Figure 3.12. Both transects were 200m long and extended from the channel edge to relatively flat floodplain beyond the scroll bar topography. Neither showed any cyclic sediment texture variations related to the variations in surface elevation, and all texture trends found were simple. Both transects showed the same trends in mean grain size and sand composition variations which peaked at the channel edge and then fell sharply, asymptotically approaching minima toward the ends of the transects (Figure 6.8). In the case of transect Sb1 sand composition and mean grain size were the only significant relationships, while Sb2 showed a significant asymptotic decline in median grain size ($r^2=0.61$, $p<0.01$) and an associated asymptotic rise in percentage clay content ($r^2=0.43$, $p<0.01$). The distal-fining trends shown here are those normally found with increasing distance from the channel (e.g. Kesel et al., 1974; Pizzuto, 1987; Marriott, 1992; Guccione, 1993; Asselman and Middelkoop, 1995; Brierley et al., 1997; Cazanacli and Smith, 1998; Walling and He, 1998) and contrast sharply with the trends shown by the levee deposits studied.

6.3.4 Summary and discussion of levee texture analysis results

The most consistent results are shown by the variation of percentage sand content and mean grain size which are closely correlated, and all levees show significant relationships between these variables and distance from the channel. Only one levee shows significant trends in median grain size and clay content away from the channel, and three show significant relationships in silt content. The inconsistency in silt and clay content variations suggest that there is usually no significant selection of these grain sizes. This is consistent with Guccione’s (1993) statement that over short distances and on small channels such as these trends in coarser sediments generally be more apparent than trends in finer sediments which are clearer over larger distances. The lack of textural trends in fine sediment fractions may be due to the transport of finer sediment here as sand-sized aggregates described by Nanson et al. (1986), Rust and Nanson (1989), Maroulis (1992) and Maroulis and Nanson (1996). As the ultimate rather than effective sediment size distribution is considered here, these results cannot register any sorting of aggregated sediment which may or may not take place in the transport of sediment across the levees. As noted previously, sometimes the aggregates contain some sand, however the strong trend in sand contents and selective deposition of sand found here suggests that the amount of sand transported in aggregates is much less than that transported free as individual, unaggregated particles. The levees differ significantly in the nature of their textural variations from levees previously described in the literature which tend to uniformly fine away from the channel (e.g. Kesel et al., 1974; Pizzuto, 1987; Marriott, 1992; Guccione, 1993;
Figure 6.8: Trends in the sediment texture of scroll bar topography, Meringhina Waterhole (for section locations see Figure 6.1.)
Asselman and Middelkoop, 1995; Brierley et al., 1997; Cazanacli and Smith, 1998; Walling and He, 1998). Also, they are not uniformly intermediate in size between channel and floodplain deposits and sections are systematically finer than both.

The tendency for sand contents to peak at and near the channel edge is explained by the transfer overbank of coarse suspended sediment transported during flood events by convection and diffusion processes, and the fact that the sand compositions near the channel of levees adjacent to large channels is higher than that of levees adjacent to small channels suggests that concentrations of sand in suspended load at flood stages are greater here than in smaller channels with slower flow. The difference in the rate of initial fining of sand appears to be due to differences in levee morphology, and the further the levee high point is from the channel edge the more rapid the initial decline in sand content is. Coarse sediment transferred from the channels adjacent to Levees A and Cw which both peak only 15m from the channel must only be transported a short distance upslope before reaching the high point of the levee, after which it appears that it can be transported some distance downslope. In contrast, Levee Cε peaks 40m and Levee G 75m from the channel edge, and so sediment must be transported upslope for much longer distances across these levees before reaching the levee high point. It appears that sand cannot be transported long distances upslope and is quickly deposited explaining the rapid fining shown by these levees. On Levee D the high point at 34m from the edge of the channel coincides with the point of minimum sand content. As one would expect, it appears that sand is transported over the surface of the floodplain as bedload and that it tends to quickly be deposited if it must be transported upslope. If the high point of the levee is close to the channel then flow may have sufficient energy to transport the sediment to the high point as on Levees A and Cw.

What is most unusual in the pattern of sand and mean grain size variations is that following fining of coarse sediment from the channels indicated by the attainment of minimum sand contents, sand contents then increase again until at levee distal margins they have attained relatively consistent values with a mean sand content of 7.4%. The analysis of 28 sediment samples from various different floodplain environments which will be reported in Chapter 7 shows that floodplain sediments generally contain significant sand contents comparable to those of samples from levee distal margins and higher than the minimum sand contents of levees. The mean sand content of sediment samples from non-levee floodplain environments was 6.4% compared to the mean of the samples from distal levee margins of 7.4%, and some sand is present in all but 2 of the 28 floodplain sediment samples. As a group levees are unusual in the high frequency of samples with low sand contents. 28% of the 62 levee samples
had sand contents below 3\% compared to only 17\% of the 28 samples from non-levee floodplain environments. Levee deposits here are not always intermediate in texture between floodplain and channel deposits and sections of the levees are systematically finer than both. The sand present towards the distal margin of levees and on the floodplain is probably derived from a number of sources including splays from contracting channels (which will be described in the next part of this chapter), the aeolian and fluvial erosion of floodplain dunes, and the aeolian supply of material from the valley sides and dune fields adjacent to the floodplain. The contrast of these trends with the normal distal fining trends shown by the transects across scroll bar topography suggest that the difference may be due to the fact that the levees are elevated above the surrounding floodplain. Flow at the margins of levees will be predominantly directed down the floodplain, deeper and faster than flow over the elevated levee surfaces, and so more competent to transport coarse sediments. This would explain why sand contents on levees dip below those of the adjacent floodplain surfaces while those of scroll bar topography do not.

The majority of sediment in levees of all sizes is the finer size fractions, and it would seem that the trends in sand content and mean grain size observed are more a response to levee form than a cause. More important in determining levee form is the deposition of the finer size fractions. The vegetation associated with anastomosing channels undoubtedly influences the accretion of levee deposits by increasing roughness, decreasing flow velocities and trapping sediment. As described in Chapter 3, anastomosing channels are generally lined by large eucalypt trees with an understory of Lignum bushes. Ephemeral vegetation also concentrates along the channels because of the ameliorating effects of the larger vegetation on the local microclimate. Vegetation densities are greatest next to the channels and decrease across the levees away from the channel. As levee high points are generally some distance from the channel edge this implies that, while important, vegetation density is not the sole determinant of accretion magnitudes here otherwise the levees would be wedge shaped deposits peaking at the channel edge which Figures 6.5 and 6.6 show that they are not. That wider levees peak further from the channel than do narrower levees is probably due to greater intensities of turbulent mixing at the boundary of larger channels impeding the deposition of low density fine-sediment aggregates and unaggregated fine sediment particles here. The accretion patterns across the levee are probably determined by a combination of vegetation density and the turbulent exchange at the channel-floodplain boundary, with the location of the point of maximum accretion and levee peak occurring where the strength of the turbulent mixing wanes sufficiently for the vegetation to trap sediment and cause accretion.
While some aspects of levee form were consistent with the model of levee evolution proposed by Cazanacli and Smith (1998), results from textural analysis are more mixed. Cazanacli and Smith state that as channels become larger the transport of coarse sediment overbank becomes more difficult because channel depth increases, however here greater amounts of coarse sediment are transported overbank next to larger channels because these channels transport more coarse sediment in suspension due to higher flow velocities. However, despite more coarse sediment being transported overbank, the sediment is deposited closer to the channel than it is on smaller levees where not as much coarse sediment is transported overbank. This is because as morphology results showed the large levees which generally occur adjacent to large channels peak further from the channel than do smaller levees, and coarse sediment appears not to be easily transported upslope across levees. Also, aspects of the morphology of the levees differs somewhat from those here. Those studied by Cazanacli and Smith were wedge shaped deposits that peaked near the channel edge and in contrast to those here showed no influence of the zone of turbulent mixing at the channel-floodplain boundary on their form. The probable reason for this is a difference in the general texture of the two sets of levee deposits. The levees studied by Cazanacli and Smith were much coarser than those here with sand contents near the channel edge being generally over 50% and as high as 93% in one case. This material is probably transferred overbank by convection currents during bank spilling and probably because of the coarseness of the deposits the levees studied by Cazanacli and Smith do not seem to be responsive to overbank flow patterns when the floodplain is completely inundated as coarse material will almost certainly be quickly deposited and will not be responsive to subtle variations in floodplain flow strength. On the Cooper Creek the influence of flow patterns when the floodplain is inundated on levee form seem to be apparent and levee form variations suggest that the transfer of sediment by diffusion is a more important determinant of levee form than it seems to be on the levees along the channels of the lower Saskatchewan River studied by Cazanacli and Smith. This is probably due to the finer texture of the deposits, and the transfer of sediment by convection is perhaps not apparent in levee form because convection currents are competent to transport all size-fractions except the coarsest sediment across the levees and so do not contribute to levee accretion. The apparent differences between the Saskatchewan and Cooper levees seem to indicate that levee morphology and evolution patterns are partially dependent upon the dominant texture of the levees, a finding which suggests that levees are more diverse and interesting features than has previously been recognised.
6.4 The form and texture of expanding and contracting channel boundaries

The channel-floodplain boundary of Didhelginna Waterhole was studied to examine the influence of the channel expansions and contractions associated with waterholes on the form and texture of the channel boundary. As stated in Chapter 3, the occurrence of what appear to be levees around channels concentrating flow would be somewhat unusual. Didhelginna Waterhole shows perhaps the most striking and systematic changes of any waterhole in the study reach, and transects along the channel were located in order to ensure that both expanding and contracting sections were sampled (Figure 6.9). The morphological variation of Didhelginna Waterhole is shown in Figure 6.10. It enlarges from a capacity of 70m$^2$ and width of 35m just below the confluence marking its inception to a maximum capacity of 353m$^2$ and width of 106m just over half the 2.5km length of the waterhole downstream. No anastomosing channels join the waterhole in this enlarging section and the only floodplain-surface channels joining are very small and contribute insignificant quantities of water when compared to the capacity of the waterhole itself. Just downstream of the widest section a stable, well vegetated, 35m long and 15m wide island is formed in the centre of the channel. Downstream of the island bifurcations begin to split off from the waterhole and channel capacity decreases both gradually in the absence of bifurcations and abruptly with each bifurcation. These bifurcations are initially large, deep anabranches but contract in size rapidly away from the waterhole and generally do not persist as Figure 6.9 illustrates. Transects D1 to D3 are taken along the expanding section of channel with D3 occurring approximately at the point of maximum width just upstream of the island, and sections D4 and D5 border the contracting channel between bifurcations with D5 near the end of the waterhole (Figure 6.9).

6.4.1 Boundary morphology

Results of the surveys are shown in Table 6.7 and Figure 6.11. Where the waterhole is expanding it is bordered by prominent levees up to 1.34m high and 149m wide. Where the waterhole is contracting levees are smaller in height and width and more subdued (D4) or absent (D5). The D4 Levee is morphologically quite different to those found where the channel is expanding, and as well as being smaller in both width and height, the point of maximum height (PMH) of the D4 Levee is markedly further from the channel in both absolute and proportional terms than those of Levees D1-3.
Figure 6.9: The location of channel-floodplain boundary transects along Didhelginna Waterhole.
Figure 6.10: Morphological variation of Didhelginna Waterhole from survey data, calculated discharge by Manning's n, (Yen, 1992).
Figure 6.11: Morphology and texture trends around Didhelginna Waterhole. Locations of sections given in Figure 6.9. Only relationships significant at $\alpha=0.05$ are shown. (Figure continued overleaf.)
Figure 6.11 (continued): Morphology and texture trends of the channel-floodplain boundary around Didhelginna Waterhole. Only relationships significant at $\alpha=0.05$ shown.
Table 6.7: Channel boundary morphology, Didhelginna Waterhole

<table>
<thead>
<tr>
<th>Levee</th>
<th>Width Wl (m)</th>
<th>Height (m)</th>
<th>H/Wl</th>
<th>Backslope</th>
<th>PMH (m)</th>
<th>PMH/ Wl</th>
</tr>
</thead>
<tbody>
<tr>
<td>D1</td>
<td>132.0</td>
<td>1.34</td>
<td>0.01</td>
<td>0.01</td>
<td>4.8</td>
<td>0.04</td>
</tr>
<tr>
<td>D2</td>
<td>79.2</td>
<td>0.57</td>
<td>0.007</td>
<td>0.007</td>
<td>6.0</td>
<td>0.07</td>
</tr>
<tr>
<td>D3</td>
<td>141.4</td>
<td>1.17</td>
<td>0.008</td>
<td>0.008</td>
<td>7.6</td>
<td>0.05</td>
</tr>
<tr>
<td>D4</td>
<td>39.0</td>
<td>0.4</td>
<td>0.007</td>
<td>0.01</td>
<td>16.0</td>
<td>0.29</td>
</tr>
<tr>
<td>D5</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
</tbody>
</table>

6.4.2 Texture variations

The expanding channel Like the anastomosing channel levees described earlier, the strongest trends in levee texture around the expanding section of Didhelginna Waterhole were shown by variations in sand composition. Unlike levees surrounding anastomosing channels, those around the expanding Didhelginna Waterhole showed simple trends in sand composition which decreased with distance from the channel. On the D1, D2 and D3 Levees sand percentages are highest at the channel edge where peak sand contents are 13%, 17.5% and 20% respectively. Sand contents then decrease away from the channel, rapidly at first and then more gradually. Results of regression analyses are shown in Figure 6.11. Regression relationships are very similar between the three transects. The relationship for D1 is

\[
\%\text{Sand} = 9.49x^{-0.07} \quad (r^2=0.67)
\]

for D2 is

\[
\%\text{Sand} = 9.86x^{-0.13} \quad (r^2=0.85)
\]

and for D3 is

\[
\%\text{Sand} = 10.84x^{-0.14} \quad (r^2=0.71)
\]

where \(x\) is the distance from the channel and \(\%\text{Sand}\) is the sand content in percent. Trends in mean grain size tend to parallel those in sand content and all three levees showed a decrease in mean grain size with distance from the channel. Like on anastomosing channel levees, trends in other variables were less consistent, and only the D1 and D2 Levees showed significant trends in median grain size and silt content which decreased linearly with distance from the channel and clay content which linearly increased away from the channel.

The contracting channel Where the waterhole is contracting textural variations are very different to those around the expanding waterhole, and both the D4 and D5 transects show no significant or consistent textural trends whatsoever. Sand contents at the channel edge are 21.5% for D4 and 10.8% for D5, which are comparable to values from the D1-3 Levees. Coarse material extends much further from the channel edge where the waterhole is contracting, and both the D4 and D5 transects show much larger maximum and wider range of mean grain size.
values than the D1-3 levees. On the D4 levee sand contents vary between 10% and 25% over the 150m length of the transect and show no sign of decreasing, while on the D4 levee contents vary between 8% and 18% over the 100m transect (Figure 6.11).

6.4.3 Summary

The morphology and texture of channel boundaries differs markedly depending upon whether the channel is expanding or contracting. Results indicate that expanding channels are flanked by relatively large, prominent levees which show simple distal-finishing textural trends in mean grain size and, most markedly, sand content. Levees are generally taken to be caused by the deposition of sediment transferred from the channel by convection currents or diffusion, however in instances where channels are enlarging explaining levee formation by this process is problematic because one would not expect flow to be forced overbank from an expanding channel which could comfortably contain whatever amount of flow was supplied to it from upstream. The levees here form due to the preferential deposition of sediment near the channel by flows converging on the channel from the floodplain in the inverse of the usual process of levee formation and the forms will be referred to as 'evels' rather than 'levees' in order to simplify the following discussion. Support for this distinction is provided by the observed differences in the textural variations of levees and evels. With one exception, evels around expanding waterholes show simple trends in mean grain size and sand content which both decrease with distance from the channel, while, as described earlier, levees around anastomosing channels show more complex trends characterised by a decline and subsequent rise in mean grain size and sand content. The most probable reason for the preferential deposition of sediment near the channel-floodplain boundary by floodplain-surface flows entering the channel is the enhanced roughness of areas these areas because of the preferential growth of vegetation there which as noted in the earlier discussion of levee formation will increase surface roughness and cause sediment to be trapped. Once this occurs, an evel will begin to form and deposition will then be accelerated due to the influence of the evel on floodplain surface flows. Evel growth will lead flows into the channel to become shallower and, with constant absolute roughness, slower, which will trigger sediment deposition on the distal flank of the evel up which sediment must be transported before it reaches the channel.

The high sand contents of the evels next to the channel are consistent with the transfer by diffusion of coarse suspended sediment from the channel. Sand contents of the evels increase consistently from 13% to 20% as the channel expands, consistent with the finding from the
study of anastomosing channel levees that sand contents at the channel-floodplain boundary tend to increase with channel size. Because there is a net convective flux of flow into the channel suspended sediment from the channel can not be transported far onto the floodplain and does not progress much beyond the channel edge which accounts for the rapid decrease in sand contents away from the channel on all the eevols studied. The increasing sand content and mean grain size of eevols towards the channel is probably due to the preferential deposition of coarser sediments as flow becomes shallow and the surface more vegetated and rougher. Finer material is more likely to be transported across this zone and into the channel.

In the absence of bifurcations, channel contraction implies that flow is being forced overbank from the channel onto the floodplain surface and where this is occurring, convective currents would be expected to carry coarser channel sediments further away from the channel onto the floodplain surface. As mentioned previously, convective currents are particularly important in the transport of coarse sediments and their omission from models based only on sediment diffusion leads such models to under-predict the extent of the transport of coarse materials away from the channel (Pizzuto, 1987; Marriott, 1992). Results from the transects along the contracting channel are consistent with this and all are coarser overall than sections from expanding areas of the channel and show coarse material penetrating far onto the floodplain surface. On the D4 transect sand content 95m from the channel is 22.8% and on the D5 transect 11% at 90m. Neither of the transects showed any significant pattern of variation in any of the measured sediment characteristics, however over larger distances fining trends like those shown by the transects from scroll bar topography will probably be found as the deposits tend towards textures typical of floodplain sediments generally. Morphologically, the forcing of sediment overbank makes levees more subdued or prevents their formation altogether. The reason for this is that when large quantities of water are forced overbank the velocity of floodplain flows is sufficient to retard or prevent the deposition of the predominantly fine material which makes up the bulk of levee deposits here and so prevent levee formation, as it was suggested in the discussion of the levee texture results is true of convection currents generally. Alternatively, extension splays are characteristic of the downstream ends of a number of waterholes (see Figures 6.9 and 3.11). It seems that the formation of large levees with complex grain size variations such as those around anastomosing channels can only occur where the strength of flow leaving the channel is lower and flow is spilling out of the channel rather than spewing out.
6.4.4 Comparison of floodplain sediment-transport implications, anastomosing channel and waterhole levees

There is a clear difference apparent between the magnitude of sand transport across the floodplain implied by the results from textural analysis of waterhole channel boundaries and those from the analysis of levees along constant width anastomosing channels reported earlier. Results from the analysis of the boundary texture of contracting waterholes indicates that significant quantities of sand are transported across the floodplain surface away from the channels which contrasts with the rapid fining of coarse sediments on anastomosing channel levees. The reason for the difference here would seem to be a difference in the strength of flow in the two situations. On anastomosing channels, water is spilling out of the channel to form the levee whereas on contracting waterholes flow, presumably large quantities of flow judging by the rapid decreases in channel size seen there, is being forced overbank by the contraction of the waterhole. Flow from contracting waterholes would be expected to be stronger and to transport coarser materials further from the channel which texture results indicate is the case. The textural fining of levees around expanding waterholes was taken to represent an increase in the preferential deposition of sand approaching the channel due to increased surface roughness caused by vegetation growth and by feedbacks initiated by levee development which cause flow to become shallower and less competent. Together with the results from levee texture analysis this implies that significant quantities of sand may be transported across the floodplain surface where flood flows are deeper and fast flowing but not where inundation depths are low. It is not possible to quantitatively evaluate the predictions of the Shield’s equation for the threshold depth of flow necessary for sand transport but, qualitatively, the results here suggest that the predictions are relatively accurate.
7 FLOODPLAIN-SURFACE CHANNELS

7.1 Types and controls on distribution

7.1.1 Introduction

As described in Chapter 3 there are two main types of floodplain-surface channel patterns found on the Cooper Creek floodplain; the braid-form pattern which occupies 44% of the floodplain surface in the study reach and the reticulate channel pattern which occupies 39%. Seventeen per cent of the floodplain-surface is unchanneled. In this chapter the determinants of floodplain-surface channel pattern formation and location will be investigated and a model of pattern formation proposed. Following this the morphology of the reticulate and braid-form patterns will be considered in more detail, and finally evidence of floodplain accretion patterns will be briefly reviewed.

7.1.1.1 Floodplain-surface channel occurrence and formation

Floodplain-surface channels are one of three levels of alluvial relief recognised by Arabian and Chalov (1998) along with the floodplain surface and inset low-water channels. A floodplain is defined here geomorphically as the alluvially constructed surfaces adjacent to a channel or channels, rather than in an engineering sense as the area of land inundated by a flood with a particular return interval (Nanson and Croke, 1992). Floodplain-surface channels can be divided into two basic types; those formed in situ as channels on the floodplain surface and those which become floodplain-surface channels following changes in the low water channel system. These include abandoned sections of low-water channel and palaeochannels. Examples of in-situ formed floodplain-surface channels are crevasse channels and erosional swales (e.g. Thornbury, 1968; Winker, 1978). Although active floodplain-surface channels tend to be smaller than adjacent low water channels, some channels such as the Solimões River form large floodplain-surface channels which may appear similar to low water channels but are only active during flood events (Mertes et al., 1996). Actively migrating floodplain-surface channels on the Solimões River may be up to 1800m wide and 15m deep in comparison to the main channel which may reach 6km in width and 20m in depth (Mertes et al., 1996). Abandoned low-water channel sections are an important component of many floodplain surfaces (e.g. Lewin, 1978, 1992; Alexander and Marriott, 1999) as their removal by either infilling or lateral migration may be very slow, leading them to persist for a long time after abandonment.

Studies of extensively channeled floodplains are as rare as those of the floodplains themselves seem to be. Winker (1978) studied the channeled floodplains of the Little River Drainage Basin...
in Texas. In the headwaters of the Little River Basin the channels are entrenched into narrow floodplains and when the rivers emerge onto the coastal plain some develop broad, extensive, fine-grained floodplains into which numerous low sinuosity, shallow, subparallel floodplain-surface channels are incised. The floodplain-surface channels range from 0.6-6.6m deep, may be tens to hundreds of metres wide, and become shallower upvalley. Most are only active when the floodplain is inundated, however the larger floodplain-surface channels may become integrated into the low-water drainage network. Energy is an important determinant of their formation, and the channels form only on floodplains with relatively high gradients (~0.00056) and are not found at lower gradients. The channels form during high magnitude flood events when, due to the susceptibility of the basins to high magnitude flash flooding, depths of floodplain inundation may average 6.6m. Although formed during extremely high magnitude events, the gross form of the floodplain is not markedly changed by a single flood and the channels develop gradually, "essentially as bedforms, which increase the efficiency of the floodplain for carrying overbank flow" (Winker, 1978, p.32).

Mertes et al. (1996) report that floodplain-surface channels occur discontinuously along the Solimões River and are confined to particular geomorphic situations. The slope of the channel is an important determinant of floodplain-surface channel occurrence and floodplain-surface channels form on steeper slopes where the main channel is mobile and where floodplain-surface channels can erode sand and be mobile (Mertes et al., 1996). Where the main channel is stable, prominent levees are formed and the channel constructs a flat, featureless floodplain by gradually burying scroll topography. Structure also controls floodplain-surface channel occurrence as structural arches promote entrenchment of the river and restrict main channel movement and floodplain-surface channel formation (Mertes et al., 1996). (It is not clear whether the reaches of the Solimões River developing floodplain channels are those classified by Baker (1978) as anastomosing, however this is likely to be the case and suggests that, interestingly, levee formation is not only associated with the development of anastomosis as stated by a number of the anastomosing models reviewed in Chapter 1, but may also retard it.)

In broad terms, floodplain-surface channel formation, like channel pattern formation generally, is controlled by flow discharge, valley gradient, sediment load and bank strength (e.g. Ferguson, 1981, 1987; Knighton, 1984; Nanson and Knighton, 1996). The above examples illustrate that energy is commonly a critical determinant of floodplain-surface channel occurrence with higher energy areas of floodplain more channeled than lower energy areas. However, floodplain-surface channel formation is not confined to high-energy floodplains and Nanson and Croke (1992) recognise five types of channeled floodplain spanning the energy
gradient. Three of the five types are high energy floodplains typically located in steep headwater areas and confined by bedrock (Nanson and Croke's (1992) type A1, A2, and A4 floodplains). These are disequilibrium floodplains responsive to extreme events which rapidly erode deep back channels and scour holes into floodplain surfaces. The fourth type of channeled floodplain is the wandering gravel-bed river floodplain (Type B2) which is a medium energy floodplain "formed by relatively regular flow events in relatively unconfined valleys" (Nanson and Croke, 1992, p.466), and the fifth type are low energy anastomosing river floodplains (Type C2). In terms of floodplain channel formation, the effective discharge is determined by the magnitude of overbank flow displacement and the channeling of anastomosing floodplains is explained by the fact that, as described in Chapter 1, anastomosing rivers are generally subject to large flood events due to high bank resistance constraining channel capacities and seasonal or highly irregular flow regimes. Channeled floodplains would seem to be relatively common on anastomosing rivers and the development of floodplain channels is included as a stage or precursor to the formation of a new anabranch in the models of Schumann (1989), Schumm et al. (1996) and the gradual avulsion process proposed in this study. However, the formation of channeled floodplains is not necessary for anastomosis as anabranches may form by an avulsion process over a single event.

7.1.1.2 Cooper Creek floodplain-surface channels

A number of factors may influence the formation and distribution of floodplain-surface channel patterns in the study reach. However, one factor that is probably not significant is valley slope which remains approximately constant over the reach at around 0.00015. The configuration and extent of the floodplain surface may be important through its influence on the strength and distribution of flood flows, and a brief examination of the study reach suggested a relationship between floodplain width and floodplain-surface pattern occurrence that will be examined in more detail here. If the floodplain is very wide then flood flows will be dispersed over a much wider area than if the floodplain is very narrow and overbank flow depth and velocity will tend to be lower. (The term 'flow power' will be used here to refer to the depth and velocity of overbank flow as they are the two main influences on the power [Wm\(^{-2}\)] of the flow as slope is constant.) The possible influence of the configuration of floodplain landforms on channel formation is illustrated by the formation of floodplain-surface waterholes at points of constriction on the floodplain surface (Chapter 3). Another factor considered to be of possible importance is transmission losses which may account for 100% of low flows and an average of 75% of high flows over the reach (Knighton and Nanson, 1994a). Such losses will cause, other
things being equal, flood flows to decrease in depth and velocity downstream over the study reach.

As reviewed in Chapter 3, the fundamental influence on reticulate channel formation was previously thought to be wetting frequency through its influence on the development of large gilgai, and for this reason the reticulate pattern was said to be confined to lower areas of the floodplain which were presumed to be wetter (Whitehouse, 1948; Rundle, 1977). Prominent gilgai do not form in braid-form or unchanneled areas where small-scale surface relief is much less pronounced than in reticulate areas. In unchanneled areas the dominant scale of surface irregularity is very small (Figure 3.4), while in braid-form areas although small scale surface irregularities are comparable to those in unchanneled areas, large scale variations in surface form occur due to the presence of braid-form islands varying in width between approximately 85m and 650m and in maximum relief up to about 1m. As well as elevation and wetting frequency, variations in soil properties and erosion rates could be expected to influence the distribution of the reticulate pattern through their influence on gilgai formation (Hallsworth et al., 1955; Elberson, 1983; Paton et al., 1995). The influence of the above factors on floodplain-surface pattern formation will be considered next.

7.1.2 Methodology

A detailed analysis of the distribution of channel types throughout the study reach was undertaken. Fifty seven cross-floodplain transects were drawn covering the study reach between Windorah and Nappa Merrie with the floodplain surfaces along the transects classified according to channel-pattern type. Transects were drawn perpendicular to the centreline of the floodplain and spaced approximately 5km apart, with the proviso that transects were not permitted to cross in order to avoid counting the same area of floodplain twice. This meant that where significant changes in valley direction occurred, the spacing between the mid-points of transects sometimes exceeded 5km, however, in all but 4 cases the spacing between transect mid-points was below 9km with a maximum of 15.4km. The classes recognised were unchanneled floodplain, aeolian dunes, reticulate and braid-form patterns. Illustrations of the reticulate and braid-form patterns are given in Figures 3.13 and 3.17 respectively. For the purposes of analysis the length of transect occupied by dunes was subtracted from total floodplain width in order to obtain a value for the maximum width of floodwaters. Anastomosing channels were not considered as a separate category because they occupy such a
small area of the floodplain surface - as little as 0.5% and a maximum of 3% of transect width. Regression analyses were undertaken to determine the nature and significance of relationships between the occurrence of channel pattern types, floodplain width and distance downreach from the head of the study reach. The study of the influence of elevation on the distribution of channel patterns utilised survey data collected by Santos Pty. Ltd. which provides detailed and extensive topographic information on the floodplain surface. Floodplain elevation and extent are not independent because a very wide floodplain may be effectively much narrower if most of it too high to be inundated and flood flows are confined to a small proportion of the full floodplain extent. However, survey data coverage is not sufficiently extensive to enable a 3-D model of floodplain topography to be constructed that would enable both relief and extent to be considered simultaneously. Firstly variations in soil properties between pattern types are considered. The influence of elevation on the occurrence of pattern types is then considered, followed by floodplain width and transmission losses. From the results of this analysis a model of floodplain-surface pattern formation is formulated.

7.1.3 Soil properties

Analysis was undertaken on soil samples from reticulate, braid-form and unchanneled areas to assess both whether there are differences in soil type between sample sites and whether large scale sediment sorting between areas of different floodplain-surface patterns occurs similar to the sorting seen over much smaller scales adjacent to anastomosing channels and waterholes. It was only possible to sample one unchanneled area because of the relative rarity of such areas, however due to the large size of the area it was possible to collect a number of samples spaced widely over distances exceeding 1km. Two areas of reticulate channels were sampled, and samples were taken from 5 braid-form areas. All soils were vertisols showing evidence of cracking and swelling and there was no difference in soil colour between pattern types which varied over a narrow range from greyish-brown (2.5Y 5/2) to olive brown (2.5Y 4/3). Ternary plots of sediment sample sizes are shown in Figure 7.1 and statistical testing using the Tukey-Kramer HSD test confirmed that there were no significant differences between areas of differing pattern type in terms of mean and median grain size or sand, silt and clay content (α=0.05). In summary, soil types and textures are similar between the different pattern areas.

7.1.4 Elevation and distribution

7.1.4.1 Reticulate areas

As stated previously the claim that reticulate networks are confined to low floodplain areas (Whitehouse, 1948; Rundle, 1977) has been made from appearance of many reticulate areas as
Figure 7.1 Sediment texture plots of surface samples from different floodplain environments: reticulate, unchanneled and braid-form areas. Note the clear lack of a textural distinction between samples from the different areas.
swampy ground rather than from actual topographic data. The recent collection of detailed survey data for large areas of the floodplain combined with satellite imagery of flood events enables the influence of elevation on the occurrence of reticulate networks to now be examined more rigorously. Figure 7.2 shows the location and topography of a prominent reticulate area just south of the floodplain divergence to Lake Yamma Yamma which receives flow only during overbank flood events. The floodplain here is relatively narrow, and the topographic profile across the area indicates that the reticulate channels are developed in the significantly higher eastern section of floodplain away from the anastomosing channels which are confined to the lower western side of the floodplain. The observation that the reticulate pattern and anastomosing channels do not co-occur is generally true for the study reach and large anastomosing channels are never found within floodplain areas occupied by reticulate patterns. An examination of satellite imagery of flood events shows that water does not pond in this area for any significant length of time following flood events.

Figure 7.3 shows the occurrence of two areas of reticulate channels across the wide area of floodplain near the confluence of the Cooper Creek and the Wilson River; one towards the lower south-eastern end of the transect, and the other in the higher, more central area of the floodplain near an area of unchanneled floodplain. An examination of satellite imagery taken following the passage of a floodwave shows that water does not pond in the reticulate area in the centre of the floodplain but that water may pond in the reticulate area towards the lower, eastern end of the transect. If the Wilson River and Cooper Creek both experience simultaneous, high magnitude flooding then this area can be inundated for months, however usually periods of inundation are much shorter.

The above data illustrate that while reticulate networks do occur on low areas of the floodplain surface they are certainly not confined to those areas as previous authors have suggested and also occur on high areas of floodplain. Floodplain elevation is clearly not the controlling factor in the distribution of the reticulate pattern, and since the reticulate pattern and the formation of large gilgai are inextricably linked here this also implies that the formation of prominent gilgai is also not confined to low areas of the floodplain where water ponds and that the determinants of prominent gilgai formation here are also more complex than previous authors believed.

7.1.4.2 Unchanneled areas

Unchanneled areas are confined to the higher parts of wider areas of floodplain which flood very infrequently. The unchanneled area in Figure 7.3 is located in the highest region of the
Figure 7.2: Location and topographic profile of a reticulated, high area of floodplain south of Lake YammaYamma. Braid-form channels occur in the lower areas of floodplain towards the western side of the transect. Dunes indicated by ‘D’.
Figure 7.3: Floodplain topographic profile showing the occurrence of reticulate and braid-form patterns in both high and low areas of the floodplain. Dotted/dashed line in topographic profile shows surface elevation trend. Survey data has been rounded to 0.5m intervals accounting for the stepped appearance of the profile.
floodplain. Figure 7.4 shows the prominent, high unchanneled area in the south of the study reach which is only dissected in one or two places by small channels such as the Watawarra Channel. Satellite imagery and gauging records were used to estimate the frequency of flooding of the area. Gauging records at Nappa Merrie extend back to 1949 although the extent of their completeness prior to 1967 is uncertain (Knighton and Nanson, 1994a). Although satellite imagery does not provide continuous coverage the frequency of imaging of an area (once every 16 days for Landsat TM imagery) is sufficient to capture either the event itself or its signature after effects (increased ephemeral vegetation growth). Satellite imagery showed inundation of the area during a flood event in 1990 when the peak gauge height at Nappa Merrie was 9.38m but not in 1989 when peak gauge height was 6.9m. Using these figures as a guide to the size of flood necessary to inundate the area suggests that the area has been flooded only 4 times since gauging records began in 1949. Another prominent area of unchanneled floodplain occurs in the north of the study reach east of Lake Yamama Yamma (Figure 7.5). Satellite imagery of the 1974 flood event described by Robinove (1978) and Baker (1986) showed that sections of this area were inundated by the 1974 flood and as the 1974 flood was the largest on record that the areas have not been inundated since 1949.

7.1.4.3 Braid-form areas
Like reticulate areas, braid-form areas may occur in both high and low areas of the floodplain as Figures 7.2 and 7.3 indicate. Braid-form areas co-occur with anastomosing channels where inundation frequencies and overbank flow powers are generally higher due to the anastomosing channels being the sources of overbank flows.

7.1.5 Floodplain width and transmission losses

Reticulate patterns Results of this analysis are given in Table 7.1 and Figure 7.6. Floodplain width is a good predictor of the width of the reticulate pattern and the strong positive relationship shows that as floodplain width increases so does the extent of reticulate networks found. Distance downreach alone is not significant in explaining the variation in reticulate pattern occurrence, but considering both floodplain width and distance downreach together produces a slightly stronger relationship than floodplain width alone ($r^2=0.71$ c.f. $r^2=0.69$). However, because this increase is so small one cannot conclude that transmission losses are a significant determinant of the extent of reticulate channels, especially as it is impossible to decrease the significance of a relationship by considering additional variables. Both floodplain width and distance downreach are very poor predictors of the percentage of the
Figure 7.4 Location, topographic profile and satellite image (SPOT false colour composite) of an unchanneled high area of floodplain in the south of the study reach. On the satellite image recently flooded, vegetated areas appear red and unvegetated areas grey-blue. Satellite imagery and gauging records suggest that the area has been inundated only 4 times since 1949.
Figure 7.5 SPOT false colour composite image of prominent unchanneled areas of floodplain just south of Lake Yamma Yamma. Images of the area were studied by Robnove (1978) and Baker (1986). The area is shown during the flood event of March 2000 after the flood peak has passed through. Blue areas are inundated, red represents vegetation, the surrounding valley side appears pink and yellow, dark areas are thought by Robnove to represent wet but not inundated floodplain, and highly reflective, lighter areas represent dry, predominantly unvegetated floodplain. Stage data suggests that the lower area has not been inundated since before 1949 when gauging records began.
Figure 7.6 The relationship between variations in floodplain and floodplain-surface pattern extent over the study reach.
floodplain occupied by reticulate channels. Considering both factors together improves the relationship significantly, however the percentage of floodplain occupied by reticulate channels is still much less predictable using these variables than is the absolute extent of the reticulate pattern ($r^2=0.32$ c.f. 0.71).

Table 7.1: Results of reticulate pattern occurrence (RET) regression analyses (n=57). * Denotes not significant at $\alpha=0.05$.

<table>
<thead>
<tr>
<th>Response variable</th>
<th>Predictor variable/s</th>
<th>Equation</th>
<th>$r^2$</th>
</tr>
</thead>
<tbody>
<tr>
<td>Width of reticulate pattern ($W_{RET}$), km.</td>
<td>Floodplain width ($W_F$), km.</td>
<td>$W_{RET}=2.943+0.49W_F$</td>
<td>0.69</td>
</tr>
<tr>
<td>$W_{RET}$</td>
<td>Distance downreach (Dd), km.</td>
<td>$W_{RET}=14.5714-0.018Dd$</td>
<td>0.04*</td>
</tr>
<tr>
<td>$W_{RET}$</td>
<td>$W_F$ and Dd</td>
<td>$W_{RET}=-7.32+0.54W_F+0.017Dd$</td>
<td>0.71</td>
</tr>
<tr>
<td>$%_{RET}$ transect reticulate</td>
<td>$W_F$</td>
<td>$%_{RET}=21.370+0.465W_F$</td>
<td>0.16</td>
</tr>
<tr>
<td>$%_{RET}$</td>
<td>Dd</td>
<td>$%_{RET}=28.120+0.04Dd$</td>
<td>0.04*</td>
</tr>
<tr>
<td>$%_{RET}$</td>
<td>$W_F$ &amp; Dd</td>
<td>$%_{RET}=0.028+0.69W_F+0.086Dd$</td>
<td>0.32</td>
</tr>
</tbody>
</table>

**Braid-form patterns** The results for this pattern differ in a number of ways to those from the analysis of reticulate channel occurrence (Table 7.2, Figure 7.6).

Table 7.2: Results of braid-form pattern (BRA) occurrence regression analyses (n=57). * Denotes not significant at $\alpha=0.05$.

<table>
<thead>
<tr>
<th>Response variable</th>
<th>Predictor variable/s</th>
<th>Equation</th>
<th>$r^2$</th>
</tr>
</thead>
<tbody>
<tr>
<td>$W_{BRA}$</td>
<td>$W_F$</td>
<td>$W_{BRA}=5.43+0.25W_F$</td>
<td>0.21</td>
</tr>
<tr>
<td>$W_{BRA}$</td>
<td>Dd</td>
<td>$W_{BRA}=21.47-0.05Dd$</td>
<td>0.34</td>
</tr>
<tr>
<td>$W_{BRA}$</td>
<td>$W_F$ and Dd</td>
<td>$W_{BRA}=15.14+0.15W_F+0.039Dd$</td>
<td>0.41</td>
</tr>
<tr>
<td>$%_{BRA}$</td>
<td>$W_F$</td>
<td>$%_{BRA}=74.86-0.81W_F$</td>
<td>0.24</td>
</tr>
<tr>
<td>$%_{BRA}$</td>
<td>Dd</td>
<td>$%_{BRA}=64.57-0.08Dd$</td>
<td>0.07</td>
</tr>
<tr>
<td>$%_{BRA}$</td>
<td>$W_F$ &amp; Dd</td>
<td>$%_{BRA}=114.55-1.22W_F-0.16Dd$</td>
<td>0.47</td>
</tr>
</tbody>
</table>

Similar to the results for the reticulate pattern, as floodplain width increases so does the extent of the braid-form pattern, however, as is clear from Figure 7.6 the relationship is much less strong ($r^2=0.21$ c.f. 0.69) and the rate of increase of pattern extent almost half as rapid (0.25*$F_p$ c.f. 0.49*$F_p$). This indicates that the extent of the braid-form pattern is less influenced by changes in floodplain width and more constant in extent downstream than is the reticulate pattern. Whereas floodplain width was the best single predictor of reticulate extent, distance downreach is the best single predictor of braid-form channel extent and the relationship indicates that the extent of braid-form pattern development decreases with increasing distance downreach and increasing transmission losses. Results of multiple regression analysis with both predictor variables shows that the predictability of braid-form extent is significantly increased but remains relatively poor compared to the results for the reticulate pattern.
Considering the percentage of floodplain occupied by braid-form channels, as with reticulate occurrence, two poor relationships combine to produce one markedly stronger relationship and the form of the relationship indicates that \( \%_{\text{BRA}} \) decreases with increased floodplain width and (very weakly) with distance downreach. To illustrate this tendency transects were divided into two approximately equal groups by length. A mean of 40\% of the larger (>25km, \( n=29 \)) transects were occupied by braid-form patterns while the smaller (<25km, \( n=28 \)) transects were dominated by braid-form channels which occupied an average of 63\% of transect width, a difference that is statistically significant (Tukey Kramer HSD test, \( \alpha=0.05 \)).

**Unchanneled areas** Individually, distance downreach is not a significant predictor of either the absolute extent or percentage of the floodplain that is unchanneled (Table 7.3).

### Table 7.3: Results of unchanneled area (UNCH) occurrence regression analyses (\( n=57 \)). *

<table>
<thead>
<tr>
<th>Response variable</th>
<th>Predictor variable/s</th>
<th>Equation</th>
<th>( r^2 )</th>
</tr>
</thead>
<tbody>
<tr>
<td>( W_{\text{UNCH}} )</td>
<td>( W_p )</td>
<td>( W_{\text{UNCH}}=-2.48+0.25W_p )</td>
<td>0.36</td>
</tr>
<tr>
<td>( W_{\text{UNCH}} )</td>
<td>( Dd )</td>
<td>( W_{\text{UNCH}}=-4.77+0.00095Dd )</td>
<td>0.0002*</td>
</tr>
<tr>
<td>( W_{\text{UNCH}} )</td>
<td>( W_p ) and ( Dd )</td>
<td>( W_{\text{UNCH}}=7.82+0.31W_p+0.021Dd )</td>
<td>0.42</td>
</tr>
<tr>
<td>( %_{\text{UNCH}} )</td>
<td>( W_p )</td>
<td>( %_{\text{UNCH}}=3.77+0.34W_p )</td>
<td>0.10</td>
</tr>
<tr>
<td>( %_{\text{UNCH}} )</td>
<td>( Dd )</td>
<td>( %_{\text{UNCH}}=7.30+0.04Dd )</td>
<td>0.05*</td>
</tr>
<tr>
<td>( %_{\text{UNCH}} )</td>
<td>( W_p ) and ( Dd )</td>
<td>( %_{\text{UNCH}}=14.52+0.53W_p+0.07Dd )</td>
<td>0.23</td>
</tr>
</tbody>
</table>

Both the percentage (\( \%_{\text{UNCH}} \)) and extent (\( W_{\text{UNCH}} \)) of transects unchanneled are positively related to floodplain width. For the percentage of floodplain that is unchanneled this relationship is very weak and in the case of the absolute extent of unchanneled floodplain the relationship is still rather poor (\( r^2=0.36 \)). The weakness of these relationships is due in part to the large numbers of zero values in the data set and, as Figure 7.6 illustrates, the actual influence of floodplain width on the extent of unchanneled areas is much stronger than these relationships suggest. Transects with no areas of unchanneled floodplain had a mean width of 23.44km compared to transects with unchanneled areas which had a mean width of 34.8km, a difference that is statistically significant (Tukey-Kramer HSD test, \( \alpha=0.05 \)). The combination of both floodplain width and distance downreach as predictor variables markedly and significantly improves the prediction of both the extent and percentage of unchanneled floodplain which increase with increasing floodplain width and distance downreach. This suggests that distance downreach is important but that its influence is not apparent individually because it is masked by the greater importance of unrelated variations in floodplain width.
7.1.6 Controls on gilgai formation and floodplain-surface channel pattern

The marked variability in the scale of gilgai development has been noted previously, and the preceding discussion of the distribution of floodplain-surface channel patterns enables a number of conclusions regarding the determinants of gilgai formation to be drawn. The highest, unchanneled alluvial surfaces which are only rarely inundated by flood flows do not develop prominent gilgai, indeed they usually have relatively smooth surfaces. As all areas possess similar soils, this implies that the development of prominent gilgai is not caused by rainfall wetting and suggests that fairly frequent inundation by flood flows is necessary for prominent gilgai to form. The ponding of flow cannot be an important control on the development of large gilgai because large gilgai and reticulate networks are developed in areas where flow does not pond, and frequent inundation by flood flows cannot be the only control on the development of large gilgai because braid-form areas which occur proximal to anastomosing channels are the most frequently inundated areas of the floodplain, and large gilgai do not form in braid-form areas. Although the expansion of floodplain width results in an expansion of the extent of all floodplain-surface patterns, reticulate and unchanneled areas expand most and occupy a greater proportion of wider than narrower floodplain transects, while the opposite is the case for braid-form channels. Particularly narrow floodplain transects may have no reticulate or unchanneled areas, while all transects, no matter their width possess areas of braid-form channels.

Results suggest a model for gilgai formation and floodplain-surface channel development based on inundation frequency and the power of flood flows (Figure 7.7). The dominance of braid-form channels in narrow areas of the floodplain suggests that this pattern tends to occur in areas which are frequently inundated and which experience relatively high depths and velocities of overbank flows. This is because of the proximity of anastomosing channels which are the source of much overbank flow and which co-occur with the braid-form pattern. In these areas the sculpting of the floodplain surface by powerful flood flows prevents the expression of large gilgai which would otherwise tend to form due to the frequent inundation of the areas. In the wider areas of floodplain, overbank flows tend to be shallower and weaker, and where the floodplain is frequently inundated pedogenic processes and gilgai formation are more significant as determinants of floodplain-surface form because the organisation of the floodplain surface by pedogenic processes is not prevented by fluvial erosion. The development of prominent gilgai leads to the concentration of floodflows in gilgai depressions, the incision of channels linking depressions and the formation of a reticulate channel pattern (Mabbutt,
Figure 7.7: The continuum model of floodplain-surface pattern distribution. The divisions between patterns are approximate only and probably show some degree of overlap. The shaded area indicates the general range of combinations of the variables and that areas subject to high frequencies of inundation tend also to be subject to more powerful overbank flows during flood events of any given size than areas less frequently inundated.
1967; Rundle, 1977). It is likely that if gilgai were absent then reticulate areas would be unchanneled as the concentration of floodwaters by the gilgai pattern seems necessary for channel formation in these areas. Sections of wide floodplain which are not frequently inundated due to their height do not form prominent gilgai because of low inundation frequencies and are not otherwise channeled because, on the rare occasions they are inundated, flow depths and velocities are very low and the low surface relief does not tend to concentrate floodwaters. This model is consistent with the distribution of floodplain-surface channel patterns at the large (reach) scale.

7.1.7 Transitions between floodplain-surface channel pattern types

As a test of the proposed model the consistency of model predictions with smaller-scale variations in floodplain-surface pattern occurrence was considered in order to examine whether relationships inferred at the large scale hold when particular local situations are considered as they should if the model is correct. Transitions between pattern types offer a good way of evaluating this as the model predicts that particular changes in overbank flow properties will accompany morphological changes. Two kinds of transition were studied; between braid-form and reticulate patterns, and between the braid-form and unchanneled floodplain.

7.1.7.1 The braid-form - reticulate pattern transition

The transition from a braid-form pattern to a reticulate pattern downstream occurs in a number of places in the study reach and was studied south of Naccowlah Waterhole where a prominent braid-form pattern can be observed to change into a reticulate pattern approaching Wilson’s Swamp. Field observations in this area were made to augment the air photograph analysis (Figure 7.8). The floodplain is expanding here and the orientation of the braid-form channels shown in Figure 7.8 suggests that, as one would expect, flow is diverging consistent with the predictions of the model. In the intermediate stage, between the reticulate and braid-form patterns, individual braid-form channels become ‘fretted’ and are joined along their length by large numbers of small channels which do not persist for long distances away from the braid-form channels (Figure 7.8). It can be inferred that these channels do not represent local floodplain drainage as if this was the case they would be expected to occur around every braid-form channel, which they do not, as floodplain slope, surface materials and probably rainfall are approximately constant between braid-form areas. Also, the fretted channels tend to occur around smaller, more subdued braid-form islands where overbank flow from local rainfall would have a smaller catchment area than the islands immediately upstream. The frets would
Figure 7.8: The transition between braid-form and reticulate floodplain-surface patterns near the confluence with the Wilson River. Note the dispersion of flow indicated by the paths of the braid-form channels accompanying the transition.
seem to represent the beginning of the influence of gilgai on the braid-form pattern and form when flowing water is diverted from the channels into depressions of the gilgai covering the braid-form islands. These frets are scoured for a short distance before the topographic rise of the braid-form island prevents flow from continuing away and forming longer channels, as Figure 7.9 from a different area of the study reach illustrates. Further downstream the relief of the braid-form islands becomes more subdued, the gilgai more pronounced, and the fretted channels grow longer, branch and become more dense, and begin to resemble a reticulate channel system. This becomes progressively more connected until eventually no trace of the braid-form pattern can be seen.

Another transition between a braid-form pattern and a reticulate pattern studied occurred adjacent to the valley side and showed similar trends to those described above (Figure 7.10). Here the reticulate pattern forms when flow expands into an embayment in the valley side, a situation in which reticulate patterns commonly occur. As well as the fact that flow must diverge in order to enter it, the embayment is essentially a cul-de-sac for flow and the fact that flow cannot pass through the embayment but must come back out after entering it restricts the velocity and energy of the flow. Like the previous example, the formation of the reticulate pattern in this and similar situations is consistent with the model proposed in that the reticulate pattern is located where floodplain morphology suggests that overbank flows will be lower in energy than over the adjacent braid-form areas.

7.1.7.2 The braid-form – unchanneled area transition

The transition between from an area of braid-form pattern to an unchanneled area downstream was studied in the lower area of the study reach near the Watarra Channel (Figure 7.11). An intermediate zone where braid-form channels are fretted is present between the two pattern types as at the transition between the braid-form and reticulate patterns. This fretted phase presumably forms by the same process described above and represents gilgai becoming a more dominant influence of floodplain surface form and channel pattern as flows weaken. The occurrence of an intermediate fretted phase is consistent with the model which suggests that the unchanneled areas, reticulate areas and braid-form areas occur along a continuum of increasing flood energy and inundation frequency, however the reason a fully developed reticulate network is not formed here is unclear. In summary, both this and the previous examples of pattern transitions are consistent with the predictions of the model presented above in Section 7.1.6.
Figure 7.9: The fretting of braid-form channels around a braid-form island and the beginning of reticulate pattern formation at the transition between braid-form and reticulate pattern areas, South Galway area north of Lake Yamma Yamma. The topographic rise of the braid-form island prevents a fully integrated reticulate network forming. This pattern represents the beginning of the influence of gilgai on channel form, further downstream the braid-form islands become more subdued and a reticulate pattern is formed across the entire surface.
Figure 7.10 Transition from braid-form to reticulate pattern, MVS area north of Lake Yamma Yamma. A reticulate pattern is formed by flow diverging into an embayment in the valley side as commonly occurs over the study reach.
Figure 7.1: Transition between braid-form and unchanneled floodplain areas, flow from the NE to SW. U denotes an unchanneled area, B braid-form channels, I braid-form islands and A an anastomosing channel running to the north of the area. The unchanneled area shown is part of the large central high area of unchanneled floodplain also shown in Figure 7.4. Note that the braid-form channels between the islands are fretted, and that the islands become less distinct downstream until fading out.
7.2 Reticulate network morphology

7.2.1 Introduction

While the formation of individual reticulate channels is understood to be due to the incision of channels in depressions between gilgai mounds (Mabbutt, 1967; Rundle, 1977), the factors controlling the morphology of networks of reticulate channels have not been investigated. Reticulate channels do not occupy every gilgai depression, and the density of reticulate channels over an area, their size, organisation and orientation varies greatly and systematically (Figure 3.13). The morphology of a number of reticulate networks over the study reach was investigated in order to help understand the causes of variations in network form. The great complexity and intricacy of the networks makes considering all the links which together make up the reticulate pattern practically impossible. It was estimated that areas where reticulate channels are densely developed typically contain over 30,000 channel links per square kilometre, most of which extend only over very short distances. An attempt was made to automatically extract the channels from remotely sensed images of the networks by applying edge extraction algorithms to SPOT images, however this was not successful. The reason for this is that when the channels are not flowing, as was the case during the period when the image was acquired, the difference in tone between the channels and surrounding floodplain seen on remotely sensed images is often slight. While the human eye can easily detect the channels, the extraction program was able to successfully detect only some links and was not able to recognise most of these as continuous channels, even if this was the case. For these reasons automated image analysis was not able to produce useful representations of network form. A novel method of network analysis was developed here which uses the identification of different classes of link to enable the representation of the networks in a simplified and tractable form.

7.2.2 Reticulate network topology

The study of drainage network form has provided a rich vocabulary of terms describing aspects of networks and network structure (e.g. Horton, 1945; Shreve, 1966, 1967), however constructed as it is to describe idealised, loopless networks this vocabulary is not suitable to adequately describe the complexities of reticulate or similar networks (Shreve, 1966). For this reason two novel terms are proposed here to enable the adequate description of reticulate network form. Three different classes of link occur in reticulate networks (Figure 7.12). Exterior links (EL), after Shreve (1967), are links which terminate at one end and at the other join another link. In dendritic drainage networks the termination point of exterior links is
Figure 7.12 Definition of reticulate network link types: LIL - looped interior links, TIL - terminal interior links, EL - exterior links.

Figure 7.13 Oblique aerial photographs of reticulate networks showing clear polygon development and the hierarchy of network links defined in Figure 7.12. Note that LIL links are larger than TIL and EL links and that in the lower figure reticulate polygons do not extend up to network margins.
generally termed a source to indicate that it represents a point at which water is fed into the
drainage network. However, in reticulate networks these terminations do not represent points at
which water is fed into the network, but rather the limit to which water is fed by the network,
and so it is inappropriate to describe them as sources. Interior links, again after Shreve (1967),
are links which terminate at both ends at junctions with other links. Two novel classes of
interior link are differentiated: looped interior links (LIL) are interior links which form part of
a closed loop of links, while terminal interior links (TIL) do not form part of a closed loop and
inevitably lead, however indirectly, to an exterior link (Figure 7.12). An area bounded by a
closed loop of LILs that cannot be subdivided into a number of smaller areas also bounded by
LILs is termed a reticulate polygon. Reticulate networks can be divided into a transitive
component which is a continuous system that transports flow across the area (LIL links) and a
spacefilling component which does not (TIL and EL links). These classes of link and the
hierarchical structure of reticulate networks are clearly visible on Figure 7.13. Representing
reticulate networks as mosaics of contiguous reticulate polygons provides a replicable and
tractable way of analysing reticulate channel patterns by reducing the number of elements
which must be considered to practically manageable levels as there are many order or
magnitude fewer reticulate polygons than there are links in any given reticulate network. The
density of reticulate polygons reflects the density of links generally because, as would be
expected, reticulate polygons are denser where links in general are denser.

7.2.2.1 Methods

Four study areas were selected to encompass all the various situations in which reticulate
channels are found within the study reach - on unconfined areas of floodplain, between dune
constrictions, and contiguous with the valley side. The width $W_p$, length $L_p$, area $A_p$ (measured
by digital planimeter) and elongation $E_p$ (defined as $L_p/W_p$) of the reticulate polygons were
measured and calculated from ortho-rectified aerial photographs as defined in Figure 7.14. The
orientation of the long axes of reticulate polygons was considered and a G.I.S. was used to
investigate the variation of reticulate polygon form over the study areas. Spatially averaged
values of polygon descriptor variables were calculated for sub-areas of the study networks and
surface plots interpolated from these values to enable trends to be displayed. The use of
braiding indices (e.g. Rust, 1978; Friend and Sinha, 1993) to describe the reticulate networks
was considered, however these measures were not applied because it was concluded that they
would not produce useful results. In contrast to braided channels which generally flow between
easily definable banks, reticulate networks are spatially extensive. Most braiding indices have
concentrated on capturing the intensity of braiding by using some measure of the total length of
Length $L$, Area $A$, Width $W$.

Figure 7.14 Definitions of reticulate polygon morphology characteristics measured and examples of polygon form from the Naccowlah area. Note that polygon width $W$, is the greatest distance that can be measured perpendicular to polygon length $L$, and the complexity and lack of apparent streamlining of polygon form.
channels or islands for a given length of braided channel, and applied to reticulate networks the application of indices derived in this manner is of limited usefulness because the results will be mostly dependent upon the extent of the pattern which is simple to measure rather than its intensity which is not. Also, any index of reticulation needs to capture the continuous spatial variation of the pattern rather than just its intensity between any two particular points.

7.2.3 Polygon form

The combined set of reticulate polygons from all study areas was analysed to investigate general trends in reticulate polygon form. Reticulate polygons show a great variety of plan morphologies as examples from the Naccowlah study area show (Figure 7.14). As stated previously (Section 3.4.5.1), one of the characteristics of the reticulate pattern is the irregular and jagged appearance of both individual channels and reticulate polygons, however such irregularities do not preclude the existence of general trends in polygon form. The morphology statistics for the complete set of reticulate polygons indicating the range of polygon forms found are given in Table 7.4 and the results of correlation analysis between the variables in Table 7.5.

Table 7.4: The variability of reticulate polygon form (n = 1122).

<table>
<thead>
<tr>
<th>Variable</th>
<th>Minimum</th>
<th>Median</th>
<th>Maximum</th>
<th>Mean</th>
<th>S.D.</th>
</tr>
</thead>
<tbody>
<tr>
<td>L_p (m)</td>
<td>63</td>
<td>221</td>
<td>2,854</td>
<td>255</td>
<td>168</td>
</tr>
<tr>
<td>W_p (m)</td>
<td>29</td>
<td>100</td>
<td>466</td>
<td>116</td>
<td>63</td>
</tr>
<tr>
<td>E_p</td>
<td>1.0</td>
<td>2.1</td>
<td>30.0</td>
<td>2.4</td>
<td>1.5</td>
</tr>
<tr>
<td>A_p (m²)</td>
<td>1,445</td>
<td>13,620</td>
<td>226,224</td>
<td>22,393</td>
<td>24,475</td>
</tr>
</tbody>
</table>

Table 7.5: Results of correlation analyses between polygon descriptor variables (n=1122)

(* indicates p not significant at α=0.05).

<table>
<thead>
<tr>
<th>Variables</th>
<th>p</th>
</tr>
</thead>
<tbody>
<tr>
<td>W_p and L_p</td>
<td>0.55</td>
</tr>
<tr>
<td>E_p and L_p</td>
<td>0.63</td>
</tr>
<tr>
<td>A_p and L_p</td>
<td>0.72</td>
</tr>
<tr>
<td>E_p and W_p</td>
<td>-0.20</td>
</tr>
<tr>
<td>A_p and W_p</td>
<td>0.86</td>
</tr>
<tr>
<td>E_p and A_p</td>
<td>0.06*</td>
</tr>
</tbody>
</table>

Polygon form is clearly widely variable. Polygons tend to be elongate forms with a mean elongation of 2.35, however show extremely variable elongations ranging from 1 to 30.
Polygon areas vary over 3 orders of magnitude as do polygon lengths. While some variables, notably $A_p$ and $W_p$, are closely related, the moderate $L_p-W_p$ relationship indicates that polygons do not tend towards a characteristic form (c.f. the extremely strong relationships shown by braid bars and natural streamlined forms described in Section 1.3.4) and the insignificant $E_p-A_p$ relationship implies that no allometric scale-form relationship exists.

7.2.4 Reticulate network morphology: polygon distribution and orientation

7.2.4.1 M2 area

The M2 study area is located away from the floodplain edge in the middle of the study reach (Figure 7.15). The area is approximately 3km by 4km and contains 383 polygons which cover a total area of 7.8km$^2$. The area is fed at its inception by two channels bifurcating from a small meandering stream to the eastern side of the reticulate network (Point F, Figure 7.15). The larger of the two channels persists some distance into the reticulate network, while the smaller channel quickly becomes indistinguishable from the surrounding LILs. Further down the network a smaller bifurcation from the same meandering channel also feeds into the reticulate area (Point G). Flows trend south-south-west from the point of network inception and the reticulate network expands to a maximum width of ~3km before contracting again. Initial impressions may be that the network resembles a rain-fed drainage system feeding into rather than away from the meandering channel, however this is not the case. Not only would this necessitate flow for many kilometres against the known floodplain slope, but if such networks occurred in one area they would be expected to form all over the floodplain surface – which they do not - as floodplain slope, surface material and probably rainfall are approximately constant over the study area.

As well as LIL links providing a throughput of water, the common occurrence of TIL and EL links means that the network also stores a lot of ponded water when at bankfull stage. It is clear from air photographs that TIL and EL links tend to be markedly smaller in size than LIL links because flow velocities and momentum through these links are low as there is no net flow through these links at bankfull stage because they terminate. TIL and EL links can only experience net flow throughput as stage changes and when the reticulate area is inundated above the bankfull level of the channels. In contrast, the continuous nature of LIL links enables flow throughput at all flow stages and the conservation of flow momentum in these channels makes them relatively large. The orientations of the reticulate polygons are clearly not random and would seem to closely reflect the flow patterns that can be inferred from network extent, floodplain slope and the paths of nearby channels (Figure 7.16). The orientations of the
Figure 7.15 Location of M2 study area. The extent of the study area is shown by the dashed line and the major channel inputs to the area with arrows. At point F there are two channel inputs while Channel G, like Channel F an offshoot from the adjacent meandering channel, is smaller and feeds into the area further downstream. The reticulate pattern grades into unchanneled floodplain and reticulate polygons do not extend right to the margins of the network.
Figure 7.16: Reticulate polygons and polygon orientations, M2 area. Arrows on orientation figure denote inferred flow directions. (For location of area see Figure 7.15.)
Figure 7.17 Surface plots of polygon morphology variables, M2 area (continued overleaf) Surfaces interpolated from spatially averaged polygon descriptor variables with ArcView © GIS using the inverse distance weighting (IDW) method and a fixed number of nearest neighbour points.

Mean length (m)
- 208 - 235
- 235 - 262
- 262 - 289
- 289 - 316
- 316 - 342
- 342 - 369
- 369 - 396
- 396 - 423
- 423 - 450

Mean width (m)
- 89 - 104
- 104 - 118
- 118 - 133
- 133 - 148
- 148 - 162
- 162 - 177
- 177 - 192
- 192 - 206
- 206 - 221
Figure 7.17 continued Surface plots of polygon form variables, M2 area.
reticulate polygons does not reflect the orientation of all channels in the network but that of the larger TIL links which are responsible for the transitive function of the networks. As the smaller TIL and EL links act mostly as stores rather than conduits for bankfull flows they would not be expected to be adjusted sensitively to the pattern of flow through the network as a whole which is the primary determinant of network form. As the reticulate area expands, polygon orientations to the west of the large feeder channel radiate away in the direction of the expansion of the network reflecting a distribution of water from the feeder channel in the direction of network expansion. As is clear from Figure 7.16 this orientation in the direction of network expansion is maintained across the network as a whole. Eventually the reticulate network appears to die away towards the southern and western margins of the network as channels become less distinct and the network grades into unchanneled floodplain. EL and TIL links extend into the margins of the unchanneled area. Surface plots of polygon descriptor variables show that polygon area tends to increase away from the point of inception as the reticulate network becomes less distinct and then grades into unchanneled floodplain, and that this increase in area is paralleled by increases in polygon length and width but not elongation ratio (Figure 7.17). The smallest polygons in terms of area, width and length occur in the upper region of the network near the main feeder channel, and the largest occur towards the downstream margins of the network where it is becoming less distinct. The density of polygon development here seems to be causally related to flow characteristics with higher flow volumes near the inception point of the network causing polygon densities to be higher there and reticulate channels and polygons are less densely developed in more distal areas of the network after the input flows have been dispersed over a wide area. The termination of the reticulate network here is inferred to be caused by a decrease in flow volume and power due to network expansion and the spreading of the input flow over an increasingly large area.

7.2.4.2 Barrolka
The Barrolka study area is located in the northern section of the reach in a wide area of floodplain where there is a large expanse of reticulate network of which the study area is the approximate centre (Figure 7.18). The study area is fed by flow from reticulate networks immediately upstream, and contains 98 polygons which cover a total area of 3.2km². The area is not extensive enough to enable meaningful study of the spatial variation in reticulate polygon form, however it was considered suitable for the study of polygon orientations (Figure 7.18). As in the M2 area a clear non-random pattern of polygon orientation is developed which approximates the orientation of the floodplain axis suggesting that orientation is adjusted sensitively to the direction of flow and floodplain slope. As was the case in the M2 area and in
Figure 7.18 Location and polygon orientations of the Barrolka reticulate area. As shown on A, to the north-west of the area the valley is bounded by a large dune field, and an area of unchanneled floodplain is found to the south and east of the reticulate area. The floodplain at this point is approximately 45km wide making it one of the wider areas of floodplain occurring in the study reach. C. shows the polygon orientations (solid lines) from the area outlined in B. Orientations are approximately aligned with the floodplain axis (dashed line) at this point suggesting an alignment with overbank flow direction.
both other study areas described subsequently TIL and EL links here are significantly smaller than LIL links.

7.2.4.3 MVS

The MVS area is located on the eastern side of the floodplain just south of the entrance to Lake Yamma Yamma where flow expands into a small embayment in the valley side (Figures 7.10, 7.19). The area is approximately 2km by 3km, contains 461 polygons which cover a total area of 5.4km$^2$, and is fed from the north diffusely by reticulate channels and from the north-west by larger, braid-form channels. There is a clearly dominant NW-SE orientation of reticulate polygons over much of the area reflecting the pattern of flow into the embayment towards the valley side (Figure 7.19). Equally striking is the deviation from this dominant direction at the south-eastern margin of the network where polygon orientation shifts to parallel the valley side. This does not occur along much of the network margin, and would seem to reflecting a redirection of flow entering the valley side which is an effective barrier to flow. Reticulate polygons extend extremely close to the valley side although like in the M2 area it is the smaller EL and TIL links which directly abut the unchanneled area. Surface plots show that there are clear and similar trends in polygon area, length and elongation ratio over the area while trends in polygon width are somewhat different (Figure 7.20). The largest polygons in area, which are also the longest and most elongate, extend from the west of the study area into the embayment to the east and south east where large elongate polygons are found close to the valley side. The smallest (in terms of area and length) and least elongate polygons generally occur on the margins of the network adjacent to the valley side.

It is difficult to infer a conclusive polygon density-flow power relationship here as it is unclear how flow into the embayment is affected by encountering the valley side that is an effective barrier to flow. The embayment may be constantly transferring flow through it, or flow may enter the cul-de-sac and back up. In the south-east of the area, polygon orientations suggest that flow is redirected along the valley side, whereas polygon orientations in other parts of the area suggest no such trend. The contrast between the tendency observed here for polygon density to increase near network margins and trends in the M2 study area where polygon densities decrease towards network margins suggests that the difference is due to the differing nature of the boundary conditions. Flow in the M2 area can cross the network boundary which marks no more than the limit of network development. Here the margin of the reticulate network is the valley side which is a barrier to flow. The reason for the more dense development of polygons near the valley side is unclear. If the relationship between flow power and polygon density
Figure 7.19: Map of polygon orientations in the MVS reticulate study area (for location see Figure 7.10). The aerial photograph shows the valley side (dashed line) and the extent of the study area (dotted line) with the eastern boundary of the study area coinciding with the valley side. Arrows in (a) indicate general floodplain flow direction. Polygons show highly organised orientations related to inferred flow directions indicated by the grey arrows (b).
Figure 7.20 Surface plots of polygon morphology variables, MVS area (continued next page)
Figure 720 (continued) Surface plots of polygon morphology variables, MVS area

Mean length (m)
- 120 - 167
- 167 - 213
- 213 - 260
- 260 - 307
- 307 - 353
- 353 - 400
- 400 - 447
- 447 - 493
- 493 - 540

Mean width (m)
- 64 - 77
- 77 - 89
- 89 - 102
- 102 - 115
- 115 - 127
- 127 - 140
- 140 - 153
- 153 - 165
- 165 - 178

Scale: 1 km
inferred from the M2 area applies here, then this suggests that flow concentrates and is deeper and faster near the valley side and that flow circulates through the embayment rather than ponding there. The occurrence of low polygon densities near the braid-form channels feeding the area also contrasts with the pattern of variation found in the M2 area where polygon densities near the point of network inception are high. This is probably due to the particular situation described earlier occurring at transitions between braid-form and reticulate patterns (Section 7.1.7.1). As flows weaken when passing from braid-form to reticulate areas, the braid bars become progressively more subdued and dissected by reticulate channels until their influence on network form is no longer apparent. The occurrence of low polygon densities near the inception of the reticulate network here is a result of the persisting influence of the braid-form pattern on reticulate network form (see Figure 7.8 and 7.9). Reticulate polygons take the form of highly elongate braid-form islands here, and it is only further away from the inception of the reticulate pattern that the braid-form islands are completely dissected and their influence on the geometry of the reticulate network desists.

7.2.4.4 Naccowlah

The Naccowlah study area is located on the eastern side of the floodplain near the confluence with the Wilson River (Figure 7.21). The area encompasses 179 polygons which cover a total area of 8.6km². The floodplain here is confined by dunes which control the direction of flood flows through the area and enables reticulate network form at points of imposed flow change to be considered. Flow enters the study area from the north-east and then bifurcates with one flow path heading east and the other heading southwards between Dunes A and C. Another southward trending flow path occurs to the west of Dune A. Like in other areas, TIL and EL links extend up to the edge of the dunes while LIL links and reticulate polygons do not. Flow from the west enters between Dune A and Dune B and combines with the flow stream between Dunes A and C before flowing south between Dunes B and C. Polygon orientations reflect known flow patterns and reticulate polygons are generally aligned with the dunes constricting the flow. Junctions between flow paths are marked by a greater variability in polygon orientation than are individual flow paths reflecting the greater complexity of flow patterns there, and polygons are also generally denser at flow path junctions as Figure 7.22 shows.
Figure 7.21. Location and polygon orientation in the Naccowlah study area. Flow path directions on orientation figure shown by grey arrows. Note the greater variability in polygon orientation at flow-path junctions.
Figure 7.22. Surface plots of polygon descriptor variables, Naccowlah area.

- **Dune Mean length (m)**:
  - 234 - 276
  - 276 - 318
  - 318 - 361
  - 361 - 403
  - 403 - 445
  - 445 - 487
  - 487 - 530
  - 530 - 572
  - 572 - 614

- **Dune Mean width (m)**:
  - 120 - 140
  - 140 - 160
  - 160 - 180
  - 180 - 200
  - 200 - 221
  - 221 - 241
  - 241 - 261
  - 261 - 281
  - 281 - 301

Scale: 1 km
Figure 7.22 (Continued) Surface plots of polygon descriptor variables

**Dune Mean area (m²)**
- 20818 - 31089
- 31089 - 41360
- 41360 - 51631
- 51631 - 61901
- 61901 - 72172
- 72172 - 82443
- 82443 - 92714
- 92714 - 102985
- 102985 - 113256

**Mean elongation ratio**
- 2 - 2.1
- 2.1 - 2.2
- 2.2 - 2.3
- 2.3 - 2.4
- 2.4 - 2.6
- 2.6 - 2.7
- 2.7 - 2.8
- 2.8 - 2.9
- 2.9 - 3

1 km

---

**Legend**
- Yellow: Dune
- Dark yellow: Mean area (m²)
- Dark yellow to dark pink: 2 - 2.1
- Dark pink: 2.1 - 2.2
- Light pink: 2.2 - 2.3
- Light pink to dark pink: 2.3 - 2.4
- Medium pink: 2.4 - 2.6
- Light pink to medium pink: 2.6 - 2.7
- Light pink: 2.7 - 2.8
- Light pink to dark pink: 2.8 - 2.9
- Dark pink: 2.9 - 3
7.2.5 The formation and adjustment of reticulate networks

Reticulate networks are not distributary systems as deltaic systems or alluvial fans are. Some do act to distribute flow, but others also act to transport flow without distribution or concentration and some concentrate flow for example at the flow path junctions of the Naccowlah study area and near the valley side in the MVS study area. Reticulate networks often appear distributary because, as revealed by the results earlier in this chapter, their occurrence requires weaker flows than occur in braid-form areas to allow gilgai expression and so the pattern often begins in locations experiencing divergent flow such as where the floodplain is widening or embayments in the valley side. The cause of formation of individual reticulate channels, the incision of channels into gilgai depressions (Mabbutt, 1967; Rundle, 1977) is well understood, and the formation of dense networks of channels is due to the complex, undulating gilgai floodplain-surface topography promoting channel divergences and recombinations. While reticulate networks probably have the effect of increasing the efficiency of overland flow transport compared to unchanneled floodplain, the persistence, not to say profusion, of terminal channel links in the network would seem to indicate that the networks are well below optimal efficiency in this respect.

The results from the four reticulate study areas enable a number of general conclusions to be drawn. Link size is related to function and LIL links are larger than EL and TIL links because they experience continuous flow throughput at all flow stages. Reticulate network form shows high levels of structure and reticulate polygons are adjusted sensitively to flow direction. Adjustment to overall direction across the network is confined to the transitive elements of the network and not the spacefilling elements. The analysis of surface plots of spatially averaged polygon form reveals that, like orientation, polygon density seems to vary systematically and to be adjusted to flow patterns. Results from the M2 area suggest that areas subject to higher flow powers tend to develop greater polygon densities, although it is not clear that this pattern is general. Reticulate networks and reticulate polygons ultimately form through the growth and enlargement of TIL and EL links which are probably cut by fluctuating flows and flows which completely inundate the channel and surrounding floodplain.

7.3 The braid-form pattern

7.3.1 Introduction

Work reported previously in this study (Chapter 5) has shown that, as Mabbutt (1967) stated, braid-form channels tend to originate from and terminate at anastomosing channels. Also found
was that the angular geometry of junctions between braid-form and anastomosing channels is sufficiently similar to the angular geometry of junctions between anastomosing channels to suggest that the mechanism of anastomosing channel dynamics operative here is the enlargement of braid-form channels into anastomosing channels by a gradual avulsion process. Results from the analysis of braid-form pattern distribution earlier in this chapter indicate that braid-form patterns and anastomosing channels co-occur and that the braid-form patterns develop in areas frequently inundated by relatively powerful flood flows. However, the previously reported analyses are incomplete because they do not address the question of whether the braid-form pattern can be regarded as an equilibrium braided channel pattern in form and behaviour as some previous authors have claimed (Nanson et al., 1986; Rust and Nanson, 1986; Maroulis and Nanson, 1996), or whether there is only a limited morphological resemblance between the two patterns. This question will be addressed here.

7.3.2 The characteristics of braided channels

As stated in Chapter 1, braided rivers are defined by Schumm (1977) as single channel bedload rivers which at low flows are divided by islands composed of either sediment or sediment and relatively permanent vegetation. Schumm explicitly distinguished braided channels from anastomosed channels by stating that anabranches of anastomosing but not braided channels may have their own individual channel patterns. It was previously thought that braided rivers were uniquely steep, high energy channels carrying coarse sediment loads (e.g. Ferguson, 1993). Most are indeed like this, however there are exceptions and braided-rivers may have low slopes like the lower Brahmaputra River which has an average water surface slope of \(0.000065\text{m/m}\) (Coleman, 1969), and may they transport relatively fine sediment like the lower Yellow River and Slims River which have silt dominated sediment loads (Rust, 1978).

Characteristic of typical braided channels is their dynamism and the rapid rate of change that they exhibit (e.g. Leopold and Wolman, 1957; Fahnestock, 1963; Bridge, 1993; Ferguson, 1993; Murray and Paola, 1994). A recent numerical simulation of braided rivers has suggested that in order for braiding to develop the channel must be transporting bedload and competent to easily erode its bed and banks, and that lateral transport of bedload downslope across the channel boundary (Parker, 1976) must occur to enable channel shifting to persist (Murray and Paola, 1994, 1997). Murray and Paola (1994) suggest that “braiding may be the fundamental instability of laterally unconstrained free-surface flow over cohesionless beds” (p.57, emphasis in original), and that other channel patterns may occur due to the imposition of “extra” factors which impede either sediment erosion such as resistant banks, or the redeposition of eroded sediment such as the transport of most sediment in suspension.
7.3.3 Braid-form pattern characteristics and analysis

The braid-form patterns of the Cooper Creek floodplains seem to differ from the characteristics of typical braided rivers listed above in a number of ways. Firstly, the braid-form pattern is not particularly dynamic. A comparison of aerial photographs from 1948 and 1988 shows no discernible change in channel position or planform over the period. It is probable that change here is a slow, gradual process occurring over many events rather than a rapid process occurring only during extremely high magnitude events as the period 1948-1988 included some very large flood events such as the 1974 flood during which rapid change, if it were to occur at all, would have been expected. While meaningful studies of braided channel change may be conducted over timescales of days or hours (e.g. Ferguson et al., 1992; Lane et al., 1996), meaningful studies of braid-form channel change on the Cooper Creek would seem to have to extend over timescales of centuries or possibly millennia. Secondly, it is clear that the channels separating braid-form islands are much smaller in scale than the islands which is inconsistent with typical braided channel morphology (Trevena and Picard, 1978). Braid-form channels range from approximately 5m-35m in width while as we shall see in the next section braid-form island widths range from 90m-620m. Thirdly, the Cooper Creek floodplain is made up of cohesive sediments which Murray and Paola (1994, 1997) suggest is incompatible with braided channel formation because such materials when eroded are generally transported in suspension, not subject to lateral transport and often not redeposited following erosion. However, Nanson et al. (1986), Rust and Nanson (1986) and Maroulis and Nanson (1996) have suggested that it is the observed production of cohesionless aggregates by wetting and drying of the cohesive sediment that enables braiding and that the fact that the bulk of the floodplain surface is made up of cohesive material does not necessarily preclude the channels from being braided in terms of behaviour as well as, perhaps, morphology. Also, while the channels may change only very slowly this does not preclude their dynamics being those typical of braided channels in nature if not pace.

In order to investigate the similarity between the braid-form and braided patterns further the morphology of the braid-form pattern was examined to determine whether it is more than superficially consistent with that of a typical braided pattern. A similarity between the form of braid-bars and braid-form islands, as well as indicating that aspects of the two patterns are morphologically similar would imply that the two patterns were similar in terms of the behavioural characteristics implied by similar morphological relationships. For example, the
attainment of a characteristic roughly lozenge-shaped bar-form would imply that the flow patterns in neighbouring channels were interdependent. Two areas of braid-form patterns were studied from ortho-rectified aerial photographs taken when the floodplain was dry and only the anastomosing channels were active (Figure 7.23). The areas were chosen because they were relatively extensive and because they appeared to be morphologically typical of braid-form areas over the reach. The length $L_B$, maximum width $W_B$, and area $A_B$ of braid-form islands were measured as defined for reticulate polygons in Figure 7.13. The elongation $E_B$ of the bars was calculated as $L_B / W_B$. Only islands bounded entirely by braid-form channels were considered and not those bounded by both braid-form and anastomosing channels. Measurement of the islands as three-dimensional forms was not possible in the field because the large size of the features meant obtaining a suitably large sample size was impractical because of the time it would have taken, and was not possible from remotely sensed imagery because of the extremely low relief of the pattern. As described in Chapter 3, it is unusual for the relief of braid-form islands to exceed 1m. The proximal and distal angles of the braid-form islands were also measured in order to compare the results with the angular geometry of braid-bars reported by Rachocki (1981) and with the findings from the study of junctions between and with anastomosing channels.

7.3.4 Braid-form island morphology

7.3.4.1 Shape and size

Examples of the planform of some randomly selected braid-form islands is shown in Figure 7.24 and it is clear from this and Table 7.6 that there is a large degree of variability in island form.

<table>
<thead>
<tr>
<th></th>
<th>Min</th>
<th>Median</th>
<th>Max</th>
<th>Mean</th>
<th>S.D.</th>
</tr>
</thead>
<tbody>
<tr>
<td>$L_B$ (m)</td>
<td>405</td>
<td>1011</td>
<td>3491</td>
<td>1193</td>
<td>615</td>
</tr>
<tr>
<td>$W_B$ (m)</td>
<td>88</td>
<td>206</td>
<td>624</td>
<td>242</td>
<td>103</td>
</tr>
<tr>
<td>$E_B$</td>
<td>2.4</td>
<td>3.7</td>
<td>13.4</td>
<td>5.1</td>
<td>2.1</td>
</tr>
<tr>
<td>$A_B$ (m$^2$)</td>
<td>28,190</td>
<td>175,077</td>
<td>955,830</td>
<td>224,819</td>
<td>196,163</td>
</tr>
</tbody>
</table>

Braid-form islands are usually, although not always, orientated with their long axes aligned approximately with floodplain slope. Relationships between morphological variables are given in Table 7.7 and a scatterplot of braid-form island length and width is shown in Figure 7.25.
Figure 7.23: (a) Location of the braid-form pattern study areas in the north of the study reach, and (b) an aerial photograph showing the pattern developed over a part of the smaller of the two study areas shown in (a).
Figure 7.24: Examples of braid-form island planform

Figure 7.25: Scatterplot of braid-form island length and width dimensions.
Table 7.7: Results of braid-form island morphology analysis (n=62). * Denotes not significant at $\alpha=0.05$.

<table>
<thead>
<tr>
<th>Predictor variable</th>
<th>Response variable</th>
<th>Equation</th>
<th>$r^2$</th>
</tr>
</thead>
<tbody>
<tr>
<td>$L_B(m)$</td>
<td>$W_B(m)$</td>
<td>$W_B = 131.95 + 0.09L_B$</td>
<td>0.30</td>
</tr>
<tr>
<td>$L_A(m)$</td>
<td>$E_a(m)$</td>
<td>$E_a = 2.475 + 0.002L_A$</td>
<td>0.42</td>
</tr>
<tr>
<td>$L_A(m)$</td>
<td>$A_a(m)$</td>
<td>$A_a = -7100 + 248.04L_A$</td>
<td>0.60</td>
</tr>
<tr>
<td>$W_B(m)$</td>
<td>$A_a(m)$</td>
<td>$A_a = -8360 + 1277W_B$</td>
<td>0.45</td>
</tr>
<tr>
<td>$W_B(m)$</td>
<td>$E_a(m)$</td>
<td>$E_a = 6.2 - 0.004W_B$</td>
<td>0.05*</td>
</tr>
<tr>
<td>$A_B(m)$</td>
<td>$E_a(m)$</td>
<td>$E_a = 4.4 + 3.21*10^{-5}A_B$</td>
<td>0.09</td>
</tr>
</tbody>
</table>

Braid-form islands are generally large, elongate features with a mean length of over 1km and a maximum elongation of 13.4. The mean elongation of 5.1 is much higher than the reported elongations of braid bars of 2.4 (Rachocki, 1981) and 3.9 (Rust, 1972). Relationships between variables were generally moderate or poor. The relationship between the length and width of the islands was very poor ($r^2=0.30$) in contrast to relationships between natural streamlined forms (see Section 1.3.4) indicating that the islands do not tend to adopt a characteristic shape, and the weakness of the relationship between island elongation and area indicates that there is no allometric relationship between island scale and form. A comparison of Figure 7.25 with Figures 1.4 and 1.5 indicates that the morphological variability of braid-form islands is much more similar to that of anastomosing islands than that of braid-bars.

7.3.4.2 Island angular geometry

The results of island proximal ($\hat{P}$) and distal ($\hat{D}$) angle measurement are given in Table 7.8. As outlined in Chapter 5, island distal angle is akin to the confluence angle of the channels bounding the bar or island while distal angle is analogous to bifurcation angle.

<table>
<thead>
<tr>
<th>P (°)</th>
<th>Min.</th>
<th>Median</th>
<th>Max.</th>
<th>Mean</th>
<th>S.D.</th>
</tr>
</thead>
<tbody>
<tr>
<td>$\hat{P}$</td>
<td>19</td>
<td>48.5</td>
<td>95</td>
<td>48.8</td>
<td>15.4</td>
</tr>
<tr>
<td>$\hat{D}$</td>
<td>17</td>
<td>38</td>
<td>85</td>
<td>40.5</td>
<td>16.1</td>
</tr>
</tbody>
</table>

Mean proximal and distal angles are relatively similar and the difference between the means of the two groups is not significant (Tukey-Kramer HSD test, $\alpha=0.05$). Mean distal angle is 10° lower than the mean angle of 50° found from braided channels by Rachocki (1981) and junction angles more variable with a range of 68° compared to approximately 60°. Notably, a quarter of braid-form islands had distal angles below 30° whereas Rachocki found no braid-bars with distal angles below 30°. A comparison of braid-form proximal and distal angles with
the form of junctions with (BF-AC) and between (AC-AC) anastomosing channels (Tables 6.3 and 6.4) shows that mean braid-form island angles are markedly more acute (by at least 10 degrees) than other junction angles. This suggests that if a braid-form channel enlarges to become an anastomosing channel as results in Chapter 5 suggest, then existing junction angles of the channel with braid-form channels become systematically larger, possibly due to the formation of levees along the anastomosing channel.

7.3.5 Summary of braided channel morphology results

7.3.5.1 The nature of braid-form channels

Morphological investigations have shown that, although it bears a superficial resemblance to a typical braided channel pattern, the braid-form pattern is dissimilar to the braided pattern in a number of ways including the rate of pattern change, the shape of the bars/islands and the relative sizes of the bars/islands and channels. The variability of braid-form island morphology is similar in nature to that shown by anastomosing channels suggesting that individual braid-form channels operate quite independently and are not co-adjusted as braided channels tend to be. The channels are, as Mabbutt (1967) stated, floodways which are formed by overbank flows taking a more direct route down the floodplain than the more sinuous and wandering anastomosing channels, and the alignment of braid-form islands with floodplain slope simply reflects the general alignment of the braid-form channels. One effect of braid-form channel formation will be to increase the efficiency of the floodplain to transport overland flow which Winker (1978) noted was also an effect of floodplain-surface channel formation on the channeled floodplains of the Little River Drainage Basin in Texas.

Floodplain behaviour and evolution may be very complex and unpredictable, and Lewin (1978) notes that floodplain swales may become foci for the slow, overbank deposition of fine sediments or may function as scour channels during flood flows. Floodplain depressions may experience elevated rates of accretion because sediment settles out from flow ponded there (Howard, 1992; Asselman and Middelkoop, 1995), however if depressions are continuous then velocities and depths of flow there will tend to be greater than on the surrounding areas of floodplain and may cause deposition to be retarded or may even cause erosion (e.g. Gretener and Stromquist, 1987). The prevalence of braid-form channels indicates that floodplain depressions here do not tend to be filled and smoothed over, and in combination with results from other chapters implies that floodplain topography tends to be magnified over time as some braid-form channels enlarge into anastomosing channels. This may occur either because channels are incised into the floodplain surface or because of lower - but still positive - rates of
channel accretion over the channels than over the surrounding floodplain. It is likely braid-form channels do form at least partly by incision which is made possible by cracking of the floodplain surface and aggregate formation which a number of authors have suggested is important in channel dynamic processes on the Cooper Creek (e.g. Nanson et al., 1986; Rust and Nanson, 1989; Maroulis, 1992; Maroulis and Nanson, 1996; Gibling et al., 1998).

7.4 Floodplain accretion rate data

Brief consideration was given to long term floodplain evolution and accretion rates in order to examine whether floodplain accretion rates differed between floodplain environments and whether a model of floodplain dynamics could be formulated. Despite extensive dating of the underlying sand sheet and study of the influence of climate change on river characteristics (Rust and Nanson, 1986; Nanson et al., 1986; Nanson et al., 1988; Nanson and Tooth, 1999; Maroulis, 2000), the evolution of the contemporary floodplain has only been considered briefly by Maroulis (2000). Maroulis (2000) studied two transects across the floodplain, one just north of Naccowlah Waterhole and the other north of Meringhina Waterhole. Surveys showed that the floodplain across both transects was flat, and thermoluminescence dates indicated that the western side of the floodplain on both transects was accreting more rapidly than the eastern side. Maroulis (2000) considered two explanations of the relief and accretion rates, the first being that

the alluvial surface in the west was lower in the past, has accreted more rapidly than the east and by coincidence is now exactly the same elevation right across the floodplain (an extremely unlikely coincidence) (p.150)

and the second being that

the western side of the valley is subsiding more rapidly than the east and that sedimentation is maintaining an equilibrium rate of accretion across an unevenly subsiding basin. (p.150)

As the quotes above suggest, Maroulis (2000) favours the second explanation and a general westward shift to the Cooper floodplain citing as evidence for this the encroachment of the Cooper floodplain into Lake Yamma Yamma. However, Maroulis’ explanation is problematic in that it is not possible to generalise from the Shire and Durham Road Transects to the entire Cooper floodplain. Surveyed transects reported in this chapter (Figures 7.2-7.4) have shown
that the floodplain is generally not flat and that cross-floodplain differences in elevation are in fact quite common. Secondly, not enough is known about the age or origin of Lake Yamma Yamma to support the contention that the existence of the Lake is evidence of a continuing western shift to the Cooper floodplain.

In order to augment the available data a number of samples for dating were taken from a pipeline trench running approximately E-W across the floodplain between Meringhina Waterhole and Lake Yamma Yamma and including both reticulate and braid-form areas (Figure 7.26). Samples were taken from the mud unit at relatively shallow depths of between 1.2 and 1.5m in order to provide information on contemporary rates of floodplain accretion. All samples were taken away from levees and channels, and were dated using TL with a comparative OSL date obtained for one sample (SF12). The floodplain here is significantly higher in the east than the west, similar to the relief of the floodplain approximately 80km to the north shown in Figure 7.2. While the highest accretion rate (0.63m/ka) did occur in the braid-form area towards the western side of the floodplain and the lowest in the reticulate area towards the eastern side, there was no consistent trend in accretion rates either across the floodplain, with elevation, or with floodplain-surface pattern type. Accretion rates in the braid-form area (2 dates) were found to be 0.14m/ka and 0.63m/ka, while those in the reticulate area (3 dates) varied between 0.02m/ka and 0.29m/ka. 3 dates were also obtained from a profile augered in the heavily reticulated Wilson’s Swamp area near the confluence of the Cooper Creek and Wilson River (Figure 7.27). The upper date in the profile seemed to have been affected by bioturbation mixing contemporary sediments into the older underlying sediments and the accretion rate of the profile is uneven. The sample from the base of the profile was dated by both OSL and TL methods which produced comparable results (TL age - 6.1 +/- 0.5ka, OSL age – 7.3 +/- 0.5ka) and an accretion rate of 0.28m/ka was obtained for the profile as whole. This is similar to the accretion rates obtained from the reticulate and braid-form areas of the transect described above and also that of the levee deposit described in Chapter 6 of 0.25m/ka.

Although the data are not sufficient to enable a model of floodplain evolution to be formulated, a number of trends are apparent. Firstly, floodplain accretion rates between reticulate and braid-form areas do not systematically differ and appear to overlap greatly. There also appears to be no tendency for accretion rates to decrease away from the main, anastomosing channels, and accretion rates shown by the reticulate areas are broadly similar to those shown by the dated levee deposit and those in braid-form areas closer to the channels. This is surprising given the abundant evidence of a general decrease in floodplain accretion rates with increasing
Figure 7.26: The floodplain accretion transect showing the position of samples and variations in elevation and floodplain-surface channel pattern occurrence over the area. Numbers in the cross section followed by the sample code are the calculated accretion rates in m/ka. Sample depths are all 1.5m except for SF 14 which is from 1.2m. The figure shows that the highest accretion rate is found in the braid-form area, but that accretion rates in the reticulate area can exceed the minimum occurring in the braid-form area. There is insufficient data to enable conclusions about general accretion patterns to be drawn.
<table>
<thead>
<tr>
<th>Surface</th>
<th>Depth (m)</th>
<th>Age (ka)</th>
</tr>
</thead>
<tbody>
<tr>
<td>SF 9</td>
<td>0.90</td>
<td>0.3 +/- 0.05</td>
</tr>
<tr>
<td>SF 10</td>
<td>1.35</td>
<td>1.2 +/- 0.1</td>
</tr>
<tr>
<td>SF 11</td>
<td>1.87</td>
<td>6.7 (ave. TL and OSL)</td>
</tr>
</tbody>
</table>

Figure 7.27: Floodplain accretion dated profile from the Wilson's Swamp reticulate area. Profile location shown by 'P'. The top date SF 9 returned an unusual TL signature and must be regarded as suspect. The accretion rate over the profile appears to be uneven and for the profile as a whole is 0.28 m/ka.
distance from the main, low-water channels (e.g. Allen, 1965; Kesel et al., 1974; Gretner and Stromquist, 1987; Asselman and Middelkoop, 1995; Simm, 1995), and suggests that sediment transport across the floodplain is very efficient. However much more evidence is needed if the evolution of the contemporary floodplain is to be well understood.
The previous chapters have described aspects of an extremely complex channel system. Following a brief summary of the results and findings of each chapter a model of anastomosing channel behaviour here will be presented. The relevance of the equilibrium concept to the operation of the channel system will be considered, some philosophical considerations arising during the study briefly described, and suggestions for future research outlined.

8.1 Summary of results

The literature review in Chapter 1 highlighted the morphological, dynamic and causal diversity of anastomosing rivers. This diversity has not been sufficiently recognised in some more recent definitions of anastomosing rivers which, in attempting to be precise, have instead been too restrictive with the effect of mischaracterising anastomosing rivers and excluding from the anastomosing class types of river that are best described as anastomosing. Not all anastomosing rivers are rapidly accreting, not all produce new anabranches by crevasse avulsions, and not all form new anabranches quickly or rapidly. The Cooper Creek is a good example of this. Where channel change is very slow and gradual the term ‘avulsion’ is inappropriate and misleading, and the term ‘gradual avulsion’ was introduced here to avoid the connotations of violence and suddenness implied by avulsion alone. In order to recognise the causal diversity of anastomosis a novel distinction was made here between autogenic and exogenic anastomosis in order to separate anastomosis resulting from channel blockages from that caused by the flow and sediment dynamics of the channel only.

The good agreement between the results of the TL and OSL dating methods reported in Chapter 2 confirms the reliability of both TL dating in this area and of the current, TL-based Quaternary history of the region.

The study of anastomosing channel planform in Chapter 4 provides a number of insights into the dynamics of the system. Slope-discharge plots and stream power considerations suggest that anabranches should be laterally stable, however all tend to develop meander geometries adjusted to their bankfull widths implying that the meandering patterns are contemporary features, and the more sinuous channels show abundant evidence of lateral dynamism in cutoff formation. Smaller channels can be much more sinuous than larger channels and seem to be more competent to meander despite having lower flow energies. This tendency for smaller
channels to exhibit a history of active meandering is even more surprising because, as they become sinuous, they reduce progressively their gradient. Clearly, the ability of a channel to migrate laterally is a function of the erosional energy relative to the boundary’s resistance to erosion. In such low energy systems channel migration is determined principally by the frequent pedogenic disruption of the channel boundary caused by wetting and drying cycles in an environment of swelling clays. These self-mulching properties have a much greater effect in smaller than larger channels, firstly because this cracking occurs most vigourously in the upper part of the floodplain stratigraphy into which these smaller channels are cut, and secondly because the fractured soil layer is larger relative to channel size. The disruption of the channel boundary seems to not only control the rate of channel migration but more fundamentally enable the occurrence of any channel migration at all. The lack of channel abandonment and underfit stream development indicates that the model of anastomosis development due to channel capacity reduction proposed by Schumm et al. (1996) does not drive anastomosis formation.

Methodologically, the fractal method was found not to be a useful tool with which to examine anabranch planform. Fundamentally, this was because it is only useful for describing fractal objects which the anabranches are not. As well as showing a characteristic scale of meandering, deviations of anabranch planform from this characteristic scale are also systematic with measured meander properties being positively skewed and approximately log-normally distributed. This means that the median rather than the mean is a more useful measure of both the central tendency of the distributions and the characteristic scale of meandering.

Results in Chapter 5 show that while confluence angles in anastomosing channels are not always larger than those in other channel network types, they are generally larger to a sufficient degree that the confluence angle distributions of anastomosing rivers may be considered a characteristic morphological signature of the anastomosing pattern. Anastomosing channel (AC-AC) confluence and bifurcation angles differ significantly as do the confluence and bifurcation angles of braid-form floodplain surface and anastomosing channels (BF-AC) junctions. AC-AC and BF-AC bifurcation angles are significantly larger than AC-AC and BF-AC confluence angles which were statistically indistinguishable. The similarity between the angular geometry of confluence types was taken to indicate that anastomosing channels here develop from the enlargement of braid-form floodplain-surface channels. The difference between anastomosing channel confluence and bifurcation angles is thought to be due both to the tendency for bifurcations to form systematically on the outside of bends where junction angles are likely to be larger while the position of confluence junctions is more random, and
the fact that bifurcations seem to increase systematically their junction angles as braid-form channels enlarge into anastomosing channels while in similar circumstances the angular geometry of confluences does not change. This is probably due to more flow energy being directed against channel banks at bifurcations than at confluences. Both the increase in bifurcation angle and the frequent occurrence of 'scorpion' bifurcation junctions at which the channels double back on themselves (which appear to be rather unique to this system) imply that junction form does not adjust to maximise efficiency here, at least as efficiency is defined in current, uniform flow models of junction geometry. The study of confluence morphology showed that a range of junction morphologies are found, that discordant (in terms of bed elevation) junctions are dominant, that scour holes are absent from all junctions observed, and that fine-grained tributary fans are formed.

Chapter 6 describes the morphology of the channel-floodplain boundary of anastomosing channels. Levees generally occur along the boundary of anastomosing channels of constant size except where scroll bar topography is formed or the levees are breached by floodplain-surface channels. Levee size and form seem to be partly determined by the size of the channel proximal to the levee and wider, more subdued levees occur along larger channels. Levee texture variations are complex and very different to those typically found. The levees here are not uniformly intermediate in texture between channel and floodplain deposits, and sections of all levees are systematically finer than both. In contrast to coarser levees like those studied by Cazanacli and Smith (1998), the predominantly fine textured levees here seem to show morphological adjustments to overbank flow patterns formed when the floodplain is fully inundated, suggesting that the transfer of sediment from the channel by diffusion rather than convection is the most important process for levee formation. The study of channels expanding in size without tributary inputs showed that, unusually, what appear to be levees are formed where flow, and so sediment, from the floodplain-surface is converging on the channels. Termed 'eevels' here, these depositional landforms form by the opposite process to normal levees with sediment from the floodplain being deposited near the channels, and this is confirmed by observed differences in the textural variations of levees and eevels. Where the channel rapidly contracts in size the strong outflow onto the floodplain suppresses levee formation and transfers large amounts of coarse sediment from the channel, causing the floodplain there to be coarse textured. Textural variations reported in Chapter 6 suggest that the system is near the threshold for sand transport over the floodplain surface which occurs only where floodplain flows are strong.

Chapter 7 highlights a number of controls on the distribution, formation and morphology of the floodplain-surface channel patterns. The distribution of floodplain-surface patterns indicates
that their occurrence is sensitively related to interactions between fluvial and pedogenic processes, and the patterns occur along a continuum of overbank flow energy and frequency with braid-form patterns forming where the power and frequency of overbank flows are relatively high, unchanneled areas where inundating flows are very weak and infrequent, and reticulate areas occurring in areas subject to intermediate frequencies and powers of overbank flows. The formation of floodplain-surface channels has the effect of increasing the efficiency of overbank flow transmission during flood events by concentrating flow and is in itself a product of flow concentration. In reticulate areas it is probable that the concentration of flow by gilgai not only controls channel location but is also necessary for channel formation, and in the absence of gilgai these areas would probably be unchanneled. The morphology of the reticulate pattern was found to be adjusted to the power, frequency and direction of floodplain flows. The investigation of the braid-form pattern on the floodplain concluded that it is not particularly similar to a typical braided river pattern in either form or process, and that the morphological resemblance of the two patterns is somewhat superficial when the resemblance is quantified. The braid-form channels are best thought of as floodways as indeed they were described as by Mabbutt (1967). It is likely that the self-mulching vertisols over the floodplain and the associated formation of pedogenic aggregates is important in enabling floodplain-surface channel formation, as well as anastomosing channel migration, by reducing the resistance of the surface material to erosion and entrainment.

8.2 Anastomosing channel formation and dynamics

Although a number of theories have stated that sediment transport considerations are of primary importance in promoting anastomosis development (e.g. Baker, 1986; Schumm et al., 1996; Makaske, 1998; Nanson and Huang, 1999) this does not seem to be the case here. The transport of sand overbank indicated by levee and channel boundary textures adjacent to steeper sided channels shows that at high discharges sand is transported in suspension. As sand is the coarsest size fraction transported by the system this suggests that at high flows much, if not all, sediment is transported in suspension. Combined with the fact that sand is a relatively small component of total sediment load, this suggest that anastomosis here is not a product of the channel system adjusting to maximise bedload transport as suggested by the theory of Nanson and Huang (1999). Anastomosis would also seem not to occur in order to store coarse bedload sediment that cannot be transported through the reach, although this would seem to be an important function of anabranches in the anastomosing reaches of the Columbia River in Canada (Makaske, 1998), and as was noted in Section 3.4.3.2 there is no significant coarse sediment aggradation in any anabranch. Measured suspended sediment loads reported in
Chapter 3 vary between 132mg/l and 427mg/l and are sufficiently low to imply that anastomosis development here is not a response to extremely high suspended sediment loads as Baker (1986) has suggested is the case for the anastomosed reaches of the Yangtze River.

Nanson et al. (1986) state that the anastomosing channels here form in response to moderate, within-bank flows, and the anastomosing channels do form a sufficiently continuous system to enable sub-bankfull flows to be transported across the study reach by anastomosing channels alone. However, as described in Section 3.4.3.1 the anastomosing network is not fully integrated and large sections of the network are active only at overbank flow stages implying that the network as a whole is not adjusted to the transport of bankfull and sub-bankfull flows. The adjustment of the planform of individual anabranches to bankfull dimensions indicates that the bankfull condition is an important morphological parameter for these channels, while the disconnectedness of the anastomosing channel system means that the bankfull of one channel does not approximate bankfull for the suite of anastomosing channels as would be the case if the network was fully integrated. The analysis of junction angles in Chapter 5 indicates that anastomosing channels form from the enlargement of braid-form floodplain-surface channels and so are a product of overbank flows and gradual avulsion events. Diffuse, shallow overbank flow is subject to higher flow resistances than concentrated channel flow and so the formation of anastomosing channels has the effect of increasing the flow transport efficiency of the channel system both by increasing the stage which flow must exceed before going overbank when the anabranches are integrated with the bankfull and sub-bankfull transporting network, and more generally by increasing the efficiency of flow transport at overbank stages. This has been noted as a function of the floodplain channels of the Little River Drainage Basin floodplains in Texas (Winker, 1978) and of the braid-form channels of the Cooper floodplain.

As stated earlier, the loss of anabranch competence and the reduction of anabranch capacity due to increasing sinuosity does not seem to drive gradual avulsion here as is evidenced by the lack of underfit anabranches that would be expected if this was the case. Channels are generally not perched above the surrounding floodplain and so this would not seem to be an important cause of avulsion and anastomosis as it is in the crevasse avulsion model applicable to channels in situations subject to more rapid accretion. More important are probably large-scale floodplain accretion variations which control the distribution of the floodwaters responsible for anastomosing channel formation. Riley (1975) notes that anabranch formation can occur irrespective of whether the existing channels are at some kind of bankfull flow and sediment transport equilibrium, and this seems to be the case here with new anabranches forming because of gradual scour at high flood stages, although their formation due to reduced but still
positive rates of accretion over the channels cannot be excluded. There appears to be no incision threshold and no knickpoints are found on the floodplain. Changes in anabranch size and prominence here are more likely consequences rather than causes of avulsion events. There is no clear morphological distinction between braid-form channels and anastomosing anabranches and, as one would expect if one channel type developed from the other, the morphologies of the two types grade into one another. Levees probably also play an important role in braid-form and so anastomosing channel formation because levee breaches are the points at which floodplain-surface channels and so eventually new anabranches form. Floodwaters invading the floodplain at levee breaches will be relatively deeper and faster than if levees were absent and only bank spilling occurred, and this will promote channel development. The apparent importance of levees to the development of anastomosis is common with other models of autogenic anastomosis development, such as the crevasse avulsion model, although obviously for different reasons. The presence of levees is probably also important to the development of high angled bifurcations here by causing convective overbank flow to be directed perpendicularly away from the channel rather than with the general slope of the floodplain. It is arguable that the importance of levees as signatures and determinants of channel dynamics has not been sufficiently recognised and that they are, as Brierley et al. (1997) state, neglected features of the fluvial landscape.

8.3 Equilibrium and adjustment of the Cooper Creek channel system

A recent paper by Tooth and Nanson (2000a) has stated that the Cooper Creek operates as a system in dynamic equilibrium and that the entire network of anastomosing and floodplain-surface channels appears to be intricately interconnected for the purpose of collecting, transporting and dispersing water and sediment across the floodplain surface. Following a discussion and description of the equilibrium concept, Tooth and Nanson's argument will be outlined and the equilibrium, or otherwise, behaviour of the Cooper Creek channel system considered.

8.3.1 Geomorphic equilibrium

In a recent, extensive review of equilibrium Thorn and Welford (1994) stated that the concept should be central to geomorphic research but that the conflict between different definitions of equilibrium proposed by geomorphologists has created confusion and reduced its power and utility. Thorn and Welford favour Ahnert's (1994) definition which defines equilibrium as
being most fundamentally an equality of process rates and distinguishes between two different types of equilibrium; ‘dynamic’ equilibrium which refers to an adjustment of process rates only, and ‘steady state’ equilibrium which implies an additional constancy of form. Ahnert (1994) distinguishes between two types of system not at equilibrium; nonequilibrium systems which tend away from rather than towards equilibrium, and disequilibrium systems which tend towards but have not yet reached dynamic or steady state equilibrium. Ahnert regards the tendency to equilibrium as being universal provided that sufficiently strong negative feedbacks capable of compensating each other exist between processes, and states that nonequilibrium represents either a short term phase of development following a major change of input to a system which in time is replaced again by the tendency towards dynamic equilibrium, or when persistent that nonequilibrium systems are local subsystems of larger systems possessing overall equilibrium tendencies. One requirement for the meaningful use of equilibrium noted by Thorn and Welford is the need to specify the system of interest prior to invoking the concept. This inevitably involves an element of artificiality and abstraction although Thorn and Welford (1994, p.691) state that ‘the step back from reality involved in the inevitable simplification associated with systems and equilibria will be more than offset by the clarity gained.’ Thorn and Welford state that equilibrium is a quantified concept and that it must be expressed mathematically with the use of mass terms being more appropriate for geomorphologists because they are simpler to quantify and more easily relatable to form changes as the alternative energy terms although not as fundamental. However, Ahnert (1994) notes that even when quantified an element of subjectivity is inherent in the recognition of equilibrium as the concept is necessarily timebound and a suitable timescale over which to consider process and form changes must be selected.

Tooth and Nanson (2000a) consider the proposal of a fully quantified equilibrium based on measurements of process rates or energy distribution to be unrealistic and note that geomorphologists generally use the term in a less mathematised way. Process rates are usually inferred from observations of form rather than directly measured, and to account for this Tooth and Nanson propose a more pragmatic definition of equilibrium which, expressed to apply to river channels, regards equilibrium as ‘a condition in which, over a period of years, the relatively stable physical characteristics of a stream indicate that the controlling variables are balanced to provide just those conditions required for the onward transport of water and sediment from upstream’ (p.3). Tooth and Nanson acknowledge that while equilibrium is a much used concept in fluvial geomorphology, there has been little explicit definition of how the equilibrium state can be recognised with the exception of Richards (1982). Richards identified four criteria as being indicative of equilibrium in river channels; constancy of
channel form, continuity of sediment transport, efficiency of channel form, and the existence of strong relationships between aspects of channel form and process. Richards stresses that these criteria are individually ambiguous and that they need to be applied together so that each is qualified by the others. Continuity of sediment transport links continuity of form with continuity of process, and in defining the efficiency criterion Richards recognises that elements of the channel system may have conflicting needs with regard to efficiency maximisation. The most efficient channel with respect to water transport may not be the most efficient with respect to sediment transport and *vice versa*. Regarding the existence of strong relationships between aspects of channel form, Richards notes that relationships are often as strong in aggrading or degrading streams as they are in streams at equilibrium. Richards doesn't specify that there is a minimum number or particular set of criteria which need to be satisfied in order for a system to be considered to be at equilibrium. The criteria are best considered as approaches for recognising equilibrium rather than a checklist, and are similar in a number of ways to those used to recognise rivers in regime or at grade (e.g. Stevens et al., 1975; Mackin, 1948). River channel equilibrium is usually assessed over relatively short timescales of a couple of years (e.g. Stevens et al., 1975).

The utility of the equilibrium concept has been much debated recently with disputes arising over the most useful and correct definition of the term and the relevance of a simple equilibrium concept in a changing world in which complexity and nonlinearity are prominent (e.g. Phillips, 1992a, 1992b; Thorn and Welford, 1994; Kennedy, 1994; Phillips and Gomez, 1994). Perhaps the most significant challenge to the equilibrium concept is the recognition that the tendency towards a simple equilibrium state is not universal and may even be unusual. Kennedy (1994) states that our current worldview is one in which change and dynamism dominate and that there are a limited number of processes which can be shown to produce definable equilibrium states. Work on nonlinear dynamical systems has shown that even systems which are very simple in terms of the number and connectedness of system elements do not necessarily tend towards stable equilibrium states (e.g. May, 1976). Phillips (1992a) and Phillips and Gomez (1994) note a trend away from work focused on the detection of a single, stable equilibrium state towards the investigation of systems away from equilibrium, disequilibrium and nonequilibrium forms, and the recognition of complex, multiple equilibrium states. As an illustration of this, a recent book on drainage basin form and the modeling of drainage basin development fails to give simple equilibrium states a single mention but frequently refers to more complex modes and models of behaviour such as self-organised criticality, punctuated equilibrium and fitness landscapes (Rodriguez-Iturbe and Rinaldo, 1997).
8.3.2 The equilibrium of dryland rivers and the Cooper Creek

Dryland rivers are generally regarded as being nonequilibrium systems which are highly responsive to individual flow events, which recover slowly following the occurrence of a large flood, and which operate discontinuously in both space and time (e.g. Schick, 1974; Wolman and Gerson, 1978; Graf, 1988; Cooke et al., 1993). Tooth and Nanson (2000a) acknowledge that this is often the case, but argue that the diversity of dryland rivers has not been sufficiently acknowledged and that they can be either equilibrium or nonequilibrium systems depending on the factors which typically determine channel morphology and behaviour, like catchment size, flow regime, slope, boundary materials and unit stream power. To illustrate the attainment of equilibrium by dryland rivers Tooth and Nanson use examples of the middle reaches of rivers from the Northern Plains in the north-west of the Lake Eyre basin and the middle reaches of the Channel Country rivers including the Cooper Creek in the study reach. Tooth and Nanson (2000a) based their identification of the equilibrium behaviour of the Cooper Creek on Richards' four criteria of which they state that three are satisfied; constancy of channel form, the existence of stable relationships between system properties, and that the system is efficient. Tooth and Nanson note that the system does not satisfy the fourth criterion of sediment transport continuity as it is aggrading and only approximately half of the sediment transported into the upstream end of the study reach leaves the downstream end. However, they state that although significant for the reach as a whole, this discontinuity is essentially insignificant for individual anabranches and sub-reaches where accretion rates are extremely low.

The stability of the channels over the period of observation has already been noted and was considered by Tooth and Nanson to indicate that there was a stable relationship between form and process in the system. All aspects of the system studied provided evidence that the system is dynamic which together with the observed stability is evidence that the Cooper is not a typical nonequilibrium system. No aspects of the system appear to respond to individual events, however large. Rather, aspects appear adjusted to the general flow regime. However, stability in and of itself is evidence of nothing except a slow rate of change or perhaps no change at all, which in this case is due to a combination of low gradients and highly resistive floodplain sediments. As an example of how constancy of form is not necessarily evidence of equilibrium, Ahnert (1994) states that nonequilibrium systems are found locally wherever erosion is weathering limited like on cliff faces. Although cliff faces may exhibit an approximate constancy of form over long periods of time this does not indicate that an equilibrium state has been attained as there is an imbalance between the rate of removal of material from the cliff face (a small, positive number) and the rate of supply of material to the
face (effectively zero).

As evidence of stable relationships between aspects of channel morphology Tooth and Nanson cite the linear relationship between anastomosing channel sinuosity and size found by Beaman (1995, unpublished). However, results in this study have shown that anastomosing channel sinuosity variations are in fact more complex than Beaman described and that a stable, simple relationship between anabranch sinuosity and channel size does not exist. Anabranch sinuosity is a complex function of the degree of fracturing of the channel boundary and age. Sinuosity does not seem to approach an equilibrium as a number of models of meander planform and sinuosity development suggest (see Section 4.3.1) but, given constancy of anabranch size, rather seems to consistently increase until abandonment. Other elements of the channel system are linked by strong relationships however. Anastomosing channel width and depth are closely related (see Figure 3.10) as are channel meander properties and channel size (Chapter 5).

The morphology of channel junctions reported in Chapter 5 suggests that not all aspects of the channel system are adjusted to maximise efficiency. Confluence angles here are generally larger than those in other systems and appear not to change as junctions evolve. This may be because sufficiently strong feedbacks between flow patterns and boundary erosion are not present here because of the relatively weak flow and highly resistive boundaries. Bifurcation junctions seem to be and to become systematically inefficient as far as this can be inferred from their angular geometry which is inconsistent with their being in equilibrium. Other aspects of the channel system do seem to be efficient and the low width/depth ratios of anastomosing channels suggest that they are efficient conduits for water flow rather than bedload transport. Given the relative absence of bedload, this would appear to be highly appropriate. The efficiency of the anastomosing network as a whole is more difficult to evaluate, and it seems to be jointly adjusted to two different flow stages. The network as a whole is partly disconnected and anastomosing channels seem to form principally in response to gradual scour during overbank flow. The formation of the anastomosing and braid-form floodplain surface channels has the effect of increasing the efficiency of overbank flow transmission. Due to the partial disconnection of the network, lower flows are restricted to a subset of the total network which forms a continuous path for low flows across the study reach. This effectively concentrates lower flows reduces transmission losses, and increases flow efficiency at low flow stages.

Overall, it is clear from the above discussion that, aside from channel junctions, many aspects of the Cooper Creek seem to be in a state of dynamic equilibrium. Those that are seem to be adjusted to the general flow regime, rather than to individual events as is typical of many desert
streams, which is an important observation for system behaviour generally. The aspects of the system which do not appear to be in equilibrium do not show typical, unstable nonequilibrium behaviour but rather, at least in the case of channel bifurcations, seem to exhibit a behaviour not described by the equilibrium/nonequilibrium terminology.

8.4 Philosophical considerations

The study utilised a number of methods among which were the examination of remotely sensed data, the dating of floodplain deposits and the direct measurement of channel and floodplain form. The slowness of change of the Cooper Creek channels and floodplain and the consequent impossibility of the direct observation of dynamics necessitated a strong reliance on inferences from planform and remotely sensed images in some sections of the thesis. As Baker (1993) notes this necessitates a striking interplay of inference and observation in order to achieve understanding, and the possibility of achieving understanding by such a method is contrary to the principles of the empiricist philosophy espoused by Hume who in *An inquiry concerning human understanding* (1955) states that 'causes and effects are not discoverable by reason but by experience' (p.42). The interplay between theory and observation noted by Baker (1993) and so clearly apparent at times during the research for this study highlighted the important role the assumption of theoretical realism plays in science (e.g. Popper, 1963; Boyd, 1973, 1983, 1985, 1990; Putnam, 1975; Hacking, 1983; Aronson et al., 1995). This can be seen in the debate surrounding the origin of drift deposits in which considerations of whether a proposed process could be real, and if real could have produced the deposits observed, were central to the debate (see Chorley et al. [1964] and Tinkler [1985]). Another example is provided by Lyell in the abstract for the paper on the glacial theory he presented to the Royal Society in 1839 which states that:

The distribution of an enormous mass of boulders on the southern side of Loch Brandy, and clearly derived from the precipices which overhang the Loch on the other three sides, is advanced as another proof in favour of the glacial theory. It is impossible to conjecture, Mr. Lyell says, how these blocks could have been transported half a mile over a deep lake; but let it be imagined that the Loch was once occupied by a glacier and the difficulty is removed. (quoted in Boylan, 1998, p.149-150)

Although Lyell almost immediately recanted his support for the glacial theory, the example nonetheless illustrates the important role of the assumption of theoretical realism in geomorphological reasoning. Arguments for realism have recently been advanced within geomorphology by Richards (1990), and while it is not appropriate to outline the full case here,
the examples above illustrate only one facet of what is a strong argument for geomorphic realism.

8.5 Suggestions for future research

The results in this study suggest a number of promising avenues for future research both within the study area and in fluvial systems generally. The data contained in this study and in Maroulis (2000) provide a good starting point for a detailed study of the accretion and dynamics of the contemporary floodplain. Such a study would shed further light on the temporal trends in channel pattern dynamics here and whether the position of the floodplain-surface channel patterns found is stable or changes in response to accretion patterns. The results in this study have shown that both OSL and TL would be suitable methods for such a study. A further beneficial use of dating would be in the investigation of channel migration rates. Chapter 4 proposed that migration rates systematically varied with channel size due to differing frequencies of channel boundary disruption by wetting and drying processes. The dating of floodplain sediments along transects extending away from the channel on the inside of meander bends should be able to quantify rates of channel migration, and if carried out across a spectrum of channel sizes, enable the findings of Chapter 4 to be tested.

Although reticulate channel systems are unusual they are not unique and are also developed throughout the Channel Country rivers and, although usually not quite so spectacularly as on the Cooper, on other Australian rivers such as the Namoi-Gwydir system (T. Pietsch, per. comm., 2001). It would be valuable to study other reticulate systems in order to ascertain whether the determinants of the occurrence and morphology of the reticulate pattern revealed here are general or vary between reticulate systems.

The direct measurement of overbank flow patterns around waterholes and in areas of differing floodplain-surface pattern would be extremely valuable, however such measurements are extremely difficult to obtain and attempts to obtain such data for this study were unsuccessful for a number of reasons. If a fortunate concatenation of events were to occur that made the direct measurements of floodplain flow patterns possible, every effort should be made to take advantage of the situation.

The study of channel junctions revealed a number of fertile areas for future research. The literature review in Chapter 5 highlighted the fact that bifurcations generally have been neglected and more study of the flow processes in and morphology of natural channel bifurcations, which was not possible here due to time limitations, is warranted. Analyses of the angles of bifurcations from other multiple channel systems would greatly benefit our
understanding of junction form and process. Particularly valuable would be observations of bifurcation evolution. Finally, the adequacy and generality of current extremal models of junction form and the assumption of junction efficiency was called into question by the findings here. Either the limits of the applicability of such models need to be investigated and clearly defined, or more suitable general models of junction form need to be developed.
Bibliography:


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**Appendix A: TL and OSL sample measurement results**

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### Appendix B: Channel planform analysis data

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Appendix C: Supplementary levee texture analysis results

Levee morphology and median grain size trends. Distance is from the channel edge. Only significant relationships (at $\alpha=0.05$) shown.
Levee morphology and clay content trends. Distance is from channel edge. Only significant relationships (at $\alpha=0.05$) shown. $r^2$ values apply to solid lines which indicate fitted regression lines.
Levee morphology and silt content trends. Distance is from channel edge.

Only significant relationships (at α=0.05) shown.