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A CATCHMENT FRAMED, PROCESS BASED APPROACH TO ANALYSIS OF THE EVOLUTIONARY TRAJECTORY OF THE TONGARIRO RIVER

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M.Sc. (1st Class Hons)

A thesis submitted as fulfilment for the degree of Doctor of Philosophy in Geography, The University of Auckland, 2013

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Declaration

I certify that this thesis has not been submitted to any other university or institution for a higher degree. Except where otherwise acknowledged, this thesis comprises solely my own work.

Helen E. Reid

September 2012
ABSTRACT

Major degradation has occurred to rivers world-wide, as channels and habitats are simplified and ecosystems impaired. Although river rehabilitation has been suggested as the best tool to stop this decline, efforts thus far have been marred by failures. A lack of process-based, geomorphic understanding is commonly given as a key reason. Multi-scalar understandings of space and time seek to address this shortcoming. This thesis applies analysis of landscape form at catchment, reach and bar scales to appraise process relationships within the Tongariro catchment, New Zealand. At the catchment scale, a DEM is used to analyse sediment conveyance within, and between, landscape units by calculating erosion indices, slope categories and drainage line stream power. Field-based analysis of the character and behaviour of streams across the catchment grounds this work. A sediment budget describes the bulk transfer of material since 1.8 ka, following the eruption of Lake Taupo. As this volcanic activity reset process zones in the catchment this presents a useful time constraint for this study. Mechanisms of channel adjustment and temporal patterns of response are assessed using 80 years of aerial photography for the 15 km reach of the lower Tongariro River. A framework to characterise river sensitivity is presented. Analyses of controls (i.e. bed material, transport capacity and valley confinement) within the wandering cobble bed reach are assessed. This provides a process-based explanation of within-reach variability in the pattern and rate of channel adjustment. At the bar scale, the distribution of bed materials was mapped using terrestrial laser scanning. Sediment entrainment during different magnitude/frequency floods was modelled. A conceptual model describing the geomorphic effectiveness of each flood event on within-bar surfaces is presented. Insights across multiple scales were synthesised using a catchment-framed, process-based, evolutionary trajectory approach. This is used to predict future pathways of geomorphic adjustment. Tongariro-specific findings include the resilience of the wandering, cobble bed reach, as a lag of lahar material retains steep slopes within the lower reach, flushing the active gravel fraction. Directly downstream, the highly sensitive braided reach captures this gravel, causing high rates of channel adjustment and widening. The meandering sand and delta reaches are narrowing towards a threshold condition, beyond which avulsion is likely. Tools which combine different scales and types of insights (i.e. qualitative landscape evolution and quantitative process based analysis) are necessary to provide a more comprehensive underpinning to rehabilitation schemes, allowing management strategies to ‘work with’ the underlying processes.
DEDICATION

This thesis is dedicated to all creatures great (e.g. Rich) and small (e.g. Poppy) which have made this thesis journey into an adventure.
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And finally, to the Tongariro River, thank you...

“Just to look at the Tongariro made us both happy. We could see bend after bend, pool after pool, rapid after rapid, all from the high bank at camp. There was a roaring rapid just above...a long narrow channel of white water, green at the edges, and just noisy enough not to be fearful...

It was what I call a gravel river, there being bars and banks, all heavy gravel, and a river bed of the same.... Perhaps the most striking feature of the Tongariro in this section... was the number and character of the islands. They were really gravel bars, under water when the river was in flood, and at low water picturesque green and grey islands around which the channels and rifts ran.”

(Grey, 1978: 247, written during a visit in 1927)
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1 INTRODUCTION

1.1 PROBLEM STATEMENT

Anthropogenic development has a long history of being centred along river corridors on alluvial floodplains, due to our dependence on rivers for drinking water, transport, irrigation, as a food source and many other uses (Karr and Chu, 2000; Lemons and Victor, 2008; Newson, 2009). However, the dynamic nature of many of these river systems, and the threat they pose to infrastructure, has increasingly posed a challenge for river managers. Past responses can be seen in attempts to stabilise and control river channels, acting to suppress natural forms of adjustment through engineering channel form in what are commonly termed ‘command and control’ approaches (Brierley and Fryirs, 2005; Ward et al., 2001). These measures suppress the ability of alluvial channels to self-adjust (Downs and Booth, 2011; Nanson and Huang, 2008). Channel aggradation or degradation commonly increases flooding or undercuts structures (Davies et al., 2003; Davies and McSaveney, 2006; Habersack and Piégay, 2007; Simon and Rinaldi, 2006). These responses cause the physical integrity of rivers to decline. Habitats become over-simplified and degraded, and ecosystem services are impaired (Beechie et al., 2010; Brierley and Fryirs, 2005; Gray, 2004; Ward et al., 2002; Wohl, 2004). This has led to calls to adopt ‘softer’, more ecologically based approaches to river management. The overwhelming consensus is that these approaches should be based on ‘working with the river’ through understanding and reinstating natural processes, particularly geomorphic processes which determine channel form and response (Beechie et al., 2010; Brierley and Fryirs, 2009; Clarke et al., 2003; Dufour and Piégay, 2009; Environment Agency, 2010; Kondolf et al., 2006; Lemons and Victor, 2008; Newson and Large, 2006; Palmer et al., 2005; Piégay et al., 2005; Ward et al., 2001; Wohl et al., 2005). However, as of yet, examples of application have been elusive. This presents a major gap within geomorphic research, as there is a need to create tools upon which management applications can ‘work with’ the specific functioning of a given system.

This thesis aims to address this knowledge gap through the development of a process-based evolutionary trajectory approach to river science and management. This framework is developed and trialled for the Tongariro River, Central North Island, New Zealand, but the approach is generic. Fluvial processes, most notably sediment transport and channel dynamics are analysed across multiple scales; namely, catchment, reach and bar scales. This multi-scalar scientific basis provides a platform to predict future trajectories of adjustment, presenting a proactive basis for management applications. Subsequent result chapters include comprehensive literature reviews at each scale of
enquiry (i.e. catchment, reach and bar) and as such, this chapter presents the key themes that run throughout the thesis and underpins this approach.

1.2 PROCESS BASED MANAGEMENT AND REHABILITATION

“... to be sustainable, river restoration projects need to effectively recreate a rivers’ functional characteristics taking into account the dynamic geomorphic characteristics”

(Lemons and Victor, 2008: 1)

Process-based rehabilitation is defined as a project which “...aims to re-establish normative rates and magnitudes of physical, chemical, and biological processes that create and sustain river and floodplain ecosystems” (Beechie et al., 2010: 209). Within river systems, this requires an understanding of key fluvial processes which drive channel adjustment, most notably sediment transport, so that management does not impede this process, and repress fluvial dynamics. This section describes the circumstances that have led to the call for adoption of process-based management applications.

Widespread recognition of the damage and simplification natural systems have undergone, and resulting disproportionately high extinctions of riverine biota, has resulted in a drive for the rehabilitation and restoration of rivers (Bernhardt et al., 2005; Wohl et al., 2005). Currently, over 1 billion USD has been spent on river restoration in the USA alone (Bernhardt et al., 2005). However, to date, successes have been limited (Bernhardt and Palmer, 2011). This is attributed, in large part, to the fact that most projects have endeavoured to recreate the form of systems, rather than working with the processes which underpin river morphology (Clarke et al., 2003; Simon et al., 2007; Wohl et al., 2005). A component of this failing can be seen in the difficulties in measuring processes, particularly bedload transport, at a spatial and temporal resolution that is relevant to management (Ashmore and Church, 1998; Brasington et al., 2003; Wilcock et al., 2009). In addition, geomorphic findings and scientific findings are commonly not presented in a format which makes them useful for river managers (Lave, 2009a; Lave et al., 2010; Palmer, 2009). Finding new techniques for a) analysing and measuring riverine processes at appropriate temporal and spatial resolutions and b) transferring these findings from geomorphic research into management agendas presents a compelling new area of geomorphic research.

Process-based approaches recognise that predicting channel adjustment incorporates an inherent level of uncertainty (Darby and Sear, 2008). Proactive management and rehabilitation schemes should strive to incorporate this uncertainty and complexity in the planning process (Brookes and Dangerfield, 2008; Lemons and Victor, 2008). This can be underpinned by process-based geomorphic insights, whereby it is recognised that different types of river may be characterised by different
degrees of sensitivity and types of response (Brunsden, 2001; Fryirs et al., 2009; Thomas and Allison, 1993). This reflects a perspective whereby rivers are considered to be better at self-adjusting in response to changes in fluxes, and creating viable habitat, than endeavours by humans to design and develop particular process-form outcomes (Darby and Sear, 2008). This view of rivers as ‘living systems’ (Everard and Powell, 2002) underpins management schemes that facilitate channel adjustment to prevailing flow and sediment fluxes by giving channels ‘room to move’ within a designated space, thereby promoting habitat turnover (Piégay et al., 2005; Rapp and Abbe, 2003). However, understanding the area necessary to contain ‘natural’ patterns of adjustment needs to be grounded by a comprehensive geomorphic survey of riverine processes across multiple scales.

Despite the growing recognition of the need for process based approaches to river restoration, examples of these have only recently started to appear in the literature. Whilst Beechie et al. (2010) list the processes that are important (i.e. sediment transport and channel migration) they do not suggest techniques for measuring these dynamics. Notable exceptions which do suggest techniques include Piégay et al. (2005) who used historical channel adjustment to delineate erodible river corridors and Rapp and Abbe (2003) who introduce a procedure for delineating channel migration zones by mapping historical migration zones, avulsion hazard zones and bank erosion hazard zones. Sear (1994) present a conceptual model for creating process based restoration schemes. Wheaton (2004) developed a model which predicts how habitats will adjust for different restoration options. While these approaches do an adequate job at estimating variability in channel location in the short-term, they do not attempt to understand process dynamics or on-going trajectories of change that drive channel behaviour across longer timescales. In addition, these approaches are largely based at the reach scale, ignoring driving influences at the catchment scale and local scale process controls. However, approaches have been developed which model sediment transport and connectivity at the catchment scale to provide a process base for management (Downs and Priestnall, 2005; Lane et al., 2008; Reid et al., 2007; Smith et al., 2011). However, linking analysis of process at the catchment and reach scales still provides a major challenge for restoration approaches and provides a major opportunity for the development of geomorphic frameworks.

Brierley and Fryirs (2005) present the River Styles Framework® which is a multi-scalar, comprehensive geomorphic assessment of channel character and behaviour which is used to predict future trajectories of change. However, this framework is largely qualitative and does not quantify sediment transport dynamics which are required to predict future rates of change. This leaves a fundamental need to create a quantitative, multi-scalar approach to guide process-based management and restoration. Downs and Gregory (2011) support this by discussing the need to
‘recast’ the role of geomorphology so that it is applicable for river management, aimed at supporting ecosystem services rather than controlling river channel and form.

Figure 1.1: An overview of processes and controls which drive bed material transport and channel adjustment. Catchment-scale controls drive sediment flux across larger scales, while reach-scale controls determine the propagation of sediment through a reach and ability of the channel to adjust. The balance of impelling (hydrological regime, slope) and resisting (bed material) forces determines whether bed material can be transported, which causes planform adjustment. This process-based underpinning can be used to develop management applications which work with the system. Guiding concepts (shown in green font) can be used to characterise patterns of sediment transfer (connectivity) and planform adjustment (sensitivity).

Figure 1.1 presents a conceptual framework that summarises the components seen as necessary to incorporate within a process-based approach to predicting the evolutionary trajectory of a river. This
Chapter 1: Introduction

outlines the framework for this thesis. Firstly, the catchment-scale controls which drive sediment generation and transfer across the catchment should be considered. Sediment connectivity across the catchment determines the volume and nature of sediment delivery for the reach of interest (Fryirs et al., 2007b; Fryirs, 2012; Kondolf et al., 2006, Chapter 3). Individual reach scale controls such as slope, channel planform and valley confinement dictate the connectivity of the reach (its ability to transport the sediment delivered), which in turn drives patterns of reach sensitivity, as reaches unable to move the sediment delivered will have to store it locally (Fryirs, 2012; Hooke, 2003b; Hoyle et al., 2008, Chapters 4 and 5). Local scale variables such as bed material characteristics and vegetation roughness are key to understanding how frequently these surfaces are able to be reworked, and thus the likely temporal pattern of channel change (Mao and Surian, 2010; Surian et al., 2009a, Chapter 6). Whilst anthropogenic influences are not mentioned on this diagram, it is recognised that they alter the characteristics and relationships between controls across all scales, providing an additional layer of complexity. Thus, anthropogenic influences should be viewed with regard to how they influence specific controls, altering patterns of sediment transport. The importance of scale and the rationale for these three specific scales is discussed in further detail both in Section 1.5 and within literature reviews at the beginning of each results chapter (Chapters 3-6).

The multi-scalar insights presented in Figure 1.1 drive bedload transport, causing planform adjustment, which in turn dictates the future locales of bedload transport. A core aim of this thesis is to assess the relationship between sediment transport and planform adjustment across multiple scales within the Tongariro catchment. This will be used to provide a process-based underpinning upon which to predict channel change in the Tongariro system. Once this foundation is established, management approaches will be suggested, based on the specific functioning (and sensitivity) of each reach.

1.3 USE OF GEOMORPHIC APPROACHES TO UNDERPIN RIVER REHABILITATION AND MANAGEMENT

In recent decades considerable attention has been given to development of approaches to analyse and describe fluvial processes for use in rehabilitation and management. This section provides a brief overview of types of approaches used to underpin the selection of the tools used within this thesis.

Significant advances in geomorphic understandings of river systems were established in the 1950s. This foundation let to the proliferation of papers and ideas that aimed to enhance understanding into how rivers work. Examples of key developments that emerged at this time included analysis of
hydraulic geometry relationships by Leopold and Maddock (1953) which used empirical mathematical equations to relate at a site and downstream changes in width, depth, velocity, slope and roughness to changes in discharge. Lane (1955) presented the ‘Lane relationship’, which demonstrated that a change in either slope and discharge or sediment load and size would result in the channel either incising or aggrading to balance this change in energy or sediment load. Leopold and Wolman (1957) defined the key river channel patterns as braided, meandering or straight and investigated how underlying controls determined these patterns using flume experiments. This identified the meandering-braided threshold, whereby increasing slopes and bankfull discharges caused the channel to adjust from a meandering planform to a braided. Wolman and Miller (1960) presented the concept of effective discharge, whereby the frequency of discharges and their sediment loads were compared to discover which floods were responsible for transporting the most sediment in a system. They found that relatively frequent floods transported the most sediment overall. Many of these concepts were combined into the seminal textbook ‘Fluvial Processes in Geomorphology’ by Leopold et al. (1964). (Schumm and Lichty, 1965) highlighted the importance of time and space in determining causes of geomorphic state at any point in time. Schumm (1960; 1968; 1977; 1980; 1985) advanced these conceptual developments through application of system principles. He conceptualised the catchment using three process zones; a source zone in the headwaters, a transfer zone in the middle reaches an a zone of deposition in the lower reaches, relating these back to the downstream shifts in energy, sediment size and valley confinement (Schumm, 1977). Hey and Thorne (1986) used multiple regression to relate width, depth, slope, velocity, sinuosity and riffle spacing to the key controls of bankfull discharge, bed and bank materials, valley slope, vegetation and estimated bedload transport rates to gain a greater understanding on controls on channel form. Many of these cross-scalar principles and applications were summarised by Church and Mark (1980) and Church (1996).

These conceptual framings set the scene for an increasing focus on quantifying processes at more local scales in the 1980s and 1990s (c.f. Lane and Richards, 1997). This built upon flume experiments which provided greater understanding of channel morphodynamics and bar development, particularly in multi-channelled gravel based systems (Ashmore, 1982; 1991b; Ashworth, 1996; Hoey and Sutherland, 1991) or analyses of confluence dynamics (e.g. Best (1988). These advances were mirrored by integrative field and modelling studies. Key examples include work by Richards (1976a; b; 1978) on the dynamics of riffle-pool sequences, Dietrich and Smith’s (1984) work on bedload transport and meander development, and work by Bridge and Gabel (1992) which characterised sediment transport across a range of flows to describe patterns of deposition and erosion and explain bar development. Notable technological advances and technique development furthered our
understandings of sediment transport dynamics (Ashworth and Ferguson, 1989; Carling et al., 2006; Ferguson, 1992; Ferguson et al., 2002; Haschenburger and Church, 1998; Laronne and Duncan, 1992). The growth in knowledge about channel form underpinned by an increased understanding of fluvial processes paved the way for the development of frameworks to describe fluvial process and underpin river management. In recent years these applications have been greatly extended through remotely sensed techniques (e.g. Carbonneau and Piegay, 2012; Fonstad et al., 2013).

These seminal works provide the underpinnings for many of the key texts and approaches which use geomorphology to guide river management (c.f. Brierley and Fryirs, 2005; Brookes and Shields, 1996; Dangerfield et al., 2005; Downes and Gregory, 2004; Knighton, 1998; Kondolf and Piégay, 2003; Petts and Amoros, 1996; Rosgen, 1996; Sear et al., 2003; Sear et al., 2010; Thorne, 1998). Particularly noteworthy is Brookes and Shields’ (1996) book ‘River Channel Restoration’ which summarises how geomorphic principles should be used to underpin restoration, including the development of catchment approaches and a technique assessing how sedimentary features indicate system instability. Sear (1996) presents a conceptual approach, whereby catchment history, actual channel response and the pre- and post-restoration boundary conditions are used to develop appropriate restoration options. Kondolf and Piégay’s (2003) ‘Tools in Fluvial Geomorphology’ reviews and summarises geomorphological techniques including remote sensing, numerical modelling, river classification, sediment budgets and procedures for measuring bedload transport. These techniques provide the basis for many studies which analyse river process, without implicitly providing a framework for how to use them to describe their use for restoration schemes.

More recently, there has been a growth of structured frameworks which analyse geomorphic principles. These approaches commonly relate the morphology observed back to the health or condition of the systems based on notions of the naturalness of the system. Table 1.1 provides an overview of geomorphic frameworks which have been used to underpin river rehabilitation. Many are based on recording geomorphic features at a site. The River Habitat Survey (RHS), specifically, has few links to the geomorphic processes underpinning the habitats being recorded and does not analyse the condition of the systems relative to its type of river in a meaningful way (Raven et al., 1998). Similarly, the Fluvial Audit requires tallying features present at a site, with a greater link to process, based on using the form of geomorphic units to indicate the processes that created them (Dangerfield et al., 2005; Sear et al., 2003). However, this is still effectively a ‘tick-box’ approach that records features present at a site, and relies on the skill of the geomorphologist to combine and use these data to assess how underlying controls may be influencing their evolution. Rosgen’s (1996) Natural Channel Design approach is particularly contentious (Lave, 2009b; Small and Doyle, 2012).
measures geomorphic variables such as slope and bed material and uses this to classify the type of channel that should exist at a location, ignoring the overlap in controls that is apparent between different types of rivers. It has been criticised as being a ‘cook book’ based approach which is easy to apply without the fundamental understanding in geomorphology necessary to create a channel restoration (Kondolf et al., 2003b). A lack of a rigorous process-base has been identified as a key component for some of the high profile restoration schemes which have failed after using this approach (Kondolf et al., 2003b; Simon et al., 2007). More recent approaches have evolved in Europe in response to the need to assess hydromorphic quality as a requirement of the European Union Water Framework Directive legislation (Table 1.1). This includes the Morphological Quality Index (MQI) (Rinaldi et al., 2013) and Index of Hydromorphic Quality (Oller et al., 2011). Both of these select geo-indicators and score them based on geomorphic condition. The MQI classifies the type of river and relates the variables used back to the applicability in different valley settings.

The use of river classification to underpin analysis of channel health is a key component of the River Styles Framework, which was used within this study (Brierley and Fryirs, 2005). However, river classification was predominately used within this work to provide an overview of how river form and the underlying processes changed across the catchment. The River Styles Framework uses an open ended approach to classification, rather than forcing river forms into pigeon holes as occurs within approaches that use pre-ordained river types. This individually classifies reaches based on the specific boundary conditions that shape the behaviour of a reach, including valley confinement, channel planform, geomorphic units and bed material. This type of approach is particularly important for use in the Tongariro River due to the variability in river type and catchment controls present in this diverse setting.
### Table 1.1: Summary of key approaches which use geomorphology to underpin river rehabilitation.

<table>
<thead>
<tr>
<th>Approach</th>
<th>Author</th>
<th>Location developed</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>Natural Channel Design</td>
<td>(Rosgen, 1996)</td>
<td>USA</td>
<td>This is the most commonly used approach for Restoration Schemes in the USA. It is based on measuring controls for rivers (e.g. entrenchment, gradient, width-depth ratio, bed material and gradient) and placing reaches within pre-existing river types. The river types and then used as guidance to decide what other restored morphologies could exist at this location.</td>
</tr>
<tr>
<td>River Habitat Survey</td>
<td>(Raven et al., 1998)</td>
<td>River Habitat Survey</td>
<td>The River Habitat Survey has been applied across Britain to record habitat features at a site. This assesses a 500 m stretch of river and records physical attributes at 10 points along 10 transects within the 500 m stretch. This is done by ticking boxes based on the presence of habitat features (i.e. macrophytes) and description of bed and bank substrate and flow characteristics. Habitat quality and habitat modification scores are used to describe the condition of the reach.</td>
</tr>
<tr>
<td>Fluvial Audit</td>
<td>(Dangerfield et al., 2005; Sear et al., 2003)</td>
<td>UK</td>
<td>This is a catchment scale approach to derive geomorphic understandings where sediment related problems have been identified. Initial stages involve collating historical information to chart how the catchment has changed. Geomorphic features are tallied in the field concentrating on indicators of sediment sources, transport and sediment sinks. This also records valley shape, channel geometry and boundary conditions. Reach stability is classified based on the presence of geomorphic features, and Potentially Destabilising Phenomena (PDPs) are identified and related to whether they increase (e.g. mining spoils) or decrease (e.g. dams) sediment supply (for full list see Sear, 1996).</td>
</tr>
<tr>
<td>River Styles</td>
<td>(Brierley and Fryirs, 2005)</td>
<td>Australia</td>
<td>Catchment scale classification of river type, which provides a basis for analysing condition, based on geomorphological indicators which describe how a specific river type would adjust in a good, moderate and poor condition. Analysis of channel evolution plays a key role in explaining causes and direction of adjustment.</td>
</tr>
<tr>
<td>Index of Hydromorphic Quality (IHG)</td>
<td>(Ollero et al., 2011)</td>
<td>Spain</td>
<td>A score is created based on the condition of the fluvial system, channel morphology and riparian margin. This assesses the naturalness of many variables including flow regime, sediment supply, floodplain functionality, channel planform and the quality of the riparian margin (width and longitudinal connectivity). Each of these variables is assessed to provide overall scores.</td>
</tr>
<tr>
<td>Morphological Quality Index (MQI)</td>
<td>(Rinaldi et al., 2013)</td>
<td>Italy</td>
<td>Rivers are separated into segments based on homogeneous geomorphic characteristics and typology is defined based in planform, geomorphic units and valley confinement. Three key components are assessed based on geomorphic functionality, artificiality and channel adjustments; these are lateral and longitudinal continuity, morphology, including channel pattern, cross-section and bed configuration, and riparian vegetation. The final output is a score which describes the condition of the overall morphology.</td>
</tr>
</tbody>
</table>
The aim of this work was not to provide a way of scoring the condition of the system as is a common theme in the approaches above (Table 1.1). Rather, classification procedures were used to analyse the geomorphic processes operating across multiple scales, providing a quantitative basis upon which to understand river adjustment. The key aim of this approach was to analyse geomorphic processes using form to infer process across the catchment, reach and bar scales. Combining the range of types of output across multiple scales presented a challenge to this work, which previous approaches had not tackled. An evolutionary trajectory approach was used, which is a key component of the River Styles Framework to combine these outputs in a coherent way (Brierley and Fryirs, 2005). This approach has appeared in the literature elsewhere (Dufour and Piégay, 2009; Fryirs et al., 2012; Ziliani and Surian, in press). However, in this study a modified approach was used, analysing channel adjustment for each reach and integrating process understandings across multiple scales using a series of diagrams displaying landscape evolution, controls on channel form and predictions of response to different natural and anthropogenic influences. In this way, the River Styles Framework (Brierley and Fryirs, 2005) provided a foundation for the approach developed within this study. However, this thesis extends this approach by using a range of tools to characterise and quantify fluvial processes and further the use of process-based geomorphology for river rehabilitation.

1.4 THE USE OF MODELLING OF GEOMORPHIC PROCESS RELATIONSHIPS TO UNDERPIN RIVER REHABILITATION

The uptake of modelling as a key tool to analyse geomorphic systems in the past two decades has been rapid. This uses datasets detailing channel form, commonly cross-sections or Digital Elevation Models (DEMs) to route flow and sometimes sediment to explore the spatial patterns of how channels and landscapes are likely to adjust. Therefore, careful consideration was given to application of these techniques in this study. Of particular interest was the use of Landscape Evolution Models (LEMs), which are coarse resolution models that assess landscapes change over geologic time scales (Coulthard et al., 2007). These are based on simplified sediment transport equations following the work by Murray and Paola (1994) which found reduced complexity equations to be surprisingly astute at predicting braided channel development, leading to the introduction of reduced-simplicity models into fluvial geomorphology (Brasington and Richards, 2007; Coulthard et al., 2007). Two approaches which successfully capture fluvial processes are SIBERIA, validated with measured erosion rates, and CAESAR, validated using dated flood deposits (Coulthard et al., 2007).

In addition, there has been a proliferation of reach scale models which attempt to mimic bedload transport, channel adjustment, flood inundation and habitat development. The production and use
of flood inundation models has become a lucrative industry, with many developments requiring a quantified assessment of flood risk. The most commonly used include MIKE 11 (Thompson et al., 2004), TUFLOW (BMT and WBM, 2010) and ISIS (Neal et al., 2009). Most of these models have been developed to tackle specific geomorphic issues, with particular emphasis upon sediment transport dynamics. Darby et al. (2002) have developed a model to characterise bank erosion so that the mechanics of channel migration can be better understood. Wheaton et al. (2004) created hydrodynamic, habitat and sediment entrainment models which predict how different restoration options would alter habitat characteristics. Models are often commonly used across multiple reaches to describe how controls alter process dynamics downstream. Hoyle et al. (2011) used HEC-RAS to model the distribution of shear stresses, so that the mobility along a reach could be assessed. Raven et al. (2011) developed a sediment routing model which incorporates channel migration, bed adjustment and sediment coarsening or fining to characterise channel response and patterns of sediment transport. Simple reduced complexity models have also been used to characterise channel dynamics in both braided rivers (Murray and Paola, 1994; Thomas and Nicholas, 2002) and meandering channels (Coulthard and Wiel, 2006). Indeed, the literature is now saturated with examples of modelling and it has become a key tool to explain geomorphic processes. However, despite its strengths, it also has weaknesses which are explored below (c.f. Darby and Van De Wiel, 2003).

Models are inherently a simplification of reality. Coulthard et al. (2007) discuss the limitations of modelling bedload transport due to our still limited understanding of the phenomena due to inherent temporal and spatial variability (c.f. Hicks and Gomez, 2003; Wilcock et al., 2009). In addition, modelling has high time and data requirements. Table 1.2 presents an overview of limitations of modelling (c.f. Darby and Van De Wiel, 2003). These limitations help explain why modelling was not pursued in the Tongariro for this thesis. Marked heterogeneity is evident within the Tongariro catchment making the prospect of modelling the catchment likely to omit significant processes (Table 1.2). This can be seen in the distinct differences in landforms and processes in the headwaters, with active volcanic cones to the west and uplifting greywacke ranges to the right. Modelling these disparate processes accurately would provide a substantial challenge. Model imperfection describes the problem that models developed in the lab may not accurately correspond to the real world. Within the Tongariro River, a lag of lahar deposits lines the bed making it less mobile compared to flume beds upon which sediment transport equations and models are based. This complexity also drives a high degree of sediment heterogeneity which would be difficult to simplify into model parameters. This diversity of grain size attributes and vegetation character would also make accurately setting initial conditions difficult (Table 1.2). Indeed, it is difficult to determine
which sediment size would be the best to add into a model when the size of material making up the bars has a large size range (boulders to sand). The banks and terraces are also very complex dependant on whether they are lahar or fluvially reworked deposits. In addition, grain size and channel type adjust rapidly downstream providing a downstream variation in sedimentological complexity in the system. River systems are non-linear and exhibit sensitivity to initial conditions within models, which can increase uncertainty for predicting future change (Table 1.2). The complexity of extreme events within the Tongariro system means the river may be recovering from floods, volcanic eruptions and lahars, making understanding of the stage of river recovery particularly complex and problematic.

In essence, the complexity of the evolution, sedimentology, and controls within the Tongariro catchment drove the selection of a non-modelling approach within this setting. Simplification of this system for modelling was seen to be likely to remove key controls that drive the unique character and behaviour of the system. Instead, it was determined that a form-process approach based on more traditional approaches to geomorphic enquiry was better suited to analysis here, as it allowed sufficient flexibility to capture the diversity of the system was necessary to analyse the dynamics within the Tongariro. Hence, a multiple lines of evidence approach was adopted, whereby different types and scales of analysis were combined to provide a comprehensive overview of the morphodynamics of the Tongariro system. This recognised the need to understand sediment transport at the catchment and landscape evolutionary timescale, in a form that would capture how both volcanic influences and floods had worked to shape the system. At the sub-reach scale, this approach analysed how sedimentology and grain size characteristics changed and how this was necessary to underpinned bar reworking and channel adjustment. Due to the dynamic nature of the river and its multiple overlapping controls, it was recognised that an understanding of channel response and how reach scale controls influenced this, was necessary to predict future channel dynamics. Thus, the approach developed within this thesis was based on building field based insights into the complexity of the setting and using these to predict channel dynamics, rather than simplifying these controls in a model in the hope that it will be able to accurately represent channel dynamics.
Table 1.2: Inherent limitations of fluvial geomorphical models (sourced from Darby and Van De Wiel, 2003) with each factor assessed based on its relevance to the Tongariro catchment.

<table>
<thead>
<tr>
<th>Limitation</th>
<th>Explanation/Example</th>
<th>Relevance to the Tongariro catchment</th>
</tr>
</thead>
<tbody>
<tr>
<td>Model Imperfection</td>
<td>This refers to the fact that incremental ‘improvement’ in models at laboratory scale do not necessarily add to our ability to make predictions at larger scales. For example, sediment transport is difficult to predict due to the uniqueness of each sediment bed.</td>
<td>Very relevant. The Tongariro displays a high diversity in sediment size across the bars due to the complex evolution which combines sediment derived from lahar flows with fluvially sourced material into process relationships. In addition, models designed from flume studies are based on fully alluvial gravel beds. The lahar lag in the Tongariro River makes it less mobile than the beds which the models are based on.</td>
</tr>
<tr>
<td>Omission of significant processes</td>
<td>The larger the scale of the fluvial system the greater the chance that more than one important process will be omitted. For example fluvial sedimentation models of debris fans that may ignore debris flow processes.</td>
<td>The Tongariro catchment is highly diverse, with markedly different processes in the Kaimanawa Ranges (i.e. debris flows) compared to the Volcanic zone (i.e. volcanic eruptions). Accurately representing both of these would be very difficult. In addition, the frequency and volume of sediment generated from volcanoes is unpredictable and therefore difficult to incorporate into modelling.</td>
</tr>
<tr>
<td>Unknown initial conditions</td>
<td>Such as the distribution of grain sizes, bank material characteristics, bed topography, etc. These conditions are often only known approximately or not at all.</td>
<td>The Tongariro has had LiDAR data flown so the morphological data is available. However, sediment distributions within the bed and banks are very complex making simplifying them for a model associated with a high area of error. This complexity is driven by channel incision into deposits created by lahars which deliver a range of material dependant on the characteristics of the lahar event (i.e. from boulders to sand). These deposits have more recently been reworked by flood events. This would make defining initial conditions that capture the complexity of the bed material complicated and prone to error.</td>
</tr>
<tr>
<td>Sensitivity to initial conditions</td>
<td>Fluvial systems are non-linear, so there can exist a sensitivity to initial conditions that effectively prohibits detailed prediction of system evolution.</td>
<td>The Tongariro is highly dynamic, undergoing channel avulsions which rework floodplain and bar surfaces, removing vegetation in response to floods. Infrequent lahars deliver large sediment load through the terraces. Time since the last event is likely to influence the extent of channel response which is particularly complex where the channel is recovering from multiple events that are characterised by different magnitudes of impact and frequency of occurrence.</td>
</tr>
<tr>
<td>Unresolved heterogeneity</td>
<td>In large-scale fluvial systems it may be impossible to define a meaningful averaging volume for each computational cell. Heterogeneity is apparent in factors such as vegetative cover and soil type.</td>
<td>This is true particularly due to bed material and vegetation distribution. In addition, much of the bed of the river is larger than the sediment size regularly transported through the system, making it difficult to estimate what sediment size should be used in the model as the bed is not fully alluvial.</td>
</tr>
<tr>
<td>External forcing</td>
<td>In fluvial systems, external forcing may be due to increases of discharge resulting from storms or dam releases. Predictive capabilities are limited if unpredictable external forcing can occur. In many cases forcing can only be incorporated statistically if the distribution of events in known.</td>
<td>The externally derived volcanic inputs provide a complicated scenario of external forcing within this catchment. Eruptions are very hard to predict.</td>
</tr>
</tbody>
</table>
1.5 THE MORPHOLOGICAL APPROACH: USING PROCESS FROM FORM

“Mutual interactions between form and process are the core of geomorphic investigation – form affects process and process affects form...The nature of the two-way connection between Earth surface process and Earth surface form has lain at the heart of geomorphic discourse”.

(Huggett, 2011: 174)

Section 1.2 describes the need to understand geomorphic processes, particularly sediment transport so that process-based management approaches can be designed. Well-functioning fluvial systems allow the formation of appropriately heterogeneous habitat assemblages, based on river type, which can support ‘healthy’ ecosystems (Beechie et al., 2010; Clarke et al., 2003; Reid et al., 2010; Wohl et al., 2005). Thus, an understanding of channel adjustment and response to changes in sediment flux is an essential underpinning for management (Figure 1.1). However, measuring sediment transport, particular bedload is notoriously difficult due to high spatial and temporal variability that does not accurately follow differences in shear stresses (Bridge, 2003; Gomez, 1991; Hicks and Gomez, 2003). As a result, tools which accurately measure bedload transport have thus far remained elusive, presenting one of the greatest challenges to geomorphology (Burt and Allison, 2010; Davies, 1987; Ferguson and Ashworth, 1992; Gray et al., 2010; Hicks and Gomez, 2003; Hubbell, 1987; Wilcock et al., 2009). However, on-going difficulties in measuring bedload transport have been paralleled by an exponential growth in techniques which can capture channel form (Hicks and Gomez, 2003). Such tools include terrestrial and airborne laser scanning, RTK-GPS, aDcp and sonar devices to measure the bed and remote sensing applications which can utilise aerial photography and satellite images (Milan and Heritage, 2012; Williams et al., 2011). These tools have greatly improved both the scale (airborne LiDAR can survey whole catchments), resolution (terrestrial laser scanning can record surfaces at >1 point per cm²), and precision of data acquisition.

The use of form to infer process is a technique that has been around for decades (Huggett, 2011; Rhoads and Thorn, 2011). However, the increase in our ability to capture form has led to a renaissance of this approach to quantify sediment transport. Ashmore and Church (1998) advocated for its importance, suggesting that the ‘morphological approach’ should form a new paradigm of geomorphic research. This suggestion has been followed by a growth in studies which use geomorphic form to measure fluvial processes, such as sediment transport.

Applications of the morphological approach have evolved as the scale and resolution with which form can be measured has increased. Early attempts can be seen in quantifying meander migration rate from aerial photographs to estimate the volumes of sediment eroded and deposited (Neill,
This evolved to comparing cross-sectional surveys over time to gain measures of erosion and deposition for each cross-section (for example see Hicks and Gomez, 2003; Warburton et al., 1993). As sediment transport rates are interpolated between each cross-section, this method was also associated with a high degree of error (Brasington et al., 2000; Fuller et al., 2003a). These approaches have been further developed by Ham and Church (2000) who combined morphological changes from aerial photographs, with field surveys of geomorphic unit height to create a sediment budget for a reach of the Fraser River, in British Columbia, Canada. Hooke (2003b) related channel connectivity and sensitivity to underlying processes by using field mapping, bed material characteristics and calculations of channel competence to determine zones of erosion and deposition within a reach. More recently, the morphological approach has been carried out by generating multiple high resolution Digital Elevation Models (DEMs) which are compared to quantify erosion and deposition within a reach. Key examples include RTK-GPS surveys by Brasington et al. (2000; 2003) of the River Feshie, Scotland and Fuller and Basher’s (2012) survey of the Motueka River in New Zealand. Terrestrial laser scanning (TLS) has once again increased the resolution, scale and accuracy of these data sets, and the quality of the data used (Milan and Heritage, 2012). Williams et al. (2011) fuse TLS point clouds with water depth derived from optical bathymetric data to create a sediment budget within a 3 km long braided river reach in New Zealand, while Milan et al. (2007) use TLS to quantify sediment erosion and deposition in a pro-glacial river in Switzerland. These examples illustrate the rapid evolution and increasing use of the morphological approach to characterise the distribution and rates of bedload transport.

This thesis argues that the use of the ‘morphological approach’ can be broadened to encompass a greater range of techniques whereby form is used to infer process. To allow this, the definition of ‘form’ and ‘process’ as used within this thesis is presented. Huggett (2011) defines form as the morphological structure and geometric properties of a landscape element. This can include a bar form, which indicates how much sediment may have accumulated in a given time period or terraces which describe net erosion. In contrast, process is associated with an expenditure of energy, including power, energy flux, force and momentum (Huggett, 2011). Within earth science this incorporates processes that facilitate the distribution of erosion and deposition of sediment across the earth’s surface, such as stream power or shear stress equations, which provide the basis for most fluvial models (Church, 2010; Finlayson and Montgomery, 2003; Whipple and Tucker, 1999). Furthering approaches which capture geomorphic form across a range of scales in efforts to model process is central to this thesis. This framing underpins the multi-scalar approach that is used.
Previous approaches to analysis of form-process approaches can be characterised in two ways. Firstly, they are nominally applied at the reach scale. However, it is possible to appraise form across a range of scales to measure process. For example, the volume of sediment in a delta (form) can be used to calculate a bulk sediment delivery rate over time (process) (Buijsman et al., 2003; Goodbred and Kuehl, 1999). In addition, catchment form from a DEM can be used to extract characteristics such as slope and drainage area to calculate the distributions of stream power across a catchment (Finlayson and Montgomery, 2003; Whipple and Tucker, 1999). Secondly, processes analysed have been reduced to direct measurements of sediment erosion and deposition over time. However, as described above, rather than directly measuring processes, measurements of form can be used to deduce process, through use of physics-based fluid mechanics. Measurements of channel cross-section and slope (form) can be used to determine potential energy, such as stream power, shear stress and sediment entrainment (process) (Church, 2006; 2010; Pitlick et al., 2009). This is based on the ability of a reach to entrain sediment delivered, and thus the likelihood of erosion or deposition for a given flood event. This broadening of the form-process approach aims to use form across a greater range of scales, including using DEMs to derive catchment scale patterns of sediment transport, down to individual bar forms to analyse sediment entrainment and bar reworking. Moreover, analyses of geomorphic form are used to infer a wider range of processes than has been traditionally applied. This includes direct measures of erosion and deposition as carried out in the past (Ashmore and Church, 1998; Brasington et al., 2003; Buijsman et al., 2003; Fuller and Basher, 2012), use of cross-sectional form to measure shear stress-based transport capacity (Church, 2010; Pitlick et al., 2009) and the use of bar roughness to map grain size and model sediment entrainment (Milan and Heritage, 2012; Rychkov et al., 2012).

The form-process approach developed in this thesis provides the quantitative foundation upon which relationships between bedload transport and planform adjustment are assessed (see Figure 1.1). This relationship is contextualised within the catchment and reach-scale controls as well as key guiding concepts such as connectivity (Chapter 3) and sensitivity (Chapter 4). Collectively, cross-scalar geomorphic applications, tied to techniques to quantify process provide a flexible tool to construct process-based evolutionary trajectories.
1.6 EVOLUTIONARY TRAJECTORIES AS CONCEPTUAL MODELS

“It has been well known for many years that a full understanding of the nature and causes of landforms requires appraisal of form, process, materials and time.”

(Richards and Clifford, 2011: 2)

The evolutionary trajectory approach is a technique that documents the timeframe, rate and type of river adjustment for a given reach, from which future pathways of adjustment can be predicted (Brierley and Fryirs, 2005; Dufour and Piégay, 2009; Fryirs et al., 2012; Surian et al., 2009b). This may be contextualised within catchment scale patterns of adjustment, so that the distribution of the timing and types of reach-scale adjustment (i.e. the response gradient), may be determined (Fryirs et al., 2009). Rivers act as archives, that if accurately interpreted, may offer insights into human, environmental and biological history (Wohl, 2004). Insights from past river adjustment provide a foundation for understanding how they are likely to function in the future (Brierley and Fryirs, 2008; Lane and Richards, 1997). Evolutionary trajectories can be used to identify past thresholds, interpret responses to events of different magnitudes, assess channel sensitivity and appraise reach to reach linkages of sediment flux (i.e. connectivity) (Fryirs et al., 2009; Hooke, 2003b; Schumm, 1980; Sear, 1994). Multiple, overlapping controls (e.g. climate and anthropogenic influences) impact upon channel dynamics and evolution (Dufour and Piégay, 2009, Figure 1.1). Using past adjustments to predict the trajectory upon which a river is moving provides a temporally dynamic geomorphic grounding for river rehabilitation initiatives (Brierley and Fryirs, 2005; Brierley et al., 2008). This extends beyond reference reach/lietbild approaches which are based upon static, temporally fixed form-based system views. Despite its importance and its support within the literature (c.f. Brierley and Fryirs, 2008; Dufour and Piégay, 2009; Fryirs et al., 2012; Sear, 2002; Wohl, 2004; Ziliani and Surian, in press), evolutionary trajectories are commonly underutilised or poorly represented within fluvial geomorphological applications.

Previous approaches to analysis of evolutionary trajectories have represented channel evolution in various ways (see Figure 1.2). Graphical representations include maps showing changes in channel position, cross-sectional changes or conceptual diagrams (Ham and Church, 2000; Rinaldi and Simon, 1998; Winterbottom, 2000). Evolution can be expressed through changes to morphological variables including channel width, sinuosity and braiding, erosion and depositional indices (Ham and Church, 2000; Hooke, 2008, c.f. Figure 1.2A). Different insights can be obtained from different techniques. Conceptual diagrams such as that seen in Simon and Thomas (2002) describe the evolution of a specific site in response to knickpoint retreat so that a site may be assessed at a particular stage along a trajectory (Figure 1.2B). Fryirs et al. (2012) similarly use ergodic reasoning (space for time
Chapter 1: Introduction

substitution) to plot the evolution of a river in the Hunter Valley, Australia. Evolution of an Intact Valley Fill reach from pre-European condition to its contemporary morphology is shown schematically in Figure 1.2C (Gregory et al., 2008). However, approaches which both quantify underlying process dynamics, map channel change and then use this as a basis for predicting future adjustment are relatively rare.

![Figure 1.2: Examples of previous evolutionary trajectory approaches. A) Shows (from top to bottom) erosion indices, deposition indices and the difference between the two as a technique for describing channel change (Hooke, 2008: 6), B) the channel evolution model from Simon and Thomas (2002: 700) shows evolution of channels in response to a knickpoint moving through the system and C) shows the evolution of a specific site derived using the River Styles Framework (Gregory et al., 2008: 233).]

These previous approaches to analysis of evolutionary trajectories represent channel adjustment through either qualitative conceptual models or quantified measures of indicators of channel form
(i.e. sinuosity, channel width). At present, description of the processes driving adjustment are rarely considered, despite recognition of the need to combine historical approaches with a process-based underpinning (Dufour and Piégay, 2009; Small and Doyle, 2012). Ziliani and Surian (in press) present the first attempt to remedy this, by combining traditional historical approaches to assessing channel change using aerial photography with CEASAR modelling, in efforts to quantify differences in bedload transport along a reach. In general, quantifying processes within evolutionary trajectories has been largely omitted due to difficulties in measuring sediment flux. However, quantifying sediment transport aids efforts to establish why a reach is evolving in the way that it is, strengthening predictions of future adjustment. To achieve this, flux relations must be tied to sediment availability, encompassing analysis of catchment-scale controls (Fryirs, 2012).

1.7 THE NESTED SPATIO-TEMPORAL HIERARCHY OF SEDIMENT FLUX

“... it appears that recognition of the scaling range of many of the most important features of flow, sedimentation, and morphology of river channels will compel a more detailed examination of almost all aspects of river processes, one consequence of which will be a radical rethinking of what are the fundamental scales, or scale limits, of river processes and forms.”

(Church, 2007: 25)

Scale has long been a key consideration of geomorphic enquiry (e.g. Schumm and Lichty, 1965). Partly this is attributable to recognition of scale dependency, whereby the scales which we use to observe a phenomenon inherently determine the patterns we observe (Church, 2007). This thesis argues for the importance of multi-scale analysis. This section briefly discusses the importance of scale and the necessity of a multi-scalar approach to riverine studies. More in-depth literature reviews of pertinent components of each specific scale can be found within the introductions of Chapters 3-6.

Rivers operate across a hierarchy of temporal and spatial scales (Figure 1.3A). The catchment scale provides a fundamental frame for understanding sediment generation, transfer and supply (Brierley and Fryirs, 2005; Chorley, 1969; Fryirs, 2012) and for grounding river management activities which aim to work with these overarching processes (Benda et al., 2011; Kondolf et al., 2006; Raven et al., 2010; Sear, 1994; Sear et al., 1995). Channel adjustment at the reach scale is driven by the supply of sediment, and internal reach scale controls (i.e. slope and channel planform) which determine the ability of the channel to transport the sediment delivered (Ferguson, 1987; Hoyle et al., 2008; Nanson and Huang, 2008, Figure 1.3B). However, to understand why a channel adjusts, smaller scales need to be considered. This allows sediment entrainment at a specific site to be related to the sedimentary make up of bar units and vegetation roughness (Corenblit et al., 2007; Mao and Surian,
2010; Surian et al., 2009a). Indeed, some studies work at the resolution of individual grains (e.g. Frey and Church, 2011 discuss bedload transport as a 'granular phenomenon') or local scale 'patches' (Hodge et al., 2009; Laronne et al., 2001). As, such adequately understanding channel behaviour is based on combining insights across scales to form a whole picture of processes driving the system (Church, 2007).

The scale of enquiry ultimately determines the types of analysis carried out and the relationships observed. Small and Doyle (2012) argue that historical landscape approaches must be merged with a local process based underpinning to gain a more rigorous understanding of system dynamics (c.f. Lane and Richards, 1997; Smith et al., 2002). Furthermore, no scale of analysis is sufficient in its own right. It is contended here that the combination of scales and types of enquiry provides the most comprehensive picture of system dynamics. Small and Doyle (2012) capture this sentiment by stating that multiple lines of evidence is the strongest way to observe a system. As such, a multi-scalar approach is necessary to combine quantitative (smaller scale analysis of sediment transport) and qualitative (larger scale landscape evolution) insights into what factors are driving change within the system.

Figure 1.3: A) The multiple temporal and spatial scales which influence sediment transport (sourced from Church, 2007) and B) multiple scales of controls which influence sediment transport (sourced from Raven et al., 2010).
Support of a multi-scale approach is not new to river research. Frissell et al. (1986) introduced the idea of river systems as nested hierarchies, whereby biota and their habitats are viewed within the context of channel form and process at progressively larger scales, from the microhabitat up to the stream system. Montgomery (1999) adapted this to a geomorphic perspective by discussing hierarchical zones dominated by similar sets of geomorphic processes. Since then, advocacy of a multi-scalar approach has been rife within the literature (Beechie et al., 2010; Brierley and Fryirs, 2005; Church, 2007; Parsons and Thoms, 2007; Reid et al., 2010; Small and Doyle, 2012; Thorp et al., 2006). Within this, the call for a catchment scale perspective to frame more finer scales of research has been particularly dominant (Benda et al., 2011; Brierley and Fryirs, 2009; Kondolf et al., 2006; Parker, 2010; Piégay and Hicks, 2005; Raven et al., 2010; Sear et al., 2009; Wohl et al., 2005). However, examples of geomorphic applications are still scarce.

1.8 RATIONALE OF THE SELECTION OF THE TONGARIRO AS THE STUDY SITE
The Tongariro catchment was selected for a number of reasons. Firstly, it has a volatile wandering cobble bed river in the lower catchment which was seen to be highly dynamic over time. This planform adjustment provided a threat to the neighbouring town of Turangi through both flooding and erosion caused by avulsion of channel migration. Therefore, it presented an ideal location to further understanding into how underlying boundary controls influence planform change. Secondly, the highly complex and unique landscape history, and types and volumes of sediment generated and transported necessitated the development of a toolkit that would assess sediment connectivity in highly diverse landscapes. This can be seen in the complexity of the upper catchment, with half being occupied by three volcanic cones and a ring plain and the other half uplifting greywacke ranges. Even at the local reach scale the Tongariro displays marked heterogeneity, with a lag of lahar boulders making up the bed material with the sediment being actively reworked by the system. It was this overwhelming diversity and complexity which made it an ideal location to develop the approach described within this PhD. A multiple lines of evidence approach was used to build up a complete picture of systems dynamics. Each type of analysis adds additional insight into how the system functions. This complexity is further enhanced by the range of influences within the catchment, including volcanic eruptions and flow regulation impacts. Finally, the systems dynamics within the Tongariro are driven by coarse sediment transport. This work is based around analysing channel form to infer bedload transport, through such approaches as using shear stress equations to calculate entrainment of bed material to the development of a morphological sediment budget. In summary, the high spatial and temporal complexity of coarse material transport across the
Tongariro catchment made it the ideal location for developing a multi-scalar approach to quantify sediment transport and predict future channel evolution.

1.9 FUNDAMENTAL RESEARCH QUESTIONS
This thesis aims to extend the evolutionary trajectory approach to geomorphic analysis of rivers by incorporating insights across multiple temporal and spatial scales. This includes establishing a process-based underpinning to enhance understanding of why the channel is adjusting in the way and at the rate that it does. This framework will be applied in the lower Tongariro River, located in the central North Island of New Zealand. The lower course of this river is primarily characterised as a wandering gravel bed river which adjusts its form in response to changes in sediment flux and flow, based on the combination of boundary conditions of a given sub-reach (i.e. confinement). Anthropogenic development including flow regulation due to dams has altered patterns of sediment flux. Sediment fluxes are analysed across multiple spatial and temporal scales: catchment scale sediment fluxes are analysed over the last 1800 years, reach scale adjustment over the past 80 years and bar reworking is assessed in relation to flood events with recurrence intervals from 2.33-100 years. This information base is used to construct multi-scalar, process-based evolutionary trajectories that capture past channel adjustments as a basis to suggest how the river might adjust into the future. The specific research questions are:

1. How and why do larger-scale spatial and longer-term temporal patterns of sediment delivery and connectivity across the Tongariro catchment provide an important framing for geomorphic adjustments of the lower Tongariro River?
2. Why does reach scale response differ along the lower Tongariro River and what are the key controls driving patterns of sensitivity?
3. Why are the patterns and frequencies of bar reworking along the lower Tongariro River an important underpinning to channel adjustment at the reach scale?
4. How can multi-scalar process-based evolutionary trajectories be used to provide a geomorphic platform for river management?

1.10 THESIS OUTLINE
The outline of the thesis is presented in Figure 1.4. The context for the thesis is presented through an Introduction chapter which describes key themes and literature that runs throughout the thesis (Chapter 1). More specific literature reviews can be found for each results chapter (Chapters 3-6). A Regional Setting chapter describes the characteristics of the Tongariro catchment (Chapter 2). Four results chapters (Chapters 3, 4, 5 and 6) analyse fluvial processes and channel response within the
Tongariro catchment at progressively more local scales (from catchment to bar). A multi-scalar discussion chapter uses insights from Chapters 3-6 to create evolutionary trajectories (Chapter 7). Finally, a Discussion and Conclusion chapter overviews the contribution of this thesis (Chapter 8). Results chapters are presented as discrete bodies of work and therefore include literature reviews, methods and discussion sections which are relevant to that study. As each ‘results’ chapter has specific aims and questions which are answered internally an overall methods chapter is omitted from this thesis.
1.10.1 Chapter 1 – Introduction
This chapter sets up the problem statement which provides the rationale for this study.

1.10.2 Chapter 2 – Regional Setting
This chapter describes the characteristics of the Tongariro catchment. This includes sections on the geology, hydrology, climate, topography, vegetation and land use history. This contextualises work presented in the results chapters.

1.10.3 Chapter 3 – Longer-term Spatio-temporal Patterns of Sediment Fluxes in the Tongariro Catchment
The first results chapter addresses the first research question, analysing the movement of sediment at the catchment scale and across larger timeframes, to provide a context for channel adjustment at the reach-scale.

The catchment is separated into Landscape Units, characterised by similar sets of processes. The distribution of River Styles across the catchment is classified. A catchment scale DEM is used to assess lateral connectivity, defined as the ability of landscape units to generate and deliver sediment from the hillslopes to the channel. This incorporates measures of slope, an erosion index (function of drainage area and slope), erosion terrains and specific suspended sediment load estimates. Longitudinal connectivity is assessed by modelling slope and stream power along drainage lines. The character and behaviour of representative sites across the catchment are surveyed in greater detail, to quantify (through shear stress equations) and ground insights into sediment transport. This section describes the supply, storage and transfer of sediment within the contemporary catchment.

The second section of this chapter deals with the long-term evolution of the Tongariro catchment. Lake Taupo is a volcanic caldera into which the Tongariro River drains. The most recent eruption of 1.8 ka. reset the process zones within the catchment, setting the current trajectory of geomorphic adjustment. Along the middle-lower catchment, volcanic deposits have been incised to create terraces and a delta has prograded beyond the terrace extent. High resolution LiDAR data are used to quantify bulk sediment volumes and construct a catchment-scale sediment budget for the past 1850 years. Spatial and temporal patterns of sediment transfer within the Tongariro catchment set the context for reach scale adjustments.

1.10.4 Chapter 4 – Reach-scale Geomorphic Adjustments over the last 80 years
This chapter addresses the second research question through analysis of planform adjustments of the lower Tongariro River over the past 80 years.
Chapter 1: Introduction

Channel form is digitised from aerial photographs and a survey map spanning the period from 1928-2007. This is used to assess controls upon the spatial patterns of adjustment and reach scale sensitivity (the rate and types of adjustment). Analysis of sensitivity is related to the natural range of variability for differing types of river.

1.10.5 Chapter 5- Controls on the Planform Adjustment of the Wandering Cobble Bed Reach

The extent of channel adjustment demonstrated in Chapter 4 was found to vary markedly for reaches of the Wandering cobble bed river. This chapter outlines process-based understandings into reach-scale variability. This is achieved by differentiating the reach into ‘wide sedimentation’ zones characterised by active sediment storage and ‘narrow’ zones with few active stores. Changes in reach scale controls such as slope, bed material, valley confinement and transport capacity (a measure of the ability of the reach to entrain the sediment present at a site) are compared for 19 sites that fall within these zones. This analysis explains why some reaches are more geomorphically sensitive to flood events than others.

1.10.6 Chapter 6- A Novel Approach to Mapping the Bed Material Distribution across Bar Surfaces and Analysing the Frequency of Reworking

This chapter presents an approach to analysis of bar reworking, addressing the third research question. A Terrestrial Laser Scanner (TLS) was used to survey four bars at high resolution. These bars were selected to represent downstream changes in slope, sediment size and confinement. Laser returns are used to map the distribution of median sediment size across the surface of each bar at 1 m resolution. Flow depths for different sized flood events (recurrence intervals between 2.33 and 100 years) were interpolated across each bar. This enabled shear stresses to be calculated and the likelihood of entrainment assessed given the ratio of shear stress to critical shear stress. This method provided a technique to compare the geomorphic effectiveness of different magnitude flood events and assess the likelihood of entrainment of within bar units for each flood event. This provided a local scale process based rationale which explained reach-scale channel adjustment.

1.10.7 Chapter 7- Process-based Evolution Trajectories for the Tongariro River

The first discussion chapter addresses the fourth research question. Process-based evolutionary trajectories are constructed for each reach. Trajectories are created using insights across multiple scales, building upon results from Chapters 3-6. Analysis of the landscape setting and channel location evolution over the past 1800 years provides a template to construct future response
trajectories, whereby the magnitude and type of response to a range of drivers are predicted based on past patterns of adjustment.

1.10.8 Chapter 8- Discussion and Conclusion
The final chapter draws together themes that run through the thesis, framing the contribution of the thesis in relation to literature. Management implications that work with the underlying fluvial processes and their evolutionary trajectory are suggested for each reach. While this section does not aim to provide a complete management plan, issues and applications considered to underpin process-based management strategies are outlined. A conclusion summarises the key themes and findings from the thesis.
2 REGIONAL SETTING FOR THE TONGARIRO CATCHMENT

2.1 INTRODUCTION
The Tongariro catchment is located in the Central North Island of New Zealand. It encompasses an area of 654 km$^2$, with the headwaters draining both an active volcanic zone and deeply dissected uplifting greywacke ranges (Figure 2.1). The Tongariro River flows into Lake Taupo, a rhyolitic caldera which was formed by a catastrophic eruption 1800 years ago (Wilson and Walker, 1985). At 47 km long, this river is the longest tributary with the largest catchment area draining into Lake Taupo (Collier, 2002).

This chapter describes the geology, climate, hydrology, topography, flora and fauna and catchment land use history of the Tongariro catchment. Primary influences upon the contemporary river include hydro-power dams in the upper catchment, volcanic influences including the deformation of Lake Taupo, flood works and gravel mining in the lower river and regulation of the levels of Lake Taupo. The chapter begins with a brief summary of the character of the catchment.

2.2 SUMMARY OF CATCHMENT CHARACTERISTICS
The Tongariro catchment is located at latitude 39° south, between elevations of 329 and 2797 m above mean sea level. The catchment can be separated into three broad sections based on topography, vegetation and geology. The western side of the catchment drains the Taupo Volcanic Zone. Parallel streams flow from the active volcanic cones of Mt Ngauruhoe (2287 m asl), Mt Tongariro (1981 m asl) and Mt Ruapehu (2797 m asl) across the flat featureless expanse of the Volcanic Plateau, formed through accumulation of ash and lahar deposits. All volcanoes are active and have erupted in the past 100 years, with the most recent eruption being Mt Tongariro in 2012 and the more prolonged eruption of Mt Ruapehu in 1995 - 1996. The lithology of this area is primarily made up of andesite with scoria and ash deposits (Collier, 2002). The steep high-altitude slopes provide an unstable and harsh environment that is unable to sustain vegetation other than mosses. However, vegetation on the flatter, lower-altitude central plateau comprises a wide range of alpine desert fauna, including some rare indigenous vegetation. The western portion of the upper catchment drains the Kaimanawa Ranges. This area is underlain by uplifting Torlesse greywacke (Genesis Energy, 2009). Rivers have cut easily into this lithology, creating a dissected, dendritic drainage pattern. The ranges remain within a forest park and have retained natural vegetation cover of native beech and Podocarp forest. The lower Tongariro, which is the main section of interest for this study, flows within terraces formed by channel incision into tephra and lahar deposits. This consists of a stretch of wandering gravel bed river, which flows into a delta at the lake margin. Land
use in this lower section of the catchment is more developed, with plantations of *Pinus Radiata* and low density sheep farming. The town of Turangi (population 3240, Statistics New Zealand, 2010) is located at the downstream extent of the terraces. The delta region includes one of the largest remaining freshwater wetlands in New Zealand (Chagué-Goff and Rosen, 2001).

![Digital Elevation Map (DEM) of the Tongariro Catchment, with important locations identified.](image)

**Figure 2.1**: Digital Elevation Map (DEM) of the Tongariro Catchment, with important locations identified.

### 2.3 CATCHMENT TOPOGRAPHY AND LANDSCAPE UNITS

The topography of the Tongariro catchment is described in detail in Chapter 3 thus only briefly summarised here. This section discusses the elevation of the Tongariro catchment, distribution of stream orders, classification of landscape units and analysis of catchment morphometrics.

The Tongariro catchment ranges from 2720 m asl at its highest point at Mt Ruapehu to 357 m asl where it drains into Lake Taupo (Figure 2.1). Other high elevation areas include Mt Ngauruhoe (2270 m asl), Mt Tongariro (1978 m asl) and the peaks of the Kaimanawa Ranges (up to 1645 m asl). The
Kaimanawa Ranges are characterised by steep dissected terrain. In contrast, the Volcanic Plateau provides a flat featureless expanse directly downstream of the volcanic cones. The mid-catchment is similarly flat and featureless and drains into the low elevation delta. Long profiles are contained with Chapter 3, as they form a key component of the catchment scale connectivity analysis (see Figure 3.13).

Drainage basin characteristics can be assessed using Stream Order (Figure 2.2), which highlights the differences between the sub-catchments. In the eastern sub-catchment, which includes the Kaimanawa Ranges, first order streams are numerous and very short, usually between 0.5 – 1 km long. These rapidly drain into short second and then third order streams, which drain the smaller valleys of the Kaimanawa Ranges into the forth order trunk streams of the Waipakihi and Whitikau Streams. This forms a pronounced dendritic drainage pattern.

In contrast, first order streams in the western, volcanic catchment are longer (up to 5 km) and drain into longer second and third order streams forming a parallel drainage pattern. Fourth order streams...
are shorter than and not as large as the Whitikau and Waipakihi Streams draining the Kaimanawa Ranges. The Tongariro starts as the fifth order stream for a short stretch of its headwaters, becoming a sixth order stream for most of its length.

The catchment is separated into ‘landscape units’ with similar characteristics (i.e. geology, climate, topography and land use) that contain similar types of rivers (Brierley and Fryirs, 2005).

![Map of Tongariro catchment with landscape units](image)

**Figure 2.3: Distribution of the Landscape Units across the Tongariro catchment.**

The Tongariro catchment can be separated into seven landscape units as shown in Figure 2.3. Photographs of each landscape unit are shown in Figure 3.4, in the following chapter.

- **Volcanic Uplands** - This includes the Ruapehu, Ngauruhoe and Pihanga Volcanic cones. Terrain is steep with many rivers within poorly formed gully complexes. Vegetation is sparse due to the cold climate and extreme conditions. Only alpine lichens and mosses and lower down tussock are able to occupy this area. Sediment comprises unconsolidated, volcanic material which has a high component of volcanic ash and sand deposited around andesitic lava flows. Streams are mainly ephemeral and have a seasonally influenced flow regime, due to increased flow from snow melt.
• **Volcanic Plateau**- This consists of the flat area downstream of the andesitic cones of Mt Ruapehu and Mt Ngauruhoe. Underlying geology mostly consists of lahar deposits and includes a wide range of poorly sorted, volcanic material. Slope is gentle with channels flowing in shallow, confined gullies. This is a volcanic desert and vegetation consists of tussock with some small patches of beech, and native shrubs such as Kanuka.

• **Steep Headwaters**- These include the Kaimanawa Ranges and consists of steep confined valleys and rugged terrain. Vegetation is predominately native, comprising beech and podocarp forest within the Kaimanawa Forest Park. Geology is greywacke, with a line of schist in the Southern reaches.

• **Tongariro Trunk**- This includes the area where the Tongariro River flows through the mid-catchment. The channel is entrenched between the uplifting Kaimanawa Ranges and the Volcanic Plateau. The terrace confined channel is steep, characterised by high energy and boulder sediment. Intermittent gorges have formed where the channel has cut down into andesitic lava flows. These act as base-level controls.

• **Rolling Foothills**- This comprises a zone of rounded, weathered hills on the margins of the Kaimanawa Ranges in the lower, eastern catchment. Geology is greywacke with minimal uplift. Vegetation is mostly low intensity pasture and pine plantations.

• **Terraceland**- This flat featureless floodplain section was covered by Taupo eruption deposits comprising mainly unwelded ignimbrite. The channel has incised into this material and is now largely disconnected from the floodplain.

• **Lowland Plain/delta**- The most downstream section of the catchment consists of a lowland alluvial plain, characterised by a very low slope. Underlying geology is pumice pyroclastic materials combined with alluvial deposits. Scattered pastoral farming is the most common land use in this area, and the town of Turangi is situated within this unit. The most downstream zone contains a delta, which has a wetland ecosystem of high to outstanding importance. Floodplains are continuous, but stopbanks and willows restrict lateral channel migration.

### 2.3.1 Catchment Morphometrics

Whilst the Tongariro catchment is considered small on a global scale, it is of moderate size within New Zealand with a catchment area of 654 km² and a perimeter of 191 km (Table 2.1). This contains approximately 1123 km of drainage lines. The total relief is 2345 m which includes the upper slopes of Mt Ruapehu at an elevation of 2720 m down to the edge of Lake Taupo at 357 asl. This change in slope occurs over a relatively short basin length of 41 km, resulting in a steep relief ratio of 0.057.
Chapter 2: Regional Setting

The Hypsometric integral also reflects this, with a high value of 89 describing a tectonically active and dynamic landscape.

Morphometric indices also describe basin shape. The Tongariro has an elongation ratio of 0.62, which describes a more elongate catchment as prescribed for a value of 0.6 (Table 2.1). In theory runoff would be expected to take longer time to travel from the headwaters to the mouth than a more rounded catchment with a ratio of about 1. The drainage density of 1.72 describes 1.72 km of drainage line per 1 km$^2$. This moderate value indicates that the catchment has a relatively dense network of drainage lines, providing minimal limitations to the delivery of sediment to river channels. The Tongariro has a form factor of 0.38, indicative of a moderate ratio of basin length to catchment area.

<table>
<thead>
<tr>
<th>Drainage Basin Morphometrics</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>General</strong></td>
<td></td>
</tr>
<tr>
<td>Area (km$^2$)</td>
<td>654</td>
</tr>
<tr>
<td>Perimeter (km)</td>
<td>191</td>
</tr>
<tr>
<td>Total Channel Length (km)</td>
<td>1123</td>
</tr>
<tr>
<td>Basin Length (km)</td>
<td>41.3</td>
</tr>
<tr>
<td><strong>Elevation</strong></td>
<td></td>
</tr>
<tr>
<td>Relief (m)</td>
<td>2345</td>
</tr>
<tr>
<td>Relief ratio (relief / length)</td>
<td>0.057</td>
</tr>
<tr>
<td>Hypsometric Integral</td>
<td>89.11</td>
</tr>
<tr>
<td><strong>Shape</strong></td>
<td></td>
</tr>
<tr>
<td>Elongation Ratio ($A^2$/basin L)</td>
<td>0.62</td>
</tr>
<tr>
<td>Drainage Density (Channel length / area)</td>
<td>1.72</td>
</tr>
<tr>
<td>Form factor ($A/L^2$)</td>
<td>0.38</td>
</tr>
</tbody>
</table>

2.4 GEOLOGY

This section describes the tectonic setting of the Tongariro catchment, the underlying geology and the imprint of volcanic eruptions and tectonic deformation.

2.4.1 The Taupo Volcanic Zone

The Tongariro catchment is located in the middle of the Taupo volcanic zone (Figure 2.4). The Hikurangi Trench to the east of the North Island marks the subduction zone where the Pacific Plate has been forced obliquely westward under the Australian Plate over the past 20 – 25 Myr (Litchfield et al., 2007). This has created a volcanic zone 300 km long and 60 km wide which includes rhyolitic calderas and andesitic cones (Wilson et al., 1995). This region has been volcanically active since 2 Ma. and is estimated to have produced more than 10 000 km$^3$ of magma over this time (Newnham
et al., 1999). This complex legacy of volcanism and uplift has shaped the nature of sediment flux within the Tongariro catchment.

2.4.2 Distribution of Rock Types across the Tongariro catchment

The Tongariro catchment can be separated into three distinct lithological areas (Figure 2.5).

2.4.2.1 Upper Eastern Catchment

The upper section of the catchment that drains the Kaimanawa Ranges (including the Waipakihi and the upper Tongariro River) drains the Torlesse group greywacke. This area is made up of relatively hard grey sandstone, with sections of argillite (sedimentary rock derived from fine grained clay particle) (Figure 2.5A). Small localised zones of lava, limestone, chert and spilitic tuff can also be found within this formation (GNS, 2009). Greywacke gravels are supplied as bed material to streams.
draining this section. A narrow section of Pelitic schist, with sections of deformed conglomerate covers the southern tributaries to the Waipakihi Stream. Ash forms overlying layers within this subcatchment. Some ash deposits that predate the Taupo eruption (> 1.8 ka) are located on some of the headlands, while reworked pumice and breccia constitute a significant proportion of the alluvial deposits within the active channel of the Waipakihi River (Figure 2.5B).

2.4.2.2 Upper Western Catchment
The very steep volcanic sections draining the west of the catchment are underlain by andesitic material from volcanic flows which is present as lavas and welded ignimbrites on the steepest flanks of the volcanoes, including Mt. Pihanga further to the south (Figure 2.6). Sections of agglomerate (most commonly from volcanic bombs) and breccias are scattered across this layer. Lahar deposits underlie the locations of contemporary river channels, whilst the remaining portion of the central plateau comprises volcanic ash material (Figure 2.5A). Most of this sub-catchment is overlain by ash from Mt. Ngauruhoe, the volcano which has been most active in recent times (Genesis Energy, 2000).

2.4.2.3 Lower Catchment
The lower section of the Tongariro River is underlain by pumice pyroclastic materials and breccia which was deposited during the Taupo (1.8 ka) and Kaharoa eruptions (27 ka) (Figure 2.5A). Sections to the east of the catchment comprise Quaternary breccias that predate the Taupo deposits (> 1.8 ka). Ashes from the Taupo and Kaharoa eruptions can be found overlying the steeper sections of the lower catchment including Mt. Pihanga and the eastern foothills (Figure 2.5B).
Chapter 2: Regional Setting

Figure 2.5: Geology maps of the Tongariro catchment. Due to the complex nature of geology within the Tongariro catchment this is presented as the A) Top-rock, which details the overlying layer and B) Base-rock describing the main basement rock underlying this. Data sourced from the New Zealand Land Resource Inventory (Landcare Research NZ Ltd, 2000). 1 – 2 shows the location of the cross-section described in Figure 2.6.
2.4.3 Volcanic Influence

The Tongariro River is located in an active volcanic zone, which has acted to shape the evolution of the catchment. Lake Taupo, the outlet for the Tongariro River is a rhyolitic caldera, which erupted most recently 1.8 ka B.P. ejecting around 60 km$^3$ of ultraplinian pumice and ignimbrite into the atmosphere (Wilson et al., 1980). This tephra fallout blanketed the landscape up to 10 m thick, killing virtually all vegetation and sterilising soils. Drainage lines incised into these deposits, with the lower Tongariro River creating terraces. A delta has formed downstream of the terrace extent.

Volcanic activity is a major control on contemporary bed material. Previously, the catchment was more volcanically active. Fingers of andesitic bedrock from past lava flows protrude across the Tongariro River into the Kaimanawa Ranges. The river has eroded into this bedrock to form gorge sections including Tree trunk gorge, Waikato Falls and the above gorge and the Pillars of Hercules. Due to their confined nature, gorges concentrate flow energy and flush sediment delivered. Gorges also control base-level at distinct points along the longitudinal profile.

The Tongariro River is a conduit for lahar material, with earliest deposits dated back to 14.7 ka (Cronin et al., 1997). Figure 2.7 presents a lahar hazard map, showing the likelihood of surfaces to be inundated by lahars based on the ages of lahar surface deposits. These rapidly flowing mixes of rock debris and water transport larger material than is generally moved by fluvial processes. Lahars deposit unconsolidated sediment, characterised by a wide range in sizes. The mid-catchment of the Tongariro River that is confined by terraces is underlain by these deposits and a boulder lag lines the bed of the channel. These lag materials provide an armour layer that slows the rate of adjustment, especially vertical change within the system (see Chapter 4).
The volcanic cones at the headwaters of the Tongariro are considered active. Mt Ruapehu is the most frequently active, erupting on average every 1 – 3 years with hazardous eruptions on average every 7 – 10 years (Department of Conservation, 2006). Mt Ngauruhoe also has an active history. It last erupted in 1975. Mt Tongariro underwent a brief, minimal eruption in 2012 with little impact on the Tongariro catchment. The most recent substantive eruption was Mt Ruapehu in 1995 to 1996.
This delivered around 6900 kilotonnes of volcanic sediment to the catchment upstream of the Rangipo dam, with two thirds consisting of fine grained sediment (< 0.5mm) (Collier, 2002; Manville et al., 1996). This material may infill pools, decreasing habitat quality (Jowett, 1980). However, most of this material has been moved through the catchment, deposited on and around the delta (Genesis Energy, 2000). The volcanoes also create a seasonal pattern in sediment load, as increased volumes of volcanic sediment are moved downstream during summer, as a result of snow melt from Mt. Ruapehu (Genesis Energy, 2000).

2.4.4 The Distribution of Grain Size across the Tongariro Catchment

Detailed analysis of grain size is presented in Chapter 3 to support the connectivity analysis (Figure 3.5 and Table 3.3). As such, this section only provides a brief overview of how the sediment calibre and volume changes across sections of the catchment.

The uplifting Kaimanawa Ranges in the eastern sub-catchment generates high loads of greywacke gravels. Median grain size ranges from 120 mm in the steep upper, first order streams with a median grain size of 63 mm draining the larger fourth order streams into the Tongariro River. Large active bar complexes (up to 100 m wide) which are regularly reworked indicate high delivery of sediment into the Tongariro River from this sub-catchment.

The western sub-catchment includes the active volcanic cones of Mt Ruapehu, Mt Ngauruhoe and Mt Tongariro. These are composed of andesitic bedrock which is slow to degrade. However, much of these cones are composed of scoria and scree mixed with volcanic ash and sand. Streams in this section of the catchment have a small grain size (~ 3 mm median) and this represents high loadings of sand and suspended sediment which are flushed through the catchment. Streams draining the volcanic plateau have incised to form gullies with coarse bed material, with the median grain size ranging from 90 – 210 mm. Active stores are minimal within this reach, with only small, infrequently reworked bar features present, indicative of lower sediment delivery into the Tongariro River.

The Tongariro River flows across cobbles in the mid catchment (Figure 3.5). Average median grain size in this area is 149 mm. The channel flows within terraces and accommodation space to store sediment is minimal within this confined setting. Within the lower Tongariro, once these terraces widen, median grain size ranges from 90 to 230 mm. Once the channel exits the terraces and flow across the delta, grain size decreases rapidly, from cobble (85 mm) to sand at the most downstream point (1 mm). This delta has been built up of high loads of volcanic ash and fluvial sediment over the past 1850 years (Smart, 1999).
Several past studies have attempted to quantify volumes of sediment moving through the catchment. Bedload capacity has been estimated at 11.7 kilotonnes/year and suspended sediment 141 kilotonnes/year at Turangi based on discharge data from 1987-1990 (Smart, 1992). This highlights the steep nature of the lower river, and its capacity to transport vast volumes of sediment. Volcanic eruptions and floods provide pulses of sediment into the lower catchment, where it is flushed and deposited on the delta.

2.4.5 Tectonic Deformation

Tectonic deformation alters the height of different surfaces within the Tongariro catchment. Following the Taupo eruption (1.8 ka) the outlet to Lake Taupo was blocked by eruption deposits. This caused the lake level to rise 30 m higher than the current level, creating a wave cut lake margin. This provides a proxy for deformation over the past 1800 years, as relative changes in vertical deformation can be observed when compared with a survey datum. Figure 2.8 presents a map of the relative deformation indicating that the Tongariro delta has subsided at a rate of between 1 and 2 mm a year, with total subsidence of between 1.8 to 3.6 m in the last 1800 years. However, these data have to be treated as indicative as the tectonically active nature of the catchment precludes determination of a static survey datum over the past 1800 years. Therefore, rather than treating these results as evidence of subsidence, they are more indicative that the northern lake margin is now higher than the southern.

Detailed short-term data of lake levels have been collected for Lake Taupo over the past 40 years. The water surface of Lake Taupo is used as a relative spirit level, with changes to bench marks around the lake compared to assess changes in height. The closest gauge to Turangi is the Waihi (WH) gauge located 2 km west of the delta. This gauge indicates slow overall subsidence of 0.4 mm/year from 1984 to 2000 (Otway et al., 2002). Between 1999 – 2001 subsidence was more rapid, at 20 mm/y as the lake tilted to the south west (Figure 2.9A). This ceased in 2001 and even reversed between 2002 and 2003 before subsiding again between 2004 – 2007 (Figure 2.10). This illustrates the non-linear nature of tectonic deformation within Lake Taupo.
Figure 2.8: Long term tectonic deformation of Lake Taupo generated by comparing the location of a wave cut terrace from when the water level was 30 m above present to a survey datum location at the Taupo control gates (sourced from Hancox, 2002).

The northern lake margin was raised by 55 mm in a year following earthquakes between 1983 – 1984 (Otway et al., 2002). Figure 2.9B compares the elevation of Acacia Bay (northern margin) with Tokaanu (2 km west of the Tongariro delta to the south). The 1984 earthquakes created slow subsidence to the south of the lake which was then subsequently reversed (Otway et al., 2002). As the gauge used to regulate Lake Taupo’s water levels is located to the north, close to Acacia Bay, an increase in height in this location could affect water levels in the Tongariro delta.
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Figure 2.9: Tectonic deformation of Lake Taupo at Waihi Gauge A) (2 km west of the Tongariro delta). Grey bands indicate time periods of increased tectonic activity (sourced from Peltier et al., 2009). B) compares deformation between the northern lake margin at Acacia Bay and the southern at Tokaanu (2 km west of the Tongariro delta) following a swarm of earthquakes in 1983 – 84 (sourced from Otway et al., 2002).

Overall patterns of deformation between 1950 and 1986 are shown in Figure 2.11. Over recent times the delta has lowered at approximately the same speed or a little slower than the northern lake margin, specifically Taupo. Control gates at Taupo set the height of the lake, which determines the inundation of the delta. These findings indicate that as the whole trough is subsiding lake level is likely to be decreasing in sync with the delta subsidence. In summary, the tectonic deformation of Lake Taupo has exhibited high diversity in spatial and temporal patterns of adjustment over the past 1800 years.
Figure 2.10: Rates of deformation around Lake Taupo. Maps describe eight separate periods characterised by similar patterns and rates of adjustment (Peltier et al., 2009).

Figure 2.11: Vertical adjustment in the Taupo region (1950 – 1986) (in mm/yr). The rate at which the delta has been subsiding in the recent past is similar to the rate for Taupo, the locality of the lake outlet. This sets the lake level height (based on data from Blick and Otway, 1995; sourced from Hancox, 2002).
2.5 CLIMATE

The climate of the Tongariro catchment may be defined as alpine in the headwaters, grading to more temperate conditions at lower elevations. Mean annual rainfall is as high as 3400 mm within the upper catchment decreasing to 1200 mm at the delta (Genesis Energy, 2009). Total monthly average rainfall is lowest in February (109 mm) and peaks at July (180 mm), with the average yearly rainfall being 1832 mm at Taupo Trout Hatchery in the mid-catchment (Figure 2.12A). Mean monthly temperatures range from 6.5°C in July to 17.3°C in February, with a yearly average of 11.8°C (Figure 3b).

![Figure 2.12: A) Mean monthly rainfall at Tongariro Trout Hatchery and B) mean monthly temperatures at Turangi with mean maximums and mean minimums (°C). Both datasets are based on 1971-2000 data (NIWA, 2009).](image)

Figure 2.12A shows the average annual temperature across the catchment. This illustrates the alpine nature of much of the headwaters including the Volcanic Plateau and the Kaimanawa Ranges (average temperatures of 6 – 8°C), especially the peaks of the volcanic cones (average annual temperatures of 0 - 2°C). The distribution of the average monthly water balance ratio across the catchment is presented. This describes the ratio of rainfall to potential evaporation, with higher values indicating a higher deficit of water. The headwaters and especially the volcanic cones are
much wetter, with cooler temperatures that limit evaporation giving them ratios of 5 - 10. This drives a flashier flood regime with soils more likely to be saturated. In the mid-catchment and especially the delta the water balance ratio is much lower, illustrating that these areas are able to evaporate much of the rainfall delivered. As such, drier soils are likely to slow the movement of water in this region.

Figure 2.13: Maps of the Tongariro catchment showing the distribution of A) average annual temperature and B) the average monthly water balance ratio derived using a water balance model which compares the ratio of rainfall to potential evaporation (calculated using temperature and solar radiation data) (sourced from Landcare Research NZ Ltd, 2000).

2.5.1 The Last Glacial Maximum

Periods of glacial advance stretch back to 75 ka, with the Last Glacial Maximum (LGM) marking the final stage of the last ice age, from which the climate has gradually warmed (Figure 2.14). In New Zealand the LGM is believed to have occurred between 26 - 18 ka B.P. and was characterised by temperatures between 3 and no more than 5°C cooler than present (McGlone and Topping, 1983; Salinger, 2001). Whilst much of the South Island was under glaciers at this time, much less of the North Island was likely to have undergone glaciation (Figure 2.15). However, as shown in a map by Newnham et al. (1999) only the volcanic cones of Mt Ruapehu and Mt Ngauruhoe were covered by glaciers, with an ice cap centred on the summit of Mt Ruapehu (Figure 2.15). This is supported through evidence of terminal moraines on the valley floor (Nolan, 2008). However, due to the localised nature of these ice fields, the volumes of sediment generated from glaciation are not a primary sediment source as seen in cooler regions of the Northern Hemisphere (c.f. Church and Slaymaker, 1989), and are minor compared to volumes of sediment generated by the volcanoes.
Figure 2.14: Marine isotope record for the past 200,000 years. Higher values indicate warmer temperatures (sourced from Sloss, 2005).

The combination of colder periods and active volcanism are associated with higher supply and transport of sediment. The volcanic ring-plain in the central plateau is formed through aggradation and provides a record of sediment generation with an estimated 110 km$^3$ of volcanic material stored (Hackett and Houghton, 1989). Cronin and Neall (1997) describe a period of high sediment productivity between 65 – 75 ka, whereby a cooler period coincided with an increase in volcanism, increasing sediment supply. During the most recent period of ice melt, around 15 ka, multiple lahars delivered high sediment loads to the ring-plain. This period is associated with explosive andesitic eruptions combined with glacial melt water. As the glaciers retreated they provided water enabling large lahars to flush sediment across the plateau, through the Whangaehu and Mangatoetoenui valleys and into the Tongariro River (Cronin and Neall, 1997). The most recent period since 9.7 ka may be characterised by low volume and magnitude eruptions, and less glacial melt represented a decrease in sediment supplied compared with before 10 ka. The climate signature within this setting plays an important role on sediment generation when combined with patterns of volcanic activity.
2.6 HYDROLOGY AND THE TONGARIRO POWER DEVELOPMENT SCHEME

The section describes general flow characteristics of the Tongariro River, how the Tongariro Power Development Scheme (TPDS) has altered hydrology and lastly, an analysis of flood magnitude and flood history.
2.6.1 Flow Characteristics of the Tongariro River

The Tongariro River has a mean annual flow of 398 m$^3$s$^{-1}$ (Table 2.2). The highest mean annual flow recorded was observed in 1962, with an average discharge of 76 m$^3$s$^{-1}$. The lowest average flow occurred during 2005 and 2007 with a discharge of 27 m$^3$s$^{-1}$. The Tongariro River has a coefficient of variability of 0.29 which places it around the mean of rivers located within temperate climates (McMahon, 1982).

Monthly mean flows indicate higher discharges during the winter months, as consistent with rainfall patterns (Figure 2.12 and Figure 2.16). The month of lowest flow is April, with an average discharge of 33 m$^3$s$^{-1}$ and noticeably lower variability around this mean. Mean monthly discharge increases in July to a flow of 42 m$^3$s$^{-1}$ and remains at this level during winter until October. Mean discharge steadily decreases across the summer months from 39 m$^3$s$^{-1}$ in December to 34 m$^3$s$^{-1}$ in May. This indicates that snow melt has minimal impact on monthly discharge, indicating a rainfall dominated regime.

Table 2.2: Flow statistics for the Tongariro River derived using hourly discharge data from the Tongariro at Turangi gauging Station (1957 - 2009).

<table>
<thead>
<tr>
<th>Characteristic</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>Average Mean Annual Flow (m$^3$s$^{-1}$)</td>
<td>39</td>
</tr>
<tr>
<td>Standard Deviation of Mean Annual Flow (m$^3$s$^{-1}$)</td>
<td>11</td>
</tr>
<tr>
<td>Max Mean Annual Flow (m$^3$s$^{-1}$)</td>
<td>76 in 1962</td>
</tr>
<tr>
<td>Minimum Mean Annual Flow (m$^3$s$^{-1}$)</td>
<td>27 in 2005 and 2007</td>
</tr>
<tr>
<td>Coefficient of Variability</td>
<td>0.29</td>
</tr>
<tr>
<td>Drainage area (km$^2$)</td>
<td>777</td>
</tr>
</tbody>
</table>

Figure 2.16: Monthly mean flows (1957-2009) at the Tongariro River at Turangi gauging station based on hourly data. Dashed lines show the distribution of the standard deviation around the mean.
2.6.2 The Influence of the TPDS on the Hydrology of the Lower Tongariro River

The hydrology of the Tongariro catchment has been altered by the development of the Tongariro Power Development Scheme (TPDS). As such, discussing contemporary flows needs to be placed in the context of this regulation. Two hydropower dams were constructed in the mid-catchment, with the Rangipo Dam located upstream of the Poutu Intake Dam (see Figure 2.1 for locations) (Genesis Energy, 2009). These structures are relatively small on a global scale, spanning valley widths of 111 m and 93 m, and creating settling ponds that directly influence channel lengths of 600 m and 300 m upstream respectively. The water from this scheme currently provides 3.5% of New Zealand’s electricity, making the station of national importance.

The Poutu Dam, the furthest downstream, was constructed in 1973 and resulted in a 40% decrease in mean flow volume in the lower reaches (Hindle, 1995). The Rangipo Dam is located in the upper catchment and was completed in 1983. Minimum flows are required to be 0.6 m$^3$s$^{-1}$ and 16 m$^3$s$^{-1}$ downstream of the Rangipo and Poutu Dams respectively (Figure 2.17). Representative records of annual flow variability of the Tongariro River at Turangi pre- and post-flow regulation are shown in Figure 2.18. Mean flows are 1.3 and 17 m$^3$s$^{-1}$, compared with natural mean flows pre-regulation of 16.1 and 33.9 m$^3$s$^{-1}$ (i.e. representing decreases of 8% and 50% for the Rangipo and Poutu Dams respectively) (Genesis Energy, 2000).

![Figure 2.17: Schematic diagram of diversion of Tongariro River water under the TPDS (adapted from Hindle, 1995: 47)](image)
Small flood events (freshes between 70-100 m$^3$s$^{-1}$) have decreased in frequency following regulation (Hindle, 1995). Pre-regulation floods were of longer duration, with gently sloped falling stages, illustrating greater geomorphic effectiveness (Costa and O’Connor, 1995). Number of flood days a year (greater than 100 m$^3$s$^{-1}$) were found to decrease following regulation, increasing again from 1993 as the managed flow rules were updated, though not to the same levels observed pre-regulation (Figure 2.19).

Figure 2.18: Tongariro River discharge and total daily rainfall at Turangi, A) 1969 and B) 2002. This illustrates the decrease in small floods (< 100 m$^3$s$^{-1}$) and lags following flood events.
The TPDS has altered the recurrence interval of small, more frequent floods. Figure 2.20 shows the decrease in discharge for floods of return intervals of less than 5 years. However, a 1440 m$^3$s$^{-1}$ flood in 2004 caused recurrence intervals to be recalculated, due to two floods occurring in 50 years which were previously calculated to have a 100 year recurrence interval. Updated values are listed in Table 2.3. This shows that a flood with a 100 year recurrence interval has increased from a discharge of 1400 m$^3$s$^{-1}$ to 1500 m$^3$s$^{-1}$.
Table 2.3: Return periods for flood events. This shows those predicted in 2003 and revised peak discharges following a 1400 m$^3$s$^{-1}$ flood in 2004. Data were derived using a Gumbel Type 1 Extreme Value distribution (Tonkin and Taylor, 2003).

<table>
<thead>
<tr>
<th>Return Period (years)</th>
<th>Annual Exceedence Probability (%)</th>
<th>Flood Peak derived in 1997 (m$^3$s$^{-1}$)</th>
<th>Revised peak discharge post 2004 (m$^3$s$^{-1}$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>2.33*</td>
<td>50</td>
<td>420</td>
<td>480</td>
</tr>
<tr>
<td>5</td>
<td>20</td>
<td>540</td>
<td>650</td>
</tr>
<tr>
<td>10</td>
<td>10</td>
<td>720</td>
<td>850</td>
</tr>
<tr>
<td>20</td>
<td>5</td>
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</tr>
<tr>
<td>50</td>
<td>2</td>
<td>1200</td>
<td>1250</td>
</tr>
<tr>
<td>100</td>
<td>1</td>
<td>1400</td>
<td>1500</td>
</tr>
<tr>
<td>200</td>
<td>0.5</td>
<td>1550</td>
<td>1700</td>
</tr>
</tbody>
</table>

* Also the mean annual flood.

The Rangipo Dam is a settling basin, trapping gravels and sand, thereby reducing sediment available to the middle Tongariro River (Smart, 1992). However, during floods greater than 100 m$^3$s$^{-1}$ the Rangipo sluice gates are opened and diversion at the Poutu Intake is stopped, allowing high discharges to flush up to 60,000 tons of sediment stored within the Rangipo reservoir down the Tongariro River (Collier, 2002). As a result, the delivery of material to the river is more pulsed than would be expected from the natural regime, with surges in supply following flood events > 100 m$^3$s$^{-1}$ and reduced loads at normal flows. More detailed appraisal of impacts of the dam on sediment transport is presented in Section 4.3.1.

### 2.6.3 Flood History of the lower Tongariro River

The Tongariro has been observed to be shaped by large magnitude events, with large floods causing channel avulsion and altering the distribution of pools (Cooper and Cooper, 1975; Scott, 1999; Sir Alexander Gibb and Partners, 1960; Smart, 2005).

Figure 2.21 presents the frequency of large magnitude events since records started in 1957. Floods with a discharge greater than 400 m$^3$s$^{-1}$ are graphed, capturing flows from around bankfull ($Q_{2.33}^1$ is 480 m$^3$s$^{-1}$) and above. This shows a general pattern of an increased frequency of large floods during the 1950s and 1960s. Large floods were rare between 1970 and the early 1980s, with none exceeding the 5 year flood. Large floods increased in frequency from the late 1980s till today, with a cluster of floods with recurrence intervals greater than 5 years occurring between 1996 - 2005. Large floods are only influenced minimally by the TPDS, as once the flow reaches 100 m$^3$s$^{-1}$ the sluice gates are opened and the flow diversion at the Poutu Intake stopped, so that all the flow is flushed down the Tongariro River. This retains similar patterns of large magnitude events pre and post regulation.
Discharge graphs of all floods with a recurrence interval of 10 years or greater are graphed (Figure 2.22). The largest flood on record occurred in 1958, with a maximum discharge of 1470 m$^3$s$^{-1}$. This completely reworked the river, destroying many of the fishing pools and transporting large volumes of boulders (Sir Alexander Gibb and Partners, 1960). The 1964 flood had a smaller, shorter peak of 1000 m$^3$s$^{-1}$ and is less documented due to fewer impacts on the channel. The 1998 flood has a smaller peak again of 877 m$^3$s$^{-1}$. However, this event was preceded by a similar sized event 7 days earlier, making it more geomorphically effective than its peak alone would indicate. Channel change and reworking was observed, with rapid aggradation of the bed below State Highway one (SH1) bridge (Scott, 1999; Tonkin and Taylor, 2003). More recently the 2004 flood had a peak discharge of 1400 m$^3$s$^{-1}$. This caused channel avulsion and reworking in the terrace confined reaches and furthered the aggradation of downstream of SH1 Bridge (Smart, 2005). High levels of Lake Taupo at the time are perceived to have amplified this aggradation (Munro, 2004).
Figure 2.22: Discharge graphs showing the maximum discharge for the four largest floods on record over 3 day periods.
2.7 FLORA AND FAUNA ACROSS THE TONGARIRO CATCHMENT

Much of the catchment area has retained high quality native vegetation (Figure 2.23). The Kaimanawa Ranges in the east of the catchment are covered with thick native vegetation. This includes sections in Podocarp forest including Totara (*Podocarpus totara*), Matai (*Podocarpus spicatus*) and Rimu (*Dacrydium cupressinum*) species. Beech forest (*Nothofagus fusca* and *N. mensiesii*) becomes dominant at higher elevations and is the most common vegetation type across the steep uplands region (Atkinson, 1981; Forsyth, 1983). The Kaimanawa Ranges support a wide range of native birds including kaka, tomtit, bellbird, tui, kakariki, riflemen and kereru (Department of Conservation, 1991). The Tongariro River in the mid-catchment provides habitat for the endangered Whio or Blue Duck which is listed as ‘nationally vulnerable’. Whio live in medium – large rivers with white water, and require high quality habitats, both riverine (good water quality, low suspended sediment) and terrestrial (native riparian vegetation with few pests) to survive (Glaser et al., 2010). While dam construction has decreased the numbers of Whio in the Tongariro River (five or six pairs fewer than in 1983) management programmes as part of dam mitigation have increased numbers in streams in other catchments to the west (Genesis Energy, 2009).

Exotic fauna threatens the Kaimanawa forest park. Red and Sika deer were introduced as far back as 1883 and can still be found in large numbers today. Other pests include possums, rats, cats, mustelids and pigs (Department of Conservation, 1991).

To the west of the catchment, the steep areas of the volcanic cones remain sparsely vegetated, with lichens and mosses such as the low creeping vegetable sheep creating alpine herbfields and subalpine scrub (Figure 2.23). The area further down the slopes, on the Volcanic Plateau is occupied by subalpine shrubland, consisting mainly of tussock grasses (Figure 2.23). Red tussock is the dominant species, though it acts to shelter other species including hebes, mosses and daisies (DOC, 2007). The rest of the region at lower altitudes consists of native scrub including mainly Manuka, Kanuka or bracken, with some small stands of Beech.

Lower down in the catchment the vegetation is dominated by human planted vegetation. Large stretches of pine plantation (*Pinus Radiata*) in the middle of the catchment are used for exotic tree harvesting (Figure 2.23). Further down, pasture is common, mostly stocked by sheep with some cattle. Following development of the TPDS, willow trees were planted alongside the river for bank stabilization and bank erosion remediation, due to concerns for increases in river-bed level and flooding (Genesis Energy, 2000). Recently, many of these have been removed in the lower wetland area as part of a rehabilitation plan carried out by the local iwi. Despite this, willows are common
within the wandering gravel bed river section, especially colonising bars. Similarly, exotic broom also alters the natural roughness and cohesiveness of these features.

The river delta is part of New Zealand’s largest remaining wetland and provides habitat to a range of threatened birds and animals (Chagué-Goff et al., 1999; Genesis Energy, 2000). It is rated as holding high – outstanding value in biodiversity and habitat value (Tonkin and Taylor, 1999a). This includes both swamp and bog vegetation types consisting mainly of flax, cabbage trees and toetoe. Other native wetland communities include raupo (reeds), *Leptocarpus* and *Carex-Juncus* (sedge-rush) land and submerged macrophyte herb-fields (Tonkin and Taylor, 1999a). The Tongariro wetland also supports a wide range of bird species including bitten, spotless crake, fernbird and water-fowl.

The Tongariro River supports a wide range of native fish species in the lower catchment. The Waikato Falls, immediately upstream of the Poutu Intake Dam provides a natural barrier to fish migration upstream of this location (Genesis Energy, 2000). Fish species include koaro, common bullies and smelt. No eels are found in the Tongariro River due to the natural barrier of the Huka Falls at the outlet of Lake Taupo. The river also supports a high diversity of macroinvertebrate species, including sensitive EPT (Ephemeroptera, Plecoptera and Trichoptera) taxa, whose abundance has undergone minimal change downstream of the dams (Storey, 2010). However, the Tongariro is most well known as a trout fishery. Brown trout was initially introduced into Lake Taupo in 1886 with Rainbow trout following in 1897 (Cooper and Cooper, 1975). These were found to thrive within this environment and especially in the Tongariro River, promoting the development of Turangi township.
Figure 2.23: Distribution of vegetation across the Tongariro Catchment. Data scoured from the New Zealand Land Resource Inventory (Landcare Research NZ Ltd, 2000).
2.8 CATCHMENT LAND USE AND HISTORY

The Tongariro catchment has a unique history, with a multitude of natural and anthropogenic controls influencing system adjustment. This section describes the catchment history, followed by an overview of these controls.

2.8.1 Land Use History

Initially the Tongariro River was named the Upper Waikato River, aptly linking it as a headwater to the greater Waikato River, which drains Lake Taupo. Due to the inland and mountainous nature of the Tongariro catchment, European colonisation was late. Initial European contact in 1830 found small Maori pas dotted around the lake, characterised by small groups of houses and a church. At this time the total numbers of Maori around the entire Lake Taupo are estimated to be between 1,100 and 5000 people (Department of Conservation, 2002). Early Europeans saw this area as a place to visit and not to live, and settlement did not occur until the early 1900s. Colonial development commenced earlier in Tokaanu, 5 km north-west of Turangi, with the establishment of an Armed Constabulary post in the 1870s (Waitangi Tribunal, 1995). This was followed by the building of Tokaanu hotel, which acted as a gateway to the volcanoes in the south and to fishing the Turangi River (Figure 2.24). Subsequently, the land surrounding Turangi was cleared for farmland, and Grace farm and homestead was built in 1856 and stocked with Merino sheep (as shown on Figure 2.25) (Department of Conservation, 1991). Fishing camps were established in the 1920s on the margins of the Tongariro River and fishermen would travel from Turangi to the camps. In the 1930s the town became more established with the creation of the first fishing lodge and post office. Land was subdivided at this time. Rangipo prison farm was also established in the 1920s upstream of Turangi township. By 1961 the town still only had a population of 489 (Waitangi Tribunal, 1995) and was limited in size (Figure 2.26). The town grew rapidly with the development of the TPDS and is estimated to have reached 4221 by 1981 (Genesis Energy, 2000). The present population has dropped to around 3,400 (Statistics New Zealand, 2010).

The Kaimanawa Ranges comprise the eastern headwaters of the Tongariro catchment, which the Waipakihi River drains. Land use in this area has been minimal. Before European settlement Maori used the Kaimanawa Ranges for settlements and travel routes (Department of Conservation, 1991). This area now comprises the Kaimanawa forest park, which remains in native vegetation. The rugged terrain has preserved these forests and restricted land use change. Uses are mostly recreational including tramping and hunting. The western sub-catchment comprises the Tongariro National Park. Land use is restricted to tramping and ski fields on Mt Ruapehu and the area has retained Native alpine desert vegetation. This area has high cultural significance to Maori (Department of
Conservation, 2006). In summary, anthropogenic influences are minimal across much of the mid-upper catchment due to their protected status as national and forest parks.

Agriculture has become the primary land use in the lower catchment over the last 40 years, dominated by sheep and cattle grazing (Strachan and Crawford, 1983). However, the intensity of farming has been restricted due to poor quality pumice derived soils, which lack soil moisture, nutrients and essential trace minerals. Regulatory restrictions upon the use of fertilizer also limits the growth of farming activities within the region (Gregg, 1960; Strachan and Crawford, 1983). As a result of the poor soil, pine plantations have been established in the mid catchment (Figure 2.23).

Due to the high proportion of the catchment which has retained its natural vegetation cover, alterations to non-point source land use and inferred to have had minimal impacts upon flow and sediment interactions upon the channel. The more direct impacts discussed below are perceived to have a greater impact upon channel dynamics.
Chapter 2: Regional Setting

Figure 2.24: Historical photographs of the lower Tongariro River. A) A gentleman fishing in a waka in the lower river with swamp native vegetation in the background (taken 1910). B) Looking downstream the Tongariro River at the location of the Turangi township showing the construction of the Turangi Road Bridge (taken 1912). This shows minimal development at this time, the braided planform downstream of the bridge and the scrubby nature of floodplain vegetation, photographers unknown (Alexander Turnbull Library, 2012).

Figure 2.25: Historic survey map of the Tongariro delta, plotted in 1900 (sourced from Smart, 2005).
Figure 2.26: A) Turangi Township looking upstream from SH1 bridge in 1951 (taken by Whites Aviation). B) Aerial photograph of the delta taken in 1964 (Whites Aviation). Note the small size of the settlement during this time-period.
2.8.2 Stopbanks, People and Property

Figure 2.27: Distribution and character of stopbanks and channel management works within the Lower Tongariro River. Maroon shaded areas indicate zones of gravel and debris clearance, green shaded areas willow trimming and removal, orange lines show locations of stopbanks, light blue areas of protection (i.e. rip rap), and the dark blue line shows the 100 year flood limit. Dashed lines indicate proposed stopbanks and protection development (Jones, 2003).
Flooding presents a significant threat to Turangi township. In 2004 a flood with a 60 year RI (previously predicted as a 100 year RI) flooded much of the lower town and delta, affecting 50 homes at a cost of $750,000 (Munro, 2004). Previous large flood events including one in 1958 and two in 1998 (Scott, 1999; Smart, 1999). Extensive flood works were constructed around Turangi in the 1970s to protect it from flooding (Tonkin and Taylor, 2003). These primarily involve the development of stopbanks and flood protection adjacent to the town (Figure 2.27). Willows within the delta and meandering reaches were trimmed and removed to decrease bank roughness and increase channel capacity. Vegetation was cleared from bars in the braided reach. Most stopbanks are located within the zone inundated by a 100 RI flood, illustrating the high risk of flooding despite these features (Figure 2.27).

2.8.3 Gravel Mining

High sediment loads in the Lower Tongariro River prompted gravel extraction in efforts to lower bed level and decrease the risk of flooding. Initially sediment was extracted to build the dams in the upper catchment. Mining commenced in the 1960s when more than one million tons of gravel was extracted below the state highway bridge. This equated to a pit that would have lowered the bed by 2 m across a 200 m wide, 1.5 km long section of river (Genesis Energy, 2000). Mining continued into the 1970s and 1980s, though exact volumes were not recorded (Smart, 1999). This location has also been mined recently. In 2004 the second largest flood on record deposited an estimated 150,000 m³ of sediment at this location, resulting in 1 m aggradation of the bed (Munro, 2004). Consents for mining 15,000m³ of gravel from downstream of the bridge were granted in 2004 (Environment Waikato, 2004). This was carried out by removing sediment from the dry bars more than 0.3 m above water level to decrease the direct influence on the form of the wetted channel as occurred during the 1960s (Tonkin and Taylor, 2003). Thus gravel mining has provided the on-going form of actively managing the planform of the braided reach directly downstream of State Highway 1 Bridge.

2.8.4 Lake Taupo Levels

The Tongariro River’s sediment and water flows are influenced by base-level controls set by Lake Taupo. Control gates installed in 1941 on the outflow of Lake Taupo increased mean lake levels from 1906 – 1940 to 1942 – 1996 by 6.5 cm (Eser and Rosen, 2000), skewing the overall distribution of lake levels towards higher levels (Smart, 1999) (Figure 2.28). However, fluctuations in level are not evenly distributed across seasons. Following installation of the control gates, mean monthly lake levels during summer months are up to 20 cm higher, while winter levels are far more similar to
those present before artificial control. Lake level variation has also altered over time, with the mean level from 1942 – 1949 being 31 cm higher than 1906 – 1940. Increasing the level of Lake Taupo has created a backwater effect that is reported to extend 3 km upstream of the delta, acting to limit sediment transport though the lower reaches of the Tongariro River (Scott, 1999; Smart, 1999) and enhancing landward movement of saturated soils and the area of wetland (Eser and Rosen, 2000; Tonkin and Taylor, 1999a).

Figure 2.29 illustrates the increase in wetland extent on the Tongariro delta that was observed following the control of Lake Taupo water levels. High lake levels due to seasonal regulation were also identified as a contributing factor enhancing the extent of flooding and deposition during the 60 year RI flood in 2004 (Munro, 2004). Thus, high lake levels provide a major factor impeding sediment delivery to Lake Taupo.

![Distribution plot of level of Lake Taupo](image-url)

**Figure 2.28:** Distribution plot of level of Lake Taupo A) prior to regulation (for 36 year before 1941) and B) since the lake level has been run on the current regime (from 1958 - 1999) (sourced from Smart, 1999).
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2.9 CONCLUSION

The Tongariro catchment is a complex environment, with a range of natural and anthropogenic pressures that alter sediment transport and channel adjustment. Tectonic deformation and the active volcanic history have shaped the catchment, creating a complex topographic and geological setting. This resulting rapid shift in boundary conditions creates a wide range of river types within a relatively short distance. The additional layer of anthropogenic controls including regulation from dams, control of the water level of Lake Taupo and gravel mining complicates the situation further. Large magnitude floods are key agents of river adjustment. This presents a considerable challenge in identifying how these controls interact and the extent to which they drive patterns of channel adjustment.

It is the complexity of the multiple, overlapping controls and rapid change in boundary conditions that requires the in-depth methodology used within this study to understand the dynamics of the Tongariro River. Simple fluvial audits or methods which record geomorphic features are not sufficient to capture the complexity of the morphodynamics. Modelling, which simplifies the real world would be insufficient in this situation due to the complexities in stratigraphy, geology, bed material and boundary conditions. Instead, the Tongariro provided the ideal location to further methodologies which build a comprehensive, qualitative and quantitative geomorphic view of the
system. This was done by creating the multi-scalar framework which analyses form to infer geomorphic processes presented in the following chapters. Thus, this chapter provides context and rationale for the further, more detailed studies undertaken in the rest of this thesis.
Chapter 3: Catchment-scale sediment flux

3 LONGER-TERM SPATIO-TEMPORAL PATTERNS OF SEDIMENT FLUXES IN THE TONGARIRO CATCHMENT

3.1 INTRODUCTION

“Rivers are the gutters, down which flow the ruins of continents”

(Leopold et al., 1964)

“Drainage basins are complex cascading systems where responses in the sediment system...may reverberate over large distances with significant time lags.”

(Dearing and Jones, 2003; pg. 148)

The catchment is the fundamental geomorphic unit (Chorley, 1969). Topographic and hydrologic controls within catchments shape the movement of sediment and water, providing a key control upon the form and functioning of rivers (Kondolf et al., 2006). Understanding the form of the catchment sets the characteristics and distribution of process domains, which describe the processes driving sediment transport. This chapter combines readily available tools including simple channel slope/stream power equations, geomorphological analysis of processes and a sediment budget to analyse patterns of sediment flux across the Tongariro catchment over the past 1.8 ka. These understandings frame the Reach and Bar scale analysis in results Chapters 4-7.

3.1.1 Aims and Objectives

This chapter has two main objectives:

1) To classify landscape units in the Tongariro catchment in relation to sediment transport processes. This presents a toolkit with which to describe the ability of each unit to generate, transport and store sediment. Connectivity within and between landscape units is used to assess sediment transport patterns across the whole catchment.

2) To create a sediment budget quantifying the bulk transfer of sediment since the Taupo eruption (1.8 ka). LiDAR data are used to calculate volumes of sediment erosion (terrace incision) and deposition (delta accumulation) by digitally recreating pre-eruption surfaces. This describes the redistribution of sediment across the catchment in the last 1800 years.

3.1.2 Conceptualising Catchment Sediment Flux

River systems are commonly conceptualised as jerky conveyer belts which sporadically transport sediment along their length and out to sea (Fergusson, 1981). However, this jerkiness is not linear along the catchment, and sediment transport becomes disconnected at points of storage, and accelerated within steep and highly connected terrains (Fryirs et al., 2007a; Fuller and Marden, 2011). Thus the sediment generation capability of a catchment may bear little relationship to
sediment volumes delivered to the mouth. This is related to either the (dis)connectivity of the fluvial system or the ability of sediment stores to buffer long-term inputs (Bartley et al., 2007; Métivier and Gaudemer, 1999; Phillips, 2003b; Phillips et al., 2004; Phillips and Slattery, 2006; Trimble, 2010). The configuration of landscape compartments and the transport of sediment within and between them provides a framework for viewing sediment cascades through a catchment (Brierley et al., 2006). This section highlights literature which describes controls upon sediment transport through landscape components, providing a conceptual frame for analysis of catchment-scale sediment flux.

Catchment processes are commonly delineated into 3 distinct zones; the source headwater zone characterised by active erosion and high sediment supply, the mid-catchment transfer zone where material delivered is approximately equal to that exported and the lower catchment depositional zone which is characterised by net sediment storage (Schumm, 1977). The conceptual catchment has an orderly downstream progression of these zones, with headwater sources zones grading to mid-catchment transfer zones and depositional zones present at the downstream extent. However, in reality catchment configuration is more complex and these process zones can be found at multiple locations across the catchment (Fryirs et al., 2007a). These discontinuities complicate sediment routing through a catchment, resulting in non-linear, dynamic relationships (Phillips, 2003a).

Montgomery (1999) presents landscapes as a mosaic of process-domains, whereby the landscape is separated into zones characterised by distinct suites of geomorphic processes. Differences in process domains are driven by the distribution of climate, geology, vegetation and relief factors, which determine the generation, transport and storage of sediment. Wohl (2010) advocates the use of process domains as a conceptual tool to aid in interpreting the spatial patterns and heterogeneity of sediment flux at a catchment scale. She illustrates their use by estimating sediment volumes and residence times in bedrock canyons. Polvi et al. (2011) found a positive correlation between process domains and riparian vegetation composition, shaped by different magnitudes and types of disturbance present in the different landforms. Brierley and Fryirs (2005) similarly delineate ‘landscape units’ based on relief, morphology, geology and landscape position, acting to separate the catchment into components with the same landscape forming processes.

Imposed and fixed boundary conditions can describe the transfer of sediment and energy in any given landscape unit (Brierley and Fryirs, 2005). The imposed boundary conditions may be defined as the underlying structure or morphology of the landscape unit, which determines the distribution of relief, valley morphology and slope. This describes controls which remain stable over long time-periods (Brierley and Fryirs, 2005). Flux boundary conditions describe the distribution of energy and sediment, representing shorter timeframes of adjustment. Flux boundary conditions can be
conceptualised as a more dynamic, process based layer which can be overlaid on imposed boundary conditions. For example, changes to climate can alter the energy regime determining sediment transport, such as seen in the alternating periods of drought and flood dominated regimes in Australia (Erskine and Warner, 1998). Similarly, the volume of sediment available to be reworked and transported also changes over time due to events such as influxes of sediment from cyclones (Marutani et al., 1999), volcanic eruptions (Gran and Montgomery, 2005) or responses to land use change (Wilkinson and McElroy, 2007).

Sediment generation and sediment storage provide the two major controls on sediment transport and output from a landscape unit. Sediment generation dictates the volume of sediment available for transport, controlled by the erodibility (lithology) and slope (available energy) of the landscape unit (Brierley et al., 2011). However, quantifying the processes which generate and transport sediment is complicated, due to high spatial and temporal variation (Burt and Allison, 2010). Within New Zealand, sediment is predominantly generated by mass-movement processes on hillslopes, including landslides, gullies and earthflows (Dymond et al., 2010). Vegetation offers an additional control on hillslope erosion, as illustrated by the high yields following deforestation associated with European settlement (Marden et al., 2011). Appraisal of sediment fluxes needs to be grounded through the specific erosion processes operating within each landscape unit or domain. Dymond et al. (2010) classified erosion terrains within New Zealand based on slope, soil type and rock type. This presents a template with which to infer mechanisms and volumes of sediment generated within different landscape units.

Uplifting mountain belts provide a major source of sediment to river systems. Many ranges have been found to be in dynamic equilibrium, where the rates of uplift and balanced by similar rates of sediment generation (Davies and Korup, 2010; Howard, 1994; Montgomery, 2001a; Selby, 1993; Whipple, 2001). This is supported by Adams (1980) in his sediment budget on the Southern Alps in New Zealand, which found that uplift and erosion were approximately equal. Davies and Korup (2010) suggest that landscapes that have undergone uplift for > 10⁶ years will have reached this state of dynamic equilibrium. Tippett and Kemp’s (1995) model of geomorphic evolution of the Southern Alps indicates that if uplift exceeded denudation by as little as 1 mm y⁻¹, the Alps would have reached altitudes of 10 km above sea level, more than three times their present height.

In essence, sediment flux in a catchment describes the relationship between sediment generation and movement. This is commonly described using the sediment delivery ratio (SDR), which describes the ratio of sediment generation to delivery at base level. For a steady state system to exist, creating a SDR of 1, slope along long profiles has to be adjusted to transport the volumes of sediment
supplied (Davies and Korup, 2010; Lu et al., 2005; Mackin, 1948). Reneau and Dietrich (1991) supported this by compared bedrock lowering rates with sediment yields in streams, finding a balance between generation and export, a relationship which remained intact independent of basin size. Thus, the relationship between the sediment volumes generated and output from a catchment describes how connected the catchment is.

Zones of sediment storage indicate a decreased efficiency in sediment transport. Significant stores of sediment can ‘buffer’ sediment output from catchments, creating constant sediment delivery rates over long time periods (Dearing and Jones, 2003; Métivier and Gaudemer, 1999; Phillips, 2003b; Phillips and Slattery, 2006). Dearing and Jones (2003) identified catchment size as a control on variability in sediment flux. Larger basins ($10^3 – 10^6 \text{ km}^2$) with significant sediment storage are more likely to be spatially and temporally decoupled, buffering changes in sediment input, compared with smaller more reactive basins. This is supported by similar work (American Society of Civil Engineering, 1975; Robinson, 1977; Walling, 1983). Catchments with higher sediment delivery ratios have a reduced ability to buffer sediment (Walling, 1999). Glacial stores (Church and Slaymaker, 1989) and the pulse of sediment generated following anthropogenic land clearance (Trimble, 2009) may exert a prolonged impact upon contemporary sediment yields. Sediment stores also have the potential to catastrophically fail, releasing large pulses of sediment into the system (Cossart and Fort, 2008; Davies and Korup, 2010; Hovious et al., 2000). This is commonly linked to ground shaking or mass movement processes such as landslides or release of former lakes following failure of glacial moraines. Differentiating whether stores of sediment are adjusting towards a state where they are likely to fail, releasing a pulse of sediment into the system, or whether they are likely to form long-term sinks is a primary consideration for the evolution of the system (Davies and Korup, 2010).

Accommodation space, defined as the area available to store sediment, provides a fixed control on the volume of sediment able to be stored in any given landscape compartment. As valleys widen downstream, accommodation space and sediment storage generally increases (Phillips and Slattery, 2006). However, departures from this pattern can be seen at localised increases in accommodation space and sediment storage at unexpected locations, further complicating the transport of sediment through a catchment (e.g. plateau landscapes, which are directly linked to tectonic setting) (Fryirs et al., 2007a).

Residence time, defined as the “average storage time in a given geomorphic unit” (Jain and Tandon, 2010; 354) (e.g. connected, short-residence time units such as riffles, bars and floodplain pockets or long-term disconnected units such as terraces or lakes) determines whether materials are accessible to be reworked by the contemporary channel. Phillips et al. (2007) contrasted the long residence...
times for terraces, which have been isolated by incision, and within terrace stores that are still available to be reworked by the river. Thus, meaningful differentiation can be made between short term stores that are prone to reworking (e.g. mobile sediments that make up mid-channel bars in river systems) and long term sinks that are spatially isolated from reworking processes (e.g. cohesive sediments that make up floodplain and/or terrace features) (Fryirs and Brierley, 2001). Residence time for sediment storage within a unit reflects the effectiveness of erosion, transport and deposition processes that fashion the behaviour of that landscape compartment. This, in turn, is affected by the position of that feature within the landscape, the surrounding topography, climatic conditions, and vegetation cover (Brown, 1987). Thus, understanding the residence time and nature of sediment storage elements across a catchment provides insight into the transfer of sediment through the system.

3.1.3 Guiding Concepts

This section presents an overview of concepts which were used to frame catchment wide sediment fluxes within this study (Figure 3.1). Connectivity describes the movement of sediment flux in the spatial dimension and landscape memory analyses past catchment evolution to determine which processes have shaped the morphology which determines patterns of sediment generation and transport today.

Figure 3.1: Key factors across multiple scales driving sediment flux in river systems (sourced from Brierley et al., 2011). Circles highlighted identify those dealt with specifically within this section.
3.1.3.1 Connectivity

Connectivity is defined as the transfer of energy and matter between two landscape compartments or within a system as a whole (Chorley and Kennedy, 1971; Fuller and Marden, 2011). As such, it describes the nature and rate of sediment movement through a landscape (Fryirs et al., 2007b; Harvey, 2002). Previous studies have shown that highly connected systems are more likely to exhibit greater sensitivity, as they respond more rapidly to events due to the rapid transfer of sediment (Brunsden, 2001; Fryirs et al., 2007a; Fuller and Marden, 2011; Harvey, 2001; Macklin et al., 2010). Features which impede sediment transfer can act to decouple catchments, disconnecting upstream processes from the receiving environment (Fryirs et al., 2007a; Phillips et al., 2004).

Lateral connectivity assesses whether sediment is delivered directly from hillslopes to drainage lines (Bracken and Croke, 2007; Harvey, 2001; Michaelides and Wainwright, 2002), while longitudinal connectivity considers the channels’ ability to transport the sediment delivered to it (Hooke, 2003b). Together, these attributes determine the extent to which a catchment flushes or stores sediment delivered to the stream network (Dearing and Jones, 2003; Fryirs et al., 2007b). In essence, this compares the within landscape unit connectivity, detailing how sediment can be transported within a process-zone and between unit connectivity, how sediment moves through the sediment cascade (c.f. Jain and Tandon, 2010).

In highly coupled systems, sediment from hillslopes is delivered directly to the channel (Schwendel and Fuller, 2011). In decoupled systems, floodplain pockets act to disconnect hillslope-derived sediments from channels, increasing the time it takes for sediment generated to be moved to the catchment outlet (Bracken and Croke, 2007; Fryirs et al., 2007b). Terraces increase the degree of disconnection by increasing the residence times of sediment (Phillips et al., 2007). Steep channels flush sediment, exhibiting high longitudinal connectivity. In contrast, reaches with low slope or that contain features which impede sediment transfer act to decrease longitudinal connectivity. Fryirs et al. (2007a) differentiate between buffers, which restrict sediment transfer from hillslopes to the channel network as sediments are stored within features such as intact valley fills, piedmont zones, alluvial fans, floodplain pockets and terraces; barriers, which inhibit downstream movement of sediment along channels, as features such as dams and bedrock steps induce base level controls along longitudinal profiles, and blankets which smother landforms, protecting underlying sediments from reworking.

Despite growing support for connectivity as a concept, tools for analysing it still remain scarce (Bracken and Croke, 2007; Fuller and Marden, 2011; Michaelides and Wainwright, 2002). Harvey (2001) used geomorphic mapping to assess slope – channel coupling for gullies. Fryirs et al. (2007b)
used a GIS based framework to identify buffers, barriers and blankets from aerial photographs to calculate the proportion of catchment area which was connected to the stream network. Hooke (2003b) assessed reach scale connectivity for a meandering gravel bed river, using stream competence and local sediment sources as determining factors on reach scale adjustment and coarse sediment connectivity. Schwendel and Fuller (2011) mapped erosional and depositional features adjacent to gullies from aerial photography, relating hillslope – channel coupling to patterns of erosion and deposition in the study reach downstream. Fuller and Marden (2011) used repeat differential GPS to survey an alluvial fan, relating changes in channel morphology of the receiving environment of the fan to the distribution and extent of activity in the upstream gully. Beel et al. (2011) physically measured suspended sediment using turbidimeters upstream and downstream of a hydrologically connected hillslope, and installed Gerlach traps upon ephemeral streams draining the hillslope to capture patterns of coarse material transfer.

Catchment scale analysis of connectivity has also been carried out using modelling approaches which characterise the transport of sediment across the catchment (c.f. Downs and Priestnall, 2003; Lane et al., 2008; Parker, 2010; Reid et al., 2007; Smith et al., 2011). However, these modelling approaches would be strengthened by being more linked to the character of the underlying landscape units, particularly in complex environments rather than just expressing modelled sediment transport. Other approaches commonly focus on small-scale processes within individual components, failing to view connectivity in terms of how these components fit together (Michaelides and Wainwright, 2002). However, many catchment scale studies which consider connectivity have been based on fine sediment transport and sediment delivery ratios, with limited consideration of how sediment moves through the proverbial ‘black box’ which is the catchment (Bracken and Croke, 2007; Fryirs, 2012; Walling, 1983). This highlights a need to use the tools from the catchment scale to create models of sediment connectivity and link this data to local scale, within unit connectivity. Connectivity provides an integral component of the framework used to assess the spatial and temporal distribution of sediment fluxes within this study.

3.1.3.2 Landscape Memory

Catchment morphology is a product of its unique history. Understanding the processes that formed landscapes can contextualise and provide insights into processes operating today. Brierley (2010) divides ‘landscape memory’ into geologic, climatic and anthropogenic components. Geologic memory determines the relief, topography and erodibility of the landscape, driven by tectonic forces and denudation over long time scales. Climatic memory refers to how past climates have shaped the land, such as the influence of glacial deposits upon contemporary sediment loads (Church and
Slaymaker, 1989) or changes to vegetation cover altering runoff dynamics and erosion. Anthropogenic memory refers to past human influences that continue to alter dynamics today, such as the surge of sediment generated from deforestation during European settlement (Trimble, 2009; Wilkinson and McElroy, 2007) or altered systems dynamics as a result of flow regulation (Vörösmarty et al., 2003). This study contextualises the morphology of each landscape unit in the Tongariro catchment with regards to the long term processes (particularly volcanism) that have shaped them.

### 3.1.4 Sediment Budgets

Sediment budgets have formed an integral part of geomorphic enquiry by accounting for sediment movement across a catchment or through a reach (Houben et al., 2009; Slaymaker et al., 2003). They are based on the simple principle of accounting for sediment movement within a defined spatial unit (e.g. hillslope, river reach or catchment) over a given timeframe. The range of tools will not be discussed within this literature review as they have been extensively reviewed elsewhere (Brown et al., 2009; Reid and Dunne, 2003; Reid and Brierley, 2011; Slaymaker, 2003), including discussion of sediment budgets as a management tool (Walling and Collins, 2008). Instead, this section discusses literature which guides the development of the sediment budget outlined later in this chapter.

Sediment budgets have been used as tools to assess catchment connectivity. Phillips et al. (2004) constructed a sediment budget for the Lower Trinity River, finding that the naturally decoupled nature of the catchment resulted in minimal impacts in sediment yield following dam construction in the upper catchment. Trimble’s (2009) sediment budget on Coon creek found that floodplain stores regulated the sediment yield, despite dramatic differences in sediment generated and supplied to the channels in the upper catchment. In connected catchments, with less reworked stores, sediment yields reflect sediment generation in the headwaters more closely (Whipple and Tucker, 2002). Work in the Loess Plateau is reported to have 100 % sediment delivery ratios, regardless of catchment size (Walling, 1983; Wei et al., 2006). Phillips (2003b) found a remarkable consistency between denudation rates and sediment yields in many catchments across the globe. Thus, sediment budgets provide a key tool to analyse the (dis)connectivity of catchment scale sediment flux.

Despite the importance of understanding catchment scale patterns of sediment transfer, examples of sediment budgets which consider sediment routing and storage across entire drainage networks and longer spatial scales are scarce (Goodbred and Kuehl, 1999; Walling, 1983). In part, this reflects the difficulties in incorporating sediment storage units into calculations, compared with the ease of measuring suspended sediment. However, GIS tools are making it easier to recreate relict surfaces and calculate the change in sediment volumes. This effectively reflects a morphological approach to sediment budgeting, as morphological units are used as proxies for volumes of sediment erosion and
deposition (Ashmore and Church, 1998). However, this approach is most commonly applied at the reach scale, using high resolution surveys (e.g. RTK-GPS or laser scanners) to accurately record changes in morphology (Brasington et al., 2000; Fuller et al., 2003a; Milan et al., 2007).

Recent studies have started using morphology of sediment storage units to calculate sediment transfer over longer timeframes. These incorporate both the coarse and fine grained fraction, a feat that is rarely accomplished within most budgets. Goodbred and Kuehl (1999) used sediment cores and radiocarbon dating to digitally reconstruct a 7 ka floodplain surface to describe rates of sediment delivery for the Holocene. Buijsman et al. (2003) used a similar approach to calculate sediment deposition in the mouth of the Columbia River since 1900 by combining LiDAR, bathymetry and digitised aerial photographs to compare DEMs from 1900 and present. Houben et al. (2006) used DEMs and geological and soil maps to calculate Holocene floodplain deposition in the Rhine over the past 7500 years, following anthropogenic influence. Whilst these studies use landforms to calculate sediment volumes, they rarely calculate volumes from each landscape component to provide a more comprehensive picture of sediment flux across the catchment (c.f. Fryirs and Brierley, 2001).

In the following section, several components of these various approaches are brought together to analyse the longer term alluvial sediment budget of the Tongariro catchment.

### 3.2 METHODS

This section describes the approaches used to characterise sediment flux between and within landscape units of the Tongariro catchment.

#### 3.2.1 Geomorphic Classification of Landscape Units and River Styles

Landscape units of the Tongariro catchment were classified based on large scale controls including geology, topography, climate and vegetation cover (Brierley and Fryirs, 2005). This provided a template to ground the analysis of sediment flux through the catchment.

The distribution of river types was also classified across within the catchment, again using the procedure outlined in the River Styles Framework (Brierley and Fryirs, 2005). Rivers were classified based on valley confinement, planform (sinuosity, number of channels), geomorphic units and bed material. Aerial photography and DEM interpretations were ground-truthed by site visits. 22 representative reaches were surveyed in greater detail. These were selected based on being evenly distributed across the catchment and capturing each of the types of rivers within different zones. Initial catchment scale visits to as many points as possible across the catchment were used to help with the selection of 22 sites which were representative of that character of rivers within the catchment.
landscape unit. Full assessment of channel characteristics included planform mapping, 50 Wolman transect counts from the coarsest locale of a bar and bank stratigraphy. This field analysis grounded desktop results.

3.2.2 Connectivity Analysis

Connectivity was quantified by a series of simple indices which were used to describe and compare sediment transport capacity for each landscape unit. These can be separated into measures of lateral channel – hillslope connectivity describing the ability of the hillslopes of erode and generate sediment and longitudinal connectivity, comparing channel competence to transport sediment. These measures were presented as catchment maps and the average and standard deviation values were extracted for each landscape unit.

Much of the analysis for the connectivity analysis is based on using existing techniques to assess sediment transport across landscapes. As such, concentration was not on developing the techniques, but rather on using them as developed by others. For example, the values for ‘n’ and ‘m’ as used in the erosion index (see Section 3.2.2.2) is used as derived from the existing literature and no attempt was made to derive alternate values (c.f. Finlayson and Montgomery, 2003; Whipple and Tucker, 1999). These were grounded by knowledge of the study site, based on where the features they identified were actively acting as sediment sinks or stores, and therefore relevance to the study location. In addition, the values of suspended sediment yields and the definitions of erosion terrains are again used as generated in the literature by Hicks et al. (2011) and Dymond et al. (2010). Thus, the approach has been to combine multiple lines of evidence within this section, derived from a range of both modelled and existing sources. This is seen as a more comprehensive approach, more applicable for this complex landscape than carrying out more in-depth analysis and sensitivity/error testing of a single approach, as is commonly used. This is because connectivity for the overall catchment is developed by combining insights derived from the multiple approaches, rather than being dependant on conclusions from a singular approach.

3.2.2.1 DEM Watershed Processing with the Arc Hydro Toolbox

The Arc Hydro toolbox within Arc-GIS was used to delineate the watershed and stream network. Stream lines from the River Environment Classification (REC) (Snelder et al., 2004) were burnt into a 25 m horizontal resolution Digital Elevation Model (DEM), as the downstream floodplain section was too flat to allow accurate stream delineation. The DEM was then processed, with sinks filled, and flow direction and accumulation models run. Steams were defined using a threshold of 0.25 km² upstream catchment area to delineate the start of drainage lines. The flow accumulation grid which calculates the number of cells upstream was converted into drainage area by multiplying by the cell
area \((25 \times 25 = 625 \text{ m}^2)\). Slope (degrees) was calculated at the same resolution using the Slope tool within Arc-GIS which creates a raster layer detailing the slope of each 25m by 25m cell. This processed dataset provided the input for the measures below.

3.2.2.2 Lateral Hillslope-channel Connectivity

Lateral hillslope connectivity was characterised in two ways. An Erosion Index was calculated as a simplified version of the stream power law of erosion described by Finlayson and Montgomery (2003). Whipple and Tucker (1999) suggest that the stream power model is the most appropriate for use in erosion modelling, as it is based on the physics of erosion:

\[ \dot{e} = kA^n S^m \]  

(1)

where \(\dot{e}\) is the incision rate, \(A\) is drainage area, \(S\) is slope and \(k, m\) and \(n\) are constants. \(k\) is based on the landscape and soils a river is moving through, meaning that it should be determined individually for each landscape. Use of this factor is important for comparing rates of erosion between landscapes. However, as the aim of the erosion index within this study is to identify where erosion is most likely to occur, this coefficient can be discarded making the model easily applied to all landscapes, without intensive soil and geology analysis. \(k\) is also scale-dependant and highly variable, and as such can introduce a high degree of error (Finlayson and Montgomery, 2003). Thus the revised equation is

\[ \dot{e} = A^{0.5} S^{0.67} \]  

(2)

\(m\) and \(n\) control the relative importance of slope or discharge, and are commonly between 0.5 and 1.0. These values differ depending on the type of model used. For example, in the unit stream power model, \(m = 1\) and \(n = 1\). However, this model is based on limited data and less frequently applied (Finlayson and Montgomery, 2003). The value for the shear stress model which was used in this study characterises \(m\) as 1/2 and \(n\) as 2/3 (Finlayson and Montgomery, 2003; Whipple and Tucker, 1999). This study utilises these values and no attempt was made to derive different values. Partly this was due to observation of the zones of erosion being supported by areas of active sediment transport in the field. For example, the highly eroded Kaimanawa Ranges generated high erosion values. In addition, this was just one of the many tools whose outputs were combined to generate the overall connectivity analysis, making use of existing values sufficient for the purposes of this study. This was calculated on a 25 m raster grid. As this model excludes geological, it cannot be used to investigate differences in the volumes of sediment generated. For this reason, it was used as a simplified indicator of which zones would generate sufficient energy to move material to the
Channel. This provided a measure of lateral connectivity, based on whether high shear stresses are all the way to the channel, or localised to the top of the hillside. More detailed erosion models developed within New Zealand incorporate geology and rainfall into erosion and sediment yield predictions (Dymond et al., 2010; Elliott et al., 2008; Elliott and Basher, 2011; Hicks et al., 2011). The erosion index presented here is designed to complement these more complex models, acting to describe hillslope connectivity simply and efficiently.

Hicks et al. (2011) modelled the specific suspended sediment yield across New Zealand using multiple regression. This was used to ground the erosion index. Suspended sediment data from 233 gauges with records that stretch up to 50 years, were correlated with rainfall, geology, soil type, slope and erosion terrains (see below) to determine coefficients to explain the strength of these controls, and predict specific suspended sediment yield across the country. This model was deemed to explain 96% of the variance seen in measured North Island rates (see Hicks et al., 2011 for further information). This dataset was used to understand the relative volumes of fine-grained sediment travelling in suspension likely to be generated by the varying lithologies within the Tongariro catchment. Hicks et al. (2011) manually decreased the predicted suspended sediment volumes for the volcanic catchment, as the modelled was found to be greater than the measured volume. This was attributed to the water draining rapidly into the poor tephra soils, and generating less runoff. It has to be noted that this does not account for sediment that is travelled as bedload, and thus would provide inaccurate predictions of sediment transport in areas dominated by bedload transport. This is particularly important for the volcanic sub-catchment, where the majority of sediment is transported as sand or gravels along the bed of the river as it does not erode to fine grained material as occurs when greywacke is eroded. This dataset was used as generated by Hicks et al. (2011) and no values have been modified within this study.

Dymond et al. (2010) classified Erosion Terrains across New Zealand using rock type, soil and slope to determine the rate and type of erosion processes. This layer was also used as a variable by Hicks et al. (2011) who used the erosion coefficients defined for each terrain for the creation of their suspended sediment yield model (discussed in the previous paragraph). In this study, this layer was used to identify zones with homogeneous erosion processes within each landscape unit, so that greater understanding could be gained about the distribution of sediment generating processes across the catchment. Again, these terrains were presented as defined by Dymond et al. (2010) and have not been modified for this work.

Slope was used to represent highly connected and disconnected areas. A value of 4° was used to identify disconnected zones of sediment storage as identified by Dymond et al. (2010). Highly
connected areas were delineated by a value of 25° as used by Fryirs et al. (2007a). These values were found to accurately represent the observed zones of storage and sediment generation based on field based observations of zones of storage and generation within the Tongariro catchment. For example, active scree slopes are common across the Kaimanawa Ranges and Volcanic uplands, which had slopes of > 25°, whilst the terraces and delta had values of < 4° and are large stores of sediment.

3.2.2.3 Longitudinal Connectivity

Longitudinal connectivity describes the transport capacity of the drainage network. This was primarily assessed by modelling stream power for all the drainage lines across the catchment. It is generally accepted that sediment transport is a function of stream power, which can be used to indicate transport capacity (Phillips and Slattery, 2006). Secondly, more in-depth analysis of sediment transport capacity was carried out at 31 representative sites across the catchment. This used field data to calculate shear stress and the ratio of dimensionless shear stress to critical shear stress as an expression of transport capacity.

Total stream power describes the rate of potential energy expenditure per unit length of channel (Bagnold, 1966). This has been linked to sediment transport capacity, planform characteristics, river channel form, vertical adjustment and initiation of floodplains (Brierley and Fryirs, 2005; Jain et al., 2008; Nanson and Knighton, 1996; Parker, 2010; Reinfels et al., 2004; Whipple and Tucker, 1999). Despite its importance, assessments of stream power at the catchment scale have been rarely carried out (Fonstad, 2003) and even more rarely related back to sediment transport (Parker, 2010). Studies which have analysed stream power at the catchment scale have found a good relationship with geomorphic variables. Reinfels et al. (2004) found that catchment scale patterns of stream power could be used to identify reaches most likely to respond to high magnitude events. Jain et al. (2006) used the distribution of total stream power along a longitudinal profile to assess erosion potential and the distribution of energy available to transport sediment across the catchment. Lawler (1995) demonstrated that bank erosion processes could be related back to changes in stream power along the catchment. Barker et al. (2009) created a method which predicts flood peaks across the catchment and then uses DEMs to analyse longitudinal variations in flood discharge, elevation, slope and flood power. Changes in stream power along longitudinal profiles were also used to identify controls upon floodplain initiation (Jain et al., 2008). As such, stream power provides a key variable for describing channel characteristics and adjustment. It was used within this study to illustrate how the longitudinal connectivity of sediment is likely to change downstream and spatial patterns of sediment transport.
Total stream power is defined by Bagnold (1966) as

$$\Omega = \rho g Q S$$  \hspace{1cm} (3)

Where \(\Omega\) is total stream power (watts), \(\rho\) is the density of water (1000 kg/m\(^3\)s\(^2\)), \(g\) is the gravitational constant (9.8m/s\(^2\)), \(Q\) is discharge (m\(^3\)s\(^{-1}\)) and \(S\) is channel slope (obtained from the DEM in m/m). A catchment area discharge relationship was created using the Water Resources Explorer New Zealand (WRENZ) model developed by the National Institute of Water and Atmospheric Pressure (NIWA, 2011). This regionalised model calculates the discharge for the mean annual flood event. Data from 343 gauging sites across New Zealand and 6324 station years are used to calibrate the model to specific regions (Pearson and McKerchar, 1989). A standard error of 22% is gained. McKerchar and Macky (2001) compare this method with 2 rainfall runoff derived models, finding the regional method to be more closely matched with gauge data. WRENZ was used to extract mean annual flood discharges and catchment areas from points along drainage lines, creating a linear relationship (equation 4), which was used to convert the drainage area raster model to discharge.

$$Q = 0.8391 \times A$$  \hspace{1cm} (4)

Total stream power was calculated for all drainage lines using raster calculator in Spatial Analysis tools, in Arc-GIS. Stream power was not converted into unit stream power, due to difficulties in accurately modelling channel width across a catchment. As a result, stream power needs to be related back to the channel width, sediment size and morphology to ground this measure (Figure 3.2). Drainage line slopes were also analysed to provide a simplistic measure of channel connectivity.

Fonstad (2003) compared patterns of stream power across a catchment with measures of transport capacity at specific sites, finding that the actual available energy could vary from the predicted by as much as 200%. This was attributed to differences in local controls such as geology. For this reason, more detailed assessment was carried out at 31 sites across the catchment. These were used to calculate mean boundary shear stress and the ratio of dimensionless shear stress to critical shear stress (from median grain size). Shear stress is widely used to relate different sized flows to sediment transport and channel form (Bridge and Gabel, 1992; Church, 2010; Coleman and Smart, 2011; Hoyle et al., 2011; Lewin and Brewer, 2001; Montgomery and Buffington, 1997). As such, this acts to incorporate sediment flux into catchment scale connectivity measures.
Sites were selected to represent each landscape unit and capture all major tributaries and sections of trunk stream (Figure 3.2). A representative cross-section and water surface slope was surveyed for each site using a Sokia SET530R total station. Between 50 and 100 bed material samples were collected using the Wolman transect method on the coarsest locale of a bar (Brierley and Hickin, 1985; Wolman, 1954). This aims to capture the largest fraction which the river is competent to transport. Sample spacing was two times the largest clast to ensure the independence of each grain sampled, with 50 samples taken at those bars that were not large enough to allow 100 independent samples to be collected (Brierley and Fryirs, 2005; Church and Kellerhals, 1978). Cross sections were input into Cross-section Hydraulic Analyser Spreadsheet (Natural Resources Conservation Service, 2011) and used to calculate the hydraulic radius for the discharge of the mean annual flood, which were predicted for each site using the WRENZ model described above (McKerchar and Macky, 2001; NIWA, 2011). The hydraulic radius was used to calculate mean boundary shear stress ($\tau_b$) using the du Boys (1879) formula:

\[ \tau_b = \frac{1}{2} g R S \]
$\tau_o = \gamma R S$ \hspace{1cm} (5)

where $\gamma$ is the specific weight of water, a product of gravity and the density of water (9807), and $R$ is the hydraulic radius. Mean boundary shear stress calculates the energy available to transport sediment for a given flood event based on wetted channel dimensions for that event.

Bed shear stress ($\tau_o$) was used to calculate Shields dimensionless shear stress ($\tau^*$) (equation 6).

$\tau^* = \frac{\tau_o}{\rho (s-1) g d}$ \hspace{1cm} (6)

$\rho$ is the density of water, $s$ is the density of sediment and $D$ is the size of sediment of a given fraction of which $D_{50}$ was used in this work (Chanson, 2004). To determine whether this dimensionless shear stress is competent to transport the bed material at a site, critical shear stress must be calculated.

Critical shear stress was calculated using Soulsby and Whitehouse’s (1997) equation based on experimental data (Coleman and Smart, 2011).

$\theta_c = \left[ \frac{0.30}{(1+1.2D_0)} \right] + 0.055 \left[ 1 - \exp(-0.020D_0) \right]$ \hspace{1cm} (2)

$D_0$ is dimensionless grain size, calculated using equation 3.

$D_0 = \left[ \frac{g(s-1)}{\nu^2} \right]^{1/3} d$ \hspace{1cm} (3)

Where $\nu$ is viscosity at a specific water temperature (10 °C is the mean temperature in the Tongariro, which yielded a viscosity of 1.2670E-6 m$^2$s$^{-1}$) and $d$ is grain size, for which $D_{50}$ was used. Finally, the ratio of dimensionless shear stress to critical shear stress was calculated.

$\frac{\tau^*}{\theta_c}$

This expresses the ability of the river to transport the sediment available and was used to express excess energy at a site. Shear stress was used rather than stream power for these specific sites. This is commonly used in engineering and provides a measure of the forces exerted on the river bed, which can be directly related to the size of sediment that is able to be entrained (Shields, 1936). This was seen as a more appropriate measure of sediment entrainment at this local scale than stream power, which is a geographically derived measure of the energy available and is more appropriate and was originally designed to look at patterns across larger scales (Knighton, 1998).

### 3.3 RESULTS

Results of this chapter are separated into two sections:
Chapter 3: Catchment-scale sediment flux

1. Description of the physical characteristics of each landscape unit is followed by analysis of sediment generation and transport, assessing the spatial patterns of sediment flux across the catchment.
2. Derivation of the sediment budget of the catchment over the past 1800 years.

3.3.1 Spatial Patterns of Sediment Transfer across the Tongariro Catchment

This section describes the distribution of landscape units and River Styles across the Tongariro catchment. Bed material characteristics and lateral and longitudinal connectivity within and between landscape units are assessed.

3.3.1.1 Landscape Units and River Styles

Headwater zones of the catchment are divided into two regions; the Kaimanawa Ranges to the east which are underlain by actively uplifting, poorly consolidated greywacke and the volcanic Tongariro National Park to the west.

The Kaimanawa Ranges are separated into three landscape units. The headwaters comprise the Steep uplands. This highly dissected landscape is composed of countless steep and short ‘V’ shaped valleys eroded into the soft rock, creating a dendritic drainage pattern (Figure 3.3A; Table 3.1). Vegetation comprises native Beech - Podocarp forest, with minimal anthropogenic influence. Downstream of this unit lies the Rolling Foothills. This weathered and eroded topography has gently sloped valleys. Further downstream, the Terrace-lands comprise remnant floodplain covered by tephra deposits from the Taupo eruption. Streams have cut down through this material, adjusting to the incision in the trunk channel. As a result, they are confined within steep banks of lahar and alluvial deposits. Despite the lower slope of this landscape unit, the streams have similar character and behaviour, in terms channel geometry, planform and geomorphic units, to those observed in the steeper landscape units upstream. Large sediment protrudes through flow.

Only two types of River Style drain these landscape units on the eastern side of the catchment. First order streams are Confined, low sinuosity, gravel bed rivers, which are actively incising into the greywacke. Steep dissected valleys in the upper reaches are transitional to floodplain and terrace materials at the downstream extent (Figure 3.3B). Many small streams drain into the trunk streams of the Waipakihi River to the north and Whitikau Stream to the south. These streams are classified as Confined, low sinuosity, gravel bed rivers with lateral bars. The lower gradient trunk streams have eroded the valley margins, increasing the accommodation space and volumes of sediment storage. Large unvegetated lateral bars consisting of very coarse gravel material (D$_{50}$ of 61 mm) are actively reworked during floods (Figure 3.5).
Chapter 3: Catchment-scale sediment flux

The western headwaters drain the active Tongariro volcanic zone. This area can be separated into two distinct landscape units; the Steep Volcanic Uplands in the headwaters and the Volcanic Plateau downstream (Figure 3.3A; Table 3.1). The Volcanic Uplands comprise the steep, andesitic volcanic cones of Ruapehu, Pihanga and Ngauruhoe, which are part of the Tongariro massif. These largely comprise andesitic lava flows, interbedded with areas of fragmental rubble and scoria (Neal et al., 2010). Ephemeral streams driven by rainfall events and snowmelt readily mobilise unconsolidated gravels and are classified as Confined, volcanic headwater reaches (Figure 3.3B). However, dense underlying andesite makes the volcanic cones resistant to erosion. These streams are confined within steep gully features. Very few instream sediment storage features are present. Sparse alpine vegetation does little to inhibit sediment movement.
Figure 3.3: A) Landscape units of the Tongariro catchment. B) DEM overlain with the distribution of River Styles across the Tongariro catchment with major rivers and landscape features named. Dams and gorges are identified.
## Table 3.1: Distinguishing attributes for each Landscape Unit within the Tongariro Catchment

<table>
<thead>
<tr>
<th>Landscape Unit</th>
<th>Physiographic character of Landscape Morphology</th>
<th>Landscape Position/Location</th>
<th>Geology</th>
<th>Vegetation/Land use</th>
<th>Relief (m asl)</th>
<th>Average Valley Slope (degrees)</th>
<th>Valley Width (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Steep Volcanic Uplands</td>
<td>Steep volcanic cones with drainage lines flowing within gullies cut into this feature</td>
<td>Headwaters of the western volcanic sub-catchment</td>
<td>Andesitic gravels, volcanic sand/ash scoria</td>
<td>Native volcanic alpine desert</td>
<td>399 – 2720</td>
<td>10</td>
<td>350 – 1500</td>
</tr>
<tr>
<td>Volcanic Plateau</td>
<td>Flat depositional unit which captures sediment delivered from the Volcanic Uplands via lahars and airborne tephra. Streams have cut down into this unit flowing within gullies</td>
<td>Downstream of the Steep Volcanic Uplands on the Central Plateau</td>
<td>Unconsolidated scoria and sand with larger volcanic gravels within channels</td>
<td>Native volcanic alpine desert with localised patches of beech forest lower down.</td>
<td>550 - 1231</td>
<td>4</td>
<td>Gullies are 100 – 300</td>
</tr>
<tr>
<td>Steep Confined Headwaters - Kaimanawa Ranges</td>
<td>Many steep, dissected 'V' shaped valleys characterised by a dendritic drainage pattern which drain into trunk streams at the base of the larger valley</td>
<td>Headwaters to the east of the Tongariro River, comprising the Kaimanawa Forest Park</td>
<td>Actively uplifting Greywacke (3 mm/y)</td>
<td>Native Beech and Podocarp forest</td>
<td>518 - 1700</td>
<td>16</td>
<td>500 – 1000</td>
</tr>
<tr>
<td>Tongariro Trunk Stream</td>
<td>Long-term incision into remnant lahar and tephra deposits flows has created terraces, confining the channel and the contemporary floodplain to less than 1 km wide</td>
<td>Runs the length of the catchment, acting as the receiving area to the Kaimanawa and the Volcanic Plateau</td>
<td>Underlining greywacke with andesitic lahar flows cutting across the channel</td>
<td>Native Beech and Podocarp forest in upper section and pine forestry and urban in lowlands</td>
<td>357 -960</td>
<td>5</td>
<td>50 – 1000</td>
</tr>
<tr>
<td>Rolling Foothills</td>
<td>Rounded more weathered hills with divided valleys drained by 1st and 2nd order streams</td>
<td>Located downstream of the Steep Confined Headwaters in the Kaimanawas and on the catchment margin to the east</td>
<td>Remnant floodplains and lahar deposits overlain with tephra deposits</td>
<td>Low intensity sheep farming</td>
<td>358 – 1000</td>
<td>4</td>
<td>&lt; 900</td>
</tr>
<tr>
<td>Terrace-Lands</td>
<td>Flat remnant floodplain which channels have incised into disconnecting this surface from the drainage network</td>
<td>In the lower eastern side of the Tongariro, downstream of the Rolling Foothills.</td>
<td>Remnant floodplains and lahar deposits overlain with tephra deposits</td>
<td>Low intensity sheep farming</td>
<td>358 – 775</td>
<td>3</td>
<td>&lt; 900</td>
</tr>
<tr>
<td>Lowland Plain</td>
<td>Prograding delta, characterised by flat, featureless floodplain with evidence of past channels and levees.</td>
<td>Receiving environment, downstream of terraces</td>
<td>Alluvial deposits including sand and small gravel.</td>
<td>Urban, low intensity agriculture and wetlands</td>
<td>350 - 400</td>
<td>0.2</td>
<td>&lt; 1300</td>
</tr>
</tbody>
</table>
A ring plain at the base of the volcanic cones that contains extensive sediment stores from lahars, lava flows and tephra deposits is named the Volcanic Plateau (Figure 3.3A). Shallow confined gully like valleys cut across the low slope of the Volcanic Plateau, resulting in *Confined, low sinuosity, volcanic plateau cobble bed rivers* (Figure 3.3B; Figure 3.4). This sub-catchment has a parallel drainage pattern, as streams drain east from the volcanic cones until they flow into the Tongariro River. Bed materials within these streams are predominantly made up of andesitic cobble materials.

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**Figure 3.4: Photographs depicting each River Style, nested within the Landscape Units that contain them.**

<table>
<thead>
<tr>
<th>Landscape Unit</th>
<th>River Style</th>
</tr>
</thead>
<tbody>
<tr>
<td>Steep Volcanic Uplands and Volcanic Plateau</td>
<td>Confined, volcanic, headwater rivers</td>
</tr>
<tr>
<td></td>
<td>Confined, low sinuosity, volcanic plateau, cobble bed rivers</td>
</tr>
<tr>
<td>Steep Confined Headwaters and Rolling Foothills and Terrace Lands</td>
<td>Confined, low sinuosity, gravel bed rivers</td>
</tr>
<tr>
<td></td>
<td>Confined, low sinuosity, gravel bed rivers with lateral bars</td>
</tr>
<tr>
<td>Tongariro Trunk</td>
<td>Terrace confined, bedrock and cobble bed river</td>
</tr>
<tr>
<td></td>
<td>Bedrock Gorge</td>
</tr>
<tr>
<td></td>
<td>Partly confined, wandering cobble bed river</td>
</tr>
<tr>
<td>Lowland Plain</td>
<td>Unconfined, braided, gravel bed river</td>
</tr>
<tr>
<td></td>
<td>Unconfined, meandering sand bed river</td>
</tr>
<tr>
<td></td>
<td>Unconfined, sand bed delta</td>
</tr>
</tbody>
</table>
Chapter 3: Catchment-scale sediment flux

(D_{50} 90 – 210 mm), as streams are competent to flush the more mobile tephra pumice and ash material (Figure 3.5). No major tributaries of the size and drainage area observed within the Kaimanawa Ranges are found within this subcatchment as the drainage pattern limits the drainage area of each stream.

The Tongariro River begins as a sixth order stream, flowing within the midcatchment down the fault between the Kaimanawa Ranges to the east and the Central Plateau to the West (Figure 3.3B). The land directly adjacent to the trunk stream was named the Tongariro Trunk landscape unit. The Terrace confined, bedrock and cobble bed river is present in the upper reaches (Figure 3.4). The channel has incised into remnants of lahar flows and tephra, forming terraces which confine the channel.

Within this section the channel is moderately steep (average slope is 4.6°), alternating between gorges formed by long-term incision in andesitic lava flows and sections of storage characterised by active bar features, where localised increases in terrace width allow short term storage of sediment. Two hydropower dams at the upper and lower extent of this reach have regulated flows (see Chapter 2). However, flood gates are opened for the geomorphically effective flows exceeding 100 m³s⁻¹ retaining similar sediment transport relationships as pre-regulation for larger floods.

Once terraces widen, the lower section of the Tongariro Trunk landscape contains the Partly confined, wandering cobble bed river (Figure 3.3B; Figure 3.4). Terrace incision into lahar deposits is shallower, creating disconnected floodplain pockets within these margins. The wandering cobble bed river contains short-term sediment stores as lateral bars and islands, which undergo cycles of stabilisation, revegetation and reworking (Chapter 6).

Once the channel exits the confines of the terraces it flows across a low gradient Lowland Plain and into the volcanic caldera of Lake Taupo. Channel form changes rapidly within this section, starting as an Unconfined, braided, gravel bed river, where gravel is abruptly deposited within bars which are actively reworked. At the downstream extent of the gravel-sand threshold, the reach transitions to an Unconfined, meandering sand bed river, with small point gravel bars on the inside of meander bends. Immediately above the outlet to Lake Taupo, the channel changes to an Unconfined, sand bed delta, characterized by multiple stable channels. Gradual decrease in slope and energy along the delta induces a high diversity of river styles, driven by changes in the ability of reaches to transport differing grain size fractions. The upper section of the delta is adjacent to the town of Turangi, whilst the lower sections drain low density farming and native wetland. Stopbanks were installed in the 1970s adjacent to Turangi, acting to limit the extent of lateral adjustment for the upper sections of
the braided reach. The delta is the receiving environment for sediment generated across the catchment.

3.3.1.2 Bed Material distribution

Bed material size was measured across the catchment. Unfortunately it is difficult to differentiate between the volcanic and greywacke material, due to both looking very similar and more in-depth analysis requiring microscopes to see the difference in crystals was needed to differentiate between the two geologies. Thus, it was not possible to determine the proportion of each geology that made up the bars in the lower Tongariro and this section only presents grain size.

Bed material within the Tongariro Trunk landscape unit is between 91 – 230 mm with an average of 143 mm (Figure 3.5, # 6-15). Incision has reworked lahar deposits, leaving a lag of boulders lining the bed (see Chapter 6). The most downstream site within this unit was noticeably finer grained (91 mm) (Figure 3.5, #6), reflecting the decrease in slope immediately upstream of the delta. Sediment size within the delta decreases rapidly, from 85 mm at the upper reaches to 1 mm at the downstream extent (Figure 3.5, #1-5).

The two contributing subcatchments display a marked difference in sediment size. Average D$_{50}$ in tributaries draining the Kaimanawa Ranges is 75 mm (Figure 3.5, # 20-22). This is much finer than the material in the Tongariro River itself, suggesting that this fraction is easily flushed over the basement of lahar lag once it is delivered to the trunk stream. Evidence for this is seen at the Sand Pool site, whereby a D$_{50}$ of 163 mm is present at the confluence with the Whitikau Stream (which delivers a D$_{50}$ of 65 mm). The influence of this material disappears rapidly, as the D$_{50}$ at Blue Pool 500 m further downstream has increased to 230 mm.

The average D$_{50}$ for streams in the Volcanic Plateau is 133 mm, ranging from 210 to 90 mm (Figure 3.5, # 25-31). Bed material is considered to reflect the frequency and magnitude of lahars. The Mangatoetoenui Stream had a much smaller D$_{50}$ of 95 mm compared to adjacent catchments (Figure 3.5, #26). This stream represents the major lahar path from Mt Ruapehu, with the shortest listed reoccurrence interval of 35 years (Cronin et al., 1997). The most recent lahar to enter the Tongariro River occurred in 1995 – 1996. This delivered between 50 - 5400 tonnes of sediment per day throughout November and December 1995 (Collier, 2002). Minimal sorting has occurred due to the short recovery time since this event, resulting in an increase of smaller material available to be transported. The most upstream site in the Volcanic Plateau captures the upstream extent of a drainage line, reflected in very small, unconsolidated material, with a D$_{50}$ of 3 mm (Figure 3.5, #25).
Two sample sites were located in the Terrace-lands. Bed material at the southern (upstream) site was primarily delivered from the Kaimanawa Ranges. This site had a $D_{50}$ of 65 mm, which is similar to the other streams draining these ranges (Figure 3.5, #23). The downstream site located close to the Tongariro Trunk has undergone incision into lahar lag materials as the channel adjusted to the base level set by the incising Tongariro River (Figure 3.5, #24). The $D_{50}$ at this site was 210 mm, reflecting access to much coarser sediments through alluvial reworking of lahar and terrace materials.

![Figure 3.5: Distribution of the median bed material size ($D_{50}$) across the Tongariro catchment.](image)

**3.3.2 Measures of Connectivity**

This section describes the ability of each landscape unit to generate and transport sediment. Connectivity relationships are analysed in terms of lateral and longitudinal components. Table 3.2 presents an overview of variables used to appraise sediment transport capacity within and between landscape units.
Chapter 3: Catchment-scale sediment flux

Table 3.2: Comparison of variables used to express differences in sediment transport capacity and connectivity for each landscape unit. (* refers to data generated from Hicks et al., 2011). Slope was calculated using the slope tool in Arc-GIS (degrees) using a 25 m DEM. Other methods are detailed in Section 3.2.2. Note: mean annual flood discharges are calculated for each 25 m raster cell using a drainage area-discharge equation (see section 3.2.2).

<table>
<thead>
<tr>
<th>Landscape Unit</th>
<th>Volcanic Uplands</th>
<th>Volcanic Plateau</th>
<th>Steep Headwaters</th>
<th>Tangariro Trunk</th>
<th>Rolling Foothills</th>
<th>Terrace lands</th>
<th>Lowland Plain</th>
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<tbody>
<tr>
<td><strong>Measures of Lateral Hillslope Connectivity</strong></td>
<td></td>
<td></td>
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<td>% Slope &lt; 4°</td>
<td>5.59</td>
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<td>509</td>
<td>346</td>
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<td>583</td>
<td>464</td>
<td>386</td>
<td>317</td>
<td>232</td>
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<td>42</td>
<td>38</td>
<td>470</td>
<td>93</td>
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<td>30</td>
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<td>Std dev of Average Annual Suspended Sediment Yield*</td>
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<td>209</td>
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<td>37</td>
<td>47</td>
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<td><strong>Measures of Longitudinal Connectivity</strong></td>
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<td>Stream Power along drainage lines for mean annual flood (w)</td>
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<td>10176</td>
<td>10753</td>
<td>561881</td>
<td>3123</td>
<td>12673</td>
<td>557009</td>
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<td>24489</td>
<td>32620</td>
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<td>Drainage line slope (degrees)</td>
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</table>
3.3.2.1 Lateral Hillslope Connectivity

Lateral connectivity describes within unit sediment transfer, based on the ability of each landscape unit to generate sediment and transport it to drainage lines.

3.3.2.1.1 Slope

This section presents average slopes for each landscape unit, as high slopes indicate available energy to transport sediment. The Steep Confined Headwaters of the Kaimanawa Ranges had the steepest terrain, with an average slope of 25.6°, indicating a highly connected landscape that is able to generate and move high sediment loads (Figure 3.6). Slope in the Volcanic sub-catchment was lower, with an average of 16° in the Volcanic Uplands decreasing to 7° in the Volcanic Plateau. The Volcanic Plateau captures and stores sediment generated in the uplands. The Tongariro Trunk stream has a high slope of 12.8°, exaggerated due to capturing the high slopes of the terrace margins, as well as the steep longitudinal profile of the channel itself. Rolling Foothills have moderate potential to transport sediment with an average slope of 9.3°, which declines to 6° downstream on the Terrace-lands. The Lowland Plain had a low slope of 3.5°, promoting sediment storage in this unit.

![Figure 3.6: Average slope with error bars showing the standard deviation of slope for each landscape unit.](image)

Landscape units were assessed based on the percentage of the unit that was highly connected and likely to generate sediment (>25°) and the percentage likely to store or buffer sediment transport to the channel (<4°). The proximity of these zones to drainage lines was used to infer slope – channel coupling. Figure 3.7 displays the extent to which each unit is connected to the channel with steep slopes (> 25°) or disconnected (< 4°).
76 % of the Steep Volcanic Uplands have hillslopes between 4° and 25° indicating that they are highly connected and able to move the loose mobile scoria and sand deposits present. The flanks of the volcanoes and gullies are especially well-connected, making up 17 % of this landscape unit. Only 5.6 % of the unit has a slope lower than 4°, indicating sediment storage, which was mostly located on interfluvies.

The role of the Volcanic Plateau as a buffer to sediment transport was clear with 36% of the unit comprising slopes less than 4°. Well-connected areas are minimal with only 0.84% of the unit characterised by slopes > 25°. 64% of land falls within the 4° - 25° category, including gully margins within which the streams flow, and the boundary between the steep volcanic cones and the plateau units.

Hillslopes within the Kaimanawa Ranges had a substantial capacity to generate and transport sediment, with 53% having slopes > 25°. Hillslopes retain this steepness to the channel margin, exhibiting high hillslope–channel connectivity. Sediment storage within this landscape unit is rare with 1.7% of the area having slopes < 4°. These are located exclusively on top of ridges which separate the sub-catchments, representing minimal sediment storage potential.

The majority of the Tongariro Trunk stream falls into the 4° – 25° category (64%) which mainly includes the area where the active channel is located. 14% has slopes above 25°, capturing the steep slopes of the terrace margins. The narrow width of the terraces, and inset nature of the channel indicates that lateral terrace erosion has been slow and does not represent a major source of sediment. 22% of the unit has a slope below 4°, capturing the low slopes that characterise the remnant floodplain on top of the terrace. However, these surfaces are disconnected from the contemporary channel.

Streams draining the Steep Headwaters in the Kaimanawa Ranges into the Rolling Foothills and then the Terrace-lands undergo an increase in the proportion of the catchment with slopes less than 4°, from 1.7% to 19% to 52% respectively. The proportion of the catchment > 25° remains low for both downstream units, at 1.9% for the Rolling Foothills and 1.7% for the Terrace-lands. This indicates limited potential to generate and transport sediment to drainage lines in this area.
Figure 3.7: A) Map and B) graph showing the lateral connectivity of sediment flux for the Tongariro catchment described through slope thresholds. Slopes with an angle greater than 25° are considered to be highly connected, whilst slope less than 4° are considered disconnected, acting as buffers impeding sediment transport. Slope was calculated using the slope tool in Arc-GIS (degrees) using a 25 m DEM.
3.3.2.1.2 Erosion Index

An erosion index, derived using a simplified stream power function of slope and drainage area, was modelled across the catchment (Section 3.2.2.2). As this does not take into account substrate
erodibility, it is used to identify zones where discharge and slope are likely to be sufficient to erode and transport sediment on hillslopes. This indicates which zones are likely to undergo erosion, and whether this erosion continues to the channel so that lateral hillslope connectivity can be assessed.

The Steep Volcanic Uplands have the second highest average erosion index value of 402, indicating a high potential to erode and transport sediment. Higher values are located on the upper flanks of the volcanoes, especially on the gully complexes which have cut down into these features. This unit has very high variability due to rapid increases in drainage area and steep slopes creating a fast increase in erosion index values.

Values are lower for the Volcanic Plateau, with an average of 233. Higher values are located on the margins of gully complexes. As channels cut down into unconsolidated volcanic sediment, the localised steep slopes on these gullies have the potential to erode and transport sediment locally, maintaining the gullies as local, active sediment sources. Lower erosion index values are present on the flat land separating the gullies indicating that these stores are disconnected from the channel network.

The Steep headwater landscape unit in the Kaimanawa Ranges has the highest average erosion index value of 508, with much of the surface able to generate and transport sediment. Lower erosion index values can be seen on the ridges separating the sub-catchments due to low slope.

The Tongariro Trunk has an average erosion index value of 345. This is noticeably higher than the Volcanic Plateau and landscape units downstream, but lower than the Headwater reaches. High slopes on the terrace margin contribute to this value, indicating potential for terrace erosion.

Localised slopes in the Rolling Foothills can undergo limited erosion, with an average value of 272. This decreases rapidly into the flat Terrace-lands characterised by a value of 169. Flat topography limits sediment transport across the terrace surface, decoupling these stores from the channels. The Lowland Plain had the lowest erosion index value of 135, as low slope disconnects the floodplain from the channel.

3.3.2.1.3 Erosion Terrains and Suspended Sediment Yield

Erosion terrains classify dominant erosion processes based on rock type, soil and slope and modelled specific sediment yield across each landscape unit. These measures incorporated geology to conceptualise sediment flux, offering insight into processes driving sediment generation across the Tongariro catchment.
The Steep Volcanic Uplands have been classified as the Volcano erosion terrain, with sections of the Upland plains/plateau with Tephra in the downstream sections (Figure 3.9A). Mt Pihanga, the oldest and most downstream volcano, was classified as Mountain land on volcanic rocks and then Hill country on airfall tephra in the lower margins. This captures the dormant volcanic nature of this mountain. Despite the steep terrain, sediment yields are predicted to be relatively low across this unit, with an average SSY of 42 t/km$^2$/y (Figure 3.10). This unit is underlain by andesitic lava, which is relatively hard, not easily eroded and characterised by low erosion coefficients, limiting sediment volumes generated (Dymond et al., 2010). However, due to the history of active volcanism, this unit may be characterised by abundant sediment supply as the landscape is mantled in loose volcanic ash, sand and scoria gravels which have been delivered directly to the system as tephra and laharls. This material is predominately moved along the bed of the river as bedload, and thus not captured in suspended sediment measurements, causing a great underestimate in sediment supplied from this landscape unit. In contrast, the greywacke forms fine grained sediment when it is eroded. In addition, directly following eruptions sediment supply increases more, resulting in a long term delivery rate that is orders of magnitude greater than represented in this value. For example, the 1995-1996 eruption and lahar from Mt Ruapehu was estimated to deliver 6900 kilotonnes of sand sized sediment to the lower catchment (Genesis Energy, 2000; Manville et al., 1996). Thus, despite this low value of suspended sediment yield, the overall yield including bedload is regarded as high within this unit.

Specific suspended sediment yields are lower for the Volcanic Plateau, with an average loading of 39 t/km$^2$/y reflecting lower slope and the high porosity of the interflues which increases infiltration and decreases runoff (Figure 3.10). Erosion Terrains are comprised predominantly of the Upland plains/plateaux, with tephra. The gullies associated with some of the larger tributaries draining the plateau are classified as terraces and low fans, and due to increases in slope, they have slightly higher erosion values (Figure 3.9A). This landscape unit has minimal potential to generate or transport sediment, though small volumes may be delivered from localised gully erosion.

Specific suspended sediment yields from the Kaimanawa Ranges are far higher than any other landscape unit, with an annual average of 470 t/km$^2$/y (Figure 3.10). Erosion terrains are classified by Dymond et al. (2010) as Mountain lands on Greywacke, with the highest elevation slopes further classified as Steeplands with sheet/wind/scree erosion, identifying them as particularly susceptible to erosion (Figure 3.9). Steep slopes and comparably soft greywacke create active gullies. As such, this highly connected landscape unit can generate and transport much higher sediment loads than observed in the Volcanic Uplands.
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The Tongariro Trunk stream has an annual average SSY of 93 t/km²/y (Figure 3.10). This is attributable to steep slopes on the terrace margins, and the inclusion of greywacke to the east of the channel. As this is such a localised section, it is unlikely to be a major contributor of sediment. This falls within the erosion category of Terraces and low fans, which is characterised by short term storage as bars and fans of sediment delivered off the hillslopes.

The Rolling Foothills has a very low average SSY of 20 t/km²/y (Figure 3.10). Erosion terrains within this unit includes Hill country with deep young airfall tephra and small patches of Downland (slopes 4-15%). The low sediment yield indicates the low sediment generation capacity of this terrain, despite its location on steeper slopes than the units downstream. The Terrace-lands had an average SSY value of 58 t/km²/y and were classified as Terraces and low fans erosion terrain (Figure 3.9B). This unit has a greater potential to generate suspended sediment than the foothills upstream due to an increase in available fine-grained sediment stored within the terraces.

The Lowland Plain is entirely situated on floodplain and has a low mean annual SSY of 30 t/km²/y indicating low sediment generation and transport capacity across its surface. This is marginally higher than the Rolling foothills as despite low slopes, this unit has an abundance of non-cohesive alluvial sediment which may be entrained.
Figure 3.9: A) Erosion Terrains of the Tongariro catchment defined using rock type, soil, expert opinion and slope to characterise dominant erosion processes (see Dymond et al., 2010). B) The modelled average annual specific suspended sediment loads displayed on a 100 m horizontal raster (see Hicks et al., 2011). This model predicts sediment yield using rainfall, geology, soil type, slope and erosion terrains and is calibrated using suspended sediment data from 233 catchments. Methods are contained within Section 3.2.2.2.
3.3.2.1.4 Lateral Connectivity Summary

The Steep headwater Reaches in the Kaimanawa Ranges have a high proportion (53%) of slopes with angle greater than 25°. This was reflected in an average erosion index value of 509, reflected a marked capacity for this unit to erode and transport sediment. Underlying greywacke is a sedimentary rock, which is moderately easily eroded (especially compared to andesite), reflected in a high specific suspended sediment yield of 470 t/km²/y. This landscape unit is a major supplier of sediment, generating and delivering large volumes of gravel to the Tongariro River.

The Volcanic Uplands have steep sections, with flatter gradients between the cones. Hill-slope channel connectivity declines downstream as slopes decrease. This unit has a moderately-high erosion index value of 402, indicating high erosion potential on the upper slopes. The specific suspended sediment yield from erosion is low (42 t/km²/y) based on slow weathering rates of andesite and ignoring the vast quantities of scoria and ash which are transported from this unit as bedload. As such, this suspended sediment volume greatly under-predicts the actual sediment volumes supplied from active volcanism which characterise lateral sediment transport within this unit.

The Volcanic Plateau has a low gradient, with 36% of the unit characterised by slopes less than 4°, indicating sediment storage. Erosion indexes are also low as little sediment is transported laterally across the plateau into the more active gullies. Specific suspended sediment yields are accordingly low at 38 t/km²/y. Thus, this landscape unit has little ability to erode surfaces and deliver sediment laterally into the river channels.
The Tongariro Trunk is located within a valley confined by terraces. Limited accommodation space within the terraces results in minimal sediment stores. As such, the major sediment source is slow erosion of the terrace margins. This zone has little active sediment sources or stores.

79% of the Rolling Foothills has slopes between 4° – 25° indicative of neither active erosion or storage. This is reflected in a low erosion index of 273 suggesting that few areas actively supply sediment. Specific suspended sediment yields are low at 20 t/km²/y. This is due to its erosion terrain definition of Hill-country with Tephra, where the tephra deposits smoother the landform, limiting erosion.

The Terrace lands have a low gradient, with 52% of the unit characterised by gradients less than 4° indicative of sediment storage. Low slopes generate low erosion index values (197) and little sediment is generated and moved across the floodplain to the channel. The specific suspended sediment yields are marginally higher than the Rolling Foothills upstream at 58 t/km²/y. This is due to its classification as Terraces and low fans, which is associated with greater loss alluvium, which is readily transported.

The Lowland Plain is located at the downstream point of the catchment. This unit has very low gradient, with 59% characterised by slopes less than 4°. This active sediment store impedes the lateral movement of sediment. Despite low slopes, this unit generates a moderate-low specific suspended sediment yield of 30 t/km²/y, due to the loose alluvial nature of the underlying sediment.

### 3.3.2.2 Longitudinal Connectivity

Longitudinal connectivity describes the capacity of drainage lines to move sediment delivered to channels. This assessment is used to characterise the transport of sediment through and between landscape units.

#### 3.3.2.2.1 Drainage Line Slope and Total Stream Power

Total stream power is calculated along all the drainage lines of each unit for the mean annual flood. Mean annual flood discharge was predicted for each grid cell using a catchment area/discharge relationship as discussed in Section 3.2.2.3.

Drainage lines in the Volcanic Uplands had an average slope of 9.5°. However, despite high slopes, low drainage areas and an average mean annual flood of only 2 m³s⁻¹ resulted in very low total stream power, with an average of 3812 Wm⁻¹. The observed ephemeral nature of these streams supports this, suggesting that sediment transport only occurs during infrequent high flows. This indicates the low sediment transfer capacity of these streams during normal flows.
Figure 3.11: A) Mean slope for drainage lines within each landscape unit. Error bars indicate standard deviation of slope. B) Average total stream power for the mean annual flood, for drainage lines within each landscape unit displayed on a log_{10} scale. Error bars display the standard deviation.
Average drainage line slope for the Volcanic Plateau is 4.1° and 76% of streams have a slope less than 5° (Figure 3.29). Total stream power is moderate, at 10175 Wm\(^{-1}\) and the average mean annual
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The Steep Uplands in the Kaimanawa Ranges have the steepest drainage line slope, with an average of 15.9°. This unit has a greater proportion of steep slopes compared to other units, with 80% of slopes > 5°. Despite the low drainage area for many of the streams the average total stream power is high at 10752 Wm⁻¹ from an average mean annual flood of 5.4 m³s⁻¹ (Figure 3.11). Total stream power gradually increases along the trunk stream of the Waipakihi, until it increases substantially once Tongariro River starts draining the Volcanic Plateau, increasing drainage area (Figure 3.13a).

![Diagram](image)

Figure 3.13: Long profiles with catchment area and stream power for A) the Waipakihi Stream draining the Kaimanawa Ranges and B) the Ohinepangu Stream draining Mt Ruapehu and the Volcanic Plateau.

Wider streams act to dissipate energy compared with the uplands, resulting in a decreased capacity to transport sediment than suggested by stream power alone. Figure 3.13 illustrates the low stream power of the Ohinepangu Stream before it joins with the Tongariro Trunk stream at 20 km downstream. Low slope limits the sediment transport capacity for streams draining the Volcanic Plateau, disconnecting the steeper volcanic headwater streams from the Tongariro River. High sediment storage in the Volcanic Plateau limits lateral connectivity, reducing the extent to which this unit can deliver sediment.

flood is 7.3 m³s⁻¹. Wider streams act to dissipate energy compared with the uplands, resulting in a
Figure 3.14: A) Slope (degrees) for drainage lines across the Tongariro catchment. B) Total Stream Power for the mean annual flood (Wm$^{-1}$) This is displayed on a log10 scale, increasing in 0.5 increments.
Figure 3.15: The Waipakihi Stream, showing the lateral bars and erosion scarp as the channel cuts down into valley deposits.

Figure 3.14A demonstrates the much lower stream power in many of the small steep tributaries within the Steep Uplands unit of the Kaimanawa Ranges. This steadily increases for trunk streams including the Waipakihi (located approximately 20 - 30 km downstream) and Whitikau Streams. As these tributaries are confined within valleys, stream power is concentrated and has a greater capacity to transport sediment than indicated by the stream power alone (usually < 1000 Wm²) (Figure 3.11B). Sediment supplied is rapidly transported indicated by the minimal storage apparent within these reaches. In contrast, the higher total stream power is observed in the Kaimanawa trunk streams (20-30 km downstream). The valleys are wider with greater active sediment stores as lateral bars (the Waipakihi has bars up to 100 m wide). During floods, flow is dissipated across these surfaces. Grain size is smaller within these reaches (average $D_{50}$ of 63 mm) indicating lower competence. Sediment transfer is slowed through this zone and stored for short periods. However, the lack of significant stores other than these bar features indicate that while this storage may slow the transfer of sediment through the reach, it is not increasing storage over time. This is supported by evidence of fluvial deposition which the channel is cutting into (Figure 3.15). Thus, in general the drainage lines within the Kaimanawa Ranges are highly connected and able to transport much of the sediment delivered due to few long-term storage features.

The Tongariro Trunk stream had the highest total stream power at 561881 Wm⁻¹ for an average mean annual flood of 307 m³s⁻¹, representing substantial capacity to transport sediment. Moderate slopes (average 4.6°) and progressively increasing discharge provide substantial capacity to transport sediment. The high energy conditions can be observed in the large sediment size (148 – 160 mm)
and steep riffle-cascade units present within this section (Figure 3.5). Terraces and gorges confine the channel, concentrating stream power during high flows and further increasing the energy available. Smaller material delivered from the Kaimanawa Ranges ($D_{50}$ of 61) is likely to be flushed, whilst the channel slowly incises into a larger basement of andesitic material ($D_{50}$ between 90 and 230 mm) delivered by lahar events (see Chapter 6).

Channel slope decreases from the Rolling Foothills to the Terrace-lands from 4.4° to 3.1° respectively. However, an increase in drainage area for the streams in the Terrace-lands results in a higher average total stream power of 12672 Wm$^{-1}$ (average mean annual flood of 10 m$^3$ s$^{-1}$) compared with 3122 Wm$^{-1}$ in the Rolling Foothills (average mean annual flood of 2 m$^3$ s$^{-1}$). This landscape unit has a moderate to low potential to transport sediment, though the larger streams including the Whitikau and Mangamawhitihiti Streams exhibit higher stream power and thus would be expected to have a greater transport capacity.

The Lowland Plain has a very low average channel slope of 0.2°. However, due to the high drainage area stream power is relatively high at 557002 Wm$^{-1}$ for an average mean annual flood of 545 m$^3$ s$^{-1}$. As this section is unconfined, high flows cannot be contained by the channel and flow energy is dissipated on the floodplain. As a result, the channel has a far lower capacity to transport sediment than stream power alone would indicate. This is reflected in a unit stream power of 11140 W/m$^2$. This is supported by much smaller grain size within this section, as it grades from gravel directly downstream of the terraces to sand at the lake edge (from a $D_{50}$ of 80 mm to less than 1 mm) (Figure 3.5).

3.3.2.2.2 Mean boundary shear stress and shear stress ratio

Mean boundary shear stress and the ratio of dimensionless shear stress to critical shear stress were analysed for 31 representative sites across the catchment. Sites were selected to be evenly distributed across the catchment, capture the range of River Styles and Landscape Units and include major trunk streams (see Figure 3.2 for map of sites). This measure relates available energy to channel width and bed material. This analysis provides specific representative examples from which to ground the desktop work presented previously within this chapter.

Mean boundary shear stresses were highest in the upper reaches of the Tongariro Trunk stream with sites 19 and 18 generating 269 and 364 N/m$^2$ respectively. These locations have the highest energy available to transport sediment due to steep slopes and confined cross-sections. Further downstream on the Tongariro Trunk, boundary shear stresses decrease as slope does. Despite this, values remain high with 123 and 129 N/m$^2$ for sites 16 and 17 at lower reaches of the terrace...
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confined reach. Due to difficulties in accessing and surveying this stretch of river, these measurements tended to be located at the flatter, slower sections of this river (i.e pools), and as such, these values can be treated as lower estimates of mean boundary shear stress for this reach. Once the channel is partly confined within terraces, boundary shear stresses varied more, based on whether a location was depositional or not, which can be seen in the values for site numbers between 6 and 15. This reflects a greater degree of variation of slope within this reach (see Table 3.3). Values varied between 42 and 203 N/m², with the upper portion of the reach, tending to be characterised by higher values of shear stress. Sites characters by lower mean boundary shear stress were located at directly upstream of the terrace extent, with sites 6 and 7 having values of 76 and 42 N/m² respectively. Despite this variation, this reach still displays high available energy.

Once the channel leaves the terraces and flows across the Lowland Plain, mean boundary shear stress decrease dramatically. Sites within the braided reach (3 - 5) had values between 24 and 30 N/m², representing low transport capacity compared with other sites across the catchment. This decreases downstream with the meandering river (site 2) and delta site (Site 1) having values of 8 and 3 N/m² respectively. Low slope limits energy, even during the largest flows.

Sites draining the Volcanic Uplands showed high variation in mean boundary shear stress. The most southern site, located in an ephemeral stream had a shear stress value of 29 N/m², due to low slope limiting sediment transport. Larger tributaries exhibited higher energy, with sites 26, 28, 30 and 31 having shear stress values of 191, 112, 110 and 238 N/m² respectively due to high slopes at these sites (Figure 3.3). Despite the flat slopes of the Volcanic Plateau, the channels within the gullies exhibited moderate sediment transport capacity, which is greater than would be indicated by the units slope alone. Two tributaries draining the Volcanic Plateau had had lower energy, with site numbers 27 and 29 generating lower shear stresses of 65 and 70 N/m².
Table 3.3: Mean boundary shear stress and the ratio of dimensionless shear stress to critical shear stress for 31 sites across the Tongariro catchment. Slope and hydraulic radius values are also included for each site, as the key inputs for calculating mean boundary shear stress. Site locations are shown on Figure 3.16 and Figure 3.17.

<table>
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<th>Site Name</th>
<th>Site #</th>
<th>Landscape Unit</th>
<th>D50 (mm)</th>
<th>Slope (m/m)</th>
<th>Hydraulic Radius (m)</th>
<th>Mean annual flood discharge (m^3 s^-1)</th>
<th>Mean boundary shear stress (Nm^-2)</th>
<th>Ratio (ςς/ςς)</th>
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<td>1.35</td>
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Streams draining the Kaimanawa Ranges have low mean boundary shear stresses. Sites 20 and 22 have especially low values of 31 and 42 N/m^2, again driven by low slopes. Site 21 is steeper and has a slightly higher shear stress of 113 N/m^2 due to greater valley confinement. As such, the Kaimanawa Ranges sites have lower shear stresses than the sites located along the Tongariro Trunk and within the Volcanic Plateau.
Two sample sites were located along streams within the Terrace-lands unit. The upstream site had a low shear stress of 50 N/m$^2$, similar to those draining the Kaimanawa Ranges. The downstream site had a much higher shear stress of 246 N/m$^2$ due to steep slopes. This site was located just upstream of the confluence with the Tongariro. As the Tongariro River has incised, the downstream section of this tributary has also incised, creating a steeper slope at the downstream extent and generating higher shear stresses than would be expected for this landscape unit.
Shear stress ratios describe whether energy generated during a flood (i.e. mean annual was used in this study) is sufficient to entrain the available sediment (median grain size). Values less than 1 indicate that the bed material would remain static, whilst values > 1 indicate excess energy.

Reaches located in the upper section of the Tongariro Trunk were all able to transport their bed material. Shear stress ratios of 2.56, 2.16 and 1.44 were calculated for sites 19, 18 and 17 indicating high longitudinal connectivity of sediment flux though the mid-catchment. At Site 16, immediately upstream of where the terraces widen, the ratio decreases to 0.94, reflecting localised increases in storage. Shear stress ratios within the partly terrace-confined reach (Sites 6 – 15) are between 0.42 and 1.60, with an average of 0.92. These reaches have a lower capacity to transport available sediment. This is reflected by storage features such as bars. Transport capacity decreases at the braided reach, down to 0.32 and 0.49 (Sites 4 and 5), and deposition increases. Site 3, located at the gravel-sand transition, exhibits an increase in competence, with a ratio of 1.35 as bed material size has decreased to 25 mm, despite a large decrease in slope. Shear stress ratios increase downstream to 5.57 at the delta mouth (Site 1) due to a decrease in grain size to 1 mm. Whilst these sites have low mean boundary shear stress compared with the rest of the catchment, the small grain size allows sediment transport to occur. Evidence that the channel is narrowing (Chapter 4) indicates that these reaches may be capacity limited and unable to transport the quantities of sediment delivered.

Within the Volcanic Plateau three sites displayed high transport ratios. Site 25 exhibited a high value of 13.77, as the fine bed material can be easily flushed. Sites 26 and 31 had values of 2.25 and 2.43. Site 26 is the Mangatoetoenui Stream which is the most recent lahar flow path from Mt Ruapehu (Cronin et al., 1997). Sediment is less sorted and smaller, and can be transported more easily. Site 31 is steep, and flows are easily able to move the bed material of 110 mm. This is located downstream of Mt Pihanga, where streams are steeper. The other sites draining the Volcanic Plateau had low transport capacities with sites 27, 28, 29 and 03 having ratios of 0.35, 1.01, 0.46 and 1.38. Moderately low slopes and low hydraulic radius limits transport capacity, especially for sites with large bed material from lahars such as the Waihohonu (Site 27) and Mangatewai (Site 29) Streams. Streams within the Volcanic Plateau displayed marked variability in their capacity to transport the available sediment, with plateau location and lahar history playing an important role in shaping the slope and sediment load available.
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Figure 3.17: Map showing the distribution of the ratio between dimensionless shear stress and critical shear stress for 31 sites across the Tongariro catchment. Site numbers correspond with those displayed in Table 3.3.

Two streams draining the Kaimanawa Ranges were just able to transport their bed material, with Sites 21 and 22 having ratios of 1.06 and 1.04, due to large bed material (120 mm) and a high hydraulic radius respectively. The Waipakihi (Site 20), which is the trunk stream draining the upper catchment had a low gradient and much lower shear stress ratio of 0.57. The Whitikau (Site 23), which is the trunk stream draining the Terrace-lands, had a similar ratio of 0.87. Both the Whitikau and the Waipakihi are confined, low sinuosity, gravel bed rivers with bars, as larger catchment areas...
and lower gradients allow sediment to deposit and form lateral bars (Figure 3.3B). Thus, these ratios capture this decrease in transport capacity. The last site in the Terrace-lands, Site 24, had a shear stress ratio of 1.32, reflecting its steep gradient as the stream has incised to meet the incised Tongariro River. This indicates that transport capacities are moderate due to either large sediment in the smaller tributaries, or smaller hydraulic radius due to wider channels with lateral bars.

3.3.2.2.3 Summary of Longitudinal Connectivity
In summary, the upper reaches of the Tongariro River within the Tongariro Trunk have the highest excess energy, easily able to transport the available sediment. As the sediment size is greater in the Tongariro Trunk, smaller sediment delivered from the Steep Uplands in the Kaimanawa Ranges and the Volcanic Plateau can be easily flushed across this basement. Transport capacity decreases on the Lower sections of the Tongariro Trunk as terraces widen and an increase in accommodation space allows short term stores of sediment as bars. Whilst this slows the transport of sediment through this reach, it does not store it over longer timeframes as evidenced in the lack of storage units within the valley margins. Once the channel flows onto Lowland Plain, a rapid decrease in slope is reflected in a decrease in available energy. However, the downstream fining within this reach means that the channel at the downstream extent of the delta is easily able to transport the calibre of sediment at this location. A high supply of sand is flushed through the terraces of the Tongariro Trunk and accumulates within the Lowland Plain where accommodation space is greater on the unconfined floodplains. Streams draining the Volcanic Plateau exhibited moderate shear stress with high variation in transport capacity which is determined by their location on the plateau and the recovery time since the last lahar, which influences the size and volume of available sediment. Several streams were characterised by high longitudinal connectivity, able to transport the sediment available. These streams flowed within confined gullies, disconnecting the drainage lines from available sediment stores. Streams draining the Kaimanawa Ranges and the Terrace lands were characterised by low mean boundary shear stress and moderate shear stress ratios due to the small size of sediment. The confined, low sinuosity, gravel bed rivers are steeper and thus more able to transport the sediment available than the confined, low sinuosity, gravel bed rivers with lateral bars, which store this fraction within the channel. Sediment size is smaller from these landscape units, so whilst the streams may be just able to transport the available sediment, the Tongariro River has a greater ability to flush this fraction. In summary, the Tongariro is a highly connected catchment, with high sediment supply delivered from the Kaimanawa Ranges and Volcanic Uplands. This material is flushed through the Tongariro Trunk in the mid-catchment and onto the Lowland Plain, where it is stored in an aggrading delta.
3.3.3 Sediment Budget for the Tongariro Catchment

In the previous section, the connectivity of sediment flux has been assessed within and between landscape units. However, much of this analysis is based on modelled energy and is unable to capture actual volumes of material moved due to temporal variations in sediment supply. The tectonically active setting of the Tongariro catchment provides a major opportunity to go one step further and build a catchment scale sediment budget, which quantifies the volumes of sediment transferred across the catchment in the past 1850 years. This is made possible due to the eruption of Lake Taupo the lake into which the Tongariro River drains, at 1.8 ka. This reset sediment process zones across the Tongariro catchment, providing a distinct landscape wide datum from which bulk transfers of sediment were quantified. This temporal dimension to sediment flux, is then combined with the spatial analysis above, and presented in Section 3.4.1.

This section initially presents the Landscape memory for the Tongariro catchment, as the evolution of sediment fluxes provides an essential foundation to support the rationale and methods of the sediment budget. Sediment budget methods are discussed, following by the results of the budget.

3.3.3.1 Landscape Memory of the Tongariro Catchment

This section discusses the evolution of the Tongariro catchment, describing the mechanisms that have formed key landforms within the catchment. This provides supporting evidence for the sediment budget.

Contemporary patterns of sediment fluxes within the Tongariro catchment are strongly influenced by the volcanic history of Lake Taupo into which the river drains. Lake Taupo is a caldera, originally formed by a rhyolitic eruption 27 ka., which shaped the northern two thirds of the lake. The most recent Hatepe eruption (also referred to as the Taupo eruption) took place 1.8 ka. and has been described as one of the “most violent eruptions of their type yet documented” (Wilson et al., 1980: 252) and the most ‘powerfully explosive’ in the last five thousand years (Wilson and Walker, 1985). It altered the atmosphere globally, with reports from China and Russia documenting ash turning the sky red (Wilson et al., 1980). This eruption deposited around 30 km$^3$ of tephra comprised predominantly of non-welded pumice over an area of approximately 20,000 km$^2$, covering the landscape with between 10 m to 15 - 30 cm distally (Manville et al., 2009). Pyroclastic flows decimated the land adjacent to the lake, depositing up to 100 m of material which travelled up to 90 km from the vent, covering most of the Tongariro catchment (GNS, 2010) (Figure 3.18).

This eruption had a dramatic effect upon the Tongariro River, as it smothered the catchment, killing all vegetation and delivering extremely high sediment loads to the system (Smith, 1991). This
effectively reset the system, providing a fixed point in time from which the river has adjusted along a new trajectory, governed by differing boundary conditions. Most notably, caldera collapse and downwarping of the northern margin altered the morphology of Lake Taupo and the base level of the Tongariro (Wilson and Walker, 1985). Initially the lake level increased to 34 m above modern levels as ignimbrite blocked the outlet of the lake. It remained at that height for ~ 20 years, before rapidly draining to a level close to present. Geomorphic response can be seen in the form of incision in the lower catchment, forming terraces (Cronin et al., 1997) and the formation of a delta downstream of these terrace which has trapped the sediment exiting the system over the past 1850 years (Rosen et al., 2002; Smart, 2005).

Figure 3.18: The extent of ignimbrite deposition from the 1.8 ka eruption showing the location of the Tongariro River and the adjacent low areas which were smothered by the ignimbrite deposition (sourced from Manville, 2002).

The eastern headwaters of the Kaimanawa Ranges have being uplifted at an average rate of 3 mm/y during the last 127 ka of the Quaternary period (Litchfield et al., 2007). This uplift maintains highly dissected, steep ‘V’ shaped valleys, with minimal sediment storage, indicating sediment generated is flushed to the Tongariro River. Figure 3.19 illustrates the steep relief in the Kaimanawa Ranges. Underlying greywacke is easily eroded creating uniform hillslopes across the sub-catchment with a
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long-term balance between uplift and erosion (c.f. Whipple and Tucker, 2002), with a smooth concave upward long profile (Mackin, 1948) (Figure 3.13A). Adams (1980) analysed uplift rates and sediment yields on the South Island of New Zealand, concluding that the Southern Alps are in steady state. He attributed ‘spiky’ mountain morphology characterised by straight slopes and sharp tops as indicative of a steady state. Similar relationships in the Kaimanawa Ranges are considered as indicative that they represent a steady state system.

In the western headwaters, the Tongariro National Park contains the active volcanic cones of Mt Ngauruhoe, Mt Pihunga and Mt Ruapehu. Whilst erosion based sediment yields off these cones are usually low, due to resilient andesitic bedrock (Figure 3.9A), they infrequently supply large pulses of sediment in the form of tephra and lahars making it particularly difficult to characterise the volumes of sediment delivered from this landscape. Eruptions have occurred on average every 2 years for the past 100 years. However, they occur in clusters, with the last 40 years being relatively quiet (Genesis Energy, 2000). Donoghue et al. (1995) identified 18 distinct tephra layers deposited within the Central Plateau over the past 1.8 ka, representing a total accumulation depth of 0.52 m, representing frequent but low magnitude eruptions. Cronin et al. (1997) mapped the distribution of lahar deposits adjacent to the Tongariro River, finding that flows post Taupo eruption were confined within the terraces and active channel, and not large enough to reach the tops of the newly incising terraces.

The middle stretches of the Tongariro River flow within an entrenched valley which has formed through the uplift of Kaimanawa Ranges and the aggradation of the Volcanic Plateau. The volcanic cones have supplied abundant volcanic sediment causing a ring plain to aggrade containing an estimated 110 km$^3$ of material (Hackett and Houghton, 1989), which includes the Volcanic Plateau unit (Figure 3.19B). Cronin and Neall (1997) document a period of high volcanic sediment productivity between 64 and 75 ka through analysis of the stratigraphy of the plateau. This coincides with a cooler period of the last glacial maximum, whereby the combination of volcanic eruptions and glacial erosion caused high rates of aggradation in the ring-plain. Growth of the ring-plain acted to trap the rivers draining the Kaimanawa Ranges, with the Waipakihi being forced to flow north along the margin between the uplifting Ranges and the aggrading ring-plain (Figure 3.19C). This forced the location of the Tongariro River. Over time, the channel has incised into its bed, setting high slopes and creating terraces. Andesitic lava flows cross the divide between the Kaimanawa Ranges and the Volcanic Plateau. Slowly, these lava tongues have been incised into, creating steep, highly confined gorges which set the base level (Figure 3.20). This history sets the planform and gradient of the Tongariro River in the mid-catchment.
Figure 3.19: A) 3-D model of topography of the Tongariro catchment. The steep dissected valley morphology of the Kaimanawa Ranges allows minimal sediment storage and is considered to be characteristic of a steady state system. This may be contrasted with the buffered, smothered landscape seen in the Volcanic Plateau. B) and C) illustrate the evolution of the system. As the volcanic cones grew, they developed a ring-plain which trapped the Waipakihi River, forcing it to flow north in a valley confined by the Kaimanawa Ranges and the ring-plain. B) and C) were sourced from Prof. P. Williams, pers. comm. (2012).
3.3.3.2 Evolution of the Lower Tongariro River since the Taupo Eruption 1.8ka

The evolution of the lower Tongariro River is discussed in greater detail here, as this reach includes two of the landscape units included in the sediment budget (the terraces and delta). This provides evidence which described the formation of these units.

Prior to the Taupo eruption, the lower Tongariro River flowed across a flat floodplain built up by lahars from the Volcanic Plateau. A lahar occurred immediately pre-Taupo eruption, as evidenced in the lack of a soil layer between pre-eruption lahar deposits and the overlying Taupo ignimbrite (Cronin et al., 1997). Once the eruption started, pyroclastic flows spread across the catchment, smothering the landscape in approximately 10 m of burning hot ash and pumice material (GNS, 2010). Drainage networks were destroyed and then re-established through the formation of rill networks cutting down into the unwelded ignimbrite (Manville et al., 2009; Smith, 1991). Initially, Lake Taupo emptied as the caldera collapsed, and incision is expected to have accelerated in response to this decrease in base level (Wilson and Walker, 1985). Lake levels increased to 30 m above present, which would have suspended this incision, until they rapidly fell to present levels. Channel adjustments can be traced back to this distinct moment in time.
Initially the channel flushed the unwelded pumice and ash material through the headward erosion of box canyons (Manville et al., 2009). No vegetation, minimal roughness and low infiltration and evaporation created high runoff, estimated to be between 80 – 90% of rainfall (Smith, 1991). This would have accelerated the speed of incision and the volume of sediment able to be flushed immediately post eruption. Manville (2009) estimated initial recovery times for channels to flush the ignimbrite of 4 – 6 years.

Once the ignimbrite layer was flushed, the channel continued to incise into underlying lahar deposits (Figure 3.22). Lahars deliver unconsolidated material comprising a range of sediment sizes. These hyper-concentrated flows are able to transport sediment larger than would be able to be delivered by the contemporary regime. Initially the finer fraction is removed, leaving a lag of the larger, in this case, boulder sized material (Gran and Montgomery, 2005) (Chapter 6). This lag slows incision, maintaining steep slopes. Gravel-sand sized sediment is rapidly flushed across this boulder lag, causing the post-Taupo eruption delta to form immediately downstream of the terraces (see Chapter 5). The low gradient delta captures all but the finest material, creating an abrupt discontinuity in slope compared with the terrace confined reach (shown in Figure 3.23). This scenario sets the context for channel adjustment described in Chapter 4.
Figure 3.22: Detrended DEM of the lower Tongariro catchment. This provides a platform to describe the evolution of this reach. B) Conceptual diagram showing the process of incision that created terraces that determine the capacity for adjustment of the contemporary channel.
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Figure 3.23: Longitudinal profile of the lower Tongariro River showing water surface measured using GPS and the bed elevation measured using a handheld sonar device. This shows the decrease in slope once the channel exits the terraces at Bridge Pool.

Over the past 1800 years smaller lahars have been flushed through the Lower Tongariro and airborne tephra has been directly deposited on the lower system. Donoghue et al. (1995) identified 18 tephra layers within the Volcanic Plateau, indicating that at least this many events have influenced the system. Smaller lahars create pulses of sediment mainly comprised of finer sand and ash material (Figure 3.24; Figure 3.25) (Manville et al., 1996). The 1995 – 96 eruption of Mt Ruapehu delivered an estimated 6900 kilotons of volcanic sediment (mostly smaller than 0.5 mm) which was flushed through the terrace confined reach to the delta within a couple of years (Collier, 2002; Smart, 2005). Steep slopes and high energy within the terrace section (Figure 3.23) ensure that these smaller eruptions have little lasting impact on the system and recovery times are swift. However, the town of Turangi is built on lahar deposits (Figure 3.22), the most recent of which are post-date the Taupo eruption (Cronin et al., 1997). This indicates that larger lahars flows of coarse material can elicit major change on the system, with much longer recovery times (Figure 3.25).
Figure 3.24: Photographs of the 2007 lahar from Mt Ruapehu which drained into the neighbouring catchment of the Whangaehu River. A) Sediment rich mixture leaving the crater lake (sourced from Galley et al., 2004) and B) Channel aggradation on the flanks of Mt Ruapehu (sourced from National Science Foundation, 2007).

Figure 3.25: Terrace exposure at Blue Pool (see Figure 3.22 for location) illustrating the chaotic unconsolidated nature of lahar deposits and the ability of a single lahar to cause metres of aggradation.
3.3.3.3 Methods used to calculate the Long-term Sediment Budget

The change in process zones following the Taupo eruption (1.8 ka) provided a platform to assess sediment transfer across the Tongariro catchment over this time. Landscape unit morphology was used to quantify sediment volumes generated (denudation measured by uplift rates in the Kaimanawa Ranges), eroded (terrace incision) and deposited (delta deposition). This sediment budget is based on the assumption that the Taupo eruption (1.8 ka) caused the formation of both the terraces and the delta. This section describes the methods used to estimate volumes and rates of sediment flux and provides evidence for these assumptions.

The Taupo eruption dramatically altered boundary conditions within the catchment. The lake existed before the 1.8 ka eruption, as it was largely formed 27 ka by the Oruanui eruption. Evidence of pre-Taupo eruption beach deposits on the northern lake margin indicate that the lake level was within +/- 10 m of the current level (Smith and Houghton, 1995; Wilson et al., 1997). The bedrock gorge of the Huka Falls 7 km downstream of the Lake outlet restricts whole-scale lake level adjustment. Following the Taupo eruption, lake levels increased to 34 m above present, creating a prominent lake terrace found between elevations of 25 – 34 m above the current level (Manville, 2002; Smith, 1988; Wilson and Walker, 1985) (Figure 3.26). Severe earthquakes, evidenced due to the presence of fissures filled with Taupo ignimbrite and large scale downwarping particularly in the north eastern areas (which is where the outlet is located) occurred in conjunction with the collapse of the caldera (Wilson and Walker, 1985). These dynamic adjustments make understanding the exact change to the morphology of the lake is complicated due to the lack of a datum to provide absolute measures of land movement (Wilson and Walker, 1985), especially due to tectonic deformation (see Section 2.4.5). The following section describes evidence of the relative changes in Lake Level at the Turangi delta.

Evidence suggests that lake level elevation decreased at the Tongariro delta, causing terrace incision and the delta to prograde. The shoreline pre-Taupo eruption is inferred to be on the Northern margin of Turangi, upstream of the State Highway (SH) 1 bridge (Figure 3.27; Figure 3.28). This comprises a lahar surface deposited between 9.8 and 10 ka (Cronin et al., 1997) (Figure 3.28B), which is inferred to demarcate the floodplain-delta surface prior to the Taupo eruption. This is supported by the presence of organic rich deltaic deposits found upstream of SH1 Bridge as shown in Figure 3.28E. Smith (1991) reported that stratigraphy 1 km upstream of SH1 Bridge had similar characteristics to the contemporary delta. A distinct step between the remnant delta terrace and the contemporary delta is indicative of lake edge erosion (Figure 3.28D). This is depicted by the 15 m
contour line on Figure 3.27 and identified as the 21 m lake terrace in Figure 3.26. This indicates that the lake level was probably 10 – 15 m higher pre eruption.

Figure 3.26: Map of terraces formed by changes to Lake Taupo’s levels post 182 eruption as created by a higher lake level after the flood as the outlet was blocked by eruption deposits (modified from Smith, 1988). Spot heights refer to elevation above contemporary lake level.

Following the Taupo eruption and the abrupt rise and then fall in lake level, the delta started to prograde beyond this delta terrace surface. Previous studies have inferred the delta to have formed post 1.8 ka. For example, Rosen et al. (2002) studied the sedimentology of the delta and Smart (1992; 2005) who estimated the rate of sedimentation following the eruption. Additional evidenced includes the lack of the Taupo ignimbrite layer in cores extracted in the delta, despite being present upstream of SH1 Bridge (Chagué-Goff et al., 1999; Chagué-Goff and Rosen, 2001; Cronin et al., 1997). Smart (1992) created a map of pre-Taupo eruption bathymetry by estimating the previous shoreline (corresponding with the upper delta surface described above) and using a seismic survey of lake stratigraphy and depth from Northey (1983) to interpolate bathymetry (Figure 3.28C). A modified version of this was used to calculate sediment input to the delta.

Due to the active volcanic nature of the region, vertical tectonic deformation of the lake margins has been observed. The Tongariro delta is estimated to have decreased by 1 – 2 mm per year relative to a survey datum on the northern lake edge (Hancox, 2002) (See Chapter 2). However, this datum is not a static landscape feature and more scientifically rigorous measures over the past 50 years have indicated marked variation in spatial and temporal rates of deformation (Otway et al., 2002; Peltier
et al., 2009). As the morphological method used is based on bathymetry and interpolation of shorelines, tectonic deformation does not impact upon these features as it occurs across much larger scales. This allows deformation to be omitted as a variable within the sediment budget.

Evidence of a decrease in lake level is supported by the incision of the channel upstream of SH1 Bridge, forming terraces on the pre-eruption floodplain-delta surface. Terrace age is supported by Cronin et al. (1997) who found that lahars pre-Taupo eruption inundated substantial areas on top of the terraces, whilst lahars in the past 1800 years have been confined within the terraces. Fluvial deposits are superimposed upon ignimbrite material from the Taupo eruption (1.8 ka) at the Sand Pool site, indicating that the channel could access these surfaces post eruption, whilst now these surfaces are inaccessible by the river (> 10 m terrace height). Lahar deposits dated immediately pre-Taupo eruption due to the lack of soil development are present within stratigraphy of the terraces. This again indicates that lahar deposits could access the terrace surface pre-eruption, but they are now disconnected from these surfaces.
Figure 3.28: Construction of the shoreline post Taupo eruption using multiple lines of evidence. A) shows detrended DEM which was used to depict the edge of the pre eruption floodplain-delta surface. B) A map of lahar inundation for the lower river consists of two surfaces, an older pre eruption lahar deposit surface and the newer delta which has reworked lahar deposits within its stratigraphy. C) Smart’s (1992) depiction of the shoreline with inferred pre eruption bathymetry. D) Distinct step between the older higher surface and the present delta and E) Bank stratigraphy above SH1 Bridge, illustrating the presence of delta deposits.

1 m horizontal resolution LiDAR data were used to characterise present terrace morphology (Figure 3.29A). The upstream extent of the terraces was defined as the Poutu intake dam due to the bedrock waterfall and gorge of the Waikato falls setting baselevel upstream (Figure 3.29B). The terrace confined reach is 20 km long, stretching from Poutu Intake dam to the culmination of the terraces at Turangi. These terraces represent the total volume of sediment that was previously stored and has subsequently been evacuated.
A TIN (Triangular Irregular Network) model was created using Arc-GIS depicting terrace height (i.e. the floodplain level pre-Taupo eruption), based on the LiDAR dataset. The contemporary channel was digitised from the LiDAR data and attributed a uniform value of 3 m, which is the average difference in elevation between the water surface and bank for the contemporary channel (Figure 3.29A). Raster calculator was used to subtract this channel from the terrace top raster, to remove this volume of sediment from the budget by assuming that a channel of a similar size would have existed pre-Taupo eruption.

Figure 3.29: A) 1 m LiDAR derived DEM with hillshade model overlain which were used to digitise river and terrace surfaces. This illustrates the ease and accuracy of delineating terrace and channel outlines due to the high level of detail generated by the LiDAR dataset. B) the Waikato Falls and gorge located at the upstream extent of A shown by *, which provides an upstream fixed point that the river grades to and the upstream extent of the terrace incision.

Terrace stratigraphy can be characterised by lahar deposits superimposed by pumice and tephra delivered from the Taupo eruption (see Cronin et al., 1997). As pumice is buoyant it is assumed that this material can be easily entrained and exit the system, with only a minimal volume stored within the delta. This is supported by Manville’s (2009) analysis of the recovery time for rivers in the central North Island following the 1.8 ka eruption (see Section 3.3.3.1). Cores extracted from the wetland in previous studies document the low presence of pumice relative to the large volumes delivered to this area (Chagué-Goff and Rosen, 2001; Cronin et al., 1997; Rosen et al., 2002). Therefore, this layer is removed from the budget by uniformly subtracting the average width from the raster. Tephra depth within terrace exposures was measured using a Laser-ace Range finder. Average depth was 2.7m (standard deviation of 1.66) with no obvious downstream trend in thickness observed. LiDAR
data were used to clearly delineate and digitise the terrace margin (Figure 3.29A). Terrace elevation and LiDAR layers were clipped to terrace extent and contemporary LiDAR DEM was subtracted from the reconstructed pre-eruption surface to calculate the volume and distribution of incision.

Sediment accumulation was calculated by estimating the volume of sediment stored in the delta. Smart (1992) created a map depicting past bathymetry based on the location of the shoreline pre Taupo eruption and interpolation of the likely slope of the lake bed (Figure 3.28C). This map was digitised, and improved shoreline locations were incorporated using the LiDAR data. Contemporary bathymetry was digitised by georeferencing a survey map created by LINZ (2006). A TIN was constructed based on the 0, 10, 30, 50, 70 and 90 m contour lines and 62 point depth measurements. This was converted to a grid raster file and merged with the 1 m LiDAR DEM of the delta region, to create a 5 m DEM combining topography of the delta and lake bed. The pre eruption DEM was subtracted from the post-eruption DEM, creating a raster of deposition volume. The data were extracted and summed, then converted to volumes by multiplying by 25 (the area of each grid cell).

Surfer 9.0 was used to create 3D models of the delta and terraces, depicting the distribution of sediment deposition and erosion.

Calculating sediment rates from the Kaimanawa Ranges involves more conjecture, based on the assumption that the system is in a steady state, with denudation equalling uplift. This section of the Kaimanawa Ranges has been undergoing steady uplift at a rate of 3 mm/y for the past 127 ka (Litchfield et al., 2007). More recent high resolution GPS work recording uplift over the last 10 years in the North Island supports this rate (Houlié and Stern, 2012). However, it is recognised that uplift is not constant, but can be punctuated, with earthquakes creating rapid increases in uplift rates and as a result, sediment generation. Therefore this component is indicative of the likely volume of material generated from the Kaimanawa Ranges, with an inherent degree of error recognised. Minimal sediment storage within this landscape unit indicates the delivery ratio from the Kaimanawa Ranges to the delta is approximately 1 (see Section 3.3.3.1). Sediment generation was estimated by multiplying the uplift rate (0.003 m) multiplied by the years the budget covers (1800) to obtain an erosion height of 5.55 m. This was multiplied by the area across which this uplift rate occurs (the area for the Steep Confined Headwaters of the Kaimanawa Ranges) to estimate overall sediment generated over this period.

Due to the high temporal variability in sediment flux from the volcanic sub-catchment, sediment input was calculated as the volume deposited in the delta, minus the input from terrace incision and
the erosion of the Kaimanawa Ranges. There are no other substantive sediment sources within the catchment. Inevitably, there is a level of uncertainty inherent within this indicative sediment budget.

3.3.3.4 Sediment Budget for the past 1850 years

Approximately 2.7 km$^3$ of sediment has been deposited within the delta since the Taupo eruption (1.8 ka) (Figure 3.30). This equates to an annual rate of 14,559,772 m$^3$y$^{-1}$ or 2.77 million tonnes per year (assuming a density of 1.9 tonnes per cubic meter (as recommended in Smart, 1992)). This number is in general agreement with Smart’s (1999; 2005) estimate of 2.6 million tonnes per year, which was derived using more simplistic, 2D methodology, incorporating greater extrapolation.

![Figure 3.30: 3-D representation of the depth of sediment deposited in the delta over the past 1850 years. The X and Y axis are coordinates in NZTM, and Z is deposition depth.](image)

Terrace incision increases upstream until the river meets the bedrock gorge of the Waikato Falls, which acts as the local base level (Figure 3.29B). High slopes within the upstream section (upstream of 5670000 on Figure 3.31) of the terraces is likely to have enabled high sediment transport and greater rates of incision. The process zone of this reach has changed from an accumulation to transfer zone, with net degradation.

The sediment volume excavated from the terraces was 0.1014 km$^3$. This represents approximately 3.8% of the sediment stored in the delta. This indicates that terrace incision and lateral reworking by the channel has not been a major contributor of sediment to the delta. Limited storage within the Tongariro Trunk indicates that the river acts as a conduit which readily flushes sediment delivered to
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it. Chapter 6 compares the size of the lahar deposits delivered from this area into the channel, finding that the lahar lag material is statistically significantly larger than the bar deposits (average D50 for the terrace is 325 mm compared with 160 mm for the bar). This indicates that the terraces are a source that delivers larger sediment than is being transported by the contemporary flow regime (See Chapter 6 for full discussion).

![Figure 3.31: 3D model representing the distribution of sediment volumes eroded from the terraces in the Lower Tongariro River over the past 1850 years. A) Depicts this from a lateral view to show relative depth of incision, whilst B) shows this from a planform view to show the comparative width of the terraces.](image)

Uplift within the Kaimanawa Ranges was identified as the major process generating sediment, with an estimated 1.5 km$^3$ of sediment generated from this landscape unit over the past 1850 years. This represents approximately 55% of the sediment volume in the delta (Figure 3.32). This equates to 813,447 m$^3$ (or approximately 1,561,876 tonnes) per year. Minimal accommodation space is available to store sediment within this landscape unit due to its steep valley sides and confined valleys. Only limited stores are present as lateral bars within the larger Waipakihi and Whitikau trunk streams, ensuring that sediment generated from this unit is transported to the Tongariro relatively rapidly. The high connectivity in the mid catchment and the small sediment size from the Kaimanawa Ranges allows this material to be flushed though the catchment to the delta.

The remaining sediment within the delta is likely to be generated by infrequent eruptions and lahars from the volcanic sub-catchments to the west. This includes 1.1 km$^3$ over the past 1850 years, comprising 41% of the material in the delta (Figure 3.32). Reducing this to an annual rate is meaningless due to the stochastic nature of sediment input from this source.
Whilst sediment storage and availability is high within this unit the nature of channels incised into gully features disconnects the channel from these stores. Volcanic input has a low frequency, but a high magnitude. 18 moderate sized eruptions have occurred since 1.8 ka (Donoghue et al., 1995), each representing an influx of sediment. Following the Taupo eruption (1.8 ka) the catchment redistributed the tephra material which covered the catchment, creating an initial pulse of sediment (Gran and Montgomery, 2005; Manville et al., 2009). Sediment yields from the Volcanic Uplands are likely to be low between volcanic events. As a result, sediment yields from the volcanic subcatchment are highly variable, but storage is prominent and disconnected on lower slopes. It is inferred that sediment supply between eruptions is mostly volcanic sand and smaller material which may be transported off slopes and through the Volcanic Plateau. Material from gully erosion may also be transported.

Figure 3.32: Morphological Sediment budget describing the bulk transfers of material across the Tongariro catchment following the eruption of Lake Taupo 1.8 ka. Methods are described in full in Section 3.3.3.3. Red arrows indicate volumes of sediment generated and blue arrows sediment deposition. All sediment volumes are in km$^3$ over 1850 years.
Smart (2005) suggests that contemporary rates of sediment input are likely to be lower than immediately post eruption, when sediment availability and associated reworking was very high (Smith, 1991). Terrace incision was likely rapid post eruption, due to a lower lake level inducing erosion though head cuts into ignimbrite (Manville, 2002). As incision continued into lahar deposits, finer sediment (i.e. sand and small gravels) would be transported, whilst large material (i.e. cobbles and boulders) would remain in the system, creating the basement of lahar lag present today. This is reworked infrequently, limiting the rate of incision (Chapter 6). As a result, sediment supply from terraces is likely to be less than during the initial period of formation. The long profile shows an abrupt change in slope on the boundary of the terraces and delta (Figure 3.23). However, records of bed level elevation taken from two flow gauging sites within this reach, which indicate minimal linear bed adjustment during the past 50 years (Smart, 1999) (Chapter 4). This supports the role of the lahar boulder deposits in maintaining steep slopes throughout this reach.

In comparison, the steep headwaters of the Kaimanawa Ranges provide a more constant source of sediment, due to the perceived constant rate of uplift in this region. However, it is recognised that uplift can be punctuated by rapid bursts, often caused by earthquakes. Thus, this catchment may have undergone rapid increases in sediment associated with bursts of uplift. The high connectivity of the mid-catchment reaches and the relatively small size of sediment generated by the Kaimanawa Ranges means that these materials can be easily transported through the catchment to the delta.

In light of these considerations, the majority of sediment moving through the system at present is expected to come from the Kaimanawa Ranges (Figure 3.32). This equates to an annual input of 813,447 m$^3$ which is flushed to the delta.
3.4 DISCUSSION

3.4.1 Patterns of Connectivity across the Tongariro Catchment

Schematic representations of sediment transfer relationships in the Tongariro catchment are shown in Figure 3.33 and Figure 3.34. Initially, the connectivity within each landscape unit is discussed. This is followed by a synthesis of the connectivity of sediment flux across the Tongariro catchment as a whole.

Figure 3.33: Patterns of sediment flux across the Tongariro catchment showing the distribution of sediment generation and storage. Drainage network is shown in terms of its transport capacity and connectivity through the different units.
3.4.1.1 Connectivity of Landscape Units

The Kaimanawa Ranges were found to be the greatest contributor of sediment to the delta (Figure 3.32). Hillslope-channel coupling is highly effective in the steep, dissected terrain of the upper catchment, and large volumes of sediment are delivered to the valley floor (Figure 3.33). Active uplift of soft greywacke rock sustains high sediment yields of approximately 813,447 m$^3$ per year. Given limited sediment storage on the valley floors, this material is delivered directly into the stream network and beyond. The larger trunk streams draining the Kaimanawa Ranges the Waipakihi and Whitikau Streams, have gentle gradients (mean of 3.0°), greater in-channel storage and lower mean boundary shear stress (31 and 50 N/m$^2$), acting to marginally slow the movement of gravels through these reaches. However, despite this, the Kaimanawa Ranges sub-catchment represents a highly
connected, efficient sediment generating machine, which represents a steady input of sediment into the system over the past 1850 years (Table 3.4).

The volcanic sub-catchment to the west represents irregular and dramatic influxes of sediment to the system. The Kaimanawa Ranges provides a fixed boundary, limiting the accommodation space that is available to store sediment generated from the volcanic cones (Figure 3.19). The forced nature of the storage unit of the Volcanic Plateau results in mass aggradation, smothering the underlying topography. As such, the unit is characterised by homogeneous features and low slope.

Actively reworked stores are evident on the boundary of the Volcanic Uplands and the Volcanic Plateau, as the rapid decrease in slope acts to capture material generated on the steeper terrain upstream (Figure 3.33; Figure 3.34). This is illustrated by the change in channel planform for the Mangatoetoenui Stream, the stream most recently influenced by lahars (1995-96), as a localised section of the channel becomes braided where slopes decrease and fine-grained material is dumped. However, the channel rapidly forms the single planform, gully confined, cobble bed material channel characteristics observed in the channels draining the Volcanic Plateau (Figure 3.35). Gullies cut into this feature have low – moderate connectivity shown by low stream power and channel slope. As a result, moderate sediment transport would be expected in this region. Flat interfluves between the gullies have very low slopes (< 4° indicating sediment storage zones) and drainage areas (< 0.02 km²). As a consequence, these streams have limited capacity to generate and transport the vast stores of material held within these features (Table 3.4). Thus, the plateau acts as major sediment sink, characterized by long sediment residence times.

The role of the plateau as a sediment sink is supported by the preservation of tephra stratigraphy dating from 75 ka to present (Cronin and Neall, 1997; Donoghue et al., 1995). This unit acts as a buffer, impeding the transport of sediment across its length and disconnecting the Volcanic Uplands from the Tongariro River between volcanic events (Fryirs et al., 2007a). However, volcanic eruptions alter sediment transport pathways, and are able to move sediment through this buffer and into the Tongariro system. Tephra and airborne deposits can be delivered directly into the stream network on the plateau and to the lower Tongariro River. Lahars act as hyper-concentrated pulses transferring material along the stream networks and into the lower catchment (Cronin et al., 1997; Cronin and Neall, 1997). Whilst volcanically sourced sediment is expected to comprise 45% of the sediment delivered to the delta (excluding lahar deposits reworked from the terraces), delivery of this material is not linear over time (Figure 3.32). Rather, it exhibits a marked stochasticity, with infrequent pulses of high sediment load. River recovery from tephra and ash deposits occurs rapidly
(Manville et al., 1996). In contrast, coarser materials from lahars may persist within the system over much longer timescales (Cronin et al., 1997; Cronin and Neall, 1997).

The Tongariro Trunk stream acts as a highly connected booster of sediment. Moderate slopes combined with high drainage area and limited accommodation space, creates high stream power and transport capacity. This is reflected in steep cascade and riffle units. Bed material within this section is coarse, with a lag of lahar deposits lining the channel. Despite the large grain size within the upper catchment, the high energy conditions are easily able to transport the median bed material, with shear stress ratios of between 1.44 and 2.56 predicted for the mean annual flood. Much of the sediment delivered from the Kaimanawa Ranges and the Volcanic Plateau is able to be flushed through this reach due to its smaller grain size. As a result, minimal sediment stores are present. This is especially true for trunk streams of the Waipakihi and Whitikau Streams that drain the Kaimanawa Ranges, where comparatively smaller material ($D_{50}$ of 61 mm compared to a $D_{50}$ in the upper Tongariro River of 230 mm) is easily transported across this lahar basement (see Chapter 6). In addition, no major changes in River Style or sediment storage was evident downstream of tributary junctions (c.f. Rice et al., 2008). Due to the high transport competence of the Tongariro Trunk, tributaries within the Tongariro Trunk are minimally geomorphically effective (defined by the potential of tributaries to geomorphically influence the trunk channel) (Benda et al., 2011).

Gorges formed by incision into lava flows provide fixed points to which the river grades, maintaining steep slopes throughout the Tongariro Trunk reach (Mackin, 1948). This is evident in the noticeably flatter slopes before the river flows through the first bedrock gorge at the upstream extent of the Tongariro River (32 km downstream on Figure 3.13A). Low slope is reflected in the presence of active
in-channel stores as lateral bars (up to 300 m wide at the widest point) and low boundary shear stress values (31 N/m²) for the Waipakihi River. This base level control disconnects the upper to the lower catchment, providing a barrier to sediment transport (Fryirs et al., 2007a). Unlike the total sediment disconnect observed in the Trinity River by Phillips et al. (2004), the disconnect in the Upper Tongariro acts to slow the rate of sediment transfer through this section, rather than stop it entirely. This assertion is supported by the lack of significant sediment stores in the upper Waipakihi, where the active channel comprises the entire valley width. As such, the lower Waipakihi acts as a transfer zone, transporting sediment less efficiently than the steep first order streams and the Tongariro River downstream.

Between these mid-catchment gorges, the lower course of the Tongariro River has incised into volcanic materials, creating a partly-confined valley that is inset between terraces (Fryirs and Brierley, 2010). These terraces are long-term sediment sinks and the sediment contained within them is effectively disconnected from the contemporary regime (c.f. small floodplain pockets are reworked). Terrace width limits the accommodation space available to store sediment. This section operates as a transfer zone, efficiently transferring the sediment delivered into the lower catchment. Whilst the dams may increase the jerky nature of sediment input to the lower catchment, their overall influence on the transfer of sediment flux into the lower catchment is minimal (see Chapter 4 for further details; Reid et al., In Press).

The Rolling Foothills and Terrace-lands play a minimal role in generating or distributing sediment flux through the catchment. The Taupo eruption (1.8 ka) deposited a thick layer of ignimbrite on these units, which remains there today. This acts to dampen these landforms, creating moderately low gradients (9.3° and 6° for the Rolling Foothills and Terrace-lands respectively). These units are accordingly less able to generate and supply sediment, reducing lateral connectivity. This is especially evident within the Rolling Foothills unit. Although hillslopes are well connected to the channel in this landscape unit, the persistence of the Taupo ignimbrite layer reduces its sediment generating ability. The Rolling Foothills has moderate - low total stream power (3122 Watts for a mean annual flood of 2 m³s⁻¹) and channel slope (4.4°), indicating a moderate capacity to flush the sediment delivered. Low accommodation space and minimal alluvial sediment stores within these valleys indicate that this moderate stream power is sufficient to transport the volumes of sediment delivered to this landscape unit. In the Terrace-lands, incision into Taupo ignimbrite has disconnected channels from reworking adjacent floodplain surfaces, creating confined valleys with limited accommodation space. Shear stress ratios of between 0.87 and 1.37 indicate moderate – low longitudinal connectivity whereby the channel is competent to flush its median grain size during the
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mean annual flood (Table 3.4). Limited sediment stores within this zone again indicate that despite low channel slopes (average of 3.1°) most of the sediment delivered is able to be transported to the Tongariro River.

Located immediately downstream of the terraces, the delta represents the total accumulation zone for the catchment. Its small size reiterates the short time frame since the catchment has redistributed its process zones. The former accumulation zone prior to 1850 years ago is now disconnected from the contemporary channel by terraces (Table 3.4). 2.77 million tons of sand and gravel have been delivered to this reach since the Taupo eruption, 1.8 ka. Due to the highly connected nature of the mid-catchment, once sediment is delivered to the conveyer belt it is likely to be rapidly transported into this storage unit. Channel adjustment within the delta is driven by this influx, and the low gradient of the reach. As a consequence, this zone experiences substantial geomorphic readjustment (see Chapter 4).

3.4.1.2 Synthesis of Connectivity across the Tongariro catchment

The Volcanic Uplands is a steep landscape unit which provides a major source of sediment. Hard andesitic rock limits the rate of erosion (Table 3.4). However, very high volumes of loose volcanic material is available for reworking within this zone, especially at the downstream boundary of the unit and frequent volcanic eruptions generate sediment. Mt Pihanga to the north has noticeably less loss sediment due to a longer time elapsed since the last eruption, and has a low sediment supply. Steep slopes create a highly connected unit, which can mobilise high sediment volumes. The abrupt decline in slope where the volcanic cones meets the volcanic ring-plain creates a zone of significant sediment storage and reworking (see Figure 3.35A; Figure 3.33). The decrease in channel gradient into the Volcanic Plateau acts to buffer the transfer of this sediment though this low slope unit, and into the Tongariro Trunk Stream (Fryirs et al., 2007a). Low slopes on the Volcanic Plateau make this unit a sediment sink, unable to access and transfer the vast sediment stores to the channel network (Figure 3.34; Table 3.4). In contrast, the streams draining the Kaimanawa Ranges are steep, highly connected, and able to generate high quantities of sediment (Table 3.4). Thus, they represent a major source of sediment into the Tongariro Trunk. Trunk streams of the Waipakihi and Whitikau have gentler gradients, increased in-channel storage as lateral bars, and only moderate channel connectivity (Figure 3.33). This reflects a slowing of sediment through this section of the catchment.
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Table 3.4: Overview of connectivity and sediment generation capacity and storage elements for each landscape unit. Cell colour indicates the connectivity of a variable, with green cells indicating high connectivity, orange moderate connectivity and red low connectivity.

<table>
<thead>
<tr>
<th>Andean Uplands</th>
<th>Slope-channel connectivity</th>
<th>Longitudinal connectivity</th>
<th>Sediment Generation</th>
<th>Stores or sinks present?</th>
<th>Connectivity of stores?</th>
</tr>
</thead>
<tbody>
<tr>
<td>Volcanic Uplands</td>
<td>Steep slopes drain into well-connected gullies. High ability to transport sediment.</td>
<td>Highly connected within unit due to steep streams, though sediment transport is intermittent due to low drainage area creating ephemeral streams. Sediment deposited when slope decreases at downstream extent of unit at the boundary with the Volcanic Plateau.</td>
<td>Despite low suspended sediment rates, abundant loose scoria and ash are present which are moved as bedload. During eruptions large volumes of ash and sediment are flushed through this unit. Note: Mt Pihanga is an exception and has notably lower sediment yields.</td>
<td>Actively reworked stores at the downstream margin with the Volcanic Plateau</td>
<td>Zone of storage actively reworked at the foot the largest volcano (Mt Ruapehu) but these zones are less active at Ngauruhoe and Pihanga.</td>
</tr>
<tr>
<td>Volcanic Plateau</td>
<td>Flat plateau deposits act as sinks as channels have incised to form gullies which are disconnected from the plateau.</td>
<td>Incised streams form gully features which cut across the plateau. Able to flush smaller sediment generated from uplands over a cobble stream bed consisting of lahar deposits.</td>
<td>Minimal. Local reworking of deposits on gully margins.</td>
<td>Whole landscape is a large sediment sink.</td>
<td>Disconnected as gullies are incised and do not laterally migrate.</td>
</tr>
<tr>
<td>Steep headwaters Kaimanawas</td>
<td>Steep slopes readily deliver sediment directly to the channel.</td>
<td>First order streams steep and well connected. Trunk streams in the Waipakihi exhibit large reworked bars indicating only moderate connectivity.</td>
<td>Active uplift (3 m m⁻¹) and soft greywacke drives high generation of sediment, estimated annual load of 813,447 m³.</td>
<td>Active bar surfaces in the trunk streams act as short term stores of sediment.</td>
<td>Connected and readily reworked meaning transport is slowed but not stopped.</td>
</tr>
<tr>
<td>Tongariro Trunk</td>
<td>Channel disconnected from stores by flat terraces. Connected to within terrace features as floodplain pockets, bars and islands are readily reworked.</td>
<td>Basement of lahar lag retains a steep channel, readily able to flush sediment supplied. Localised gorges have an increased capacity to flush sediment.</td>
<td>Low through slow incision of gorges and terrace incision/erosion.</td>
<td>Channel has incised and this feature is surrounded by old floodplain, representing a long term sink.</td>
<td>Sink disconnected from channel as terrace incision means the channel cannot inundate the floodplain and the flat topography of the terrace results in minimal sediment transport across this surface.</td>
</tr>
<tr>
<td>Rolling foothills</td>
<td>Moderate as no stores impede transport and moderate slopes.</td>
<td>Moderate as moderate channel slope and stream power.</td>
<td>Low sediment generation as much of this area remains covered in tephran deposits from the Taupo eruption decreasing erosion.</td>
<td>None.</td>
<td>N/A</td>
</tr>
<tr>
<td>Terrace-lands</td>
<td>Flat terrain covered with tephran creates limited connectivity.</td>
<td>Low - moderate, with flat slopes and limited drainage area.</td>
<td>Low. Slightly higher suspended sediment generation but little morphological impact.</td>
<td>Terrace is a sediment sink.</td>
<td>Small streams have incised into lahar deposits to meet the base level of the Tongariro, disconnecting these stores from being reworked.</td>
</tr>
<tr>
<td>Lowland Plain</td>
<td>Flat terrain so little connectivity.</td>
<td>Progressive decrease in slope along this unit traps sediment, until only finest fraction is able to be transported to the mouth.</td>
<td>Low.</td>
<td>Delta is a sediment sink capturing sediment as it exits the terraces.</td>
<td>Well-connected sink which is actively aggrading and reworked by channel adjustment and avulsion.</td>
</tr>
</tbody>
</table>
In the mid-lower catchment large stores of sediment are tied up in sinks, which are disconnected from the channel (Figure 3.33). In the upper catchment this is due to streams incising into materials forming gullies in the Volcanic Plateau, whilst in the lower catchment, streams have incised into ignimbrite deposited during the Taupo eruption (1.8 ka), converting previously accessible floodplains into inaccessible terraces. Streams that drain these deposits have moderate-low connectivity, driven by low channel gradients. The Rolling Foothills has moderate slopes. However, the ignimbrite layer blanketing this unit limits the amount of sediment that is generated (Table 3.4). However, steeper channel slopes allow moderate connectivity of sediment delivered. The highly connected Tongariro River runs through the middle of these sinks. It is readily able to flush much of the smaller sediment delivered from tributaries. Lateral connectivity is limited to the narrow zone within the terraces, where limited in-channel stores are present at lateral bars which are frequently reworked (Table 3.4). In the lower Tongariro, the terraces widen, and channel competence was found to become varied and more frequently lower, as the channel has more space with which to adjust its form and store sediment. Competence declines rapidly downstream of the terraces, with sediment dumped in an active braided reach. The sediment sink of the Lowland Plain remains connected, though a low gradient limits the movement of sediment across this surface and into the channel (Figure 3.33). Rapid downstream fining across this unit reflects the decline in sediment transport capacity, illustrating the extent to which this unit is able to capture all but the smallest sediment generated and transported through the rest of the catchment.

3.4.1.3 Patterns of Sediment Flux across the Tongariro Catchment

Conceptual models of catchment sediment flux suggest that catchments either operate as well connected steady states, which are graded to deliver sediment from the headwaters to the mouth (Adams, 1980; Davies and Korup, 2010; Reneau and Dietrich, 1991; Whipple and Tucker, 2002) or disconnected systems where storage elements act to buffer sediment delivery, effectively decoupling source and accumulation zones (Dearing and Jones, 2003; Métivier and Gaudemer, 1999; Phillips, 2003b; Trimble, 2009). Essentially these models describe whether the catchment has adjusted to transport the load generated, or whether on-going perturbations or landscape features hinder the establishment of a graded profile.

Stark contrasts in the character of the eastern and western halves of the Tongariro catchment can be explained by these alternate models. The Kaimanawa Ranges operate as a steady state system where sediment generation is driven by steady rates of uplift during the late Quaternary (Litchfield et al., 2007). Few zones of storage between the headwaters and the Lowland Plain are present, indicating a sediment delivery ratio of around 1 (c.f. Walling, 1983; Wei et al., 2006).
In contrast, sediment transport through the volcanic catchment is impeded by the significant sediment sink and flat slopes that make up the Volcanic Plateau, supporting the theory that long-term sediment yields are regulated by sediment stores (Phillips, 2003b). A major control driving this difference is more frequent, high magnitude landscape forming events in the volcanic sub-catchment. This results in a landscape unit that is always recovering from extreme events, compared with the more constant rate of change observed in the Kaimanawa Ranges. The contrasting tectonic controls of uplift in the east and volcanism in the west creates geologically induced differences in accommodation space. The uplifting Kaimanawa Ranges have minimal accommodation space, as steep slopes drive incision and flush sediment. Immediately to the west of the fault boundary, minimal uplift and high sediment supply from the growth of the volcanic cones has created a large accommodation space, as many layers of volcanic sediment build up to form a near-featureless sediment sink. Thus, high sediment supply has created low slope and forced a high accommodation space.

The role of sediment stores in regulating sediment yield has been well documented, through either increasing storage during times of high supply, or acting as a source of sediment during periods with lower yields (Dearing and Jones, 2003; Métivier and Gaudemer, 1999; Phillips, 2003b; Trimble, 2009). Notable examples include Trimble (2009) who documented the role of stores in capturing high loads generated from agriculture, resulting in consistent sediment yields despite large fluctuations in supply. Wilkinson and McElroy (2007) describe the deposition of post-settlement alluvium as the most “important geomorphic process...shaping the landscape of the earth” (Wilkinson and McElroy, 2007: 140). Church and Slaymaker (1989) identified glacial stores as contributing on-going high sediment loads to rivers in Canada. Very few studies relate landscape scale sediment storage units created by volcanism to contemporary sediment yields. Instead, most studies deal with sediment yields directly following eruptions (Gran and Montgomery, 2005; Manville, 2002; Simon, 1992). Smith and Swanson (1987) discuss sediment storage following the Mt St Helens eruption in 1980, finding that approximately 88 % of tephra remained in storage on hillslopes. The landscape and airborne form of delivery is reflected in hillslope mantling, rather than discrete storage in sediment sinks. Thus the influences on the drainage network are distributed across the catchment, rather than concentrated within a single major landscape unit as seen in the Volcanic Plateau. Palmer (1991) describes phases of aggradation and dissection and erosion of the Tongariro ring-plain during alternating phases of high and then low volcanic activity. Fryirs et al. (2007a) view stores as landscape features which impede the sediment transport across
Chapter 3: Catchment-scale sediment flux

contemporary landscapes, disconnecting hillslopes from channels. Thus, the patterns of stores in a landscape and catchment provide a major influence on contemporary patterns of sediment transfer.

The Tongariro catchment has vast stores of sediment in the Volcanic Plateau and the terraces, with the ring-plain estimated to store 110 km$^3$ of material (Hackett and Houghton, 1989). However, most of these stores are decoupled from the channel network due to terrace incision in the mid catchment and incision forming gullies in the Volcanic Plateau. This decoupling of stores and channels limits their ability to buffer or influence sediment transfer, as channels cannot reach these features to deposit sediment, while low slopes inhibits generation or transport of sediment across their surfaces. Fryirs et al. (2007a) describes terraces as buffers, which retain sediment for long residence times (> 10$^3$ years), and are considered external to the channel network. However, as not all stores are connected to the channel network, and therefore able to play a role in ‘buffering’ sediment transport, the nature of this relationship needs to be explored. Whilst the sediment sinks of the terraces and the Volcanic Plateau are effectively ‘disconnected’, the reworked stores at the foot of the Volcanic Uplands and the sink of the Lowland Plain remain connected to the fluvial network, as they are able to be reworked to supply or store material (Figure 3.34). Understanding the role of stores on catchment patterns of sediment flux requires interpretation of the (dis)connectivity of these storage elements, and consideration of the types of on-going interactions between stores and the drainage lines that dissect them.

The lack of connected sediment stores within the mid-catchment of the Tongariro results in more dynamic, responsive patterns of sediment flux. Once sediment is delivered to the conveyerbelt of the Tongariro Trunk stream it is likely to be flushed through this mid-section with relative ease. This is supported by the high boundary shear stress values (between 123 and 364 N/m$^2$) within this middle section, and the lack of accommodation space within the terraces which can store sediment. Dearing and Jones (2003) analysed sediment archives, finding that small basins (< 1000 km$^2$) underwent greater variation in sediment flux with faster reaction times, due to greater spatio-temporal coupling. In contrast, Reneau and Dietrich (1991) found that basins between 1 and 1500 km$^2$ exhibited spatially uniform erosion rates and a steady state between hillslope erosion rates and sediment yield. The Tongariro displays a trend more similar to the Dearing and Jones model. Limited connected stores which buffer the drainage network, mean that once sediment is delivered into the drainage network, and particular the Tongariro itself, it is readily mobilised and transferred to the delta. The trunk streams draining the Kaimanawa Ranges, the Waipakihi and Whitikau Streams, present an exception to this, as their gentler gradients, wide channels and low mean boundary shear
stresses (31 and 50 N/m$^2$ respectively) slow the movement of sediment through these sections of the catchment (Figure 3.33).

Temporal patterns and rates of sediment reworking vary for differing landscape units. The Kaimanawa Ranges are inferred to be undergoing steady uplift over the last 127 ka (Litchfield et al., 2007). This is likely to generate relatively consistent patterns of sediment supply. Limited storage within the Kaimanawa Ranges supports this. In contrast, sediment delivery from the Volcanic sub-catchment is driven by high magnitude, low frequency events including lahar flows and airborne tephra deposition. 18 moderate sized lahars have inundated the Volcanic Plateau in the last 1800 years (Donoghue et al., 1995). This represents pulsed inputs of sediment which the channels rework and adjust to over long timeframes. Whilst smaller material may be flushed rapidly, larger clasts remain in the system (Gran and Montgomery, 2005). Terrace incision provides the final sediment source. As discussed above, this is inferred to have been rapid initially, slowing once finer sediment had been removed (Manville, 2002; Smith, 1991). Intermittent lahars flow though the terraces reinitiating the process of recovery, whilst the coarser fraction is reworked and transported across longer timescales (1 - 100 years) (Gran and Montgomery, 2005). These findings illustrate the important of incorporating the temporal component of sediment flux when considering patterns of sediment sources. This allows likely future changes to sediment sources to be conceptualised.

Catchments are commonly represented by a linear downstream transition in process zones which go from source to transfer to accumulation zones as accommodation space and sediment storage increases (as represented by Church, 2002; Schumm, 1977). Sediment transfer within the Tongariro is at odds to this pattern, with the large storage zone of the Volcanic Plateau located in the upper reaches of the catchment. In addition, limited instream or floodplain sediment storage in the lower catchment is also at odds to the commonly conceptualised pattern of increasing accommodation space and sediment storage downstream (Brierley and Fryirs, 2005; Schumm, 1977). This supports the premise that many catchments are characterised by patchy and discontinuous patterns in sediment generation, transport and storage (Phillips, 1999; 2003a). Therefore simple linear models of sediment flux are unable to capture the variables driving or impeding sediment transport, especially in highly disturbed landscapes.

3.4.2 Sediment Budget as a Tool for measuring Sediment Flux

The sediment budget provided a quantitative framework measuring the volumes of sediment moving through landscape components. Previous studies that have used the morphology of units to quantify sediment deposition have concentrated on quantifying sediment transport into a single unit

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(i.e. delta) not relating this back the sediment transport paths through the catchment (Goodbred and Kuehl, 1999; Walling, 1983). For example, Goodbred and Kuehl (1999) used floodplain and delta morphology to calculate that a third of suspended sediment delivered from upstream was deposited on these units. Other morphological sediment budget approaches have been based at the reach-scale (c.f. Brasington et al., 2000; Brasington et al., 2003; Fuller and Basher, 2012; Ham and Church, 2000). This chapter linked deposition volumes back to the relative volumes of sediment derived from the different sources. This was enabled by framing the sediment budget within a conceptual framework of connectivity and process zones, effectively clarifying the processes within the ‘black box’ which is catchment scale sediment transport (De Vente et al., 2007; Fryirs, 2012; Walling, 1983).

Application of landform scale sediment budgets are commonly constrained by the lack of a continuous datum (Houben et al., 2006). Tephra layers within the Tongariro catchment provided a distinct landscape wide marker that provided a continuous datum. This provides a distinct point in time from which the catchment has adjusted upon its current trajectory.

Sediment budgets inherently incorporate a level of uncertainty. This is especially true for budgets estimating sediment volumes across wider spatial (catchment) and temporal (> 100 year) scales. Main sources of uncertainty within this budget can be seen in the assumption that the Kaimanawa Ranges is operating as a steady state system, undergoing a steady rate of uplift over the past 1850 years. The exact shoreline and bathymetry of Lake Taupo post eruption also involves a level of interpolation. Hence, these volumes need to be viewed as approximations.

3.4.3 Connectivity

Connectivity provided an overarching framework with which to conceptualise sediment transport. Fryirs et al. (2007b) present a simple conceptual model describing the influence of buffer, barrier and blankets in impeding sediment transport across a catchment. However, within the Tongariro catchment, identification of these storage features was a less important variable due to minimal sediment storage units. Instead, the ability of hillslopes to generate and transport sediment, and the ability of the stream network to transport it was a more appropriate measure. This is relevant for many New Zealand landscapes, which are highly dissected by small streams, with high sediment generation capacity and minimal accommodation space where sediment is able to be stored (Davies and Korup, 2010; Fryirs et al., 2007a; Kasai et al., 2005). The tools used in this thesis provided a more comprehensive approach for describing catchment-scale connectivity in active landscapes.

Stream power or stream power based derivatives (i.e. the Erosion index) present a commonly used technique for analysing sediment transport capacity for both drainage lines and hillslopes (Brierley
and Fryirs, 2005; Finlayson and Montgomery, 2003; Phillips and Slattery, 2006; Whipple and Tucker, 1999. The parsimonious nature of this approach (i.e. simpler theories are based on few assumptions and less error) represents its strength. It also allows it to be applied easily across wide spatial scales, with minimum data required (Finlayson and Montgomery, 2003). Slope also presented a simple, yet effective measure of connectivity. Identification of slope categories that corresponded with sediment storage and generation were effective at delineating process zones within the catchment, and yielded results consistent with geomorphic observations. This approach mirrors that applied by Fryirs et al. (2007b) who applied slope ‘thresholds’ to identify buffers and calculate the effective catchment area. However, if connectivity is to consider volumes of sediment generated rather than locations of erosion, geology and landscape history must be included. This was supplied through analysis of sediment yields and erosion terrains (Dymond et al., 2010; Hicks et al., 2011). For example, this captured the high suspended sediment yield from the Steep headwater reaches, due to the soft greywacke eroding high loads of fine grain sediment. This illustrates the need to incorporate landscape history and the processes generating sediment to support these simplistic measures.

This study highlights the need to combine a conceptual framework of rivers and landscape morphology with quantitative measures of the transport capacity (c.f. Small and Doyle, 2012). For example, qualitative assessment of the Volcanic Plateau indicated that it had limited sediment generation and transport capacity. However, more in-depth analysis identified that individual streams crossing the plateau had variable transport capacity, related to the time since last inundation by lahar flows. For example, Mangatoetoenui Stream had a greater ability to transport its load (shear stress ratio of 2.25) due to smaller grain size and steeper slopes attributed to aggradation by lahar debris. Channels which are disconnected from recent lahar paths tended to have low shear stress ratios (e.g. 0.35, 1.01 and 0.46 for the three streams north of the Mangatoetoenui Stream). Measures of connectivity should be selected to measure sediment transfer across a range of spatial and temporal scales, to capture variability in processes across landscapes.

3.5 CONCLUSION

A conceptual framework was developed from the literature by ascertaining key variables and concepts which influence patterns of sediment flux. This includes connectivity notions, the influence and distribution of process zones of sediment generation and storage and appraisal of simple measures of transport capacity for each unit (i.e. slope and stream power), linked to assessment of SSY and erosion terrains. Representative sites were used to collect field data to describe the
character and behaviour of the types of rivers and measure shear stress and bed material across the
catchment. This grounded the desktop analysis. A catchment wide sediment budget determined the
transfer of sediment across the past 1850 years, providing a temporal dimension to sediment
transport as well as quantifying bulk volumes transferred. This toolkit generated substantive insight
into sediment fluxes across the Tongariro catchment.

Contemporary patterns of sediment flux within the Tongariro catchment are driven by its volcanic
history. Differential landscape histories have created a stark difference between the eastern and
western sub-catchments, which act to determine different capacities to generate and transport
sediment. The eastern Kaimanawa Ranges is acting as a steady state system, whereby uplift
generates steep slopes and high loads of greywacke. Eroded materials are rapidly moved through
the system due to minimal accommodation space to store sediment. In contrast, the volcanic
western catchment generates high sediment loads sporadically. Lahar deposits have blanketed the
landscape, creating a flat featureless sediment storage feature of the Volcanic Plateau. Sediment
stores are largely disconnected from the drainage network as the streams have incised into gullies
and do not rework this surface. The evolution of these two sub-catchments pins the Tongariro River
into a narrow, entrenched valley, with limited accommodation space and high longitudinal
connectivity. The delta, at the downstream extent of the catchment, captures sediment flushed
through the catchment.

This study illustrates the strength in combining analytical and conceptual approaches to frame
sediment flux, whereby quantitative measures need to be supported by a geomorphic based
conceptual framework to strengthen interpretation of processes. The marked difference in sub-
catchment character within the Tongariro presents a complex and dynamic system where
contemporary sediment fluxes and are undeniably tied to past landform development.
Chapter 4: Spatial and Temporal patterns of Reach-scale adjustment

4 REACH SCALE GEOMORPHIC ADJUSTMENTS OVER THE LAST 80 YEARS

4.1 INTRODUCTION

“One cannot ever forget the presence of the Tongariro...Green-white thundering Athabasca river of New Zealand”

(Grey, 1978: 58, written during a visit in 1927)

The reach provides the scale at which river management is most commonly applied (Hoyle et al., 2008). For this reason, it is pertinent to understand how a channel is likely to adjust at this scale. This is especially true as the types and rates of adjustment (referred to herein as channel sensitivity) and channel form change downstream, in response to alterations in reach scale controls, most notably slope and sediment flux (Ferguson, 1987; Leopold and Wolman, 1957). For this reason, channel form derived from aerial photography and survey maps is commonly used to quantify river response to changes in prevailing fluxes (Fryirs et al., 2009; Gurnell, 1997; Hooke, 2007; Richardson and Fuller, 2010; Surian, 1999; Winterbottom, 2000). This allows evolutionary trajectories describing the direction the system is moving to be determined (see Section 1.4) and the sensitivity of individual reaches to be assessed (Brierley and Fryirs, 2008; Dufour and Piégay, 2009; Fryirs et al., 2012; Surian et al., 2009b). Fryirs et al. (2009) describe ‘response gradients’, which assess channel change as the sequence and timing of channel response across the catchment. Such analyses provide insights into thresholds, historical contingency and complex responses that may be inherent to the reach or system (Phillips, 1999; Phillips, 2012; Schumm, 1980; Toone et al., in press). For example, the role of event sequencing can be determined, whereby the history of the system can alter reach sensitivity to a subsequent event, creating marked variability in the geomorphic effectiveness of similar flood events (Beven, 1981; Marutani et al., 1999; Toone et al., in press). These inherent complexities illustrate the importance of determining reach-specific past responses to predict likely future adjustment (Lane and Richards, 1997).

The ease with which a channel adjusts reflects the character and behaviour of a river. This supports the need to determine the ‘natural capacity for adjustment’ of each river type by assessing the “…range of process activity that is possible for that setting” (Brierley and Fryirs, 2005: 144). Meaningful differentiation can be made between river behaviour and river change (Brierley and Fryirs, 2008). River behaviour describes “… adjustments around a characteristic assemblage of geomorphic units over timeframes of tens to hundreds of years” (Brierley and Fryirs, 2005: 145). Process-form relationships remain the same, and morphological adjustments define the behavioural regime for that type of river. In contrast, river change describes a system that has undergone a shift in
biophysical flux or boundary conditions, such that the balance of process-form relationships is altered and the morphology changes (Brierley and Fryirs, 2005). The distribution of floodplains along a river course determines the accommodation space within which the channel is able to adjust (Fryirs and Brierley, 2010). Imposed boundaries such as bedrock valley margins and terraces limit vertical and lateral adjustment, resulting in rivers that are resilient to change (i.e. they have low sensitivity). Fully alluvial channels alter their state in response to the balance between sediment characteristics and energy (slope and discharge), and thus represent more responsive, sensitive systems that are able to undergo greater change (Nanson and Huang, 2008). This energy – sediment balance, and resulting alluvial river morphology, also reflects the ability of the energy supplied to rework material and do geomorphic work (Eaton et al., 2010). Thus, assessment of channel change is strengthened when the character and behaviour of the river type is considered with regard to their natural capacity for adjustment.

Contemporary channel adjustments should be contextualised within longer term patterns of landscape evolution. Past patterns of geology, climate and anthropogenic controls shape landscapes and determine the distribution and availability of sediment stores (Brierley, 2010; Lane and Richards, 1997). As Chapter 3 describes the evolution of the Tongariro catchment, this chapter uses this context to provide insights on channel adjustment. This provides a knowledge base upon which future trajectories of channel adjustment can be predicted (Chapter 7).

This chapter analyses spatial and temporal adjustments of the lower Tongariro River over the past 80 years. Across this time, sediment and water fluxes have undergone changes due to natural (flooding and volcanic eruptions) and anthropogenic (regulation and gravel mining) influences. This chapter aims to determine how responses to these events differed, with consideration of the type of river and the types and rates of adjustment (sensitivity) that is particular to a specific reach type. With this aim in mind, this chapter initially describes changes to sediment flux over time. Patterns of vertical bed adjustment and the adjustment of channel form are described, and adjustments observed are related to changes in sediment and water fluxes. A framework to classify the sensitivity of each reach is presented. Building upon these analyses, the spatial and temporal patterns of channel response are discussed.

4.2 METHODS

Planform adjustment of the lower Tongariro River was analysed using aerial photographs (1941, 1958, 1964, 1973, 1984, 1993 and 2007) and one survey map (1928) (Table 4.1). The 1928 survey map provides a baseline dataset, prior to anthropogenic development, as only two small Maori settlements were present in the catchment (Waitangi Tribunal, 1995). Discharge during periods of
aerial photography displayed a greater variability pre-regulation, between $32 - 47 \text{ m}^3\text{s}^{-1}$ compared with $27 - 29 \text{ m}^3\text{s}^{-1}$ following regulation in 1973 (Table 4.1). However, as analysis combined the wetted channel and active gravel surfaces, the influence of flow variability on the data extracted is negligible. Figure 2.20 shows the flood distribution with each aerial photograph date superimposed.

Aerial photographs were acquired in digital form and geo-referenced to 2002/03 ortho-imagery using Arc-GIS. Rectifying required a minimum of 15 ground control points and 1st or 2nd order polynominal equations to acquire a good fit. Root mean square (RMS) errors below 6.9 m were deemed acceptable as they are similar to those published in the literature (Gurnell et al., 1994; Gurnell, 1997; Winterbottom, 2000). Channel change within this study was primarily assessed by comparing differences in channel width, rather than migration rates based on the key mechanisms of adjustment for the specific River Styles. As such, it is perceived that the error margin described above has limited influence upon the output. Fewer ground control points (a minimum of 8) were acquired for the delta region in 1941 due to the small area covered by each photo and the undeveloped, wetland nature of the area at that time which limited the placement of control points.

Table 4.1: List of aerial photographs used to analyse bar adjustment including the discharge at the time of each survey. Discharge is measured for the Tongariro River at the Turangi gauging station, which is located within Reach F.

<table>
<thead>
<tr>
<th>Date</th>
<th>Map/photo reference</th>
<th>Coverage</th>
<th>Scale</th>
<th>Discharge m$^3$s$^{-1}$</th>
</tr>
</thead>
<tbody>
<tr>
<td>1928</td>
<td>SO18742-45</td>
<td>Delta to Whitikau confluence</td>
<td>1: 3960</td>
<td>Surveyed to bankfull discharge</td>
</tr>
<tr>
<td>1941 Sept</td>
<td>SN 178</td>
<td>Delta to Admirals Pool</td>
<td>1: 16000</td>
<td>Pre hydrologic gauging</td>
</tr>
<tr>
<td>1958 05-Jun</td>
<td>2774/4 2775/4 2776/5 2776/6 2777/7 2778/4 2779/4</td>
<td>Delta to Whitikau confluence</td>
<td>1: 18290</td>
<td>47.60 41.54</td>
</tr>
<tr>
<td>1964 21-May</td>
<td>SN 1690: Run A/1–6 Run B/1–7 Run C/1–4</td>
<td>Delta to Whitikau confluence</td>
<td>1: 2500</td>
<td>32.53</td>
</tr>
<tr>
<td>1973 26-Nov</td>
<td>SN 3688</td>
<td>Delta to Whitikau confluence?</td>
<td>1: 16000 and 1: 17500</td>
<td>40.47</td>
</tr>
<tr>
<td>1984 26-Dec</td>
<td>&gt; SN 8440</td>
<td>Delta to Whitikau confluence</td>
<td>1: 2500</td>
<td>29.392*</td>
</tr>
<tr>
<td>1993 12-Jan</td>
<td>230267, 16.2, 17.2, 18.2, 19.2, 230251 16.3</td>
<td>Delta to Whitikau confluence</td>
<td>1: 2500</td>
<td>27.776*</td>
</tr>
<tr>
<td>2007 25-Oct</td>
<td>SN 50660c 49_5 50_8</td>
<td>Delta to Whitikau confluence</td>
<td>1: 40000</td>
<td>28.532*</td>
</tr>
</tbody>
</table>

* indicates photographs taken post regulation of the river, effectively halving the flow.

The channel boundary and morphological units (e.g. bars and islands) were digitised for each of the aerial photographs. Past attempts to avoid inconsistency in identifying channel margins due to water levels being surveyed at different dates are described by Winterbottom (2000) and Wishart et al. (2008). These suggest that the definition of channel boundary includes wetted area, unvegetated
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gravel, and sparsely vegetated gravel within the active channel (cf. Hicks et al., 2008). Vegetation thickness upon each unit provides an indication of the frequency of reworking and the stability of the unit. This was used to underpin three measures which relate channel width to different temporal frequency of reworking. These are the short-term active channel which includes wetted channel and unvegetated, active gravel surfaces, the long-term active channel which includes the addition of partly vegetated, less frequently reworked features, and lastly the total channel area which also includes fully vegetated units such as islands or floodplain connected to the channel by fluvially-created flood channels (Table 4.2). This allowed the lateral changes in channel planform to be related back to the temporal patterns of reworking. The combination of the three measures was designed to provide a more advanced description of adjustment to the wandering gravel bed planform, by incorporating a temporal component (i.e. vegetation cover indicates time elapsed since a surface was reworked by a flood). In addition, commonly used indices to describe channel planform change including sinuosity and braided indexes were calculated (Brice, 1960; Gurnell, 1997; Knighton, 1998) (see Table 4.2).

The lower Tongariro River was delineated into ten reaches, based on differences in the nature/rate of adjustment observed. These reaches also reflected similarities in underlying controls and dominant geomorphological processes which determined responses to flood events. These encompassed four distinct river styles, exhibiting a wide range of character and behaviour (Table 4.3). The wandering cobble bed river comprised 7 reaches (A-G as shown in Figure 4.1). Each reach was seen to undergo large differences in the type/rate of adjustment, with highly active reaches reworking large areas of floodplain, whilst less active reaches underwent very little change in the 80 year period observed. In recognition of the differences in adjustment and channel width this separation into sub-reaches was used to assess why some reaches are sensitive and others not. However, it should be noted that despite differences in behaviour, the variables which are used to differentiate River Style type (i.e. confinement, planform, geomorphic units and bed material) remain the same across these sub-reaches. The three River Styles downstream of the terraces each comprise a single reach (Reaches H, I and J). Overall, each reach (A-J) varied between 1.3 and 2.8 km of valley length.
Table 4.2: Measures of channel planform adjustment. This is based on areas extracted from the digitised aerial photographs displayed in Figure 4.3 and Figure 4.5. Surface area types (unvegetated gravel, partly vegetated bars) correspond to the classifications of surfaces areas presented there. All area measures have been standardised by valley length to make them comparable, removing the bias of reach length and providing a measure of average channel width for a point in time.

<table>
<thead>
<tr>
<th>Measure</th>
<th>Notation in Table 4</th>
<th>Explanation as to how it is physically measured and what it describes</th>
<th>River Style Relevance</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Short-term channel area</strong></td>
<td>STA</td>
<td>Areal measure (m$^2$) of the wetted channel + unvegetated gravel bar and island units. Signifies the portion of the channel that has been reworked within the last 6 months. As this measure records the impact of recent events, it may provide a misleading representation of longer-term adjustments because of differences in flood histories.</td>
<td>Partly confined, wandering, cobble</td>
</tr>
<tr>
<td><strong>Long-term channel area</strong></td>
<td>LTA</td>
<td>Areal measure (m$^2$) including the short-term channel area and partially vegetated bars and islands. This is a summary measure of all units that have been reworked over timescales &lt;2 years.</td>
<td>Unconfined, braided gravel</td>
</tr>
<tr>
<td><strong>Total channel area</strong></td>
<td>TA</td>
<td>Areal measure (m$^2$) of the channel area incorporating Long term channel area + full vegetated units. This can include floodplain units collected by flood runners that may not have been actively reworked. This measure expresses the total land influenced by the channel at a given point in time.</td>
<td>Unconfined, Meandering, sand</td>
</tr>
<tr>
<td><strong>Sinuosity</strong></td>
<td>S</td>
<td>Thalweg length / valley length, describing reach sinuosity.</td>
<td>Unconfined delta</td>
</tr>
<tr>
<td><strong>Braided Index</strong></td>
<td>BI</td>
<td>2 x the length of lateral bars and islands / reach length taken by the thalweg (Brice, 1960).</td>
<td></td>
</tr>
<tr>
<td><strong>Sensitivity Index</strong></td>
<td>SI</td>
<td>The total channel area (divided by valley length to make it comparable across reaches) is calculated and the standard deviation for each reach across all time periods is extracted. This provides an expression of the variability in adjustment to channel width over time for a specific reach.</td>
<td></td>
</tr>
</tbody>
</table>

The relevance of these measures for differing River Styles is appraised using the following ranking scheme: 1 Poor measure of change for that River Style, 2 moderately important measure of change, and 3 very important measure for that River Style.
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The extent to which a specific measure of planform change was used to describe adjustment in a reach was selected based on the types of adjustment that type of river would be expected to undergo given its natural capacity for adjustment (see Table 4.2). For example, the braided index is a poor assessor of change in a meandering gravel bed river due to the lack of barforms, while sinuosity is far better at capturing changes to the planform as the channel migrates to adjust its sediment load and its slope (Brierley and Fryirs, 2005). Adjustment within wandering gravel bed rivers is driven by changes to sediment storage and bar extent, as large floods can rework floodplain surfaces and introduce pulses of sediment which is stored in active bar complexes. In addition, the size of the braidplain can reflect the volume of active sediment storage within a reach. High sediment delivery would be expected to increase braiding, whilst periods of lower yields are associated with narrowing and a decrease in bar area (Church, 2006; Ferguson, 1981; Leopold and Wolman, 1957). As such, each measure was related back to its ability to describe planform change and the mechanisms of adjustment for a specific type of river, and its implications to document changes to sediment fluxes and storage (Table 4.2).

Additional information on reach scale controls (i.e. slope, floodplain pocket width) was extracted from 1 m resolution LiDAR imagery, captured in 2006.
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Figure 4.1: Study reaches set within a detrended DEM, showing the location and relative height of the terraces and surrounding floodplains (Westaway et al., 2003). River styles are annotated and photographs are presented to display their character. Channel planform is taken from the 2007 aerial photographs.
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<table>
<thead>
<tr>
<th>River Type</th>
<th>Study Reaches</th>
<th>Valley Confinement</th>
<th>Planform</th>
<th>Geomorphic Units</th>
<th>Bed Material</th>
<th>River behaviour</th>
</tr>
</thead>
<tbody>
<tr>
<td><em>Partly Confined, wandering cobble bed river</em></td>
<td>A - G</td>
<td>Located within terraces between 50m-1000m.</td>
<td>Typically 1-2 channels with adjacent active gravel areas that are commonly larger than the wetted area. Low sinuosity.</td>
<td>Terraces, discontinuous floodplain pockets, lateral bars, point bars, mid channel bars, islands, riffles, runs, pools.</td>
<td>Gravel to Cobble. D$_{50}$ ranges from 212-91 mm down the reach. Larger material from past lahars result in local boulders up to 1110mm.</td>
<td>Bankfull flows transport gravel from bar to bar, reworking bar features and scouring pools. High magnitude flows strip the floodplain, which becomes part of the active channel. Extended periods of low flow increase vegetation on surfaces, increasing roughness and decreasing the ease with which it can be reworked by the next large event.</td>
</tr>
<tr>
<td><em>Unconfined, braided gravel bed river</em></td>
<td>H</td>
<td>Unconfined, with localised constructed levees.</td>
<td>1-5 channels. Low sinuosity.</td>
<td>Lateral bars, point bars, mid channel bars, islands, riffles, runs, pools, floodplain, created levee floodworks</td>
<td>D$_{50}$* ranges from 91-25 mm. Few large clasts present, up to 500mm.</td>
<td>The unconfined valley allows greater lateral adjustment via thalweg shift, locally reworking the very flat floodplain. Levees in the upper section artificially confine short sections, trapping overbank flows on the floodplain.</td>
</tr>
<tr>
<td><em>Unconfined, meandering, sand bed river</em></td>
<td>I</td>
<td>Unconfined, flat floodplains on both sides</td>
<td>Single channel, high sinuosity</td>
<td>Runs, pools, point bars, wetland floodplain</td>
<td>D$_{50}$ 8 mm, range 1-23 mm.</td>
<td>Bankfull flows move sediment from bar to bar. Due to the small size of sediment in this reach, sub-bankfull flows are able to transport sediment and do geomorphic work. The lack of ridge and swale topography indicates that the channels are relatively fixed in place (i.e. this is a passive meandering river).</td>
</tr>
<tr>
<td><em>Unconfined, multi-channelled delta</em></td>
<td>J</td>
<td>Unconfined, flat floodplains on both sides</td>
<td>Multiple branching channels, 2 for most part and 4 at the downstream end</td>
<td>Islands, wetland floodplain, glide.</td>
<td>80 % of material &lt; 1 mm, so D$_{50}$ = 1 mm</td>
<td>Low slope and frequent overbank flows result in a depositional river. Sediments accumulate on the floodplain and prograde into the lake. Fluctuations in lake level induced by control gates influence the rate of deposition. In-channel deposition has narrowed the channel. Palaeochannels are indicative of former avulsion events.</td>
</tr>
</tbody>
</table>

* Sediment sizes are from D$_{50}$ Wolman counts (100 clasts) or sieving taken from the coarsest locale of a bar. As a result these represent the largest material a reach can transport, rather than reflecting the overall mean sediment size and thus may differ from the overall bed material classification.
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The flood history was analysed using 50 years of data from the Turangi flow gauge located within Reach F. This provided a measure of the changes that flow regulation from the hydro-power dams in the upper catchment have had on large, geomorphically effective flood events.

Assessment of vertical bed adjustment was carried out using average bed elevation data surveyed at two flow gauges. The Puketerata gauge is located 1 km upstream of the upper limit of the study reach (Figure 4.1) and has 51 years of data, whilst the Turangi gauge is located within Reach F and has 47 years of data. This provided a rigorous way to assess how anthropogenic and natural events have influenced bed adjustment.

4.3 RESULTS

This section analyses key controls upon planform adjustment within the lower Tongariro River over the past 80 years. The first section presents a summary of natural and anthropogenic influences upon sediment and water fluxes for each aerial photographic period. This allows channel response in the subsequent sections to be related back to changes in flux. Vertical bed adjustment is related to erosional and depositional responses. Patterns and rates of planform adjustment across space and time are analysed. A sensitivity framework is developed as an interpretive tool with which to rank the extent of adjustment observed for differing types of river.

4.3.1 Changes to Fluxes within the Tongariro Catchment

Despite the low proportion of the catchment that is directly influenced by human land use, anthropogenic influences have elicited major controls upon the sediment and water fluxes within the lower Tongariro over the past 80 years (Table 4.4). Note, a graph of the flood frequency between each aerial photograph survey is provided in Figure 2.20.

There are two small hydropower dams in the mid-catchment. Poutu Intake Dam, the furthest downstream, was constructed in 1973. This caused a 40% decrease in mean discharge in the lower reaches (Hindle, 1995). Rangipo Dam is located in the upper catchment and was completed in 1983. Average flows downstream of the dams are 1.3 and 17 m³s⁻¹ for the Poutu Intake and Rangipo dams respectively, compared with natural mean flows pre-regulation of 16.1 and 33.9 m³s⁻¹ (Genesis Energy, 2000). However, changes to larger flood events are more important for understanding alterations to bedload transport. Small flood events (freshes between 70 - 100 m³s⁻¹) decreased in frequency following regulation (Hindle, 1995). Number of flood days a year (> 100 m³s⁻¹) also decreased following regulation, increasing again from 1993, though not to the same levels observed pre regulation (Figure 2.19). The frequency of large magnitude events decreased in the 1970s and early 1980s but increased to higher than pre-regulation levels since 1993 (Figure 2.21).
Both dams trap sediment, decreasing sediment delivery to mid-catchment reaches. The Poutu Intake allows suspended and bed sediment smaller than 200 mm to be transported through a screen into settling chambers (Jowett, 1980), from which it is scoured out approximately once a week (Christmas et al., 1996). During floods greater than 100 m³ s⁻¹ the Rangipo sluice gates are opened and diversion at the Poutu Intake is stopped. This allows high discharges to flush up to 60,000 tons of sediment stored within the Rangipo reservoir down the Tongariro River and through the Poutu intake (Collier, 2002). As a result, the delivery of material to the conveyer belt is more jerky than would be expected from the natural regime. In summary, whilst there are fewer ‘flood’ days per year, the number of large flood events are unaffected and are influenced by natural rainfall rather than regulation.

Sediment and water movement is influenced by base-level, which is set by Lake Taupo for the Tongariro Catchment. Control gates on the lake outlet have increased the frequency of above average levels, creating a backwater effect that limits sediment transport though the lower reaches (Scott, 1999; Smart, 1999). As a result, an increase in the size of the wetland and frequency of flooding has been observed in the delta (Eser and Rosen, 2000; Smart, 2005).

Gravel has been extracted intermittently from the braided reach directly downstream of the terraces, with over 1 million tons removed in the 1960s (Genesis Energy, 2000). Following the 2004 flood, gravel mining recommenced, with consent granted for the removal of 15,000 m³ of gravel (Environment Waikato, 2004). Floodworks have been constructed adjacent to Turangi (primarily stopbanks), improving conveyance of material through these sections of the reach (Jones, 2003).
Table 4.4: History of changes to fluxes within each time period between Aerial Photos

<table>
<thead>
<tr>
<th>Reach Name</th>
<th>Reaches influenced</th>
<th>Adjustment</th>
<th>Change to fluxes*</th>
</tr>
</thead>
<tbody>
<tr>
<td>Pre-1928</td>
<td>1- All</td>
<td>1- Some land clearance and development of low intensity farming in the downstream reaches.</td>
<td>1- Minimal</td>
</tr>
<tr>
<td>1928-1941</td>
<td>1- All</td>
<td>1- Little change. Small growth of fishing accommodation on the banks upstream of State Highway 1 bridge.</td>
<td>1- Minimal</td>
</tr>
<tr>
<td>1941-1958</td>
<td>1- I–J</td>
<td>1- Control gates installed on the outlet of Lake Taupo, altering patterns of base-level fluctuations, resulting in an increase mean level and the lake spends more time above the mean, creating a backwater effect and facilitating deposition in the delta.</td>
<td>1- ↓ sediment transport and flow velocities in delta section and increase in flooding. 2- ↑ in energy to move sediment and rework greater surface areas (s)</td>
</tr>
<tr>
<td></td>
<td>2- All</td>
<td>2- 1958 saw the largest flood on record with a max flow of 1470m^3^s at Turangi</td>
<td></td>
</tr>
<tr>
<td>1958-1964</td>
<td>1- All</td>
<td>1- 3rd largest flood with a peak flow of 1000 m^3^s at Turangi though this was of relatively short duration.</td>
<td>1- ↑ in energy to move sediment and rework greater surface areas (s)</td>
</tr>
<tr>
<td>1964-1973</td>
<td>1- G–I</td>
<td>1- Gravel extraction between 1964-1972 where 800,000m^3^ of sediment was taken from the braided section below the bridge.</td>
<td>1- ↓ sediment load (m)</td>
</tr>
<tr>
<td>1973-1984</td>
<td>1- All</td>
<td>1- Both Poutu and Rangipo Dams constructed during this period resulting in a 40% decrease to mean flow volume and discontinuous supply of sediment load, with the dam being flushed when flows exceed 100m^3^s scouring up to 60,000 tonnes from the dam. Shorter flood peaks decrease the bedload transport capacity of floods.</td>
<td>1- Pulsed sediment load, which takes longer to reach downstream (l) ↓ mean flow, number of small floods and transport capacity of large events</td>
</tr>
<tr>
<td></td>
<td>2- N/A</td>
<td>2- More sediment extracted for use in the TPD. Exact amounts are not documented</td>
<td>2- ↓ sediment load (m)</td>
</tr>
<tr>
<td></td>
<td>3- G–H</td>
<td>3- Bank protection constructed on both banks below SH1 bridge.</td>
<td>3- ↑ confinement and decreases areas available to be reworked (l)</td>
</tr>
<tr>
<td></td>
<td>4- All</td>
<td>4- No significant flood events, with none even close to bankfull flow (reoccurrence interval of 2.33yrs).</td>
<td>4- ↓ in energy available to move sediment and rework surfaces (s)</td>
</tr>
<tr>
<td>1984-1993</td>
<td>1- F–H</td>
<td>1- Further development of stopbanks above SH1 bridge.</td>
<td>1- ↑ confinement and decreases areas available to be reworked (l)</td>
</tr>
<tr>
<td>1993-2007</td>
<td>1- All</td>
<td>1- Volcanic eruption of 1995-1996 delivered an estimated 6900 kilotonnes of volcanic sediment, most was very fine &lt; 0.5mm and flushed relatively rapidly.</td>
<td>1- ↓ sediment load (m) 2- ↑ in energy to move sediment and rework greater surface areas (s)</td>
</tr>
<tr>
<td></td>
<td>2- All</td>
<td>2- 2004 flood, 2nd largest flood on record with peak discharge of 1400 m^3^s at Turangi</td>
<td>3- ↓ sediment load (m)</td>
</tr>
<tr>
<td></td>
<td>3- G–I</td>
<td>3- Following the 2004 flood 15,000m^3^ of gravel extracted</td>
<td>4- ↓ Channel roughness, increase in bank erodibility. Anticipated increase in channel capacity in the lower reaches.</td>
</tr>
<tr>
<td></td>
<td>4- F–J</td>
<td>4- Channel clearance in the lower reaches, including removing willow species and accumulated woody debris and extracting localised gravel shoals that are above water level to minimise impact upon channel morphology.</td>
<td>5- ↑ in energy to move sediment and rework greater surface areas (s)</td>
</tr>
<tr>
<td></td>
<td>5- All</td>
<td>5- Increased frequency of high magnitude flow events.</td>
<td></td>
</tr>
</tbody>
</table>

* (s) indicates short term effect of < 1 year, (m) indicates medium term influences of greater than a year and (l) indicates ongoing long term effect.
4.3.2 Vertical Channel Change within the Lower Tongariro

Prolonged changes to sediment and water fluxes would be expected to bring about vertical adjustments to bed level. Bed adjustment is analysed from long-term records of cross-section surveys at the Puketerata gauging station situated 1 km upstream of Reach A and the Turangi gauging station located within reach F.

Figure 4.2: Bed adjustment for A) the Puketerata gauging station 2.5 km upstream from Reach A and B) at the Turangi gauging station located within Reach F. Bed level height is normalised for a discharge of 25 m$^3$s$^{-1}$.

At the Puketerata gauging station the channel bed fluctuated, exhibiting a small overall increase from 465.4 m to 465.65 m between 1960 and 1971. The bed remained fairly static at around 465.65 m until 1995. Following the eruption of Mt Ruapehu the river bed decreased in elevation by around 0.1 m. This is surprising given the increase in sediment load on the river. However, an increase in flood
days (flow greater than 100 m$^3$s$^{-1}$) as more water was released to flush the dams during this period (Genesis Energy, 2000) may explain this degradation (Figure 2.19A). The bed remained static until the 60 year RI flood in 2004 which decreased bed elevation by 0.25 m. Following this, the bed has been aggrading to the level observed prior to the 2004 flood.

Bed level at the Turangi gauging station shows a similar pattern. The 1958 flood caused a 0.6 m increase in bed level indicating a pulse of sediment delivered to this downstream reach. The bed slowly aggraded by 0.3 m until 1986, where it plateaued, and has subsequently undergone slow incision (of about 0.2 m). The period over which the dams were constructed shows slight increase in bed height of 0.1 m. However, as this is well within the natural range of adjustment due to the large size of the bed material, no causal link is evident.

Observed changes in bed elevation are relatively minor, especially given the size of bed material and wavelength of bedforms. It is likely that much of the change falls within the margin of error for this technique and therefore must be treated as indicative. Smart (1999) suggests that the gentle rise and fall in bed level is indicative of a slug of sediment moving through this lower reach. If this is the case, it is a relatively small sediment slug, and could represent bedform migration.

Data from the two gauging stations exhibit marked similarities and differences in adjustment, indicating that responses to some events may operate separately from reach to reach. The main observable change in both records can be seen in the incision that occurred following the 1995 eruption of Mt Ruapehu and the 2004 60-year RI flood event. This period also coincides with a period of increased frequency in high magnitude events (7 with a greater than 5 year RI). However, the aggradation following the 2004 flood at the Puketerata gauge was not observed at the Turangi gauge. A decrease in large floods during this period may have increased aggradation at this point, and such responses may not yet have been transferred to the Turangi gauge lower in the catchment.

4.3.3 The Pattern of Spatial Channel Adjustments along the Lower Tongariro

Types, patterns and rates of channel adjustment along the lower Tongariro River have varied markedly over the last 80 years. This section describes adjustments for each reach.

Reaches A to G are located within the terraces, comprising the Partly confined, wandering, cobble bed river style (Table 4.3 and Figure 4.1). A, D and F underwent minimal, slow adjustment to channel character and sediment storage through mid-channel bar growth, lateral bar growth and the transition from mid-channel bars to lateral bars (Figure 4.3 and Table 4.5). Boundary condition controls upon these reaches are similar. These reaches have low sinuosity (a maximum of 1.2), narrow average terrace width (185-352 m) and average channel width of <90 m (Table 4.6). Reach B
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exhibited the greatest degree of adjustment for this River Style (Figure 4.4 and Table 4.6). Its terrace width of 678 m is noticeably wider than other reaches (Table 4.6). Despite this, the channel still has a high degree of contact with the left terrace margin. The channel expanded in area and increased in sinuosity between 1928 and 1958, through meander extension, chute cutoffs and reworking of floodplain, transforming local areas into lateral bars (Table 4.5). Total channel area exhibited an overall decrease from 1958 to 1993 (Figure 4.4). Channel area increased substantively from 1993-2007 through channel straightening and reworking of the floodplain area adjacent to the channel, creating bars. Reaches C and E underwent moderate degrees of adjustment. Both consist of meander bends that are forced by the terrace margin. Reach E is a larger meander, as the terrace is wider at this point. Adjustment within Reach C consists primarily of the reworking of a lateral bar and a point bar, wherein floodchannels increase the active area at high flows. Mid-channel bars form at the downstream extent of the reach and at the apex of the bend. These bars increased in size from 1958, but remained relatively stable features from 1973 onwards. Reach E exhibited an increase in total active channel between 1928 and 1958 through reworking of floodplain surfaces, forming lateral bars (Table 4.5). A mid-channel bar formed at the apex of the bend by 1958, increasing in size till 1964. This feature remained stable until 1993. Between 1993 and 2007 it became attached to the floodplain, transitioning to a lateral bar. Total channel area (which includes active, wetted and vegetated surfaces) increased over this last period (Figure 4.4).

Reach G has a terrace on the right bank. Stopbanks were constructed to increase confinement in 1973 (Figure 4.3). The active area for this reach has remained relatively stable over time, whilst the area within this has been reworked. A meander bend translated and extended from 1973-2007, with the primary channel eventually switching to occupy a smaller channel, fixing the bend against the right terrace margin. This has split a large island into smaller island features, instigating the development of a point bar on the convex side of the bend and small mid-channel bars at the upstream extent of the reach.
Figure 4.3: Digitised aerial photographs showing the history of planform adjustment for the upstream reaches of the lower Tongariro River.
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Table 4.5: Mechanisms of change for each reach between 1928-2007. Changes to characteristics of the channel are presented on the top row, and the mechanisms of change in the rows beneath in italicised text. Adjustments that had a major influence upon the character of the reach are annotated in capital letters. “+” and “−” represent relative increases and decreases, respectively. Shaded squares for the period from 1928-1941 indicate reaches for which the 1941 aerial photography was not available. Reaches A – G are Partly confined, wandering cobble bed rivers, H is an Unconfined, braided, gravel bed river, I is an Unconfined, meandering, sand bed river and J an Unconfined, multi-channelled delta.

<table>
<thead>
<tr>
<th>Time Period</th>
<th>A</th>
<th>B</th>
<th>C</th>
<th>D</th>
<th>E</th>
<th>F</th>
<th>G</th>
<th>H</th>
<th>I</th>
<th>J</th>
</tr>
</thead>
<tbody>
<tr>
<td>1928-1941</td>
<td></td>
<td>+ta,</td>
<td>+TA,</td>
<td>+STA</td>
<td>+TA,</td>
<td>+TA,</td>
<td></td>
<td>-TA</td>
<td>-ta</td>
<td>-ta</td>
</tr>
<tr>
<td></td>
<td>cc,</td>
<td>BI,</td>
<td>M-E,</td>
<td>mbg</td>
<td>M-CBG</td>
<td>LBG</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>1941-1958</td>
<td>-TA</td>
<td>+STA</td>
<td>+MBG</td>
<td>-bi,</td>
<td>-TA</td>
<td>+TA</td>
<td></td>
<td>+TA</td>
<td>-ta</td>
<td>-ta</td>
</tr>
<tr>
<td></td>
<td>mcbg</td>
<td></td>
<td></td>
<td></td>
<td>mcbg</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>1958-1964</td>
<td>-TA</td>
<td>+MBG</td>
<td>+TA</td>
<td>+BI</td>
<td>+TA</td>
<td>+TA</td>
<td></td>
<td>+TA</td>
<td>-ta</td>
<td>-ta</td>
</tr>
<tr>
<td></td>
<td>mcbg</td>
<td>VS,</td>
<td>V,</td>
<td></td>
<td>mcbg</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>1964-1973</td>
<td>+ta</td>
<td>VS,</td>
<td>T</td>
<td>+MBG</td>
<td>+TA</td>
<td>+TA</td>
<td></td>
<td></td>
<td>-ta</td>
<td>-TA</td>
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<tr>
<td></td>
<td>T</td>
<td></td>
<td>m-L</td>
<td></td>
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<td></td>
<td></td>
</tr>
<tr>
<td>1973-1984</td>
<td>+ta</td>
<td>VS,</td>
<td>T</td>
<td>+MBG</td>
<td>+TA</td>
<td>+TA</td>
<td></td>
<td></td>
<td></td>
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<tr>
<td></td>
<td>T</td>
<td></td>
<td>m-L</td>
<td></td>
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<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>1984-1993</td>
<td>M-L</td>
<td>-TA</td>
<td>-MBG</td>
<td>+MBG</td>
<td>+TA</td>
<td>+TA</td>
<td></td>
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<tr>
<td></td>
<td>VS,</td>
<td>T</td>
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<td></td>
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<td></td>
</tr>
<tr>
<td>1993-2007</td>
<td>+ta</td>
<td>+TA</td>
<td>+MBG</td>
<td>+MBG</td>
<td>+TA</td>
<td>+TA</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>LBG</td>
<td>CC,</td>
<td>LBG,</td>
<td>mcbg</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Measure</td>
<td>STA</td>
<td>Changes to frequently reworked surfaces measured through wetted channel area and unvegetated gravel.</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Notation</td>
<td>LTA</td>
<td>Changes to area of surface influenced by the channel, including floodplain areas connected by flood channels.</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Explanation</td>
<td>S</td>
<td>Measure of channel sinuosity.</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Braided index</td>
<td>BI</td>
<td>Number of channels.</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Vegetation stability</td>
<td>VS</td>
<td>Indicates changes to the stability of the reach based on changes to density of vegetation cover. Recovering vegetation indicates infrequent or low reworking and the transition of active units back to floodplain.</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Bend migration</td>
<td>BM-E or BM-T</td>
<td>Bends migrate through extension (BM-E) or translation (BM-T).</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Channel narrowing</td>
<td>CN</td>
<td>Decrease in the area of active channel (wetted channel + unvegetated gravel).</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Terrace forcing bend</td>
<td>T</td>
<td>Bend is pinned against the terrace, creating a forced morphology and sinuosity.</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Primary channel switch</td>
<td>PCS</td>
<td>The primary channel switches to occupy an existing channel. This is commonly due to a channel bend being forced against the terrace and switching course creating a straighter channel.</td>
<td></td>
<td></td>
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<td></td>
</tr>
<tr>
<td>Chute channels</td>
<td>CC</td>
<td>Chute channels form on bars floodplain surfaces, creating mid-channel bars and islands and increasing braiding.</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
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<td></td>
</tr>
<tr>
<td>Mid channel bar growth</td>
<td>MCBG</td>
<td>Islands form by material being deposited in the channel. This grows, horizontally and vertically, causing channel widening and increasing braiding.</td>
<td></td>
<td></td>
<td></td>
<td></td>
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<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Transformation of mid channel bar to lateral bar</td>
<td>M-L</td>
<td>The primary channel dominates one side of the island, causing the island to reattach to the floodplain becoming a lateral bar. This is commonly associated with an increase in stability and a decrease in braiding, and provides a step in the transformation of active channel to floodplain.</td>
<td></td>
<td></td>
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<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Lateral Bar growth</td>
<td>LBG</td>
<td>Lateral bars form through deposition or floodplain reworking, commonly causing an increase in sinuosity and reworking of the valley margin.</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
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<td></td>
</tr>
</tbody>
</table>
### Table 4.6: Controls on adjustment for each reach.

Terrace outlines were delineated from the LiDAR derived 1m resolution DEM. Areas were extracted using polygons in Arc-GIS. Valley slopes were also extracted from the DEM using the centre line down the valley for the length of each reach. Reaches A – G are Partly confined, wandering cobble bed rivers, H is an Unconfined, braided, gravel bed river, I is an Unconfined, meandering, sand bed river and J is an Unconfined, multi-channelled delta.

<table>
<thead>
<tr>
<th>Reach Name</th>
<th>Valley Slope (2007) m/m</th>
<th>Average Terrace width* (m)</th>
<th>(D_{50}) (mm)</th>
<th>Sinuosity (2007)</th>
<th>Range in Sinuosity</th>
<th>Braided Index (2007)</th>
<th>Range in Braided Index</th>
<th>Average Total Channel Width(^ ^) (m)</th>
<th>Range in average total channel width(^ ^) (m)</th>
<th>St dev average total channel width (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>A</td>
<td>0.0111</td>
<td>352</td>
<td>163</td>
<td>1.20</td>
<td>1 - 1.2</td>
<td>1.56</td>
<td>0.32 - 1.56</td>
<td>75</td>
<td>54 – 108</td>
<td>22</td>
</tr>
<tr>
<td>B</td>
<td>0.0104</td>
<td>678</td>
<td>213</td>
<td>1.18</td>
<td>1.12 - 1.34</td>
<td>3.08</td>
<td>0.93 - 3.11</td>
<td>195</td>
<td>119 – 264</td>
<td>61</td>
</tr>
<tr>
<td>C</td>
<td>0.0063</td>
<td>422</td>
<td>140</td>
<td>1.16</td>
<td>1.08 – 1.27</td>
<td>2.11</td>
<td>1.49 – 3.15</td>
<td>145</td>
<td>85 – 172</td>
<td>28</td>
</tr>
<tr>
<td>D</td>
<td>0.0071</td>
<td>249</td>
<td>120</td>
<td>1.06</td>
<td>1.02 – 1.1</td>
<td>0.61</td>
<td>0.49 – 2.05</td>
<td>90</td>
<td>74 - 107</td>
<td>13</td>
</tr>
<tr>
<td>E</td>
<td>0.0081</td>
<td>374</td>
<td>130</td>
<td>1.40</td>
<td>1.29 – 1.45</td>
<td>2.45</td>
<td>1.58 – 3.75</td>
<td>162</td>
<td>124 – 209</td>
<td>29</td>
</tr>
<tr>
<td>F</td>
<td>0.0056</td>
<td>185</td>
<td>113.5</td>
<td>1.04</td>
<td>0.99 – 1.04</td>
<td>0.59</td>
<td>0.29 – 1.37</td>
<td>65</td>
<td>44 – 89</td>
<td>17</td>
</tr>
<tr>
<td>G</td>
<td>0.0046</td>
<td>281</td>
<td>91</td>
<td>1.14</td>
<td>0.97 – 1.14</td>
<td>2.04</td>
<td>1.29 – 2.76</td>
<td>157</td>
<td>123 – 180</td>
<td>17</td>
</tr>
<tr>
<td>H</td>
<td>0.0026</td>
<td>Unconfined</td>
<td>85</td>
<td>1.26</td>
<td>1.18 – 1.33</td>
<td>2.31</td>
<td>1.39 – 8.23</td>
<td>288</td>
<td>95 – 589</td>
<td>196</td>
</tr>
<tr>
<td>I</td>
<td>0.0016</td>
<td>Unconfined</td>
<td>8</td>
<td>2.66</td>
<td>2.5 – 2.67</td>
<td>0.11</td>
<td>0 – 0.5</td>
<td>201</td>
<td>118 – 265</td>
<td>59</td>
</tr>
<tr>
<td>J</td>
<td>0.0006</td>
<td>Unconfined</td>
<td>1</td>
<td>1.00</td>
<td>0.96 – 1.02</td>
<td>1.46</td>
<td>0.57 – 2.26</td>
<td>239</td>
<td>147 - 330</td>
<td>76</td>
</tr>
</tbody>
</table>

*Terrace area/valley length  
\(D_{50}\) taken from coarsest locale of a representative bar within each reach  
\(^ ^\)Total channel area/valley length

Figure 4.4: Changes to the total area for each reach between 1928 and 2007. Reaches H, I and J are not included as they have different mechanisms of adjustment making direct comparison misleading.
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Figure 4.5: Digitised aerial photographs showing the history of planform adjustment for the downstream reaches of the lower Tongariro River.
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Reach H is an unconfined, braided gravel bed river directly downstream of the termination of the terraces. This reach has experienced the greatest degree of channel change in the study area (Figure 4.5). Channel area increases dramatically between 1928 and 1958 by over 1 km$^2$ (Figure 4.6). Flood channels reworked floodplain surfaces, increasing lateral and mid-channel bars and increasing braiding (from 2.3 to 8.2). By 1964, the channel area had decreased, with large sections reverting back to floodplain. By 1973, the channel had narrowed significantly, adopting a predominantly single channel of low sinuosity. Gravel extraction removed around 800,000 m$^3$ of material from this section. This narrow channel remained from 1984-1993, exhibiting little change, although bars increasingly stabilised through vegetation regeneration. By 2007 there had been an increase in channel area and braiding, with the channel expanding at the widest point from <100 m in 1973 to >300 m (Figure 4.6). These adjustments were located slightly upstream of earlier adjustments, increasing the threat to the adjacent town of Turangi (Figure 4.5).

![Figure 4.6: Changes in active channel over time for Reach H. This is displayed as ‘total channel area’, which includes wetted unvegetated, partly vegetated and fully vegetated surfaces, ‘long term active channel’ which excludes fully vegetated surfaces as they are indicative of less reworking and ‘active’ channel area, which only includes the most frequently reworked surfaces comprising the wetted channel and unvegetated gravels.](image)

Downstream of the braided reach is an unconfined meandering, sand bed reach (reach I). Although banks comprise non-cohesive sand materials, the fairly tight meanders (average sinuosity of 2.62) have exhibited minimal or no lateral migration. The channel has undergone long term contraction, consistently decreasing the channel width with an overall decrease of 26 m channel width per metre thalweg length (Figure 4.5; Figure 4.7). This reflects a cycle of formation and stabilisation of lateral and point bars, creating on-going narrowing. The delta (reach J) has shown a similar pattern of
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channel narrowing, including the abandonment of a secondary channel in the upper reaches by 1964, and the almost complete infilling of an additional channel at the delta mouth by 2007 (Figure 4.5). Delta extension has occurred via the formation of bars and islands at the delta mouth. Stabilisation of these features has induced channel contraction at the mouth. Both the meandering and delta reaches increased in channel width from 1984-1993, before narrowing once more to 2007, providing an exception to the long-term trend (Figure 4.7). The establishment of exotic willow species has accelerated narrowing by increasing bank roughness and channel deposition. Over the past decade a secondary channel has been forming to the east of the delta. A head-cut is slowly migrating upstream during large floods, forming a new path where the channel is likely to avulse to once the contemporary channel becomes too narrow (Smart, 2011).

![Graph showing channel width for meandering and delta reaches](image)

**Figure 4.7:** Change in channel width for the meandering and delta reaches depicting on-going channel narrowing. Width for the meandering reach is calculated using average thalweg between 1928 – 2007. However, due to the multiple channels, width for the delta is calculated using valley length.

### 4.3.4 Temporal Patterns of Planform Adjustment in Response to Changes in Fluxes

This section discusses patterns of channel response in the lower Tongariro River, identifying which natural and anthropogenic influences are driving this response, and how long a reach takes to recover from a given event.

#### 4.3.4.1 Response of the Wandering, Cobble bed River Reaches (A-G)

The period from 1928 - 1958 saw an overall increase in total channel area across reaches A - G (Figure 4.8 a, b). The largest changes were observed in increases in average channel width of 145 and 85 m within reaches B and E. Over this period, meanders extended, and floodplain adjacent to these bends was reworked, substantially increasing the active channel area (Figure 4.3). The 1958 photographs were taken relatively soon after a 100 year flood event, capturing the response to this high magnitude event.
Between 1958 and 1964 minimal alterations occurred to sediment and water fluxes, and all wandering cobble bed reaches exhibited a small decrease in channel area (Figure 4.8b). The 1964 photographs were also taken following a large event (3rd largest on record) so it is likely that the channel had adjusted to accommodate high discharges. Further decreases in channel area were observed between 1964 and 1973 and 1973 - 1984 (excluding reach D, which underwent a slight increase in the later period) (Figure 4.8b). Minimal adjustment occurred across this time period. Despite the construction of the Poutu and Rangipo Dams, which resulted in a 40% decrease in mean...
flow volume, few large events occurred, limiting the reworking of bar and island surfaces. In-depth analysis of the effect of dams on planform adjustment is shown in Figure 4.10.

The period between 1984 and 1993 saw an increase in the variability in channel response. Overall a decrease in long term active channel area was observed, reflecting a decrease in actively reworked surfaces (Figure 4.8a). Reaches A, B and F were an exception to this pattern and underwent minor increases in long-term active channel area. Reaches B and C underwent a decrease in total channel area, seen in the stabilisation and return of bar surfaces back to floodplain (Figure 4.3). Reaches D, F and G exhibited an increase in total area, seen in the connection of vegetated to the channel through flood-channels reworking the outer limits of this surface (though this did not increase the area of reworked surfaces). This period was characterised by a marked decrease in braided index averaged across reaches, reflecting a decline in in-channel storage (Figure 4.9). Few large floods and a decrease in the number of flood days occurred within this period, bringing about channel narrowing and a decrease in braiding.

The final period from 1993 – 2007 saw an increase in reworking and channel width for all reaches. Large floods including the 2004, 60 year RI flood reworked surfaces. The number of flood days per year also increased (Figure 2.19). The sensitive reaches, especially Reaches B, C and G, exhibited the greatest increases in channel area, with B and C exhibiting increases in total channel area of 139 and 88 m respectively whilst the long-term channel area of reach G increased by 48 m (these measures are normalised by valley length).

Channel characteristics of the Wandering cobble bed reaches were compared pre- and post-dam construction to assess how flow regulation impacted channel planform.
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![Diagram](image)

Figure 4.10: A) Average channel width, B) braided index, and C) sinuosity for reaches located within the terrace constrained, wandering, cobble bed reach pre-dams (1928, 1941, 1958, 1964, 1973) and post dam construction (1984, 1993, 2007). Error bars indicate standard deviation. No significant difference was observed for any variable at any site.

The difference between pre- and post-dam widths ranged from -19 m and +20 m for reaches A – G and demonstrated no linear direction of change (Figure 4.10). Standard deviations in width overlapped, and widths for the two time periods were not significantly different. Planform characteristics, including sinuosity (which had an average difference pre- and post-dam of -0.017)
and braided index (average difference of 0.18) display minimal change to channel planform downstream of the dams after flow regulation (Figure 4.10B and C). Thus, the magnitude of change is very similar to that measured in the 45 years preceding dam development and flow regulation is perceived to have exerted negligible influence upon planform adjustment.

4.3.4.2  Response of the Braided, Gravel bed River Reach (H)

The braided reach underwent a dramatic increase in total channel area between 1928 and 1958. In 1928 the total channel area was 0.67 which increased to 1.10 in 1941 and 1.69 in 1958 (Figure 4.5; Figure 4.11). While the 1928 data incorporate increased error as is inherent with survey maps, they describe the approximate nature of the system at that time. In contrast, the 1941 aerial photographs capture the channel immediately pre-regulation of the water levels of Lake Taupo (i.e. both regulation and dates for the aerial photographs occurred within September 1941) (McConchie et al., 2008). Following regulation a marked increase in braiding and substantial reworking of the floodplain was observed (see Figure 4.12) (Hindle, 1995). Whilst the 60 year RI flood in 1958 directly before the aerial photograph was taken plays an important role in reworking floodplain surfaces, it is likely that higher water levels increased the geomorphic effectiveness of the flood and the area of floodplain that was inundated and reworked. This parallels the large areas of land inundated during the 60 year RI flood in 2004, whereby high water levels of Lake Taupo at the time were identified as a contributing factor to the high geomorphic effectiveness of the flood event (Munro, 2004).

Following the regulation of Lake Taupo, lake level is on average higher than under natural conditions, causing a backwater effect and impeding sediment transport through the delta (Smart, 1999; Tonkin and Taylor, 1999b). This impedes sediment transport through this reach. The high reworking of this zone up to 1964, where total channel area was 1.71, supports the causal influence of high lake levels on this increase in braiding (Figure 4.5; Figure 4.12). Increased flooding following the regulation of Lake Taupo levels is also supported by anecdotal evidence. The increase in flooding has led to the abandonment of Graces Farm, the original farming homestead on the delta which is shown on the 1901 survey map (Genesis Energy, 2000; Henderson, 2009; Smart, 2005; Waitangi Tribunal, 1995). A treaty claim by the local Iwi (Maori tribe) is currently in process for compensation to loss of farming land from the increase in lake levels (Waitangi Tribunal, 1995). Additional work by Eser and Rosen (2000) found that significantly higher lake levels in the summer have led to increased soil moisture and encroachment of native wetland vegetation further up the delta, replacing pastoral land.
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Figure 4.11: Changes to the channel area for the braided reach. This is further delineated into active areas comprising unvegetated gravels, semi-vegetated areas which have partial vegetation cover and fully vegetated island/floodplain sections which are considered part of the channel due to back channels or flood runners on their outer margin, connecting these features to the channel.

Gravel mining commenced in 1964 (after the 1964 aerial photograph was taken) for use in the development of the Tongariro Power Development Scheme (TPDS). This caused significant channel narrowing, whereby the braided planform was replaced by a single channel with fewer active bar surfaces. Total channel area decreased across this period from 1.71 to 0.48 km$^2$ (Figure 4.6). Gravel extraction continued through the 1970s and early 1980s masking any influence the TPDS may have had on channel planform of this reach. The volume of sediment removed is estimated to be equivalent to six years of sediment delivery for this river (Smart, 1999). Few alterations to fluxes were experienced between 1984 and 1993 and the channel was narrowest during this time (0.27 and 0.28 km$^2$ respectively). The number of flood days was relatively low, with few large magnitude events (all under 10 year return period) able to deliver sediment to the reach and rework bar surfaces (Figure 2.19). The channel underwent substantial widening by 2007. An increase in flood events, including the 60 year RI flood in 2004, increased the channel area back to 0.51 km$^2$ with 50,000 m$^3$ of gravel estimated to be delivered to this reach (Smart, 2005). As a result 15,000 m$^3$ of gravel was extracted from the reach. However, whilst in the past the zone of braiding stretched down to the meandering reach, it has established higher up, closer to Turangi. The volcanic eruption of Mt Ruapehu in 1995-1996 delivered high sediment loads to this reach (6900 kilotons). The small size of this material (< 0.5 mm) meant it was able to be rapidly flushed to the delta (Genesis Energy, 2000). The small grain size also ensured that this material did not play a role in shaping the channel planform observed within this reach, as bars are predominantly composed of cobble-gravel clasts.
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Figure 4.12: Change in the reworking of the braided reach. 1941 shows the planform immediately pre-regulation of Lake Taupo water levels, 1958 and 1964 shows post regulation, 1973 following gravel mining, 1993 after the construction of dams in the upper catchment and finish of gravel mining and 2009 illustrates the present state of the channel.
This reach has the capacity to undergo dramatic change in response to sediment fluxes. However, direct management of sediment through gravel mining has suppressed channel response.

4.3.4.3  Response of the Meandering Sand and Multi-channelled Delta Reaches (I-J)

The meandering reach has been steadily narrowing over the past 80 years. The channel has remained fixed in place, with little or no lateral adjustment observed. Channel width remained stationary between 1928 and 1964 at between 62 - 68 m wide (Figure 4.7). This period includes the installment of the control gates on the outlet of Lake Taupo. A marked decrease in width is occurred between 1964 and 1973, when the channel narrowed to 49 m. Gravel extraction took place in the upstream reach at this time, and this was a period of few large flood events. It seems likely that gravel extraction events mobilised sand which was captured within this reach, as few flood events were competent to flush these materials into the lake. Minor channel narrowing continued through to 1984 with average width declining to 45 m (Figure 4.7). By 1993 the channel had widened to 62 m. This reflects a few moderate sized floods, only two of which had RIs > 5 years and a stabilisation of sand supply from the upstream reach as gravel mining ceased. This is seen in the reworking of floodplain directly adjacent to the channel, though overbank vertical accretion may play a role in making this widening look greater than it is. By 2007, the channel had narrowed again to 47 m reflecting a return to the longer-term narrowing trajectory observed over the past 80 years (Figure 4.7). The eruption of Mt Ruapehu in 1995-1996 delivered an estimated 6900 kilotons of fine grained sediment (< 0.5 mm) into the Tongariro River, and much of this material is likely to have been captured within this reach (Collier, 2002). High magnitude floods increased in frequency during this time and additional smaller freshes were released to flush the dam. However, given the low slope and small channel capacity, these flood events were unable to flush the volumes of sediment delivered into the lake.

The delta shows similar patterns of adjustment. The channel has been narrowing from an average channel width of 103 m in 1928 to 51 m in 1984 (Figure 4.7). The width increased to 60 m in 1993 most likely due to the presence of moderate floods events that mobilised sediment within the channel without acting as a major source into the reach (see paragraph above). By 2007 the channel had narrowed to 40 m. This reflects the infilling of the channels within the delta, as on-going aggradation narrows the channels. This has been progressive and on-going over the period of record.
4.3.5 Assessment of Reach Sensitivity within the Lower Tongariro River

This section uses the analysis of planform adjustment presented in Sections 4.3.3 - 4.3.4 as a basis to classify reach sensitivity along the lower Tongariro River. Reaches display marked variability in response due to internal (i.e. slope and bed material) and external (i.e. bedrock outcrops) controls. This analysis addresses the nature of channel response (e.g. the area of land reworked following a flood), to disturbance events for different types of river.

A classification scheme detailing differences in reach scale sensitivity was created. This involved i) classifying the sensitivity of each River Style based on its ‘natural capacity of adjustment’ (as either Low, Moderate or High) and ii) classifying reach-scale adjustment based on the magnitude and rate of response observed in aerial photographs, contextualised within the sensitivity that a specific river style can exhibit.

Rates and types of adjustment (i.e. sensitivity) reflect the type of river and its natural capacity for adjustment. Reaches along the lower Tongariro River displayed differing capacities to adjust. Sensitivity rankings were designed to reflect both the types and rates of adjustment, with the initial letter (e.g. H-) representing the sensitivity of the River Style (this reflects the capacity to rework and influence larger areas of land) and the following number (e.g. -I) indicating the relative sensitivity of adjustment for a given reach, given the potential range of adjustment for that River Style (Table 4.7). The sensitivity of River Styles is ranked as High (H), Moderate (M) or Low (L) with reaches further categorised using numerical ranks from I to III, indicating increases in sensitivity (See Table 4.7 for further description). These variables are then combined to provide a full expression of sensitivity (e.g. H-II would indicate a reach of a highly sensitive River Style which is exhibiting moderate sensitivity) (Table 4.7).

Reach scale sensitivity of the wandering cobble bed river was found to be captured in a measure termed the ‘sensitivity index’.

\[ SI = \sigma(TA/(valley length)) \]  

This was derived by calculating the total channel area (TA) for each aerial photograph survey (1928, 1958, 1964, 1973, 1984, 1993, 2007) and dividing each reach by valley length to normalise the values making them comparable. The standard deviation (\( \sigma \)) for each reach was calculated, which expresses the variation in width for each reach over the time period. This captured differences in the variation of adjustment, providing a direct measure of sensitivity for rivers, especially those with a wandering and braided planform. Hence, this measure is used herein as a proxy for sensitivity within this system, and is referred to as the Sensitivity Index (Table 4.2). This index can be used to quantify and
support qualitative assessments of sensitivity observed during the reach scale analysis of planform change.

Table 4.7: An approach to analysis of river sensitivity. The first part of the procedure assesses the manner and capacity of reach scale adjustment (i.e. the ability of that River Style to rework large areas of land), while the second component appraises the degree and rates of adjustment displayed by any given reach based on analysis of historical channel adjustments. The overall river sensitivity ranking is a combination of these two measures (e.g. M-III).

<table>
<thead>
<tr>
<th>Sensitivity of River Style</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>L-</td>
<td>River Style displays minimal adjustment and undergoes little change in the area of active channel. Rivers Styles are commonly confined or pinned in place by imposed boundaries. Low slope and sediment input may also limit the degree of adjustment the reach undergoes.</td>
</tr>
<tr>
<td>M-</td>
<td>These River Styles display moderate levels of adjustment and are responsive to changes in fluxes over time and large magnitude events, however the rate and extent of adjustment does not rework or influence large floodplain areas as seen in the ‘H’ sensitive River Styles.</td>
</tr>
<tr>
<td>H-</td>
<td>Highly active River Style, with a wide range of adjustment that has the ability to rework or flood large swathes of land adjacent to the channel. Sensitive River Styles are commonly unconfined and types of adjustment include lateral adjustment, avulsion and reactivation of paleo/flood channels. Changes to sediment fluxes can elicit major channel response and planform change.</td>
</tr>
</tbody>
</table>

Reach sensitivity given potential range of adjustment imposed by River Style

| -I                        | Channel undergoes minimal changes to active channel area, and rates of adjustment are slow, given the potential range of adjustment available for that River Style. |
| -II                       | The channel undergoes moderate rates of adjustment and change, displaying changes to the areas of surfaces reworked. The channel does not access the full potential range of adjustment possible for the River Style. |
| -III                      | The channel responds rapidly to events, undergoing a high degree of change. The channel incorporates the whole range of adjustment available for that River Style. |

Latent forms of adjustment

| T                         | Thresholds represent a different type of sensitivity, as the change is not necessarily underpinned by high rates of adjustment. Therefore if a system is nearing a threshold that makes it sensitive a T is added to the end of the ranking. |

The Wandering cobble bed River Style (reaches A-G) has moderate sensitivity and is rated as M (Table 4.7; Table 4.8). Terraces control reach slope and the potential range of adjustment, creating a well-connected river which flushes most of the sediment delivered from upstream.

Within this, reaches A, D and F exhibited the lowest sensitivity, reflected in low sensitivity indices of 22, 13 and 17 respectively. Site G also underwent minimal variation in channel area (sensitivity index of 17). The channel was wider at this site with a greater area that was reworked, but with minimal variation over time (Figure 4.4). These reaches displayed a limited response to large floods, with the channel adjusting within active lateral bar deposits. Average total channel widths for these reaches increased by 22 m on average during the two Q_60 floods (1958 and 2004) illustrating the little influence these large floods have on the floodplain adjacent to these reaches. As such, they are attributed a ranking of grade M-I (Table 4.6; Figure 4.13).

Reaches C and E had sensitivity indexes of 28 and 29. Both reaches comprise forced bends that are pinned against the terrace margin. Channel adjustment in each case was driven by large events that
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reworked floodplain areas adjacent to the meander. Channel response to the 1958 and 2004 floods was seen in an average increase of 56 m in total channel width. These were ranked as moderate sensitivity, with a ranking of M-II (Figure 4.13).

Site B underwent the greatest adjustment exhibited by this River Style, with a sensitivity index of 61, double the value seen for any of the other terrace confined sites. This reflects the wider floodplain pocket, and steep slopes, that allow energy and accommodation space to generate a high degree of reworking. Average increase in channel width following the two Q₆₀ floods (1958 and 2004) was 142 m, representing a significant increase in the area of active floodplain. This site was given a sensitivity ranking of M-III (Figure 3.13).

Reach H had the highest sensitivity index of any reach (196), suggesting a dramatic variability in channel area (Figure 4.6). This unconfined, braided gravel bed river has a high degree of adjustment and response to changes in fluxes. The small-sized, homogeneous sediment is readily reworked at
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high flows, whilst the low slope of the reach limits the transport of this material out of this reach, creating a zone of deposition. Average response to the two Q60 floods in 1958 and 2004 was a 216 m increase in total channel width. Changes to fluxes have had a major impact upon this reach, including base level changes due to variations in the level of Lake Taupo and gravel mining. As a result, this highly sensitive system has undergone significant adjustment, giving it a sensitivity ranking of H-III (Figure 3.13).

The meandering (Reach I) and delta (Reach J) reaches were attributed an M- sensitivity grading, as despite low slopes, grain size is small and the river adopts a sinuous form. However, both reaches underwent minimal changes to all measures of sensitivity, other than long-term narrowing and a decrease in braiding (i.e. islands became attached to the floodplain). This was poorly represented through the sensitivity index, which had high values of 59 and 76. The sinuosity was compared for the meandering channel and the multiple distinct channels for the delta reach, but was misleading, as it resulted in a greater length of channel being recorded when area is averaged for valley length. The low rates of adjustment for both reaches would usually be recorded as systems that have low sensitivity. However, channel narrowing has pushed the system closer towards a threshold where channel capacity is unable to hold the discharge and avulsion will occur. Evidence for this longer-term channel evolutionary trait is provided by paleochannels adjacent to the contemporary channel and head-cut migration forming a secondary channel, which generates shear stresses a magnitude higher than the main channel during a Q₅₀ flood (Smart, 2011). Given the high impact avulsion would have on the channels and land adjacent to both reaches, these River Styles are designated sensitivity rankings of M-III-T (Figure 3.13).
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Table 4.8: Types, patterns and rates of adjustment for the River Styles within the lower Tongariro River and appropriate measures of sensitivity.

<table>
<thead>
<tr>
<th>River Style</th>
<th>Sensitivity Rank</th>
<th>Dominant Patterns of Adjustment</th>
<th>Rate of Adjustment</th>
<th>Measures of Sensitivity</th>
</tr>
</thead>
<tbody>
<tr>
<td>Partly Confined, wandering cobble bed river</td>
<td>M</td>
<td>Floodplain reworking and incorporation back into the active channel through channel straightening and chute cut offs. Mid-channel bars cause widening then reattach to the floodplain as lateral bars. During periods of few large events, lateral bars become vegetated and return to floodplain.</td>
<td>Large magnitude events can create dramatic changes in the channel area reworked (e.g. between 22 – 142 m increase in total channel width in response to Q&lt;sub&gt;60&lt;/sub&gt; flood). Adjustment is greatest in reaches with wider floodplain pockets and more sinuous planform.</td>
<td>- Change in areal measures averaged over valley length to make them comparable. Total channel area measures incorporate all land reworked by the channel over periods &gt;1 year and are less influenced by responses to the last flood event relative to measures such as unvegetated gravel. - Braided index</td>
</tr>
<tr>
<td>Unconfined, braided gravel bed river</td>
<td>H</td>
<td>Increases in sediment causes aggradation and increases braiding and widening, increasing channel number and bar surface area. At high flows, flood channels form across the floodplain, connecting sections of vegetated floodplain to the main channel.</td>
<td>Rapid alterations to channel extent. This reach can undergo significant and rapid adjustment (e.g. 355 m increase in average total channel width following the 1958 Q&lt;sub&gt;60&lt;/sub&gt; flood).</td>
<td>- Change in areal measures. Total channel area is especially useful as it represents the full area of floodplain affected. - Braided index</td>
</tr>
<tr>
<td>Unconfined, meandering, sand bed river</td>
<td>M-T</td>
<td>In most systems, adjustment is dominated by meander migration through translation and extension and cut offs. However, this system is undergoing minimal adjustment, and channel narrowing and contraction is the dominant process. Paleo-channels (which indicate past avulsion) and channel contraction which has decreased channel capacity and increased flooding indicate proximity to a threshold condition.</td>
<td>Slow on-going narrowing with decrease in total channel width of 21 m between 1928 - 2007. Minimal adjustment observed.</td>
<td>- Total channel area -Sinuosity - Rate of meander movement - Likelihood of avulsion- Ability of channel to contain frequent floods, presence of head-cuts creating secondary channels.</td>
</tr>
<tr>
<td>Unconfined, multi-channel delta</td>
<td>M-T</td>
<td>On-going narrowing and infilling of channels (though delta length has been extending, so increasing). Primary channels are still flowing, but secondary channels have become in-filled and incorporated into the floodplain in the past 80 years. Islands form at the mouth of the delta, stabilising and increasing its extent. This reach is nearing an avulsion threshold.</td>
<td>Slow on-going narrowing with decrease in total channel width of 63 m between 1928 - 2007. Minimal adjustment observed except infilling of channels.</td>
<td>- Channel area records extent of infilling (the extent was clipped to remove the influence of delta growth). - Braided index can indicate the formation of islands, which may induce channel narrowing. - Likelihood of avulsion (see above)</td>
</tr>
</tbody>
</table>
Chapter 4: Spatial and Temporal patterns of Reach-scale adjustment

4.4 DISCUSSION

This section initially explains the variability in channel adjustment along the lower Tongariro River. Following this, temporal patterns of adjustment are discussed to identify which natural and anthropogenic drivers elicit the greatest effect within each reach. Finally, the approach to assess sensitivity used within this chapter is discussed.

4.4.1 Explanation of Variability in River Adjustment along the Lower Tongariro River

The sensitivity of River Styles along the lower Tongariro River has been shaped by the long-term evolution of the system, particularly its geologic memory (Brierley, 2010; Brierley et al., 2011). As discussed in Chapter 3 (Section 3.3.3), sediment eroded from the terraces formed by river incision initiated at 1.8 ka has been deposited to form a delta unit, with morphology similar to an alluvial fan. This delta has prograded, resulting in a rapid transition in process zones, from a sediment transfer zone within the terraces, as the channel flushes sediment across the steep gradient composed of lahar lag, to a sediment accumulation zone in reaches downstream of the terraces. This change governs the distribution of River Styles and their character and behaviour.

The terraces act as an imposed boundary that limits the capacity for lateral channel adjustment for the Partly confined, wandering, cobble bed river. Incision has lined the channel with a larger andesitic lahar lag, which can only be transported at very high flows (e.g. > Q20 as illustrated in Chapter 6). As the sensitivity of this reach is determined by its ability to mobilise and rework bar and floodplain surfaces, the large sediment limits the extent of reworking and the ease with which the channel can grade to attain a concave long-profile. As a consequence, the channel within the terraces retains relatively high slopes that range from 0.0111 to 0.0046, decreasing to 0.0026 once the channel extends beyond the terraces (Table 4.6). This change induces a shift in underlying processes, from a competence to a transport-limited system (c.f. Cowie and Brierley, 2008).

The Unconfined, braided, gravel bed river is the most upstream of the River Styles located on the alluvial fan. Lower slopes capture finer gravel material with a homogeneous distribution. The increase in braiding reflects the inability of the channel to transport all the material delivered (i.e. the capacity limit is exceeded). The gravel-sand transition provides a maximum limit to the river’s ability to transport gravel sized clasts. Low stream power conditions disconnect gravel transfer from the receiving environment of Lake Taupo. This depositional environment captures sediment transported through the terrace reaches, creating a very sensitive system which adjusts in response to sediment delivery from upstream, as well as changes to the base-level of Lake Taupo.

The most downstream River Styles (Unconfined, meandering sand bed river and the Unconfined, multi-channelled delta) are characterized by further decreases in slope (0.0016 and 0.0006
respectively). Despite the non-cohesive sand bed and bank materials there is insufficient energy to bring about channel adjustment, and channel movement has been negligible in recent decades. However, longer-term evolution via avulsion is evidenced by palaeochannels. This indicates that channel narrowing promotes a threshold-induced pattern of adjustment. In light of these considerations, this reach is considered to be moderately sensitive to adjustment.

Vegetation is not a key control on adjustment within the lower Tongariro. Partly this can be attributed to the longer temporal time scale considered in this study. Floods which are large enough to rework the boulder sized bed material in the terrace confined section, and elicit planform change, which are the events considered in this study are easily able to uproot the shrubby vegetation which colonises bars, islands and floodplain surfaces within the Terrace. The steep nature of this reach drives the high energy of these floods. A study looking at smaller scales and smaller, more frequent floods would need more consideration of vegetation. In the lower section, downstream of the terraces, willows are more important and are likely to constrain river adjustment in this area. Significant adjustment in the braided reach shows that this reach is competent to remove vegetation at the large floods. However, vegetation likely played a key role in the rate of narrowing in the meandering and delta reaches, due to the low slope as discussed in the following section.

Valley confinement is a key determinant of channel morphology and adjustment through directly fixing and controlling the channel boundaries (Brierley and Fryirs, 2005; Cowie and Brierley, 2008; Fryirs and Brierley, 2010; Jain et al., 2008; Tinkler and Wohl, 1998). In this instance, terrace width reflects long-term reach scale adjustments following the Taupo eruption (186AD). Variations in terrace width provide a major control on contemporary channel adjustment along the lower Tongariro River. A strong positive relationship ($R^2 = 0.93$) was observed between the sensitivity index (calculated as the standard deviation of average channel width over time) and terrace width for each reach (this is fully discussed in Chapter 5). However, both channel banks rarely have direct contact with the imposed boundary of the terrace, and accommodation space is available which would allow wider channels to exist for all reaches within the terraces (A–G). Investigating how process-based controls drive variability in the sensitivity of the wandering gravel bed reach is considered fully in Chapter 5.

**4.4.2 Spatio-temporal Patterns of Influence and Response**

Figure 4.14 provides a summary of the magnitude and direction of the response of each reach to changes in fluxes. Response is measured by changes to long-term (wetted, active and partly vegetated gravel) channel width, providing a measure that relates directly to changes in channel form. This highlights that despite the close proximity of each reach, internal characteristics drove
distinct differences in the mix of anthropogenic and natural influences that elicited change within each reach and the resulting type, extent and rate of changes. These patterns are described below.

Figure 4.14: Channel responses within the lower Tongariro to changes in fluxes. This shows the response of A) the wandering gravel bed reach which exhibited High (H), medium (M) and low (L) levels of sensitivity, B) the braided reach and C) the meandering and delta reaches. Discharge for events with a recurrence interval above 5 years is shown in D) and E) displays changes to fluxes over the time of the study. Rates of change are based on long-term active channel widths from aerial photographs.

The active channel width of the Wandering, cobble bed river was primarily determined by the time elapsed since that last major flood event. For example, narrowing occurred across all reaches between 1964 and 1993, coinciding with few large floods (only three floods occurred, with Recurrence Intervals (RI) of 5 - 8 years). Following this period, an increase in large floods after 1993,
including the 2004 flood (2nd largest flood on record) drove a marked increase in channel area and braiding (Figure 4.14). While the type of response was consistent between all reaches, the rate and extent of change varied markedly, as described through the assessment of sensitivity above. This difference in response is illustrated for the wandering reach in Figure 4.14a.

The impact of flow regulation upon channel planform was assessed for the wandering cobble bed reach. This found minimal evidence for changes in planform characteristics pre and post regulation (Figure 4.10). Moreover, there is no evidence for sustained aggradation or degradation in the period after flow regulation (Figure 4.2). This finding contrasts with the global literature which describe decreases in sediment yields and impacts upon physical integrity following dam construction (Graf, 2001; Kondolf, 1997; Pearse, 2006; Vörösmarty et al., 2003; Williams and Wolman, 1984). Major changes to the flood hydrographs pre and post regulation show a decrease in the number of flood days (with flows > 100 m$^3$s$^{-1}$), whilst the pattern of large magnitude events (> 400 m$^3$s$^{-1}$) remain unchanged. Once flows exceed 100 m$^3$s$^{-1}$ control gates are opened and dams are flushed, maintaining the pre-existing distribution of large floods. As such, the distribution of large floods (with RI > 5 years) have been seen to be responsible for eliciting planform adjustment, whilst smaller floods (RI ~ 100 m$^3$s$^{-1}$) have little impact on planform adjustment (see Chapter 6). The TPDS has retained the geomorphically effective flows that are responsible for the periodic transport and reworking of bed and bar sediments. As a result, little change to patterns and rates of planform adjustment are evident pre- and post-regulation. Although dams created more pulsed patterns of sediment delivery due to the flushing of the materials trapped behind the dams and the regulated flow regime it appears that a range of flows has been retained, and reductions to smaller more frequent floods have not significantly disrupted patterns of sediment transport and planform adjustment within the Tongariro River.

Volcanic eruptions are a major natural impact upon the Tongariro River. Infrequent, large pulses of sediment, are redistributed by the river (Gran and Montgomery, 2005; Major et al., 2000). The ease and speed with which the channel of the Wandering, cobble bed reach flushed the lahar debris following the 1995-96 eruption reflects a resilient system that is able to recover rapidly from events that deliver high loads of fine grained sediment. In comparison, the delta and meandering reaches were observed to narrow over this time, likely a consequence of the sediment (mostly < 0.5 mm) becoming trapped at this location (Collier, 2002). This is supported by observations that this sediment fraction was redistributed across the delta (Genesis Energy, 2000; Smart, 1999). This illustrates the variability in response to a single event for the different types of rivers. Higher slopes (0.011 - 0.005 m/m) and larger bed material (91 – 213 mm) within the terraces are easily competent
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to flush finer material, increasing the resilience of these reaches to this sediment fraction (Table 4.6). Very low slopes on the meandering (0.0016) and delta (0.0006) make these reaches more sensitive to influxes of finer materials.

Alterations to base-level provided a major source of change for the delta reaches of the lower Tongariro River. The regulation of Lake Taupo water levels from 1941 elicited an increase in average water elevation. This is reported to cause a backwater effect, that extended 3 km upstream of the delta, enhancing landward movement of saturated soils and the area of wetland (Eser and Rosen, 2000; Tonkin and Taylor, 1999b). Channel responses can be seen in the on-going narrowing of the sand bed meandering reach, with an overall decrease in width of 21 m. The delta has also exhibited narrowing across this time, with rapid sediment deposition following regulation (Figure 4.15). The increase in deposition in this region has been inferred to be a result of alteration to the level of Lake Taupo, and it is unlikely to be linked to the dam development as the greatest rates of adjustment and infilling occurred pre-TPDS. Whilst the channel may have been narrowing naturally, the rate of adjustment is likely accelerated, increasing the speed at which the delta moves along the narrowing trajectory. Colonisation of the banks of this reach by Willow species may have also increased the rate of narrowing. Once channel capacity has decreased beyond a threshold, the channel is likely to avulse creating a new path into the lake (Smart, 2011). This can be seen in the presence of previous deltas, particularly to the east of the contemporary delta.

Figure 4.15: Aerial photographs of the delta in 1941 and 1958 illustrate the increase in deposition and delta progradation following the regulation of the Levels of Lake Taupo. The change in vegetation with an increase in wetland species is also evident over this time period. Arrows show the same locations on each photograph so the effects can be clearly observed.

The average decrease in base-level also caused widening in the braided reach, as noted by the increase in braiding in the 1958 aerial photograph. The large flood directly before this aerial
photograph was taken is likely to have played a role in reworking much of this surface. However, an increase in base-level is likely to have increased the geomorphic effectiveness of this flood event. Gravel extraction from the 1960-1970s caused channel narrowing within the braided reach, effectively resetting the system. Cessation of mining has resulted in the system moving back along a widening trajectory, though the point of widening is now upstream, closer to Turangi than previously (see 2007 survey). This mining halted the original adjustment, making it difficult to assess what the new channel morphology would have been and what the channel is now adjusting towards. This reach is highly sensitive to changes in flux and provides a direct threat to the town of Turangi.

Despite the complexity of external and internal controls that impact upon the Lower Tongariro River, the pattern and distribution of geomorphically effective flood events provided the major driver of change, especially for the partly confined, wandering, cobble bed river. The delta section responded to a greater number of controls, with gravel mining and alterations to base level also influencing channel planform. However, the extent of change, especially within the braided reach was still driven by the pattern of large flood events. This was illustrated through the rapid increase in channel area following the 1958 flood. This illustrates the importance of the natural flood regime for driving channel adjustment.

Channel adjustment is commonly attributed to pulses or slugs of sediment which are transported downstream, causing a downstream propagation of effects (Bartley and Rutherfurd, 2005; Beven, 1981; Church and Rice, 2009; Macklin et al., 2010; Passmore and Macklin, 2000; Toone et al., in press). Within the terrace confined reach of the Tongariro, evidence of such a relationship between reaches was notably absent. Instead the magnitude of reach scale response was a function of the magnitude of the driving flood event and the internal reach scale controls that determine the sensitivity of the reach to a flood. This also identifies a lack of event sequencing within the system, whereby similar flood events were found to elicit similar degrees of response over time (Beven, 1981). This lack of connection between reaches is best explained by the mechanisms of adjustment. Planform change within the Wandering, cobble bed reach is driven by the reworking of bars, chute channels and floodplain surfaces. While sediment transport is inherently a part of this process, the degree of change is related to the capacity of floods to rework these surfaces. Sediment delivered from upstream is noticeably finer than the bed of the Tongariro River at this point (a $D_{50}$ of 65 mm from the Kaimanawa Ranges compared with 90-230 mm grain size of the lower Tongariro River bed material supports this assertion). In contrast, sediment supply is a more important agent for driving change in the River Styles draining the delta. Narrowing of the meandering and delta reaches was linked to increases in fine grained sediment (i.e. volcanic eruptions) and increases in braiding in the
braided reach was attributed to pulses of gravel delivered during large floods. As each River Style was sensitive to a specific grain size fraction, this limited the extent to which channel response was propagated through the system.

4.4.3 Analysis of River Sensitivity

Measures of channel change and sensitivity should be related to the underlying mechanisms that bring about adjustment and the magnitude of formative events (Hooke and Yorke, 2010). In this chapter, this is achieved by separating the capacity for adjustment for differing types of river from the ease with which adjustment takes place. From this, process-form interactions are linked to derive quantitative measures of channel change. While the designation of River Style reflects the potential range of adjustment for any given type of river, antecedent controls within a particular reach determine the degree and rate of adjustment that is likely to be experienced within that reach. Such thinking builds upon previous approaches to analysis of river sensitivity (e.g. Brierley and Fryirs, 2009; Downs and Gregory, 1993; Fryirs et al., 2009; Rapp and Abbe, 2003; Sear and Arnell, 2006).

Marked differences in river type along short reaches of the Tongariro River demonstrate pronounced variability in the nature, type and rate of adjustments driving channel change. Terraces restrict the degree of adjustment for the Partly confined, low sinuosity, wandering, cobble bed river. The standard deviation of average total channel width was found to accurately express the degree of channel reworking. This relationship is far less evident in the meandering and delta reaches, where averaging the channel width over valley length resulted in different lengths of channel being compared. As a result, total changes in area were used to more realistically reflect the trend of ongoing channel narrowing in this reach. Surian (2008) similarly combined channel width, braiding and bed elevation to provide a comprehensive description of channel change for three rivers in Italy, representing the need to relate measures of channel planform back to the specific mechanisms of adjustment for each River Style.

Assessments of river sensitivity are strengthened when they are framed in relation to longer-term adjustment, as this enables analysis of threshold conditions that may bring about channel change (Lane and Richards, 1997; Schumm, 1980). In this instance, subtle channel adjustments such as the narrowing seen in the meandering and delta reaches of the lower Tongariro River are considered to represent high sensitivity reaches, as slow progressive adjustments are moving the system towards a threshold condition whereby wholesale channel avulsion will occur at some stage in the future. In contrast, the unconfined braided reach is also a sensitive reach, as it displays dramatic changes in channel area. For measures of sensitivity to accurately interpret the likelihood of high channel
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adjustment, they need to consider both active types of adjustment and more latent change which may elicit channel metamorphosis. Longer term observations of channel dynamics (such as the presence of paleo-channels on the floodplain) can provide essential guides to predicting channel change and therefore sensitivity over a variety of timescales and not just those easily observed within short-term records (Sear and Arnell, 2006). Assessments of instability and sensitivity can be misleading if they are not appropriately contextualised within multiple temporal and spatial scales of sediment movement (Macklin et al., 1998).

Predicting future adjustment of a river should be grounded through understanding of the controls upon character and behaviour which drives the system. The capacity for adjustment, and associated rates of channel movement, vary markedly for differing River Styles. These considerations, alongside appraisals of reach-specific sensitivity, provide a platform to integrate channel adjustment and floodplain reworking more effectively within ‘space to move’ rehabilitation initiatives (Piégay et al., 2005; Rapp and Abbe, 2003). Sensitivity describes the threat that a channel has on adjacent land. Sensitive rivers are able to flood and/or rework large areas of floodplain, representing a substantive threat to adjacent infrastructure (Sear et al., 1995). River sensitivity can also be used to pinpoint the most likely nodes of adjustment in response to changes in fluxes, based on the reach-scale response observed. This includes anthropogenic controls such as dams, gravel extraction and floodwork installation, as well as natural controls such as large floods and landslides. Grounding notions of river sensitivity with assessment of River Style adjustment and reach scale evolutionary trajectories provides a dynamic, system-specific template with which to frame management applications.

4.5 CONCLUSION

These findings illustrate that despite the relatively short reach of river (15 km), each River Style had a unique pattern of response. In addition, the extent to which anthropogenic and natural changes to fluxes influenced adjustment was River Style specific. For example, change within the Wandering Cobble bed river was driven primarily by large magnitude floods. However, in the lower reaches, control of Lake Taupo’s water level elicited far greater response. This finding illustrates the need to understand channel response specifically for difference reaches and types of river, as their internal controls may yield them particularly responsive or unresponsive to changes in fluxes. In addition, the approach to sensitivity analysis captured reach-scale differences in response based on the natural capacity of adjustment of each River Style. Such analyses can be used to guide management activities, such as the designation of space to move initiatives, as insensitive reaches require minimal areas of land to support their adjustment whilst more sensitive reaches may require large (possibly unrealistic) areas to support natural functioning. This chapter provides a foundation for analysis of
reach scale controls upon adjustment in the Wandering bed reach (Chapter 5) and the creation of process-based evolutionary trajectories (Chapter 7).
Chapter 5: Controls on Adjustment of the Wandering Cobble bed river

5 CONTROLS ON PLANFORM ADJUSTMENT ALONG THE WANDERING COBBLE BED REACH

5.1 INTRODUCTION
The Wandering, cobble bed reaches were observed to exhibit marked variability in channel adjustment. This chapter analyses internal sub-reach scale controls (i.e. slope, transport capacity) to relate channel adjustment back to underlying processes. This section reviews literature on river channel form, followed by a summary of previous studies that discuss controls on the channel adjustment of wandering river channels.

5.1.1 Controls on Channel Form
Studies which link channel form to key controls date back to the 1950s. This includes, most notably, Leopold and Wolman’s (1957) paper which relates slope to bankfull discharge, finding that braided rivers are characterised by steeper slopes than meandering rivers for a given discharge. Flume studies by Schumm and Khan (1972) further substantiated this, finding increasing slope caused channels to adjust from straight to meandering to braided planforms. Lewin and Brewer (2001) related the degree of braiding to higher excess shear stress and Froude numbers. Church (2006) linked the morphology of alluvial rivers to bed material transport relationships using the Shields number, a process-based measure of sediment transport capacity. Shield’s numbers tended to be lower for wandering and braided systems, due to transport being bedload dominated, requiring more energy to move sediment compared with fine grained suspended-load systems. Additional work by Schumm (1985) recognised the importance of sediment characteristics by relating channel planform to sediment transport regime and bed and bank cohesivity. He found that bed-load dominated systems had a high width-depth ratio and braided planform, whilst mixed-load and suspended-load systems tended to be narrower and meandering. Ferguson (1987) attributed an increase in braiding to increases in bedload, stream power and shear stress. He also cites examples where a change from a meandering to braided planform has been observed due to localised increases in sediment load either across time (e.g. volcanic eruption, deforestation or mine waste disposal) or space (e.g. highly sediment loaded tributary or aeolian dunefield). These studies reflect the difficulties in understanding controls on channel change, due to the dominant and interacting roles both slope and sediment supply and size have to play in determining channel form (as noted by Schumm and Lichty, 1965).

Recent work reflects a further expanded scope, to consider additional types of channels and controls upon form. Eaton et al. (2010) expanded the thresholds presented by Leopold and Wolman (1957) to include intermediate, anabranching river types (which are analogous with the wandering gravel bed
planform), relating the critical slope associated with a change in channel pattern to dimensionless discharge and relative bank strength. Zanoni et al. (2008) documented a decrease in channel braiding in response to natural vegetation regeneration increasing bank strength and island roughness. However, channel form also drives patterns of vegetation age and succession on floodplain surfaces (Beechie et al., 2006). As such, channel form and response indicate and reflect key controls and processes operating within the system.

5.1.2 Wandering Gravel Bed Rivers

This chapter specifically analyses planform adjustment for the Wandering cobble bed reach. As yet, few studies have been carried out which discuss influences upon adjustment within this river type. The wandering planform is defined as exhibiting “irregular sinuous channels, sometimes split around channel islands and in some places braided” (Desloges and Church, 1989: 360). However, incorporation of this river type into assessments of controls upon channel form tended to view this as a transitional form between braided and meandering systems, rather than a type of river in their own right (Eaton et al., 2010; Ferguson, 1987; Leopold and Wolman, 1957). Partly this reflects the diversity in form that is found within the wandering planform, with straight, meandering and braided sub-reaches (Schumm, 1985). These channels have also commonly been confused with anabranching or anastomosing channels. However, mechanisms of adjustment are fundamentally different as these rivers are lower energy, suspended-load aggradational (through vertical accretion) systems. Adjustment is characterised by channel bifurcation due to stable vegetated islands and not active bars as present in wandering systems (Eaton et al., 2010; Knighton and Nanson, 1993; Latrubesse, 2008; Makaske, 2001). In addition, wandering gravel bed systems are characterised by larger sediment than anastomosing rivers and a mixed-load sediment regime (Church, 2002). Nanson and Knighton (1996) defined a type 5 anabranching river form, which is analogous to a wandering gravel bed river. This is described as a laterally active, gravel based system, characterised by higher specific stream power than other anabranching rivers and placed between the braided and meandering forms on the river continuum in mountainous regions.

Few studies discuss controls upon channel adjustment of wandering gravel bed rivers. Desloges and Church (1989) identify multi-channels zones of instability in the Bella-Coola River in B.C., Canada, differentiating stable, single thread zones which alternate with dynamic and sensitive ‘sedimentation zones’ that are characterised by increased bar area, sediment storage and channel adjustment. These multi-channelled zones were found to be located downstream of tributaries and upstream of valley constrictions, where channel deposition is maximised. Sedimentation zones were wider and steeper than the straighter, less dynamic reaches. In contrast, the location of sedimentation zones within the larger Fraser River was linked to the propagation of slugs of
Chapter 5: Controls on Adjustment of the Wandering Cobble bed river

sediment mobilised from reaches upstream (Desloges and Church, 1989). Ham (2005) on the Fraser River predicted that the location of future zones of instability could be assessed by following the sequence of morphological development, whereby bar formation causes erosion and increases sediment supply to the downstream reach.

Passmore and Macklin (2000) analysed the sedimentology of a wandering gravel river in England over the late Holocene, finding that adjustment was driven by the progradation of bars downstream. Sub-reach controls including valley floor morphology, slope and sources of sediment elicited an additional control on adjustment. Periods of large floods drove adjustment by reworking sedimentation zones. Ferguson and Werritty (2009) identified progradation of diagonal bars as the key process in the River Feshie, Scotland. They also noted that change within the system was not a regular predictable process, especially as avulsions created new bars, complicating the patterns of channel adjustment. Bartholdy and Billi (2002) linked the size of flood to the development of the wandering planform. They demonstrated that moderate floods increased sinuosity and caused migration, whilst large floods caused avulsion and chute cut-offs, which maintained the low sinuosity of the reach. White et al. (2010) highlighted the importance of valley width on planform in a confined wandering gravel bed river, whereby width determined riffle and pool locations, with pools present at valley constrictions and riffles located at localised wider sections associated with lower transport capacity during floods. Wishart et al. (2008) documented the influence of gravel mining on the planform of a wandering gravel bed river in northern England, finding that channel adjustment was similar in both the extraction and control reaches, citing the difficulties of differentiating natural and anthropogenic influences on channel form. Winterbottom (2000) discussed channel narrowing in response to flood embankment construction and then flow regulation. This drove flow incision followed by stabilisation of islands and bars by vegetation, illustrating the sensitivity of the wandering planform to changes in flow regime. Understanding which controls drive the location of sensitive zones provides an essential underpinning for managing this river type.

This chapter analyses the key controls upon the location of sensitive sedimentation zones within the lower Tongariro River.

5.2 METHODS

Controls on channel adjustment were analysed at two scales within the Wandering, cobble bed reach of the lower Tongariro River. The largest scale coincides with the reaches that exhibited similar rates and types of adjustment introduced within Chapter 4, included Reaches A-G (Figure 4.1). This primarily assessed the control of valley confinement upon underlying processes including the slope, braiding and channel widths. Whilst the terrace margin plays a key role in directing the path of the
Chapter 5: Controls on Adjustment of the Wandering Cobble bed river

river, at no point is the channel fully confined, and as such it does not provide a direct influence upon channel variables such as channel width and braiding.

This study uses multiple terms to describe different channel and floodplain units. Table 5.1 provides the definition of these terms as they are used within this study.

<table>
<thead>
<tr>
<th>Morphological Descriptor</th>
<th>Explanation</th>
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<tbody>
<tr>
<td>Terrace width</td>
<td>The floodplain width within the terrace margins. This is referred to as terrace width so that the unit is linked back to its evolution, formed from incision creating terraces following the Taupo eruption (1.8 ka.). This also links the floodplain to the type of valley confinement, where steep terraces form the valley margins, and are a source which delivers sediment direct into the channel.</td>
</tr>
<tr>
<td>Floodplain width</td>
<td>The width of the floodplain. This may be the width between the terraces or for the unconfined reaches downstream the description may be unlimited.</td>
</tr>
<tr>
<td>Valley width</td>
<td>Valley width describes the room to adjust and is in most places the same as the floodplain width. If there is no floodplain then valley width is used as it describes the space within the valley which the river can adjust within. For example, in a confined river this will be the width of the river channel and benches if present.</td>
</tr>
<tr>
<td>Floodplain pocket</td>
<td>This refers to discrete, disconnected pockets of floodplain which exist between the channel and the terraces.</td>
</tr>
<tr>
<td>Channel width</td>
<td>Channel width includes the wetted channel and active, unvegetated or partly vegetated bar surfaces and islands. This is also referred to as the long-term channel area.</td>
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Regression analyses are used to explore the relationship between floodplain pocket width between the terraces and channel characteristics (i.e. slope, braiding, channel width and the degree of change in channel width). This aimed to understand the extent of influence exerted by valley confinement upon reach scale controls. In addition, the relationship between floodplain width and terrace top slope, as a proxy for the valley slope pre-Taupo eruption (1.8 ka), was also assessed. This aims to understand whether reach-scale differences in slope (and energy) influenced the width of the floodplain pocket that was created within the terraces. The slope of the terrace surface and floodplain pocket width was measured using 1 m horizontal resolution LiDAR data.

A greater resolution of analysis was undertaken for 18 sites within these larger scale reaches. 15 sites are in the Wandering, cobble bed river, and the 3 downstream sites describe the braided reach beyond the terrace constraints. These sub-reach sites acknowledge high within reach variability in cross-section, bed material and slope characteristics.
Figure 5.1: Map of channel features digitised from 2007 aerial photographs. Study reaches as presented within Chapter 4 are shown as A-H. Sub-reaches are categorised as either single-channelled or multi-channelled sedimentation zones (see Table 5.2). Field analysis was carried out at pool geomorphic units at each reach, with sediment sampled from adjacent lateral bars.
These reaches could be clearly delineated into sedimentation zones with active stores and multiple channels or single channelled reaches with minimal active storage (only small lateral bars). Widths were measured from the 2007 aerial photographs (Figure 5.1) and included the wetted channel and unvegetated and semi-vegetated bar surfaces. 2007 aerial photographs were used as they were most relevant to the time of survey (in 2009 – 2010) and they capture an active phase whereby the river was recently reworked by the 60 year recurrence interval flood in 2004 (Figure 2.21). In this study multi-channelled, sedimentation zones had widths of 118 - 365 m due to substantial sediment storage as bars and islands. These zones exhibited a high capacity to adjust across these surfaces and were considered more sensitive (Figure 5.1; Chapter 4). In contrast, single channelled reaches had widths of 38 – 75 m. These reaches have limited active sediment storage, with small lateral bars present, and were observed to undergo minimal channel adjustment (i.e. low sensitivity) in the past 80 years (Table 5.2; Chapter 4). As such, this captured channel storage, which was found to be directly related to past magnitudes of adjustment, as observed in the aerial photograph history (Chapter 4).

<table>
<thead>
<tr>
<th></th>
<th>Single channelled zones</th>
<th>Sedimentation zones</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Width (m)</strong></td>
<td>38 - 75</td>
<td>118 - 365</td>
</tr>
<tr>
<td><strong>Sediment stores</strong></td>
<td>Minimal in-channel sediment stores are present. Narrow lateral or mid-channel bars may be present. Predominately single channelled reaches.</td>
<td>More complex geomorphic structure with greater active sediment storage than in the single channelled reaches. This includes a complex arrangement of islands, chute channels and mid-channel and lateral bars. The channel has multiple channels or well defined flood channels.</td>
</tr>
<tr>
<td><strong>Reach behaviour</strong></td>
<td>Single channelled zones are observed to undergo lower rates of adjustment than sedimentation zones. Large floods rework localised adjacent floodplain surfaces and channel location undergoes minimal adjustment.</td>
<td>Far more dynamic than single channelled zones. Large magnitude events elicit greater change, including the reworking of floodplain areas, avulsion of primary channels, formation of secondary chute channels, and stripping of island surfaces.</td>
</tr>
</tbody>
</table>

Sites were evenly spaced along the 10 km reach, with each site located adjacent to a pool (for cross-section surveying) and a lateral bar (for bed material analysis) (see Figure 5.1 for site locations). 100 Wolman transect counts were conducted on the coarsest fraction of the lateral bar, providing a measure of the maximum transport capacity of the reach (Brierley and Hickin, 1985; Church and Kellerhals, 1978; Cowie and Brierley, 2008). Water surface slope was extracted using LiDAR flown in 2006 by averaging slopes from 250 m upstream and 250 m downstream of the site. Cross-sections were surveyed for each reach. Due to the fast flows and deep nature of the Tongariro River these required a mix of approaches. Dry and wadeable sections were surveyed using a Sokkia SET530R3 Total station. Deep fast flowing sections were surveyed using a white water kayak which was used to
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take depth measurements using a Hawk digital DF2200PX handheld sonar device. Distance from the water’s edge to the sonar was measured using a MDL Laser Ace 300 range finder so that all measures were taken within the same datum.

Bedload transport equations were used to calculate the ratio of shear stress to critical shear stress, termed the ‘transport capacity’ (Pitlick et al., 2009). Shear stress was deemed as a more appropriate measure of sediment entrainment and bedload transport than stream power as it is a direct measure of the force exerted on the bed (Knighton, 1998; Shields, 1936). This allows field data to be used to express the capacity of a reach to transport the sediment present. The Bedload Assessment for Gravel-bed Streams (BAGS) software was used (Pitlick et al., 2009). This allows the choice of multiple transport equations. Wilcock and Crowe’s (2003) equation was selected as most appropriate for the study site. This is because it calculates sediment entrainment based on a sediment distribution curve, rather than median grain size, and incorporates a hiding function, which is important for the Tongariro system, given the high variability in grain-size (Chapter 6). This equation is ‘surface-based’ meaning it calculates entrainment using the top layer of sediment on the bar. As this work aims to understand how easily bar surfaces can be entrained as an underpinning for channel adjustment the use of a ‘surface-based’ equation is essential. Due to the high uncertainty that is attributed to bedload transport estimates gained from bedload transport equations (Gray et al., 2010; Wilcock et al., 2009), this model was used to calculate transport stage, which describes the ratio of available shear stress to the critical shear stress necessary to transport the sediment in a reach (shown in equation 1).

$$\phi = \frac{\tau}{\tau_r}$$

(1)

$\phi$ is the transport stage, $\tau$ is the shear stress acting on the bed and $\tau_r$ is the reference shear stress (analogous to the critical shear stress) which describes the energy necessary to create “a small but measurable transport rate” (Pitlick et al., 2009: pg. 8). The transport stage was calculated for the $Q^{2.33}$ and $Q^{100}$ recurrence interval floods with discharges of 480 m$^3$s$^{-1}$ and 1500 m$^3$s$^{-1}$ respectively. The $Q^{2.33}$ is commonly referred to as bankfull stage and is commonly represented as the most geomorphically effective flood that is responsible for shaping the system (Andrews, 1980; Emmett and Wolman, 2001). However, within wandering gravel bed rivers, channel width and therefore bankfull discharge varies dramatically along a reach. As such, it is important to assess the competence of a large event, as it is able to highlight how differences in channel width and confinement may influence sediment transport. The influence of $Q^{60}$ on the Tongariro has been recorded twice (Munro, 2004; Smart, 1999). These two flood events in 1958 and 2004 have driven
significant planform adjustment, transferring large volumes of bed material (see Chapter 4). As such, the $Q_{100}$ flood represents the impact of these large, infrequent but highly geomorphically effective floods. The $Q_{100}$ is marginally larger than the 1958 $Q_{60}$ flood ($1500 \text{ m}^3\text{s}^{-1}$ compared with $1417 \text{ m}^3\text{s}^{-1}$ for the $Q_{60}$). However, it is especially important to understand as it is a commonly used measure of large floods as stopbanks and bridges are commonly designed to accommodate the $Q_{100}$ event. Thus, within this work it is used as an expression of the amount of energy that would be generated during infrequent and large flood events within the Tongariro River.

Regression analysis was used to test relationships between controls for these sites. This includes assessing how controls changed downstream for the sedimentation zones and single channelled reaches, including slope and grain size. In addition, the relationship between grain size and slope is assessed for both types of sites. The downstream change in transport stage is also compared between the two types of sub-reach to assess differences in transport capacity.

5.3 RESULTS

Whilst Chapter 4 described spatial and temporal patterns of channel adjustment, this chapter aims to explain which controls are responsible for differences in reach-scale sensitivity. Initially controls on the larger reaches are discussed (Reaches A-G from Chapter 4). Following this, the influence of local scale controls on the distribution of single-channelled reaches and sedimentation zones is explored using data from the 18 sub-reach sites.

5.3.1 Valley Slope Post-Taupo Eruption (1.8 ka) as a Control on Terrace Width

The terraces were created following the eruption of Lake Taupo, 1.8 ka. This section assess whether pre-eruption valley slope (i.e. the slope of the terrace top) influenced the width of the floodplain pockets created between the terraces. This has a process-based rationale, as slope influences energy, and differences in floodplain width represent different volumes of material that have been removed.

A positive relationship was found between the valley slope pre Taupo eruption (1.8 ka) and the width of the floodplain within the terraces, with an $R^2$ value of 0.87 following the removal of an outlier (Reach A) (Figure 5.2). This relationship was much stronger than was observed between terrace width and contemporary valley slopes ($R^2 = 0.38$, Figure 5.4C). High slopes indicate increases in the potential energy within this reach, allowing much larger volumes of sediment to be excavated and creating wider valley widths. The treatment of Reach A as an outlier can be explained due to its position at the upstream location of the reach. Upstream of Reach A, the river has more imposed boundaries and greater bedrock control. It is likely that this reach may have additional upstream influences upon slope which determine its form, creating a different relationship than observed in
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the downstream reaches. However, its existence requires that these results be considered as indicative of a general trend rather than definitive.

![Graph A](image1.png)

![Graph B](image2.png)

Figure 5.2: Correlation between the valley slope pre Taupo eruption (1.8 ka) which is derived from the slope on top of the terrace and the average terrace width (i.e. the floodplain width within the terraces) for each reach. Reach names are annotated. A) shows this relationship with all reaches included, whilst B) shows it with reach A removed as an outlier.

Reach A was an outlier, as slope (0.015) is much steeper than would be predicted for a terrace that narrow (352 m). In comparison, Reach B had much higher slopes (0.015) and terrace widths (678 m). Widths decrease with slope through Reaches C and D (Figure 5.3). An increase in slope at reach E is mirrored by an increase in terrace width. Slopes and width decreases again for Reach F and then both increase at reach G. Sites do not show a linear downstream decrease in slope, with terrace width exerting a greater control (Figure 5.2).
These findings indicate that the valley slope after the Taupo eruption (1.8 ka) played a key role in determining the terrace width and setting patterns of valley confinement for the reaches within.

5.3.2 The role of Terrace Confinement on Reach-scale Adjustment

This section analyses the relationship between floodplain pocket width between terraces and channel characteristics. A positive correlation was observed between the average terrace width for each reach and the braided index, which describes the degree of bifurcation for a reach ($R^2$ value of 0.73). This indicates that as the floodplain width increases, the channel widens and is characterised by more active sediment stores (Figure 5.4). This finding was supported by a positive correlation between terrace confinement and the average width of the active channel (including wetted channel and active unvegetated bar surfaces) ($R^2$ value of 0.6). Weaker correlations were observed between long-term active channel width, which includes partly-vegetated surfaces and the total channel width (both have $R^2$ values of 0.57). As the floodplain width does not directly influence channel width (i.e. there is accommodation space for the channel to be wider at all points), this implies that floodplain width may influence underlying processes which determine channel width.
Figure 5.4: Linear trend between the average terrace confinement and A) the braided index and B) the reach average active channel width (including wetted channel and active bar surfaces) and C) the average valley slope for each reach contained within the terraces (A – G).
A weak positive relationship is evident between slope and terrace width ($R^2 = 0.38$). Naturally slope would be expected to decrease downstream. However, despite noticeable downstream changes in gradient, valley confinement does appear to exhibit some relationship to slope. Reaches are labelled on Figure 5.4C to illustrate that valley confinement has a greater influence upon slope than catchment position. This indicates that reaches located within narrow floodplain pockets tend to have flatter slopes.

Valley confinement was found to have a strong positive relationship ($R^2 = 0.926$) with the sensitivity (i.e. magnitude of response) of each reach, measured using the sensitivity index (see Section 4.2). This describes the variability in channel width by calculating the standard deviation in average channel width between 1928 and 2007 (Figure 5.5). This finding illustrates that floodplain pocket width has a greater influence upon the magnitude of adjustment (i.e. fluctuations in width) than it does on overall channel form (e.g. braiding and width). Valley width represents a primary determinant of channel response within this system.

![Graph image](image.png)

**Figure 5.5:** The relationship between the average terrace confinement and the standard deviation in average total channel width over the period 1928 – 2007. This expresses the sensitivity or the degree of adjustment a reach has undergone.

### 5.3.3 Controls on Sub-Reach Sites

This section describes downstream changes in controls for the 18 sub-reach sites within the wandering – braided reach of river (Figure 5.1). This aims to understand controls upon the location of multi-channelled sedimentation sub-reaches and single-channelled sub-reaches (Table 5.3).
Table 5.3: Description of the controls on each site.

<table>
<thead>
<tr>
<th>Bar Name</th>
<th>Channel Width (m)</th>
<th>Width category</th>
<th>D50 (mm)</th>
<th>Slope (m/m)</th>
<th>Transport Stage (Q^2.33)</th>
<th>Transport Stage (Q^100)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Sand</td>
<td>55</td>
<td>Single channel</td>
<td>162.5</td>
<td>0.0119</td>
<td>2.71</td>
<td>4.53</td>
</tr>
<tr>
<td>Blue</td>
<td>52</td>
<td>Single channel</td>
<td>230</td>
<td>0.0077</td>
<td>1.47</td>
<td>2.09</td>
</tr>
<tr>
<td>Big Bend</td>
<td>189</td>
<td>Sedimentation</td>
<td>212.5</td>
<td>0.0122</td>
<td>2.25</td>
<td>3.09</td>
</tr>
<tr>
<td>Boulder</td>
<td>135</td>
<td>Sedimentation</td>
<td>182.5</td>
<td>0.0111</td>
<td>1.20</td>
<td>2.13</td>
</tr>
<tr>
<td>Poutu</td>
<td>365</td>
<td>Sedimentation</td>
<td>134</td>
<td>0.0064</td>
<td>1.63</td>
<td>1.67</td>
</tr>
<tr>
<td>Red Hut</td>
<td>38</td>
<td>Single channel</td>
<td>140</td>
<td>0.0051</td>
<td>1.86</td>
<td>3.29</td>
</tr>
<tr>
<td>Silly</td>
<td>68</td>
<td>Single channel</td>
<td>115</td>
<td>0.0066</td>
<td>2.44</td>
<td>4.01</td>
</tr>
<tr>
<td>Birch</td>
<td>56</td>
<td>Single channel</td>
<td>120</td>
<td>0.0071</td>
<td>3.30</td>
<td>3.71</td>
</tr>
<tr>
<td>Stag</td>
<td>164</td>
<td>Sedimentation</td>
<td>130</td>
<td>0.0065</td>
<td>1.85</td>
<td>1.94</td>
</tr>
<tr>
<td>Admiral</td>
<td>118</td>
<td>Sedimentation</td>
<td>140</td>
<td>0.0078</td>
<td>1.79</td>
<td>2.35</td>
</tr>
<tr>
<td>Never fail</td>
<td>64</td>
<td>Single channel</td>
<td>115</td>
<td>0.0073</td>
<td>2.23</td>
<td>3.01</td>
</tr>
<tr>
<td>Hydro</td>
<td>62</td>
<td>Single channel</td>
<td>127.5</td>
<td>0.0061</td>
<td>1.36</td>
<td>1.72</td>
</tr>
<tr>
<td>Breakfast</td>
<td>37</td>
<td>Single channel</td>
<td>113.5</td>
<td>0.0030</td>
<td>1.44</td>
<td>1.44</td>
</tr>
<tr>
<td>Island</td>
<td>148</td>
<td>Sedimentation</td>
<td>91.5</td>
<td>0.0044</td>
<td>0.95</td>
<td>1.13</td>
</tr>
<tr>
<td>Bridge</td>
<td>75</td>
<td>Single channel</td>
<td>91</td>
<td>0.0048</td>
<td>1.90</td>
<td>3.60</td>
</tr>
<tr>
<td>Swirl</td>
<td>187</td>
<td>Sedimentation</td>
<td>80</td>
<td>0.0036</td>
<td>0.75</td>
<td>1.01</td>
</tr>
<tr>
<td>Upper island</td>
<td>301</td>
<td>Sedimentation</td>
<td>84.5</td>
<td>0.0028</td>
<td>0.48</td>
<td>0.71</td>
</tr>
<tr>
<td>Bain</td>
<td>75</td>
<td>Single channel</td>
<td>84.5</td>
<td>0.0030</td>
<td>0.86</td>
<td>1.49</td>
</tr>
</tbody>
</table>

5.3.3.1 Median Grain-size and Slope

Downstream trends in median grain size (D_{50}) and slope were compared for single-channelled and sedimentation sub-reaches. Only small tributaries are present within this section of river making external inputs of sediment minimal. Poutu Stream is the largest which flows in directly above Red Hut site. This tributary is regulated, and as a result, sediment supply is diminished.

Median grain size (D_{50}) decreased downstream more rapidly for sedimentation zones compared with single-channelled reaches (Figure 5.6a). In addition, the single-channelled reaches exhibited a greater variation in grain size around the linear trend (R^2 of 0.79 compared to 0.89). As such, sedimentation zones exhibited tighter, more linear and sorted patterns of downstream fining than single-channelled reaches.

Slope exhibited similar downstream trends to D_{50}. Again, sedimentation zones exhibited a slightly greater decrease in slope, with on average higher slopes in the upstream sections (Figure 5.6b). The correlation around this trend was again tighter (R^2 = 0.84) compared with the single-channelled
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reaches ($R^2 = 0.60$), again indicative of a tighter downstream trend in slope for sedimentation reaches.

These findings indicate that slope and sediment size are more in sync with their catchment position for the sedimentation zones. In comparison, the single-channelled reaches exhibit greater variability in slope and grain size relative to what would be expected given their catchment position.

![Figure 5.6: Relationship between A) median grain-size (D50) and B) slope and distance downstream from the most upstream drainage point of the Waipakihi River.](image)

The relationship between slope and grain size was compared for the single-channelled and sedimentation sub-reaches. Single-channelled reaches displayed a highly scattered distribution around a linear trend with a weak $R^2$ value of 0.31 (Figure 5.7A). In contrast, the sedimentation sub-reaches exhibited a tight linear trend with a $R^2$ value of 0.97 (Figure 5.7B). This indicates that
sediment size and slope is graded to form a balance within these reaches. It is likely that sedimentation zones allow greater reworking, wherein the channel has the capacity to adjust its slope relative to its grain size.

Figure 5.7: Relationship between slope and grain size for A) single channelled reaches which are characterised by minimal sediment storage features and B) sedimentation zones composed of active bar surfaces.

5.3.3.2 Transport Capacity
The Wilcock and Crowe bedload equation was used to calculate the transport stage (ratio of shear stress to critical shear stress) for each reach (Pitlick et al., 2009; Wilcock and Crowe, 2003). This describes capacity of a reach to transport the size of sediment present at a location.

Figure 5.8 compares transport stages for single-channelled sub-reaches which undergo minimal channel adjustment and sedimentation zones with greater storage of sediment. Single-channelled
reaches were found to exhibit greater excess shear stress on average than the sedimentation reaches during the $Q^{2.33}$ (flood with a recurrence interval of 2.33 years) flood. This gap becomes more evident for the larger $Q^{100}$ flood. In general, sites with higher excess shear stresses and transport capacities are more able to transport their bed material. Sites with lower excess shear stresses store excess sediment supplied as bars. While scatter was evident, correlations were tighter for the sedimentation zones, with $R^2$ values of 0.6 and 0.7 for the $Q^{2.33}$ and $Q^{100}$ discharge events respectively compared with 0.22 and 0.24 for the single-channelled reaches (Figure 5.8). This again indicates that bed material within the sedimentation zones has adjusted to reflect the transport capacity of a reach (determined by its slope) and its catchment position.

![Figure 5.8: Correlation between the transport stage (ratio of shear stress to critical shear stress) and distance downstream from the most upstream point of the Waipakihi River. A) shows transport stage for a 480 m$^3$s$^{-1}$ flood event with a recurrence interval of 2.33 years, and B) is for a 1500 m$^3$s$^{-1}$ flood event with a recurrence interval of 100 years. Sites are differentiated based on in-channel sediment storage into single channelled reaches and sedimentation zones (Table 5.2).](image)

Figure 5.9 shows downstream patterns in transport stage. Single-channelled reaches exhibited a greater increase between the $Q^{2.33}$ and $Q^{100}$ events compared with sedimentation zones. Whilst
single-channelled reaches underwent an average increase in transport stage (dimensionless) of 0.93, the sedimentation reaches only increased by 0.39. During large floods, flow is able to be dissipated across bar surfaces in the sedimentation zones, limiting the energy available to entrain sediment. In contrast, single-channelled reaches concentrate energy, and an increase in flood magnitude, greatly increases energy.

Figure 5.9: Map of the transport stage (ratio of dimensionless shear stress to critical shear stress) for two flood events. The Q$^{2.33}$ has a discharge of 480 m$^3$s$^{-1}$ and the Q$^{100}$ flood which has a discharge of 1500 m$^3$s$^{-1}$. This is underlain by a detrended DEM and the channel planform in 2007, representative of the time of survey data used to calculate transport stage.

Transport capacity decreased overall downstream. Overall, single-channelled reaches (e.g. Sand, Red Hut, Silly, Birch, Never Fail and Bridge) had noticeably higher transport capacities than their neighbouring sedimentation zones (e.g. Boulder, Poutu, Stag, Admiral, Island, Swirl and Upper...
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Island), especially during the $Q^{100}$ event (Figure 5.9). The highly sensitive Reach B which includes the sub-reach sites of Big Bend, Boulder and Poutu Lower had noticeable lower transport capacities than adjacent reaches. This was especially pronounced at Poutu site, as the island dissipates energy (transport stage value of 1.7). Similarly, values within the reworked meander bend (Reach E) were lower than those in the single-channelled section upstream, with values of 1.9 - 2.4 compared with 3.3 - 4.0. Thus, these sedimentation zones indicate a decreased capacity to flush the sediment delivered during large flood events.

Exceptions to this trend include Blue Pool (the transport capacity of this bar is dealt with extensively in Chapter 6) as a decrease in slope from 0.0119 in the upstream reach to 0.0077 m/m limits the transport capacity (Figure 5.9). Hydro Pool in the lower reach has a moderate slope but larger grain size (slope of 0.0061 m/m and $D_{50}$ of 127.5 mm) which creates a low transport capacity of 1.72 during the $Q^{100}$ (Table 5.3). Breakfast Pool also exhibited a low transport stage of 1.44, again controlled by low slope of 0.003 m/m within this reach. Sediment is delivered from terrace incision, making it larger than sediment which is delivered from hydrological forces (i.e. sourced from upstream, see Chapter 6). In addition, floodworks and riprap on the outside of the bend at Hydro Pool and Breakfast Pool have acted to retain the reaches in their current position since the 1970s, altering the ‘natural’ patterns of confinement.

Artificial confinement at the Bridge site is reflected in a high transport capacity of 3.6. The rapid decrease in energy downstream of the bridge can be seen in the reduction of transport capacity from 1.01 at the top of the braid to 0.7 in the mid braid at Upper Island Pool (Figure 5.9). Bain Pool exhibited an increase in transport capacity of 1.49 due to slightly higher slopes at this point and fewer active bar surfaces to dissipate energy during higher flow events. These relationships help to explain the downstream transition in the type of river and associated behavioural regime.

5.4 DISCUSSION

This section discusses the influence of local controls on the distribution sedimentation zones and single-channelled reaches which exhibited less active storage and planform adjustment. Findings are compared with the literature.

Terrace width (i.e. the width of the floodplain within the terrace) exerted a primary control on channel adjustment. This was shown through a strong positive relationship ($R^2 = 0.926$) between the sensitivity index (standard deviation of average channel width from 1928 - 2007) and terrace width. This indicates a direct control on the range of adjustment each reach underwent (Figure 5.5). In comparison, the correlation between terrace width, and channel width and the braided index were less well defined ($R^2$ values of 0.6 and 0.73 respectively). This indicates that the control of terrace
width upon channel adjustment was greater than on channel geometry. In most systems, valley confinement influences channel morphology and adjustment through directly fixing and controlling the channel boundaries (Brierley and Fryirs, 2005; Cowie and Brierley, 2008; Tinkler and Wohl, 1998). For example, White et al. (2010) document a case of terrace width within a terrace confined wandering gravel-bed river exerting a control upon the location of pools and riffles. However, within this system at no point are both channel boundaries dictated by the valley margins as the channel does not accommodate the full floodplain pocket available. As such, the underlying processes which created this difference in width play a key role in determining contemporary adjustment.

Patterns of valley confinement are most commonly controlled by downstream changes in process zones, allowing the deposition and storage of material once slope falls below a specific threshold (Jain et al., 2008). However, within the Tongariro system, valley confinement is determined by the volume of sediment that has been eroded since the Taupo eruption (1.8 ka). Hence, terrace width reflects long-term reach-scale differences in the capacity of the channel to rework and erode sediment. This was supported by the relationship between valley slope pre-Taupo eruption (1.8 ka) and terrace width, as slope determines the potential energy within a reach, creating marked differences in erosion and width for each reach.

Terrace width constrains the type of adjustment a reach can undergo by influencing channel alignment, ultimately shaping the sensitivity of a reach. Narrow terrace widths limit channel migration and its ability to adopt a sinuosity which balances sediment and water fluxes. This forces the location of channel meanders where they abut the terrace margin, creating a straighter channel between these forced bends. These sub-reaches are less prone to adjustment, as exhibited in the less sensitive reaches A, D and F (Figure 5.10). Sub-reach C, in comparison, underwent a greater degree of adjustment. However, this was concentrated in the sub-reach section that runs from one terrace margin to the other, allowing less forced adjustment and the formation of chute cut-offs and island formation adjacent to the primary channel (Figure 5.10). Hooke (2007) describes the adjustment of a meander pinned against a terrace, finding that these bends reduced the transfer of sediment. However, she was unable to observe a relationship between sensitivity and reach controls. In the Tongariro River these forced bends are unlikely to inhibit sediment transport, especially as the narrower reaches tend to have higher transport capacities and slopes (Figure 5.8; Figure 5.9). However, they are a primary control on the channel alignment (i.e. sinuosity), which could explain a forced higher transport capacity for narrower reaches.
Existing literature most commonly indicates that the balance between sediment size and supply, for a given slope and discharge are the key drivers which determine channel form (Church, 2002; Lane, 1955; Lane and Richards, 1997). Competing views exist on whether an increase in braiding is due to an increase in flow strength (determined by discharge and slope which can be combined to express shear stress) or the supply and type of sediment delivered to the channel (Kleinhans, 2010; Leopold and Wolman, 1957). It is likely that channel form is a product of the balance of both these factors.
Interestingly, there is little evidence that sediment size or slope were greater at the wider (and more braided) sedimentation zones along the lower Tongariro River. However, a strong correlation with an $R^2$ value of 0.97 was observed between slope and median sediment size ($D_{50}$) for the sedimentation zones with bars compared to only 0.3 for the single-channelled reaches. This indicates that sedimentation zones have adjusted their slope through altering their planform to balance the size of sediment available at a given point. Correlations between transport capacity and distance downstream were tighter for sedimentation zones, with $R^2$ values of 0.59 and 0.69 for the $Q_{2.33}$ and $Q_{100}$ floods, compared with 0.22 and 0.24 for the single-channelled reaches. Sedimentation zones exhibited greater grading to their catchment position, with grain size and slope characteristics that were more appropriate for that position. Valley confinement is wider for these reaches, allowing the channel to adjust its form so that sediment size and slope obtain a balance. In contrast, the terrace margins control that planform of the single-channelled reaches impeding their ability to adjust their form, creating greater variability in slope and grain size (Figure 5.10). Previous studies suggest that rivers adjust their form to transport the load delivered (Davies and McSaveney, 2006; Ferguson, 1981; Nanson and Huang, 2008). Valley confinement impedes the ability of the Tongariro to adjust its form in response to the sediment flux, accordingly altering its sediment transport capacity.

Many studies analysing controls on channel planform associate an increase in braiding with an increase in flow strength or sediment yield (Church, 2006; Eaton et al., 2010; Ferguson, 1987; Leopold and Wolman, 1957; Lewin and Brewer, 2001; Schumm and Khan, 1972). Whilst this may determine the ‘type’ of river, the opposite relationship of controls were found to determine localised increases in braiding. For example, local decreases in transport capacity and slope within this type of river determine where sediment is deposited and thus the extent of braiding. This makes it especially important to assess sub-reach scale attributes when considering the sensitivity of a particular site.

This raises the question of what controls the transport capacity of the single-channelled reaches, such that they have enough energy to flush the sediment, yet they do not tend to erode and rework their margins. Most commonly, low adjustment is driven by imposed boundaries such as bedrock or valley margins. Whilst the valley margins do not directly restrain the width of the channel, they force the channel planform (as described in Figure 5.10), creating low sinuosity reaches which undergo minimal lateral reworking within this system. Straight channels are symmetrical and energy is distributed evenly across the channel, rather than exhibiting the energy concentration on the outer bank created by helical flow patterns observed at channel bends (Bridge, 2003; Knighton, 1998).
Thus, symmetrical channel geometry lacks the erosive potential of channel bends, explaining why greater floodplain reworking was observed adjacent to the channel bends. Previous studies including Hoyle et al. (2008) and Rinaldi (2003) associated sinuous reaches with greater sensitivity. Thus, at single-channelled reaches, narrower valleys and pinch-points fix the channel in place, creating a higher transport capacity than would be predicted for a given catchment location. This may be linked to path dependency, whereby, once the river has incised, it is difficult for the channel to adjust its form, unless a meander develops. This highlights the fundamental role of flow alignment on channel adjustment within this system.

Downstream bars are fixed in place by floodworks including riprap and levees, such as Hydro and Breakfast Pools (Figure 5.10). Bridge Pool is constrained by the development of the State Highway 1 Bridge, confining it and creating a very high transport capacity (Figure 5.9). Bedrock outcrops act to confine the river at Red Hut and force a straight planform at Birch and Silly Pools, creating high excess shear stresses (Figure 5.9; Figure 5.10). These high shear stresses convey sediment delivered through the sub-reach to downstream areas with lower competence where it is stored (c.f. Hooke, 2003b). Other reaches showed minimal reworking due to low slopes that are unable to entrain the boulder lag creating low transport stages, such as observed at Blue and Breakfast Bars (see Chapter 6) (Figure 5.9). As such, these bars can flush the active fraction, but are less competent at reworking their bed and driving planform adjustment.

This diversity in controls highlights how local geomorphic factors provide a complicating factor in understanding what is driving channel change (Lane and Richards, 1997). This is especially true within the Tongariro River, due to its complex setting and history. These can be separated into visible controls such as bedrock outcrops which are easily observed and hidden controls such as the lahar lag, which are more difficult to observe and assess, but they are just as important for controlling contemporary adjustment.

Understanding channel adjustment within a wandering gravel bed system relies on determining which controls influence the location and sensitivity of sedimentation zones. Within the literature these zones have been associated with the propagation of sediment waves which migrate as bars gradually moving sedimentation zones downstream (Desloges and Church, 1989; Ferguson and Werritty, 2009; Passmore and Macklin, 2000). However, sedimentation zones within the Tongariro River did not migrate downstream. Instead, reaches could be separated into sensitive, frequently reworked reaches with bars or static narrow channels with minimal sediment stored. These active reaches exhibit similarities to the sedimentation zones in the Bella-Coola River described by Desloges and Church (1989). They identified locations upstream of valley constrictions and
downstream of tributary confluences as sedimentation zones, which tended to be straighter and steeper than less active reaches. Within the Tongariro system, tributary confluences are minimal, and those that are present were not geomorphically effective. Localised valley constrictions were less important than patterns of valley confinement, which provided a key control on the capacity of adjustment of a reach. Within this broader setting, local controls impede reach adjustment, such as bedrock outcrops and imbalances between lahar lag materials and the energy necessary to entrain it.

Wider terrace widths indicate zones of greater reworking and energy in the past, as the channel was able to migrate across the surface, entraining and reworking more sediment than in the single-channelled reaches. As a result, the contemporary channel is able to adjust within these floodplain pockets, adjusting its form by adopting more channels or altering its sinuosity so that the bed material is in sync with the form and slope of the channel. This provides a process-based rationale as to why adjustment is concentrated within these zones, and the influence of valley confinement upon sediment transport processes.

5.5 CONCLUSION

The terrace confined, wandering cobble bed stretch of the lower Tongariro River demonstrated the importance of understanding the evolution of the system, to underpin analysis of contemporary adjustment. Pre-terrace floodplain slopes influenced the volumes of sediment eroded and the width of the floodplain pockets inside the terraces. This set the boundary conditions for the contemporary system. Wider floodplain pockets allow a greater range of adjustment and sinuosity and the channel may adopt its form to balance slope and the input of sediment. In contrast, narrow floodplain pockets impede this adjustment, creating rivers with forced, higher slopes and energy, which flush sediment, limiting the volumes stored and the size of bars. Thus, landscape evolution and history is key for understanding these processes. Wandering gravel bed rivers are still poorly understood due to their location on a continuum between the meandering and braided planforms. In addition, the wandering morphology has been found to adjust differently in different settings, with some systems undergoing gradual progradation of bars, whilst other display stochastic, complex relationships which are not the product of gradual adjustment (see Discussion above). The Tongariro fits within the latter category. This complexity highlights the importance of assessing morphodynamics of wandering gravel bed rivers, so that the sensitivity dynamics of a specific system can be understood, and appropriate management can be designed.
6 A NOVEL APPROACH TO MAPPING BAR BED MATERIAL SIZE ACROSS BAR SURFACES AND ANALYSING BAR REWORKING

6.1 INTRODUCTION

“Bars are key expressions of river behaviour providing information about active processes and the sediment regime at the reach scale”

(Rice et al., 2009; pg. 709).

Analysis of channel adjustment can be strengthened through local-scale process-based studies which offer insights into what controls drive channel adjustment. In this vein, interpreting the process-form relationship of bar complexes has been a long-term aim of fluvial research (Bluck, 1979; Bridge, 2003; Church and Jones, 1982). Bar topography and grain-size composition underpin insights into the mechanisms of channel adjustment, reach-scale sediment regime and bedload transport. These linkages are especially apparent in wandering-braided systems, where bar morphology and reworking drive planform adjustment (Surian et al., 2009). Church and Jones (1982: pg. 292) proposed that “description of these features and consideration of their origin is tantamount to describing the form and enquiring into the stability of the entire river channel”. However, despite their importance, challenges associated with capturing precise and spatially continuous grain size data across these surfaces have limited approaches to modelling sediment entrainment and, as a result, bar reworking (Verdú et al., 2005)

River systems which store material as bars can be differentiated into either ‘competence-limited’ systems, where sediment size means it is only mobilised infrequently and thus is stored within the system, or ‘capacity (or transport) -limited’ systems, where sediment is stored due to the channel being unable to transport the volume of material supplied to the channel (Church and Jones, 1982; Cowie and Brierley, 2008; Hicks and Gomez, 2003). Bars may be further differentiated into ‘hydraulic’ elements that form in response to deformation of the channel bed, or ‘storage’ elements which store sediment being transported through the reach, or bars that reflect a combination of both processes (Church and Jones, 1982). Different mechanisms of bar formation play a key role in determining the range of sediment sizes within a bar, and thus the size of event (or events) necessary to entrain materials that makeup these different units. This is not always straightforward as the sediment distribution across bars is commonly heterogeneous reflecting distinct differences in depositional and erosional environments (Rice and Church, 2010).

Bars are commonly considered to be made up of a resilient, stable ‘nucleus’ around which more mobile units are deposited on the falling limb of floods (Bluck, 1976; Leopold et al., 1964). This node
of coarse sediment indicates flow competence during high magnitude events, whereas the smaller, more active fraction is reworked more frequently (Brierley and Hickin, 1985). Materials that makeup the coarse node, also referred to as a ‘unit’ bar, may subsequently be reworked across the bar surface, as secondary channels redistribute sediment across the bar (Rice et al., 2009; Rice and Church, 2010). Ashworth (1996) presents a model of mid-channel bar development based on a flume experiment. Flow convergence causes scour, resulting in a decrease in transport capacity behind this scour and deposition at the head of the bar. Flow is then diverted around these materials, increasing deposition, particularly at the more sheltered downstream extent in the lee of the bar, causing further bar growth. These models present a rationale for the processes driving bar development and reworking. Accordingly, residence times and reworking differ for within bar units (Brown, 1997). However, attempts to quantify or measure this have been scarce due to the high diversity in topography and grain size across bar surfaces.

Modelling sediment entrainment and the reworking of bar surfaces is an on-going challenge within fluvial geomorphology. In part, this is attributable to the difficulties in characterising grain size across bar surfaces in the field, due to the heterogeneity of these surfaces (Verdú et al., 2005). This is essential as, within gravel bed rivers, the distribution of grain-size across a bar is a greater reflection of bed shear stress than more simple models of depth (Bridge and Gabel, 1992; Bridge, 2003). Previous attempts to measure bar reworking include the use of sediment tracers or painted patches to measure transport on bar units (Carling et al., 2006; Laronne and Duncan, 1992; Laronne et al., 2001; Mao and Surian, 2010; Surian et al., 2009a), capture of sediment in traps or pits (Bridge and Gabel, 1992), topographic surveys for sediment budgeting (Brasington et al., 2000; Fuller et al., 2003a; Williams et al., 2011) and aerial photography to depict rates of adjustment (Church and Rice, 2009; Rice et al., 2009). However, these approaches measure sediment transport as or after it occurs. Models which predict sediment entrainment at the resolution with which sediment varies across bar surfaces are notably absent.

Analysis of bar reworking is a key component in determining the geomorphic effectiveness of flood events, defined as the “ability of an event or combination of events to shape the landscape” (Wolman and Gerson, 1978: 190). Ascertaining which floods drive geomorphic adjustment supports analysis of channel stability within river restoration and management agendas (Doyle et al., 2007). Bankfull flows are commonly cited as channel forming events, as flow is concentrated within the banks, before being dissipated on the floodplain during higher flows (Leopold et al., 1964; Sambrook Smith et al., 2010; Williams, 1978). However, recent studies have suggested that multiple discharges may be responsible for shaping channel morphology. For example, Phillips (2002) found that ‘bi-
modal dominant discharges shaped channels, whereby flows below bankfull flush fine sediment and maintain channel shape, whilst rarer floods rework coarse sediment and erode channel banks. Surian et al. (2009a) measured bedload transport in a braided system, finding that floods with a discharge of 20 - 50 % of bankfull could mobilise the wetted channel, whilst bankfull discharge was necessary to rework higher elevation bar surfaces. Predicting future adjustment is reliant on understanding the role played by different magnitude/frequency flood events in shaping channel morphology through the reworking of bar surfaces.

Primary approaches to analysis of bed material size are framed around manual measurements of grain-size using techniques such as Wolman transect counts or sieving (Church and Jones, 1982; Hoyle et al., 2007; Kondolf et al., 2003a; Rice and Church, 2010; Wolman, 1954). These techniques rely on measuring grain-size (typically between 100 – 400 clasts or 1% of bar for sieving) to establish sediment size distributions for a specific unit or bar (Rice and Church, 2010). This limits grain size comparison to between units or bars, reducing the ability to analyse sediment size distribution within units and across the bar as a whole. Recent advances in geomatics and image processing technology and associated processing techniques, including Terrestrial Laser Scanning (TLS), and photogrammetry, have greatly increased the resolution and scale of data capture which can be used to derive grain size information (Carbonneau et al., 2005; Milan and Heritage, 2012). However, to date, most TLS based investigations have been based on methodological development and ‘patch’ scale grain size mapping (Heritage and Milan, 2009; Hodge et al., 2009). Rychkov et al. (2012) expand this scale by demonstrating changes to grain size over three flood events on a 80 by 40 m patch in a braided river. However, this work primarily concentrates on the development rather than geomorphological applications of this method. In contrast, other remotely sensed data are extracted at lower resolution, such as using photogrammetry on aerial photography to map grain size distributions at the reach scale (Carbonneau et al., 2005; Verdú et al., 2005). These recent investigations represent progression towards characterising grain size across bar surfaces. Nonetheless, as yet, applications of this technology to describe grain size distribution at the bar scale and consideration of the geomorphic effectiveness of flood events at reworking these deposits is absent.

This study uses TLS to (i) map grain size distribution at a high-resolution across four bars and (ii) develop a novel analytical approach to calculate sediment entrainment and bar reworking across a range of flood magnitudes. This approach is used to offer insight into wider geomorphic questions including (i) what are the mechanisms of bar formation and how does this determine bed material size distribution across the bar, (ii) how do reach scale controls (slope, valley confinement) alter
patterns of bar reworking and (iii) which flood events are geomorphically effective to rework bars within the Tongariro River.

6.2 METHODS

6.2.1 Site Selection

Four bars were selected along a 15 km reach of the Tongariro River (Figure 6.1). The upstream three bars (Blue, Red Hut and Breakfast Bars) represent upper, middle and lower sections of the Partly confined, wandering cobble bed river. The most downstream bar (Bain Bar) is located downstream of the terraces within the braided reach. Sites were selected to represent downstream changes in grain-size, slope, valley confinement and the degree and rate of historical channel adjustment (see Chapters 4 and 5; Table 6.2). All bars fit the classification of lateral, compound bars as they are situated on relatively straight sections of channel (sinuosity < 1.07), and their morphology has been shaped by both depositional and erosional processes and events, creating diverse, complex features (Bridge, 2003; Rice et al., 2009). The Tongariro River has a long history of trout fishing, which has resulting in the naming of all the pools along the river. As such, each bar studied in this work is named based on the pool that it is adjacent to, to provide consistency with other work carried out on the lower Tongariro River (c.f. Cooper and Cooper, 1975; Grey, 1978; Hindle, 1995; Smart, 2005).

Aerial photographs were analysed to assess the recent evolutionary adjustments of the bar, formative events and the mechanism of bar development. This reach-scale historical context is used to contextualise in-depth bar sediment relationships. Details of aerial photographs used are supplied in Table 4.1.

Each bar was classified into geomorphic units which describe areas with distinct morphology and grain size distributions (Table 2). This recognises that bars are not homogeneous sedimentological entities, but a product of multiple formative events and processes (Bluck, 1987; Brierley, 1989; Church and Jones, 1982; Leopold et al., 1964; Rice and Church, 2010). RTK-GPS was used to map within bar units and Arc-GIS was used to map the distribution of units based on topography and grain size data from the TLS scans.
Table 6.1: Description of geomorphic units used to define the different morphological and sediment characteristics across the bar surfaces. These are based on units described by Bluck (1971; 1982).

<table>
<thead>
<tr>
<th>Bar unit</th>
<th>Morphological and sediment distribution characteristics</th>
<th>Process-form relationship</th>
</tr>
</thead>
<tbody>
<tr>
<td>Bar head</td>
<td>Coarsest locale of the bar commonly found at the bar head or on the apex of bends next to the channel. Surface commonly slopes up to meet the flatter supra-platform.</td>
<td>This unit is scoured during high flow, representing the highest energy environment. It may be an extension of a coarse riffle located in the wetted channel.</td>
</tr>
<tr>
<td>Bar tail</td>
<td>Depositional feature at the downstream extent of the bar which is made up of smaller grains often predominantly sand.</td>
<td>This unit is protected by the coarser, higher elevation deposit upstream. Smaller grains are deposited here during the falling limb of floods.</td>
</tr>
<tr>
<td>Supra-platform</td>
<td>Middle, flatter section of the bar, between the bar head and bar tail, which is essentially the bar top. Commonly exhibits a downstream fining trend in grain size (Rice and Church, 2010). Bars which exhibited pronounced downstream fining within this unit were separated herein into upstream (US) and downstream (DS) supra-platform units.</td>
<td>Comprises the main bulk of the deposit which is protected by the bar head and has comparatively lower shear stress during floods. May have vegetation starting to colonise more protected sections which can create topographic irregularities. Preferential pathways of flow can also carve out lower zones with coarser sediment.</td>
</tr>
<tr>
<td>Back channel</td>
<td>Channel which is cut around the back of the bar and only inundated during high flow. This surface is lower than the adjacent supra-platform and may have fine grained material superimposed which is captured during the falling limb of floods.</td>
<td>During floods, flow scours the channel. In wandering gravel bed rivers this channel may have been a previous primary channel which has become depositional as a mid-channel bar has transitioned to a lateral bar.</td>
</tr>
<tr>
<td>Fine-grained deposit</td>
<td>Localised area of fine grained material or smaller clasts deposited on the bar at a location which is not the tail. This was sometimes observed at the upstream location of the bar when the larger material of the bar head was located at the apex of the bar rather than the upstream section.</td>
<td>Localised protected areas cause a depositional environment to form within the bar surface.</td>
</tr>
</tbody>
</table>
Table 6.2: Characteristics of each bar.

<table>
<thead>
<tr>
<th>Bar Name</th>
<th>Blue</th>
<th>Red Hut</th>
<th>Breakfast</th>
<th>Bain</th>
</tr>
</thead>
<tbody>
<tr>
<td>Distance downstream(\text{&amp;})</td>
<td>57.63</td>
<td>60.44</td>
<td>65.15</td>
<td>68.04</td>
</tr>
<tr>
<td>Bar Area (m(^2))</td>
<td>2163</td>
<td>7775</td>
<td>4374</td>
<td>4641</td>
</tr>
<tr>
<td>Valley Width (m)</td>
<td>246</td>
<td>80(^*)</td>
<td>80</td>
<td>Unconfined</td>
</tr>
<tr>
<td>D50 (mm)(^A)</td>
<td>230</td>
<td>140</td>
<td>113.5</td>
<td>84.5</td>
</tr>
<tr>
<td>Slope (m/m)</td>
<td>0.00766</td>
<td>0.00508</td>
<td>0.00298</td>
<td>0.00304</td>
</tr>
<tr>
<td>Sensitivity index(^*)</td>
<td>22</td>
<td>61</td>
<td>17</td>
<td>196</td>
</tr>
<tr>
<td>River Style</td>
<td>Partly confined, wandering, cobble</td>
<td>Partly confined, wandering, cobble</td>
<td>Partly confined, wandering, cobble</td>
<td>Unconfined, braided, gravel</td>
</tr>
<tr>
<td>Process zone</td>
<td>Transfer</td>
<td>Transfer</td>
<td>Transfer</td>
<td>Accumulation</td>
</tr>
<tr>
<td>Terrace height</td>
<td>25 m</td>
<td>12 m</td>
<td>7 m</td>
<td>None.</td>
</tr>
</tbody>
</table>

\(^*\) This is done at the sub-reach scale and expressed as the standard deviation of active channel width over time. See Chapter 4.

\(^A\) 100 Wolman clasts sampled at the coarsest locale of the bar

\(^\text{Red Hut}\) is located at a localised pinch in the valley. The channel width is up to 1km upstream and 250 m downstream.

\(^\text{&}\) Distance downstream from the most upstream drainage point in the Waipakihi River.

Figure 6.1: Location of study bars with contemporary channel form from 2007 aerial photographs.
6.2.2 Field Work

Each bar was surveyed using a Leica ScanStation 2000 to map bar morphology and surface roughness. Three reflective targets were used to georeference scans. To minimise error each scan had at least two targets that were present in the previous scan. Scans were set to a maximum point spacing of 5 cm at a distance of 50 m. Scan view fields were manually set in the field using a laptop and run through Cyclone 7.0. Scanning stations were positioned close to each bar’s perimeter and were configured to ensure considerable overlap between adjacent point clouds, thus maximising line of sight and reducing the effect of larger clasts creating shadow areas in their lee. This configuration required between 7 and 11 TLS scans per bar. A Trimble RTK-GNSSR8 Global Positioning System was used to create a benchmark network which established an accurate survey mark on each bar and a Sokia SET530R Total Station was then used to survey scanner and target positions. The benchmark network was configured using existing LINZ geodetic survey marks (with an order accuracy of 3) which were logged using the base station for 5 hours within this study (see Land Information New Zealand (LINZ), 2009 for further details). The rover was used to establish new survey marks by logging for 30 minutes on each bar, resulting in a horizontal precision of between 0.005 - 0.008 m, a vertical precision of between 0.007 - 0.12 m and a root mean squared error of between 0.007 - 0.011. A second location on each bar was logged for thirty minutes to provide a backsight location,
allowing for surveying with the total station in coordinate mode. Each scan and target location was marked as a temporary survey point on an underlying stable boulder or stake.

Leica Cyclone 7.0 was used to georeference scans by attributing the coordinates captured using the total station to each scan station and target location. Scans were then resectioned combining the data for all the scans to one dataset for each bar. Bars were rigorously cleaned, to remove unwanted points. This was carried out to two specifications; 1) bar vegetation retained within the scan, but bars clipped to size and 2) bars cleaned of vegetation for sediment size to be analysed. This reduced the dataset to clouds of 6,644,717 - 16,902,852 points with vegetation (for bar areas of 2163 m$^2$ - 7775 m$^2$) and 3,895,780 - 10,180,366 once cleared of vegetation. Point cloud data were unified to 5 cm resolution, so that each 5 cm$^3$ was represented by a single point. Data were then extracted in text format as raw coordinates, and processed using the PC Tools software library (Rychkov et al., 2012). This tool is designed to process high resolution and large coordinate point clouds to produce statistics for a specified resolution.

DEM$s$ were constructed using the minimum elevation statistic (derived from PC Tools) at a 0.1 m$^2$ resolution (i.e. one elevation point was generated every 0.1 m$^2$). These data were imported into ArcGIS as raw points with an elevation attribute and used to create a DEM with a 0.1 m horizontal resolution. Vegetation distribution was mapped using the standard deviation of laser returns, again at 0.1 m$^2$ resolution. Patches where vegetation was present created a high standard deviation as the laser return was spread across a greater height than sediment yielded. These standard deviations can be used to identify zones of vegetation and exclude them from the DEM so sediment size characteristics can be viewed more clearly. The standard deviation value used to identify areas of vegetation was different for each bar due to differences in sediment size (i.e. higher values used for bars with larger bed material), and was identified visually. These thresholds were standard deviations of 0.26 for Blue Bar, 0.14 for Red Hut Bar, 0.08 for Breakfast Bar and 0.03 for Bain Bar.

Grain size was mapped using a linear correlation between the average standard deviation of point returns within 1 m$^2$ patches and field measured grain-size. Bars were qualitatively delineated into units with different sedimentological characteristics, and a sample was taken for each unit. 27 surface Wolman transect b-axis counts of 100 clasts were collected from across the four bars (Wolman, 1954). Sample spacing was two times the largest clast, to ensure the independence of each grain sampled (Brierley and Fryirs, 2005; Church and Kellerhals, 1978). However, to retain this independence, only 50 grains were able to be measured from four of the smaller units on Red Hut bar. Samples were truncated at 8 mm, as sediment smaller than this is considered to be transported as suspended load and not of primary concern for the processes shaping the morphology of the
system (Brierley and Hickin, 1985; Kondolf et al., 2003a; Parker, 2008). Sampling locations were selected by qualitatively separating each bar into units based on differences in grain size characteristics and morphology and sampling from each unit. Unit outlines and transects were georeferenced using the Trimble RTKGNSSR8 Global Positioning System (Figure 6.3).

The use of photogrammetric techniques, specifically the sedimetrics gravelometer software package was trialled in this study (Graham et al., 2005). However, fractures, air bubbles and discolouration in the sediment resulted in inaccurate grain outlines (referred to as watersheds) being produced, even on painted patches. For this reason, this approach was not used in favour of the traditional manual Wolman transect method described above.

Figure 6.3: Location of transects used to collect Wolman b-axis data. Bars were separated into different sedimentary units and a transect surveyed for each unit. Transect labels are qualitative descriptors. Bars are located on Figure 6.1.
Figure 6.4: Correlation between median grain-size ($D_{50}$) and the mean detrended standard deviation of TLS returns for 1 m$^2$ patches that underlay each Wolman transect. The resulting linear relationship $y = 3.335x - 31.18$ was used to convert detrended standard deviation of TLS returns to express grain size.

Measured grain size was correlated to 1 m resolution detrended standard deviation extracted from the TLS data. The PC toolkit produces a detrended standard deviation, whereby the cell is flattened, removing error due to slope. Arc-GIS was used to overlay transects on the detrended standard deviation raster grid. Points that fell within 1m of each transect were selected, and the mean standard deviation calculated. A linear correlation between the median grain size from the field data ($D_{50}$) and the mean standard deviation for each transect was observed (Figure 6.4). This produced an $R^2$ value of 0.92, deemed to be an acceptable strength trend and similar or stronger to those used in other studies (Carbonneau et al., 2005; Rychkov et al., 2012). The resulting linear equation of $y = 3.335x - 31.18$ was used to convert the 1 m detrended standard deviation grid to express the $D_{50}$ using Raster Calculator in the Spatial Analysis toolbox in Arc-GIS. This provided the basis for the analysis of bar reworking.

Alternative correlations of $D_{50}$ with the standard deviation of laser returns ($\delta$) have previously been determined. In this regard, measurements in five New Zealand waterways studied by Smart et al. (2004) found $D_{50}-\delta = 1.2 - 3.3$. Coleman et al. (2011) found that these and other water-worked gravel beds could be described on average using $D_{50}/\delta = 2.2$. The value of $D_{50}/\delta = 3.3$ derived in this study is consistent with these results identified within the literature. The relation presented in Figure 6.4 was used herein as it is specific to the river reaches studied and the large size and characteristics
(i.e. shape, sphericity) of the present sediment. Variations in the slopes of $D_{50}/\delta$ correlations can be attributed in part to differences in local surface structure, with finer particles present filling in surface hollows and reducing $\delta$ but not affecting Wolman assessment of $D_{50}$, which is typically based on larger particles (e.g. Strom et al., 2010).

A single scan of a 35 m long unvegetated section of vertical terrace was completed to compare sediment size within the terrace with that of the adjacent bar (Blue Bar). PC tools needs a horizontal plane to process data. Therefore, cleaned and unified point clouds were exported and the columns reordered from x (easting coordinate), y (northing coordinate), z (elevation) to z y x, so that the vertical terrace was projected as a gently sloping ramp. PC tools was then able to process the data to detract a detrended standard deviation at 1 m horizontal resolution. The linear equation described above was used to extract grain size statistics across the terrace exposure.

### 6.2.3 Calculating Relative Erodibility

The maps of median grain size were used to model sediment entrainment and bar reworking for flood events of different magnitudes and frequencies. This was achieved by calculating the ratio of dimensionless shear stress, $\tau^*$ (impelling forces) for a flood event to critical shear stress, $(\tau^*)_c$, (resisting forces) for a sediment to be entrained.

$$\frac{\tau^*}{(\tau^*)_c}$$

This measure is based on the assumption that if $\tau^*$ is greater than $(\tau^*)_c$ (the threshold of energy necessary for sediment of a particular size to be entrained) then sediment will be entrained (Chanson, 2004; Church, 2010; Shields, 1936).

Bar reworking was calculated across a range of flood events to measure the geomorphic effectiveness of the event. A Gumbel Type 1 Extreme Value distribution was used to calculate the Recurrence Interval (RI) for different discharge events using 50 years of discharge data from the Tongariro River at the Turangi gauging station, which is located 400 m downstream of Breakfast Bar. This describes the statistical probability of the frequency of an event of that size occurring. Each event can be described using the Annual Exceedence Probability (AEP), which describes the probability of an event of that size occurring within 1 year (Figure 6.4). This study uses flood events with RIs of 2.33 (the arithmetic mean flood commonly referred to as the ‘mean annual flood’), 10 years, 20 years, 50 years and 100 years to compare the differences in bar reworking across events associated with different magnitudes and frequencies (Figure 6.4). Flow depths for these events are illustrated on cross-sections (Figure 6.5 and Figure 6.6). This assumes that the discharge for floods with different RIs does not differ significantly between sites. While some discharge variation is likely,
it is unlikely to be large enough to provide a major source of error because (1) bed shear is calculated using depth, so small variations in discharge are spread across the cross-section, translating to only incremental differences in flow depth; (2) the lack of major tributaries along this section minimises the extent to which discharge changes along a reach.

Tributaries that flow into the Tongariro River within the reach are small. These deliver low flow and sediment loads, reflected in minimal change to discharge along this reach. The largest, Poutu Stream is located between Blue and Red Hut Bars. However, due to the Tongariro Power Development Scheme its flow is regulated, with natural mean flows at the confluence with the Tongariro decreasing from 10 m$^3$s$^{-1}$ to 4 m$^3$s$^{-1}$. The 100 year RI flood discharge for this stream in a natural condition is projected to be in the vicinity of 30 m$^3$s$^{-1}$ (Genesis Energy, 2000). As drainage area has decreased from 145 km$^2$ to 49 km$^2$ discharge would be expected to be much less, providing a minimal impact on flood discharges within the Tongariro River. This makes it possible to characterise flood events using the same discharge across the reach.
Table 6.3: Recurrence Intervals and Annual Exceedence Probabilities for different flood events, based on 50 years of discharge data measured at the Tongariro at Turangi Flow gauging station.

<table>
<thead>
<tr>
<th>Recurrence Interval of the flood (ARI) (years)</th>
<th>Discharge ($m^3$s$^{-1}$)</th>
<th>Annual Exceedence Probability (AEP) (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>2.33</td>
<td>480</td>
<td>42.9%</td>
</tr>
<tr>
<td>10</td>
<td>850</td>
<td>10%</td>
</tr>
<tr>
<td>20</td>
<td>1000</td>
<td>5%</td>
</tr>
<tr>
<td>50</td>
<td>1250</td>
<td>2%</td>
</tr>
<tr>
<td>100</td>
<td>1500</td>
<td>1%</td>
</tr>
</tbody>
</table>

Figure 6.5: Cross-sections with flood inundation levels for the floods modelled within this study for A) Blue Bar and B) Red Hut Bar. Inundation levels are labelled using to flood discharge ($m^3$s$^{-1}$). Location of the bar surface and normal flow levels are also included.
Figure 6.6: Cross-sections with flood inundation levels for the floods modelled within this study for A) Breakfast Bar and B) Bain Bar. Inundation levels are labelled using to flood discharge (m³s⁻¹). Location of the bar surface and normal flow levels are also included.
Bedload transport calculations were carried out using the raster calculator in Arc-GIS. This allowed data from multiple georeferenced raster layers (i.e. median grain size and flow depth rasters) to be used as calculations were carried out at the patch scale, i.e. within each 1 m raster grid cell. The output can be extracted as maps showing the distribution of dimensionless shear stress across the bars, or the raw data extracted for further analysis.

Critical shear stress ($\tau_c$), which describes resistance due to grain size was calculated using Soulsby and Whitehouse’s (1997) equation which uses 18 experimental datasets, including Shield’s (1936) original data which describe the shear stress necessary to entrain clasts of a given size to model sediment entrainment (Coleman and Smart, 2011). This has been found to make better predictions than the Sheild’s curve for large and small sediment sizes, which is important given the large material of the Tongariro River. While Soulsby and Whitehouse (1997) go on to derive equations for the coastal environment, the sediment entrainment curve equation is based on data from riverine flume experiments with unidirectional flow.

$$ (\tau)_c = \frac{0.30}{(1+1.2D_5)} + 0.055 [1 - \exp(-0.020D_5)] \quad (2) $$

$D_5$ is dimensionless grain size, which is calculated from:

$$ D_5 = \left[\frac{g(s-1)}{\nu^2}\right]^{1/3} d \quad (3) $$

$g$ is gravitational acceleration, $s$ is sediment specific gravity, $s = \rho_s/\rho$, $\rho_s$ is the sediment density and $\rho$ is the water fluid density (for quartz sediments, $s = 2.65$) and $\nu$ is fluid viscosity at a specific water temperature (a value of $1.267 \times 10^{-6} \text{ m}^2\text{s}^{-1}$ was used for the lower Tongariro River based on a mean temperature of $10^\circ\text{C}$) and $d$ is grain size, for which the 1 m$^2$ resolution raster grid of median grain size ($D_{50}$) was used.

Equation 2 assumes minimal impact of hiding and protrusion upon entrainment and relatively homogeneous grain size. Due to the local patch scale (1 m$^2$) which sediment entrainment is calculated over herein, these assumptions are reasonable and similarities of grain size within 1 m$^2$ patches are supported by field observations (Laronne et al., 2001) and the TLS derived map of $D_{50}$ within which patterns of sediment sorting are evident (Figure 6.7). This assumption was not made for sediment calculations using cross-sections (i.e. Chapter 5) due to the variability in grain size across this larger scale compared with the patch scale used within this analysis.
The dimensionless shear stress $\tau^*$ which describes impelling forces was calculated using the depth-slope product to derive the bed shear stress, $\tau_b$:

$$\tau_b = g\rho h S$$  \hspace{1cm} (4)$$

Where the product $\rho g$ gives the specific weight of water, $h$ is water depth and $S$ is the water surface slope averaged across a 500 m reach, which was derived from 1 m LiDAR imagery, flown in 2006 during a flow discharge of 30.8 m$^3$s$^{-1}$.

Water surface elevations across each bar, for each flood event, were estimated using a simple, single cross-section steady state model. Each representative cross-section (incorporating wetted channel, bar and floodplain surfaces) was surveyed in the field (Figure 6.5 and Figure 6.6). Due to the steep nature of the Tongariro River, only one cross-section was able to be surveyed for each bar within pool geomorphic units, as riffles are too steep and fast to wade in. Each section was input into the Cross-section Hydraulic Analyser Spreadsheet (Natural Resources Conservation Service, 2011). This uses Excel to apply the Manning’s equation for uniform flow in a cross-section to create a rating table, calculating water surface elevation for different discharge events (See Figure 6.5 and Figure 6.6).

$$Q = \frac{1.486}{n}AR^{2/3}S^{1/2}$$  \hspace{1cm} (5)$$

Where $Q$ is discharge, $n$ is the Manning’s roughness coefficient, $A$ is cross-sectional area perpendicular to flow, $R$ is hydraulic radius and $S$ is reach scale slope measured from the aforementioned LiDAR dataset. A reach-scale Manning’s ‘$n$’ value of 0.03 was selected using the measured relationship between Manning’s ‘$n$’ and discharge for the Tongariro River at the Turangi gauging station as established by Hicks and Mason (1991). They found that ‘$n$’ decreases from 0.06
at a discharge of 28 m$^3$s$^{-1}$ to 0.03 for a flow of 161 m$^3$s$^{-1}$. As all flows used in this study are larger than 161 m$^3$s$^{-1}$, a constant value of $n = 0.03$ has been used herein. This assumes a constant influence of bed roughness on flow during high magnitude events at this site. This only takes into account bed roughness, ignoring floodplain roughness and assumes that flow is uniform. Inevitably this is a simplification of reality. This equation was used to determine the water surface elevation based on the discharge of each flood event (Figure 6.5 and Figure 6.6).

Water surface elevation for each flood event was extrapolated across the bar surface using Arc-GIS to create a sloped triangulated irregular network (TIN), which was converted to a raster grid. Flow depth was depicted by subtracting a 1 m$^2$ Digital Elevation Model (DEM) derived from the TLS data from the water surface elevation raster using raster calculator (in the spatial analyst toolbox). Figure 6.8 shows the distribution of depth across the bar surfaces. This assumes that the water surface
elevation is flat transversely across the bar during floods, which may incorporate some error due to routing of flow across the bar. However, as bar surfaces are completely inundated for all modelled flood events, the impact of underlying topography on flow routing is likely to be less significant compared to that for smaller floods (Bridge and Gabel, 1992).

Shields dimensionless shear stress ($\tau^*$) was calculated using the bed shear stress ($\tau_b$) estimated from equation 4 as

$$\tau^* = \frac{\tau_b}{\rho(s-1)gD_{50}}$$  \hspace{1cm} (6)

Finally, a measure of relative erodibility was derived by dividing bed shear stress ($\tau^*$ of equation 6) by critical shear stress ($\tau_c$ of equation 2). This measure does not attempt to account for the inherent complexities of sediment transport, including turbulence and local spatial variation of flow through eddies. However, the approach is designed as a simple technique to investigate the relative susceptibility of difference geomorphic units to sediment entrainment, and to characterise and compare patterns of bar reworking.

6.3 RESULTS

This section analyses i) bar evolution using aerial photography, ii) describes the topography and grain size distribution for each bar and iii) analyses patterns of bar reworking.

6.3.1 Bar Evolution and Historical Adjustment

Analysis of aerial photographs allows bar evolution to be assessed, offering insights into bar age, sensitivity and the mechanisms of bar development and destruction. Long-term trends in sediment flux can be reflected in changes to bar extent, providing a link between local and reach scale dynamics.

6.3.1.1 Blue Bar Evolution

A bar has been present on and off at this location since 1958. It became part of the floodplain between 1973-1993 following a decrease in large floods over this period, before being stripped of vegetation and reformed as a bar following the 2004 flood, giving it a recently reworked age of 8 years, but an initial formation age $> 50$ years (Figure 6.9). The bar was most likely created or reworked by the Q^60 flood in 1958, directly before the aerial photograph was taken (Figure 6.9). The Q^20 flood in 1964 stripped the bar of vegetation. Bar size and vegetation cover at a given time reflect the period elapsed since the last large magnitude flood event (e.g. $> Q_{60}$). This was seen in the revegetation and incorporation of the bar into the floodplain between 1973 and 1993, which coincided with a period of no flood events greater than the Q^10 (Figure 2.21). The Q^60 flood in 2004
stripped the bar of vegetation and reworked the bar surface, resulting in the reappearance of an unvegetated bar in the 2007 survey. In summary, the Blue Bar appears as a small, homogeneous bar which has undergone little morphological adjustment and requires flood events $> Q^{20}$ to strip and mobilise the surface.

6.3.1.2 Red Hut Bar Evolution
The Red Hut Bar has had a dynamic history, with evidence of significant adjustment to barforms during the past 80 years (Figure 6.9). The head of the contemporary lateral bar was present in the survey map from 1928. However, by 1958 a back-channel had formed and the Red Hut bar became a mid-channel bar. By 1973, a lobe of sediment had been deposited in the lee of the upstream Poutu Island. This became partly vegetated, until the $Q^{60}$ flood in 2004 stripped vegetation and mobilised this surface, creating a mid-channel bar as seen in 2007. By 2009, aerial photographs show that the bar had formed its contemporary shape as a lateral bar, giving it a present age of between 3 - 4 years. This history describes a zone of frequent reworking, where the large 2004 $Q^{60}$ flood played a key role in mobilising and redistributing sediment deposits within the reach.

6.3.1.3 Breakfast Bar Evolution
The Breakfast Bar has been relatively stable since 1958, undergoing minimal adjustment in the past 80 years (Figure 6.9). This bar exhibited similar patterns of adjustment to the Blue Bar. Vegetation cover increased and the bar was integrated into the floodplain between 1973 and 1993 as few flood events occurred. The 2004 $Q_{60}$ flood again stripped vegetation. Thus, the Breakfast Bar has an age of $> 50$ years and has undergone minimal adjustment. This bar again illustrates the necessity of large magnitude flood events ($> Q^{20}$) to remove vegetation, and rework and deliver sediment.

6.3.1.4 Bain Bar Evolution
The Bain Bar is located downstream of the terraces, within a braided reach that drains an actively reworked alluvial fan. Channel widening occurred between 1958 and 1964. This actively adjusting braidplain reworked the surface now accommodated by the contemporary bar (Figure 6.9). Gravel extraction in 1973 caused artificial narrowing, with the channel creating a low sinuosity, single channel planform with fewer active bar surfaces. Between 1973 and 2007 this reach widened again, increasing the area of actively reworked bar surfaces. A bar surface was first observed at Bain Bar’s present location in 1993, giving the bar a present age of $> 18$ years. Frequent reworking and gravel extraction have shaped this active reach of river, making Bain Bar a relatively young and dynamic feature. The mechanisms of bar development are difficult to ascertain within this reach due to human disturbance, particularly gravel extraction.
Figure 6.9: Bar development extracted from aerial photography (1941 – 2007) and a survey map (1928). See Table 4.1 for details for each aerial photograph.

### 6.3.2 Topography and Grain-size Distribution across Bar Surfaces

This section describes patterns of sediment distribution (Table 6.4; Figure 6.10) and topography (Figure 6.11) across the four bars.
Table 6.4: Grain size and area characteristics for each within bar unit for each bar. Standard deviation is used to describe sediment sorting.

<table>
<thead>
<tr>
<th>Bar</th>
<th>Unit</th>
<th>Area (m$^2$)</th>
<th>Unit area as % of total bar area</th>
<th>Average D$_{50}$</th>
<th>Grain-size classification</th>
<th>St-dev of D$_{50}$</th>
</tr>
</thead>
<tbody>
<tr>
<td>Blue</td>
<td>Head</td>
<td>792</td>
<td>41</td>
<td>214</td>
<td>Coarse cobble</td>
<td>92</td>
</tr>
<tr>
<td></td>
<td>Supra-platform</td>
<td>291</td>
<td>15</td>
<td>171</td>
<td>Coarse cobble</td>
<td>67</td>
</tr>
<tr>
<td></td>
<td>Back channel</td>
<td>504</td>
<td>26</td>
<td>132</td>
<td>Coarse cobble</td>
<td>64</td>
</tr>
<tr>
<td></td>
<td>Tail</td>
<td>365</td>
<td>19</td>
<td>74</td>
<td>Coarse cobble</td>
<td>67</td>
</tr>
<tr>
<td>Red Hut</td>
<td>Head</td>
<td>978</td>
<td>15</td>
<td>173</td>
<td>Coarse cobble</td>
<td>82</td>
</tr>
<tr>
<td></td>
<td>US supra-platform*</td>
<td>1928</td>
<td>29</td>
<td>128</td>
<td>Cobble</td>
<td>60</td>
</tr>
<tr>
<td></td>
<td>DS supra-platform*</td>
<td>1804</td>
<td>27</td>
<td>110</td>
<td>Cobble</td>
<td>48</td>
</tr>
<tr>
<td></td>
<td>Back channel</td>
<td>1058</td>
<td>16</td>
<td>78</td>
<td>Cobble</td>
<td>56</td>
</tr>
<tr>
<td></td>
<td>Tail</td>
<td>809</td>
<td>12</td>
<td>36</td>
<td>Very coarse gravel</td>
<td>47</td>
</tr>
<tr>
<td>Breakfast</td>
<td>Head</td>
<td>1577</td>
<td>40</td>
<td>143</td>
<td>Coarse cobble</td>
<td>62</td>
</tr>
<tr>
<td></td>
<td>Supra-platform</td>
<td>1318</td>
<td>34</td>
<td>101</td>
<td>Cobble</td>
<td>38</td>
</tr>
<tr>
<td></td>
<td>Fine-grained deposition</td>
<td>241</td>
<td>6</td>
<td>63</td>
<td>Very coarse gravel</td>
<td>33</td>
</tr>
<tr>
<td></td>
<td>Tail</td>
<td>760</td>
<td>20</td>
<td>22</td>
<td>Coarse gravel</td>
<td>34</td>
</tr>
<tr>
<td>Bain</td>
<td>Head</td>
<td>1251</td>
<td>33</td>
<td>67</td>
<td>Cobble</td>
<td>33</td>
</tr>
<tr>
<td></td>
<td>Supra-platform</td>
<td>810</td>
<td>21</td>
<td>42</td>
<td>Very coarse gravel</td>
<td>19</td>
</tr>
<tr>
<td></td>
<td>Back channel</td>
<td>883</td>
<td>23</td>
<td>21</td>
<td>Coarse gravel</td>
<td>22</td>
</tr>
<tr>
<td></td>
<td>Fine-grained deposition</td>
<td>168</td>
<td>4</td>
<td>15</td>
<td>Medium Gravel</td>
<td>18</td>
</tr>
<tr>
<td></td>
<td>Tail</td>
<td>684</td>
<td>18</td>
<td>46</td>
<td>Very coarse gravel</td>
<td>42</td>
</tr>
</tbody>
</table>

* The Breakfast Bar and the Bain Bar both exhibited localised zones where fine grained sediment had superimposed the gravel by filling interstices (referred to as ‘fine grained material’ on Figure 6.10). As these do not reflect reworking and were local they were excluded from analysis, resulting in total bar areas of < 100 % for these bars.

* The supra-platform of Red Hut bar was large, exhibiting marked downstream fining and was separated into two different units, the US (upstream) and DS (downstream) supra platforms.

6.3.2.1 Blue Bar Topography and Grain-size distribution

The bar head at Blue Bar includes most of the upstream portion of the bar, sloping up to a topographic high along the length of the bar (Figure 6.11). Sediment distribution follows this grade with an increased frequency of boulders adjacent to the channel, grading to coarse cobble on the ridge and in the downstream locations (Figure 6.10). Sparse vegetation has established on the ridge, indicating less frequent reworking. A back channel has formed between this ridge and the more densely vegetated back of the bar characterised by D$_{50}$ ranging from very coarse gravel to coarse cobble. Sand fills interstices. The supra-platform is downstream of the bar head and exhibits a decrease in sediment size, with coarse cobble adjacent to the channel. The bar tail is small and
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characterised by cobble and very coarse gravel. A sand sheet has been deposited at the back of the tail of the bar, which can be attributable to anthropogenic clearance of vegetation for a walkway, and as such has been excluded from analysis. Blue Bar displays classic patterns of within bar downstream fining (Powell, 1998; Rice and Church, 2010). The topography of this bar is simple with only few units present and quite low variability in grain size seen across the units.

6.3.2.2 Red Hut Bar Topography and Grain-size distribution

Red Hut Bar has more complex morphology. The bar head, made up of coarse cobble and boulders, is located at the apex of the bar, rather than the upstream location, reflecting the more sinuous nature of the channel at this point (Table 6.4). Aerial photographs indicate that in 2007 this unit was a steep riffle, explaining the boulder deposit (Figure 6.9). This sediment distribution is a divergence from the common conception of the coarse nucleus being located at the bar head (Bluck, 1976; Powell, 1998). Directly downstream of the bar head and adjacent to the channel is a strip of vegetation, which stabilises this face, creating a steep bar edge (Figure 6.11). The presence of vegetation indicates that during most flows, water would be routed around this ridge and onto the supra-platform surface. The supra-platform is separated into upstream (US) and downstream (DS) surfaces due to marked downstream fining, from predominantly coarse cobbles to predominantly cobbles (Figure 6.10). Behind this unit, a back channel has formed, with a 1 m wide riparian margin. This is a remnant of past channel configuration as a secondary channel existed at this location prior to 2007 (Figure 6.9). The lower elevation of this surface allows greater inundation and scour during flood events, and the downstream section is below the water table, maintaining stream flow at low flow stage. The tail of the bar exhibits rapid downstream fining, from cobble to sand. This unit has a low elevation and is thus protected by higher bar surfaces upstream (Rice and Church, 2010). The morphology and sediment size distribution for Red Hut Bar is influenced by valley shape, and the channel flows against the valley margin, creating a curved bar.
Figure 6.10: Distribution of median grain-size ($D_{50}$) derived from TLS data and location of within bar geomorphic units. Grain sizes are displayed using the Wentworth grain-size scale (Kondolf et al., 2003a). Bar position is from most upstream on the left to most downstream bar on the right.
Figure 6.11: 0.1 m$^2$ DEM of bar morphology from TLS data for the four bars, A) Blue Bar, B) Red Hut Bar, C) Breakfast Bar and D) Bain Bar. Vegetation distribution was plotted using standard deviation of TLS returns at 0.1 m$^2$ resolution.
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6.3.2.3 Breakfast Bar Topography and Grain-size distribution

Breakfast Bar has a fairly homogenous morphology. The bar head may be characterised as a lower elevation surface around the edge of the bar, which comprised of predominantly coarse cobble, with some cobble and boulders (Figure 6.10 and Figure 6.11). This unit grades to the back of the bar, more steeply at the head than at the tail forming the supra-platform and sediment too accordingly grades to smaller material, becoming mostly cobble. The tail of the bar is made up of sand, with a cobble ridge adjacent to the channel which has homogeneous, low elevation morphology.

6.3.2.4 Bain Bar Topography and Grain-size distribution

The topography of Bain Bar is complex, with vegetation complicating patterns of sediment transport. The bar head is a topographic low located on the upstream section of the bar. This is made up of very coarse gravel, fining to medium gravel towards the back of the bar (Figure 6.10). A back channel unit runs along the back of the bar, where sediment is coarse gravel at the upstream section and sand downstream of this. A localised deposit of finer material (fine gravel and sand) is located at the most upstream section of the bar. Sediment size decreases downstream of the bar head to the supra-platform composed of very coarse gravel. The bar tail exhibits noticeable variation, characterised by a range of gravel size-classes. This feature is relatively flat and of low elevation, with vegetation covering around 50% of the surface leading to a complex array of units, which are the product of alternating reworking and vegetation stabilisation (Figure 6.11). Vegetation is dominated by an exotic, invasive Willow species, which grows rapidly and is resistant to frequent flood events. As this vegetation has not colonised other reaches, the increase in vegetation density is not an indication of stability. Rather, this bar is frequently reworked which is reflected in the complex patterns of morphology and grain size.

6.3.3 Grain-size Distribution

Grain size statistics are derived using two methods. The first extracts the D_{50} for each 1 m^2 of the TLS data. As this is based on the D_{50}, it does not document the whole range of sediment sizes. This was done to compare grain-size distributions for the whole bars and for the units within each bar. The second method uses the full grain size distribution gained from Wolman transects (see Figure 6.3). Each bar was delineated into zones of similar grain-size distributions and morphology and a transect completed for each zone. This was completed at a finer resolution than the bar unit classification and as such, bar units may have multiple transects. This was necessary to gain an appropriate number of samples with which to correlate TLS data and measured grain size. Grain-size distributions for each transect within a bar are combined and weighted based on unit area. This is done by creating histograms for each unit and multiplying the proportion of the sample in each bin with the proportion of unit area to bar area. The resulting proportions are summed for each bin, and
a histogram constructed which describes the frequency of sediment size across the whole surface. This technique is based on full grain-size distributions, rather than the distribution of $D_{50}$ across the bar, thereby capturing a greater range of clast sizes.

### 6.3.4 Grain size Derived from TLS data

The four bars show a decrease downstream in median sediment size, consistent with downstream fining (Powell, 1998; Rice and Church, 2010). Coarse cobble makes up the majority of the surface area on Blue Bar, comprising 52%, with 23% cobble and 13% boulder (Table 6.5). Sand is minimal, with a cover of 1.2%. Median $D_{50}$ decreases significantly from 156 mm at Blue Bar to 105 mm at Red Hut Bar within a downstream distance of 2.9 km. 40% of Red Hut Bar comprises cobble, and 32% coarse cobble, and 3.8% boulder, exhibiting a significant decrease in the coarse cobble and boulder fractions compared with Blue Bar. In contrast, sediment distribution at Red Hut and Breakfast Bar are remarkably similar, despite a downstream distance of 4.7 km. Breakfast Bar exhibited a 4% increase in cobble (comprising 44%) and a 5% decrease in coarse cobble (down to 27%) (Table 6.5). The proportion of sand doubled from 4% at Red Hut bar to 8% at Breakfast bar. Breakfast Bar exhibited a decrease in larger clasts, with the max $D_{50}$ decreasing from 682 mm at Red Hut bar down to 490 mm at Breakfast Bar (Table 6.6; Figure 6.12). The standard deviation decreases from 70 to 65, an indication that these bars are similar (Table 6.6). Bain Bar is located 2.8 km downstream of Breakfast Bar and the median ($D_{50}$) has decreased markedly from 98 mm to 41 mm. Sediment size is homogeneous and has a standard deviation of 35, compared to standard deviations between 65 – 92 mm for the other sites (Table 6.6). This reflects a change in confinement, as the channel leaves the terraces, adopting a more mobile braided planform. 40% of the bar is made up of very coarse cobble, with 21% and 20% of coarse gravel and cobble respectively. Proportion of sand is similar to Breakfast Bar, at 7% (Table 6.5).
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Table 6.5: Proportion of $D_{50}$ within each sediment category as derived from TLS data. Greyscale bars underlying the data indicate the relative proportion of grain sizes for each bar.

<table>
<thead>
<tr>
<th>Wentworth Sediment size classes</th>
<th>B-axis measurement (mm)</th>
<th>Blue</th>
<th>Red Hut</th>
<th>Breakfast</th>
<th>Bain</th>
</tr>
</thead>
<tbody>
<tr>
<td>Sand</td>
<td>0 - 4</td>
<td>1.22</td>
<td>4.16</td>
<td>7.98</td>
<td>7.27</td>
</tr>
<tr>
<td>Fine gravel</td>
<td>5 - 8</td>
<td>0.73</td>
<td>0.81</td>
<td>1.54</td>
<td>2.88</td>
</tr>
<tr>
<td>Medium gravel</td>
<td>9 - 16</td>
<td>1.03</td>
<td>1.61</td>
<td>2.83</td>
<td>6.38</td>
</tr>
<tr>
<td>Coarse gravel</td>
<td>17 - 32</td>
<td>2.31</td>
<td>4.04</td>
<td>4.03</td>
<td>21.3</td>
</tr>
<tr>
<td>Very coarse gravel</td>
<td>33 - 64</td>
<td>6.08</td>
<td>13.2</td>
<td>10.11</td>
<td>38.97</td>
</tr>
<tr>
<td>Cobble</td>
<td>65 - 128</td>
<td>23.05</td>
<td>39.99</td>
<td>44.29</td>
<td>20.01</td>
</tr>
<tr>
<td>Coarse cobble</td>
<td>129 - 256</td>
<td>52.13</td>
<td>32.34</td>
<td>26.93</td>
<td>3.03</td>
</tr>
<tr>
<td>Boulder</td>
<td>257 - 512</td>
<td>13.02</td>
<td>3.81</td>
<td>2.3</td>
<td>0.16</td>
</tr>
<tr>
<td>Coarse boulder</td>
<td>513 - 1024</td>
<td>0.43</td>
<td>0.04</td>
<td>0</td>
<td>0</td>
</tr>
</tbody>
</table>

Table 6.6: Grain-distribution statistics derived from TLS data. All sizes are presented in mm.

<table>
<thead>
<tr>
<th>Bar Name</th>
<th>Distance Downstream (km)</th>
<th>Median ($D_{50}$)</th>
<th>Mean</th>
<th>St Dev</th>
<th>Min</th>
<th>Lower quartile</th>
<th>Upper quartile</th>
<th>Max</th>
</tr>
</thead>
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<tr>
<td>Blue</td>
<td>57.6</td>
<td>156</td>
<td>166</td>
<td>92</td>
<td>1</td>
<td>109</td>
<td>215</td>
<td>780</td>
</tr>
<tr>
<td>Red Hut</td>
<td>60.5</td>
<td>105</td>
<td>113</td>
<td>70</td>
<td>1</td>
<td>66</td>
<td>150</td>
<td>682</td>
</tr>
<tr>
<td>Breakfast</td>
<td>65.2</td>
<td>98</td>
<td>101</td>
<td>65</td>
<td>1</td>
<td>61</td>
<td>136</td>
<td>490</td>
</tr>
<tr>
<td>Bain</td>
<td>68.0</td>
<td>41</td>
<td>46</td>
<td>35</td>
<td>1</td>
<td>23</td>
<td>62</td>
<td>295</td>
</tr>
</tbody>
</table>
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6.3.5 Grain Characteristics of each Within-bar Unit

Grain size data were extracted for each geomorphic unit to compare how distributions change across these surfaces. Essentially this reflects the distribution of shear stress during flood events (Bridge, 2003).

6.3.5.1 Blue Bar within unit grain size distribution

Grain-size distributions for the head and the supra-platform are very similar, except the head has a greater proportion of boulders and coarse boulders (24 % compared to 12 %). Both comprise between 63 % and 64 % of coarse cobble. The supra-platform has more cobble with 22 % compared with 9 % for the head. This shows that while downstream fining is evident, the downstream section is still made up of large sediment. The back channel shows a decrease in boulders (to 3 %) and coarse cobbles (to 45 %) and an increase in cobbles (to 41 %), representing a finer environment than the head and supra-platform. The tail is much finer with 51 % being very coarse gravel or finer.
Figure 6.13: A-D) Histograms showing sediment size distribution for each of the units within Blue Bar. E) shows the distribution of each unit, with the unit colour indicating the average $D_{50}$ value.
6.3.5.2 Red Hut Bar within unit grain size distribution

Figure 6.14: A-E) Histograms showing grain size distribution for each of the units within Red Hut Bar. F) shows the distribution of each unit, with the unit colour indicating the average $D_{50}$ value. N/A indicates the area where gravels were not visible due to either vegetation or surface water.
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The head of Red Hut Bar was the coarsest locale. 56 % was coarse cobble and 15 % boulder and 21 % cobble. The US supra-platform displayed a marked decrease in coarse cobble to 41 % and an increase in cobble to 45 %. This fining was continued to the DS supra-platform where coarse cobble and cobble were 27 % and 58 % respectively. This unit displayed an increase in very coarse gravel to 11 %. The back channel exhibited fining, with a normal distribution present centred around 36 % for both very coarse gravel and cobble. The tail comprises 31 % sand with a normal distribution apparent for the rest of the clasts, centred around 22 % of very coarse gravels, representing a low energy environment where sediment is deposit during the falling limb of floods.

6.3.5.3 Breakfast Bar within unit grain size distribution

The bar head Breakfast Bar is composed of cobble (40 %) and coarse cobble (48 %) representing an area of large material. The supra-platform at the back of the bar exhibits the marked decrease in coarse cobble (to 19 %) and an increase in cobble (to 68 %). The grain size distribution of the tail is far flatter, with 11 - 14 % of clasts sized medium gravel to cobble, and 41 % sand, representing an area were fine grained material is deposited on the falling limb of floods. The deposition of finer material at the head of the bar has larger sediment with 77 % between very coarse gravel and cobble. Whilst this material may be smaller than the adjacent bar head, it is significantly larger than the majority of the bar tail.
Figure 6.15: A-D) Histograms showing the grain size distribution for each of the units within Breakfast Bar. E) shows the distribution of each unit, with the unit colour indicating the average D$_{50}$ value.
6.3.5.4  Bain Bar within unit sediment distribution

Figure 6.16: A-E) Histograms showing grain size distribution for each of the units within Bain Bar. F) shows the distribution of each unit, with the unit colour indicating the average D$_{50}$ value.

The bar head at Bain Bar is mostly composed of very coarse gravel and cobble, with few fine grained sediments present (< 64 mm). At the supra-platform, the proportion of cobble has decreased from 38 % to 12 %, and the proportion of coarse and very coarse gravel has increased, representing a fining of sediment. The back channel displays substantive fining, with only 21 % > 64 mm and peaks in sediment of 34 % coarse gravel (16 - 32 mm) and 22 % sand (< 4 mm). Whilst the bar tail is finer than the head and the supra-platform, it consists of more coarse grains than seen on the back channel and fines. 59 % of grains are coarse gravel – very coarse gravel and only 9 % makes up the smaller fraction of sand and fine gravel which is commonly found on this unit. 83 % of the deposition
of finer grained material at the upstream extent of the bar was < 32 mm, representing coarse gravel and smaller.

6.3.6 Grain-size from Area-weighted Wolman Counts

The data derived from the Wolman counts showed similar trends, but captured a greater range of sediment compared to that derived from the TLS data. This is presented using a linear mm scale rather than the logarithmic phi scale as used for the TLS derived data, to allow differences in the distribution of the larger fraction to be clearly displayed. This analysis aims to identify the different fractions of sediment moving through the system.

The sediment distribution for Blue Bar was spread, with multiple peaks, indicating a tri or quad modal distribution (Figure 6.17). The smallest peak is at 40 - 60 mm indicating coarse gravel. The next peak occurs in the bin between 160 - 180 mm, constituting the coarse cobble fraction. The largest peak is at 240 - 260 mm, on the cusp of classification of boulders which is skewed to the right with another slight peak at 340 - 360 mm.

Grain size distribution for Red Hut bar is strongly skewed to the left (Figure 6.17). The highest peak is between 20 – 40 mm, capturing the finer medium to coarse gravel fraction. This peak flattens to include sediment between 40 - 80 mm (very coarse gravel to cobble) with an additional peak at 80 – 100 mm (cobble). Grain size distribution exhibits a long tail to the right, with minor peaks present at 140 - 160 mm, 200– 220 mm and 280– 300 mm, following a similar pattern to the peaks observed at Blue Bar.

The grain-size distribution at Breakfast Bar is flat and wide, indicating equal proportions of grains of different sizes. The frequency of grains between 40 mm and 200 mm is between 7.9 % and 9.2 %, with a peak for grains between 80 - 100 mm (cobble) of 11.8 %. This peak is consistent with a peak in Red Hut Bar. The peak of small sediment (between 20 – 40 mm) observed for the other bars is minimal for Breakfast Bar, indicating a reduction in readily reworked material. The tail of the distribution stretches from 220 - 400 mm, slowly decreasing in proportion present, indicating that large boulder material, while still present is not as common.

Bain Bar exhibits a uni-modal distribution with 26.6 % of sediment being between 20 - 40 mm (Figure 6.17). 65 % of the sediment distribution is less than 60 mm defined as gravel, a substantial reduction in size compared with the upstream bars. The frequency of sediment decreases rapidly to 220 mm, indicating minimal presence of material larger than coarse cobbles.
Figure 6.17: Histograms displaying the weighted grain-size distribution for each bar (in mm).
6.3.7 Grain-size Delivered from the Terraces

A single terrace adjacent to Blue Bar was scanned to compare the grain size of sediment delivered from terrace incision with that of the adjacent bars (i.e. Blue Bar) which incorporates the active fraction. The terraces are made up of lahar deposits, delivered from the volcanic cones of Mt Ruapehu and Mt Ngauruhoe. As materials are derived from hyper-concentrated flows, they comprise a greater range of sediment, including larger material than transported through fluvial processes (Cronin et al., 1997; Fagents and Baloga, 2005). This presents an opportunity to compare the size of material delivered from this fraction through terrace incision, with that being transported through the system.

![Figure 6.18: Terrace section scanned with TLS. A) Photograph of a representative section of terrace. B) The clipped point cloud coloured by a photograph taken by the scanner. C) Distribution of sediment size across the terrace, based on converting the detrended standard deviation to represent D_{50}.](image)
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Figure 6.18 shows the terrace scan and the distribution of $D_{50}$ across this surface. Sediment delivered from the lower portion of terrace is mostly boulder sized (between 257 - 512 mm), with localised coarse boulder (513 - 1,024 mm) and cobble (129 - 265 mm) sections. In comparison, the upper portion is smaller, comprising predominantly coarse cobble. This indicates that at this location, more recent incision is likely to deliver larger material than the sediment delivered during earlier stages of incision.

![Figure 6.19: Comparison of grain-size distribution between A) the surface of Blue Bar and B) the terrace exposure directly across the channel from Blue Bar.](image)

Grain size was compared for the terraces and Blue Bar. Grains derived from the terrace are significantly larger ($p$ value = 0) than those that make up the surface of Blue Bar. Figure 6.19 shows a marked increase in both boulder (48 % compared with 13 %) and coarse boulder (11 % compared to <1 %) respectively. Figure 6.20 supports this, showing similar shaped distributions which are strongly skewed to the left, with a tail of larger sediment to the right. However, the peak for the bar surface is around 150 mm compared to the larger peak of 300 mm for the terrace. In addition, the tail of the
distribution for the terrace comprises very large clasts, which are rare or absent from the bar surface. Figure 6.21 shows even larger material coming from the terrace deposits 500 m upstream. This illustrates the very coarse and variable nature of sediment delivered to the channel from this source.

![Image](image_url)

**Figure 6.21**: Size of sediment excavated from terraces 500 m upstream of Blue Bar. This illustrates localised areas of much larger sediment. Scale model is 1.9 m tall. Largest clasts have a b-axis > 1 m.

### 6.3.8 Assessment of Bar Reworking

The ratio of dimensionless shear stress to critical shear stress was calculated across each bar to model bar reworking across a range of flood events. Bar reworking results shown in Figure 6.22 and Figure 6.23, detail the distribution of patches likely to undergo sediment entrainment across the range of modelled flood events. Quantification of entrainment is provided in Table 6.7, which presents the percentage of each geomorphic unit entrained during each modelled flood event. Within the following text, numbers within brackets following a geomorphic unit indicate the percentage of the unit that is entrained for the given flood event.

#### 6.3.8.1 Spatial distribution of Relative Erodibility

71% of Blue Bar remains static during the Q^{2.33} event. However, 80 % of the bar tail is entrained during this flood, including the coarse gravel to sand fractions that make up the tail, and the coarse cobble on the channel margins (Figure 6.22; Table 6.7). During the Q^{10} flood, cobbles to coarse cobbles in the back channel (77 %) and the supra-platform (78 %) are entrained. Larger floods are needed to mobilise the boulder-sized grains on the bar head, with entrainment increasing by 12 % between the Q^{10} and Q^{30}, and Q^{10} and Q^{50} floods, up to 82 % for the Q^{50} flood.
During the Q^{2.33} flood, gravel to sand sized sediment located on the bar tail (94 % entrained) and upstream portion of the back channel (38 %) of Red Hut Bar is entrained (Figure 6.22; Table 6.7). The Q^{10} flood causes a marked increase in the reworking of the total bar, which increases from 34 % during the Q^{2.33} to 75 %. This is primarily due to a marked increase in reworking of the supra-platform (68 % on the upstream supra-platform and 80 % on the downstream supra-platform), which makes up 56 % of the bar surface. The bar head is a high elevation ridge of large sediment (coarse cobble to boulder) running along the middle to top of the bar. Increases in flood magnitude increase the proportion of this surface that is entrained, with 63 % entrained during the Q^{20}, 76 % during the Q^{50} and 83 % during the Q^{100} floods, illustrating the role of large events in reworking this unit. However, as the bar head is small, making up only 15 % of the total bar area, the size of flood necessary to rework this unit is not likely to be a limiting factor for planform adjustment.

During the Q^{2.33} flood, fine gravel and sand is entrained from the tail (95 % entrainment) of the Breakfast Bar (Figure 6.23; Table 6.7). The Q^{10} flood mobilises cobbles from the supra-platform, as 60 % of the unit is entrained. The head is characterised by moderate elevation and relatively coarse material (predominately coarse cobbles). The head runs along the upper-middle section of the bar adjacent to the channel, undergoing minimal reworking. Much of this surface remains static up to the Q^{100} flood, undergoing only 60 % entrainment during the Q^{100} compared with between 78 % and 90 % for bar heads on the other bars. However, as flood magnitude increases from Q^{10} to the Q^{100} flood, a greater number of patches within this unit are selectively entrained.

During the Q^{2.33} flood on the Bain Bar, smaller grains on the back channel (73 %) and bar tail (62 %) are entrained, leaving clasts larger than very coarse gravel static (Figure 6.23; Table 6.7). The Q^{10} flood increases entrainment in the supra-platform (from 44 % during the Q^{2.33} to 82 %), while the areas comprising cobble are stationary. As flows increase to the Q^{100} flood, patches on the bar head are entrained, with 57 % entrainment during the Q^{50} up to 71 % for the Q^{50}. The bar tail undergoes less entrainment during large flood events than seen at the other reaches, with 88 % entrained during the Q^{50} compared with > 99 % for the other bars (Table 6.7). This is a combination of comparatively large sediment within this unit (very coarse gravel) given its reach position and the low slope of this reach (0.003) (Table 6.2). Localised patches of coarse cobble remain static with only 78 % entrainment predicted during the Q^{100} (the second lowest head entrainment for the Q^{100} flood after Breakfast Bar).
Figure 6.22: Relative erodibility calculated as the ratio of dimensionless shear stress to critical shear stress at a 1 m² resolution for the two most upstream bars, Blue Bar and Red Hut Bar. Black indicates any value less than 1, where sediment remains immobile.
Figure 6.23: Relative erodibility calculated as the ratio of dimensionless shear stress to critical shear stress at a 1 m² resolution for the two most downstream bars, Breakfast Bar and Bain Bar. Black indicates any value less than 1, where sediment remains immobile.
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Table 6.7: Grain size and area characteristics of within-bar geomorphic units and the percentage of sediment entrainment on each unit and bar during different magnitude/frequency flood events. The location of geomorphic units on each bar is shown on Figure 6.4.

<table>
<thead>
<tr>
<th>Bar</th>
<th>Unit</th>
<th>Average $D_{50}$ (mm)</th>
<th>Percentage of each unit entrained during each flood event</th>
<th></th>
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</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td></td>
<td>$Q^{2.33}$ ($480 \text{ m}^3\text{s}^{-1}$)</td>
<td>$Q^{10}$ ($850 \text{ m}^3\text{s}^{-1}$)</td>
</tr>
<tr>
<td>Blue</td>
<td>Head</td>
<td>214</td>
<td>13</td>
<td>58</td>
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<td></td>
<td>Supra-platform</td>
<td>171</td>
<td>24</td>
<td>78</td>
</tr>
<tr>
<td></td>
<td>Back channel</td>
<td>132</td>
<td>23</td>
<td>77</td>
</tr>
<tr>
<td></td>
<td>Tail</td>
<td>74</td>
<td>80</td>
<td>96</td>
</tr>
<tr>
<td></td>
<td>Total Bar Entrainment</td>
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<td>81</td>
</tr>
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<td>Tail</td>
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<td></td>
<td>Total Bar Entrainment</td>
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<td>Head</td>
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<td></td>
<td>Supra-platform</td>
<td>101</td>
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<td></td>
<td>Tail</td>
<td>22</td>
<td>95</td>
<td>98</td>
</tr>
<tr>
<td></td>
<td>Total Bar Entrainment</td>
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<td>60</td>
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<td>Head</td>
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<td>13</td>
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<td></td>
<td>Supra-platform</td>
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<td>82</td>
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<td></td>
<td>Tail</td>
<td>46</td>
<td>62</td>
<td>80</td>
</tr>
<tr>
<td></td>
<td>Total Bar Entrainment</td>
<td>44</td>
<td>72</td>
<td>79</td>
</tr>
</tbody>
</table>

* The supra-platform of Red Hut bar was large, exhibiting marked downstream fining and was separated into two different units, the US (upstream) and DS (downstream) supra platforms.
Figure 6.24: The proportion on each unit at each bar reworked for the A) $Q^{2.33}$, B) $Q^{20}$ and C) $Q^{100}$ floods.

Figure 6.24 displays the proportion of each unit reworked between the bars across three floods. This highlights similarities in the reworking of bar heads during the $Q^{2.33}$ flood (13 – 19 %). The supra-platform also displayed similar reworking, though entrainment was noticeably higher at Bain Bar with 10 % higher reworking. This difference was even more pronounced for the back channel at Bain Bar, with > 35 % more of its surface entrained compared with the other bars. The high percentage of gravel and sand that makes these units ensures that they are easily reworked during frequent floods. In contrast, high reworking is predicted for the bar tail with > 80 % entrainment. Bain Bar presents an exception, with only 62 % reworking predicted.
The patterns of unit reworking are very similar for the $Q^{20}$ and $Q^{1500}$ floods. A greater proportion of the head of Blue Bar and Red Hut Bar (58 and 51 % respectively) can be reworked. Breakfast Bar exhibited much lower reworking for the head and the supra-platform compared to the other bars. This indicates that large flows are not sufficient to rework the coarse material of these units. Entrainment of bar tails was > 97 % for all bars excluding Bain Bar which was lower at 83% and 89% for the $Q^{20}$ and $Q^{100}$ floods respectively.

### 6.3.9 Percent of Bar Reworked for Different Magnitude Events

Figure 6.25 displays the percentage of each bar entrained across the range of flood events. This allows downstream changes in reworking to be assessed.

The two upstream bars, Blue Bar and Red Hut Bar, are located in reaches characterised by steeper slopes and larger sediment (Table 6.2) than the downstream bars and are seen to exhibit similar trends in reworking (Figure 6.25). During the $Q^{2.33}$ flood, less of the bar surface was mobilised with 29 % at Blue Bar and 34 % at Red Hut Bar compared with 41 % and 44 % entrainment at Breakfast Bar and Bain Bar respectively (Table 6.7). These latter bars store a greater proportion of active sand and gravel, which is more easily flushed during frequent floods. As such, Bain Bar in the braided reach is predicted to undergo the greatest reworking during frequent, low magnitude floods.

![Figure 6.25: The percent of each bar reworked for different magnitude/frequency flood events. Recurrence intervals are annotated in years.](image)

With increasing discharge from $Q^{2.33}$ to $Q^{10}$, sediment entrainment increases rapidly in the upstream bars, going from 29 % to 73 % for Blue Bar and 34 % to 75 % for Red Hut Bar for the $Q^{2.33}$ to the $Q^{10}$ floods respectively. For the same flow increase, the percentage of bed materials entrained from Bain
Bar increased to 72%, placing it slightly below the upstream sites. The Breakfast Bar displays less change, with percentage entrained increasing to 60%, well below the values for the other bars.

For floods > $Q^{20}$, the trend becomes more linear, with all bars exhibiting similar rates of increase in percentage entrainment with flow increase. 94% of the surface of both of the upstream, higher energy bars is predicted to be entrained during the large $Q^{100}$ flood. Entrainment for Bain Bar increases more slowly, with a lower proportion entrained (89% for $Q^{100}$) compared to the upstream bars. Bar reworking was lower for Breakfast Bar across all events excluding the $Q^{2.33}$ flood, with only 78% entrainment predicted during the $Q^{100}$ flood. Predictions of reworking for the largest flood events are more uncertain as the bed may become fully mobile, enabling the transport of clasts that would be predicted to remain static. This may cause more sediment to be entrained than predicted, making the uncertainty of sediment transport higher for larger events.

### 6.4 DISCUSSION

#### 6.4.1 Use of Bar Erodibility to Inform Larger Scale Processes

This section compares bar reworking with channel adjustment from aerial photographs to relate local scale processes back to adjustment at the reach scale.

##### 6.4.1.1 Blue Bar Reach-scale adjustment

Blue Bar was created pre-1958, and has alternated between periods of stabilisation and reworking. Lahar derived boulder lag from terrace erosion supplies large sediment. As a result, the less frequently reworked bar head makes up half the bar surface. A flood event greater than the $Q^{10}$ is necessary to entrain between 50–60% of this unit, highlighting the role of large floods in reworking this bar. This was supported through the aerial photograph history. Blue Bar was created during the 1958 flood, with a discharge of over 1400 m$^3$s$^{-1}$ (about a $Q^{60}$), which this study found was sufficient to rework the majority of the bar surface. The 1964 flood created a similar response, with a discharge of 1000 m$^3$s$^{-1}$ ($Q^{10}$) stripping vegetation. Few large floods occurred between 1973 - 1993, with the largest discharge recorded at 800 m$^3$s$^{-1}$ ($\sim Q^{10}$) (six years before the following aerial photograph), resulting in an increase in vegetation. This indicates that frequent, smaller flows were not sufficient to rework this surface. A flood with a discharge of 1400 m$^3$s$^{-1}$ ($Q^{60}$) in 2004 stripped the vegetation and reworked the bar again. Steep slopes (0.005) ensure that all but the largest material can be moved during these high magnitude events. This history of adjustment supports the degree of reworking predicted in the relative erodibility analysis.
6.4.1.2 Red Hut Bar Reach-scale Adjustment
Red Hut Bar has had a more dynamic history of adjustment compared with Blue Bar. This is reflected in a short age of < 5 years. The reach it is contained within is similarly sensitive, undergoing dramatic change over the past 80 years. Relative erodibility reflects this, as this bar had the highest bar percentage entrained for all events, except the Q\textsuperscript{2.33} flood. The node of coarse material, formerly a riffle head, now forms the head of the lateral bar, undergoing little entrainment until large flood events > Q\textsuperscript{20} flood. Bar heads have commonly been found to be an extension of riffles (Church and Jones, 1982). Finer cobble and very coarse gravel has been deposited around this node to form a lateral bar. This fraction can be reworked across a range of flows from the Q\textsuperscript{2.33} flood, and is responsible for the highly sensitive nature of the reach. This was displayed in the rapid bar evolution, as large floods in 1958, 1964 and 2004 with discharges between 1000 - 1400 m\textsuperscript{3} s\textsuperscript{-1} (Q\textsuperscript{10} – Q\textsuperscript{60}) significantly reworked the channel, whilst small floods yielded negligible geomorphic response.

6.4.1.3 Breakfast Bar Reach-scale Adjustment
Breakfast Bar exhibited the lowest percent of bar reworking for all flows except bankfull. Whilst the sand fraction can easily be removed from the tail of the bar during frequent Q\textsuperscript{2.33} floods, the coarse cobble material adjacent to the water’s edge remains static across all flows modelled. This was reflected in the age of the bar of > 50 years. Channel adjustment within this reach has been minimal. Low slope (0.0029) is not sufficient to move large material delivered from terrace incision, except at very large events (i.e. > Q\textsuperscript{50}).

6.4.1.4 Bain Bar Reach-scale Adjustment
Bain Bar is located within the most active reach of the river, where sediment exiting the terraces is deposited to form an aggrading braidplain. A greater proportion of sediment can be entrained during the Q\textsuperscript{2.33} flood compared with the other bars, indicating that this bar can be more frequently reworked. This is supported by the evolutionary history of the reach, which has undergone significant adjustment through channel migration and braiding over the past 80 years. For a larger flood event (> Q\textsuperscript{10}) a greater proportion of Bain Bar was entrained compared with Breakfast Bar. In both instances this is a lower proportion than the two upstream sites. This reflects the trapping of cobble material within this reach, as slopes are not sufficient to transport it further downstream.

6.4.2 Grain-size Fractions
This section relates variability in sediment sizes within the Tongariro catchment back to their sources, residence times and depositional environments. Understanding characteristics of the active fraction provides key insights into the processes forming and reworking these bar surfaces.
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The largest fraction is the boulder lag, which makes up the head of the bars. This can be seen in the peak larger than 280 mm at Blue Bar and Red Hut Bar (Figure 6.17). At Blue Bar this peak is less evident, around 180 mm and above, whilst at Bain Bar it is largely absent. Analysis of sediment size within terrace deposits at Blue Bar indicates that the larger material is likely to be reworked lahar deposits delivered from the terraces. This sediment is deposited on the bar head, where it is infrequently reworked. The largest clasts are predicted to remain static during the $Q^{100}$ flood. Lichen covers their surfaces, indicating very long residence times.

A second grain-size fraction can be seen in the coarse material which makes up the majority of the bar, in particular, the supra-platform and back channel. This has a bi-modal distribution for Blue and Red Hut Bars with peaks at 260 and 180, and 220 and 160 mm respectively. At Breakfast Bar the distribution is relatively flat, including clasts between 100 – 200 mm. At Bain Bar this is sediment greater than 80 mm. The smaller clasts in located in areas of high bed shear may be reworked during the $Q^{33}$ flood. However, major reworking requires an event of $Q^{10}$ or above. The reworking of this fraction is key to morphological adjustment in bar shape and extent around the more fixed bar head. The large size of the material is most likely a combination of the smaller clasts delivered from terrace incision combined by the coarser fraction moving through the river. Median grain size of the tributaries draining the western Volcanic Plateau ranges from 95 – 210 mm (Table 3.3) indicating evidence of some overlap in clast size.

The smallest peak in grain size is likely comprised of sediment delivered from the trunk streams draining the Kaimanawa Ranges. This was seen in a peak between 40 - 60 mm at Blue Bar and 20 - 40 mm at the other three bars, classified as coarse gravel to very coarse gravel. The $D_{50}$ of the two trunk streams in the Kaimanawa Ranges before the confluence with the Tongariro is 65 mm for the Whitikau Stream and 61 mm for the Waipakihi Stream. As Blue Bar is downstream of these confluences, this explains a decrease in size for this fraction (Powell, 1998). This fraction was deposited within bar tails, fine-grained deposits and back channels at the downstream sites. It is a transient and frequently reworked, characterised by short residence times (a few months to a year) and is easily flushed through the terraces to the lowland plain. Predicted supply is approximately 800,000 m$^3$ per year (Chapter 3).

The wide range of grain size creates bars with complex patterns of sorting. Primary sediment sources within the Tongariro catchment have a distinctive size associated with it, with boulders delivered from the terraces and coarse gravels from the Kaimanawa Ranges. The volumes delivered from these sources and their residence times are a key influence upon bar configuration and channel adjustment.
6.4.3 Mechanisms of Bar Development

Mechanisms of bar development along the Tongariro River differ considerably between reaches. Studied bar platforms conform aptly to bars characterised by Church and Jones (1982) as (i) hydraulic elements, where bar morphology is formed by the deformation of the bed, (ii) storage elements, where bar morphology is formed by the storage of active material as it is transferred through the reach, or (iii) a combination of both hydraulic and storage elements. This relates the form of the bar to its creation by erosional or depositional processes.

Blue and Breakfast Bars, which are located at the upper and lower sections within the terraces (Figure 6.1) were formed by large floods (> Q$^{50}$). These events stripped vegetation and fine grained sediment from the floodplain, uncovering a lag of coarse boulder material to form the bar head. As such, the bulk of each bar is composed of hydraulic elements. This is reflected in a high proportion of the bar area which is classified as bar head (Table 6.7). This unit is characterised by coarse cobbles. It makes up 40 – 41 % of the area for these bars compared with < 33 % for the other bars. The greater spatial extent of the bar head is morphodynamically significant since it indicates that a greater proportion of the bar will undergo minimal reworking during frequent floods. Little or no evidence of systematic increase in size and linear growth over time (Figure 6.9) further supports the characterisation of these bars as hydraulic elements, as storage elements would be expected to undergo on-going adjustments to their shape, reflecting variability in sediment supply from upstream (Bridge, 2003; Church and Rice, 2009). In contrast, bar size reflected time elapsed since the last flood event large enough to rework the bar and floodplain surface, stripping vegetation (> Q$^{20}$).

The second most upstream bar in the study area, Red Hut, has a more complex evolutionary history (Figure 6.9). Storage elements have existed at this location since 1928. These elements have been extensively reworked, from a lateral bar, to a mid-channel bar, and back. Deposits have existed at the present location of the bar head throughout this time, indicating that the bar head is likely a hydraulic element. This is supported by the Q$^{20}$ size of flood event noted to be necessary to rework this unit, which is the same magnitude as that for the Blue and Breakfast bars. The rest of the bar is more transient, made up of a sediment fraction that can be more readily entrained and transported during Q$^{10}$ flood flows. Ashworth (1996) presents a model for mid-channel bar growth, where big sediment is deposited downstream of a confluence, and flow converges causing scour and then diverges causing deposition. Similar processes are likely to have caused the deposition of the coarse nucleus of Red Hut Bar, as flow converges downstream of Poutu Island (Figure 6.9). This island is large by Tongariro River standards, at 300 m wide and 500 m long. It has existed in some form or another at this location since the earliest survey map in 1928, acting to protect the area within
which Red Hut Bar is located, and allowing storage of finer fractions in this vicinity. As such, secondary units that comprise the lower half of the Red Hut Bar are frequently reworked, causing the high sensitivity of the bar (c.f. Church and Rice, 2009).

The whole of Bain Bar, the farthest downstream bar in this study, can be classified as a storage element. The floodplain-delta containing the bar was formed following the Taupo eruption, 1.8 ka BP (Rosen et al., 2002; Smart, 1999; Wilson and Walker, 1985) and, as such, the bar consists of sediment which has been delivered by the contemporary regime. Bar adjustment is driven by channel shift, as pulses of sediment are delivered during floods > $Q^{2.33}$, and the configuration of bars is altered. Thus, bar reworking is a function of the switching and alignment of primary channels during these large magnitude events, rather than the formation of a unit bar head followed by the addition of secondary units as commonly suggested in the literature (Ashmore, 1982; Ashworth, 1996; Bridge, 2003; Church and Rice, 2009; Lunt and Bridge, 2004).

Models of mid-channel bar development describe coarse sediment being deposited at a locale of decreased shear stress, facilitating further deposition of smaller material downstream (Ashmore, 1982; Ashworth, 1996; Bluck, 1987; Bridge, 2003; Hein and Walker, 1977; Lunt and Bridge, 2004; Smith, 1974). However, for the terrace-confined bars within the Tongariro River, the active fraction delivered from upstream is combined with the lag of lahar boulders which is reworked to form the bar heads. As such, bar development is not simply a function of changes in shear stress capturing mobile fractions, but also the moulding and redistribution of this larger fraction which may be retained within the reach across long residence times (100-1000 years). This makes the role of hydraulic elements in bar formation especially pertinent within this system. Most existing models of bar development are created for rivers with active beds, especially within flume work (Ashmore, 1982; Ashmore, 1991a; Ashworth, 1996; Bridge, 2003; Knighton, 1998). This highlights the need to further investigate the influence of lag sediment from other sedimentary influences (volcanic, glacial or due to decreased stream capacity following regulation) on bar development and channel adjustment.

6.4.4 Influence of Reach Position on Bar Reworking

The data assembled for this study indicate that reach-scale controls, such as slope and valley confinement, influence patterns of sediment reworking on bar surfaces. The $Q^{2.33}$ flood entrained a lower proportion of the two upstream bars, Blue Bar and Red Hut Bar, compared to the two downstream bars, Bain and Breakfast Bar. These upstream bars are characterised by steeper slopes of 0.0077 and 0.0051 and larger median grain sizes of 230 mm and 140 mm (Table 1), compared to the two downstream bars which have slopes < 0.003 and $D_{50} < 114$ mm. Blue Bar and Red Hut Bar
are also located within partly confined floodplain pockets, with adjacent terraces between 25 and 12 m high respectively (Fryirs and Brierley, 2010). Terrace heights are an indicator of the volumes of lahar deposits that have been delivered to the river. Despite steep slopes, this source of boulder-sized sediment decreases the ability of frequent floods to rework the bar and elicit planform change.

The proportion of Breakfast Bar able to be reworked by $Q^{10}$ floods and greater was well below the other bars. Again, this is considered to reflect the coarse lag of sediment from terrace incision. Whilst only 7 m of incision has taken place at this site, the low slope (0.00298) of the reach restricts the capacity of floods to rework and redistribute this material. As such, this creates a very coarse bar head which restricts entrainment even during the largest floods, with only 60% of bed materials being entrained during the $Q^{100}$ flood.

Bain Bar is located within the braided reach on a depositional floodplain and thus has different process characteristics compared to the other bars. Whilst Bain Bar does contain remnant or reworked lahar deposits (Cronin et al., 1997), the grain-size distribution is more strongly influenced by sediment delivered by the contemporary flow regime. This is reflected in a high proportion of smaller, more mobile material (77% coarse gravel or finer). This fraction is frequently reworked by smaller, more frequent flood events, which is reflected in the high entrainment predicted for the $Q^{2.33}$ flood (Figure 9). In comparison, the larger fraction which makes up the bar head is infrequently reworked, as $>Q^{10}$ magnitude events elicit less reworking here than at the other sites (excluding the Breakfast Bar). The marked decrease in slope along the reach containing Bain Bar is reflected in the gravel-sand transition being located immediately downstream of Bain Bar. As such, this zone traps larger cobble sediments which the channel is not competent to transport further downstream. In addition, the unconfined nature of the reach limits stream power generated by flood events greater than bankfull. Once flows exceed bankfull, water is dissipated onto the floodplain, and flow depth and shear stresses do not increase substantially (c.f. Sambrook Smith et al., 2010).

The volume of lahar lag delivered to the channel influenced the magnitude of flood necessary to rework it. Bluck (1987) identifies the additional complexity in understanding channel adjustment when the bed material is derived from a range of sedimentary events (e.g. such as delivered from fluvial entrainment and landslides). Most commonly this lag is delivered at discrete points (e.g. tributaries) (Rice et al., 2008). However, because lahar materials are delivered from terraces throughout the reach, it is more difficult to monitor its influence on channel dynamics. In addition, local variations in incision and the size of sediment within the former lahar deposits further complicate matters. Brummer and Montgomery (2006) discuss the role of lag from landslides in inhibiting bed incision and protecting the underlying fraction from reworking. Within the Tongariro
system the lag is reworked to form bar heads that are characterised by long residence times, as during the $Q^{100}$ flood 10% - 40% of bed materials are not entrained. As such, this material plays an important role in fashioning the reworking of bar surfaces and channel adjustment.

Dams and associated flow regulation are associated with major impacts to the sediment transport of rivers, usually due to a decrease in sediment transport capacity (Kondolf, 1997; Vörösmarty et al., 2003). Regulation within the Tongariro reduces frequent freshes (between 70 - 100 m$^3$s$^{-1}$) and low flows with a more natural flow regime observed once flows exceed 100 m$^3$s$^{-1}$, when control gates are opened to flush sediment (Collier, 2002; Hindle, 1995). The minimum discharge event studied in this work was 480 m$^3$s$^{-1}$, substantially larger than the floods which have been reduced. This flood event was able to entrain ~ 40% of the downstream bars and ~ 30% of the upstream bars. However, the locales of reworking were limited to tail and backchannel units, representing minimal influence upon bar reworking and bedload transport across these units. The lahar lag in the system increases the magnitude of flood necessary to rework bar surfaces. This means that a reduction in the smaller, more frequent 100 m$^3$s$^{-1}$ floods has had minimal influence on planform adjustment and bar reworking, as these flows are not geomorphically effective. This was supported in the limited change to planform adjustment observed since regulation (Chapter 4). However, these smaller flows are important for transporting finer fractions such as sand and fine gravels and an increase in this fraction filling the interstices of boulders have been observed (Jowett, 1984).

6.4.5 Conceptual Model of Bar Reworking

Insights from patterns of sediment entrainment on geomorphic units were used to create a conceptual model of bar reworking that describes the geomorphic effectiveness of each flood (see Costa and O’Connor, 1995). Usually geomorphic effectiveness is related to maintenance of the channel shape or cross-section, with effectiveness measured as the volume of sediment transported for different sized floods (Andrews, 1980; Lenzi et al., 2006; Wolman and Gerson, 1978). More recent approaches have started to relate sediment transport on gravel bars to geomorphically effective flows (e.g. Laronne and Duncan, 1992; Laronne et al., 2001; Surian et al., 2009a). However, these studies are spatially and temporally limited since they only consider geographically discrete patches upon a bar. In addition, analysis is dependent upon the magnitude of flood events which occur during a sampling period, which may not capture the system’s morphologically formative events. Thus, the bar reworking approach developed within this chapter considers geomorphic effectiveness across the whole of the bar for the range of flood events.

Each flood event was found to mobilise different sediments within bar units (Figure 6.26). From this, a conceptual model of bar reworking was created (Figure 6.27). The $Q^{2.33}$ flood flushed fine grained
material from the tail of all bars, as well as mobilising 60-70 % of the backchannel for Bain Bar, the most downstream bar located in the braided reach. The $Q^{10}$ flood greatly increased entrainment for all bars, mobilising > 50 % of the supra-platform and > 70 % of back-channels. Floods of $Q^{20}$ and above mobilise a greater proportion of all units, including much of the bar head. The $Q^{20}$ flood entrains > 70% of bar heads at the high slope, terrace confined bars and > 40 % of bar heads at the two downstream bars. This indicates that bar heads in the downstream bars are competence limited, with larger clasts trapped within these reaches due to low slopes. Flood discharge increases above the $Q^{20}$ level increases the proportion of the head and supra-platform units that are entrained.

Investigations into geomorphic effectiveness have traditionally identified a single discharge, typically the effective discharge, which transports the greatest volume of sediment over time (Andrews, 1980; Emmett and Wolman, 2001; Wolman and Gerson, 1978). These flows are usually perceived to be around bankfull flow, with a recurrence interval similar to the mean annual flood (Andrews, 1980; Emmett and Wolman, 2001; Wolman and Miller, 1960). However, more recent work has suggested the importance of ‘multiple forming discharges’ or ‘bi-modal’ discharges, whereby smaller floods shape within-channel environments, while larger floods are necessary to rework higher bar surfaces and the macro-channel (Lenzi et al., 2006; Phillips, 2002; Surian et al., 2009a). The evidence collated in this chapter suggests that a range of discharge magnitudes is necessary to explain the geometry and bed material distribution on bar surfaces. Within the Tongariro River, small floods ($Q^{2.33}$) flush and deposit smaller gravel fractions on the tail and back-channel of bars. These floods are likely responsible for the majority of sediment transport through the reach, especially moving the smaller active material delivered from the eastern sub-catchment which includes the Kaimanawa Ranges. However, these floods have minimal influence on channel change. In contrast, larger floods ($> Q^{20}$) are required to entrain the coarse cobble-boulder fraction on the supra-platform and head of the bars. Whilst these floods may not cumulatively transport as much sediment as the smaller, more frequent events (c.f. Wolman and Miller, 1960), they are far more geomorphically effective at reworking the lahar boulder lag deposits and eliciting planform change.

The conceptual model shown in Figure 6.27 highlights the high similarities in reworking of bar heads during the $Q^{2.33}$ flood. All bar head undergo between 13 - 19 % entrainment. This is surprisingly similar given the marked differences in slope, grain-size and mechanisms of bar adjustment between the reaches. This indicates that the bar head deposit is graded to each reach position to obtain a balance between the coarse fraction and the shear stress generated during these frequent floods. Indeed, Brummer and Montgomery (2006) discuss the role of coarse clasts in providing a ‘stable
nuclei’ for step formation, and Carling et al. (2006) suggest that bars act as ramps, which feed finer-grained sediment to the body and tail of the bar, providing a rationale for fining. Within the Tongariro system, bar formation requires the deposition or concentration of a nucleus of clasts large enough to be resistant to frequent flood events. These relative ‘nodes of stability’ then precipitate deposition and protect more transient material to their lee (as suggested by Carling et al., 2006), creating units that are reworked by different magnitude floods. Whilst it is well documented that bar heads are reworked less frequently than the downstream units (Bluck, 1976; Bridge, 2003; Leopold et al., 1964), understanding which flows bar heads are adjusted to has not been previously documented to the writer’s knowledge.

Channel adjustment at the reach scale was found to be linked to the proportion of bar made up of more frequently entrained units (tail, backchannel, supra-platform) relative to the more static units at the bar head. Red Hut Bar underwent greater channel and bar adjustment compared with Blue Bar (Figure 6.9) despite both bars displaying similar patterns of sediment entrainment for each unit (Figure 6.26). This can be explained by 41 % of Blue Bar being classified as bar head compared with only 15 % of Red Hut Bar. This difference drives 5 % more sediment entrainment on Red Hut Bar (the head and overall) during the Q$^{2.33}$ event (Table 6.7). Once flows exceed the Q$^{2.33}$ flood discharge, the high reach slope at Blue Bar (0.0077 m/m) is competent to entrain a greater proportion of the bar head, minimising the impeding role of this large sediment. Similar to Blue Bar, Breakfast Bar also underwent minimal adjustment during the past 80 years and has a comparable percentage of bar head (40%). Due to low slopes within this reach (0.0029), the large proportion of coarse cobble material impedes entrainment across all events modelled (Figure 6.10). This is reflected in only 60 % of the bar head predicted to be entrained during the Q$^{100}$ flow, causing much lower predictions of overall bar reworking for Breakfast Bar than the other bars (i.e. 78 % overall entrainment compared to > 89 % for the other bars). This demonstrates that the proportion of the bar composed of active units (particularly the tail and back-channel) compared to the more static bar head exerts a direct influence upon the extent of adjustment of the bar and the reach as a whole.
Figure 6.26: The percentage of entrainment for each geomorphic unit across the range of flood events modelled. Not to scale.
6.4.6 Consideration of Methods

Recent advances in technology, such as TLS and photogrammetry, are revolutionising our ability to gain high resolution measures of grain size and river morphology (Carbonneau et al., 2005; Graham et al., 2005; Heritage and Milan, 2009; Hodge et al., 2009; Rychkov et al., 2012), ultimately changing the way we can quantify bar geometry and surface sedimentology. To date, most investigations have focused upon methodological developments of these techniques. As a result, spatial scales of investigation are limited, commonly to the ‘patch’ scale of a bar, and few examples exist whereby
the technologies are applied to aid the understanding of geomorphological form and processes at broader scales of enquiry. This study extends the spatial scale of morphological and grain size mapping using TLS to the scale of multiple bars, contextualising contemporary observations within a historical narrative. Moreover, this study increases the range of grain sizes previously mapped with this technology to include the cobble and boulder fraction (c.f. Hodge et al., 2009; Rychkov et al., 2012). Findings have been used to interpret mechanisms of bar evolution, providing a foundation to analyse bar reworking.

A major strength of the bar reworking approach is the ability to calculate sediment entrainment at the scale at which grain size varies. In addition, modelling entrainment for specific flood magnitudes provides bench-marks of adjustment, from which likely future pathways of adjustment can be predicted. This recognises that while it is not possible to predict future distributions of flood events, it is possible to predict the likelihood of how a system will respond to flood events of a given size. The aerial photographs and flood analysis used herein provided a historical database with which to compare the model predictions made with past patterns of bar response. The database and predictions were found to correspond well, with the necessity for large floods (>$Q^{10}$) to alter bar form evident both in the model and real world.

Limitations of this study arise in the simplifying of real-world complexity. Sediment transport is notoriously difficult to measure and model due to local-scale complications including turbulence, bed roughness and the effects of hiding and protrusion on entrainment (Laronne et al., 2001; Wilcock and Crowe, 2003). Further limitations can be seen in the phenomenon of event sequencing (Beven, 1981; Hassan and Woodsmith, 2004; Marutani et al., 1999). In the Tongariro system, vegetation growth and density (and roughness) reflect the time elapsed since the last large flood, which may influence sediment entrainment. However, the large floods modelled within this study are expected to be easily competent to strip the vegetation and entrain underlying surfaces. For these reasons, this measure is indicative of process, measuring the potential work, rather than work itself. Its strength lies in using fewer variables and assumptions than are contained within more comprehensive modelling approaches. As such, this parsimonious approach presents a simplification of sediment transport relationships for which to investigate patterns of sediment reworking. For this reason, results should be regarded as indicative of process, rather than definitive, as indeed all models should be.

This study also did not include analysis of bank erosion. This was largely ignored due to the nature of the study site. The banks are coarse deposits from bars with sand above. As such, the sand is entrained easily compared to the coarser material of the bars. Minimal cohesion limits the types of
bank erosion that can occur (for example slumping) and instead when material is eroded it is transported immediately. Therefore calculating bar erodibility is far more meaningful than bank erosion for this situation. In other catchments, where bank erosion mechanisms are more important for fluvial dynamics it may be more important to include this as a factor influencing channel adjustment.

In summary, this work successfully created a novel approach for analysing bar reworking, extending both the type and scale of approach available to characterise bar reworking and predict future channel adjustments.

6.5 CONCLUSION

Bars are a commonly overlooked, yet integral component of gravel-based river systems. In part, the intricacy of bar sedimentology has been a major limiting factor in our ability to capture the complexity of these surfaces, and predict how they are likely to evolve into the future. This is especially true of sedimentary bimodal systems, where material is delivered from non-hydraulic processes, in this case lahars, where a wider range of flood events bring about complex patterns of reworking. This study highlights the importance of understanding bar features as indicative of stability and sediment transport at the reach scale (i.e. Church and Jones, 1982). The exponential growth in new technologies has dramatically increased the scale and resolution with which we can capture and characterise these features and their process-form relationships. These new techniques allow us to address old questions in new and exciting ways, widening the scope and our ability to derive insight from detailed field-based analyses.

Results from this chapter extend previous attempts to characterise bar reworking by developing a relatively simple approach that couples sedimentology and hydraulic forcing to explain bar morphodynamics. The key findings are (i) ‘multiple bar forming discharges’ exist within the Tongariro River system, whereby frequent flood flows rework tail and back-channel areas, whilst much larger, less frequent floods are required to mobilise coarse-boulder and cobble fractions on the bar head; (ii) bar heads across all reaches analysed were found to be resilient to the mean annual flood; and (iii) reach-scale sensitivity is largely a product of the proportion of the bar that is made up of active, frequently reworked units compared to the more static bar head materials. This work provides a process-based appraisal for predicting future patterns of adjustment. For large, dynamic systems where planform change is driven by the reworking of bar deposits, rather than upstream sediment supply, these types of enquiry are fundamental. This work acts to push the boundaries of how we model bar surfaces, presenting a ‘novel’ approach that quantifies the geomorphic effectiveness of flows to rework bar surfaces.
Chapter 7: Evolutionary Trajectories

7 PROCESS-BASED EVOLUTIONARY TRAJECTORIES FOR THE LOWER TONGARIRO RIVER

7.1 INTRODUCTION
An evolutionary trajectory is a conceptual model which uses appraisals of past system adjustments and responses to disturbance events, to predict likely future pathways and rates of change, identifying threshold induced changes that may occur. In the past, most applications of the evolutionary trajectory approach have been two dimensional conceptual models which show the evolution of the planform or cross-section of a reach over time (Beechie et al., 2008; Brewer and Lewin, 1998; Brierley and Fryirs, 2005; Comiti et al., 2011; Dufour and Piégay, 2009; Fryirs et al., 2009; Gurnell, 1997; Hooke, 2003b; Hoyle et al., 2008; Simon and Rinaldi, 2006; Surian et al., 2009b; Wishart et al., 2008) (see Chapter 1 for full discussion). At present, description of the processes driving adjustment are rarely considered, despite recognition of the need to combine historical approaches with a process based underpinning (Dufour and Piégay, 2009; Small and Doyle, 2012). Exceptions include Hooke (2003b) who related channel competence to channel adjustment and Ziliani and Surian (in press) who used CEASAR modelling to calculate bedload transport to support future predictions of channel evolution. However, a greater understanding of the processes driving channel change and sediment transport, are essential to strengthen conceptual models which predict channel change. In addition, many previous examples lack a catchment scale framing, which contextualises the reach within catchment-scale connectivity (e.g. the patterns and rates of sediment fluxes across the catchment), and thus they ignore the importance of sediment storage and delivery in driving channel adjustment downstream (c.f. Brierley and Fryirs, 2009; Kondolf et al., 2006; Raven et al., 2010; Sear et al., 2009; Walling and Collins, 2008; Wohl et al., 2005). This chapter synthesises insights from previous chapters across the catchment, reach and bar scales to create evolutionary trajectories for the lower Tongariro River that are grounded within underlying processes operating across multiple scales. This provides a more comprehensive, process-based foundation upon which to estimate channel change that has previously been constructed using an evolutionary trajectory approach. The lower Tongariro River was selected as the study site, due to its high capacity to adjust once the terraces widen and its role as a threat to the Turangi township.

7.2 RESULTS
Evolutionary trajectories are presented for each of the four River Styles on the lower Tongariro River, due to similarities in response to disturbance events and underlying controls that characterise adjustment for each river types. Future response trajectories are then presented, which suggests how a river is likely to adjust in response to different natural and anthropogenic influences.
7.2.1 Process-based Evolutionary Trajectory for the Wandering Cobble bed River

The Wandering, cobble bed river reach began evolving along its current trajectory following the eruption of Lake Taupo 1.8 ka. Prior to this, the channel adjusted across an unconfined floodplain, built up by lahars, which delivered pulses of coarse-grained sand-boulder material from the Volcanic Plateau (Cronin et al., 1997) (Figure 7.1). The Taupo eruption (1.8 ka) smothered the floodplain and channel with approximately 10 m of pyroclastic tephra material, killing all vegetation (GNS, 2010). Drainage lines were reinitiated rapidly, incising into the loose tephra and then the underlying lahar deposits (Manville, 2002; Manville et al., 2009; Smith, 1991). This incision a) caused the channel to become partly confined within terraces and b) mobilised lahar deposits to form a lag which lines the contemporary channel. Both of these factors are key controls on the evolution and adjustment of the contemporary channel.

The formation of terraces set valley confinement for this reach. Floodplain width within the terraces fluctuates downstream creating marked differences in imposed boundary conditions. Patterns of valley slope immediately post-Taupo eruption (1.8 ka) shaped differences in the potential energy (and transport capacity) for each reach. A strong positive correlation with an $R^2$ of 0.87 was observed between the slope on the top of the terrace (i.e. pre-eruption valley slope) and terrace width (Chapter 5). This indicates that steeper reaches removed a greater volume of sediment, creating wider floodplain pockets within the terraces. Reaches that were characterised by flatter gradients 1.8 ka are now more confined, and adjust within much narrower floodplain pockets. Valley pinch-points within these reaches directly influence channel position and alignment at these points. Thus, valley confinement set by the terrace locations provides a key control on the evolutionary trajectory observed within this reach in the past 80 years.

On-going incision mobilised underlying, unconsolidated lahar deposits which ranged from sand to boulder in size (D-max of 1 m). Finer grained fractions ranging from sand to very coarse gravels can be rapidly flushed by this system (i.e. the $Q^{2.33}$ flood is competent), whilst the boulder sized fraction became trapped in the system, remobilised to form riffles and bar heads. This lines the channel with cobble-boulder sized material which requires large flood events ($> Q^{26}$) to rework the bed and banks of the channel. Frequent floods (i.e. the $Q^{2.33}$) were found to rework between 29-41 % of bars, flushing finer sand and gravel fractions on the bar tails and back channels. In comparison, the $Q^{10}$ flood could mobilise 60-72 % of bar surfaces, reworking the body of the bar. However, even larger floods are necessary to rework the largest of the lahar lag and cause dramatic changes to channel planform. The $Q^{50}$ flood is predicted to entrain 73 % of the downstream (low slope) bar and 90-91 % of the upstream (steeper) bars within the terraces, reflecting a marked increase in the ability of this flood to elicit planform change. Thus, the lahar boulder lag requires large, infrequent floods to fully
mobilise it and drive planform change. This is supported by the high adjustment observed following two Q_{60} floods in 1958 and 2004, including floodplain stripping and substantial change to channel and bar units (note: due to the flood frequency distributions the Q_{60} is only 100 m^3s^{-1} smaller than the Q_{100} at 1400 m^3s^{-1} and 1500 m^3s^{-1} respectively).

Large sediment size and infrequent reworking limits the rate of incision and the ability of the river to maintain a graded long profile. As a result this reach has retained steep slopes (0.0046 - 0.0111 m/m), especially compared with the unconfined reaches downstream (0.0026 m/m) (Table 4.6). The Kaimanawa Ranges provide the major sediment source, supplying an estimated 800,000 m^3 annually (Chapter 3). The majority of this fraction is smaller than that lining the bed of the Wandering, cobble bed reach, with D_{50} of 60 - 65 mm delivered from the trunk streams, compared to the 90 - 233 mm material which lines the bed of the wandering reach. The fraction from the Kaimanawa Ranges is easily flushed by frequent floods (i.e. the Q_{2.33} flood). This material makes up bar tails and back channels. These transient units are characterised by short residence times (estimated to be < 3 years given the frequency of flows around the Q_{2.33} events, recognising that smaller flows are also likely to move much of the finer sediment). As such, upstream supply of sediment does not drive change within this system. Rather sediment which is delivered is transferred through this reach relatively rapidly, with minimal influence upon channel planform adjustment.

Evolutionary trajectories based on channel response for reaches within the wandering, cobble bed river can be separated into sedimentation zones which were between 118 – 365 m wide, characterised by large bar complexes and usually multiple channels and single-channelled reaches between 38 – 75 m wide characterised by minimal in-channel storage and low sensitivity (Chapter 5). This descriptive classification captured marked variability in in-channel storage and sensitivity (rate and extent of adjustment), essentially describing the multi-channelled sedimentation zones and relatively static single channelled reaches as described in Desloges and Church (1989).
Chapter 7: Evolutionary Trajectories

Figure 7.1: Evolutionary trajectory for the Wandering, cobble bed river. This describes evolution since the Taupo eruption (1.8 ka) and controls that drive contemporary channel response.

**Long-term Channel Evolution**

**Pre Taupo eruption (1.9 ka)**
- Channel underlain by lahar deposits
- Channel connected to floodplain
- Native vegetation on floodplain

**Post Taupo eruption (1.8 ka)**
- Channel smoothened by approx. 10 m tephra from Taupo eruption.
- Vegetation removed increasing discharge and the rate of incision until vegetation returns.
- Channel incises into tephra.

**Present day**
- Incision created terraces, laterally confining the channel.
- Terrace width determined by valley slope pre-Taupo eruption.
- Incision into lahar deposits lines the channel with a lag of boulders ($D_{50} = 91 - 230$ mm). These require large, infrequent floods to rework (entrainment for $Q_{50} = 73\%$ ds bar, 81% us bars).
- Lahar lag limits incision, retains steep slopes that flush the mobile fraction from Kaimanawa Ranges ($D_{50} = 60$ mm) during frequent floods (e.g., $Q^{2.23}$).

**Single-channelled reaches**

**Sedimentation zones**

**Controls on single-channelled reaches**
- Forced low sinuosity planform from valley pinch points creates high slopes with limited zones of reworking.
- Lower storage of mobile fraction.
- High and more varied transport capacity* (us reaches 3.5, ds reaches 2.25).
- Limited response to high magnitude events ($Q^{60} = \text{avg. width increase of 23 m}$).

**Controls on sedimentation zones**
- Channel planform (sinuosity/braiding) has adjusted to balance sediment and slope.
- Greater storage of mobile fraction.
- Low transport capacity* (us 2.23, ds 0.95).
- Tend to have gentler gradients than adjacent narrow reaches.
- Greater sensitivity of response to high magnitude events ($Q^{60} = \text{avg. width increase of 80 m}$).

*Shear stress ratio for $Q^{100}$ flood; us indicates sites in the upper, steeper zone of the reach, and ds those at the downstream zone characterised by gentler slopes and smaller bed material.
Floodplain width within the terraces (terrace width) provided a key control on the distribution of sedimentation zones and single-channelled reaches despite never directly controlling the channel width (i.e. there is enough room within each floodplain pocket for all reaches to create a wider active channel). However, pinch-points or localised narrow valley sections up- and downstream of single-channelled reaches acted to force a less sinuous, steeper planform, which tended to be characterised by higher transport capacities. In contrast, the sedimentation zones located within wider valleys had the capacity to adjust their form to create a balance between slope (altered by sinuosity), cross-section (altered by width of bar and island complexes) and grain size (reflecting lahar lag and the proportion of storage from the supply upstream). As such, slope and grain size were in sync, forming a tight linear trend ($R^2 = 0.97$) at sedimentation zones compared to greater scatter observed for single-channelled reaches ($R^2 = 0.31$). In addition, sedimentation zones tended to have low transport capacities. For example, the reaches classified as sedimentation zones in the upstream reach exhibited an average shear stress ratio for the $Q_{100}$ flood of 2.23 compared with 3.5 for the single-channelled reaches, and the downstream reaches exhibited ratios of 0.95 and 2.25 respectively. This suggests that single-channelled reaches are able to flush active fractions more readily than sedimentation zones which store excess material in active bar complexes.

Future adjustments for sedimentation zones and single-channelled reaches can be best described through differences in response to past events. Single-channelled reaches were found to undergo a width increase of 23 m following the $Q_{60}$ flood, showing limited change in bar extent and reworking. In contrast, the average response for sedimentation zones was 80 m, undergoing substantial increase in channel area through floodplain stripping, avulsion, chute cutoffs and bar and island reworking. The difference in past response sets an underpinning for how this reach is likely to respond to future events.

### 7.2.2 Future Response Trajectories for the Wandering Cobble bed River

This section uses past patterns and rates of response to create likely future trajectories of response to multiple anthropogenic and natural scenarios. While it is recognised that multiple overlapping controls can act to complicate responses, understandings of likely future response to a single influence is necessary to underpin more complicated scenarios.
Floods were shown in Chapters 4 and 6 to be the largest, most frequent cause of channel change for the Wandering, cobble bed river. This is because to alter the channel, high shear stresses (e.g. $Q^{50}$) are required to rework the cobble-boulder bed material and strip vegetation off bar and floodplain surfaces. As discussed above, more sensitive sedimentation zones are more likely to undergo greater change (e.g. width increases of 80 m) than single-channelled reaches which exhibit dampened responses (only 23 m increase in width).

Volcanic eruptions provide an additional high magnitude influence, especially in the form of lahar flows, whereby high loads of unconsolidated volcanic material is delivered to the lower catchment.
In 1995 a lahar delivered an estimated 6900 kilotons of fine grained sediment (two thirds < 0.5 mm) to this reach (Collier, 2002; Manville et al., 1996). However, minimal impact on planform characteristics was observed. Indeed, the short residence time of this material within the reach illustrates the resilience of the system to pulses of fine grained material (~2 years) (Collier, 2002; Genesis Energy, 2000). Steep slopes (> 0.004, with an average of 0.007) allow minimal residence time for finer fractions within this reach, flushing it rapidly to the delta. In comparison, lahars comprising much larger cobble-boulder material are projected to cause substantial widening and aggradation of the active channel, creating long recovery times ~ 100 - 1000 years. This is illustrated in the longevity with which these deposits are stored within this reach (e.g. within terrace exposures). While finer sand and gravel fractions within this mix can be rapidly flushed during < Q_{1/33} events, much larger floods (i.e. Q_{50} - Q_{100}) are necessary to rework the largest cobble – boulder fraction, creating these extended recovery times. In addition, the sedimentation zones are predicted to undergo longer recovery times. These zones were characterised by lower transport capacities (a mean of 1.75 compared with 2.89 in the single-channelled reaches) allowing greater storage of lahar lag and longer recovery times for the channel to mobilise and redistribute the larger fraction of boulder sized grains.

The impact of multiple anthropogenic controls has been observed within this reach. An increase in the base level of Lake Taupo was found to have minimal/no effect upon the adjustment of this reach. The backwater effect this creates is predicted to extend 3 km above the margin of Lake Taupo (Tonkin and Taylor, 1999b). The wandering reach is 7 km above the lake margin, decreasing the influence of lake levels on sediment transport at this point.

Reduction of small floods had little effect upon this reach, as smaller freshes are competent to redistribute small gravels and sand, but are not competent to rework the bars and islands necessary to alter planform characteristics. A reduction of these freshes whilst maintaining large floods (400 m$^3$s$^{-1}$) is predicted to elicit minimal change. However, a reduction in the larger geomorphically effective flood events (> 400 m$^3$s$^{-1}$) has been observed to cause channel narrowing. Floods of this magnitude are geomorphically effective, able to strip vegetation and cause floodplain and bar turnover.

### 7.2.3 Long-term Evolutionary Trajectories of the Delta

This section describes delta evolution in response to the Taupo eruption 1.8 ka. This sets the boundary conditions for the braided, meandering and delta River Styles. More in-depth process-based trajectories for each of these reaches are described in the following sections.
Chapter 7: Evolutionary Trajectories

Pre-Taupo eruption (1.9 ka)
- Delta located 6 km upstream of contemporary position.
- Underlain by lahar deposits.

Present day
- Taupo eruption (1.8 ka) smothers delta with ignimbrite.
- Lake level drops ~10-15 m.
- Delta rapidly progrades to new level capturing reworked Taupo ignimbrite (deposited 1.8 ka) and ongoing supply from the Kaimanawa Ranges and Volcanic cones.
- Delta characterised by marked decrease in slope downstream.
- Drives rapid progression of distinctly different River Styles.
- Avg sediment input of 2.77 million tons/year since 1.8 ka with contemporary rates estimated to be one twentieth of this.

Figure 7.3: 1800 year evolution of the delta which sets the boundary conditions for the three River Styles contained within it.

Following the eruption of Lake Taupo (1.8 ka) lake levels were lowered relative to the shoreline at Turangi by 10-15 m. Vast volumes of ignimbrite were draped across the catchment up to 10 m thick. The post-eruption period could be characterised by rapid reworking and redistribution of sediment across the catchment, including the initial phase of delta creation downstream of the pre-eruption delta, which now forms the terraces. Sediment would have rapidly accumulated and built up this delta, which has prograded lakewards since then. Average rates of sediment delivery into the delta over the last 1.8 ka. are estimated to be 2.77 million tons annually. However, contemporary rates are estimated to be around a 20\textsuperscript{th} of this rate (Smart, 2005), illustrating the extremely high loadings immediately post eruption.

As the delta evolved, it has developed distinct changes in slope which control the abrupt progression of River Styles. The steeper delta (0.0026 m/m) at the upstream point is actively aggrading, capturing all the gravel- cobble fraction delivered from across the catchment. Sand sized material is easily moved through this reach and makes up the bed and most of the bars of the meandering, sand bed river downstream, as slopes decrease to 0.0016 m/m. Finally at the downstream extent, the channel forms a classic birds foot delta, which was classified as the multi-channelled delta, characterised by
especially low slopes of 0.0006 m/m. These boundary conditions control the evolutionary trajectories and transport capacity of these systems.

7.2.4 Process-based Evolutionary Trajectories for the Braided Gravel bed River
The braided, gravel bed reach has undergone high rates of adjustment during the past 80 years, especially following the increase in the base level of Lake Taupo in 1941. A Q$_{60}$ flood in 1958 augmented this response by supplying a pulse of sediment and water, and reworking the floodplain. Gravel mining from 1964 directly altered channel characteristics, forcing the river to adopt a single channel planform. Due to the short time period (23 years) between regulation of the base level and gravel mining, the extent of channel response to an increase in base level was never fully observed. In addition, two of the three largest floods in history occurred during this period, amplifying the extent of response seen. This makes it difficult to know whether the post-Lake Taupo regulation planform would have been the wider channel observed in 1958, or a narrower channel. This makes it particularly important to interpret controls on sediment transport through this reach, so that direction the channel is moving towards can be understood, following the cessation of gravel mining.

Channel evolution since 1984 records the trajectory that the braided reach has moved along following the cessation of mining in the 1980s. This records a legacy of on-going increases to braiding and the active channel area. The increased frequency of high magnitude flood events between 1993-2003, including five floods with a discharge greater than the Q$_{5}$, substantially increased braiding and active channel area, with an average total channel width increase of 103 m (Figure 7.4). Increased sediment supply from the eruption of Mt Ruapehu may have played an additional role in increasing braiding, despite the small size of sediment delivered (most < 0.5 mm) and the short residence time this material had within the channels of about 3 years (Collier, 2002; Manville et al., 1996). During this period a node of braiding re-established within this reach. As a result the 2004 flood (Q$_{60}$) only increased the average total channel width by 0.5 m, whilst increasing the average active, unvegetated gravel width by an average of 22 m. This indicates that despite significant aggradation and reworking of within channel surfaces, the channel did not respond by increasing the extent of braiding as was evident following the 1958 flood. The increasing trajectory of braiding seen between 1993 and 2003 has been halted by further gravel mining and few large flood events between 2005-2011.
Figure 7.4: A) Channel planform adjustment in the braided reach since gravel mining stopped pre-1984. B) Quantified response in active (including wetted channel and unvegetated gravel bars) and total channel width over the past 27 years.

Spatial patterns of adjustment can be attributed to the changing slopes, grain size distributions and as a result transport capacities along the braidplain (Figure 7.5). The gravel-sand transition immediately downstream of Smallmans Reach (Figure 7.5) illustrates the role of this reach to capture all gravel-cobble material that is flushed through the terraces upstream. As a result this reach is particularly sensitive to changes in the supply of this fraction. The location of the gravel-sand transition is most likely caused by the decrease in slope. As Figure 7.3 illustrates the slope within the braided reach is 0.003 compared to a much lower slope of 0.0008 for sites within the meandering reach. Figure 7.5 demonstrates the decline in slope from each site from 0.0048 down to 0.0013 at Smallmans Reach showing this relationship in greater detail. Sambrook Smith and Ferguson (1996)
describe how as slope declines, depth increases and velocity and shear stress decrease causing sand to rapidly aggrade. Similarly within the Tongariro River, reducing transport capacity along the reach was seen to cause marked downstream fining.

Slopes and as a result grain size decrease rapidly along this reach (Figure 7.5). Bridge site describes the characteristics at the most downstream point within the terraces. Steeper slopes (0.0048) and confinement drive high transport capacity (shear stress ratio of 3.59 predicted for the Q100 event). This is reflected by larger sediment size at this point ($D_{50}$ of 91 mm and $D_{95}$ of 260 mm). Directly downstream the channel widens and sediment is dumped. $D_{50}$ decreases to between 80 - 85 mm within this braidplain and $D_{95}$ decreases from 227 mm at the upstream site to 164 mm in the middle of the braided reach, down to 170 mm at the downstream extent. Despite similarities in the median grain size, the largest fraction decreases rapidly. Downstream of Bain Pool the channel adopts a single planform and sediment size fines. Smallmans Reach at the most downstream location within the historically braided reach, and immediately upstream of the meandering reach has a $D_{50}$ of 25 mm and a $D_{95}$ of 50 mm, illustrating a marked decrease in sediment size between Bain Pool and Smallmans Reach.

Transport capacities decrease downstream to reflect changes in sediment size. Bridge Pool has a much higher transport stage (ratio of bed shear to critical shear stresses) of 3.59 during the Q100 flood and is thus competent to transport the sediment present. Once the channel flows onto the unconfined floodplain this decreases rapidly. Swirl Pool has a ratio of 1.0 reflecting that the Q100 is only just competent to move the median grain size present. This decreases into the middle of the most braided locale, down to 0.71, where low confinement and slope makes the channel unable to entrain the median grain size. Slightly higher slopes at Bain Pool drive a higher shear stress ratio of 1.49. Smallmans Reach has a shear stress ratio of 1.33, whereby despite low slopes, the smaller grain size at this location is able to be entrained more frequently. This pattern highlights the very low transport capacity at the node of braiding. This provides a process-based explanation for the role of this locale in capturing sediment. As this reach aggrades, slope is expected to increase and the node of braiding move downstream, increasing the extent of braiding and accommodating more of the area occupied by the 1958 channel. However, before this can occur, slopes and transport capacities must increase at this upstream section, so that gravel-cobble sized sediment can be flushed to the downstream reaches, further from the township of Turangi. Gravel mining in this area is acting to create a flat bowl which is not able to flush sediment through to the downstream reaches.
Figure 7.5: Channel outlines in 1958 and 2007 annotated with shear stress ratios for the Q\textsuperscript{100} flood generated using the methods described in Chapter 5. D\textsubscript{50} and D\textsubscript{95} are also displayed to illustrate changes downstream to the median and largest fraction. Inset shows long profile of water surface and bed elevation.

7.2.5 Future Response Trajectories for the Braided Gravel bed River

Future trajectories of the braided reach are shown on Figure 7.6. If gravel mining has ceased and the zone where sediment has been removed aggraded, then the channel is expected to widen, obtaining a similar planform to that observed in 1958. Large floods provide a key mechanism in causing this
aggradation and channel widening. This agrees with modelling results from Smart (2011) which show high shear stress and depths during a Q50 flood in the zone occupied by the 1958 channel. Shear stress ratios and slopes in Figure 7.7 also support this. However, once the channel obtains the 1958 channel area, it is unlikely to get any larger, due to low transport capacity on the surrounding floodplain (Figure 7.7).

Volcanic eruptions are inferred to increase braiding. Sand and finer grained deposits as delivered in the 1995-1996 eruption are inferred to have a minimal, short-term influence. This is because the channel is easily able to flush this fraction, with sand transport observed during ‘normal’ flows in the flatter, smaller channel within this reach. If a lahar delivered larger gravel-cobble sized sediment, as makes up the terraces upstream, this fraction would remain within this reach. The terrace confined
reach has a high capacity to rework and transport this fraction, creating an on-going supply to the braided reach, until the supply upstream is stabilised. However, the installation of a bund in the upper catchment acts to protect the Tongariro River from future lahars from Mt Ruapehu, which has a crater lake from which lahars flow. Whilst this decreases the likelihood of this threat, it does not act to extinguish it, as a lahar could be possible from another volcanic cone. For example, Mt Ngauruhoe has been the most recent to erupt in 2012 and has multiple streams which drain into the Tongariro River.

Anthropogenic influences have had major impacts upon the character and behaviour of this reach. The increase of base level in 1941 profoundly altered the character of the reach. This created a backwater effect, which greatly increased the extent of braiding by increasing total channel width to 265 m on average (Figure 7.6). This influence has been superseded by that of gravel extraction. However, if gravel extraction is stopped, it is likely to push the channel back to the 1958 planform with the help of high magnitude floods (> Q50) (Figure 7.7).

Gravel mining provides an imposed, direct impact upon channel planform. In the past, this has acted to force the channel into a single channel, causing significant narrowing (average width decrease of 83 m). More recently, gravel mining has allowed the channel to maintain its form, with sediment extracted from the top of bar surfaces. This acts to retain the area of braiding at its location, as it increases the accommodation space and decreases slopes and transport capacities so sediment is captured. Once mining ceases, the channel is likely to widen back to the extent of the 1958 planform.

Reduction of large floods caused narrowing (average decrease of 30 m of unvegetated surfaces) and an increase in vegetation on bar surfaces within this reach. Total channel areas may also decrease if chute channels are not maintained and become overgrown, decreasing the extent to which large floodplain areas are connected to the channel.
Figure 7.7: Process relationships within the delta reaches. A) 1 m resolution DEM derived from LiDAR data. Median grain size ($D_{50}$) and Shear stress ratios for the mean annual flood are annotated (methods in Chapter 3). The likely future path of channel evolution is also annotated. B) Detrended elevation map describing the height of surfaces above, or below the water surface. C) Distribution of slope (m/m) calculated across 100 m horizontal grid cells. D) Shear stress ratio generated using slope grid for the mean annual flood assuming that each cell had a channel with the properties of Bain Pool (hydraulic radius of 1.34) with critical shear stress derived for a grain size of 85 mm (the median for the node of braiding). This exploratory technique aims to describe differences in energy across the delta, rather than explicitly model sediment transport.

The future spatial distribution of braiding is likely to follow past patterns. This is illustrated by the patterns of slope and shear stress throughout this reach (Figure 7.7). Gravel mining has created a
localised zone which generates low shear stresses and traps gravel-cobble material ($D_{50}$ around 85 mm) that has been flushed through the terraces. In the absence of this, steeper slopes would be able to form, and the extent of braided area would likely increase, particularly in response to large flood events which deliver slugs of sediment into this reach. Findings from this study indicates that the upper section directly to the north-west of the current braided area is most likely to be the locale of future adjustment (Figure 7.7A). Higher shear stress ratios (>1) and slopes (> 0.007) within this zone (Figure 7.7C, D) means flow would be able to rework this zone easily. This zone was also active in 1941, before the braidplain spread downstream as shown by the 1958 outline (Figure 7.5). The downstream section of the braided area which was active in 1958 has deep flows due to flow convergence from multiple directions, though comparatively low shear stresses due to low slope (< 0.003). However, once braiding is established in the upstream node, it is likely to inundate and rework these surfaces during large floods (as shown on Figure 7.7) though not to the intensity observed in the upstream section.

7.2.6 Process-based Evolutionary Trajectories for the Meandering, Sand bed River and Multi-channelled Delta

Due to the connected nature of the Meandering, sand bed river and the Multi-channelled delta, it is inferred that no change can occur in one, without major implications for the other. For this reason the evolutionary trajectory of these two reaches is dealt with in tandem.

Lateral adjustment has been minimal for the meandering reach over the past 80 years, despite non-cohesive banks. The floodplain shows evidence of paleochannels and a previous delta mouth, but there is no evidence that the channel freely meandered across this surface (such as ox box lakes or ridge and swale topography) (Figure 7.8). The sunken delta to the east indicates that in the past the channel flowed eastwards of the current delta (Chapter 3). Over time, high loads of sand predominantly from the Volcanic Plateau caused the channel to narrow and aggrade. This caused the channel to avulse to its present position. Higher shear stresses generated by steeper slopes of the new channel position along the delta allowed lateral migration of the channel during floods. Over time, as the channel became more sinuous, slopes decreased as did the energy to drive lateral migration. As a result, the channel began to narrow and infill, entering the phase it is in presently (Chapter 5). The more the main channel narrows and aggrades, the less shear stress can be generated during high floods as channel capacity has substantially decreased. At present, flows are greater than bankfull 2 % of the time, with the delta inundated on average 7 days a year (Smart, 2011). Lower channel capacity delivers greater volumes of water to the floodplain. Levees have formed as the channel has continued to aggrade above much of the surrounding floodplain. In other systems, once a channel reaches this point it commonly creates cut-offs, straightening the channel.
and commencing a cycle of lateral meander migration (Hooke, 2003a). However, slopes along this section of the delta were too low to generate shear stress necessary to create a cut-off. In addition, the minimal adjustment within the delta provides no evidence of changes to the location of the boundary between the meandering and delta reaches over the past 80 years.

Figure 7.8: Channel evolution of the Tongariro delta over the last ~200 years.

Overland flow during large floods has started to form a channel to the west of the current delta, creating a steeper path than that of the trunk stream (Figure 7.8). Smart (2011) found that shear stresses in the new channel were an order of magnitude larger than observed in the main channel during a $Q_{50}$ flood. Therefore, it is estimated that several more large floods and continued narrowing
will further develop this channel, driving the system closer to avulsion and the creation of a new delta (Figure 7.7).

Figure 7.7A shows downstream changes in median grain size ($D_{50}$) and shear stress for reaches within the delta. This illustrates a pattern of decreasing shear stress ratios through the braided reach (0.32 - 0.49), which increases into the meandering (1.35 - 1.10) and especially delta (5.6) reaches. This reflects the downstream decrease in grain size which increases the competence of the mean annual flood to transport the 1 mm fraction found on the delta. However, despite this, the channel at this point has been steadily narrowing, with secondary channels becoming infilled. This illustrates that channel infilling is not due to the size of the sediment, but rather the high supply, delivered across a range of flows which the channel is not able to flush. As such, these reaches are particularly sensitive to increases in this fine fraction which may act to increase the rate of narrowing.

![Figure 7.7A](image-url)

**Table 7.1: The likely response of natural and anthropogenic influences on the meandering and delta River Styles.**

<table>
<thead>
<tr>
<th>Natural and anthropogenic process</th>
<th>Likely response in the meandering and delta reaches</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Natural</strong></td>
<td></td>
</tr>
<tr>
<td>Volcanic eruptions</td>
<td>This reach is particularly sensitive to finer sediment generated from volcanic eruptions, making this a major cause of narrowing as this fraction gets trapped in the reach. This may cause the gravel-sand transition to move upstream (c.f. Knighton, 1999).</td>
</tr>
<tr>
<td>Tectonic deformation which may lead to the delta subsiding relative to the output</td>
<td>Increased narrowing as the base level increases.</td>
</tr>
<tr>
<td>Moderate flood events (RI of $&lt; Q^5$)</td>
<td>May cause some widening, as these flows are competent to transport available sediment sizes. These flows do not greatly increase the sediment loads as observed during larger flood events due to less reworking in upstream reaches than would be carried out by a larger flood event (e.g. $&gt; Q^{10}$).</td>
</tr>
<tr>
<td>Large flood events (RI of $\sim Q^{100}$)</td>
<td>Large floods can increase sediment supply and may increase narrowing (seen in the 2004 and 1958 floods) as an increase in sand fraction is flushed from upstream as bars are reworked. These floods also increase scour in the secondary channel which has higher velocities than the current channel. This furthers headcut migration of this new channel.</td>
</tr>
<tr>
<td><strong>Anthropogenic</strong></td>
<td></td>
</tr>
<tr>
<td>Increase of base level of Lake Taupo</td>
<td>Increases the rate of narrowing as the volumes of sediment flushed into the lake are decreased.</td>
</tr>
<tr>
<td>Gravel mining upstream</td>
<td>Releases sand which underlies the armoured layer in the braided reach. This could increase sediment supply, and increase the rate of narrowing.</td>
</tr>
<tr>
<td>Clearing willows from the lower reaches</td>
<td>Bank erosion has been observed in response to clearance and hence may cause localised widening. This decreases bank roughness and cohesivity, though this has only been seen to have a minimal impact. The response would likely be greater in response to a large flood ($&gt; Q^{10}$).</td>
</tr>
</tbody>
</table>

Response trajectories are not created for this channel as the exact timing of avulsion is difficult to estimate. Instead past adjustment can be used to interpret whether different natural and
anthropogenic influences are likely to either push the channel closer to avulsion or slow the process (Table 7.1). This illustrates that most influences increase the speed of narrowing, especially increased sediment supply from gravel mining upstream, large flood events and volcanic eruptions, whilst the base level of Lake Taupo may enhance narrowing by causing a backwater effect. The channel needs to avulse in order to continue to transport its load into the lake.

### 7.3 DISCUSSION:

Marked variability is evident in the inferred evolutionary trajectory for different reaches of the lower Tongariro River. Commonly, channel change and adjustments are driven by impacts propagating through the system, such as sediment slugs or knick points (Bartley and Rutherfurd, 2005; Crosby and Whipple, 2006; Desloges and Church, 1989; Simon and Rinaldi, 2006). However, within the Tongariro, the marked differences in controls, particularly slope (and thus sediment transport capacity), created varying responses for each River Style to different pressures. This is because each pressure either influenced a distinct spatial locale (i.e. gravel mining in the braided reach) or altered the transport of a specific grain size fraction (e.g. volcanic eruptions increased the volume of sand delivered to the delta). As such, the wandering cobble bed reach was sensitive to pressures that altered the entrainment of the cobble-boulder bed material, the braided reach to changes in the transport of the gravel fraction and the meandering reach to changes in sand transport and delivery loads. This means that an influence may drive dramatic change in one River Style, whilst the neighbouring river type remains relatively unaffected. For example, the increase in base level drove dramatic channel widening in the braided reach, while the terrace confined reach was unaffected due to steep slopes and upstream position reducing the backwater effect this far upstream. In contrast, the fine grained sediment delivered from volcanic eruption is easily flushed through terrace confined reaches, but it is captured in the meandering sand and delta reaches, causing narrowing. This highlights the importance of viewing reaches with regard to their sediment transport capacity, and internal characteristics that determine the extent of response. In a sense, variable patterns of process connectivity are evident along the lower course of the Tongariro River (c.f. Hooke, 2003b; Jain and Tandon, 2010).

The history of adjustment showed that boundaries between the River Styles have remained the same over the past 80 years. However, evolution across longer timeframes, such as the progradation of the delta during the past 1800 years, has caused rapid evolution and adjustment of these boundaries. The terrace extent provides a fixed endpoint for the Wandering, cobble bed river. It is considered that this reach is unlikely to undergo significant adjustment over the next 1000 years (unless there is another volcanic eruption). However, River Styles draining the delta are more likely to change. Perhaps the most dynamic boundary is the gravel-sand transition, which demarcates the
border between the braided gravel and meandering sand River Styles. As the delta evolves, underlying controls continue to adjust. An increase in sand would shift the boundary upstream (Knighton, 1999), whereas, an increase in slope (as may occur from aggradation) can increase bed shear and shift the margin downstream (Sambrook Smith and Ferguson, 1996). Wholesale avulsion, as is likely, would cause the reach to re-establish these River Style zones, driven by local scale differences in underlying controls (i.e. slope) and sediment transport processes.

Differences in River Style specific responses can also be attributed to the magnitude of geomorphically effective flows for any given reach. Reaches within the terraces need large floods (> Q^{20}) to rework the cobble-boulder sized sediment and rework bars and islands to drive planform change. Within the Braided reach, frequent floods (Q^{1.33}) are able to rework 44% of bar surfaces, mobilising and reworking a greater proportion of bars within the braidplain. Even larger floods (e.g. Q^{50}) are more geomorphically effective as they rework floodplain areas that were occupied by the channel in 1958. In addition, these floods deliver high loads of sediment to this reach, causing substantial aggradation (95,000 tonnes in 2004) (Smart, 2005). However, even the largest floods are not able to entrain full bar surfaces, as this section traps and retains all of the gravel-cobble material, meaning that slopes are too low to move this fraction further down the delta. Lastly, large floods (Q^{50}) in the meandering and delta reaches do not have greater transport capacities within the channels than the frequent bankfull flood which occurs around 7 days a year (Smart, 2011). As such, the main channel has become hydrologically inefficient. Instead these larger events rework the floodplain, furthering the establishment of a secondary channel which will eventually replace the present channel.

The future response trajectories are presented as likely future adjustments to fluxes. It is understood that rivers and their responses to changes in sediment and water fluxes, are complex phenomenon (Chapter 8). Influences rarely occur in isolation. Instead one influence (i.e. a large flood) can amplify the channel response to another control (i.e. base level). Commonly multiple, overlapping influences interact to drive complex patterns of response (Phillips, 1999). In addition, the state of the system at a given time will influence how it responds to the same event (Lane and Richards, 1997). This is particularly true if the system is close a threshold which may cause a change in state or amplify the response (Schumm, 1980). Event sequencing or contingency is also key, as previous flood events may act to decrease or increase sediment supply, decreasing or increasing the geomorphic effectiveness of a following flood event (Marutani et al., 1999; Phillips, 2012). For example, large floods in the Tongariro remove vegetation off bar surfaces and rework bed material, increasing the capacity of subsequent floods to elicit change (i.e. recovery time is a critical control). Predicting river
futures is complex, as we can never accurately predict the future frequency and magnitude of controls. Future response trajectories provide a technique to infer how the river might respond to a single control. As this is based on past adjustment, it includes consideration of thresholds. In addition, this study grounds estimates of future change with the quantification of sediment transport and channel change processes across multiple scales. By building a process-base to build into the conceptual evolutionary trajectory model, this study provides a more comprehensive basis for estimating change to a range of influences. This acknowledges the complexity of the system by estimating how a reach may respond to an event, as opposed to how it is likely to evolve. This provides a guiding image to underpin how these rivers may adjust in the future based on a range of scenarios, rather than trying to explicitly state what a system will look like. As such, this incorporates uncertainty into the assessment which is essential given the complex nature of these systems.

This form of analysis is strengthened in its ability to consider multiple influences. Modelling has a common focus on flood events or deriving how a single influence (often a dam or form of regulation) may influence the transport capacity of this single driver. However, rivers are influenced by multiple natural and anthropogenic drivers. Each driver may be characterised by a different frequency/magnitude of influence (e.g. floods occur more often than lahar flows). In addition this allows identification of feedback loops and thresholds that may be present within a system. Likely responses combined with acknowledgement of how frequently they occur, provide an underpinning to estimate the likelihood and direction of change of the future trajectory of the river.

The strengths and weakness of process-based evolutionary trajectories as an approach to estimating river futures is discussed fully within the discussion chapter (Chapter 8).
8 DISCUSSION

This chapter discusses key findings of the thesis, framing these contributions in relation to the international literature. The four fundamental research questions presented in Chapter 1 are addressed, summarising findings at the catchment, reach and bar scales. This is followed by discussion of how the evolutionary trajectory approach was strengthened by incorporating a process-based underpinning. Finally, implications for management of the lower Tongariro River are considered, based on insights from this thesis.

8.1 KEY FINDINGS

This thesis demonstrates the importance of a multi-scalar approach to fluvial investigations. The scale of a study is inherently linked to the phenomena observed, and the type of enquiry (large scale qualitative vs small scale quantitative). For this reason, using multiple scales and lines of evidence provides a more comprehensive view of the system, characterised by less bias due to both scale dependency and methodology (Church, 2007; Small and Doyle, 2012). A catchment framing is necessary to determine what landscape scale processes are driving change within individual reaches. This allows the nature and type of sediment delivery into a specific reach to be determined. Considering long-term landscape evolution illustrates how these sources have changed over time. This allows differentiation between landscape units that supply on-going sediment loads and those that have fluctuating loads which are generated from extreme events (i.e. volcanic eruptions) and generally decrease over time. Channel change should be assessed at the reach scale, as rates and types of adjustment can be readily observed. The sensitivity in spatial patterns can be related back to differences in internal controls and evolutionary history, which drives the nature of channel response. Management plans can be designed at this scale to ‘work with’ underlying sediment transport processes. More detailed, local (bar) scale insights are necessary to provide a process-based understanding to underpin river morphodynamics observed at the reach scale. Within this study, this scale of analysis was used to measure the grain size distributions for individual reaches, so the magnitude of flood events necessary to rework these surfaces could be assessed. This provided insights into channel response to the pattern of flood events and explained variation in adjustment between reaches.

This analysis was underpinned by the use of form across multiple scales to infer process. Measuring process has been an on-going challenge within geomorphic research. However, technological advances have dramatically altered the scale and resolution with which we can measure channel form. This work extends the ‘morphological approach’ from a reach scale, direct measurement of
sediment erosion and deposition to a larger range of scales and process measurements. For example, bulk sediment volumes in the delta were used to measure average annual rates of sediment delivery, channel cross-section and slope indicates shear stress, and high resolution surveys of bar topography were used to map grain size and model sediment entrainment. This illustrates the increase in scope that technological advances have provided for scales at which the process-form approach may be applied. To fully utilise these advances, the morphological approach should be broadened to include physics-derived measures of energy (shear stress, stream power) rather than limited to direct measures of sediment transport. Energy measures can be combined with past patterns of channel adjustment to quantify why differences in adjustment may be observed. For example, within the Tongariro River, zones with lower transport capacities stored more material and thus were more responsive to large floods, providing a direct link from process to form.

The datasets and approaches described above allowed the creation of a catchment-framed, process-based evolutionary trajectory approach. Prediction of future trajectories should be based on past patterns of channel adjustment, incorporating notions of thresholds, event contingency and non-linear dynamics into systems response. In order to present likely future pathways of adjustment, analysis needs to consider not how the river has adjusted, but why did it adjust, and what controls were key in driving the adjustment in a particular reach. This is implicitly underpinned by knowledge of the processes driving channel morphodynamics, which is grounded through combining the multi-scalar, form-process analysis described above. These analyses provide an insightful basis with which to consider river management and rehabilitation schemes.

### 8.2 Fundamental Research Questions

Within this section each fundamental research question is answered with regard to findings from the thesis.

#### 8.2.1 Why do larger-scale spatial and longer-term temporal patterns of sediment delivery and connectivity across the Tongariro catchment provide an important framing for the adjustment of the lower Tongariro River?

Many studies have advocated the need for an appreciation of sediment transport at the catchment scale to underpin analysis of reach scale adjustment, and associated implications for river management (Benda et al., 2011; Brierley and Fryirs, 2009; Fryirs et al., 2007b; Kondolf et al., 2006; Lane and Richards, 1997; Parker, 2010; Raven et al., 2010; Sear et al., 1995; Sear et al., 2009; Walling and Collins, 2008; Wohl et al., 2005). This section discusses how this thesis built upon these approaches and the insights that were derived from their application in the Tongariro catchment.
Catchments have traditionally been viewed as systems which undergo linear downstream progressions in slope, sediment storage and bed material as they progress from source through transfer to accumulation zones (Church, 2002; Schumm, 1977; Vannote, 1980). However, the Tongariro displays a more complex pattern in sediment process zones, with vast sediment stores in the headwaters of the Volcanic Plateau effectively disconnecting sediment transfer from the volcanic cones to Lake Taupo. This supports previous work by Phillips (1999; 2003a) whereby the distribution of sediment storage within catchments is non-linear, patchy and discontinuous. This highlights the importance of developing conceptual models which detail catchment-specific configuration of sediment process zones.

The classification of ‘landscape units’ provided a coherent structure upon which to understand sediment transfer and dominant geomorphic processes (Brierley and Fryirs, 2005; Montgomery, 1999). This supports previous work which argues that sediment transfer through landscape components needs to be assessed before patterns across the whole catchment can be ascertained (Brierley and Fryirs, 2005; Fryirs et al., 2007a; Jain and Tandon, 2010; Montgomery, 1999; Polvi et al., 2011; Wohl, 2010). The high diversity of landscape units within the Tongariro catchment (i.e. uplifting greywacke ranges adjacent to an active volcanic field) illustrates the importance of combining insights into landscape history with present distributions of geology, topography and land use to interpret contemporary sediment fluxes.

Whilst ‘landscape units’ and ‘process domains’ have been used before, applications have been largely descriptive, based on qualitatively separating zones with different landscape characteristics and processes. Wohl (2010) presents process domains as a first-level delineation of sediment dynamics which can be used to underpin and explain more complex mathematical models. Brierley and Fryirs (2005) use landscape units to contextualise and classification river type by including variables such as geology, topography and valley confinement. However, these previous approaches do not incorporate quantification of sediment transport into these conceptual models, despite advocating for its importance (c.f. Small and Doyle, 2012). In contrast, this thesis combined qualitative historical insights, including landscape unit evolution, with quantification of characteristics (i.e. bed material and slope) and processes such as stream power and erosion indexes, which characterise sediment generation and transfer. This parsimonious approach presents a simple framework which can be easily applied in other catchments to understand sediment transfer.

This thesis highlighted the importance of assessing the (dis)connectivity of sediment stores. Within previous studies, sediment storage zones are assumed to disconnect sediment propagation through
a landscape (Fryirs et al., 2007a; Fryirs, 2012; Métivier and Gaudemer, 1999; Phillips, 2003b; Phillips and Slattery, 2006). However, within the Tongariro catchment, some sediment storage zones were disconnected from the active channels, limiting the ability of the reach to buffer excess sediment delivered, improving connectivity. For example, within the terrace confined reach, incision has disconnected the channel from the floodplain. As a result, the longitudinal connectivity of sediment through the reach has been increased, as minimal accommodation space was available between the terraces to store sediment. In contrast, active sediment stores were located at the boundary between the steep volcanic cones and Volcanic Plateau. Low longitudinal slope captures much of the sediment delivered from the cones and the low gradient channel limits transport of all but the finest sand sized sediment across the Volcanic Plateau. As such, assessing the connectivity of sediment stores is a key underpinning to ascertaining their role within the system.

Existing conceptual models of landform development characterise catchments as either well-connected steady states, which are graded to deliver sediment from the headwaters to the mouth (Adams, 1980; Davies and Korup, 2010; Reneau and Dietrich, 1991; Whipple and Tucker, 2002) or disconnected systems where storage elements act to buffer sediment delivery, effectively decoupling source and accumulation zones (Dearing and Jones, 2003; Métivier and Gaudemer, 1999; Phillips, 2003b; Trimble, 2009). The Tongariro is a system whereby half acts as a steady state system (the Kaimanawa Ranges) and half as a disconnected system (due to storage within the Volcanic Plateau). This finding is key to explaining the type, volume and size of sediment delivered to the lower catchment. Uplift of the highly connected Kaimanawa Ranges creates an on-going source of gravel to the lower catchment. Its small size (60-65 mm) is readily flushed over the boulder lag (91-230 mm) derived from terrace incision. As such, these materials exert a minimal role upon adjustments within the terrace confined reach. However, lower slopes in the braided reach capture this fraction, driving high sensitivity and channel change in this area (Chapter 4). In contrast, the disconnected nature of the Volcanic Plateau limits the role of this highly productive sediment source in supplying material to the lower Tongariro River, making this source a lesser supply of material which is large enough to drive planform change (i.e. greater than coarse gravel). However, lahar flows are competent to deliver pulses of coarse sediment into the terrace confined reach. As a result, sediment supply from the volcanic sub-catchment is driven by high magnitude events that shape the system, with long relaxation times following. These conceptual models were key for understanding the sources, grain size and volume of sediment moving through catchment components. Channel response in the lower catchment is driven by these catchment scale patterns of sediment transfer.
Consideration of landscape evolution is essential in unravelling controls upon these relationships. Brierley (2010) discusses the need to consider how past geological, climatic and anthropogenic influences have left imprints on the system. These determine the contemporary distribution of sediment stores and geomorphic processes. Lane and Richards (1997) suggest that processes measured over small spatial and temporal scales are not sufficient to understand system behaviour. Instead these need to be contextualised within “an understanding of the sequence of configurational states through which the system evolves” (Lane and Richards, 1997: 249). These notions were found to be key within the Tongariro system. The Taupo eruption (1.8 ka) acted to reset boundary conditions and redistribute the process zones across much of the catchment. This realignment or reframing of the system underpins contemporary controls on channel adjustment. The most dramatic example of this is the incision of the terraces and the progradation of the delta immediately downstream. This has created an abrupt decrease in slope immediately downstream of the terraces which captures the gravel fraction, creating the highly sensitive braided gravel bed river. Understanding how these zones have evolved is essential for predicting future trajectories of river adjustment.

To quantify landscape scale morphological evolution, a novel approach to creating a sediment budget was developed. This builds upon ‘morphological’ approaches at the reach scale, whereby changes to channel form are used to infer sediment erosion and deposition over time (Ashmore and Church, 1998; Brasington et al., 2000; Fuller et al., 2003a; Fuller and Basher, 2012; Ham and Church, 2000). Applications of a morphological approach at the landscape scale are scarce. Examples include Buijsman et al. (2003) who constructed past and present DEMs to calculate sediment deposition in the mouth of the Columbia River since 1900 and Houben et al. (2006) who used DEMs to calculate Holocene floodplain deposition in the Rhine over the past 7500 years. These applications describe changes within a single unit, and are not able to characterise sediment transfer across the catchment as a whole, creating a closed sediment budget. In fact, examples of applications of sediment budgets which consider sediment routing and storage across entire drainage networks and longer spatial scales are scarce (Goodbred and Kuehl, 1999; Walling, 1983). Application of the ‘morphological’ approach in this study demonstrated its strength for developing catchment-scale sediment budgets.

Development of sediment budgets are commonly constrained by a lack of continuous datum (Houben et al., 2006). However, the tephra from the Taupo eruption provided a distinct point in time from which catchment evolution could be measured. Within volcanic regions, this presents an under
used technique for creating sediment budgets and quantifying sediment deposition at the catchment scale.

This sediment budget illustrated the bulk movement of sediment across the Tongariro catchment over the past 1800 years. The Kaimanawa Ranges provide the major source of sediment (55%). It is inferred that these materials have consistently been flushed through the mid-catchment to the delta. Terrace incision of the Tongariro River provides a minor source of sediment (4%). It is likely that this source has diminished over time as the boulder lag and gentler slopes slow the rate of incision. The Volcanic Plateau contributes 41% of sediment through high magnitude events which drive pulses of material through the system. This can be contrasted with patterns of sediment delivery in other catchments. During periods of high sediment delivery, storage was found to play a key role in reducing sediment output. Trimble’s (2009) work in Coon Creek described the storage in the upper catchment of sediment yields generated during early agricultural practice, resulting in minimal changes to sediment delivery over time. The Tongariro catchment displays a similar pattern with the storage from Volcanic Plateau (estimated at 110 km$^3$) reducing loads delivered to the lower catchment (Hackett and Houghton, 1989). High sediment yields delivered from lahar flows have swamped existing landforms, creating low slopes and inhibiting further reworking and transfer of this material. In contrast, sediment stores in other landscape settings have contributed on-going high yields, such as seen in the reworking of para-glacial stores in Canada (Church and Slaymaker, 1989). The low slope and low drainage areas created by parallel drainage pattern (valley margins are not present to create dendritic drainage networks) across the Volcanic Plateau limit stream power, reducing the ability of the channel to rework these stores.

Combining the sediment budget with the conceptual framework of connectivity and process zones acted to offer insights into the ‘black box’ which is catchment scale sediment transport (De Vente et al., 2007; Fryirs, 2012; Walling, 1983). Volumes of material generated from sediment sources and deposited in sediment sinks were described. Beyond this, the ease with which material was able to be transferred between these elements was appraised. This approach combined spatial (catchment scale connectivity analysis) and temporal (sediment budget) analysis of sediment transfer, creating a dynamic platform with which to contextualise reach scale adjustment.

The strength of this approach can be seen in its ability to combine multiple tools which describe sediment transfer at the catchment scale. This approach is supported by Small and Doyle (2012) who argue that both conceptual and process-based numerical models are essential to gain system-wide understanding. The conceptual framework adopted in this thesis separated the catchment into landscape units, characterised by distinct suites of geomorphic processes. These processes were
Chapter 8: Discussion

described through quantification of sediment transport by calculating stream power, slope
categories and erosion indexes. This work was then tested in the field, through analysis of thirty
representative sites which analysed channel characteristics, bed material, shear stress and the
capacity of the reach to transport the available bed material. This grounded findings from the
remote sensing analysis, linking stream power to the specific character and behaviour of a reach.
Lastly, the sediment budget added an explicit temporal component to the framework, describing
how sediment has been redistributed across the catchment over the past 1800 years.

Small and Doyle (2012) suggest that geomorphologists have developed an armoury of tools, but it is
the combination of multiple tools and approaches which offers the most appropriate technique to
create a system specific understanding of sediment flux. This work developed a comprehensive,
system-wide understanding of sediment flux with which to contextualise and explain geomorphic
processes and sediment transport at the reach scale (see also Benda et al., 2011; Brierley and Fryirs,
2009; Kondolf et al., 2006; Raven et al., 2010).

8.2.2 Why does reach scale response differ along the lower Tongariro River and what
key controls drive variability in sensitivity?
The eruption of Lake Taupo (1.8 ka) set the Tongariro River along a new trajectory of adjustment,
resetting boundary conditions. This geological control has been the key determinant of the high
diversity of river type and channel adjustment along the Lower Tongariro River. Terrace confinement
was a key control upon the sensitivity of channel responses to recent flood events. Valley
confinement has long been regarded as a control upon fluvial dynamics. This study built upon
existing knowledge regarding the control of partly confined valleys on underlying processes. Fryirs
and Brierley (2010) discuss the role of antecedent controls within partly confined valleys. They found
that differences in geomorphic memory (i.e. geology) determine the characteristics of valley
confinement and floodplain pockets, including width and alignment. Their landscape unit scale of
enquiry aimed to explain why floodplain pockets have evolved to obtain certain characteristics. In
contrast, this study provides a more direct and local reach-scale example of how controls pre-
eruption have driven variations in the formation of the terraces. This was seen in a positive
correlation between valley slope pre-Taupo eruption (1.8 ka) and the width of the valley between
the terraces ($R^2$ value of 0.87). This relationship indicates that steeper reaches (with greater energy
during floods) had an increased transport capacity, enabling the removal of a greater volume of
sediment, and creating wider floodplain pockets within the terraces. This illustrates how past
controls drive contemporary system evolution.
Findings from Chapter 4 showed that floodplain pocket width provided the key control on the sensitivity of channel response. This was captured in the strong positive correlation ($R^2$ value of 0.93) between floodplain width within the terraces and the standard deviation of average channel width describing the range of adjustment over time. Previous studies have highlighted valley confinement as a key control on channel characteristics. For example, White et al. (2010) found that valley width within a fully confined valley acted to fix the location of riffles and pools despite high on-going incision due to regulation. Cowie and Brierley (2008) demonstrated the influence of valley confinement on downstream fining of grain size, with the confined reach exhibited greater fining than the unconfined reach. However, these studies describe systems which are fully confined by the valley (or terrace) margin, forcing the channel width. In comparison, the lower Tongariro River flows within a partly confined valley where there is sufficient accommodation space for the channel to create a wider planform. However, floodplain width within the terraces has an antecedent control on the channel alignment and sinuosity that is able to form within the terrace margins. Reaches within narrower valleys adopt a straighter planform, as lateral migration is limited, decreasing the ability of the channel to adopt a more sinuous form and dissipate excess energy. Reaches characterised by greater sinuosity were observed to undergo a greater degree of floodplain reworking. The helical nature of flow, concentrates energy on the outside of bends, (Bridge, 2003; Knighton, 1998), providing a mechanistic-based explanation for the increased lateral adjustment and floodplain reworking observed at more sinuous reaches (see also Hoyle et al., 2008). Straighter reaches would be expected to have energy spread more evenly across the channel bed and banks, decreasing bank migration and lateral erosion.

Narrow valley confinement impeded the ability of the channel to adjust to a form that balances sediment load and channel slope. Reaches located in wider floodplain pockets exhibited characteristics of a graded system, whereby slope and grain size were in sync (c.f. Mackin, 1948). This was demonstrated by a strong linear correlation for sedimentation zones between reach slope and median grain size which generated an $R^2$ value of 0.97. In comparison, the correlation for single-channelled reaches generated high scatter, created a lower $R^2$ value of 0.31. In addition, sedimentation zones were characterised by lower average excess shear stress during the $Q_{100}$ flood, with values of 1.75 compared to 2.89 for single-channelled reaches (calculated by ratio of shear stress to critical shear stress). This indicates that sedimentation zones adjust their form to increase resistance and decrease excess energy. Previous work on channel form has discussed how channels continually adjust their form, particularly sinuosity (slope) and braiding (roughness and number of active channels) to balance prevailing sediment load and energy-flow conditions (Church, 2006; Ferguson, 1981; Lane, 1955; Nanson and Huang, 2008). However, partly confined valleys represent a
marked variability in accommodation space and therefore the capacity for geomorphic adjustment (Fryirs and Brierley, 2010). As such, the lower Tongariro River switches between being a fully alluvial and non-alluvial system multiple times within a short reach. This creates marked differences in transport capacity. This study demonstrates the role of valley confinement in controlling sediment transport processes which underpin channel form and adjustment.

Terrace incision and the progradation of the delta following the Taupo eruption (1.8 ka) have created pronounced changes to boundary conditions along the lower Tongariro River. Wide variation in slope and valley confinement has created four distinct types of river within the 15 km study reach of the lower Tongariro River. Each River Style is characterised by different rates and types of response to a range of natural and anthropogenic pressures. Channel response is commonly characterised through the adjustment observed in response to a driving influence or control. Examples of studies include channel response to land clearance and urbanisation (Gregory, 2006), vegetation removal (Brooks and Brierley, 2004), vegetation recovery (Zanoni et al., 2008), dam construction (Kondolf, 1997; Williams and Wolman, 1984), gravel mining (Surian, 1999; Wishart et al., 2008), large flood events (Kasai et al., 2004; Marutani et al., 1999) and volcanic eruptions (Gran and Montgomery, 2005; Simon, 1992). Other forms of response illustrate the propagation of change through a system, such as that caused by sediment slugs (Bartley and Rutherfurd, 2005; James, 2010), or knick-point migration (Crosby and Whipple, 2006; Simon and Rinaldi, 2006). However, within river systems, channel responses to a given event are rarely linear and predictable, due to the multiplicity of variables affecting the system (Schumm, 1991). This necessitates the need for more complex, process-based insights to guide interpretations of channel response.

Within the lower Tongariro River rapid change in boundary conditions (decrease in slope) creates disconnected patterns of river response for each type of river. The bed of each River Style is characterised by a different grain size and therefore sensitive to different natural and anthropogenic pressures which alter the transport of each grain size fraction (Figure 8.1). For example, the terrace confined, wandering, cobble bed river requires large flood events ($Q_{50}$) to rework the bars and drive planform adjustment (Figure 8.1). This reach is also resilient to increased loads of fine-grained sediment, such as pulses of sand and ash from volcanic eruptions which move rapidly through the system. In contrast, the meandering and delta River Styles are sensitive to controls which influence the transport of this finer fraction through the system. For example, volcanic eruptions increased the rate of channel narrowing, whilst having negligible influence upon the upstream wandering reach (Figure 8.1). Increases in the loads of fine grained sediment (i.e. volcanic ash/sand) can be trapped and stored by exotic willow trees which colonise the bank margins, enabling willows to encroach
further into the channel margins. Ascertaining which grain size drives the greatest level of response for each River Style can be related to an understanding of how different natural and anthropogenic controls can alter the transport of this fraction. This is key to identifying the types and locations of adjustment that are likely to occur.

The various insights into variability in sediment flux for differing reach types were used to create a novel framework with which to analyse and interpret sensitivity. The sensitivity concept has been well established within geomorphology since the work of Brunsden and Thorne (1979) who discussed landscape evolution with regards to the response time of the system to events. Subsequent studies presented sensitivity as a descriptive measure of landscape (Crozier et al., 1990; Harvey, 2001) and then more recently channel response (Brierley and Fryirs, 2009; Downs and Gregory, 1993; Fryirs et al., 2009; Hooke, 2003b). However, no previous studies provide frameworks or techniques for assessing and classifying river sensitivity, or using it to underpin management applications.

Figure 8.1: Summary of the key grain sizes that drive channel adjustment for each River Style, and the key controls which were seen to be responsible for altering the fluxes of each grain size fraction.

The framework designed within this study was strengthened by contextualising channel response within the ‘natural capacity for adjustment’ for a particular River Style, by assessing the “… range of process activity that is possible for that setting” (Brierley and Fryirs, 2005: 144)(see Table 8.1). For
example, the unconfined, braided gravel bed river was ranked as ‘high’ sensitivity as the unconfined floodplain allows unimpeded lateral adjustment. In contrast, the partly confined, wandering cobble bed river was classified as moderate sensitivity, as terrace margins set a maximum capacity of adjustment for reach this River Style (Table 8.1). The sensitivity of specific reaches was then ranked based on adjustment observed during the past 80 years, contextualised within the natural range of adjustment that for a specific River Style. This set a process-based context for the assessment of sensitivity for individual reaches. Marked variability in channel response was observed in the partly confined, wandering cobble bed river. Sensitivity was found to be accurately expressed using the standard deviation of channel width. Values between 13-22 m were found for reaches of low sensitivity, 28 and 29 indicated moderate sensitivity and a value of 61 was obtained for a highly sensitive reach. Channel response to large floods is driven by floodplain and bar reworking, increasing channel width and braiding. As such, this measure analysed the mechanisms of adjustment and sensitivity, as the variability in width describes the magnitude of channel response.

Trajectories of adjustment were used to estimate the future sensitivity of the system. The delta and meandering reaches were classified as sensitive despite undergoing minimal lateral adjustment over the past 80 years. Channel narrowing is moving the system closer to a threshold where avulsion is necessary to maintain transport to the outlet, indicating a high likelihood of channel change in the future. This incorporates latent forms of adjustment into this framework, so that sensitivity rankings consider both contemporary adjustment and likely future rates and types of adjustment, based on consideration of longer-term process dynamics.

The creation of an applied tool with which to classify and communicate sensitivity provides an important framework for management applications that act to accommodate the natural functioning and adjustment of the system. This includes ‘room to move’ or ‘erodible corridor’ approaches which are discussed in more detail in Section 8.3 (Piégay et al., 2005; Rapp and Abbe, 2003).
Table 8.1: Overview of the forms of adjustment, how frequently it occurs, the process-based explanation and sensitivity ranking for each River Style.

<table>
<thead>
<tr>
<th>River Style</th>
<th>Mechanisms of adjustment</th>
<th>Frequency/magnitude of adjustment (sensitivity)</th>
<th>Process-based explanation</th>
<th>Sensitivity</th>
</tr>
</thead>
<tbody>
<tr>
<td>Partly confined, wandering cobble bed river</td>
<td>Chute cutoffs, mid-channel bars aggrading to form bank attached bars, floodplain stripping and reworking to form lateral bars.</td>
<td>The $Q^{C}$ flood reworks bar and island surfaces, causing the channel to adjust within its macro-channel. The $Q^{B}$ flood elicits greater change, stripping vegetation, activating chute channels and reworking the floodplain, causing an increase in active bar surfaces. During periods with minimal floods (&lt; $Q^{L}$) channel narrowing and revegetation of bar and island surfaces occurs. Within reach magnitude of change is dependent on internal reach scale controls (i.e. wide floodplain pocket and low transport capacity creates sedimentation zones).</td>
<td>Change is driven by large floods which inundate and rework the cobble-boulder material that makes up floodplain, bar and island surfaces and removes vegetation. Upstream sediment supply is not a primary cause of adjustment within this River Style.</td>
<td>M</td>
</tr>
<tr>
<td>Unconfined, braided gravel bed river</td>
<td>Floods deliver slugs of sediment which are stored as mid-channel and lateral bars. Channels adjust their alignment and may erode banks as bars extend and grow.</td>
<td>Adjustment occurs &gt; yearly with bars and channels readjusting after each flood event. Large floods (e.g. &gt; $Q^{10}$) increase the volumes of sediment delivered. The $Q^{60}$ flood has been observed to deliver large pulses of sediment and inundate and rework large areas of floodplain, substantially increasing the magnitude of adjustment.</td>
<td>Channel adjustment is driven by the volume of gravel sized sediment delivered to the reach. Large floods flush sediment through the terraces to this reach. Due to the low slope of the floodplain, large floods can also reactivate vegetated areas of braid plain, delivering sediment and removing vegetation, increasing the area of the active channel.</td>
<td>H</td>
</tr>
<tr>
<td>Unconfined, meandering sand bed river</td>
<td>The channel has exhibited on-going narrowing during the past 80 years. However, in the past the channel meandered laterally eroding outer banks and depositing on the point bars.</td>
<td>Minimal adjustment over the past 80 years with a steady rate of narrowing observed (average of 0.8 my$^{-1}$).</td>
<td>Following the last avulsion, the channel meandered until slopes were not sufficient to drive lateral erosion. Low slopes cause in-channel deposition and narrowing, decreasing channel capacity and causing further deposition, until the channel infills to the point where avulsion occurs.</td>
<td>M-T</td>
</tr>
<tr>
<td>Unconfined, multi-channelled delta</td>
<td>The channel has undergone narrowing with secondary channels infilled, islands growing and becoming bank attached while the delta has prograded out into Lake Taupo.</td>
<td>Less narrowing is observed with an overall decrease in channel width of 21 m over the past 80 years. However, channel avulsion is expected which would cause a greater magnitude of channel with the river having to prograde and reform this delta unit.</td>
<td>This River Style slowly narrows until stream power becomes too low to transport its load, blocking the outlet and avulsion occurs. This resets the river and delta formation recommences through deposition and progradation.</td>
<td>M-T</td>
</tr>
</tbody>
</table>
8.2.3 Why are the patterns and frequencies of bar reworking along the lower Tongariro River an important underpinning to channel adjustment at the reach scale? How has this approach furthered the ways in which we quantify local scale processes?

Bars have long been considered as important elements for interpreting the behaviour of river systems (Bluck, 1979; Bridge, 2003; Church and Rice, 2009; Rice et al., 2009). This is largely because “...description of these features and consideration of their origin is tantamount to describing the form and enquiring into the stability of the entire river channel” (Church and Jones, 1982: 292).

Understanding how these surfaces are reworked provides an underpinning for channel adjustment. However, up until now, there has been a dearth of information regarding detailed sedimentology and the reworking of these features. This is largely due to the difficulties of capturing grain size data (Verdú et al., 2005). Chapter 6 presents a novel approach which advances modelling of bar reworking, and provides a meaningful process-based underpinning for understanding channel change within the Tongariro River. The following section describes the development of the technique, insights about bar reworking derived from the Tongariro system and finally discusses the role of this dataset in providing a meaningful measure of process to underpin and compliment adjustment at the reach scale.

The method used within Chapter 6 extended the resolution and scale with which the sedimentology and topography of bar surfaces could be captured. Traditionally, grain size has been derived by manual Wolman transect counts or bulk sieving (e.g. Brierley and Hickin, 1985; Hoyle et al., 2007; Kondolf et al., 2003a; Powell, 1998; Rice and Church, 2010; Wolman, 1954). These techniques are time consuming and tend to characterise whole surfaces or units through a single distribution, rather than capturing the variation over space (Church and Jones, 1982; Hoyle et al., 2007; Rice and Church, 2010). Rapid developments in technology have exponentially increased the scale with which bar surfaces can be captured, especially the use of laser, particularly Terrestrial Laser Scanning (TLS), to create high resolution surveys of bar sedimentology. TLS has recently been used to measure grain size, though as yet, these applications tend to be limited in space to discrete 1 m² patches (Heritage and Milan, 2009; Hodge et al., 2009). Rychkov et al. (2012) extended this scale to a larger 50 x 100 m patch of a braided channel. This thesis extended this scale again, by characterising grain size across four bars, and using this to compare how grain size varies downstream within the Tongariro system. Moreover, this supported the use of TLS to characterise larger grain sizes than previously documented. For example Rychkov et al. (2012) used TLS to map grains up to 120 mm whilst this work was able to include grains up to 250 mm and used data derived from four sites within a 10 km reach of river. As such, this work furthered the use of TLS to map grain size distribution, both with
regards to the scale mapped and the range of grain size captured. This dataset provided a foundation for the modelling of bar reworking.

Understanding the magnitude of flood necessary to rework a bar surface underpins notions of which floods are geomorphically effective to rework the channel as a whole. Previously, these types of questions have been assessed through use of tracers or the reworking of painted gravel patches (c.f. Brummer and Montgomery, 2006; Carling et al., 2006; Haschenburger and Church, 1998; Hassan and Ergenzinger, 2005; Hicks and Gomez, 2003; Laronne and Duncan, 1992; Laronne et al., 2001; Mao and Surian, 2010; Surian et al., 2009a). While these studies can offer considerable insight into sediment mobility during different sized flood events, they are inherently constrained by the size of floods that occur during a sampling time period and the length of time available for sampling. Surian et al. (2009a) analysed the geomorphic effectiveness of different magnitude flood events by combining channel adjustment from aerial photographs with assessment of bed mobility from painted gravel patches. In comparison, the tool developed within this thesis measures geomorphic effectiveness by modelling sediment entrainment based on the high resolution survey of bar topography and sedimentology described above. This extends the tools available by allowing a greater range of flood flows to be considered.

Tracer studies inherently have a limited spatial scope, as it is only possible to paint or tag representative patches of bars (Haschenburger and Church, 1998; Hassan and Ergenzinger, 2005; Surian et al., 2009a). The technique described within this thesis allowed consideration of the entire bar, including the easily flushed material at the tail and the more static bar head. This whole bar perspective allowed greater understanding into the morphological effectiveness of floods at reworking different within bar surfaces, rather than representing a bar through a single grain-size distribution. This allowed the development of the conceptual model presented in Chapter 6. This model describes the proportion of within bar units (e.g. bar head and bar tail) that are reworked during each flood event. As a result, the geomorphic effectiveness of each flood is quantified, at much greater resolution than previously published. As a result, the impact of future floods can be predicted based on the magnitude of the flood and the within bar units able to be entrained.

Surian et al. (2009a) found that multiple discharges were responsible for shaping the river channel, with smaller floods around bankfull (RI of Q[^1.1]) reworking lower surfaces close to the channel, while floods with an RI of around Q[^5] were necessary to cause full gravel transport across all surfaces and drive significant morphological changes of the channel. Phillips (2002) similarly found bimodal flood distributions were responsible for forming channels, where frequent floods are competent to flush fine grained material, whilst large floods are necessary to rework the coarser fraction, causing
channel change. Other studies have traditionally viewed flows close to bankfull as the most geomorphically effective (Andrews, 1980; Emmett and Wolman, 2001; Sambrook Smith et al., 2010).

Findings from this study showed that floods much larger than bankfull were necessary to be geomorphically effective. More frequent floods ($Q^{2.33}$) were only able to flush finer sand and very coarse gravel fractions located on the tail and back channels of the bar. Floods $> Q^{20}$ were necessary to rework the cobble on the bar supra-platform and discharges $> Q^{50}$ were needed to entrain the boulder sized sediment on the bar head and drive significant channel reworking. This conceptual model of bar reworking furthers understanding into how these features are reworked, and the role of multiple discharges in entraining different grain size fractions within bar surfaces.

Variability in bar reworking was observed between the reaches surveyed. Frequent floods ($Q^{2.33}$) entrained a greater proportion of the downstream bars (between 41-44 % compared to 29-34 % for the upstream bars), which were situated in reaches characterised by gentler slopes and smaller grain-size (slopes 0.0029-0.0030 and $D_{50}$ 85-114 mm). The upstream bars characterised by larger bed material ($D_{50}$ of 140-230 mm) needed larger, less frequent floods to rework the bed. For example, the $Q^{10}$ entrained 73-75 % of the surface of these bars.

Sources of sediment played an important role in the size of sediment and its frequency of reworking. Most notably, material delivered from incision into lahar deposits forming terraces created a lag of large cobble-boulder sediment that lines the bed of the river. This boulder material requires larger, less frequent floods to rework the bed and elicit planform adjustment across larger scales (as discussed in Section 8.2.2). This builds on existing insights from Bluck (1987) who commented on the additional complexity in understanding channel adjustment and grain size sorting when the channel is reworking material generated from a different sedimentary event. Brummer and Montgomery (2006) document the role of lag delivered from landslides in creating stable nuclei which inhibited bed incision and increased residence times of underlying fractions. The lag in the Tongariro is a function of its volcanic past. The degree of incision determines the volume of lahar deposits delivered to a site, whilst reach scale slope controls the ability of a reach to rework and redistribute this material. Within the Tongariro, this material was remobilised to form bar heads and riffles, increasing the magnitude of events necessary to rework this fraction (described below) and slowing the rate of incision. This response is similar to observations made by Brummer and Montgomery (2006). As such, the input of this larger fraction plays an important role in the frequency of reworking for the Tongariro River. Most existing models of bar development are created for braided rivers with active beds (Ashmore, 1982; Ashmore, 1991a; Ashworth, 1996; Lunt and Bridge, 2004). This highlights the need to further investigate the influence of lag sediment from other sedimentary
influences (such as volcanic, glacial or due to decreased stream capacity following regulation) on bar development and channel adjustment. The role of lag and the history of sediment sources is an underplayed, yet vital, underpinning to understanding how river systems function in complex sedimentary environments.

Previous studies have discussed the occurrence of ‘stable’ nuclei or bar heads which act to protect a more mobile fraction of deposits (Ashmore, 1982; Ashmore, 1991a; Ashworth, 1996; Bluck, 1976; Bridge, 2003; Carling et al., 2006; Church and Jones, 1982; Church and Rice, 2009; Leopold et al., 1964). This study provides additional insights into the frequency of reworking and thus ‘stability’ of bar heads in a way not previously quantified. Despite significant variation in slope, bed material characteristics and valley confinement, all bar heads surveyed displayed an inherent resilience to entrainment (13-19% entrained) during the frequent bankfull ($Q^{1.33}$) flood. This indicates that all bar heads within this system were graded to be resilient to the prevailing flow during these frequent floods.

The assessment of bar reworking used local scale modelling of process to aid in describing the mechanisms of adjustment at the reach scale. This aims to understand which flows are responsible for driving the greatest rates of geomorphic adjustment. This local scale measure of process is linked to adjustment at the reach scale (e.g. aerial photographs of channel adjustment as presented in Chapter 6).

Lane and Richards (1997) advocate the need to question the value of intensive, local scale laboratory studies when they are carried out without implicit links to processes and morphodynamics across larger reach and watershed scales. Studies which have recently crossed these scales include the work by Surian et al. (2009a) that combines reach scale adjustment and full or partial transport on patches. Verdú et al. (2005) discuss the inherent limitations involved with hydraulic modelling of gravel bed systems due to a lack of high resolution grain size data. This thesis developed a technique which was able to map the distribution of grain size across bar surfaces, at the scale that the median grain size varies. This allowed a simple modelling approach to be applied which related bar reworking to analysis of channel adjustment at larger scales.

8.2.4 Why do multi-scalar process-based evolutionary trajectories provide a more comprehensive underpinning for the management of river systems?

“Thus, changes between average states can only be understood through an understanding of the sequence of configurational states through which the system evolves”

(Lane and Richards, 1997: 249)
This thesis extents previous approaches to creating evolutionary trajectories. Most commonly, evolutionary trajectories are a qualitative tool which describes channel response to anthropogenic influences (e.g. Beechie et al., 2008; Brewer and Lewin, 1998; Brierley and Fryirs, 2005; Comiti et al., 2011; Dufour and Piégay, 2009; Fryirs et al., 2009; Gurnell, 1997; Hooke, 2003b; Hoyle et al., 2008; Simon and Rinaldi, 2006; Surian et al., 2009b). These studies all implicitly argue for the importance of using a historical perspective to underpin management (Brierley and Fryirs, 2008; Wohl et al., 2005). However, the necessity of combining process-based morphodynamics with historical analysis has recently been recognised. Dufour and Piégay (2009: 576) support this by stating “we will have a better idea of the potential functioning of a given reach by integrating the historical trajectory into a process-based understanding”. Schumm (1977: 10) similarly states “… it is possible to view the [fluvial] system either as a physical system or a historical system. In actuality it is a physical system with a history’. These sentiments summarise the key aim of this thesis, promoting efforts to incorporate process-based measures of morphodynamics into analysis of observed changes to channel character and behaviour.

The lack of previous applications of process-based evolutionary trajectories most likely lies in the division that has formed between larger-scale, descriptive, qualitative conceptual approaches and quantitative, smaller-scale process-based approaches to geomorphic enquiry. This has been recently highlighted by describing the evolution of geomorphology as a discipline, whereby Davis’s landscape evolution approach has been viewed as subservient and largely replaced by Gilbert’s more analytical process-based approach (Lave, 2009a; Rhoads and Thorn, 2011; Richards and Clifford, 2011; Small and Doyle, 2012). However, as Small and Doyle (2012) contend, these approaches are not mutually exclusive, but rather complementary. This void between conceptual framings and the measurements which could and should support them, ignores the implicit strength and complementarities of these two approaches. This thesis adopts the perspective that in order for applied geomorphology to progress and underpin management in more comprehensive and progressive ways, techniques which combine historical, larger scale insights and measurement of local/reach-scale process need to be developed. Determination of evolutionary trajectories provides a conceptual tool with which quantification of process at the catchment and local scales can be integrated.

The calls to combine process and historical approaches can be paralleled by increased recognition for riverine management to be process-based (Beechie et al., 2010; Dufour and Piégay, 2009; Environment Agency, 2010; Lemons and Victor, 2008; Small and Doyle, 2012; Wohl et al., 2005). Wohl (2005: 1) suggests that “…restoration of process is more likely to succeed than restoration aimed at a fixed endpoint”. However, applied examples and frameworks to achieve this aim remain
scarce. Small and Doyle (2012) suggest that combining landscape modelling with classification techniques such as the Rosgen method (c.f. Rosgen, 1996) would provide a more comprehensive underpinning for management, by combining historical and process-based perspective. They suggest that “it is clear that multiple viewpoints and approaches triangulate towards a more thorough understanding of a system and will increase the probability of successful restoration” (Small and Doyle, 2012: 138). This sets the context and the need for a framework such as the process-based evolutionary trajectories presented within this thesis. Table 8.2 summarises the strengths of multi-scalar, process-based evolutionary trajectories. Each of the components has value, as supported in the list of references that support or provide an example of the use of each value. While each concept is supported by multiple studies, attempts to combine them have been scarce.

In this thesis, the ‘evolutionary trajectory’ conceptual model provides an empty tool box which is supported by process-based tools (catchment scale connectivity, reach-scale channel sensitivity and bar reworking). Combining key tools provides a more substantive basis for river management.

A key consideration within this study was which measures of process are most appropriate to underpin conceptual models of evolution? Recent debate has highlighted a disillusionment in the use of methods to predict and measure bedload transport due to high temporal and spatial variability in the process (Ashmore and Church, 1998; Ferguson and Ashworth, 1992; Gomez, 1991; Gray et al., 2010; Hicks and Gomez, 2003; Lane et al., 1995; Pitlick et al., 2009; Wilcock et al., 2009). As such, the creation of a tool which is able to provide a spatially and temporally accurate representation of sediment transport has provided one of the greatest challenges to geomorphology and has thus far remained elusive (Burt and Allison, 2010; Davies, 1987; Hicks and Gomez, 2003; Hubbell, 1987). As a result, there has been an increase in using form to derive and calculate process, a technique that has been a core principle of geomorphology for decades (Huggett, 2011; Rhoads and Thorn, 2011). This has especially increased as our ability to capture high resolution datasets of channel and catchment form is advancing exponentially, whereas our ability to measure process has not (c.f. Milan and Heritage, 2012; Rychkov et al., 2012). Ashmore and Church (1998) highlighted the importance of this approach, by arguing that the use of form to infer the process of sediment transport should be the new paradigm of geomorphic research. In response, a proliferation of studies apply this approach at the reach-scale to quantify erosion and deposition (Brasington et al., 2000; Brasington et al., 2003; Fuller et al., 2003a; Fuller et al., 2003b; Fuller and Marden, 2011; Fuller and Basher, 2012; Ham and Church, 2000; Lane et al., 1995; Schwendel and Fuller, 2011; Williams et al., 2011).
### Chapter 8: Discussion

#### Table 8.2: Key advantages of the process-based evolutionary trajectory approach as presented within this thesis.

<table>
<thead>
<tr>
<th>Key strength</th>
<th>Reason for importance</th>
<th>References which support the use of each approach</th>
</tr>
</thead>
<tbody>
<tr>
<td>A catchment scale consideration of sediment fluxes</td>
<td>The catchment scale sets the context for sediment delivered at more local scales. This describes how much sediment is being delivered and where it is sourced from.</td>
<td>(Benda et al., 2011; Brierley and Fryirs, 2009; Fryirs et al., 2007b; Kondolf et al., 2006; Lane and Richards, 1997; Parker, 2010; Piégay and Hicks, 2005; Raven et al., 2010; Sear et al., 1995; Sear et al., 2009; Walling and Collins, 2008; Wohl et al., 2005)</td>
</tr>
<tr>
<td>An appreciation of landscape evolution and the redistribution of sediment fluxes across longer, geological (10^3) timescales</td>
<td>Highlights how the catchment has evolved and changes in sediment loads and process zones. This larger-scale evolution sets the boundary conditions for local scale evolution and indicates the direction the landscape components are evolving in.</td>
<td>(Brunsden and Thornes, 1979; Burt and Allison, 2010; Harvey, 2002; Houben et al., 2006; Houben et al., 2009; Piégay and Hicks, 2005; Schumm and Lichty, 1965; Walling and Collins, 2008; Wilkinson and McElroy, 2007)</td>
</tr>
<tr>
<td>Uses historical planform adjustment to assess channel response</td>
<td>Used to assess differences in sensitivity, nodes of adjustment, timescales of response and identification of thresholds. Basis to predict future trajectories of evolution.</td>
<td>(Beechie et al., 2008; Brewer and Lewin, 1998; Brierley and Fryirs, 2005; Comiti et al., 2011; Dufour and Piégay, 2009; Fryirs et al., 2009; Gurnell, 1997; Hooke, 2003b; Hoyle et al., 2008; Simon and Rinaldi, 2006; Surian et al., 2009b; Wishart et al., 2008)</td>
</tr>
<tr>
<td>Incorporates process-based assessment of reach scale morphodynamics and controls on channel form</td>
<td>Explains how controls drive reach adjustment. Important for identifying which grain-size fractions drive change within a reach, and why zones may be more sensitive to particular fluxes.</td>
<td>(Church, 2006; Eaton et al., 2010; Ferguson, 1981; Perşoiu and Rădoane, 2011; Schumm and Khan, 1972)</td>
</tr>
<tr>
<td>Incorporates local-scale measures of bar reworking and sediment entrainment for different magnitude/frequency flood events</td>
<td>The reworking of bar surfaces provides a process based underpinning that explains which floods are geomorphically effective and their ability to elicit change.</td>
<td>(Bridge, 2003; Carling et al., 2006; Lane et al., 1994; Laronne and Duncan, 1992; Laronne et al., 2001; Mao and Surian, 2010; Rice et al., 2009; Surian et al., 2009b)</td>
</tr>
<tr>
<td>Uses measures of form (catchment topography, reach planform, long profile and cross-sections, bar topography) across multiple scales to infer process (sediment transport, shear stress, bar reworking)</td>
<td>Direct measures of process vary in magnitude over spatial and temporal scales. Instead, form provides a way to measure process and the outcome of processes across multiple scales.</td>
<td>(Ashmore and Church, 1998; Brasington et al., 2003; Ferguson, 1986; Huggett, 2011; Lane et al., 1995; Reinfelds et al., 2004; Whipple and Tucker, 1999; Wilkinson and McElroy, 2007)</td>
</tr>
<tr>
<td>Combines a range of approaches across multiple scales to answer questions posed by the specific characteristics of the system.</td>
<td>By combining insights across multiple scales, this thesis acts to decrease problems of scale-dependency and create a more comprehensive information base which describes how the system operates. i.e. multiple viewpoints create better understanding of the system (Small and Doyle, 2012).</td>
<td>(Church, 2007; Frissell et al., 1986; Montgomery, 1999; Parsons and Thoms, 2007; Small and Doyle, 2012; Thorp et al., 2006)</td>
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This thesis extended the morphological (form-process) approach by increasing the range of scales at which it was applied (beyond the reach-scale), and the measures of process that were used (rather
than just erosion and deposition). Form refers to the morphological structure and geometric properties of a landscape element, whilst process is associated with an expenditure of energy (i.e. power, energy flux, force and momentum, which can translate into geomorphic measures of shear stress, stream power and sediment entrainment) (Huggett, 2011). High resolution surveys of bar forms in Chapter 6 were used to map the distribution of grain size and water depth for different flood events. This measure of form was used to calculate sediment entrainment using physics-based bedload transport equations. At the reach-scale (Chapter 5), cross-sections and slope (form) were used to quantify process by calculating mean boundary shear stresses and transport capacity. At the catchment scale (Chapter 3), slope and upstream drainage area measured from a DEM were used to calculate stream power and erosion indices, detailing expected energy during a flood. Finally, the morphology of landscape units (i.e. terrace and delta morphology) was used to quantify sediment erosion and deposition since 1.8 ka for each unit respectively. These applications demonstrate that form across multiple scales (catchment, reach and bar) can be used to provide measures of process. This illustrates the importance of not only directly measuring sediment transport, but using measures of energy (i.e. stream power and shear stress) to directly explain fluvial patterns of response.

The use of multiple scales to derive evolutionary trajectories was a key strength within this study. ‘Which scale?’ is a commonly asked question within geomorphology, acknowledging that the scale of inquiry will inherently determine the results and phenomena observed (Church, 1996; 2007; Frissell et al., 1986; Harvey, 2002; Lane and Richards, 1997; Nadal-Romero et al., 2011; Richards and Clifford, 2011). Moreover, analysis of larger spatial scales will inherently describe processes which occur across larger temporal scales (de Boer, 1992).

Each scale of inquiry within this thesis had a unique contribution to the creation of evolutionary trajectories within the Tongariro catchment. Catchment-scale framings of sediment flux to describe inputs and controls upon individual reaches are increasingly being promoted as vital to management and rehabilitation schemes (Benda et al., 2011; Brierley and Fryirs, 2009; Fryirs et al., 2007b; Kondolf et al., 2006; Lane and Richards, 1997; Parker, 2010; Raven et al., 2010; Sear et al., 1995; Sear et al., 2009; Walling and Collins, 2008; Wohl et al., 2005). The reach scale is essential to assess the evolution of the river, including consideration of trajectories the channel is moving along, geomorphic thresholds, notions of contingency, ‘morphological conditioning’ and the inherent sensitivity (rates and nature) of response displayed by a specific system (Brunsden, 2001; Downs and Gregory, 1993; Fryirs et al., 2009; Lane and Richards, 1997; Schumm, 1980). Local, bar scale insights describe what the channel bed is composed of. This can be used to assess channel behaviour,
including determining which floods are geomorphically effective and how local scale patterns of sediment transport can drive channel adjustment across larger scales (Bridge, 2003; Carling et al., 2006; Lane et al., 1994; Laronne and Duncan, 1992; Laronne et al., 2001; Mao and Surian, 2010; Rice et al., 2009; Surian et al., 2009b). This demonstrates the necessity of each scale, and as such, the need to combine the scales of enquiry, providing a more comprehensive picture of the system (i.e. Aristotle’s famous quote ‘the whole is greater than the sum of the parts’, c.f. Kirby, 2008).

Widespread recognition supports the need to consider multiple scales within geomorphology (Brierley and Fryirs, 2005; Church, 1996; 2007; Frissell et al., 1986; Kondolf et al., 2006; Montgomery, 1999; Parsons and Thoms, 2007; Thorp et al., 2006; Ward et al., 2002; Wohl et al., 2005). Despite the volume of literature advocating for research across multiple scales, examples of its application are rare. In part, division between sub-disciplines within earth science may act as a barrier (Benda et al., 2002; Montgomery, 2001b; Newson and Large, 2006). This thesis crosses disciplinary boundaries, incorporating insights from geology (landscape evolution and response to volcanism), reach scale traditional geomorphology (channel planform pattern and adjustment), applied geomorphology (sensitivity framework) and process-based geomorphology/hydrology (bar reworking and channel competence calculations). Whilst other people have worked within these different disciplinary components, the strength of this thesis is that it pulls them together. The evolutionary trajectory approach provides a central conceptual framework within which these disparate scales and types of information may be combined. In essence, this thesis combines qualitative, landscape evolution approach with quantitative, process-based analysis to create more comprehensive, grounded conceptual models as argued for by Small and Doyle (2012). This fills a void that has been created between modelling applications and more qualitative, descriptive, conceptual approaches. Thus the core argument of this thesis is that the future evolution of applied geomorphology lies not in advances in the discrete fields of modelling or qualitative approaches, but in finding more coherent ways to develop tools which cross these boundaries, creating more comprehensive datasets based on multi-scalar, multi-disciplinary approaches which combine qualitative and quantitative techniques for understanding the processes driving the system.

Lastly, the multi-scalar, process-based, evolutionary approach provides a tool which works with the uncertainty and complexity of the real world. This recognises that managing these dynamic systems inherently contains a degree of uncertainty (Darby and Sear, 2008). Thus we need to design techniques and approaches that seek to find order out of chaos (Phillips, 2003a). The evolutionary trajectory approach estimates future states and directions of adjustment, rather than fixed, static endpoints, which are misleading given this uncertainty (Brierley and Fryirs, 2008; Dufour and Piégay,
In addition, the multi-scalar, process-based approach provides a more comprehensive basis upon which to base these future estimates of response. The ‘future response trajectories’ built upon this by suggesting that different influences are likely to push the system in different directions, recognising the diversity in pressures as well as response. For example, the influence of ash-based volcanic eruptions into the wandering cobbles bed reach was minimal, as the channel was competent to flush the material rapidly. In contrast, lahars were able to deliver much larger cobbles-boulder sized sediment which remains in the system, creating recovery times of 1000s of years. Thus this approach, tackles the problem of uncertainty by presenting multiple trajectories based on the type and magnitude of channel response to different natural and anthropogenic controls. This allows the most likely response to be estimated based on the likelihood of the occurrence of a control. For example, the Q20 flood is moderately likely, whilst a large lahar event is unlikely, as all volcanoes in the Tongariro catchment have 0-1 alert levels, indicating minimal signs of volcanic unrest (GNS, 2012). Channel response to a flood is likely (i.e. widening and erosion) whilst response to a lahar (massive aggradation and infilling of the channel) is unlikely. Such a dynamic and diverse approach for estimating future channel adjustment is essential in such complex environments.

In the face of such complexity and uncertainty, application of a diversity of geomorphic approaches is necessary (Small and Doyle, 2012). Tools applied need to be selected with regards to the specific questions and knowledge gaps that a catchment or site may have (Brierley et al., 2010). For example, Taupo eruption (1.8 ka) caused an abrupt change in process-zones and provided the rationale for the creation of the sediment budget to describe how the system had evolved from this time-point. As such, an ‘adaptive’ approach that uses multiple lines of evidence generated through the selection of a range of tools may be a more appropriate way to understand how a system is functioning given the inherent uncertainty (c.f. Brierley et al., 2010; Darby and Sear, 2008; Phillips, 2003a; Small and Doyle, 2012).

### 8.3 IMPLICATIONS FOR MANAGEMENT

“Realizing ... opportunities requires geomorphologists to re-cast ‘applied geomorphology’, wherein the goal for environmental management is not the control or manipulation of the natural environment, but rather the maximizing of ecosystem services.”

(Downs and Booth, 2011: 98)

River management has traditionally endeavoured to fix the river in place, in engineering based procedures referred to as ‘command and control’ approaches (Brierley and Fryirs, 2005). However, recent advances in mind-set question this philosophy favouring methods which support ecosystem and geomorphic processes by allowing the channel to self-adjust. As Downs and Booth (2011)
suggest above, there is a need to ‘recast’ applied geomorphology so that it sustains the ecosystems contained within, rather than concentrate on stabilising river channels. The process-based approach to restoration or management is being promoted as the best future direction to maximise ecosystem services (Beechie et al., 2010; Dufour and Piégay, 2009; Environment Agency, 2010; Kondolf et al., 2006; Lemons and Victor, 2008; Small and Doyle, 2012; Wohl et al., 2005) and may be defined as “re-establishing normative rates and magnitudes of physical, chemical, and biological processes that create and sustain river and floodplain ecosystems” (Beechie et al., 2010: 209). In order to do this, insights from multiple scales must be combined to understand how a system has changed and what has caused this change. Beechie et al. (2010) identified key processes which need to be sustained, with processes identified at the watershed scale including sediment delivery, hydraulic regime and fluxes of organic matter, whilst key reach scale processes were consideration of the fluxes which maintain habitat morphology and geomorphic structure.

Process-based approaches can be complemented by ‘room to move’ approaches. Example applications of this approach include Piégay et al.’s (2005) ‘erodible river corridor concept’ and Rapp and Abbe’s (2003) ‘room to move’ approach. Alluvial rivers are self-adjusting entities and the above approaches suggest that given a sufficient area to adjust within, channels are more competent to reset their form to changing prevailing flow and sediment conditions than humans are at creating it. As such, riparian corridors are left for the channel to adjust within. However, these approaches need to be designed considering the area of floodplain necessary to contain future adjustment. This thesis provides a geomorphically based rationale to start to tackle some of these questions.

This section applies a multi-scalar, catchment-specific geomorphic template to highlight implications for process-based management of the Tongariro River. It is not the role of this thesis to create a management plan for the Tongariro. However, the information base compiled within this thesis can be used to guide the management of the Tongariro River and ‘recommend’ actions. This includes discussion of catchment scale insights which contextualise reach scale management, building upon the evolutionary trajectories presented in Chapter 7 to create geomorphically acceptable patterns of channel behaviour.

8.3.1 Catchment-scale Context for Process-based Management

Comprehensive management plans need to consider prevailing sediment and water fluxes across the catchment. This section shows how catchment scale sediment fluxes influence reach scale management.

The Tongariro catchment had two major sources of sediment. The volcanic western sub-catchment has generated vast volumes of sediment. However, this material is delivered during high magnitude,
infrequent events (namely lahars) which greatly exceed the transport capacity of the streams, so that the landscape is smothered with sediment, creating a low slope plateau. Streams have incised into these deposits, disconnecting drainage lines from reworking these stores. Thus, despite, or because of, the capacity of this sub-catchment to generate high volumes of sediment, the volume of bedload delivered to the Tongariro River is relatively minor. However, finer grained sediment is able to be moved more easily across this gravel base. Sediment supply is increased during volcanic eruptions, with 6900 kilotons delivered during the 1995 eruption (Collier, 2002). Higher energy gravel bed reaches can mobilise this fraction easily, and it has a short residence time of a few years in the system (Genesis Energy, 2000). However, this fraction can be deposited in the meandering and delta reaches due to the low slopes and energy of these River Styles.

In contrast, the Kaimanawa Ranges represents a major source of material, as the uplifting greywacke basement provides an on-going source of sediment. The sediment budget predicted a volume of around 800,000 m$^3$yr$^{-1}$ is generated (Chapter 3). The Tongariro River in the upper-mid catchment has been trapped between this uplift, and the growth of the volcanic ring-plain As a result, the channel flows within terraces around 30 m high. Minimal accommodation space is available to store sediment, and stream power is high and bed material large, indicating that much of the material delivered must be flushed. This indicates that much of the sediment supplied into lower Tongariro River is the gravel sized material delivered from the Kaimanawa Ranges (45 - 65 mm). Generation and transport of this fraction is expected to increase during high rainfall, as flood events can move material from gullies and off mountain sides, through the highly connected Kaimanawa Ranges uplands and down into the lower Tongariro. In addition, large floods can rework active stores within the terrace confined reach, mobilising this material and moving it downstream. Dams in the mid-catchment may increase the pulsed nature of sediment delivery, as flood gates are opened, and the dams flushed once discharges exceed 100 m$^3$s$^{-1}$, releasing up to 60,000 tons of sediment (Collier, 2002). Whilst this may influence the timing, it does not affect the overall volumes of bedload sediment delivered downstream. Managing the transfer of this gravel fraction through the lower Tongariro River, given the sensitivity of each River Style to this material, is a key consideration for process-based management.

A key catchment scale control can be seen in the influence of the base level of Lake Taupo on the transfer of sediment through the delta. This has been regulated since 1941 by control gates on the outlet of the Waikato River (Genesis Energy, 2000). Mean lake levels from 1906 - 1940 to 1942 - 1996 have increased by 6.5 cm (Eser and Rosen, 2000), skewing the overall distribution of lake levels towards higher levels (Smart, 1999). This has created a backwater effect, which is estimated to
influence the transport of bed material in the lower 3 km of delta (Tonkin and Taylor, 1999b). High lake levels have been attributed to the channel widening in the braided reach both during the 1958 flood (Chapter 4) and 2004 (Munro, 2004) Q10 flood events. Base level control is also likely to impede the ability of the meandering and delta reaches to flush fine grained sediment and may contribute to the rate of narrowing and deposition observed within these reaches. As this fraction can be transported across a range of flows due to its non-cohesive nature, sediment delivery is on-going. However, large floods can rework larger surfaces upstream, which may mobilise this material from interstices and increase supply. High on-going supply of volcanic sand is likely in these reaches.

### 8.3.2 Implications for Managing the Wandering Cobble bed Reach

Channel adjustment for the Wandering, cobble bed river was contained within land allocated to contain the river. The channel rarely adjusted its position on the valley floor beyond that of reworking recently occupied flood channels and islands (Figure 8.2). Whilst the channel responded to flood events by increasing braiding and active channel widths, highly sensitive sedimentation zones characterised by greater in channel stores and channel response were found to occur at the same locations. In other wandering gravel bed systems these zones have been found to migrate, providing a threat to infrastructure as they move downstream (Desloges and Church, 1989; Passmore and Macklin, 2000). However, within the Tongariro River, the terraces margins influenced the transport capacity and channel alignment (i.e. sinuosity) determining the type and magnitude of response. As a result, sedimentation zones were reactivated at the same locations which were characterised by low slope, low transport capacity relative to catchment position, wider terrace margins and a more sinuous planform, as greater volumes of sediment were able to be stored, eliciting greater channel adjustment.

Given their location, sedimentation zones are a low threat to infrastructure along the lower Tongariro River. The only zone where a possible threat is likely is upstream of the SH1 Bridge. However, this reach has retained the same planform since the initial survey map in 1928, even before stopbanks were constructed in the 1970s (Figure 8.2). Turangi sits on lahar deposits into which the channel has incised. As a result, the channel is inset and partly disconnected from the surface upon which Turangi is located. However, the edge of these deposits have been sculpted by fluvial erosion. This is evidenced by a lower surface with smooth edges created by river meanders (Figure 8.2). This line is also consistent with the 100 year flood line (see Chapter 2, Jones, 2003), indicating that whilst the channel has not occupied this zone within the last 80 years (the time periods over which we have survey map and aerial photograph evidence), it is still at risk of flooding. However, levees that constrain the channel decrease this possibility and are viewed as sufficient to manage the position of the channel within this reach.
In previous studies on sediment loaded rivers within New Zealand, levees and stopbanks have been found to be linked to bed aggradation (Davies and McSaveney, 2006). However, within the lower Tongariro River, steep slopes retained by the lag of cobble-boulder size bed material are competent to transport the gravel-sized active fraction delivered from the Kaimanawa Ranges. As the sediment supplied to the reach can be transported, the risk of bed aggradation is minimal. This is supported by minimal changes to bed level measured at the Turangi Gauging station, in the mid-section of this reach (Chapter 4). As such, negative effects commonly associated with levees are unlikely to occur within this reach.

Figure 8.2: Legacy of channel adjustment around the location of SH1 Bridge overlain upon 1 m LiDAR derived detrended DEM showing height above water surface. This shows that channel adjustment has occurred within the same zone over the past 80 years. In the past it is likely that it occupied land now containing Turangi township. This is illustrated by the fluvial shaping of the edges of the lahar deposit upon which Turangi is situated.
Recommended management actions for this reach are minimal. The channel is able to function naturally within the constraints imposed by human development. Channel adjustment is driven by high magnitude events ($> Q^{20}$) as these are necessary to rework bars (including lahar lag materials) and drive planform adjustment. Whilst flows are regulated, the well managed regime allows a near natural distribution of the larger geomorphically effective floods, providing minimal impact on patterns of channel adjustment. However, less frequent freshes ($\sim 50 - 100 \text{ m}^3\text{s}^{-1}$) may increase fine grained sediment deposition in the interstices of boulders, decreasing habitat quality (Jowett, 1984).

The main recommended management action within this reach is the maintenance of the riparian strip which has been replanted by the Turangi community. Ensuring that these plantings survive and thrive and controlling invasive weed species is the only recommended management action.

### 8.3.3 Implications for Managing the Braided Gravel bed reach

The unconfined, braided gravel bed reach was found to be highly sensitive, undergoing significant changes to the active channel area over the past 80 years. This is attributed to the low slope of this reach (0.0026), which captures all the gravel-cobble sized material which is flushed through the terraces. Gravel is delivered by large floods from the Kaimanawa Ranges, with volumes estimated to be around 800,000 m$^3$yr$^{-1}$ (Chapter 3). The gravel-sand transition at the downstream boundary of this reach illustrates the propagation of this material further downstream (slopes decrease to 0.0016), whilst steep slopes in the terrace confined reach upstream reflect an ability to rapidly transport this fraction to the braided reach (slopes between 0.0046 - 0.0111). Gravel mining is the main existing form of management. However, this has created a discrete zone of lower elevation which captures and retains sediment delivered. Aerial photographs indicate that the braided reach extended further downstream in the 1940-1960s (Figure 8.3). Gravel mining has reduced slopes and sediment transport capacity directly downstream of the SH1 bridge, concentrating the location of future aggradation directly adjacent to Turangi. In this instance, gravel mining has impeded the natural development of steeper slopes which aggradation creates. As a result, the braided area which stores this fraction has decreased significantly, from 1.7 km$^2$ at it maximum extent in 1958 to 0.5 km$^2$ in 2007 and migrated upstream, closer to Turangi township.
Figure 8.3: Management of the braided reach that works with the adjustment of the system. The red line indicates the room allocated within which the channel can move and coincides also with the channel outline in 1958 and predicted Q$_{100}$ flood inundation. Blue arrows indicate the area of most likely to become braided if the channel was left alone based on past aerial photographs (particularly 1941 immediately before lake regulation).

Process-based management schemes suggest that rivers need ‘room to move’ to allow the systems to self-adjust to the material delivered (Piégay et al., 2005; Rapp and Abbe, 2003). The area occupied by the braidplain in 1958 remains undeveloped, and has been retained as reserve. By allowing aggradation and slopes to steepen directly downstream of Turangi, the braided reach would propagate further downstream, to occupy previously active braidplain. Increasing the area which transports and stores the active gravel-cobble fraction would make the braided reach more resilient to pulses of sediment as it could store sediment across a greater aerial extent. This would also extend the locale of braiding further away from Turangi township (Figure 8.3). As such, management options that encourage braiding to increase and propagate further downstream are suggested:

1. Cease gravel mining directly downstream of Turangi where the channel first adopts a braided planform. This increases braiding and the sensitivity of this reach.
Chapter 8: Discussion

2. If gravel mining is to be carried out, then lowering sections at the downstream extent of the current braided reach would aid in increasing the area that can store the gravel size fraction and decrease the sensitivity of the reach.

3. Low flow channels on the braidplain have been in-filled and bunds erected (Jones, 2003), again amplifying storage directly downstream of Turangi. Removal of these would aid in extending the section of braiding downstream and moving floodwaters away from Turangi.

4. Exotic willows increase bank strength and choke the channel, impeding sediment transport (Jones, 2003; Tonkin and Taylor, 2003). Willow removal would allow the braided reach to migrate further downstream. This is particularly important immediately downstream of the most active zone of braiding where the channel is pinned in place, forcing the river into a single channel (Figure 8.3).

In summary, management that works with the system relies on reactivating parts of the ‘natural’ braidplain and increasing the area which stores the gravel-cobble fraction.

8.3.4 Implications for Managing the Meandering Sand and Multi-channelled Delta Reaches

The unconfined, meandering sand bed river and multi-channelled delta are moving along a narrowing trajectory towards avulsion. The channel has become perched above the floodplain and channel capacity has been steadily decreasing over the past 80 years (Chapter 5; Figure 7.7B). This has led to the creation of a secondary channel to the west of the main channel, whereby a head-cut is migrating upstream, causing channel enlargement and generating shear stresses an order of magnitude greater than those in the primary channel during floods (Smart, 2011). However, the placement of this new channel presents a threat as the location of the new delta would coincide with the Tokaanu Tailrace, which delivers water used in the Tongariro Power Development Scheme (TPDS) (Figure 8.4). Due to the highly sediment charged nature of this catchment it would be expected that the delta would prograde rapidly, blocking this outlet. Another lesser channel is forming to the east of the main channel, which would form a delta within a wetland region, minimising the threat to infrastructure and allowing the channel to adjust within this region ‘naturally’. However, flow velocities during floods are lower within this channel compared to the developing eastern secondary channel (Smart, 2011). This requires flow to be redirected to the east in order for this channel to become the main channel during an avulsion event.

Recommended management actions should act to block the establishment of the western secondary channel, allowing flow instead to move east and erode the eastern channel. Suggested means are to establish stopbanks which block flow to the west, redirecting flow back to the eastern portion of the
delta (Figure 8.4). This uses ‘room to move’ or ‘erodible corridor’ notions by allocating an area of land within which the river can undergo the type of adjustment that is appropriate for the river type, in this case avulsion (Piégay et al., 2005; Rapp and Abbe, 2003). Native vegetation could be used as a tool to strengthen stopbanks to the west. Clearing vegetation along the eastern secondary channel could accelerate the rate that the river erodes and forming a channel within this vicinity, as suggested by Smart (2011). It is especially important that any willow species are removed as they increase roughness and soil cohesion. If a channel does not form naturally, then constructing a channel may be an option. However, it is more beneficial to let the river form a new channel itself, under the premise that rivers are better at creating appropriate channel planform and geometry than humans are (Rapp and Abbe, 2003). While establishing a channel to the east will protect the Tokaanu Tailrace at the moment, it is worth noting that as the floodplain aggrades in this region, it increases the likelihood of the channel switching back to the west in the future. However, due to the minimal adjustment of channels in the present delta over the last 80 years, it is predicted that the time to the next avulsion would be > 100 years (Chapter 4).

If the above approaches fail to decrease flooding, then reconsideration of the management of the water level of Lake Taupo may be the most appropriate form of management. This could be based around maintaining levels as close as possible to pre-regulation (i.e. lake level predictions based on rainfall), rather than using statistical annual means which allow greater variation. The difficulties of this are recognised, given the politically and economically loaded nature of such a management strategy.
Chapter 8: Discussion

Figure 8.4: Process-based management for the Tongariro delta.
9 Conclusion

This conclusion summarises the key findings from the fundamental research questions as set up in Chapter 1. It then discusses the overall contributions of the thesis including applicability of the approach elsewhere and future work that could be beneficial within the Tongariro.

9.1 Key Findings from Fundamental Research Questions

Key answers to the Fundamental Research Questions from this thesis are outlined below.

9.1.1 Why do larger-scale spatial and longer-term temporal patterns of sediment delivery and connectivity across the Tongariro catchment provide an important framing for the adjustment of the lower Tongariro River?

- This work illustrated the importance of a catchment scale framing to understanding adjustment in the lower Tongariro River. This showed a complex arrangement of sediment process zones, with storage units in the headwaters buffering sediment delivery from the highly productive volcanic region to the lower catchment and terraces in the mid-catchment disconnecting significant sediment stores.

- A morphological sediment budget approach quantified volumes of sediment generated and delivered to and from different units within the catchment. This temporal component to analysis of catchment scale sediment flux aided in understanding the relative importance of different sources for delivering sediment to the lower river over differing timeframes.

- The use of multiple approaches including DEM processing to generate statistics on slopes, stream power and erosion indexes, and pre-existing analysis of suspended sediment supply and erosion terrains, grounded by local site specific analysis provided a comprehensive picture of the connectivity of sediment across and within the landscape units of the Tongariro catchment.

9.1.2 Why does reach scale response differ along the lower Tongariro River and what key controls drive this variability in sensitivity?

- The width of the floodplain pockets within the terraces was set by the valley slope pre-Taupo eruption, 1.8 ka., whereby steeper slopes generated more energy to erode sediment, creating wider floodplain pockets. Terrace width directly influenced reach sensitivity of the wandering cobble bed river, whereby reaches located with wider terrace margins underwent greater adjustment in response to flood events. While terrace width did not directly influence the width of the channel, it had an antecedent control on channel alignment, with more sinuous reaches undergoing greater adjustment.
Chapter 9: Conclusion

- Wider, more active sedimentation zones were characterised by lower stream power for their catchment position than single channelled reaches, providing a process-based rationale for the increased sediment storage at these locales. Sedimentation zones were also able to adjust their planform to balance grain size and slope, seen in a tight correlation with a $R^2$ value of 0.97 between these two variables, whilst single channelled reach displayed far greater variation, seen in a correlation with a $R^2$ value of 0.31. Thus, sedimentation reaches are graded to be in-sync with their valley position.

- Rapid changes in boundary conditions (i.e. valley confinement and slope) create the high diversity in River Styles in the lower Tongariro. As the beds of each River Style have different grain size mixes, each river type was found to be sensitive to different natural and anthropogenic influences based on the extent of their influence on the bedload transport of specific grain size fractions. For example, ash and sand generated from volcanic eruptions had minimal impact upon channel adjustment of the wandering, cobble bed river, but they had a high impact on the rate of infilling for the sand bed meandering river.

9.1.3 Why are the patterns and frequencies of bar reworking along the lower Tongariro River an important underpinning to channel adjustment at the reach scale? How has this approach furthered ways in which we quantify local scale processes?

- The use of TLS to capture grain size allowed sediment size across the whole of the bar to be mapped, providing previously unprecedented understanding into the sedimentology of bar surfaces.

- This dataset underpinned a conceptual model which assessed the geomorphic effectiveness of different flood events to rework the bar surfaces for reaches with different boundary conditions.

- Frequent floods ($Q^{2.33}$) entrained sediment at the tail and back-channel of the bar, whilst larger, more infrequent floods were necessary to generate whole bar reworking and to entrain the bar head.

- The lower energy bars (slope ~ 0.003) located downstream underwent greater reworking during frequent floods ($Q^{2.33}$) compared with the higher energy bars (slopes 0.005 - 0.0076) which needed greater floods ($Q^{10}$) to undergo significant reworking.

- All bar heads displayed very similar rates of reworking with 13 – 19 % entrained during the frequent flood ($Q^{2.33}$). This indicates that in order for bars to form within the Tongariro a proportion of the bar must have an inherent degree of resistance to frequent floods.
• The extent of bar reworking related well to the aerial photographic history, which captured past response to flood events. This provided a process-based rationale for predicting how a reach is likely to respond to future flood events.

9.1.4 Why do multi-scalar process-based evolutionary trajectories provide a more comprehensive underpinning for the management of river systems?
• This thesis extends previous approaches to evolutionary trajectories by combining information across multiple spatial and temporal scales, and most notably providing a catchment-scale framing. This allowed channel response to be more readily related back to underlying causes.
• Findings from this study demonstrate the need for process-based understanding of channel adjustment to ensure that predictions of future adjustment are underpinned by an appreciation of the processes which drive channel response within a particular reach. The flexibility to apply understandings and measurements of landscape form across multiple scales aids inferences of process relationships.
• The core argument of this thesis is that if we are to improve our understanding of riverine morphodynamics, then we need to find better techniques for combining insights across multiple scales. Specifically, combining information from the larger catchment scale, gained using historical approaches, with smaller scale analysis of process allows a more comprehensive view of the whole system, offering significant insights into process dynamics.

9.2 Overall Contribution of the Thesis
This work illustrates the complex, multi-scalar nature of river systems and the catchments within which they are contained (Montgomery, 1999; Parsons and Thoms, 2007; Phillips, 1999; 2003a). In the face of this complexity, adaptive management approaches that embrace uncertainty are required to devise management plans that work with natural processes (Brierley et al., In Press; Newson, 2009; Wohl et al., 2005). Previous work has advocated the need for river management and rehabilitation to be underpinned by catchment specific understanding (Beechie et al., 2010; Benda et al., 2011; Downes and Gregory, 2004; Kondolf et al., 2006; Raven et al., 2010; Sear et al., 2009; Wohl et al., 2005). This study illustrates that for this to occur, forms of analysis and techniques must be set by the questions that arise within a specific catchment (c.f. Brierley and Fryirs, 2009). For example, the influence of the lahar lag on bed reworking gives a rationale for analysis of bar reworking to underpin channel adjustment in the Tongariro River. To truly appreciate the complexity of the natural world, practitioners need to ‘know their system’ before choosing what forms of assessment are best applied to answer the specific questions that arise in a setting (Brierley and Fryirs, 2005). Combining landscape evolution scale approaches with more local-scale process
provides a particularly advantageous approach to gaining a comprehensive and holistic understanding of the specific system (c.f. Brierley and Fryirs, 2009; Kondolf et al., 2006; Small and Doyle, 2012). This thesis has analysed how the Tongariro adjusted over 2000 years, 100 years and in response to individual events of specific sizes and reoccurrence intervals, including the mean annual flood. This was applied at the catchment, reach and bar scales. These multiple layers of understanding show how landscape evolution inherently sets the boundary conditions which drive marked differences in channel type and response in the lower catchment. As such, evolutionary analyses over multiple temporal and spatial scales are essential underpinnings for predicting future trajectories in complex, dynamic environments.

A major gap in river research is the establishment of open-ended frameworks which synthesise information generated across different scales and approaches (Brierley et al., In Press). Catchment framed, process-based evolutionary trajectory approaches provide coherent frameworks to address this concern. Such framings use analyses of channel processes and evolution to predict future types and rates of adjustment. The greater the range and quality of data that are combined as an evolutionary trajectory, the more substantive the output and insights become (i.e. additional data can be used to increase insights and comprehension). This contrasts to some modelling applications wherein the greater the number of data inputs the more uncertainty is introduced, prospectively resulting in less accurate results, as demonstrated by the avocation for simplified reduced complexity modelling (Brasington and Richards, 2007; Coulthard et al., 2007). However, the approach adopted in this study processes each form of data and understandings across the range of scales analysed here, and then uses these multiple findings to create a more complete picture of the system. In contrast, within models, data are combined to provide a single answer, which may, or may not, have a known degree of uncertainty. This thesis has developed and applied a tool which synthesizes these multiple findings, demonstrating how multiple scale and types of enquiry can be combined as a comprehensive whole. The use of form to infer process is a key strength of this approach, incorporating multiple techniques to quantify both sediment transport and potential energy. As it is quicker and easier to record morphology than process, this approach allowed the incorporation of a significant number of measures across a wide range of scales.

Insights into the history and evolution of the Tongariro River emphasise management perspectives that frame analyses of rivers as living systems, with a past which influences their future (Harper and Everard, 1998). If we want to improve our ability to understand these systems, we need to become more astute at relating past stages of adjustment to contemporary controls, and likely future trajectories. When linked to contemporary processes, such insights can be used to indicate the
direction and rate of geomorphic adjustment that set the response gradient for any given system (Fryirs et al., 2009).

### 9.3 APPLICABILITY OF THE APPROACH OUTSIDE OF THE TONGARIRO CATCHMENT

Specific attributes of the Tongariro catchment made this approach particularly applicable for this setting. Firstly, the Taupo eruption which reset the system provided a distinct point in time from which the channel started along its current trajectory, allowing changes to landscape units to be quantified. As such discrete landscape forming events are unique, the exact methods developed and applied in this thesis are not directly transferable to other settings. However, the underlying principles are transferable, and the skill is for the user to identify opportunities to work out what techniques may be the most applicable for any given situation. For example, many of the landscapes within the Northern Hemisphere have been shaped by glaciers (Church and Slaymaker, 1989). Following the retreat of the ice sheet, sediment supply has decreased and channels have incised into valley fills. These provide a time scale (e.g. 12,000 years since the retreat of the ice sheet in Britain) and bulk volumes of sediment can be quantified to describe this evolution (c.f. Tunnicliffe and Church, 2011). However, quantifying bulk volumes of sediment transport was only one aspect of this work. The key is that the combination of layers from multiple types of information, as supplied at the catchment scale though the combination of the connectivity analysis and the sediment budget, were used to create a single combined, system-wide understanding.

The high complexity of the Tongariro system generated the need for many types of analysis to be undertaken in order to build a comprehensive view of the system. In other, simpler systems this level of data may not be necessary to guide management. The dynamism and flexibility of evolutionary trajectory approaches to geomorphic analysis are key attributes that support this approach to management applications (see Surian et al., 2009). Application of a blanket approach applied in all settings is unlikely to yield appropriate results, as selected approaches must reflect the specific process relationships that occur in that setting (Brierley et al., In Press). This can be reflected in the number of ‘threads’ and extent of the process-base regarded as necessary to underpin the evolutionary trajectory analysis based on the complexity of the system. For example, this approach could be applied in a simpler system using fewer ‘threads’ or datasets than was deemed necessary to build a complete systems understanding within the Tongariro scenario. This is underpinned by the need to know the system before selecting the types and extent of data that are required to characterise the behaviour of the system. Thus, this approach could be qualitative, as used elsewhere (e.g. Brierley and Fryirs, 2005; Dufour and Piégay, 2009), or can become increasingly quantitative and process-based dependant on the complexity of the system, and the specific systems understanding requirements of proposed management schemes (see Small and Doyle, 2012). As
such, an inclusive approach has a wide breadth to combine as many or as few types of data as are deemed necessary to answer the questions proposed by the specific site, making this approach applicable to a wide range of situations.

9.4 FUTURE WORK

The inherent diversity of the Tongariro lends itself to a range of future research projects that could further build understanding of the processes and their management implications. This section briefly summarises future work that would benefit the management and understanding of processes within this catchment.

This thesis did not quantify the proportion of different lithologies in the Lower Tongariro River to determine the extent to which bars were made up of andesitic material from the volcanoes or greywacke from the Kaimanawa Ranges. This form of analysis would have allowed direct quantification of which sub-catchments are most responsible for making up the bed material in the Lower Tongariro River. This thesis assesses this question based on the size of material coming from different sub-catchments, due to the much larger bedload delivered from the volcanic sub-catchments. However, this assertion has not been quantified. Additional information regarding abrasion rates of the different sediment types could offer insight into controls on downstream fining and sediment dynamics within this geological complex environment.

This thesis did not consider channel adjustment at local spatial and temporal scales. For this reason, further work on the evolution of the braided reach could be beneficial to understanding how to understand and manage controls upon channel geometry and planform in this section (e.g. Eaton and Church, 2004; Eaton et al., 2010). An attempt to map bedload transport paths within this reach was made within this study. However, flows during the study period were insufficient to mobilise the bed, and thus this section was not completed. Further work using aDcps to gather data about the volumes and paths of bedload transport (Rennie and Church, 2010), combined with analysis of how vegetation is shaping channel adjustment at the local, short time scale (Corenblit et al., 2007; Gilvear et al., 2007; Gurnell, 2012) would be especially beneficial to gain more evidence for predicting how this active sub-reach is likely to adjust in the next few decades. As this is close to the town of Turangi, gaining more in-depth information about the likelihood of planform adjustment impacting the town could be important for the future management of this localised reach which has been shown to be particularly active within this study and has been actively managed in the past. In particular, concerns for management of stopbanks in relation to sediment flux, and appraisal of avulsion history (and likelihood of future events) are important considerations.
The natural complexity of processes within the Tongariro catchment is mirrored by a complexity in stakeholders and land and water management. Local Iwi (Maori tribes) were the original settlers in the area, and changes to the management of the lake levels and the increase in the flooding of the delta as the river had infilled has led to a loss of valuable farming land. This would be an ideal location for work on stakeholder engagement and environmental management. The Iwi play a role in managing the river, and have received funding to remove the willows in the delta, whilst the active fisheries group in Turangi have been replanting the reach within town with native vegetation. Understanding more about complex stakeholder interactions, and how science could be better used to inform on the ground management would be beneficial for guiding management elsewhere.

9.5 CLOSING COMMENT

Furthering the management of river systems is dependent upon finding appropriate ways of measuring the geomorphic processes which drive channel change (Beechie et al., 2010; Clarke et al., 2003; Kondolf et al., 2006; Newson and Large, 2006; Wohl et al., 2005). A multi-scalar, process-based, evolutionary trajectory approach offers considerable insight into the mechanics of a specific system. Progressing river management agendas relies on firstly, the generation, and secondly, the uptake of such approaches. The process-based approach developed in this thesis describes how a reach is adjusting and what variables are driving changes. Only when management starts to work with the unique reach specific forms and directions of adjustment will the goal of maintaining sustainable, appropriately functioning systems be realised. Given the massive, continuing degradation of rivers across the globe, such approaches are vital to improving the health of river systems and sustaining their ability to support essential ecosystem services. This thesis argues we need to broaden approaches and the scale at which we measure fluvial processes, creating techniques which transfer these insights for applied use, thereby providing a superior underpinning for proactive management.
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