The Transient Response of Bedrock River Networks to Sudden Base Level Fall

by

Benjamin T. Crosby

B.A. Geology
University of California at Berkeley, 1999

SUBMITTED TO THE DEPARTMENT OF EARTH, ATMOSPHERIC AND PLANETARY SCIENCES IN PARTIAL FULFILLMENT OF THE REQUIREMENTS FOR THE DEGREE OF

DOCTOR OF PHILOSOPHY IN GEOLOGY
AT THE
MASSACHUSETTS INSTITUTE OF TECHNOLOGY

SEPTEMBER 2006

Signature of Author: ______________________________
Department of Earth, Atmospheric and Planetary Sciences
August 31, 2006

Certified by: ______________________________
Kelin X Whipple
Professor of Geology
Thesis Supervisor

Accepted by: ______________________________
Maria T. Zuber
E.A. Griswold Professor of Geophysics and Planetary Science
Department Head

The Transient Response of Bedrock River Networks to Sudden Baselevel Fall

by

Benjamin T. Crosby

Submitted to the Department of Earth, Atmospheric and Planetary Sciences
On August 31, 2006 in Partial Fulfillment of the
Requirements for the Degree of Doctor of Philosophy in Geology

Abstract

Following a change in the factors that determine landscape form, a transient signal of adjustment propagates through the river network, progressively adjusting channels and hillslopes to the new conditions. When conditions favor incision, the rate and mechanism by which this signal propagates throughout the network determines basin response time. This ultimately influences the growth and stability of mountain ranges and the tempo of sediment delivery to depositional basins. As this incision signal propagates through the river network, its upstream extent is often recognized as a discrete, steep convexity in the channel profile, defined here as a knickpoint.

In this thesis, theoretical, numerical, and field-based research techniques are utilized to study the initiation and distribution of knickpoints within fluvial networks. Field observations are derived from the Waipaoa River on the North Island of New Zealand. In the Waipaoa, 236 knickpoints distributed throughout the network define the upstream extent of a large magnitude incision signal initiated ~18,000 years ago. These features, frequently located at the confluence between small drainage area tributaries and trunk-streams, are characterized as near-vertical single or multi-step waterfalls. Though flights of trunk-stream strath terraces document a prolonged incision history, the single step knickpoints upstream of tributary junctions suggest that trunk-stream incision can outpace tributary response, producing fluvial hanging valleys. Variations in knickpoint form observed in non-hanging tributaries demonstrate that many factors, both internal and external to the tributary, ultimately determine knickpoint form.

We use a two dimensional numerical landscape evolution model (CHILD) to test a suite of bedrock channel incision rules to determine whether instantaneous or prolonged base level fall can trigger erosion thresholds. These experiments reveal that for bedrock channel incision rules where the relationship between slope and erosion rate is not positive and monotonic, incision rate can decrease with increasing slope until the incision rate is less than the background uplift rate. This provides a mechanism for hanging valley formation when over-steepened tributaries cannot incise at rates equal to the incision in the mainstem. This suggests that if trunk-stream incision outpaces tributary response, hanging valleys form, thus limiting the upstream propagation of incision signals and ultimately increasing basin response time.

Thesis Supervisor: Kelin X Whipple,
Title: Professor of Geology

Acknowledgements

I am indebted to Kelin Whipple for his guidance and patience throughout my years in graduate school. His capacities to both remain open to my diverse interests while helping me stay the course toward thesis completion seasoned my skills as a multi-tasking research scientist/teacher/spouse/parent. I also want to thank my thesis committee, David Mohrig, Rafael Bras, Leigh Royden and Tom Dunne, for their contributions throughout this project. And, of course, I offer gratitude to the wonderful cohort of graduate, undergraduate and post-doctoral students who not only helped strengthen my skills as a scientist and a teacher, but provided an essential support network that extended well beyond academic affairs. Your names and deeds would fill a document twice the thickness of this thesis. I also owe a heartfelt thanks to my parents, my grandparents and my brother who have always supported my crazy ambitions…even graduate school on the other side of the country! And most importantly, I must acknowledge my family, Cana, Dylan and Wells Crosby, who filled me with all the love and encouragement I needed to complete this project. Being a graduate student parent is no easy task, but I fully acknowledge that being a spouse or child of a graduate student must be ten-fold more difficult. I offer my profound thanks to you all!

Table of Contents

Chapter 1: Introduction

1.1 Motivation

The Earth's topography evolves as a dynamic interface, reflecting the interactions among tectonics, climate and erosion. Where these three processes can be assumed stable over a significant period of time, workers have explored whether topographic form can be used to infer the magnitudes or relative influences of those processes (Hilley and Strecker, 2004; Kirby and Whipple, 2001; Pazzaglia and Brandon, 2001; Roe et al., 2002; Roe et al., 2003; Snyder et al., 2000; Snyder et al., 2003; Willett and Brandon, 2002; Wobus et al., 2006b). This thesis arises out of the recognition that in many settings tectonics and climate are not constant and the frequency of variation in tectonic or climatic forcing often outpaces the catchment's topographic response time (Whipple, 2001). Under this circumstance, the landscape retains a record of the ongoing transient adjustment to the present condition. Transient landscapes offer not only a record of changing tectonic and climatic conditions but also provide one of the only methods of distinguishing between competing models for bedrock river incision (Whipple, 2004). Though the steady state forms of these incision models are indistinguishable, they have distinctly different transient responses that can be used to identify which is most viable in a particular landscape (Whipple and Tucker, 2002). An improved understanding of sediment production during transient adjustment may also aid our interpretation of sedimentary records in depositional basins (e.g. Paola et al., 2001).

In an incising non-equilibrium or transient landscape, the catchment is composed of two separable regions: a downstream region that has begun to respond to the incision signal ('the adjusting region') and an upstream region that retains the pre-disturbance form ('the relict region'). The mobile boundary between these two regions migrates upstream or upslope through time. The fluvial component of this boundary is known as a knickpoint. Most previous experimental and field studies of knickpoint form and function focus on the transmission of the incision signal only in trunk streams (Bennett et al., 2000; Gardner, 1983; Haviv et al., in press; Hayakawa and Matsukura, 2003; Miller, 1991; Seidl et al., 1994; Zaprowski et al., 2001) but some recent work has begun to address knickpoint distribution within fluvial networks (Bigi et al., 2006; Bishop et al., 2005; Weissel and Seidl, 1998; Wobus et al., 2006a). The upstream propagation of the knickpoint and its subsequent distribution of the incision signal through the fluvial network determine basin response time following a change in climatic or tectonic conditions. In this thesis, we pair field observations and numerical modeling to explore what sets the rate and form of basin-wide knickpoint propagation.

1.2 Chapters 2-5

The chapters in this thesis summarize our efforts regarding both topographic analysis from the Waipaoa River on the North Island of New Zealand and numerical modeling inspired by observations from the Waipaoa and other field sites. We selected the Waipaoa as an ideal location to examine the distribution and form of knickpoints because the catchment is actively responding to a pulse of incision initiated ~18,000 years ago (Berryman et al., 2000; Eden et al., 2001). In Chapter 2, we take advantage of this natural experiment by analyzing the network position of 236 knickpoints known to be related to the 18 ka pulse of incision and not located on a resistant lithology or structure. Using this dataset of knickpoint positions within the network, we evaluate how well two end-member models for knickpoint retreat in fluvial networks replicate the observed distribution. The first model assumes that knickpoints travel upstream from the basin outlet at a rate that is a power-law function of their local upstream drainage area. This dynamic model utilizes the drainage area structure of each individual channel's path from the outlet to the knickpoint. This model searches over parameter space to find the prefactor and exponent for the power-law that minimizes the residual between all observed and modeled knickpoint positions along the channel. The second model recognizes that though the knickpoints are distributed throughout the basin, most are at around the same drainage area and located very close to tributary junctions. This model posits that the knickpoint distribution is not a reflection of a continuously propagating, discreet signal, but instead the knickpoint positions are a consequence of a threshold process where knickpoints form near tributary junctions when trunk stream incision outpaces the tributaries' capacity to keep up with the mainstem. This 'threshold' model is not dynamic but simply evaluates the mean drainage area for the entire observed knickpoint population and locates where in each tributary the drainage area is at or below the mean value. The residuals between the two models are compared and we find that though both models perform very well, their residuals cannot be distinguished from one another. We suggest that, based on field evidence, it is unlikely that the incision signal swept upstream as a single step-like feature and that the abrupt, steep form of the knickpoints suggest that fluvial processes are no longer the most effective process driving knickpoint retreat. Based on these observations, we conclude that the 'threshold' model for knickpoint initiation at tributary junctions most appropriately describes the formation of the Waipaoa knickpoints.

In Chapter 3, we further explore the notion that knickpoints are unstable in mainstem channels but form at tributary junctions as a consequence of erosional thresholds. Motivated by further observations of knickpoints hanging at tributary junctions (Wobus et al., 2006a), we pursue whether thresholds fluvial incision process laws explain the observed position of knickpoints at tributary junctions. Utilizing a two-dimensional numerical model for landscape evolution (CHILD, Tucker et al., 2001), we compare the response of initially steady-state landscapes to both instantaneous and prolonged base level fall signals. Each model run evaluates the response using one of four different incision models: (1) a sediment-flux dependent incision model based upon abrasion by saltating bedload (Sklar and Dietrich, 2004), (2) a sediment-flux dependent incision model that accounts for sediment functioning as both armor and tools but does not include saltation dynamics (Parker, 2002), (3) the stream power law for bedrock channels {Howard, 1983 #178} and (4) a simplified transport-limited incision model (Paola et al., 1992). Both theoretical and numerical exploration of these models' response to base level fall reveal that only the sediment-flux dependent incision model that includes saltation dynamics can create permanent hanging valleys at tributary junctions. This behavior results from the fact that the modeled relationship between incision rate and slope is non-monotonic and shaped like a convex hump. In this model, at steep slopes the incision rate can decrease below the background uplift rate, thus creating a hanging valley. Note that any model (for instance a plucking model with a hump) with this same functional relationship between incision rate and slope would exhibit this behavior. Our results do not provide a explicit verification of the specific formulation proposed by Sklar and Dietrich, but rather demonstrate that incision rules that have non-monotonic relations between incision rate and slope can create permanent hanging valleys. The Parker model which has a positive, monotonic relationship between slope and incision rate does create temporary hanging valleys, but because the incision rate does not decrease with increasing slope, permanent hanging valleys are never created. Field observations from the Waipaoa show that most knickpoints are so steep that fluvial processes (abrasion by water and sediment) are no longer effective at eroding the knickpoint face. This observation suggests that, among the models tested, the Sklar and Dietrich model most accurately describes the formation and morphology of knickpoints in the Waipaoa River catchment.

Chapter 4 provides an analysis of field survey data hand-collected from 27 channels that contain knickpoints throughout the Waipaoa River catchment. Data collected include the longitudinal profile and measurements of bankfull width, mean sediment grainsize and percent bed exposure upstream and downstream of the each knickpoint. The goal of this study was to determine if there were observable changes in profile geometry, channel morphology and bed state above and below the knickpoint. We were surprised to find that the differences between the two reaches were less severe than anticipated. Though channel steepnesses are higher in the downstream adjusting reach, the concavities of the two reaches are very close and similar to other measurements in reaches of the Waipaoa that are not associated with knickpoints. Measurements of bankfull width are observed to increase downstream of the knickpoint in contrast to the expectation that the channel would narrow in the confined, incised gorge. We also were surprised to find that the mean grainsize increased between ~300m upstream of the knickpoint to ~400 m downstream of the knickpoint. We had anticipated coarser bed material downstream of the knickpoint where the oversteepened gorge walls would be shedding material into the channel, but had not anticipated that there would be a trend in the upstream reach as well. Measurements of percent bed exposure were highly variable along the channel, but we found a discrete increase in bed exposure ~150 m upstream and downstream of the knickpoint. Though there are significant trends within the aggregate data from all 27 channels, there is significant variability suggesting that knickpoint form, channel morphology and bed state do not support the idea that knickpoints contain deterministic traits, but rather that each reflects local conditions within the tributary and the unique base level fall signal experienced at the junction with the trunk stream.

Chapter 5 applies the same survey techniques used in Chapter 4 in a focused study of the mainstem Waihuka River and 15 of its tributaries. The Waihuka River, a sub-catchment of the Waipaoa River, is an ideal location to test sources of variability in knickpoint form because the basin has a well constrained incision history at its outlet (Berryman et al., in preparation), a relatively uniform substrate, tributaries of relatively similar size and well exposed mainstem terraces that record the along-stream incision history. Given the above expectations, we anticipated tributaries to share similar morphological traits but instead found that the form knickpoints reflected unique internal (from within the tributary) and external (from the mainstem base level fall signal) influences at each tributary. We found that the magnitude of the bedrock incision signal and degree of lateral mobility in the mainstem were highly variable along-stream. These variations influenced whether tributaries formed hanging valleys at the confluence with the mainstem Waihuka, retreated upstream as a single step or contained a diffuse record of multiple small steps. Though these three knickpoint-types share common morphologies, the forms of the individual knickpoints were found to reflect local hillslope instabilities or slightly harder substrates. In conclusion, we found that even in a discrete field location with a well known (albeit locally constrained) incision history, a high degree of variability in knickpoint morphology persists, suggesting that knickpoints are too sensitive to variations in their internal and external influences to exhibit deterministic behavior.

Together these 4 chapters suggest that in basins that have experienced a discrete, large magnitude incision signal, knickpoints positioned at or just upstream of tributary junctions may result from threshold conditions in the fluvial incision processes responsible for the upstream propagation of the incision signal. By limiting the propagation of the incision signal into minor tributaries by fluvial erosion processes, the response time of the basin is increased. Variations in the form of knickpoints is a observed to be direct consequence of both along stream variations in the base level fall signal and the high sensitivity of knickpoints to factors internal to the tributary such as substrate strength and local sediment availability.

References

Bennett, S. J., Alonso, C. V., Prasad, S. N., and Romkens, M. J. M., 2000, Experiments on headcut growth and migration in concentrated flows typical of upland areas: Water Resources Research, v. 36, no. 7, p. 1911-1922.

Berryman, K., Marden, M., Eden, D., Mazengarb, C., Ota, Y., and Moriya, I., 2000, Tectonic and paleoclimatic significance of Quaternary river terraces of the Waipaoa River, east coast, North Island, New Zealand: New Zealand Journal of Geology and Geophysics, v. 43, no. 2, p. 229-245.

Berryman, K., Marden, M., Palmer, A. S., Wilson, K., Mazengarb, C., and Litchfield, N., in preparation, The post-glacial downcutting history in the Waihuka tributary of the Waipaoa River, Gisborne District, New Zealand, and implications for tectonics and landscape evolution.

Bigi, A., Hasbargen, L. E., Montanari, A., and Paola, C., 2006, Knickpoints and hillslope failures; interactions in a steady-state experimental landscape: Special Paper - Geological Society of America, v. 398, p. 295-307.

Bishop, P., Hoey, T. B., Jansen, J. D., and Artza, I. L., 2005, Knickpoint recession rate and catchment area: the case of uplifted rivers in Eastern Scotland: Earth Surface Processes and Landforms, v. 30, p. 767-778.

Eden, D. N., Palmer, A. S., Cronin, S. J., Marden, M., and Berryman, K. R., 2001, Dating the culmination of river aggradation at the end of the last glaciation using distal tephra compositions, eastern North Island, New Zealand: Geomorphology, v. 38, p. 133-151.

Gardner, T. W., 1983, Experimental study of knickpoint and longitudinal profile evolution in cohesive, homogeneous material: Geological Society of America Bulletin, v. 94, no. 5, p. 664-672.

Haviv, I., Enzel, Y., Whipple, K. X., Zilberman, E., Stone, J., Matmon, A., and Fifield, K. L., in press, Amplified erosion above waterfalls and oversteepened bedrock reaches: Journal of Geophysical Research – Earth Surface.

Hayakawa, Y., and Matsukura, Y., 2003, Recession rates of waterfalls in Boso Peninsula, Japan, and a predictive equation: Earth Surface Processes and Landforms, v. 28, no. 6, p. 675-684.

Hilley, G. E., and Strecker, M. R., 2004, Steady state erosion of critical Coulomb wedges with applications to Taiwan and the Himalaya: Journal of Geophysical Research, B, Solid Earth and Planets, v. 109, p. no.1, 17.

Kirby, E., and Whipple, K., 2001, Quantifying differential rock-uplift rates via stream profile analysis: Geology (Boulder), v. 29, no. 5, p. 415-418.

Miller, J. R., 1991, The Influence of Bedrock Geology on Knickpoint Development and Channel-Bed Degradation Along Downcutting Streams in South-Central Indiana: Journal of Geology, v. 99, no. 4, p. 591-605.

Paola, C., Heller, P. L., and Angevine, C. L., 1992, The large-scale dynamics of grain-size variation in alluvial basins, 1: Theory: Basin Research, v. 4, p. 73-90.

Paola, C., Mullin, J., Ellis, C., Mohrig, D. C., Swenson, J. B., Parker, G. S., Hickson, T., Heller, P. L., Pratson, L., Syvitski, J., Sheets, B., and Strong, N., 2001, Experimental stratigraphy: GSA Today, v. 11, no. 7, p. 4-9.

Parker, G., 2002, Somewhat less random notes on bedrock incision: Internal Memorandum 118, St. Anthony Falls Laboratory, University of Minnesota.

Pazzaglia, F. J., and Brandon, M. T., 2001, A fluvial record of long-term steady-state uplift and erosion across the Cascadia forearc high, western Washington State: American Journal of Science, v. 301, no. 4-5, p. 385-431.

Roe, G. H., Montgomery, D. R., and Hallet, B., 2002, Effects of orographic precipitation variations on the concavity of steady-state river profiles: Geology (Boulder), v. 30, no. 2, p. 143-146.

-, 2003, Orographic precipitation and the relief of mountain ranges: Journal of Geophysical Research, v. 108, no. B6, p. 2315, doi:10.1029/2001JB001521.

Seidl, M. A., Dietrich, W. E., and Kirchner, J. W., 1994, Longitudinal Profile Development into Bedrock - an Analysis of Hawaiian Channels: Journal of Geology, v. 102, no. 4, p. 457-474.

Sklar, L. S., and Dietrich, W. E., 2004, A mechanistic model for river incision into bedrock by saltating bed load: Water Resources Research, v. 40, p. no.6, 21.

Snyder, N. P., Whipple, K. X., Tucker, G. E., and Merritts, D. J., 2000, Landscape response to tectonic forcing; digital elevation model analysis of stream profiles in the Mendocino triple junction region, Northern California: Geological Society of America Bulletin, v. 112, no. 8, p. 1250-1263.

-, 2003, Channel response to tectonic forcing: field analysis of stream morphology and hydrology in the Mendocino triple junction region, northern California: Geomorphology, v. 53, no. 1-2, p. 97-127.

Tucker, G. E., Lancaster, S. T., Gasparini, N. M., and Bras, R. L., 2001, The channel-hillslope integrated landscape development model (CHILD), *in* Harmon, R. S., and Doe, W. W. I., eds., Landscape Erosion and Evolution Modeling: New York, Kluwer Academic/Plenum Publishers, p. 349-388.

Weissel, J. K., and Seidl, M. A., 1998, Inland Propagation of Erosional Escarpments and River Profile Evolution Across the Southeast Australian Passive Continental Margin, *in* Tinkler, K. J., and Wohl, E. E., eds., Rivers Over Rock: Fluvial Processes in Bedrock Channels: Washington, D. C., American Geophysical Union, p. 189-206.

Whipple, K. X., 2001, Fluvial landscape response time: How plausible is steady-state denudation? American Journal of Science, v. 301, p. 313-325.

-, 2004, Bedrock Rivers and the Geomorphology of Active Orogens: Annual Review of Earth and Planetary Sciences, v. 32, p. 151-185.

Whipple, K. X., and Tucker, G. E., 2002, Implications of sediment-flux-dependent river incision models for landscape evolution: Journal of Geophysical Research, Solid Earth, v. 107, no. B2, p. 3-1 - 3-20.

Willett, S. D., and Brandon, M. T., 2002, On steady states in mountain belts: Geology (Boulder), v. 30, no. 2, p. 175-178.

Wobus, C. W., Crosby, B. T., and Whipple, K. X., 2006a, Hanging valleys in fluvial systems: Controls on occurrence and implications for landscape evolution: Journal of Geophysical Research – Earth Surface, v. Vol. 111, no. No. F2, p. F02017, 10.1029/2005JF000406.

Wobus, C. W., Whipple, K. X., Kirby, E., Snyder, N. P., Johnson, J., Spyropolou, K., Crosby, B. T., and Sheehan, D., 2006b, Tectonics from topography: Procedures, promise and pitfalls, *in* Willett, S. D., Hovius, N., Brandon, M.T., and Fisher, D. M., ed., Tectonics, Climate and Landscape Evolution: Geological Society of America Special Paper 398: Penrose Conference Series, p. 55-74.

Zaprowski, B. J., Evenson, E. B., Pazzaglia, F. J., and Epstein, J. B., 2001, Knickzone propagation in the Black Hills and northern High Plains: A different perspective on the late Cenozoic exhumation of the Laramide Rocky Mountains: Geology, v. 29, no. 6, p. 547-550.

ARTICLE IN PRESS

+ MODEL

ELSEVIER

Available online at www.sciencedirect.com

ScienceDirect

Geomorphology xx (2006) xxx–xxx

GEOMORPHOLOGY

www.elsevier.com/locate/geomorph

Knickpoint initiation and distribution within fluvial networks: 236 waterfalls in the Waipaoa River, North Island, New Zealand

Benjamin T. Crosby *, Kelin X Whipple [1]

Department of Earth, Atmospheric and Planetary Sciences, Massachusetts Institute of Technology, Cambridge, MA 02139-4307, USA

Received 13 July 2004; received in revised form 14 January 2005; accepted 3 June 2005

Abstract

If knickpoints transmit signals of base level fall in river networks, then improvements in our understanding of their retreat rate and basin wide distribution helps constrain the transient response following perturbation. Many studies of knickpoint retreat focus on the response of trunk streams to base level fall. Here we examine the response of an entire fluvial network, as recorded by 236 active knickpoints distributed within the Waipaoa River on the North Island of New Zealand. Base level fall within the Waipaoa catchment initiated 18,000 years ago in response to a climatically triggered and tectonically exacerbated pulse of incision. Using observations from field work, aerial photo analysis and a digital elevation model (DEM), we study the knickpoint positions within the network. We find that ~70% of the knickpoints are located at drainage areas between 1×10^5 m^2 and 1×10^6 m^2 and more than half are <1 km upstream of a large change in drainage area. For the knickpoints <1 km upstream of large tributary junctions, we find that the retreat distances were well correlated with the tributaries' drainage areas. In order to determine how a pulse of incision distributes itself throughout a fluvial network, we develop two simple, end-member models and compare their behavior to the observed knickpoint distribution in the Waipaoa. In the first model, we propose that a knickpoint initiated at the basin outlet retreats upstream and distributes the signal throughout the network at a rate that is a power law function of drainage area. In the second model, we propose that knickpoints form near a threshold drainage area, below which channels cannot incise with the same efficiency as possible in downstream reaches. Though neither model addresses the along-stream variability in substrate or knickpoint form, the misfit between the modeled and the observed knickpoints' along-stream positions are surprisingly low (<1 km; ~1.25% of the total stream length) for knickpoints with drainage areas $<1 \times 10^6$ m^2. Large misfits (up to 3.5 km) are observed for knickpoints with present-day drainage areas greater than 1×10^6 m^2. The large, single step in channel elevation that characterizes knickpoints presently observed in tributaries of the Waipaoa River may not characterize the base level fall signal that propagated through the trunk streams. Evidence for progressive (rather than instantaneous) incision in the trunk streams, the knickpoints' vicinities to tributary junctions and the equivalent success of the two-end member models lead us to conclude that the present positions of the 236 observed knickpoints are largely a consequence of thresholds in channel incision at low drainage areas.
© 2006 Elsevier B.V. All rights reserved.

Keywords: Knickpoint; Waipaoa River; Bedrock incision; Fluvial network; Transient phenomena; Waterfall

* Corresponding author. Now at: Department of Geosciences, Idaho State University, Pocatello, ID 83209-8072, USA. Tel.: +1 208 282 2949; fax +1 208 282 4414.
E-mail address: crosbenj@isu.edu (B.T. Crosby).
[1] Now at: Department of Geological Sciences, Arizona State University, Tempe, AZ 85287-1404.

0169-555X/$ - see front matter
doi:10.1016/j.geomorph.2005.08.023

GEOMOR-01949; No of Pages 23

1. Introduction

1.1. Transient adjustment in fluvial networks

River longitudinal profiles of both alluvial and bedrock channels tend toward a form which best facilitates transport of the sediment load and erosion of the bed (Mackin, 1948; Hack, 1957; Howard et al., 1994; Whipple and Tucker, 1999). Following a change in boundary conditions, the efficiency of erosion or the efficiency of sediment transport, the channel adjusts to a different form appropriate for the new conditions. In studies of bedrock rivers using field, experimental and modeling techniques, most workers have examined the adjustment of the trunk stream to new conditions (Brush and Wolman, 1960; Holland and Pickup, 1976; Gardner, 1983; Seidl and Dietrich, 1992; Seidl et al., 1994; Sklar and Dietrich, 1998; Stock and Montgomery, 1999; Whipple and Tucker, 2002). Though modeling studies have made significant progress toward understanding how this transient adjustment extends throughout an entire fluvial network (e.g. Howard, 1994; Moglen and Bras, 1995a,b; e.g. Tucker and Slingerland, 1997; Crosby, 2006, Ch. 2), few have examined this aspect of the problem using laboratory or field techniques (Weissel and Seidl, 1998; Hasbargen and Paola, 2000; Bishop et

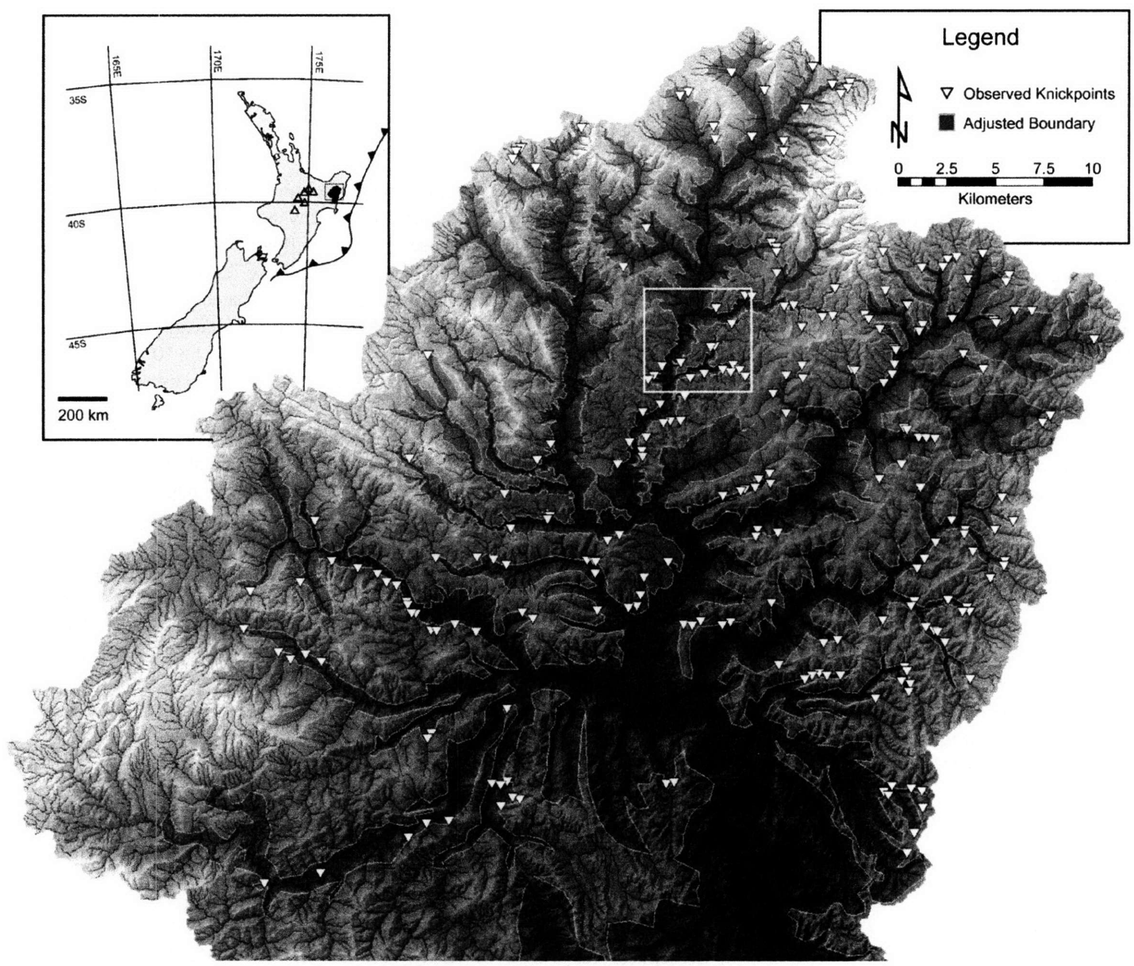


Fig. 1. Location map for the northern two thirds of the Waipaoa drainage basin. On the inset regional map (upper-left), the Waipaoa catchment is located on the northeastern coast of the North Island of New Zealand, ~150 km east of the central North Island volcanic centers (triangles) and ~100 km west of the Hikurangi subduction trench. In the catchment view (central image), 236 knickpoints (inverted triangles) are identified along a mobile boundary that distinguishes which portions of the landscape have adjusted to a pulse of incision initiated 18,000 years ago (dark overlay with thin white outline) and which portions retain their relict, pre-disturbance form. The white square locates Fig. 3.

al., 2005). The rate of this basin-wide transient response determines the landscape response time to external forcing, the history of sediment delivery to offshore depositional centers and ultimately mediates the dynamic coupling between tectonics, climate, and erosion — a coupling that is hypothesized not only to affect landscape form but to strongly influence orogen evolution (e.g. Beaumont et al., 1992; Koons, 1995; Willett, 1999). To better understand how bedrock river networks respond to a change in the conditions that govern their form, we study the actively adjusting, transient landscape of the Waipaoa River drainage basin on the North Island of New Zealand Fig. 1).

As adjustment following perturbation is not contemporaneous throughout the basin but rather migrates upstream (Howard, 1994; Tucker and Slingerland, 1994), a mobile boundary develops, that separates portions of the landscape that are adjusting to the new conditions from those that retain their relict, pre-disturbance form. Although many analyses have purposefully considered the idealized and useful concept of topographic steady state (in which erosion balances rock uplift) (Adams, 1985; Willgoose et al., 1991; Howard et al., 1994), it is likely that most landscapes through geologic time have existed in a state of constant perturbation away from steady state conditions. One of the difficulties in understanding the transient response of fluvial systems arises from the diversity of perturbations capable of initiating similar responses. Some perturbations are regional, such as where a long term change in bedrock uplift rate or climate requires the landscape to assume a new form. Other perturbations are local, such as base level fall and do not require the landscape to assume a new form, but rather to just adjust the present form to a new elevation. A rapid change in base level can occur due to surface rupture on faults, changes in sea level, stream capture or incision rate contrasts at the junctions of tributary streams. Whether the perturbation is regional or local, during the transient response the network of upstream channels can change their elevation, slope, width, or bed state in order that the final form is adjusted to the new conditions. We use the term 'bed state' to encompass bed morphology, bed roughness, grain size distribution and the thickness and percent of alluvium covering the channel bed (Whipple, 2004).

1.2. Knickpoints as transient features

We will define the fluvial portion of the transient boundary between adjusting and relict topography as a knickpoint (Fig. 2). These enigmatic features provided early examples of the dynamic response of bedrock rivers (Gilbert, 1896; Penck, 1924; Davis, 1932) and continue to offer compelling evidence for disequilibrium conditions in field, experimental and theoretical studies. Despite the frequency with which knickpoints are discussed in the literature, fundamental questions regarding their origin, mobility and form remain unanswered. Some of this difficulty results from the diversity of landforms and processes implied by the term. The term knickpoint can be used morphologically to describe an abrupt change in river gradient (Whipple and Tucker, 1999), but in this work we will use the 'classic' definition where a knickpoint describes a discrete, steep reach

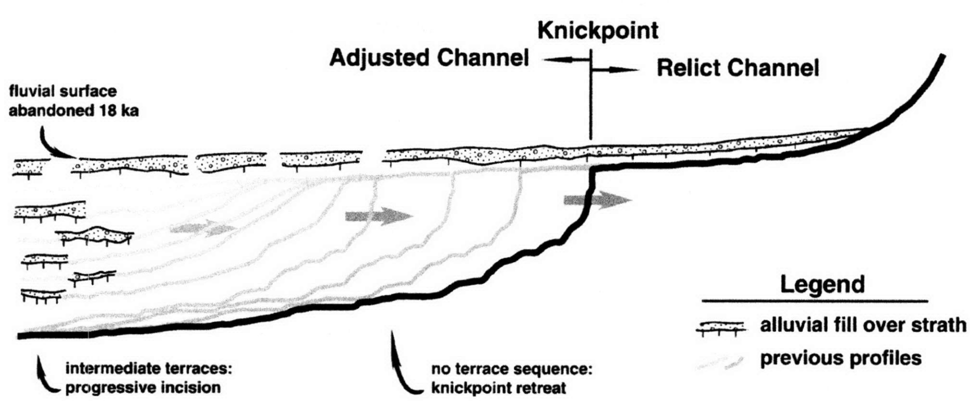


Fig. 2. Schematic illustration of a typical knickpoint on a long profile of a tributary of the Waipaoa River. The knickpoint defines the mobile boundary between the adjusting and relict portion of the channel. The 18 ka discontinuous strath terrace surface is shown, as well as strath terraces formed during progressive vertical incision. In reaches where base level lowering is achieved through knickpoint propagation, no intermediate terraces are created. Grey sequential channel profiles show two potential ways that incision may occur along stream. The grey profiles also demonstrate how these different styles of adjustment may influence the channel's ability (or inability) to generate terrace surfaces. Arrows denote the direction of knickpoint retreat.

which creates a local convexity in the generally concave-up equilibrium channel profile (Fig. 2). These landforms can range from high gradient rapids to waterfalls.

In many field studies, the upstream retreat of knickpoints has been suggested as the dominant mode of channel adjustment in response to either regional or local perturbation (Gilbert, 1896; Holland, 1974; Wolman, 1987; Seidl and Dietrich, 1992; Wohl, 1993; Seidl et al., 1994; Weissel and Seidl, 1998; Stock and Montgomery, 1999; Zaprowski et al., 2001; Haviv et al., 2003; Hayakawa and Matsukura, 2003; Bishop et al., 2005). Multiple mechanisms have been suggested for knickpoint retreat. In regions where the substrate is horizontally stratified or a resistant cap-rock prevents erosion of the knickpoint lip, circulation of water and sediment in a plunge pool at the base of the drop undercut the lip and the step retreats upstream (Gilbert, 1896). In other settings, groundwater seepage along a permeability contrast is suggested to accelerate weathering and facilitate erosion by undercutting the knickpoint lip (Laity and Malin, 1985; Dunne, 1990). Others have suggested that the impact of water and sediment against the bedrock base and steep face of the knickpoint drive retreat (Miller, 1991; Hayakawa and Matsukura, 2003). The stream power erosion model (Howard and Kerby, 1983) has also been applied to describe knickpoint retreat and provides the most popular, quantitative tool for comparing the form and position of modeled and observed knickpoints (Rosenbloom and Anderson, 1994; Seidl et al., 1994; Weissel and Seidl, 1998; Stock and Montgomery, 1999).

Experimental studies of knickpoints demonstrate that the mechanical properties of the substrate (homogeneous, stratified, jointed, cohesive vs. noncohesive) influence whether the knickpoint face progressively decreases gradient (typically noncohesive) (Brush and Wolman, 1960; Begin et al., 1981; Gardner, 1983), or whether the knickpoint migrates upstream with varying degrees of incision upstream of the lip and/or gradient reduction on the face (typically cohesive) (Holland and Pickup, 1976; Gardner, 1983). Retreat without substantial modification of the knickpoint form has only been observed under the special circumstance where either pervasive and extensive bedrock jointing exists or a strongly cohesive cap-rock prevents erosion of the knickpoint lip (Holland and Pickup, 1976; Stein and LaTray, 2002). Gardner (1983), recognized that under certain conditions, as the flow approaches the free overflow, it accelerated and enhanced erosion of the knickpoint lip. The region over which this enhanced erosion occurred was referred to as the draw-down reach. Another experimental study using a non-cohesive, homogeneous substrate observed autogenic knickpoint-like waves of base level fall migrating throughout a channel network (Hasbargen and Paola, 2000). Another flume-scale experimental study examined knickpoint retreat across vertically oriented beds of variably cohesive substrate (Frankel et al., 2001). They found that the knickpoint diminished in height when translating across the weakly cohesive sections, then reformed, migrated and diminished again as it passed through highly cohesive sections. Though experimental studies offer workers the opportunity to observe the time-evolution of a transient form, there are numerous scaling concerns that indicate that the rate-limiting erosion process at the flume-scale may be very different from those occurring in the field.

Theoretical or modeling studies of knickpoints provide workers with both the capacity to observe the time-evolution of the transient response and to effectively test the influence of different variables on the result. Studies of knickpoint behavior in systems where erosion is limited by the channel's capacity to transport the imposed sediment load (transport limited) are typically analogous to experimental studies in noncohesive substrates where the knickpoint face progressively decreased gradient (Brush and Wolman, 1960; Begin, 1988). If these knickpoints migrate upstream, the initial step in elevation could be described as diffusing, decaying, degrading or rotating as it migrates upstream. Studies of knickpoint behavior in systems where erosion is limited by the channel's capacity to incise into bedrock (detachment limited) are typically analogous to experimental studies in cohesive substrates where the knickpoint migrated upstream with varying degrees of reduction in height and/or gradient. Detachment limited channels are often described by the stream power incision model (Howard and Kerby, 1983), where incision rate along the channel is a power function of both upstream drainage area and local slope. Using this model, others have derived the celerity, or wave speed of a knickpoint (Rosenbloom and Anderson, 1994; Weissel and Seidl, 1998; Whipple and Tucker, 1999). Knickpoint celerity is a power law function of both upstream drainage area and local slope. The propagating knickpoint's form is invariant if the celerity is not dependent on slope (e.g., the slope exponent is zero).

Generalizing the field, experimental and theoretical findings above, we find that if the steep slope of the knickpoint face enhances erosion, then the local convexity migrates upstream (Fig. 2). The expectation that a migrating step persists from the point of initiation throughout the fluvial network is complicated by two concerns. First, the conditions that set the rate of retreat and form of the migrating knickpoint change as the knickpoint retreats upstream. For example, water and sediment discharge decrease stepwise at tributary junctions and substrate erodability often varies within a drainage basin. The second concern is that steps in channel elevation are unstable and are rapidly modified. If knickpoints are susceptible to modification as they translate upstream, then caution must be used when associating steps in fluvial profiles with basin-wide, upstream migrating transients. In addition, non-transient, static knickpoints can be fixed in space by local decreases in bed erodability such as an erosion-resistant intrusive body or coarse sediment armoring downstream of a tributary junction. These static knickpoints must be distinguished before the distribution of transient knickpoints is analyzed. Provided the controls on knickpoint form and retreat rate can be determined, the distribution of transient knickpoints may provide insight into how pulses of incision are communicated throughout fluvial networks.

1.3. Our approach

The landscape in the Waipaoa River catchment (Fig. 1), records the transient response following a large magnitude episode of rapid incision (Berryman et al., 2000; Eden et al., 2001). Incision began ca 18 ka and initiated a pulse of incision that extended into all large tributaries of the Waipaoa. The present position of this headward advancing wave of incision is marked in many tributaries by a knickpoint, often in the form of a waterfall (Fig. 1). Field observation and historical aerial photo analysis offers clear evidence for headward migration of these knickpoints, at least over short distances ($\sim$1 km). This leads us to first test a simple model of knickpoint propagation that assumes knickpoint retreat rate is a power law function of drainage area. This simple model assumes that a knickpoint initiated at the basin outlet swept upstream and branched into tributaries, without substantially changing form or incision process. However, there is reason to expect more complex behavior. For instance, if mixed bedrock–alluvial channels are increasingly alluvial in character at large drainage areas as suggested by some models and data (see review by Whipple, 2004), one might anticipate diffusive attenuation of knickpoints in downstream channel segments, and an increasing tendency to form coherent, detachment-limited bedrock knickpoints as the wave of incision reaches upstream channel segments. Moreover, one might anticipate changes in the dominant processes of incision and knickpoint retreat as drainage area decreases upstream. Indeed, abrupt changes in drainage area, water discharge, and sediment flux at tributary junctions might trigger knickpoint initiation. This behavior has been suggested by theoretical river incision models that include a dual sediment flux dependence (Gasparini, 2003; Crosby, 2006, Ch 2; Gasparini et al., 2006), such as the saltation–abrasion model of Sklar and Dietrich (1998, 2001, 2004). In these models, the efficiency of bedrock incision increases with additional sediment flux up until the point that the increasing sediment no longer functions as tools to abrade the bed, but rather covers and armors the bed from further incision. In addition, there is some direct evidence (discussed below) against a simple basin-wide propagation of knickpoints in the Waipaoa. In order to test the possibility that knickpoints develop high in the network (at small drainage areas or at tributary junctions), we also evaluated a second, alternative model that simply predicts knickpoint formation at a threshold drainage area. We compare the results of these two, simple, end-member models against the present distribution of 236 knickpoints observed within the Waipaoa fluvial network.

2. Field area

2.1. Geologic setting

The 2150 km^2 Waipaoa drainage basin is etched into forearc sediments accreted along the east coast of New Zealand's North Island (Fig. 1). Along this subduction margin, the thinned, leading edge of the Australian plate obliquely overrides the west-dipping and overthickened portion of the Pacific Plate called the Hikurangi Plateau (Davy, 1992). As the two plates converge at a rate of $\sim$45 mm/year (DeMets et al., 1994), seamounts entering the trench deform and destabilize the trench slope suggesting that the oceanic plate topography and thickness strongly influence the time evolution of wedge deformation (Lewis and Pettinga, 1993; Collot et al., 2001; Eberhart-Phillips and Chadwick, 2002). In the Waipaoa drainage basin, the exposed forearc is composed of latest Cretaceous to Pliocene shelf and slope sediments. These are dominated by clay-rich siltstone and mudstone units, interbedded with sandstone and limestone lenses (Black, 1980; Mazengarb and Speden, 2000). Due to different structural histories (sheared argillite to undeformed sandstone) and different chemical compositions (clay rich mudstones to calcite cemented sandstones) the bedrock of the Waipaoa drainage basin exhibits significant variability in strength.

Faulting and folding of the exposed forearc sediments is of late-Miocene to recent age (Mazengarb et al., 1991; Mazengarb and Speden, 2000). Active structures in the region have a dominantly normal sense, but low preservation potential for fault scarps in this rapidly eroding landscape suggests that available field observations likely sample only a small fraction of the recently active faults (Berryman et al., 2000). Rock uplift rates in the region are reported between 0.5 and 4 mm/yr using the near-shore record (Ota et al., 1992; Brown, 1995), river terrace sequences (Berryman et al., 2000) and summit accordance along the crest of the Raukumara Range (Yoshikawa, 1988). In the Waipaoa itself, analysis of strath river terrace sequences suggested that the rock uplift rate ranged between 0.5 and 1.1 mm/yr (Berryman et al., 2000). Located 100–200 km east and downwind of the central North Island volcanic centers (Fig. 1), the Waipaoa drainage basin has received frequent deposits of airfall tephra throughout the Quaternary (Eden et al., 2001). As will be discussed in the proceeding section, tephra deposits of known age are found on terraces or other stable landforms and provide valuable stratigraphic markers (Berryman et al., 2000).

2.2. Present landscape morphology

At present, the Waipaoa drainage basin is actively adjusting its form following a large magnitude pulse of incision initiated 18,000 years ago (Berryman et al., 2000; Eden et al., 2001; Berryman et al., in preparation). The prevalence of retreating knickpoints, incised gorges, unstable hillslopes and fill and strath terraces associated with this pulse of incision provides dramatic evidence for the dynamic response of this landscape (Fig. 3). Because the pulse of incision is so large (50 to 120 m), it is straightforward to distinguish the mobile boundary that separates portions of the landscape that are adjusting to the new base level from those that still retain their relict, pre-disturbance form. An initial investigation using ~1:25,000 stereo aerial photographs revealed that >50% of the areal extent of the Waipaoa drainage basin retains this relict topography (Fig. 1). This relict topography is found upstream of the knickpoints and on the hillslopes left hanging above the incised gorges downstream of the knickpoints (Fig. 3). Today, the headward retreat of knickpoints and failure of over-steepened hillslopes continues to adjust the relict surface to the lower base level. Downstream of the knickpoints, terrace remnants of the relict surface either trace continuously to or are graded to the relict surface upstream of the knickpoint (Fig. 2). If knickpoint retreat was the dominant mechanism for propagating base level fall, then most of the incision would occur abruptly as the knickpoints pass; there would be no record of progressive incision (Fig. 2). However, in some large Waipaoa tributaries ($\sim > 5 \times 10^7$ m^2), geomorphic features indicative of progressive fluvial incision exist between the 18 ka relict surface and the present river elevation. These features include flights of strath surfaces and abandoned meander cutoffs (Fig. 2), and have been documented in some detail in the Waihuka tributary. In other reaches of this tributary, there is no evidence preserved indicating progressive incision. In these locations, narrow, incised gorges that terminate at waterfalls suggest that as the knickpoint retreated, base level was lowered step-wise. This record leads us to suggest that a pulse of incision can assume multiple forms as it translates upstream and encounters variations in substrate erodability or decreases in sediment and water flux at locations such as tributary junctions (Fig. 2).

Preservation of the relict surface throughout the basin is sufficient enough to reconstruct the paleotopography prior to incision ~18 ka (Fig. 1). Though the Waipaoa drainage basin was never glaciated, the colder temperatures during the last glacial maximum had a profound effect on the local landscape; vegetation changed from forest to shrub, sediment production increased and precipitation decreased (Pillans et al., 1993; Newnham et al., 1999; Berryman et al., 2000). These conditions are suggested to have produced the thick strath terrace cover-beds and the subdued, sediment mantled topography that we see today in relict portions of the landscape (Fig. 3). Within this relict landscape there exists evidence for multiple older generations of transient adjustment such as terraces and knickpoints that predate the 18 ka surface (Berryman et al., 2000). This strongly indicates that in this drainage basin, fluctuations in climatic or tectonic conditions outpace the landscape's response time for readjusting to the new conditions (Whipple, 2001).

Berryman et al. (2000) suggest that incision was initiated by two consequences of the transition to postglacial climatic conditions ~18 ka. First, as climate became warmer and wetter, more substantial vegetation provided greater soil cohesion and reduced both sediment supplied from the hillslopes and physical weathering. Second, the warmer and wetter conditions increased stream discharge and raised the transport capacity and the erosive potential of the channels. We also note that at the time climate began to change, sea level was >100 m below its present elevation (Shackelton, 1987; Chappell, 2001) and thus a pulse of incision could have initiated as sea level dropped below the shelf-slope break. This could have generated an unstable step in the channel profile that would have propagated toward the headwaters.

On active fluvial surfaces, airfall tephra deposits are only preserved once river incision adjacent to that surface is significant enough to prevent overbank flows from reworking the tephra. The first tephra deposit preserved on the terraces created by the pulse of incision that we study in this work was the Rerewhakaaitu (Eden et al., 2001), which erupted from the Okataina Volcanic Center ~17,700 years before present (Lowe et al., 1999). The previous tephra-producing event, (the Okareka tephra, ~21,000 years before present), is not preserved on the relict fluvial surface, suggesting that the river was actively reworking the surface ~21 ka.

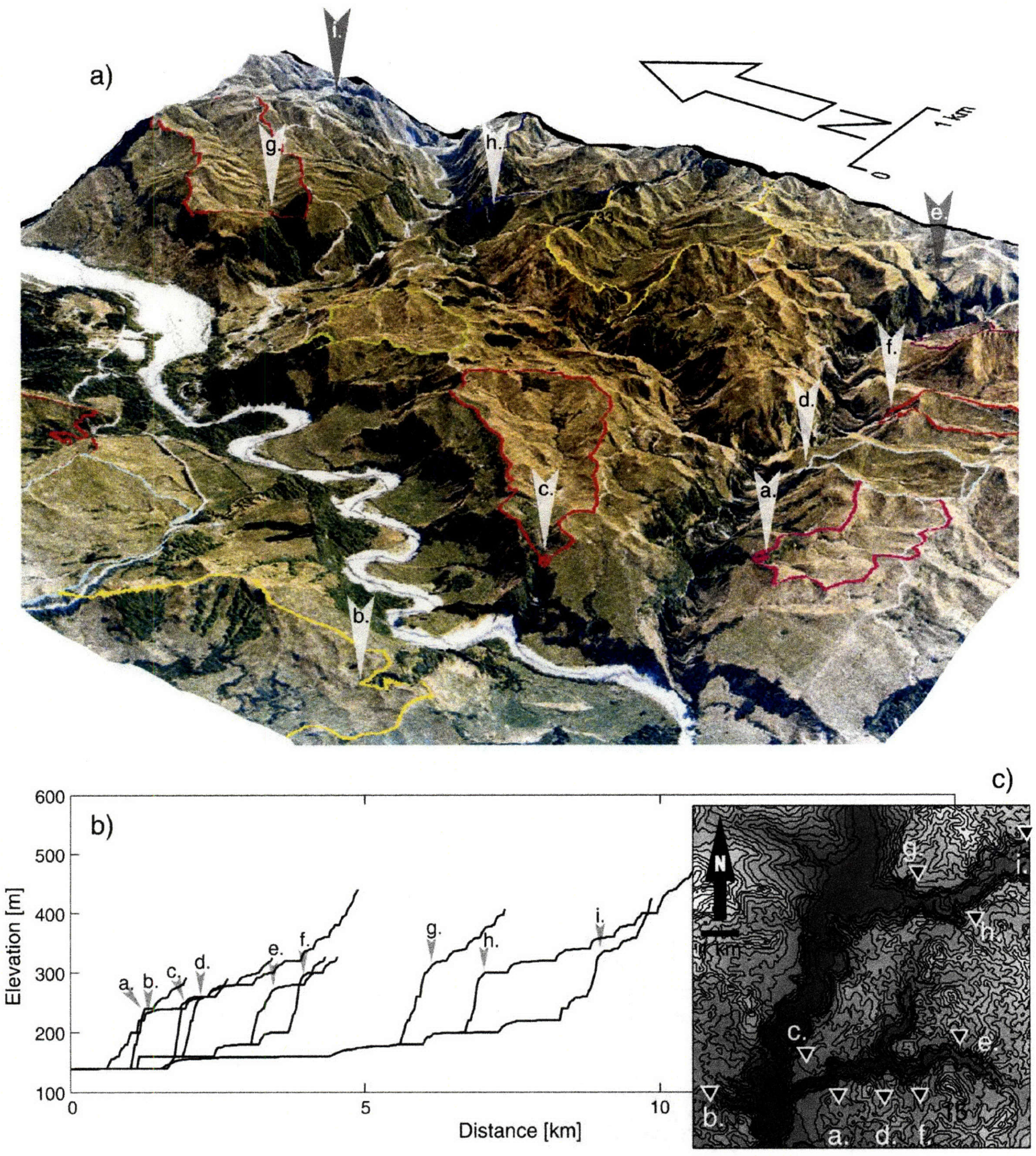
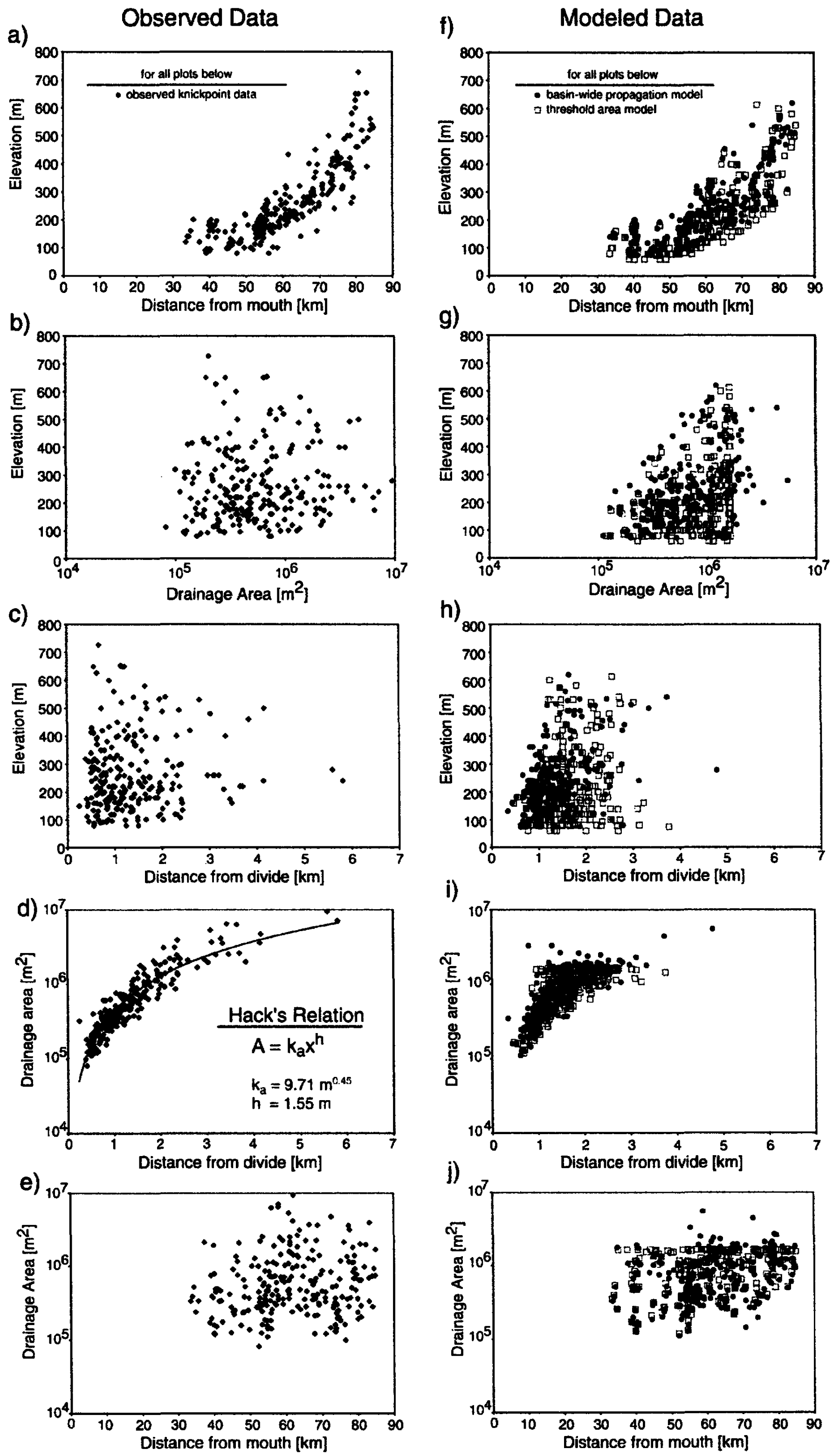


Fig. 3. The topography of the Waipaoa catchment is visualized in a color aerial photograph, (3a), draped over a 25 m digital elevation model (DEM) from the upper catchment (for location see Fig. 1). The mainstem Waipaoa flows downstream, left to right. Note that the two large tributaries entering from the east are adjusted to the elevation of the mainstem, while smaller tributaries are elevated above and separated from the mainstem by knickpoints (located by lettered arrows). The same knickpoints are shown, (3b), along their channel profiles as extracted from the 25 m DEM. Besides the large knickpoints, also note the 20 m convex step-like artifacts in the DEM. A contour map, (3c), of the same region locates the knickpoints in map view (contour interval: 10 m).


Observed Data
Modeled Data
a)
b)
c)
d)
e)
f)
g)
h)
i)
j)
for all plots below
• observed knickpoint data
• basin-wide propagation model
□ threshold area model
Elevation [m]
Distance from mouth [km]
Drainage Area [m2]
Distance from divide [km]
Drainage area [m2]
Hack's Relation
A = kaxh
ka = 9.71 m0.45
h = 1.55 m

This brackets the initiation of incision between ~21 ka and ~18 ka.

2.3. Knickpoint morphology

Despite broad variations in lithology, structure, position within the basin, upstream drainage area and bed state, our observations (collected over five field seasons) indicate that knickpoint morphologies are strikingly similar throughout the Waipaoa drainage basin (Figs. 2 and 3b) (Crosby, 2006, Ch 2). Most trunk streams are low gradient (between 0.003 and 0.01) and alluviated at drainage areas greater than $\sim 5 \times 10^7$ m^2. Farther upstream, bare bedrock reaches become more common as channels' slopes increase to values between 0.02 and 0.1. These bedrock reaches are punctuated by discrete alluvial patches often associated with temporary storage of sediment shed from nearby over-steepened banks. In the first kilometer downstream of knickpoints, most of the elevation gain is achieved over 1–2 m bedrock steps. Less than 500 m downstream of knickpoints, coarse debris begins to litter the channel beds as mass failures become frequent in the narrow, steep-walled gorges (Fig. 3). The knickpoint faces are 15–50 m tall and are most often composed of one or two near-vertical drops. Though not undercut, the faces are so steep that at high flows, water may not make direct contact with the knickpoint face. There are no plunge pools or cap-rocks observed that would maintain the knickpoint lips. We hand-surveyed water surface elevation and noted bed elevation of ~27 channels (Crosby, 2006) and found that knickpoints in the Waipaoa lack drawdown reaches (Gardner, 1983), as introduced earlier. Upstream the knickpoint lip there is no enhanced erosion near the free overflow. Instead, upstream of the knickpoint, the channels are low gradient, sediment poor and have only incised through the alluvial fill of the relict surface and 1–2 m into the underlying bedrock (Fig. 2). Contrary to the expectation for a drawdown reach, this incised morphology is not localized to the region near the knickpoint lip but rather continues for more than a kilometer upstream of many knickpoints. This observation suggests that the initial stripping of sediment was localized within channels and did not extend across the broad valley floors or to the hillslopes. This has implications for the history of sediment delivery during the incisional event. At present, the channel beds above the knickpoints contain very little sediment and their slopes range between 0.008 and 0.02. The morphology of the knickpoint and the relict reach is relatively consistent throughout the basin and suggests that these channels are presently behaving in a systematic fashion following incision in the the trunk streams.

3. Observed knickpoint distribution

3.1. Methods

Multiple observational tools were used to identify and confirm the location and form of knickpoints within the Waipaoa drainage basin (Fig. 1). Though a 25 m digital elevation model (DEM) proved extremely valuable to other aspects of this project, the channel profiles extracted from this dataset could not be used reliably to identify knickpoint form or location because of the artificial 20 m steps inherited from the original contour data (Fig. 3b). Therefore, an accurate mapping of knickpoint location and form was accomplished using aerial photo analysis and field verification. We used color stereo aerial photography flown at a scale of ~1:25,000 between 1997 and 1999 by Air-Maps New Zealand. We analyzed only the northern 3/4 of the basin due to limitations in the aerial photo coverage available to us at the time of the study. This was not a concern because aggradation during sea level rise had buried most of the relict landscape in the southern 1/4 of the Waipaoa basin. Only knickpoints with discrete steps in channel elevation were selected for analysis as it was too difficult to identify the upstream extent of adjustment in diffuse knickpoints. Controls on the variability of knickpoint form and the distribution of these forms will be addressed in future work (Crosby, 2006).

Of the ~350 potential knickpoint locations identified using the aerial photos, ~55 were visited, characterized and located with GPS. This exercise confirmed the accuracy of our air photo technique for locating knickpoints and revealed the striking consistency in knickpoint form as discussed in the previous section. After observed knickpoint locations were transferred onto digitized 1:50,000 topographic maps and incorporated into a GIS database, their spatial distribution could be examined relative to topographic, lithologic and

Fig. 4. Plots comparing channel parameters (elevation, distance from outlet, distance from divide and drainage area) at observed (4a–e) and modeled (4f–j) knickpoints. The presence or lack of a correlation between the parameters is discussed in the text. Note that the both the basin-wide retreat and the threshold area models give similar results and match the observed knickpoint parameters well. In both observed and modeled environments, the lack of correlation between the parameters (except for in the long profile (4a,f) and in Hack's relation (4d,i)), demonstrates how each channel's unique course across the landscape influences knickpoint position.

structural parameters. For every channel containing a knickpoint, we extracted its longitudinal channel profile from the 25 m DEM and measured for each pixel along the profile the along-stream distance from outlet, the along-stream distance from divide, the elevation, the upstream contributing area and the underlying lithology. For a subset of 161 channels we also measured the knickpoint's distance upstream from the closest significant tributary junction.

Approximately 350 potential knickpoints were then filtered to remove those whose form or location appeared to be a consequence of special circumstances. This was necessary in order to study how a single incisional event is distributed throughout the many channels of a fluvial network. First, we separated out the 33 knickpoints that predate the pulse of incision initiated 18,000 years ago and were found within the relict surface. Second, we separated out 76 knickpoints whose locations or morphologies were attributable to their proximity to a lithologic or structural contact. The third population removed from the dataset was 5 composite knickpoints that consisted of multiple, broadly spaced knickpoints. In this population, the farthest downstream knickpoint was often anchored by a resistant lithology, leaving the upper basin to adjust to a base level fall of lesser magnitude. The remaining 236 knickpoints that we examined were: (1) attributed only to the pulse of incision initiated ~18 ka, (2) not lithologically or structurally constrained and (3) composed of a single major knickpoint.

3.2. Results and discussion

The topographic and hydrologic characteristics of the 236 selected knickpoints were examined to determine if their distribution within the basin exhibited systematic behavior (Fig. 4a–e). Knickpoints in the Waipaoa drainage basin were predominantly located in the uppermost regions of the network and were thus found primarily in first or second order blue-line channels. For channels that are roughly 80 km long, the mean distance from the divide was 1369.09 m (Fig. 4c). Normalizing for the different channel lengths in the basin, we calculated that the population of 236 knickpoints was 97.79±1.33% (1σ) of the way from the present outlet to their headwaters. As a consequence of being so far up within the network, knickpoints are found within a narrow range of characteristically low drainage areas. 75% of the knickpoints (176 of the 236) were found to have upstream drainage areas between 1×10^5 m^2 and 1×10^6 m^2 (Fig. 4b). This observation that channel adjustment extends almost to their headwaters seems in conflict with our earlier observation that >50% of topography still retains its relict form (Fig. 1). This apparent conflict arises from the large percentage of the landscape that is composed of unadjusted hillslopes above gorges and in low-order tributary basins. Many channels in the Waipaoa have incised deep, narrow gorges, leaving most of the hillslopes hanging, still graded to the elevation of the 18,000 year old terrace surface (Fig. 3a).

Knickpoint elevations are distributed between 100 and 750 m above sea level (asl) with ~55% of the knickpoints found at elevations between 150 and 300 m asl (Fig. 4a). Theoretical studies of detachment limited channels responding to an increase in uplift rate predict that knickpoints communicating this adjustment should remain at like elevations throughout their passage through the basin (Whipple and Tucker, 1999; Niemann et al., 2001). Instead, in the Waipaoa river we find that the elevations of the knickpoints appear to be a function of where low drainage area reaches are encountered along the individual stream profiles (Fig. 4b).

The upstream drainage area, distance from divide, distance from outlet and elevation of each knickpoint were compared to evaluate whether some systematic relation exists between these parameters (Fig. 4a–e). The correlation between distance from the mouth and elevation only demonstrates where the knickpoints are located on the longitudinal profiles of the channels (Fig. 4a). The correlation between knickpoint drainage area and distance from divide (Fig. 4d) is a consequence of Hack's relation (Hack, 1957) which states that the distance from the divide is a power law function of drainage area. All other comparisons between the measured knickpoint parameters showed no statistically significant correlation (Fig. 4b,c,e).

Though we filtered the knickpoints to remove those retained on lithologic contacts, we wanted to determine whether particular lithologies favored the formation or preservation of knickpoints. To do this, we compared knickpoint location with the mapped geologic units in the basin (Mazengarb and Speden, 2000). First, we measured how often a knickpoint occurs within a particular lithology (ex. 23% of the knickpoints are in the early-Miocene, Tolaga group mudstones, (eMt)). Second, we measured how often a particular lithology occurs within the study area (ex. 22% of the study area is underlain by the eMt). These measurements were used to normalize the frequency of knickpoint occurrence in a particular lithology relative to the areal distribution of that lithology. Some erosion-resistant geologic units were moderately 'over-populated,' having a disproportionately large number of knickpoints relative to their areal extent. Conversely, erosion-prone lithologies (as evidenced by lower slopes, less relict surface

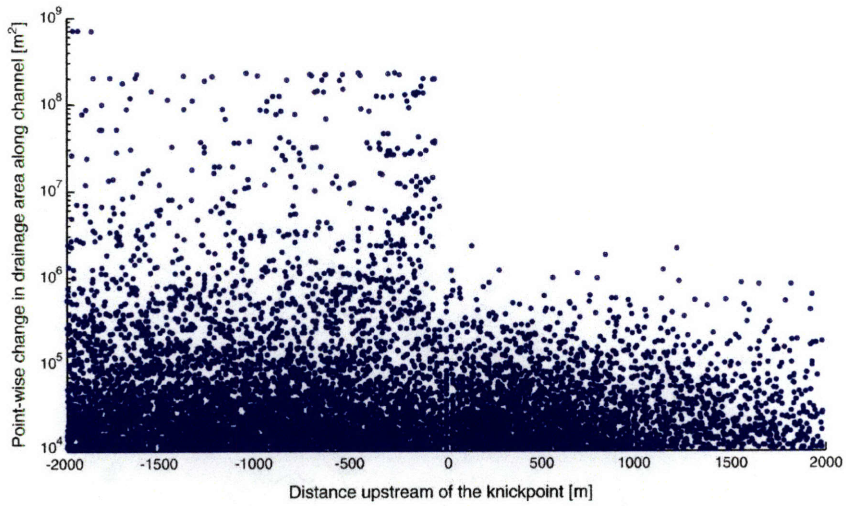


Fig. 5. For each of the 236 knickpoint-containing channels, we examined the point-wise change in drainage area 2 km downstream and upstream of the knickpoint. Each point above the x-axis notes the location of a step-wise increase in drainage area at a tributary junction. The larger the point-wise change in drainage area, the larger the tributary. The most striking aspect of the plot is the abrupt decrease at the knickpoint of the magnitude of the maximum drainage area changes. Upstream of the knickpoint, the maximum drainage area changes decrease by 2 orders of magnitude. Also note the moderately dense cluster of large magnitude drainage area changes 0–500 m downstream of the knickpoints.

preservation and greater structural deformation) were found to be 'under-populated.' This demonstrates that lithology influences but does not determine the distribution and preservation of knickpoints in the Waipaoa.

A large subset ($n=161$) of the 236 knickpoints were further examined to determine if their position in the network was related to their proximity to tributary junctions. We found that ~78% of the knickpoints examined were less than 1 km upstream of large tributary junctions. At these large tributary junctions, drainage areas dropped on average from $\sim 5\times10^7$ m^2 in the trunks to $\sim 1.5\times10^6$ m^2 in the tributaries. This drop was most evident when we plotted the point-wise change in drainage area for each of the 236 knickpoint-containing channels against the distance upstream of the knickpoint (Fig. 5). Looking at the data for reaches 2 km downstream of knickpoints, there are often large changes in drainage area, but in the upstream direction, changes in drainage area are significantly smaller. We then compared the drainage area at the knickpoint to its distance upstream from the tributary junction but found no statistically significant correlation. However, a good correlation was observed between the tributary's drainage area at the junction and the distance the knickpoint had traveled upstream of that junction (Fig. 6). Though this is consistent with the expectation that knickpoints in larger tributaries would have retreated greater distances upstream, such a correlation is also expected if knickpoints simply form at some small threshold drainage area, as a consequence of Hack's law (dashed line on Fig. 6). We will explore this point at greater depth in Section 5.2. In

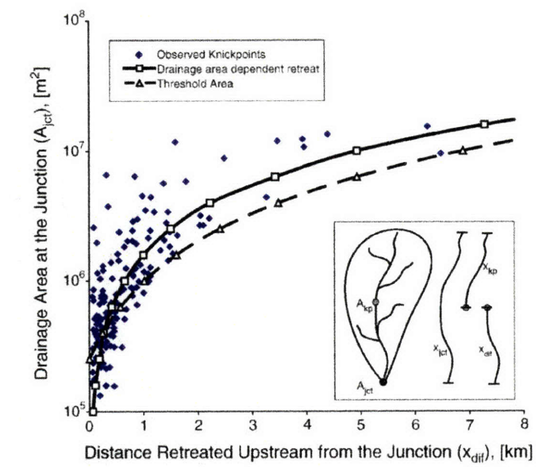
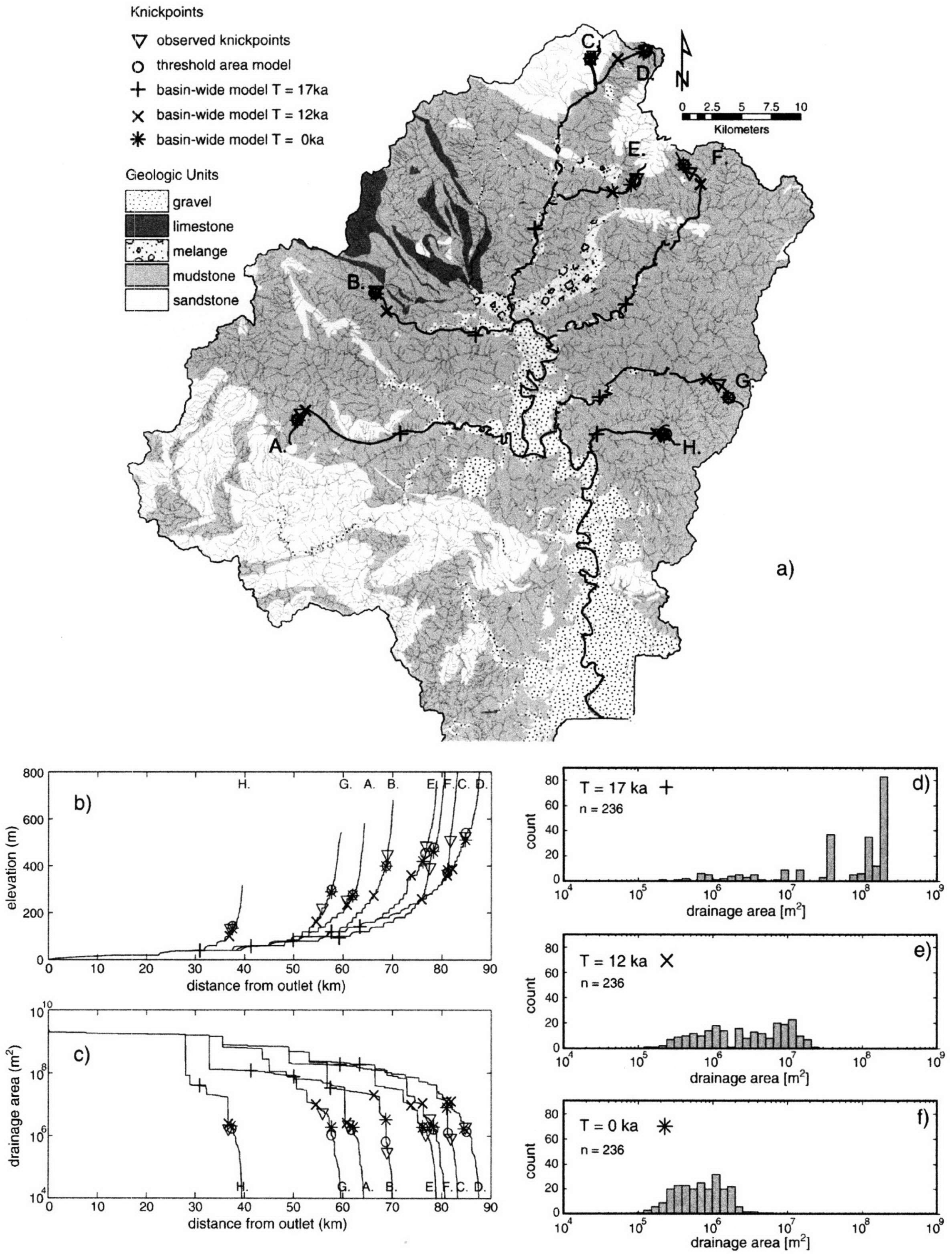


Fig. 6. We observe that knickpoints have retreated farther upstream in tributaries with large drainage areas. As explained further in Section 5.2, this observation could result from two different behaviors. The data suggest that knickpoints could retreat at a rate that is a function of drainage area (solid line with squares) but a similar relation could result from knickpoints forming at a threshold drainage area in basins with a Hack's relation-type geometry (dashed line with triangles). This observation limits our ability to say whether or not knickpoint retreat rate is function of drainage area. The inset figure defines variables used in the discussion section.


Knickpoints
observed knickpoints
threshold area model
basin-wide model T = 17ka
basin-wide model T = 12ka
basin-wide model T = 0ka
Geologic Units
gravel
limestone
melange
mudstone
sandstone
0 2.5 5 7.5 10
Kilometers
A.
B.
C.
D.
E.
F.
G.
H.
a)
b)
elevation (m)
distance from outlet (km)
c)
drainage area (m^2)
distance from outlet (km)
d)
T = 17 ka
n = 236
count
drainage area [m^2]
e)
T = 12 ka
n = 236
count
drainage area [m^2]
f)
T = 0 ka
n = 236
count
drainage area [m^2]

order to explore controls on knickpoint position and gain insight into the processes responsible for their retreat or initiation, we evaluate two simple end-member models motivated by the above discussion of the characteristics and spatial distribution of the observed knickpoints.

4. Modeled knickpoint distributions

Consistencies in the location and form of knickpoints in the Waipaoa drainage basin suggest that their present distribution may reflect some systematic behavior. There is evidence for rapid incision and adjustment to a lower base level throughout larger drainage area portions of the network (Fig. 1), but at present, most knickpoints are found at relatively small drainage areas (Fig. 4b) and provide some evidence for area-dependent migration (Fig. 6). In accordance with these observations, we developed and tested two plausible, end-member models of knickpoint behavior. In the first, we use a forward model to predict how knickpoint positions evolve through time. In this model, a knickpoint initiated at the basin outlet retreats upstream and distributes itself throughout the channel network at a rate that is a function of the local upstream drainage area. In the second model, we do not model knickpoint migration, but instead examine if knickpoints initiate at some threshold drainage area. The second model does not evolve through time or provide information about the rate of response to the perturbation. It instead tests for some stable equilibrium condition where knickpoints would form at low drainage areas and then retreat by a different set of processes that may be independent of drainage area (Weissel and Seidl, 1997; Weissel and Seidl, 1998).

The models treat upstream drainage area as a proxy for both sediment and water discharge, each of which strongly influences the channel's erosional efficiency (Sklar and Dietrich, 1998, 2004). Because both models use drainage area to determine knickpoint positions, their predicted distribution depends strongly on the irregular, step-wise manner that drainage area decreases in the upstream direction in each individual channel. The pattern and frequency with which these abrupt changes in drainage area (Fig. 5) at tributary junctions are encountered ultimately depends on the network geometry and ridge spacing of the basin. For each model, we will discuss the basis for their formulation, explain the mechanics of the model and discuss model results.

4.1. Basin-wide knickpoint propagation

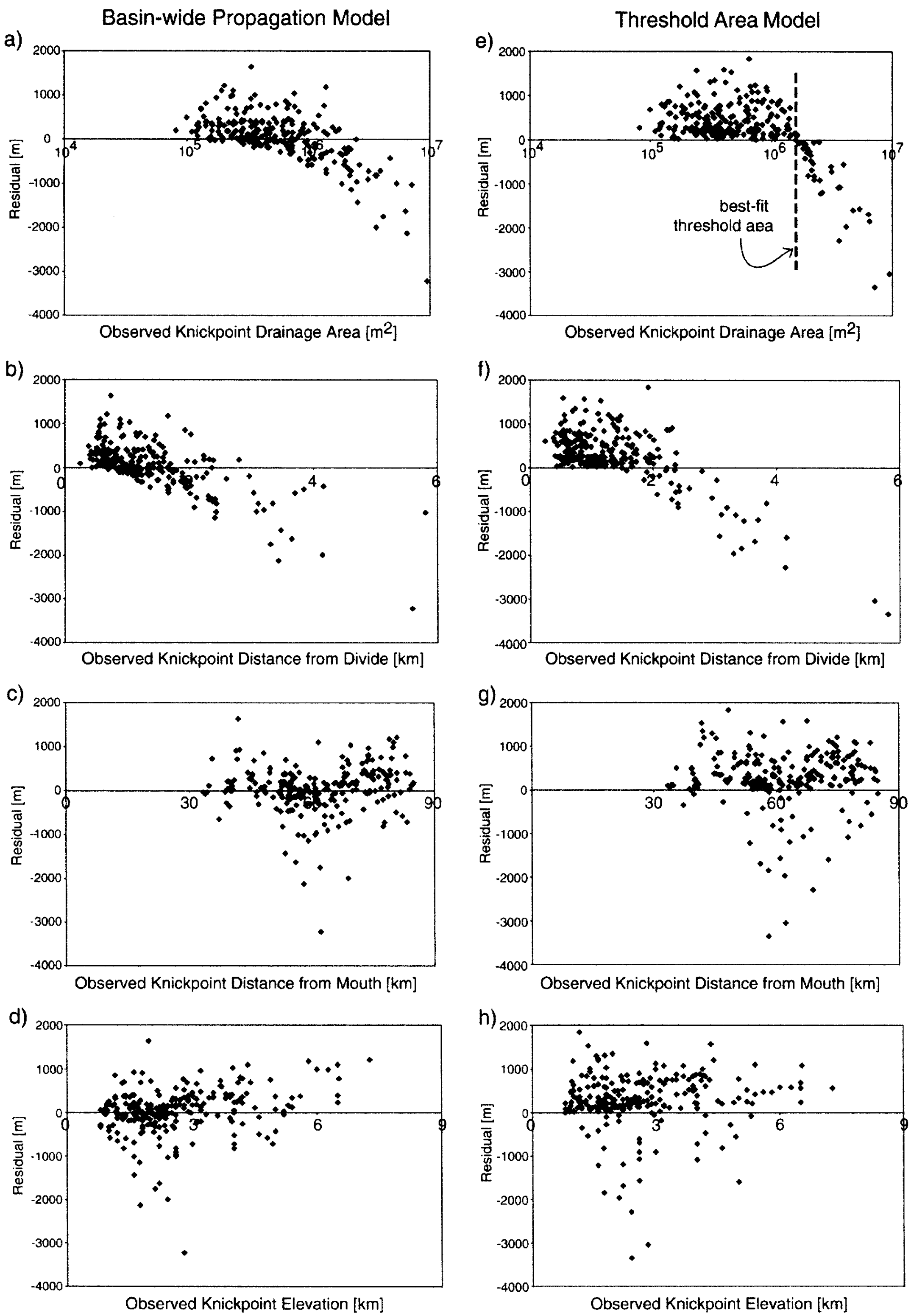
4.1.1. Model concept and mechanics

Since the late-nineteenth century, the upstream retreat of knickpoints has remained a popular mechanism for fluvial readjustment following base level fall. Though others have examined the conditions under which knickpoints are initiated and retreat, few have examined how this adjustment distributes through a fluvial network. Understanding how a pulse of incision is distributed within a fluvial network may prepare us to better approximate basin response times and sediment flux histories following base level fall. Our first model proposes that following the initiation of an incisional pulse at a downstream location, knickpoints propagate upstream throughout a fluvial network at a rate that is a power law function of the upstream drainage area. Modeled knickpoints thus migrate quickly along the large drainage area trunk streams and slow significantly upon entering tributaries of lesser drainage area.

The relationship we chose between retreat rate and drainage area bears a similar form to knickpoint celerity (Rosenbloom and Anderson, 1994; Weissel and Seidl, 1998; Whipple and Tucker, 1999). As discussed earlier, knickpoint celerity treats the rate of knickpoint propagation as analogous to speed of a kinematic wave (Whitham, 1974) and is derived from the well-known detachment-limited stream power incision model (Howard and Kerby, 1983). In the celerity derivation, the stream power incision model is combined with a statement of mass balance and rearranged to resemble the wave equation. In this form, the change in elevation with time (incision rate) is directly proportional to the change in elevation with distance (slope). The terms in the proportionality "constant" vary along the channel and define the wave speed. These terms include bed erodability, drainage area raised to an exponent and slope raised to another exponent.

Our approach recognizes the utility of this formulation, but also recognizes that the assumptions inherent in the formulation of the stream power model are violated at very steep reaches and at waterfalls in particular. The stream power model explicitly assumes that fluid flow is uniform

Fig. 7. Comparison between 8 selected knickpoint positions according to field observation, the basin-wide propagation model and the threshold area model. Plots compare knickpoint locations relative to basin geology (7a), channel profiles (7b) and drainage area structure (7c). Histograms also show the distribution of drainage areas of the 236 modeled knickpoints at three time steps during the 18,000 year long basin-wide propagation model run (7d–f). Note that some knickpoints in the basin-wide propagation model have already traveled from the outlet and into sub-tributaries within 1000 years. Also note that in the last 12,000 years of the model run, the knickpoints travel only a very short distance as decreasing drainage area dramatically decreases knickpoint velocities. In many of the channels, the final modeled positions (stars and circles) do a good job of matching the observed knickpoint positions (triangles).


Basin-wide Propagation Model
Threshold Area Model
a)
b)
c)
d)
e)
f)
g)
h)
Residual [m]
Observed Knickpoint Drainage Area [m2]
Observed Knickpoint Distance from Divide [km]
Observed Knickpoint Distance from Mouth [km]
Observed Knickpoint Elevation [km]
best-fit
threshold aea

(spatially invariant), steady (time invariant) and over a low slope. The slope must be low enough that the shear stresses of the fluid and sediment acting on the bed can drive incision, and in the case of knickpoints, drive retreat. For knickpoints at both large and small drainage areas, we found that waterfall faces were too steep to interact much with the falling water and sediment. Moreover, the assumption of uniform flow is violated as the fluid accelerates as it approaches and flows over the knickpoint lip. Because of these concerns, we do not associate our preliminary model with the stream power model and instead simply posit that the rate of knickpoint propagation is a power law function of drainage area:

$$\text{velocity} = \frac{dx}{dt} = CA^p \quad (1)$$

where dx/dt is the upstream knickpoint velocity, C is an unknown constant dimensional coefficient representing the efficiency of retreat [meters^{1-2p}/years], A is the upstream drainage area [m^2], and p is an unknown constant describing the power law dependence on drainage area [non-dimensional]. Though we acknowledge that the efficiency of retreat may vary throughout the basin as a function of local lithology or climate, we assume in this simple model that these factors are not dominating the basin wide behavior and thus treat C as a constant (see Section 5.1 for a discussion of possible influences of lithologic heterogeneity). Some factors that ought to affect the rate of retreat, such as water discharge, sediment supply and sediment character, are known to vary throughout the basin as a function of drainage area. In restricting our analysis to a constant value of the exponent, p, we are effectively assuming that these factors vary with drainage area in a consistent way within the basin. The data do not justify a more complex analysis in this regard. We also examined a variation on this formulation that includes a critical drainage area (A_{crit}) below which the knickpoint cannot propagate. This variation addresses some of the observations made earlier regarding the potential formation of knickpoints at tributary junctions and the upstream decrease in sediment flux. Requiring that A must be greater than A_{crit}, this variation takes the form:

$$\text{velocity} = \frac{dx}{dt} = C(A - A_{crit})^p \quad (2)$$

These two models for area-dependent knickpoint retreat were then tested in the Waipaoa drainage basin. We used the 25 m DEM to extract the 236 knickpoint-containing longitudinal profiles. Each channel profile starts at the basin outlet at the Pacific Ocean, extends upstream and past the knickpoint and stops at the drainage divide (Fig. 7a,b). For each pixel along each longitudinal profile, the upstream drainage area was calculated so that changes in drainage area were known along the entire path from the outlet to the divide (Fig. 7c). Using these data, modeled knickpoints initiated at the basin outlet traveled together up the mainstem, progressively slowing and branching off into smaller and smaller sub-basins, finally entering the unique portion of their path that contains the observed knickpoint (Fig. 7). This model thus mimics the way a single pulse of incision propagates up a trunk stream, and progressively distributes itself into lesser and lesser tributaries.

Each model run starts with particular values for C and p and then calculates the time required to translate across each pixel along the channel's path,

$$dt = dx(CA^p)^{-1} \quad (3)$$

where dt is the time necessary to translate across each pixel and dx is the diagonal or straight line distance across the pixel. These pixel travel times are then cumulatively summed from the outlet upstream. The predicted position of the knickpoint is where the cumulative travel-time equals or first exceeds 18,000 years. We calculated the misfit for each model by comparing the predicted and the observed knickpoints' along-stream distances from the divide. Because we have two unknowns and only one equation, an infinite number of pairs of C and p values provide a perfect fit for any individual channel. To find a single pair of C and p values that minimize the sum of the squares of these misfits for all 236 channels, we perform a brute-force search of C and p parameter space in each channel. The final C and p combination that results in the lowest sum of squares of the misfits in all the channels is the regional best fit (method similar to Stock and Montgomery, 1999).

We could have chosen to define the misfit by comparing the modeled and the observed knickpoint's drainage area or elevation but we found that when compared to the 'distance from divide' result, these best-fit model parameters were indistinguishable within uncertainty. In addition, the measurement of misfit using elevation is complicated by artificial 20 m steps observed within the DEM-derived stream profiles and

Fig. 8. Residuals for the best-fit basin-wide propagation model (8a–d) and best-fit threshold area model (8e–h). By definition, a positive residual indicates that the modeled knickpoint is downstream of the observed knickpoint. For drainage areas $>1\times10^6$ m^2, model residuals are strongly correlated with drainage areas of observed knickpoints. Because of the Hack relation, the same relationship is evident in comparisons between residuals and observed knickpoint distance from divide. Otherwise, residuals are uncorrelated and quite similar for the two models.

the discreet jumps in elevation at waterfalls (Figs. 3b and 7b). Therefore, we present only the more intuitive stream-wise, linear measure of the misfit. The magnitude and structure of residuals to the best-fit models are then used to evaluate the relative successes and failures of the model (Fig. 8a–d). By definition, a positive residual indicates that the modeled knickpoint is downstream of the observed knickpoint.

4.1.2. Basin-wide model result

The C and p combination that minimized the misfit between the observed and predicted knickpoint positions in the Waipaoa drainage basin was 7.9×10^{-9} $(m^{1.25}\,yr)^{-1}$ and 1.125 respectively. The difference between this power law dependence and a linear dependence on area is not statistically significant or within the error of the analysis. Using these values of C and p, the mean residual was $\sim40\pm$ 560 m (1σ) downstream of the observed knickpoint (Fig. 8a–d). In the case where a critical area was included in the model (Eq. (2)), there was no improvement in the fit; i.e. the lowest residuals were for a critical area of zero with the same best-fit C and p values as the standard model (Eq. (1)). The basin-wide model residuals were uncorrelated at smaller drainage areas, but for knickpoints with upstream drainage areas greater than 1×10^{6} m^2, the residuals were well correlated with area (Fig. 8a). Because drainage area and the distance from the divide can be related through Hack's relation, this same correlation exists between the distance from the divide and the residual (Fig. 8b). No correlations were observed between the residuals and the knickpoint elevations or the distances from the mouth (Fig. 8c–d). For all measured parameters, the residuals were well distributed around zero.

4.1.3. Basin-wide model discussion

The basin-wide propagation model provides valuable insight into how a drainage area dependent signal distributes itself within a fluvial network. The model makes its greatest contribution by demonstrating how each channel's unique, step-wise drainage area structure (Fig. 7c) ultimately sets the knickpoint distribution. Given the complexity of drainage networks, it is not surprising that no simple correlations (besides the long profile form (Fig. 4f) and Hack's law (Fig. 4i)) exist among drainage area, distance from divide, distance from outlet or elevation (Fig. 4g,h,j).

In order to evaluate how the modeled knickpoint positions varied through time, we tracked the distribution of knickpoint drainage areas at 1000 year time steps using the best-fit parameters of the basin-wide propagation model. We found that just after model initiation (17 ka), knickpoints demonstrate a narrow distribution strongly weighted toward high drainage areas (Fig. 7d). As the signal propagates upstream (12 ka), the distribution of drainage areas broadens (Fig. 7e). This broadening of the distribution is attributed knickpoints leaving the mainstem and branching into tributaries of varying drainage area. As the model reached the present time (0 ka), we found that the knickpoints in small, low drainage area tributaries near the outlet had retreated very little during the lifetime of the model. In the upper basin, the knickpoints that had once traveled so quickly in the mainstem have since slowed upon reaching drainage areas similar to the knickpoints near the outlet. As knickpoints throughout the basin all slow in low drainage area regions, the distribution of knickpoint drainage areas begins to narrow (Fig. 7f). If the basin-wide model was allowed to run beyond the present, we found that the distribution narrowed even further. We also used the model to track knickpoint velocities through time. One startling result is that in channels with drainage areas $>1\times10^{9}$ m^2 (the first 35 km of the mainstem), knickpoints were found to travel at rates greater than 100 m/yr. By the time they reach their present positions, the mean modeled velocity has dropped to ~4 cm/yr. Though the modeled retreat rate at present is reasonable given the observed frequency with which knickpoint faces collapse, the retreat rates at large drainage areas may be unrealistically fast.

Though the sequential terrace elevations observed between the 18,000 year old terrace and the active channel argue against the likelihood of a single, headward-propagating knickpoint lowering base level 50 to 100 m, we suggest two alternative scenarios to explain this circumstance. First, multiple mid-size knickpoints could have progressively swept through the system, generating sequential terrace elevations but culminating into a single step at tributary junctions where a large decrease in drainage area limits their retreat. Alternatively, if along-stream variations in substrate erodability can temporarily slow the pulse of incision, an episodic base level fall signal could culminate into a single step in harder lithologies. As this harder substrate reach is passed, the large, single-step knickpoint may become unstable and degrade progressively into multiple pulses of lowering, thus generating flights of terraces upstream of the resistant unit. Under these conditions, some reaches of the channel may experience knickpoint retreat, incising in a step-like manner and producing no intermediate terraces, while in others the base level lowering would be more gradual, producing sequential terraces (Fig. 2). In the second scenario, even if the knickpoint forms and degrades, the mean retreat rate of the pulse of incision may also still depend on upstream drainage area. Though we acknowledge that terrace preservation plays a large role in how we interpret the incision history, we presently favor the second model because the evidence for progressive incision is only observed within short, local reaches. In this model, the success with which the prediction matches the observed knickpoint distribution may be attributed to the tendency for most erosional responses, whether transport or detachment limited, to migrate upstream as some function of drainage area (Whipple and Tucker, 2002). Another positive attribute of this model is that the propagation of base level fall continues slowly into the upper-most reaches of the basin and could eventually reach drainage divides.

4.2. Knickpoint initiation at a threshold drainage area

4.2.1. Model concept and mechanics

Although the basin-wide knickpoint retreat model matches the Waipaoa knickpoint locations relatively well, there are reasons to believe, as briefly discussed above, that the present knickpoint distribution is not the end-result of a basin-wide process of knickpoint retreat. Instead the present distribution may be simply a consequence of a threshold in the channel's capacity to communicate large magnitude incision signals. As discussed earlier, previous workers only observe retreat without modification for knickpoints that have plunge pools, strong seepage erosion or horizontally stratified substrates with a resistant cap-rock. In the Waipaoa these conditions generally are not met. More importantly, as previously discussed, some trunk streams in the Waipaoa contain flights of terraces that suggest that the channel's elevation lowered progressively, not step-wise as would occur during the passage of a single knickpoint. These concerns lead us to develop an alternative model for the present distribution of knickpoints in the Waipaoa. This model does not address how the pulse of incision may have translated upstream, but rather suggests that stable knickpoints are only generated once the processes responsible for their retreat are no longer dominantly fluvial. First we will discuss the reasoning for why knickpoints might be initiated at a threshold drainage area and then we will discuss which processes might be responsible for the subsequent evolution of the knickpoint.

Some models for bedrock channel incision suggest that sediment flux plays a fundamental role in setting the rate and style of channel response to a pulse of incision (Sklar and Dietrich, 1998, 2001; Whipple and Tucker, 2002; Sklar and Dietrich, 2004). These models predict that as the base level fall signal reaches farther into the basin, the upstream drainage area can drop below a threshold where sediment discharge is not sufficient to continue rapid incision (Gasparini, 2003; Gasparini et al., 2006). Whether base level fall is progressive or step-wise, erosion continues downstream of the threshold drainage area as sediment shed from the over-steepened banks provides the tools necessary to erode the channel downstream of the knickpoint. Upstream of the threshold drainage area, low water and sediment flux suppress fluvial incision and knickpoint retreat becomes a function of other, non-fluvial processes.

Following Weissel and Seidl (1997), we hypothesize that once the knickpoint reaches the threshold drainage area, the primary mechanisms responsible for knickpoint retreat switch from dominantly fluvial processes to processes such as rock-mass failure and weathering. Our field observations suggest that the retreat rates of many of the 236 knickpoints are determined by the frequency of large block failures on the steep faces of the knickpoints. Because much of the substrate in the basin is clay-rich, these failures are most active on the knickpoint faces where the seasonally wetting and drying rocks experience significant shrinking and swelling. This accelerated physical weathering prepares the planes of weakness along which most failures occur. As the knickpoints slowly retreat upstream from the critical drainage area, they continue to define the boundary between the adjusting and relict portions of the basin. This theory of a threshold drainage area is also consistent with our earlier observation that many knickpoints are found at drainage areas between 1×10^5 m^2–1×10^6 m^2 (Fig. 4b) and just upstream of tributary junctions where the channel experiences a large rapid change in drainage area (Fig. 6).

We tested the behavior of the threshold drainage area model in the Waipaoa drainage basin and compared the model's predicted distribution with the observed knickpoint distribution (Fig. 8e–h) and with the predictions of the basin-wide retreat model (Fig. 4f–j). Like the basin-wide knickpoint propagation model discussed above, this model also used the 236 unique knickpoint-containing longitudinal profiles that run from outlet to divide. Because the upstream drainage area is known at every pixel along each channel profile, we could rapidly locate the point along each channel where the drainage area first decreases below the threshold drainage area (circles in Fig. 7a–c). Due to the efficiency of the model, we were able to test over 300 different threshold drainage areas between 1×10^4 m^2 and 1×10^8 m^2. For each model run we calculated the misfit in each channel by comparing the predicted and the observed knickpoints' distances from the divide. The best-fit threshold drainage area provided the lowest sum of squares of these misfits.

4.2.2. Threshold area model result

Though a threshold drainage area of 1.65×10^6 m^2 minimizes the sum of the squares of the 236 residuals (Fig. 8e), values between 1.1×10^6 m^2 and 2.0×10^6 m^2 all provided low residuals that were not statistically separable. For the best-fit threshold area, the mean modeled knickpoint position was $\sim 270 \pm 670$ m (1σ) downstream of present day observed knickpoint position (Fig. 8e–h). When minimizing the sum of squares for the residuals, the result was strongly influenced by a small population of large drainage area knickpoints. The threshold area model residuals were uncorrelated at smaller drainage areas, but knickpoints that occur at relatively large drainage areas ($> 1 \times 10^6$ m^2) demonstrate a strong correlation between the residuals and drainage area (Fig. 8e). Because drainage area and the distance from the divide can be related through Hack's law, this same correlation exists between the distance from the divide and the residual (Fig. 8f). There is no correlation, however, between the residual and the distance from the mouth or the knickpoint elevation (Fig. 8g,h). For all measured parameters, the residuals were not well distributed around zero, but were skewed downstream; the best-fit threshold drainage area (1.65×10^6 m^2) is larger than the mean drainage area at the observed knickpoints ($\sim 9 \times 10^5$ m^2).

4.2.3. Threshold area model discussion

Though the model locates knickpoints where the drainage area first falls below the threshold area, the model predicted knickpoints positions over a broad range of sub-threshold drainage areas ranging between $\sim 1 \times 10^5$ m^2 and 1.65×10^6 m^2 (Fig. 4g,i,j). We found that this broad range of predicted drainage areas was due to the model locating knickpoints just upstream of tributary junctions. At junctions, the drainage area decreases sharply to sub-threshold values which vary with tributary size (Fig. 7c). This modeled behavior is consistent with the fact that the observed knickpoints are also distributed over a similar range of drainage areas and that they are also located close to tributary junctions. This result suggests that if knickpoints do form at threshold drainage areas, then the dimensions of landscape parameters such as network density or ridge spacing (which determine the drainage area of the first order tributary basins) may play an important role in determining where threshold knickpoints are initiated.

It is interesting that for bedrock channels around the world, the transition between fluvial to arguably debris-flow dominated erosion processes occurs at drainage areas similar to our best-fit threshold area, $\sim 1 \times 10^6$ m^2 (Montgomery and Foufoula-Georgiou, 1993; Snyder et al., 2000; Stock and Dietrich, 2003; Whipple, 2004; Wobus et al., 2006). In the Waipaoa, the reaches upstream of knickpoints are at low gradients (< 0.020), receive a very small amount of coarse sediment from hillslopes and debris flows are not observed to be an important process. This suggests that though the drainage areas are similar, debris-flow dominated erosion is not likely determining knickpoint position or form. Also, knickpoints in the Waipaoa are not located at positions along the channel where the slope increases rapidly (Figs. 2 and 3b), as was suggested by Seidl and Dietrich (1992) and Schumm and Hadley (1957). However, the transition to debris-flow dominated conditions in other landscapes and the apparent stability of knickpoints in the Waipaoa both speak to a marked decrease in the efficiency of fluvial erosion processes at drainage areas less than $\sim 1 \times 10^6$ m^2.

One important and unique attribute of this model is that it does not require a rapid change in base level to create a knickpoint. Though steady channel incision can create knickpoints at threshold areas just as readily as a rapid pulse of incision, it is important to note that the theoretical models find that the rate of base level fall dictates the magnitude of the threshold drainage area (Gasparini, 2003; Gasparini et al., 2006). The quality of the model fit is not likely influenced by knickpoint retreat beyond the threshold area since, for the subset of tributary junctions studied, 78% are less than 1 km upstream of the large tributary junction where they may have initiated. It would be valuable for future models to allow the threshold area to vary as some function of substrate properties, water discharge or sediment flux and character. In addition, models for non-fluvial retreat processes are necessary to propagate the knickpoint upstream once the threshold area is reached.

5. Discussion

5.1. Lithologic influence on knickpoint formation and retreat

Although it seems intuitive to expect that the variability of the substrate strength in the Waipaoa should play a role in setting knickpoint location and retreat rate, we observed no relation between the residuals to either best-fit model and any particular lithology. However, this is largely a consequence of the fact that we filtered out knickpoints that are directly influenced by local lithologic contacts or conditions. The subset of knickpoints we fit our models to are found in relatively comparable substrates (Fig. 7a) and thus exhibit a similar form. If the models were run for all the channels in the Waipaoa, not just the 236 with well-defined knickpoints associated with the 18,000 year old incision event, the quality of the fit would decrease dramatically and a strong lithologic influence would become apparent. In soft or strongly deformed substrates we observed very few knickpoints, but in harder, more coherent units, most tributaries with sub-threshold drainage areas contained knickpoints. We surmise that knickpoints in the softer units either degraded in a diffusive manner as seen in Gardner's (1983) experiments, or never formed in the first place.

5.2. Comparison between models: knickpoint propagation vs. threshold area

For end-member models whose underlying mechanics are fundamentally different, both predict a surprisingly similar knickpoint distribution and match the observed knickpoint distribution quite well (Figs. 4 and 7). Both models do a good job of fitting observed knickpoints at small drainage areas near tributary junctions, but both models have a poor fit to the large drainage area knickpoints (Fig. 8a,e). These similarities in model performance are likely a consequence of the mature stage of the fluvial component of the transient response in the Waipaoa. Because of this maturity, most knickpoints are at similar, small drainage areas. In the basin-wide propagation model, it may be the re-narrowing of the knickpoint drainage area distribution (Fig. 7f) that gives the model such striking similarity to the observed and threshold model distribution of knickpoints. Unfortunately, while the fluvial adjustment is so nearly complete, it is not possible to unequivocally differentiate between the basin-wide and the threshold models. To successfully compare the performance of the models, observations of the transient response need to be made at some earlier time when the relative differences in the performance of the two models would be more apparent.

A striking example of this inherent difficulty in tests of knickpoint formation and propagation models is found in our data for the distance between major tributary junctions and knickpoints (x_{dif}) as a function of drainage area at the tributary junction (A_{jct}). As shown in Fig. 6, there is a strong correlation between these variables. This correlation is of course expected for the scenario in which knickpoint retreat is driven by fluvial processes, as such processes are drainage-area dependent. This is illustrated on Fig. 6 by plotting a solid line of knickpoint positions predicted by the basin-wide propagation model (using our best-fit model parameters [$C=7.9\times10^{-9}$ $(\mathrm{m}^{1.25}\ \mathrm{yr})^{-1}$, $p=1.125$]) in different sized tributary drainage basins. This calculation assumes (1) knickpoints initiate simultaneously at tributary junctions, and (2) each drainage basin is characterized by the same Hack's relation parameters [typical values for the Waipaoa; $k_a=9.71$, $h=1.55$ (Fig. 4d)]. It is tempting to take this as strong confirmation of knickpoint retreat as a function of drainage area (at least upstream of tributary junctions where a knickpoint may actually initiate). However, a simple analysis reveals that a correlation between x_{dif} and A_{jct} is in fact also predicted using the simple threshold area model for knickpoint formation in basins following a Hack's relation-type geometry.

Using Hack's relation we can write distance from divide to the junction, x_{jct}, as a function of A_{jct}:

$$x_{jct} = k_a^{-1/h} A_{jct}^{1/h} \tag{4}$$

The distance from divide to the junction, x_{jct}, is by definition (Fig. 6) also equal to the sum of the upstream distance from the junction to the knickpoint, x_{dif}, and the distance from the divide to the knickpoint, x_{kp}.

$$x_{jct} = x_{dif} + x_{kp} \tag{5}$$

The value of x_{kp} is determined according to Hack's relation and the threshold drainage area A_{kp}. Solving Eq. (5) for x_{dif}, we can plug Eq. (4) in for x_{jct} and rewrite an analogous expression for x_{kp}. Thus we find a simple, direct relation is expected between tributary drainage area and the distance between the tributary junction and the knickpoint, even in the case where no fluvially-driven knickpoint retreat has occurred:

$$x_{dif} = [k_a^{-1/h} A_{jct}^{1/h}] - [k_a^{-1/h} A_{kp}^{1/h}] \tag{6}$$

This predicted relation using a threshold drainage area is plotted in Fig. 6 as a dashed line. This result suggests that for a set of basins with drainage areas and drainage network structures similar to those in the Waipaoa, we would find similar relations between A_{jct} and x_{dif} whether knickpoints were formed at some threshold drainage area, A_{kp}, or whether their locations in the basin were a function of drainage-area-dependent knickpoint retreat (Fig. 6).

5.3. Basin-wide knickpoint retreat: toward a general theoretical model

Our field observations from the Waipaoa drainage basin and the preliminary models presented above offer some insight into which factors may influence the distribution of knickpoints in a fluvial network. Actively retreating knickpoints may play an important, sometimes dominant role in fluvial adjustment following perturbation, yet there is little field evidence as to whether they retain their form while translating upstream throughout entire drainage basins. To account for the complications outlined above we propose a much more dynamic response to perturbation. If rapid base level fall generates a step in the channel that propagates upstream, this form may be unstable during retreat and degrade and re-form multiple times as it encounters regions of higher and lower erodability (Frankel et al., 2001). This behavior would allow knickpoints to temporarily exist in reaches where sediment and water discharge are large. In addition, a knickpoint that forms and degrades multiple times may generate a complicated terrace record. In locations where base level fall is distributed over time, incision may be recorded by sequential terrace levels (Fig. 2). In other reaches where the knickpoint retreats as a single step, no record of progressive incision would be recorded (Fig. 2). One might anticipate that at any given time, even migrating knickpoints would most likely appear to be "hung up" on hard bedrock ribs or at tributary junctions where there is a step decrease in sediment and water discharge. As a consequence, the very idea of knickpoint migration has been challenged and debated episodically for over 100 years (Gilbert, 1896; Penck, 1925; von Engeln, 1942; Leopold et al., 1964; Gardner, 1983; Higgins and Gardner, 1984; Wohl et al., 1994). Because this conceptual model allows knickpoints to form and degrade multiple times during their translation through the downstream reaches, it is likely that multiple small pulses of incision or even steady base level fall would be held up at a resistant substrate and collect into a single step. Moreover, if knickpoint formation occurs at locations where the erosive capacity cannot keep pace with downstream rates (such as at the junction of a small tributary and a large trunk stream), it does not require rapid base level fall to generate knickpoints. These two concerns suggest that our interpretation of knickpoints as responding to pulses of incision must be made with caution as they could result from either gradual base level fall or a pulse of incision.

As the pulse of incision extends into the upper reaches of the network or encounters a lesser tributary, the erosive capacity of that channel may not be capable of keeping pace with downstream incision rates. The reduced erosive capacity may be due to an insufficient upstream sediment supply or water discharge to move the available sediment load. Lower sediment supply would limit the erosive potential of the accelerating flow upstream of the knickpoint lip, thus explaining the lack of incised drawdown reaches in the Waipaoa. In this case, the knickpoint would not lip degrade as there would be too few tools to develop a drawdown zone. With a diminished supply of erosive tools, the rate-limiting process for knickpoint retreat is no longer fluvial, but instead relies on rock-mass stability, weathering and hillslope erosion processes (Weissel and Seidl, 1997). As the knickpoint face collapses and retreats upstream, the downstream incised gorge is supplied with sediment eroded from the knickpoint face and over-steepened banks. This material provides tools to continue incision in the over-steepened reach downstream of the knickpoint.

It must be emphasized that though much of the Waipaoa river and its tributaries has responded to the base level lowering, much of the aerial extent of the basin remains disconnected from the base level fall signal. If channel form depends on sediment supply, channels will remain in disequilibrium until the relict surface has been removed and hillslope–channel connectivity is reestablished.

5.4. Research needs

A model describing the multi-process, transient response of a fluvial network to a pulse of incision would greatly contribute to our understanding of landscape evolution. One of the greatest challenges in developing this model would be providing the flexibility for detachment and transport limited erosional processes to trade-off as local substrate, sediment and flow conditions change along the channel. As well, in the upper reaches of the channels, the transition between fluvial bedrock incision and rock-mass wasting at waterfalls must be better explored. If we desire to use a knickpoint retreat model to approximate the sediment flux history following an incisional pulse, we must also better understand the response times of hillslopes following base level fall. Though the Waipaoa provides a great resource to study knickpoint behavior in the upper reaches of a basin, it would also be valuable to examine a suite of younger transient systems where the incisional pulse has not extended as far into the network. If the issues concerning the scaling of basin-wide processes can be better constrained, it would be most valuable to use laboratory experiments to examine how bedrock incision in a trunk stream is propagated into tributaries whose discharge is some fraction of the trunks.

6. Conclusions

The present distribution of 236 knickpoints within the Waipaoa drainage basin can be well approximated using models that either allow knickpoints to propagate upstream at a rate that is an approximately linear power law function of drainage area or initiate knickpoints at a threshold drainage area. These models may succeed at matching observed knickpoint locations because they link the basin's response to incision to the network's drainage area structure. Alternatively, the consistency of our low residuals between the two models could be a consequence of the maturity of the pulse of incision and the knickpoints' consistent upper-network, lower-drainage area position.

Combining observation from field analysis and model behavior, we propose that the response to a pulse of incision may consist of three stages. First, large drainage area channels incise as knickpoints form and degrade multiple times as the pulse of incision propagates upstream. Second, as this incisional wave sweeps up the network, lower order, small drainage area channels are unable to keep pace with the lowering and thus develop steep knickpoints. Third, these knickpoints slowly migrate upstream and decay as a consequence of primarily weathering and mass-wasting driven erosion processes.

Further developments to modeling the transient response of landscapes to a pulse of incision will need to acknowledge (1) the variability in erodability along each channel's path and the dynamic response of knickpoints to this variability, (2) the erosional processes active at low drainage areas and (3) the lag-time between channel incision and hillslope response. With these improvements, future landscape evolution models may be capable of determining more accurate fluvial response times and sediment fluxes following a pulse of incision.

Acknowledgements

The authors would like to thank those who provided great assistance in the field: especially Mike Marden (Landcare Research), The Meban Family (Te Hau Farmstay), Simone Hood Hills, Cana Crosby and Ben Gross. Early discussions with and reviews by Jeremy Boyce, Doug Jerolmack, Joel Johnson, Blair Schoene and Cameron Wobus helped sharpen the focus of this work. Later versions of the manuscript we considerably improved following reviews by Nicole Gasparini, Simon Mudd and an anonymous reviewer. The authors also extend thanks to Landcare Research, the Institute of Geological and Nuclear Sciences, Air Maps New Zealand Ltd. and Eagle Technology Ltd. for kindly sharing their mapping resources. This work was supported through an NSF grant, EAR-0208312 (to KXW) and an NSF Graduate Research Fellowship (to BTC).

References

Adams, J., 1985. Large-scale tectonic geomorphology of the Southern Alps, New Zealand. In: Morisawa, M., Hack, J.T. (Eds.), Tectonic Geomorphology. Allen and Unwin, Winchester, Mass, pp. 105–128.

Beaumont, C., Fullsack, P., Hamilton, J., 1992. Erosional control of active compressional orogens. In: McClay, K.R. (Ed.), Thrust Tectonics. Chapman and Hall, New York, pp. 1–18.

Begin, Z.B., 1988. Application of a diffusion–erosion model to alluvial channels which degrade due to base-level lowering. Earth Surface Processes and Landforms 13, 487–500.

Begin, Z.B., Meyer, D.F., Schumm, S.A., 1981. Development of longitudinal profiles of alluvial channels in response to base-level lowering. Earth Surface Processes and Landforms 6, 49–68.

Berryman, K., Marden, M., Eden, D., Mazengarb, C., Ota, Y., Moriya, I., 2000. Tectonic and paleoclimatic significance of Quaternary river terraces of the Waipaoa River, east coast, North Island, New Zealand. New Zealand Journal of Geology and Geophysics 43 (2), 229–245.

Berryman, K., Marden, M., Palmer, A.S., Wilson, K., Mazengarb, C., Litchfield, N., in preparation. The post-glacial downcutting history in the Waihuka tributary of the Waipaoa River, Gisborne District, New Zealand, and implications for tectonics and landscape evolution.

Bishop, P., Hoey, T.B., Jansen, J.D., Artza, I.L., 2005. Knickpoint recession rate and catchment area: the case of uplifted rivers in Eastern Scotland. Earth Surface Processes and Landforms 30, 767–778.

Black, R.D., 1980. Upper Cretaceous and Tertiary geology of Mangatu State Forest, Raukumara Peninsula, New Zealand. New Zealand Journal of Geology and Geophysics 23 (3), 293–312.

Brown, L.J., 1995. Holocene shoreline depositional processes at Poverty Bay, a tectonically active area, northeastern North Island, New Zealand. Quaternary International 26, 21–33.

Brush Jr., L.M., Wolman, M.G., 1960. Knickpoint behavior in noncohesive material — a laboratory study. Geological Society of America Bulletin 71 (1), 59–73.

Chappell, J., 2001. Sea level change through the last glacial cycle. Science 292, 679–686.

Collot, J., Lewis, K., Lamarche, G., Lallemand, S., 2001. The giant Ruatoria debris avalanche on the northern Hikurangi margin, New Zealand; results of oblique seamount subduction. Journal of Geophysical Research, Solid Earth 106 (B9), 19,271–19,297.

Crosby, B.T., 2006. The Origin and Evolution of Waterfalls. Ph.D. Thesis, Massachusetts Institute of Technology, Cambridge, MA, USA.

Davis, W.M., 1932. Piedmont bench lands and primaerruempfe. Geological Society of America Bulletin 43 (2), 399–440.

Davy, B.W., 1992. The influence of subducting plate buoyancy on subduction of the Hikurangi–Chatham Plateau beneath the North Island, New Zealand. AAPG Memoir 53, 75–91.

DeMets, C., Gordon, R.G., Argus, D.F., Stein, S., 1994. Effect of recent revisions to the geomagnetic reversal time scale on estimates of current plate motions. Geophysical Research Letters 21 (20), 2191–2194.

Dunne, T., 1990. Hydrology, mechanics, and geomorphic implications of erosion by subsurface flow. In: Higgins, C.G., Coates, D.R. (Eds.), Groundwater Geomorphology: the Role of Subsurface Water in Earth-Surface Processes and Landforms. Geological Society of America, Special Paper, vol. 252, pp. 1–28.

Eberhart-Phillips, D., Chadwick, M., 2002. Three-dimensional attenuation model of the shallow Hikurangi subduction zone in the Raukumara Peninsula, New Zealand. Journal of Geophysical Research, Solid Earth 107 (B2, 2033). doi:10.1029/2000JB000046.

Eden, D.N., Palmer, A.S., Cronin, S.J., Marden, M., Berryman, K.R., 2001. Dating the culmination of river aggradation at the end of the last glaciation using distal tephra compositions, eastern North Island, New Zealand. Geomorphology 38, 133–151.

Frankel, K.L., Vaughn, J.D., Pazzaglia, F.J., 2001. Knickpoint retreat and long profile evolution through a vertically bedded substrate; a flume study. Geological Society of America — Abstracts with Programs 33 (6), 314.

Gardner, T.W., 1983. Experimental-study of knickpoint and longitudinal profile evolution in cohesive, homogeneous material. Geological Society of America Bulletin 94 (5), 664–672.

Gasparini, N.M., 2003. Equilibrium and transient morphologies of river networks: discriminating among fluvial erosion models. Ph.D. Thesis, Massachusetts Institute of Technology, Cambridge, 232 pp.

Gasparini, N.M., Bras, R.L., Whipple, K.X., 2006. River longitudinal profiles and bedrock incision models: stream power and the influence of sediment supply. In: Willett, S.D., Hovius, N., Brandon, M.T., Fisher, D.M. (Eds.), Tectonics, Climate and Landscape Evolution: Geological Society of America Special Paper 398, Penrose Conference Series, pp. 127–141.

Gilbert, G.K., 1896. Niagara Falls and their history. National Geographic Society. The Physiography of the United States. The American Book Co., New York, pp. 203–236.

Hack, J.T., 1957. Studies of longitudinal stream profiles in Virginia and Maryland.U.S. Geological Survey Professional Paper 294-B, 45–97.

Hasbargen, L.E., Paola, C., 2000. Landscape instability in an experimental drainage basin. Geology 28 (12), 1067–1070.

Haviv, I., Enzel, Y., Whipple, K.X., Zilberman, E., Stone, J., Matmon, A., Fifield, L.K., 2003. Knickpoint retreat and drawdown reach evolution along the arid escarpments of the Dead Sea basin. Geological Society of America — Abstracts with Programs 35 (6), 333.

Hayakawa, Y., Matsukura, Y., 2003. Recession rates of waterfalls in Boso Peninsula, Japan, and a predictive equation. Earth Surface Processes and Landforms 28 (6), 675–684.

Higgins, C.G., Gardner, T.W., 1984. Experimental study of knickpoint and longitudinal profile evolution in cohesive, homogeneous material; discussion and reply. Geological Society of America Bulletin 95 (1), 122–123.

Holland, W.N., 1974. Origin and development of hanging valleys in the Blue Mountains, New South Wales. Ph.D. Thesis, Sydney University, Sydney, Australia, 416 pp.

Holland, W.N., Pickup, G., 1976. Flume study of knickpoint development in stratified sediment. Geological Society of America Bulletin 87 (1), 76–82.

Howard, A.D., 1994. A detachment-limited model of drainage basin evolution. Water Resources Research 30 (7), 2261–2285.

Howard, A.D., Kerby, G., 1983. Channel changes in badlands. Geological Society of America Bulletin 94 (6), 739–752.

Howard, A.D., Dietrich, W.E., Seidl, M.A., 1994. Modeling fluvial erosion on regional to continental scales. Journal of Geophysical Research, Solid Earth 99 (B7), 13971–13986.

Koons, P.O., 1995. Modeling the topographic evolution of collisional belts. Annual Reviews of Earth and Planetary Sciences 23, 375–408.

Laity, J.E., Malin, M.C., 1985. Sapping processes and the development of theater-headed valley networks on the Colorado Plateau. Geological Society of America Bulletin 96 (2), 203–217.

Leopold, L.B., Wolman, M.G., Miller, J.P., 1964. Fluvial Processes in Geomorphology. Freeman, San Francisco.

Lewis, K.B., Pettinga, J.R., 1993. The emerging, imbricate frontal wedge of the Hikurangi margin. In: Ballance, P.F. (Ed.), Sedimentary Basins of the World: South Pacific Sedimentary Basins. Elsevier Science Publishers B.V, Amsterdam, pp. 225–250.

Lowe, D.J., Newnham, R.M., Ward, C.M., 1999. Stratigraphy and chronology of a 15 ka sequence of multi-sourced silicic tephras in a montane peat bog in eastern North Island, New Zealand. New Zealand Journal of Geology and Geophysics 42, 565–579.

Mackin, J.H., 1948. Concept of the graded river. Geological Society of America Bulletin 101, 1373–1388.

Mazengarb, C., Speden, I.G., 2000. Geology of the Raukumara Area: Institute of Geological and Nuclear Sciences 1:250,000 geological map 6. Institute of Geological and Nuclear Sciences Limited, Lower Hutt, New Zealand.

Mazengarb, C., Francis, D.A., Moore, P.R., 1991. Sheet Y16-Tauwhareparae, Geological map of New Zealand, scale 1:50,000, Department of Scientific and Industrial Research, Lower Hutt, NZ.

Miller, J.R., 1991. The influence of bedrock geology on knickpoint development and channel-bed degradation along downcutting streams in south-central Indiana. Journal of Geology 99 (4), 591–605.

Moglen, G.E., Bras, R.L., 1995a. The effect of spatial heterogeneities on geomorphic expression in a model of basin evolution. Water Resources Research 31 (10), 2613–2623.

Moglen, G.E., Bras, R.L., 1995b. The importance of spatially heterogeneous erosivity and the cumulative area distribution within a basin evolution model. Geomorphology 12, 173–185.

Montgomery, D.R., Foufoula-Georgiou, E., 1993. Channel network representation using digital elevation models. Water Resources Research 29, 1178–1191.

Newnham, R.M., Lowe, D.J., Williams, P.W., 1999. Quaternary environmental change in New Zealand: a review. Progress in Physical Geography 23, 567–610.

Niemann, J.D., Gasparini, N.M., Tucker, G.E., Bras, R.L., 2001. A quantitative evaluation of Playfair's law and its use in testing long-term stream erosion models. Earth Surface Processes and Landforms 26 (12), 1317–1332.

Ota, Y., Hull, A.G., Iso, N., Ikeda, Y., Moriya, I., Yoshikawa, T., 1992. Holocene marine terraces on the northeast coast of North Island, New Zealand and their tectonic significance. New Zealand Journal of Geology and Geophysics 35, 273–288.

Penck, W., 1924. Die Morphologische Analyse: ein Kapital der Physikalischen geologie. J. Engelhorns nachf, Stuttgart.

Penck, W., 1925. Die piedmontflachen des sudlichen schwarswaldes. Gesellschaft fur Erdkunde, Berlin 1, 81–108.

Pillans, B., McGlone, M., Palmer, A., Mildenhall, D., Alloway, B., Berger, G., 1993. The Last Glacial Maximum in central and southern North Island, New Zealand; a paleoenvironmental reconstruction using the Kawakawa Tephra formation as a chronostratigraphic marker. Palaeogeography, Palaeoclimatology, Palaeoecology 101 (3–4), 283–304.

Rosenbloom, N.A., Anderson, R.S., 1994. Hillslope and channel evolution in a marine terraced landscape, Santa Cruz, California. Journal of Geophysical Research, Solid Earth 99 (B7), 14,013–14,029.

Schumm, S.A., Hadley, R.F., 1957. Arroyos and the semiarid cycle of erosion. American Journal of Science 255, 161–174.

Seidl, M.A., Dietrich, W.E., 1992. The problem of channel erosion into bedrock. Catena Supplement: Functional geomorphology; Landform Analysis and Models, vol. 23, pp. 101–124.

Seidl, M.A., Dietrich, W.E., Kirchner, J.W., 1994. Longitudinal profile development into bedrock — an analysis of Hawaiian channels. Journal of Geology 102 (4), 457–474.

Shackelton, N.J., 1987. Oxygen isotopes, ice volumes and sealevel. Quaternary Science Reviews 6, 183–190.

Sklar, L.S., Dietrich, W.E., 1998. River longitudinal profiles and bedrock incision models: stream power and the influence of sediment supply. In: Tinkler, K.J., Wohl, E.E. (Eds.), Rivers Over Rock: Fluvial Processes in Bedrock Channels. American Geophysical Union, Washington, D.C, pp. 237–260.

Sklar, L.S., Dietrich, W.E., 2001. Sediment and rock strength controls on river incision into bedrock. Geology (Boulder) 29 (12), 1087–1090.

Sklar, L.S., Dietrich, W.E., 2004. A mechanistic model for river incision into bedrock by saltating bed load. Water Resources Research 40. doi:10.1029/2003WR002496.

Snyder, N.P., Whipple, K.X., Tucker, G.E., Merritts, D.J., 2000. Landscape response to tectonic forcing; digital elevation model analysis of stream profiles in the Mendocino triple junction region, Northern California. Geological Society of America Bulletin 112 (8), 1250–1263.

Stein, O.R., LaTray, D.A., 2002. Experiments and modeling of head cut migration in stratified soils. Water Resources Research 38 (12).

Stock, J.D., Dietrich, W.E., 2003. Valley incision by debris flows: evidence of a topographic signature. Water Resources Research 39. doi:10.1029/2001WR001057.

Stock, J.D., Montgomery, D.R., 1999. Geologic constraints on bedrock river incision using the stream power law. Journal of Geophysical Research, Solid Earth 104 (B3), 4983–4993.

Tucker, G.E., Slingerland, R.L., 1994. Erosional dynamics, flexural isostasy, and long-lived escarpments — a numerical modeling study. Journal of Geophysical Research, Solid Earth 99 (B6), 12229–12243.

Tucker, G.E., Slingerland, R.L., 1997. Drainage basin response to climate change. Water Resources Research 33 (8), 2031–2047.

von Engeln, O.D., 1942. Geomorphology. MacMillan and Company, New York.

Weissel, J.K., Seidl, M.A., 1997. Influence of rock strength properties on escarpment retreat across passive continental margins. Geology 25 (7), 631–634.

Weissel, J.K., Seidl, M.A., 1998. Inland propagation of erosional escarpments and river profile evolution across the southeast Australian passive continental margin. In: Tinkler, K.J., Wohl, E.E. (Eds.), Rivers Over Rock: Fluvial Processes in Bedrock Channels. American Geophysical Union, Washington, D.C., pp. 189–206.

Whipple, K.X., 2001. Fluvial landscape response time: how plausible is steady-state denudation? American Journal of Science 301, 313–325.

Whipple, K.X., 2004. Bedrock rivers and the geomorphology of active orogens. Annual Review of Earth and Planetary Sciences 32, 151–185.

Whipple, K.X., Tucker, G.E., 1999. Dynamics of the stream-power river incision model; implications for height limits of mountain ranges, landscape response timescales, and research needs. Journal of Geophysical Research, Solid Earth 104 (B8), 17,661–17,674.

Whipple, K.X., Tucker, G.E., 2002. Implications of sediment-flux-dependent river incision models for landscape evolution. Journal of Geophysical Research, Solid Earth 107 (B2), 3–1–3–20.

Whitham, G.B., 1974. Linear and Non-Linear Waves. Wiley, New York.

Willett, S.D., 1999. Orogeny and orography: the effects of erosion on the structure of mountain belts. Journal of Geophysical Research, Solid Earth 104 (B12), 28957–28981.

Willgoose, G., Bras, R.L., Rodriguez-Iturbe, I., 1991. A coupled channel network growth and hillslope evolution model. 1. Theory. Water Resources Research 27 (7), 1671–1684.

Wobus, C.W., Whipple, K.X., Kirby, E., Snyder, N.P., Johnson, J., Spyropolou, K., Crosby, B.T., Sheehan, D., 2006. Tectonics from topography: Procedures, promise and pitfalls. In: Willett, S.D., Hovius, N., Brandon, M.T., Fisher, D.M. (Eds.), Tectonics, Climate and Landscape Evolution: Geological Society of America Special Paper 398, Penrose Conference Series, pp. 55–74.

Wohl, E.E., 1993. Bedrock channel incision along Piccaninny Creek, Australia. Journal of Geology 101 (6), 749–761.

Wohl, E.E., Greenbaum, N., Schick, A.P., Baker, V.R., 1994. Controls on bedrock channel incision along Nahal-Paran, Israel. Earth Surface Processes and Landforms 19 (1), 1–13.

Wolman, M.G., 1987. Sediment movement and knickpoint behavior in a small Piedmont drainage basin. Geografiska Annaler. Series A. Physical Geography 69 (1), 5–14.

Yoshikawa, T., 1988. Pattern and rate of tectonic movement and late Quaternary geomorphic development in the Raukumara Peninsula, northeastern North Island, New Zealand. Bulletin of the Department of Geography, University of Tokyo 20, 1–28.

Zaprowski, B.J., Evenson, E.B., Pazzaglia, F.J., Epstein, J.B., 2001. Knickzone propagation in the Black Hills and northern High Plains: a different perspective on the late Cenozoic exhumation of the Laramide Rocky Mountains. Geology 29 (6), 547–550.

Formation of Fluvial Hanging Valleys: Theory and Simulation

Benjamin T. Crosby[1,2]

Kelin X Whipple[1,3]

Nicole M. Gasparini[4]

Cameron W. Wobus[1,5]

1 Department of Earth, Atmospheric and Planetary Sciences, Massachusetts Institute of Technology, Cambridge, Massachusetts 02139
2 Now at: Department of Geosciences, Idaho State University, Pocatello, ID 83209
3 Now at: School of Earth and Space Exploration, Arizona State University, Tempe, AZ
4 Now at: Department of Geology and Geophysics, Yale University, New Haven CT 06511
5 Now at: Cooperative Institute for Research in Environmental Sciences (CIRES), University of Colorado, Boulder, CO 80309

Primary Contact: Benjamin Crosby at crosby@alum.mit.edu
In Review at JGR-Earth Surface

Abstract

Although only recently recognized, hanging tributary valleys in non-glacial landscapes are common in tectonically active regions. Standard river incision models do not allow for the formation of fluvial hanging valleys; these disequilibrium landforms present an opportunity to advance our understanding of river incision processes. In this work, we demonstrate that thresholds apparent in sediment-flux dependent bedrock incision rules provide mechanisms for the formation of hanging valleys in response to transient pulses of river incision. We simplify recently published river incision models in order to derive analytical solutions for the conditions required for hanging valley formation and use these results to guide numerical landscape evolution simulations. Analytical and numerical results demonstrate that during the response to base level fall, these incision rules may create either temporary or permanent hanging valleys. These hanging valleys form as a consequence of (1) rapid mainstem incision over-steepening the tributary junctions beyond a threshold slope or (2) tributary sediment flux response lagging behind the pulse of incision, temporarily limiting the tributary's capacity to keep pace with the mainstem. The distribution of permanent and temporary hanging valleys results from four competing factors: the magnitude of base level fall, the upstream attenuation of the incision signal, the lag time of the sediment flux response and the non-systematic variation in tributary drainage areas within the stream network. The development of hanging valleys in landscapes governed by sediment-flux dependent incision rules limits the transmission of base level fall signals through the channel network, ultimately increasing basin response time.

Motivation

Following a change in tectonic or climatic forcing, hillslopes and channels adjust their form as the landscape moves toward equilibrium with the new boundary conditions. The form and response time of this transient adjustment exerts a fundamental influence on the growth and development of mountain ranges, the timing and delivery of sediment to depositional basins and other fundamental processes in tectonically active landscapes. In non-glacial, tectonically active landscapes, it has recently been discovered that many tributary basins are observed "hanging" above their junction with the incised trunk stream [*Crosby and Whipple*, in press; *Snyder, et al.*, 1999; *Wobus, et al.*, in press]. The over-steepened channel reach, often referred to as a knickpoint, separates the relict hanging tributary basin from the incised, adjusting portion of the landscape (figure 1). We identify the Waipaoa River on the North Island of New Zealand as an excellent location to study knickpoint distribution within fluvial basins because of the known time of disturbance [*Berryman, et al.*, 2000; *Eden, et al.*, 2001] and the abundance of transient landforms such as terraces, incised inner gorges and knickpoints [*Crosby and Whipple*, in press]. Even with a field site as well constrained as the Waipaoa, we found it is difficult to discern whether hanging valley formation requires discrete knickpoints migrating up the mainstem to their present position or whether hanging valleys can develop in-place at tributary junctions when prolonged mainstem incision outpaces tributary adjustment [*Crosby and Whipple*, in press]. The presence and persistence of knickpoints at tributary junctions limits the upstream communication of subsequent signals of base level change into the upper portions of the channel network, thus extending basin response time following disturbance.

The most broadly utilized formulation for fluvial bedrock incision, the detachment-limited stream power model [*Howard*, 1994; *Howard and Kerby*, 1983; *Whipple and Tucker*, 1999] (henceforth simply termed the stream power model), does not predict the formation of hanging tributaries[e.g. *Niemann, et al.*, 2001]. This discrepancy between field observation and model behavior suggests a clear inadequacy in standard river incision models. Recently developed sediment-flux dependent bedrock incision relations allow sediment to behave either as a tool for incising the bed or as armor, inhibiting erosion [*Gasparini, et al.*, 2006; *Parker*, 2002; *Sklar and Dietrich*, 1998; *Sklar and Dietrich*, 2004; *Whipple and Tucker*, 2002]. We find that these relations provide mechanisms for explaining the formation and persistence of hanging tributaries at threshold drainage areas [*Crosby and Whipple*, in press; *Gasparini, et al.*, submitted; *Wobus, et al.*, in press]. In particular, the prevalence of hanging tributaries in tool-starved environments may provide the strongest existing field evidence for the applicability of sediment-flux dependent incision rules and their clear superiority to standard detachment-limited and transport-limited stream power models.

Approach and Scope

In this work, we utilize analytical and numerical models to explore whether thresholds in existing relations for stream incision by sediment abrasion provide a plausible mechanism for the formation of hanging valleys. The theoretically demonstrated consequences of these thresholds are then considered relative to existing field studies of hanging valley distributions [*Bishop, et al.*, 2005; *Crosby and Whipple*, in press; *Weissel and Seidl*, 1998; *Wobus, et al.*, in press]. Though plausible alternative explanations for the formation of hanging tributaries exist, thresholds in sediment-flux dependent erosion relations provide an excellent opportunity to compare theoretical predictions against field data.

Though it is difficult to distinguish between steady state channel profiles predicted by different stream incision rules, each incision rule demonstrates a unique behavior during its transient response to disturbance [*Howard and Kerby*, 1983; *Stock and Montgomery*, 1999; *van der Beek and Bishop*, 2003; *Whipple, et al.*, 2000; *Whipple and Tucker*, 2002]. To evaluate the real-world applicability of any particular incision rule, it thus becomes necessary to compare the predicted transient response with field observations from a disequilibrium landscape with a know age and type of disturbance. The direct comparison between modeled and observed transient landscapes thus provides an excellent opportunity to recognize the strengths and weaknesses of the present formulations for stream incision (figure 1).

In this paper, we provide a comparative analysis of how four different stream incision rules respond to two different scenarios for base level fall. Two of the four incision rules utilize simplified versions of two recent sediment flux dependent incision rules [*Gasparini, et al.*, 2006; *Gasparini, et al.*, submitted; *Parker*, 2002; *Sklar and Dietrich*, 1998; *Sklar and Dietrich*, 2004]. For reference, we also model the channel response to base level fall using the detachment-limited stream power incision rule e.g.[*Howard and Kerby*, 1983; *Whipple and Tucker*, 1999] and a simplified transport-limited incision rule [*Paola, et al.*, 1992a; *Tucker and Bras*, 1998; *Willgoose, et al.*, 1991].

Because the incision rate in three of the four studied incision rules is directly dependent on sediment flux, we find it advantageous to employ CHILD [*Tucker, et al.*, 2001a; *Tucker, et al.*, 2001b], a 2-dimensional landscape evolution model (figure 1d) where changes in sediment production and transport capacity are explicitly accounted for during the transient response [*Gasparini, et al.*, 2006; *Gasparini, et al.*, submitted]. This study provides an in depth utilization of a landscape evolution model to examine the interaction between the behavior of the trunk stream and the consequent response of the tributary channels.

The two base level fall scenarios represent end members for the range of forcing an incising river may experience in response to a sudden, but finite pulse of base level fall. Gasparini et al. [submitted] considers primarily mainstem response to a sustained increase in rock uplift rate, as opposed to the finite base level falls considered here. Our two scenarios help evaluate the relative sensitivities of the 4 models to the type of base level fall and allow us to mimic tributary networks nested within a larger catchment. In the first scenario, we subject the modeled channels to an instantaneous drop in base level at the outlet. In order to provide comparison to typical field settings (including the Waipaoa River basin, introduced earlier), we elect to drop base level 100 meters or ~1/2 of the modeled basin's steady-state fluvial relief. Instantaneous base level fall could result from stream capture or surface rupture along a fault [*Sklar, et al.*, submitted]. In the second scenario, we examine the channel response following a finite prolonged period of base level fall. In this scenario, we impose a 10 fold increase in uplift rate and allow the model to run until the accumulated rock uplift is equivalent to the magnitude of the instantaneous base level fall. This type of type of disturbance could result from sea level fall over the shelf-slope break [*Snyder, et al.*, 2002] or a temporary increase in uplift rate and mimics the base level fall signal that would be felt at the outlet of a sub-catchment nested within a basin too large for us to simulate that is subject to an instantaneous base level fall at its outlet.

We begin by our analysis by introducing the 4 stream incision rules utilized in the model and discussing the simplifications we made to these rules to allow us to write analytical solutions. We then outline the dependence of each model's incision rate on local slope and discuss the presence or absence of theoretically predicted instabilities. After a brief discussion of the utility and mechanics of the CHILD model, we present, for each of the 4 incision rules, the transient response of the trunk and tributaries following the two base level fall scenarios. Our subsequent discussion focuses on the interaction between trunk stream incision and the development of hanging tributary valleys.

Stream Incision Rules

Though our primary motivation is to understand the transient response to base level fall of channels governed by sediment-flux-dependent bedrock incision rules, we also provide, for comparison, a discussion of the transient response in both the standard detachment-limited stream power incision rule and a simplified transport-limited stream incision rule. This comparative analysis emphasizes the unique attributes and sensitivity of the sediment-flux dependent incision models. In the following section we introduce all four of the stream incision rules employed in the CHILD landscape evolution model.

In all cases, we are modeling the potential development of extremely steep channel reaches. Under these conditions, the small angle approximation, where $\sin(\alpha) = \tan(\alpha)$ = Slope, and α is the channel gradient in radians, is no longer valid. In all equations that follow in this text, the variable S stands for $\sin(\alpha)$. In figures, we plot data relative to the more intuitive variable, slope, which is equal to $\tan(\arcsin(\sin(\alpha)))$.

Transport-Limited and Stream Power Incision Rules

We only briefly discuss of the transient behaviors of the transport-limited and the detachment-limited stream power incision rules as these behaviors have been well explored by other workers, [*Howard and Kerby*, 1983; *Whipple and Tucker*, 1999, 2002]. In this analysis, we use a simplified rule for transport-limited incision where the volumetric sediment transport capacity is a power law function of unit stream power. We assume a negligible threshold of motion for the floods of interest. In this model, incision is a consequence of the stream's excess transport capacity [*Paola, et al.*, 1992b; *Willgoose, et al.*, 1991] . This relation does not include a critical shear stress, τ_{crit} and posits that Q_t, the volumetric sediment flux, is a power law function of A, the upstream drainage area, and S, the sine of the channel slope:

$$Q_t = K_t A^{m_t} S^{n_t} \qquad (1)$$

where K_t is a dimensional transport coefficient describing the erodability of the stream and m_t and n_t are dimensionless positive constants. We hold $n_t = 1$ in all simulations, consistent with bedload transport conditions. Given that the change in elevation is a consequence of the difference between rock uplift rate, U, and the downstream divergence of sediment flux, we can solve for how incision rate varies as a function of slope,

$$\frac{dz}{dt} = U - \left(\frac{1}{1-\lambda_p}\right)\frac{d}{dx}\left[\frac{1}{W}(K_t A^{m_t} S^{n_t})\right] \tag{2}$$

where λ_p is the sediment porosity and W is the channel width. All calculations of channel width in this paper (both in analytical and numerical analysis) utilize a power-law relation between width and drainage area, $W = k_w A^b$. Given the form of the incision relation formulated above and $n_t = 1$, incision rate at a particular drainage area is a linear function of slope (figure 2). Because this incision relation has the form of a nonlinear diffusion equation, an over-steepened reach created by base level fall is expected to decay rapidly as it propagates upstream.

The stream-power incision rule assumes that transport capacity well exceeds the imposed sediment load and thus the rate of channel incision is limited simply by the channel's capacity to detach bedrock from the channel bed [e.g. *Howard and Kerby*, 1983; *Whipple and Tucker*, 1999]. In this relation, there is no explicit functional dependence of the incision rate on the sediment flux. Changes in bed elevation are determined by differencing the background rock uplift rate, U, and the bedrock incision rate. Bedrock incision rate is a power law function of A, the upstream drainage area, S, the local slope:

$$\frac{dz}{dt} = U - KA^m S^n \tag{3}$$

where K is a dimensional coefficient describing the erodability of the channel bed as a function of rock strength, bed roughness and climate, and m and n are dimensionless positive constants.

This relation can be rewritten into the same functional form as used to describe kinematic waves [*Rosenbloom and Anderson*, 1994]. As a consequence, a base level fall signals modeled with the stream-power incision rule propagate upstream through the network as discrete waves. The form of the over-steepened reach may evolve as it propagates upstream depending on the value of n [*Tucker and Whipple*, 2002; *Weissel and Seidl*, 1998]. As discussed for the transport-limited incision model, the stream power incision rate at any given drainage area is linearly dependent on slope (for $n = 1$, as used here) (figure 2).

Sediment-Flux Dependent Models for Channel Incision

Sediment-flux dependent incision rules introduced by Sklar and Dietrich [1998; 2004], Whipple and Tucker [2002] and Parker [2002] explicitly model the dual role of sediment in bedrock channel incision. In these incision rules, when sediment flux exceeds transport capacity, an immobile blanket of sediment covers and armors the bed, inhibiting incision. For low volumes of sediment flux, insufficient tools are available to impact and abrade the channel bed.

We distinguish two classes of sediment-flux dependent incision rules: one where saltation dynamics (on a planar bed) plays a fundamental role in channel incision and another where although incision is still accomplished by abrasion alone, saltation dynamics are not explicitly modeled. We utilize simplifications (derived in Gasparini et al., in review) of the river incision models proposed by Sklar and Dietrich [1998; 2004] and Parker [2002] in order to facilitate analytical exploration and to provide direct comparability to the behavior of the previously discussed incision rules [e.g. *Gasparini, et al.*, 2006]

Saltation-Abrasion Incision Rule

The incision rule developed by Sklar and Dietrich [1998; 2004] provides a process-specific, mechanistic relation for bedrock incision by the abrasion of saltating bed load of a single grain size on a planar bed. In this rule, the incision rate is an explicit function of the flux of kinetic energy normal to the bed (impacts) and the uncovered fraction of the bed exposed to those impacts. The incision rate for this model, I_{SA}, is written as the product of three measurable terms: the volume of rock prepared for transport per particle impact, V_i, the rate of particle impacts per unit area, per unit time, P_r, and the fraction of the channel's bedrock bed exposed to incision, F_e.

$$I_{SA} = V_i P_r F_e . \tag{4}$$

Each term in this expression can be expanded and expressed as a function of excess sheer stress. Combining the three above terms back into a simple expression, Sklar and Dietrich express the the bedrock incision rate as

$$I_{SA} = \left[\frac{R_b g}{25\varepsilon_v}\right] \cdot \left[\frac{Q_s}{W}\left(1-\frac{Q_s}{Q_t}\right)\right] \cdot \left[\left(\frac{\tau^*}{\tau_c^*}-1\right)^{-0.52}\right] \tag{5}$$

where R_b is nondimensional buoyant density of the sediment, g is gravitational acceleration, ε_v is the energy required to erode a unit volume of rock, Q_s is volumetric sediment flux, W is channel width, Q_t is volumetric sediment transport capacity, τ^* is nondimensional shear stress and τ^*_c is nondimensional critical shear stress. We can observe the dual role of sediment flux in determining incision rate in this expression by recognizing that as sediment flux, Q_s, goes to zero, so does I_{SA}. As well, if Q_s at any time exceeds the transport capacity, Q_t, then the system becomes depositional and the incision rate becomes negative. These are important considerations during the transient response as sediment flux can vary dramatically as hillslopes and tributaries respond to trunk stream incision. The negative exponent on the third term, excess transport stage, is a consequence of the explicit saltation dynamics and provides that incision rate decreases for increasing excess bed shear stress, all else being equal.

In our simplification, which we generically refer to as the saltation-abrasion rule, we group the terms in the first bracket of equation 5 as constants that characterize the channel's erodability, K_{SA}. We group the two terms in the second bracket as $f(Q_s)$, an expression reflecting the role of sediment flux in setting the incision rate. The first term in this second bracket, (Q_s/W), quantifies the volume of sediment per unit width available as tools and the second term, (1-(Q_s/ Q_t)), describes the cover-effect, where increasing sediment flux relative to transport capacity diminishes the incision rate. The terms in the third bracket describe the excess transport stage and can be rewritten as a power function of shear stress and thus expressed as a power law function of drainage area and slope (derivation provided in Gasparini, et al., in review). Our simplified expression for equation 5 shares the form of the generic sediment-flux dependent incision rule presented by Whipple and Tucker (2002),

$$I_{SA} = K_{SA} f(Q_s) A^{-1/4} S^{-1/2} \tag{6}$$

where, following Sklar and Dietrich [1998; 2004],

$$f(Q_s) = \frac{Q_s}{W}\left(1 - \frac{Q_s}{Q_t}\right). \tag{7}$$

The steady state channel profile, slope-area relation and dependence of incision rate on slope for this simplified expression closely approximate those predicted by the original Sklar and Dietrich incision rule. The slope of the channel at large drainage area asymptotically approaches that predicted by the purely transport-limited model. The steady state longitudinal form of the channel deviates from the transport-limited slope only at small drainage areas in upper reaches of the channel network (Gasparini et al., figure 1, in review). However, under steady state conditions, there is a critical drainage area below which channel slopes become infinite. This critical area, as described in detail by Gasparini et al., [equation 30; in review], is a consequence of two characteristics of small drainage areas: low sediment supply and high slopes. Because incision depends on saltating bedload, these two factors limit the interaction of sediment with the bed, thus limiting incision. In the CHILD model, diffusive hillslope processes are responsible for sediment production and transport at drainage areas less than this critical area. We reserve further discussion of the dependence of incision rate on slope for the proceeding section on model instabilities.

Simplified Abrasion Incision Rule

Though the full formulation of the Parker bedrock incision rule addresses both bedload abrasion and plucking of weakened bed material [*Parker*, 2002], we simplify the model by focusing only on the bedload abrasion process (Gasparini, et al., in review). This facilitates simpler analytical solutions and a more direct comparison between the Parker incision rule and that of Sklar and Dietrich. Our simplified abrasion rule, modified from the Parker incision rule [*Parker*, 2002], is very similar to the modified version of the Sklar and Dietrich incision rule presented above, except that the exponents *m* and *n* are equal to zero:

$$I_{AD} = K_{AD} \frac{Q_s}{W}\left(1 - \frac{Q_s}{Q_t}\right) \tag{8}$$

where K_{AD} is a dimensional constant describing the erodability of the channel bed. Note that in order to have a constant incision rate, I_{AD}, at steady state , the $f(Q_s)$ term in equation 8 (everything but the K_{AD}) must be constant. Like the saltation-abrasion incision rule, the Simplified Abrasion incision rule predicts that at large drainage areas, the channel slope is determined by the system's sediment transport capacity. The steady state Simplified Abrasion incision rule also predicts a critical drainage area at which slopes become infinite as a consequence of high slopes and low sediment fluxes [*Gasparini, et al.*, submitted]. Like the saltation-abrasion incision rule, the Simplified Abrasion rule's threshold drainage area increases with decreasing values of the dimensional constant, K_{AD} (stronger rocks). Unlike the saltation-abrasion incision rule, the Simplified Abrasion rule's critical drainage area is insensitive to uplift rate. We reserve further discussion of the dependence of incision rate on slope for the proceeding section on model instabilities.

Predicted Transient Instabilities in the Incision Rules

As there are no transient instabilities in the transport-limited and the stream power incision rules as we have formulated them above, this section outlines instabilities predicted to occur in the two sediment-flux dependent incision rules. In each case, we assume that during the initial response to base level fall, changes in sediment flux lag behind the profile adjustment (or $Q_s = Q_{s\text{-}old}$). If this assumption is violated and sediment flux does not lag during adjustment, the instabilities predicted below likely underestimate threshold slopes or overestimate threshold drainage areas.

Instabilities in the Saltation-Abrasion Incision Rule

In figure 2, we demonstrate that the steady state incision rate predicted by the saltation-abrasion incision rule of Sklar and Dietrich (1998, 2004) initially increases from extremely small values at low slopes, past the steady state slope at S_{ss}, to a maximum incision rate at S_{peak}. For slopes greater than S_{peak}, the incision rate decreases monotonically with increasing slope, reaching an incision rate equal to the background uplift rate at S_{hang}. The form of the solid-line curve in figure 2 provides the basis for the instabilities observed in the saltation-abrasion incision rule. Note that incision rate is a cubic function of channel slope.

The first of two instabilities in the saltation-abrasion incision rule occurs if a channel's slope exceeds the slope required for the incision rate to keep pace with the background uplift rate, S_{hang} (figure 2). Beyond this critical slope, the stream can no longer effectively erode the bed at a rate sufficient to keep pace with the background uplift rate, U_{old}, and there is a runaway negative feedback that leads to the formation of a permanent hanging valley (a waterfall). Both steady-state channel slope, S_{ss} and S_{hang} can be recognized as two of the three roots of equation 6 when $I = U$ (figure 2). Thus we can derive and expressions for S_{ss} and S_{hang} by setting $I = U$ and thus $Q_s = \beta A U$, and then solve equation 6 for slope. To do this, we use a power-law width-discharge relation and a power-law discharge-area relation to set $W = k_w k_q^{\,b} A^{bc}$ as outlined by Whipple and Tucker [1999]. The solution is found by making the substitution $x = S^{1/2}$ and then rearranging equation 6 (the saltation-abrasion incision rule) into a relatively simple cubic form,

$$x^3 - S_t^{-1} x = -S_t^{-1}\left(\frac{1}{K'\beta A^{1-bc+m}}\right). \qquad (9)$$

In equation 9

$$S_t = \frac{\beta U}{K_t} A^{1-m_t} \tag{10}$$

where S_t defines the transport-limited slope [e.g. *Whipple and Tucker*, 2002], $K' = K/(k_w k_q^b)$, and the cubic form of equation 9 is $x^3 + px = j$. The three real roots of this cubic equation are:

$$S_1 = \left(2\sqrt{-\left(\frac{1}{3}p\right)} \cos\left(\frac{\theta + 2\pi}{3}\right)\right)^{-2} , \tag{11}$$

$$S_{ss} = \left(2\sqrt{-\left(\frac{1}{3}p\right)} \cos\left(\frac{\theta}{3}\right)\right)^{-2} \text{ and} \tag{12}$$

$$S_{hang} = \left(2\sqrt{-\left(\frac{1}{3}p\right)} \cos\left(\frac{\theta + 4\pi}{3}\right)\right)^{-2} \tag{13}$$

where

$$\theta = \cos^{-1}\left(\frac{\frac{1}{2}j}{\sqrt{-\frac{1}{27}p^3}} \right). \tag{14}$$

These roots define the three slopes at which the incision rate is equivalent to the background uplift rate (figure 2). The first root, S_1, is unphysical and is not highlighted in our figure. The second root, S_{ss}, is a physically meaningful solution that describes the channel slope at steady state. At this slope, the channel exhibits a stable, partially covered bed and can erode and transport material at a rate sufficient to keep pace with the background uplift rate. The third root, S_{hang}, is also physically valid but describes the critical slope above which the incision rate decreases below the background uplift rate (figure 2), facilitating the formation of a permanent hanging valley. At slopes higher than S_{hang}, the saltating bedload no longer impacts the bed effectively or frequently enough to maintain a sufficiently high incision rate. Basins upstream of these over-steepened reaches are terminally divorced from the lower reaches. Combining equation 9, 13 and 14, we derive the slope of S_{hang} to be

$$S_{hang} = \left[2\sqrt{\frac{1}{3S_t}} \cdot \cos\left(\frac{1}{3}\cos^{-1}\left(\frac{\frac{-1}{2S_t K' \beta A^{1-bc+m}}}{\sqrt{\frac{1}{27S_t^3}}} \right) + \frac{4}{3}\pi \right) \right]^{-2} . \tag{15}$$

As presented in figure 3, the slope required to form a permanent hanging valley increases with drainage area (the increase in S_{hang} with drainage area is exactly commensurate with the decrease in S_{ss} with drainage area). This implies that larger tributaries will be able to keep up with a given pulse of mainstem incision while smaller tributaries experiencing the same magnitude pulse would not.

The second transient instability associated with the saltation-abrasion incision model creates temporary hanging valleys that fail to keep pace with mainstem incision for a period of time, but eventually recover and equilibrate to the mainstem. These occur when the transient pulse of incision increases channel slopes to values greater than S_{peak}, but less than the previously discussed S_{hang} (figure 2). We can derive the value for S_{peak} at any particular drainage area by solving for where the change in incision rate with slope equals zero. We find that this value is three times the initial transport limited slope or

$$S_{peak} = \frac{3\beta}{K_t} U_{old} A^{1-m_t} . \tag{16}$$

Because m_t is greater than one (table 1), the negative exponent on area in equation 16 means that S_{peak} increases with decreasing drainage area. This implies that the difference between S_{peak} and S_{hang} decreases with drainage area, as does the difference between S_{ss} and S_{hang}. This implies that smaller tributaries are far more susceptible to the formation of permanent hangs than larger tributaries which will either respond quickly in an essentially transport-limited manner or will develop temporary hangs and then recover. For an instantaneous pulse of base level fall, a temporary hanging valley forms as the pulse of incision creates tributary outlet slopes between S_{peak} and S_{hang}. The gradual subsequent incision of this over-steepened reach exceeds the background rock uplift rate, and thus decreases the height and maximum gradient of the over-steepened reach, resulting in a positive feedback loop in which gentler slopes promote greater incision rates and thus more rapid decay of the temporary hanging valley knickpoint. This positive feedback leads to the eventual recovery of the temporary hanging valley to a graded condition. Our derivations of S_{peak} and S_{hang} assume we are observing the initial response of the channel (where $Q_s = Q_{s\text{-}old}$); any partial communication of the incision signal to the upper basin will increase sediment delivery and ultimately increase the predicted values of both S_{peak} and S_{hang}.

Instabilities in the Simplified Abrasion Incision Rule

Unlike the saltation-abrasion incision rule [adapted from *Sklar and Dietrich*, 1998; *Sklar and Dietrich*, 2004], the simplified abrasion incision rule [adapted from *Parker*, 2002] does not have a hump or discrete maximum value in the relation between incision rate and slope (figure 2). Though n, the exponent on slope in equation 8, is zero, there is still a functional dependence on slope through the Q_t term defined in equation 1. In the Simplified Abrasion incision model, the steady state incision rate increases asymptotically with slope toward a maximum value, I_{max} (see figure 2):

$$I_{\max} = \frac{K_{AD}\beta}{k_w k_q^b} A^{1-bc} U_{old} . \tag{17}$$

This asymptotic approach to I_{max} prevents the incision rate from ever declining below the background uplift rate in response to channel steepening. Thus the only way the channel could create a permanent hanging valley (or waterfall) would be in the unlikely case where the sediment flux in the upper basin decreases to zero.

Similarly to the saltation-abrasion model, temporary hanging valleys will form in the Simplified Abrasion incision rule if the transient incision rate exceeds the maximum incision rate associated with the initial sediment flux. Solving the equation 17 for drainage area, we find the maximum drainage area at which a temporary hanging valley could form:

$$A_{temp} = \left(\frac{k_w k_q^b}{K_{AD}\beta} \frac{I_{\max}}{U_{old}} \right)^{\frac{1}{1-bc}}. \tag{18}$$

Following a pulse of incision, the over-steepened reach at A_{temp} decreases in slope because the mainstem incision rate returns to just balancing the background uplift rate (U_{old}) while the tributary continues to incise at a rate equal to I_{max}, resulting in a progressive decay of hanging valley height and the maximum slope of the over-steepened reach. In addition, sediment flux from upstream increases in response to the incision, resulting a positive feedback, accelerating the upper-lip of the over-steepened reach. This results in the eventual, if asymptotic, readjustment of the tributary to graded conditions following the transient pulse of incision.

Evidence from the Waipaoa River and experimental studies suggest that through most of the channel network, the base level fall signal experienced at a point along the channel is not a discrete, on/off pulse of incision but rather gradually builds toward a maximum incision rate and then declines [*Berryman, et al.*, in preparation; *Crosby and Whipple*, in press; *Gardner*, 1983]. This rise and fall in the wave of incision experienced at a tributary junction allows an initial signal of incision to propagate up into the tributary before the large magnitude incision rate potentially results in the formation of a knickpoint, or hanging valley, that effectively isolates the tributary from the mainstem. This partial communication of the initial incision signal increases tributary sediment flux enough so that when the large magnitude incision rate creates an over-steepened reach, the hanging tributary's elevated sediment flux is sufficient to allow recovery.

Numerical Simulations of Transient Landscape Response

We utilize the CHILD numerical landscape evolution model to compare the transient responses of landscapes governed by four different stream incision rules [*Gasparini, et al.*, submitted; *Tucker, et al.*, 2001a; *Tucker, et al.*, 2001b]. The CHILD numerical landscape evolution model provides an explicit accounting of sediment production and transport at every model node, thus offering an excellent tool for exploring the transient response of sediment-flux-dependent channel incision rules. Triggered by base level fall, channel incision destabilizes hillslopes and generates a spatially and temporally complex sediment flux response. During each model run, at any point along the mainstem channel, the sediment flux fluctuates as a consequence of both the local adjustment of destabilized hillslopes and the integrated, upstream network response to the pulse of incision. This unsteady sediment flux response directly impacts the continued fluvial communication of the transient base level fall signal through the system.

Though numerical landscape evolution models are useful for studying the interaction between sediment production, sediment transport and channel incision at the network scale, they also present limitations. In our application of the CHILD numerical model, we study the transient response within large basins (2.5×10^7 m^2) (figure 1d) and thus use a large node spacing of ~100m. This large node spacing, and the subsequent numerical diffusion, results in the progressive decay of the over-steepened reach as it propagates upstream. Though this influences evolution of the form of the transient signal in the stream power incision model (figure 4e-h), we are less concerned in the other models. As discussed in Whipple and Tucker [2002] and Gasparini et al. [2006; submitted], large drainage area regions in sediment flux dependent incision rules respond to pulses of incision largely in a transport-limited manner. Because this diminishes the impact of numerical diffusion on the modeling of the sediment-flux-dependent incision rules (our main focus), we are confident that numerical diffusion does not limit the findings of this study. Another unintended consequence of large node spacing is that the instantaneous base level fall at the outlet creates a step whose slope is the ratio between the vertical base level fall and the horizontal node spacing. For example, an instantaneous 100m base level fall, instead of being near-vertical, has a gradient on the order of ~1.

In order to test the response of each incision rule to the base level fall scenarios discussed above, we first created four steady state landscapes, each fully adjusted to a particular incision rule. To do this, each model landscape starts as a 5000 meter by 5000 meter square of random, low amplitude topography with the outlet in one corner and zero-flux edges. Each initial surface is then subjected to uniform and steady rock uplift until the network and hillslopes stabilize. These four steady state landscapes provide the initial condition for the base level fall experiments. As we only study the basin's response to finite base level fall, and not a sustained change in rock uplift rate [as explored by *Gasparini, et al.*, submitted], all model parameters are identical before and after the disturbance.

For each incision rule, we first examine the responses of the trunk stream to the two base level fall scenarios. Next, we focus on how the trunk stream's response influences tributary response as a function of their size and position within the drainage network. The modification of the incision signal as it propagates up the trunk stream provides each tributary with a unique base level fall signal, potentially resulting in significant along-stream variation in tributary response.

Transport Limited Channel Incision Rule

The transport-limited incision rule, though computationally expensive in CHILD, is relatively robust over a wide range of parameters (table 1). Because the transport-limited incision rule is a nonlinear diffusion equation, the over-steepened reach created by the instantaneous base level fall decays rapidly as the convex kink in the channel profile propagates upstream (figure 4a-b, 6a). As a consequence, the highest incision rates of all four incision models are observed during the initial transport-limited adjustment to instantaneous base level fall. As the base level fall signal propagates upstream, the location of highest incision rate moves surprisingly slowly upstream (figure 4c-d). Though the maximum incision rates are extremely high in this model, the location of the peak incision rate does not sweep upstream at the same rate as was observed in the other models. This is a consequence of the lower reaches being overwhelmed by the high sediment fluxes generated during the transient adjustment.

For the scenario where base level lowers progressively through time as outlined in the Approach and Scope section (figure 4b,d; gray lines), much of the channel network efficiently steepens as the transport-limited channel responds to the temporarily higher uplift rate. No distinct knickpoints are created along the channel profile (figure 4b). Once the uplift rate returns to the lower background value, a diffuse pulse of incision sweeps up the mainstem channel and gradually decreases the slope of the transiently over-steepened channels (figure 4d, 6b). As in the instantaneous base level fall example above, the highest incision rates occur near the outlet.

The response of the landscape to the two base level fall scenarios is very similar and it could be argued that the instantaneous fall scenario, after a number of time steps, closely resembles the initial form of the progressive base level fall scenario. The only significant difference between the two cases is that the response in the progressive base level fall model is slightly faster than the response to an instantaneous base level fall. Neither of the two models creates temporary or permanent hanging valleys during the transient response. The only delay in response is a consequence of the lag in hillslope sediment delivery following the pulse of incision.

Stream Power Incision Rule

The detachment-limited stream power incision rule runs much more efficiently than any of the other incision rules in CHILD because sediment flux does not need to be explicitly accounted for at every node. The CHILD model is relatively robust to the parameters chosen for the stream power incision rule (table 1). As discussed above and expanded upon below, the only limitation of modeling the stream power incision model in CHILD is that numerical diffusion rapidly attenuates the sharp breaks in slope that should persist during the transient response (figures 4e-h).

For conditions where $n = 1$, the transient response following instantaneous base level fall should resemble a shock wave where the imposed step in the channel profile retreats upstream with out changing form (figure 1e, dashed lines). For values greater than or less than one, the form of the step is modified as it retreats upstream [*Tucker and Whipple*, 2002; *Weissel and Seidl*, 1998]. In the CHILD model, the step propagates upstream, at a rate that is a power law function of drainage area [*Rosenbloom and Anderson*, 1994], and slows at tributary junctions but never creates permanent or temporary hanging valleys (figure 6c) [see also *Niemann, et al.*, 2001]. If numerical diffusion could be avoided, the step created by instantaneous base level fall would propagate through out the basin without diminishing in magnitude (figures 4e-h, dashed lines). The transient signal propagates throughout the entire extent of the channel, reaching the headwaters without ever forming a hanging valley.

During progressive base level fall (figures 4f, 4h, 6d), the channel steepens to a higher slope appropriate to the temporarily higher rock uplift rate (relative to base level). Once the base level fall period has ended, the over-steepened reach propagates upstream in the same manner as the discrete step did in the instantaneous base level fall scenario. The over-steepened reach maintains its form as long as $n = 1$. Just as with the instantaneous base level fall, the over-steepened reach is modified during its upstream migration as a consequence of numerical diffusion (figure 4f-h, compare to dashed lines). For tributaries, there is insignificant differences between the instantaneous and prolonged base level fall responses (figures 6c,d)

Saltation-Abrasion Incision Rule

Both the steady state form and transient response of CHILD landscapes governed by the saltation-abrasion incision rule are highly sensitive to the parameters used. Modeled landscapes with realistic channel slopes and network densities are created only by a narrow combination of values for hillslope diffusivity, K_t and K_{SA} (table 1). This sensitivity is a direct consequence of the inclusion of saltation dynamics in the model (and thus the negative exponents on drainage area and slope, equation 6).

When subjected to a 100 meter instantaneous base level fall (figure 5a), the discrete step created between the first and second nodes of the mainstem fails to retreat. As the model runs forward, the step very slowly begins to decrease elevation. For each increment of lowering at the top of the step, the small incision signal is rapidly transmitted through the upper portion of the basin (figures 5c and 6e). The reason the step fails to retreat and instead slowly lowers is because the slope created by the 100m base level fall is between S_{peak} and S_{hang} on the plot of incision rate versus slope (figure 2). If base level fall were extreme (creating a gradient larger than ~14), the resulting slope at the outlet would exceed S_{hang} (figure 3, see S_{hang} for 2.5 x 10^7 m^2) and the incision rate at the step would drop below the background uplift rate and the entire network would be permanently hung above the outlet (a result confirmed in several simulations not reported here). For instantaneous base level falls significantly less than 100m, given our 100m cell size, the slope at the outlet is smaller and closer to S_{peak} and consequently the step is worn down faster than observed for the 100m step. In our model run for instantaneous base level fall, because only a fraction of the total signal propagates quickly upstream, these adjustments are too small to create steep slopes or hanging valleys in any tributary (figure 6e). We found it impossible for an instantaneous base level fall to create a temporary or permanent hanging tributary using the saltation-abrasion incision model, except at the basin outlet itself. This reflects mostly the restricted size of drainage basins we can simulate with the moderate resolution of 100-m cells. If we consider that the outlet corresponds to a tributary confluence into a much larger drainage basin, an instantaneous base level fall at the outlet of this larger basin would be felt as a progressive base level fall at the outlet of our simulated catchment; a scenario we consider next.

For progressive base level fall (figures 5b, d and 6f), only moderate slopes develop at the outlet. Because these moderate slopes never exceed $S_{peak,}$ there is a rapid upstream propagation of the base level fall signal without ever creating a permanent or temporary hanging valley at the basin outlet. If the prolonged mainstem base level fall produces an outlet slope close to S_{peak}, then the mainstem channel incises at its maximum potential rate, propagating the incision signal rapidly through the network. Under this circumstance, rapid incision in the mainstem can create tributary slopes in excess of their local S_{hang} or

S_{peak}, resulting in the formation of permanent or temporary hanging valleys (figures 7 and 6f respectively). For slow rates of progressive base level fall, the outlet slope is significantly less than S_{peak} and the transient signal propagates quickly through the network without forming hanging tributaries except in the smallest of tributaries (figure 6f). It is important to note that for all progressive base level fall scenarios, the mainstem incision signal decreases in both amplitude and retreat rate as it propagates upstream through the network (figure 5d). This means upper-basin tributaries feel a smaller maximum incision rate for a longer duration. As discussed earlier, S_{hang}, the slope necessary to create a permanently hanging valley decreases with drainage area. This allows smaller, upper-basin tributaries experiencing relatively low mainstem incision rates to form hanging valleys while larger tributaries maintain a graded connection to the mainstem, consistent with field observations in New Zealand [*Crosby and Whipple*, in press] and Taiwan [*Wobus, et al.*, in press].

We can adjust parameters in the saltation-abrasion model so that rapid mainstem incision creates slopes in excess of S_{hang} at tributary junctions, allowing permanently hanging tributaries develop. As seen in figure 7, tributaries become elevated above the mainstem when their junction is only capable of incision at a rate below the background uplift rate. Because the base level fall signal is attenuated as it propagates upstream, we observe variations in response at different upstream distances, at different tributary drainage areas and for different base level fall magnitudes.

Simplified Abrasion Incision Rule

For all CHILD model runs using the Simplified Abrasion incision rule [adapted from *Parker*, 2002], permanent hanging valleys never form as a consequence of finite base level fall, as expected from the analytical solutions discussed above. Relative to model runs using the saltation-abrasion incision rule, landscapes predicted by the Simplified Abrasion incision rule are much more robust and stable over a wider range of parameter space (utilized values in Table 1). In the Simplified Abrasion incision model, the mainstem response to both instantaneous and progressive base level fall behaves similar to the channels governed by the transport limited incision rule (figures 5e, f). The step created by the instantaneous base level fall signal retreats and decays quickly as it propagates upstream (figures 5e). The plot of normalized incision rate shows the decrease in incision rate as the mainstem signal decreases slope during its upstream retreat (figure 5g). The tributaries (figure 6g) contain temporary hanging valleys that relax through incision at the upstream end of the over-steepened reach. By the end of the main wave of adjustment, the channel profile appears to have reached a steady form but it is still steeper than the equilibrium slope. This is a consequence of the excessive sediment flux still being delivered off the adjusting hillslopes. Though the mainstem channel may stabilize, it remains at disequilibrium until all aspects (channels and hillslopes) of the landscape return to equilibrium. Opposite of the response to the pulse of incision, this return to equilibrium is a top-down process where the hillslopes have to adjust before the channels can complete their adjustment.

During progressive base level fall, the mainstem channel rapidly adjusts to the imposed higher uplift rate by increasing gradient (figure 5f). When the initial conditions are restored, the form of the profile resembles one from halfway through the instantaneous base level fall response. This gives the progressive base level fall scenario a "head start" in its adjustment back toward steady state. This is reflected in a faster basin response time and a smaller maximum incision rate (figure 5h) and the lack of hanging tributaries (figure 6h).

Temporary hanging valleys do form around A_{temp} in landscapes modeled in CHILD with the Simplified Abrasion incision rule as a consequence of the lag time in sediment flux delivery from tributary hillslopes (figure 6g). Without sufficient sediment delivered from upstream, the incision rate at an over-steepened tributary junction decreases and a temporary hanging valley forms. As the tributary's hillslopes increase sediment delivery in response to the initial partial communication of the incision signal upstream of the over-steepened junction, the incision rate in the hanging valley increases. Incision at the lip of the over-steepened reach eventually leads to the decay of the hanging valley and re-establishment of graded conditions.

Discussion

Our analytical results and numerical simulations demonstrate that sediment-flux dependent incision rules present viable mechanisms for the formation of both permanent and temporary hanging valleys. In landscapes governed by the two sediment-flux dependent incision rules used here, the progressive modification of the base level fall signal as it propagates up the mainstem determines the magnitude and duration of the incision signal at each tributary junction. Lower-basin tributaries, located near the outlet where the base level fall initiates, feel the fastest incision rates for the shortest duration and thus have the greatest potential to create large drainage area hanging valleys. The pulse of incision is attenuated in a different manner for each incision model as it propagates upstream (figure 4c,d,g,h and 5c,d,g,h), thus upper-basin tributaries experience a long-duration signal at a lower incision rate. This upstream attenuation of the mainstem incision signal plays an important role in determining whether particular tributaries become either temporary or permanent hanging valleys. In the saltation-abrasion model, we demonstrated earlier that a hanging valley forms when the tributary channel gradient exceeds S_{hang}. This threshold gradient decreases at lower drainage areas (figure 3). We also demonstrated that the slopes required to form temporary hanging valleys in both the saltation-abrasion and the Simplified Abrasion incision rules decrease at lower drainage areas. Consequently, following a single pulse of incision, the resulting basin-wide distribution of permanent and temporary hanging valleys is governed by a complicated competition between four factors: the magnitude of the initial base level fall at the outlet, the rate of upstream attenuation of that incision signal in the mainstem, the lag time of the sediment flux response and the non-systematic variation in tributary drainage areas within the channel network.

If the magnitude and duration of the incision signal changes as it propagates upstream, the sediment-flux dependent incision rules suggest that tributaries of identical drainage area positioned at different locations in the basin will not have the same propensity to form hanging valleys. In a study of hanging valleys in the Coast Ranges of Taiwan, Wobus et al. [in press] recognized that in this field area, most hanging tributaries had a trunk-to-tributary ratio greater than ~10:1 [*Wobus, et al.*, in press; figure 11]. Lower in the mainstem, our analysis does support the observation that mainstem drainage area must be significantly greater than tributary drainage area to create high enough tributary slopes to form hanging valleys. Our analysis also provides evidence that this trunk-to-tributary drainage area ratio for hanging valley formation is sensitive to the four factors listed above, especially in the upper portion of the network where the tributaries and the trunk stream may be of similar drainage area and both close to forming hanging valleys.

Comparing the channel response to both instantaneous and progressive base level fall signals, we find that extending the time of base level fall diminishes the magnitude of the maximum incision rate and extends the duration of the pulse of incision. By distributing the same magnitude mainstem base level fall signal over a longer time period, tributary junctions have a greater probability of keeping pace with mainstem incision and thus a lower likelihood of creating temporary or permanent hanging valleys.

As anticipated, the transport-limited and detachment-limited stream power incision rules did not generate either permanent or temporary hanging valleys. In these model runs, the base level fall signal was communicated all the way to the headwaters. In the transport-limited model, the response to base level fall was extremely rapid with the only significant lags due to the armoring of the channel by high initial sediment fluxes and the delay in hillslope adjustment to the incising streams. The model basin governed by the detachment limited incision rule did not create hanging valleys at tributary junctions that grew in magnitude or slowed retreat rate any more than anticipated by the celerity model for detachment-limited knickpoint retreat [*Crosby and Whipple*, in press; *Rosenbloom and Anderson*, 1994; *Whipple and Tucker*, 1999].

Limitations and Recommendations for Future Work

In future analyses, it would be advantageous to decrease node spacing to limit the affect of numerical diffusion on channel evolution. This could be achieved by running the code on faster machines or clusters or by simply budgeting longer run-times for each analysis. Although the CHILD model offers an excellent environment for testing the behavior of these sediment flux dependent incision rules, we recognize that the treatment of hillslope processes requires improvement. With more explicit hillslope erosion processes active, simulation of the rate and magnitude of sediment delivery to the channels following incision would be improved. As well, improved parameterization of hillslope processes will allow hanging tributaries created by thresholds in the sediment-flux dependent incision rules to continue to evolve due to non-fluvial processes (such as: sapping, mass wasting, weathering, debris flows, etc.), as observed in field sites.

Though all of our theoretical derivations and findings (Figures 2 and 3) explicitly treat slope as $\sin(\alpha)$, the numerical model simulations at present do not. We need to explore how the violation of the small angle assumption affects the results of the numerical model. At present, our findings are general, demonstrating that hanging valleys do or do not form depending on whether the tributary slopes predicted by the saltation-abrasion model exceed S_{hang}. At this point, the violation of the small angle approximation in the numerical model simply limits the applicability of parameters such as K_s or K_t to real world scenarios.

Future analyses would also benefit from examining the impact of spatially variable substrate erodability on the attenuation of pulses of incision and the retention or decay of hanging valleys. Adding this dimension to the analysis, brings it closer to representing the field sites we hope to better understand. Along these lines, it would be advantageous to consider more complicated signals of base level change. For example, how would multiple base level fall events staggered in time influence the sensitivity of tributaries to forming hanging valleys? The pre-disturbance slope of a tributary channel is an important factor in determining its sensitivity to a particular base level fall signal. Variations in tributary slope could occur as a consequence of local lithology in the tributary or as a consequence of incomplete adjustment to a previous transient signal. If the tributary is already over-steepened (but not hanging) from a previous incision event, then it might only require a lesser mainstem incision event to cause the tributary to hang. A well-adjusted tributary of similar drainage area, sediment flux and position in the network but lower slope would not share the same sensitivity to the lesser incision event.

Implications for Landscape Evolution and Field Observations

Analytical and numerical investigations of the response of sediment-flux dependent incision rules to base level fall provide significant evidence for the formation of over-steepened reaches at tributary junctions as a consequence of exceeding threshold conditions. Because modern field observations provide an incomplete record of past events, the utilization of theoretical analyses, as presented above, provide an effective mechanism for testing sparse field data against continuous modeled data. Our work in the Waipaoa River, New Zealand [*Crosby and Whipple*, in press] and in the Coast Ranges of Taiwan [*Wobus, et al.*, in press] provide useful datasets to compare with modeled behavior.

Numerous characteristics of the modeled behavior align well with our observations from the Waipaoa River basin [*Crosby and Whipple*, in press]. In this fluvial catchment, a pulse of incision initiated ~18,000 years ago has lowered base level throughout the main channels in the 2150 km^2 basin on the North Island of New Zealand. Some ~236 knickpoints separate hanging valleys from downstream adjusting reaches. Though the cause of the pulse of incision is unknown, observations from fluvial terraces suggest that trunk stream incision was progressive rather than instantaneous, as there is evidence for a gradual rise and fall in incision rate [*Berryman, et al.*, in preparation]. This evidence for progressive incision fits well with the modeled channel behavior because, even for a signal initiated as instantaneous base level fall, the transport limited nature of large drainage area reaches attenuates discrete signals as they propagate upstream.

In the Waipaoa, large magnitude (50-100m) knickpoints with well defined upstream lips are most frequently found just upstream of tributary junctions, as predicted by both sediment-flux dependent incision models. There are no observations of undercutting plunge pools or cap-rock conditions that would facilitate the maintenance of a discrete kink at the knickpoint lip. Field observations suggest that the minor retreat of the over-steepened reaches from the tributary junctions is dominated by non-fluvial processes including physical and chemical weathering, block failures and other mass wasting processes. This is contradicted by the fact that in both the Waipaoa River and Taiwan, we find that the distance the knickpoint retreated back from the mainstem is drainage area dependent[*Crosby and Whipple*, in press]. This observation may result from the combined contribution of fluvial and non-fluvial processes in the retreat and decay of hanging valleys.

In the Waipaoa, the tributary reaches upstream of the knickpoints have two distinguishing characteristics: (1) there is very little incision into either the bed of the hanging tributary or the knickpoint lip and (2) the hanging channels are bare with almost no sediment on planar bedrock beds. The first observation aligns well with the saltation-abrasion model for the case where the over-steepened reach is greater than S_{hang}. Alternatively, the observation could fit well with both models under the assumption that sediment production in the upper basins decreased dramatically and the hanging valleys have not produced sufficient tools to erode either the channel bed or the knickpoint lip. This is supported by the observation that the hillslopes in many of the hanging valleys only deliver sediment to their foot-slopes, not the channel. Extensive fill deposits border the hillslopes, storing the sediment derived from the hillslopes and inhibiting the transfer of material to the fluvial network.

Model simulations provide strong evidence that for channels governed by sediment-flux dependent incision rules, two mechanisms extend the basin's response time following disturbance. First, because sediment delivery from hillslopes appears to lag behind fluvial incision, the channel's return to equilibrium is prolonged as the hillslopes slowly reestablishes equilibrium sediment delivery. Analysis of slope-area data downstream of hanging tributaries reflects this disequilibrium condition as the channels are still adjusting to the insufficient sediment flux from the upper hanging basins. The second mechanism that extends the response time is the formation of hanging valleys in which potentially slower, lithology-dependent, non-fluvial processes dominate the knickpoint retreat process.

Our simulations also suggest that the concept of "knickpoint retreat rate" needs to be carefully evaluated as the communication of a pulse of incision is not a continuous or single-process phenomenon. Because of potential changes in erosion processes, measured retreat rates in the mainstem may have little bearing on those observed in tributaries. Multiple observations of transient features such as terraces and knickpoints provide a basin-wide context to help differentiate between actively retreating knickpoints and stagnant hanging valleys. Even with this basin-wide context, it is difficult to take locally measured retreat rates and scale them appropriately to constrain the rate of adjustment in basins complicated by network structure and non-uniform substrate.

Further analysis of how tributaries and trunk streams interact will provide not only a better understanding of how rivers communicate signals of incision through their network of streams, but also shed light on the timing of sediment delivery to depositional basins. Because of the abrupt, large magnitude changes in important parameters such as slope, water discharge and sediment flux, the likelihood of process transitions at these junctions complicate the transmission of these signals between trunk and tributary. The development of hanging valleys not only impedes the upstream transmission of subsequent incision signals and delays the equilibration of sediment delivery from hillslopes, but also limits the connectivity of pathways for mobile species such as fish. The loss of this connectivity can segregate and isolate biological communities, potentially affecting species evolution [*Montgomery*, 2000].

Conclusions

We present evidence from both theoretical and numerical analyses that sediment-flux dependent incision rules predict the formation of temporary and permanent hanging valleys in fluvial networks responding to a discrete pulse of incision. We find that four factors determine the distribution of hanging valleys in fluvial networks: the magnitude of the pulse of incision, the rate of decay of the incision signal as it propagates upstream, the lag time of the sediment flux response and the drainage area of the tributary. The slopes required to create permanent or temporary hanging valleys decrease with decreasing drainage area, thus allowing upper-basin tributaries receiving an attenuated base level fall signal to become hung above the mainstem. Though the Simplified Abrasion incision rule provides mechanisms for the formation of temporary hanging tributaries, the formation of permanent hanging valleys in the saltation-abrasion incision rule provides a better fit to our field observations of both knickpoint form and process-transitions at knickpoints. The development of hanging valleys in tributaries extends the response times of landscapes well beyond those predicted by the stream power or transport-limited incision rules. An improved understanding of the erosion processes that subsequently modify the over-steepened reach will improve our predictions of landscape response time and the delivery of sediment to depositional basins.

References

Berryman, K., et al. (2000), Tectonic and paleoclimatic significance of Quaternary river terraces of the Waipaoa River, east coast, North Island, New Zealand, *New Zealand Journal of Geology and Geophysics*, *43*, 229-245.

Berryman, K., et al. (in preparation), The post-glacial downcutting history in the Waihuka tributary of the Waipaoa River, Gisborne District, New Zealand, and implications for tectonics and landscape evolution.

Bishop, P., et al. (2005), Knickpoint recession rate and catchment area: the case of uplifted rivers in Eastern Scotland, *Earth Surface Processes and Landforms*, *30*, 767-778.

Crosby, B. T., and K. X. Whipple (in press), Knickpoint Initiation and Distribution within Fluvial Networks: 236 waterfalls in the Waipaoa River, North Island, New Zealand, *Geomorphology*.

Eden, D. N., et al. (2001), Dating the culmination of river aggradation at the end of the last glaciation using distal tephra compositions, eastern North Island, New Zealand, *Geomorphology*, *38*, 133-151.

Gardner, T. W. (1983), Experimental-Study of Knickpoint and Longitudinal Profile Evolution in Cohesive, Homogeneous Material, *Geological Society of America Bulletin*, *94*, 664-672.

Gasparini, N. M., et al. (2006), Numerical modeling of non-steady-state river profile evolution using a sediment-flux-dependent incision model, in *Tectonics, Climate and Landscape Evolution: Geological Society of America Special Paper 398*, edited by S. D. Willett, et al., pp. 127-141.

Gasparini, N. M., et al. (submitted), Predictions of Sediment-Flux-Dependent River Incision Models in Steady State and Transient Landscapes, *submitted to Journal of Geophysics - Earth Surface*.

Howard, A. D. (1994), A detachment-limited model of drainage basin evolution, *Water Resources Research*, *30*, 2261-2285.

Howard, A. D., and G. Kerby (1983), Channel changes in badlands, *Geological Society of America Bulletin*, *94*, 739-752.

Montgomery, D. R. (2000), Coevolution of the Pacific salmon and Pacific Rim topography, *Geology (Boulder)*, *28*, 1107-1110.

Niemann, J. D., et al. (2001), A quantitative evaluation of Playfair's law and its use in testing long-term stream erosion models, *Earth Surface Processes and Landforms*, *26*, 1317-1332.

Paola, C., et al. (1992a), The large-scale dynamics of grain-size variation in alluvial basins, 1: Theory, *Basin Research*, *4*, 73-90.

Paola, C., et al. (1992b), Downstream Fining by Selective Deposition in a Laboratory Flume, *Science*, *258*, 1757-1760.

Parker, G. (2002), Somewhat less random notes on bedrock incision, 20 pp, Internal Memorandum 118, St. Anthony Falls Laboratory, University of Minnesota.

Rosenbloom, N. A., and R. S. Anderson (1994), Hillslope and channel evolution in a marine terraced landscape, Santa Cruz, California, *Journal of Geophysical Research, Solid Earth*, *99*, 14,013-014,029.

Sklar, L. S., and W. E. Dietrich (1998), River longitudinal profiles and bedrock incision models: Stream power and the influence of sediment supply, in *Rivers Over Rock: Fluvial Processes in Bedrock Channels*, edited by K. J. Tinkler and E. E. Wohl, pp. 237-260, American Geophysical Union, Washington, D. C.

Sklar, L. S., and W. E. Dietrich (2004), A mechanistic model for river incision into bedrock by saltating bed load, *Water Resources Research*, *40*, no.6, 21.

Sklar, L. S., et al. (submitted), Evolution of fault scarp knickpoints following 1999 Chi-Chi earthquake in West-Central Taiwan, *JGR-ES*.

Snyder, N., et al. (1999), Evidence for an equilibrium between main-trunk-channel incision and tectonic uplift: Mendocino triple junction region, northern California, *Abstracts with Programs, Geological Society of America*, *31*, 444-445.

Snyder, N. P., et al. (2002), Interactions between onshore bedrock-channel incision and nearshore wave-base erosion forced by eustasy and tectonics, *Basin Research*, *14*, 105-127.

Stock, J. D., and D. R. Montgomery (1999), Geologic constraints on bedrock river incision using the stream power law, *Journal of Geophysical Research-Solid Earth*, *104*, 4983-4993.

Tucker, G. E., and R. L. Bras (1998), Hillslope processes, drainage density, and landscape morphology, *Water Resources Research*, *34*, 2751-2764.

Tucker, G. E., et al. (2001a), The channel-hillslope integrated landscape development model (CHILD), in *Landscape Erosion and Evolution Modeling*, edited by R. S. Harmon and W. W. I. Doe, pp. 349-388, Kluwer Academic/Plenum Publishers, New York.

Tucker, G. E., et al. (2001b), An object-orineted framework for distributed hydrologic and geomorphic modeling using triangulated irregular networks, *Computers and Geosciences*, *27*, 959-973.

Tucker, G. E., and K. X. Whipple (2002), Topographic Outcomes Predicted by Stream Erosion Models: Sensitivity Analysis and Intermodel Comparison, *Journal of Geophysical Research-Solid Earth*, *107*.

van der Beek, P., and P. Bishop (2003), Cenozoic river profile development in the upper Lachlan catchment (SE Australia) as a test of quantitative fluvial incision models, *Journal of Geophysical Research*, *108*, 2309, doi:2310.1029/2002JB002125.

Weissel, J. K., and M. A. Seidl (1998), Inland Propagation of Erosional Escarpments and River Profile Evolution Across the Southeast Australian Passive Continental Margin, in *Rivers Over Rock: Fluvial Processes in Bedrock Channels*, edited by K. J. Tinkler and E. E. Wohl, pp. 189-206, American Geophysical Union, Washington, D. C.

Whipple, K. X., et al. (2000), Rates and processes of bedrock incision by the upper Ukak River since the 1912 Novarupta ash flow in the Valley of Ten Thousand Smokes, Alaska, *Geology*, *28*, 835-838.

Whipple, K. X., and G. E. Tucker (1999), Dynamics of the stream-power river incision model; implications for height limits of mountain ranges, landscape response timescales, and research needs, *Journal of Geophysical Research, Solid Earth*, *104*, 17,661-617,674.

Whipple, K. X., and G. E. Tucker (2002), Implications of sediment-flux-dependent river incision models for landscape evolution, *Journal of Geophysical Research, Solid Earth*, *107*, 3-1 - 3-20.

Willgoose, G., et al. (1991), A coupled channel network growth and hillslope evolution model. 1. Theory, *Water Resources Research*, *27*, 1671-1684.

Wobus, C. W., et al. (in press), Hanging valleys in fluvial systems: Controls on occurrence and implications for landscape evolution, *Journal of Geophysical Research – Earth Surface*.

Figure Captions

Figure 1: Landscapes responding to a discrete period of incision often possess hanging tributaries or subcatchments isolated from the mainstem by a large step in channel elevation. In this topographic rendering from a portion from the Waipaoa River on the North Island of New Zealand (1a), channels draining the outlined tributary basins are elevated above and segregated from the trunk streams by steep knickpoints (1b). In order to understand how these features develop during a transient pulse of incision, we utilize analytical and numerical methods to model channel (1c) and landscape (1d) responses to discrete base level fall events. (1a and 1b modified from Crosby and Whipple, in press)

Figure 2: Analytical solutions for four steady-state channel incision rules demonstrate variations in the dependence of incision rate on channel slope at a fixed drainage area (1 x 10^6 m^2). Note that all calculations were made without applying the small angle approximation. Instead of plotting the data relative to sin(α), we plot it relative to a more intuitive variable, slope, where slope = tan(arcsin(sin(α))). Though model parameters used in this figure differ from those used in the numerical modeling, the relationships between slope and incision rate demonstrated here are directly comparable. In both the Transport-Limited and Stream-Power incision rule, incision rate is linearly dependent on channel slope up to a value of ~0.4. At higher slopes, incision rate continues to increase with slope, but less rapidly. In the saltation-abrasion model, incision rate initially increases with slope, exceeding the background uplift rate at S_{ss}. After reaching a maximum at S_{peak}, the incision rate monotonically decreases with increasing slope, eventually dropping below the background uplift rate (gray horizontal line) at S_{hang}. The incision rate in the Simplified Abrasion model initially increases rapidly with slope, but levels off as it asymptotically approaches the maximum incision rate at I_{max}.

Figure 3: In the saltation-abrasion (S-A) model, the dependence of incision rate on slope varies with drainage area. Note that all calculations were made without applying the small angle approximation. Instead of plotting the data relative to sin(α), we plot it relative to a more intuitive variable, slope, where slope = tan(arcsin(sin(α))). All subsequent mentions of slope are calculated using the preceding equation. The slope, S_{hang}, at which incision rate falls below the background uplift rate and permanent hanging valleys form, decreases with decreasing drainage area. This reveals that small tributaries have a greater probability of creating hanging valleys than large ones. Though difficult to perceive in this figure, the slope at which the S-A model achieves peak incision rates, S_{peak}, increases with decreasing drainage area. Note that for drainage basins with the drainage areas 5 x 10^6 m^2, the value for S_{hang} is so large that does not plot on these axes.

Figure 4: Numerical simulations of mainstem profile evolution and normalized incision rates observed following base level fall. Plots 4a-4d reflect simulations governed by the transport-limited (T-L) incision rule while plot 4e-4h reflect simulations governed by the stream-power (S-P) incision rule. In plots 4e-4h,

thin dashed lines demonstrate the S-P rule's transient evolution independent of numerical diffusion. In plots 4b, 4d, 4f and 4h, the thick grey lines reveal the behavior of the channel during the progressive base level fall. Note that the incision signal in the S-P case migrates up the mainstem as a form of discrete adjustment while in the T-L case the increase in high sediment delivery during the transient prevents the downstream reaches from re-equilibrating. Jagged kinks in plots 4b and 4d are a consequence of deposition-triggered numerical instability.

Figure 5: Numerical simulations of mainstem profile evolution and normalized incision rates observed following base level fall. Plots 5a-5d reflect simulations governed by the simplified saltation-abrasion (S-A) incision rule while plots 5e-5h reflect simulations governed by a Simplified Abrasion (A-D) incision rule. In plots 5a and 5c, instantaneous base level fall creates high outlet slopes (between S_{peak} and S_{hang}) in the S-A model and retards the upstream transmission of the full incision signal. In plots 5b and 5d, progressive base level fall in the S-A model creates an over-steepened channel (with slopes between S_{ss} and S_{peak}) that rapidly readjusts toward its pre-disturbance form. Plots e and g demonstrate the rapid decay (at I_{max}) of the step created by instantaneous base level fall in the A-D model. In plots 5f and 5h, the response following the disturbance is fast, but maximum incision rates are limited by the maximum slopes created as a consequence of the progressive base level fall. Though the signal propagates rapidly upstream and establishes a pseudo-equilibrium incision rate in figures 5d, 5g and 5h, the channel does not return to pre-disturbance slopes or incision rates until the hillslopes and upper portions of the network deliver sediment at the equilibrium rate.

Figure 6: Numerical simulations of tributary profile evolution observed following base level fall. In plots 6a and 6b, tributaries in the transport-limited model adjust in concert with the incision in the mainstem, never developing hanging valleys. Plots 6c and 6d demonstrate that in the stream power model, during passage of the incision signal from trunk to tributary, there is no growth in tributary elevation as a consequence of hanging valley formation. Plot 6e demonstrates that for instantaneous base level fall in the saltation-abrasion model, the over-steepened reach creates a locally low incision rate at the outlet that buffers the rapid upstream communication of the base level fall signal. Tributaries consequently experience slow mainstem incision rates and never create hanging valleys. Plots 6f, 6g and 6h demonstrates the formation of temporary hanging tributary valleys as the pulse of incision migrates up the mainstem.

Figure 7: The formation of permanent hanging valleys in the saltation-abrasion model depends on sufficiently fast mainstem incision rates over-steepening tributary junction to slopes beyond S_{hang}. In these channel profiles from the CHILD model, we display the formation and persistence of permanent hanging tributaries elevated above the mainstem. The upstream lips of the knickpoints are eroding slower than the background uplift rate and the elevations of tributary channels increase indefinitely.

Table 1

Parameters from each of the 4 CHILD model runs.

Table 1

Model Run	K_t	K_{model}	m	n	K_D
Transport-Limited	2.00E-05	-	-	-	0.3
D-L Stream Power	4.00E-05	2.00E-05	0.5	1	0.025
Saltation-Abrasion	2.00E-05	5.00E-02	-0.25	-0.5	1.5
Simplified Abrasion	2.00E-05	5.00E-03	0	0	1

Parameters Identical in All Models	
Basin Dimensions [m]	5000 x 5000
Basin Area [m^2]	2.5 x 10^7
Node Spacing [m]	100
Background Uplift Rate, U, [m/yr]	1.00E-03
m_t	1.5
n_t	1
k_w	1
b	0.5
k_q	1
c	1
β	1
Instantaneous Base Level Fall [m] (discrete elevation change basin outlet)	100
Prolonged Base Level Fall [m] (increase uplift rate 10x and run model until 100m new of material is exhumed/uplifted.)	100

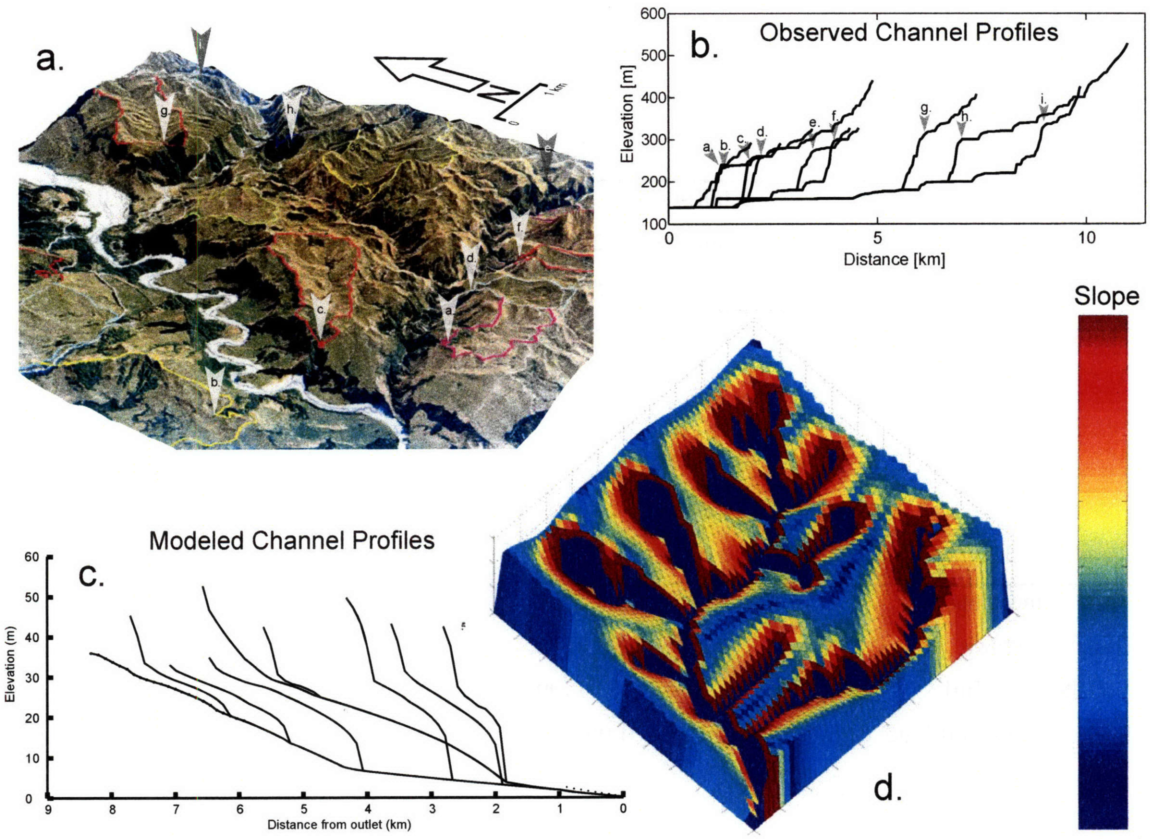
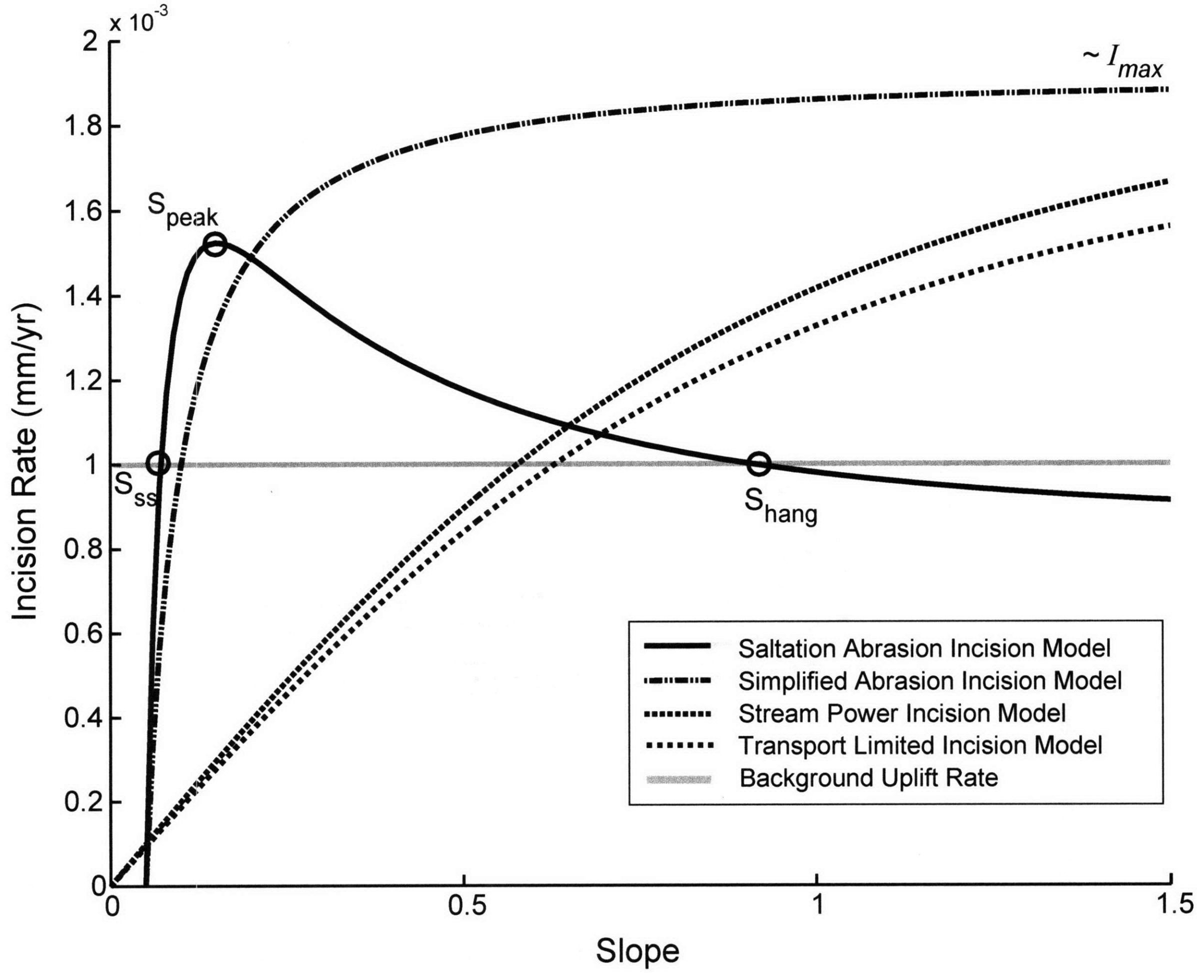

a.
b.
Observed Channel Profiles
Elevation [m]
Distance [km]
c.
Modeled Channel Profiles
Elevation (m)
Distance from outlet (km)
d.
Slope

Figure 1



x 10^-3
2
1.8
1.6
1.4
1.2
1
0.8
0.6
0.4
0.2
0
Incision Rate (mm/yr)
0
0.5
1
1.5
Slope
~ Imax
Speak
Sss
Shang
Saltation Abrasion Incision Model
Simplified Abrasion Incision Model
Stream Power Incision Model
Transport Limited Incision Model
Background Uplift Rate

Figure 2

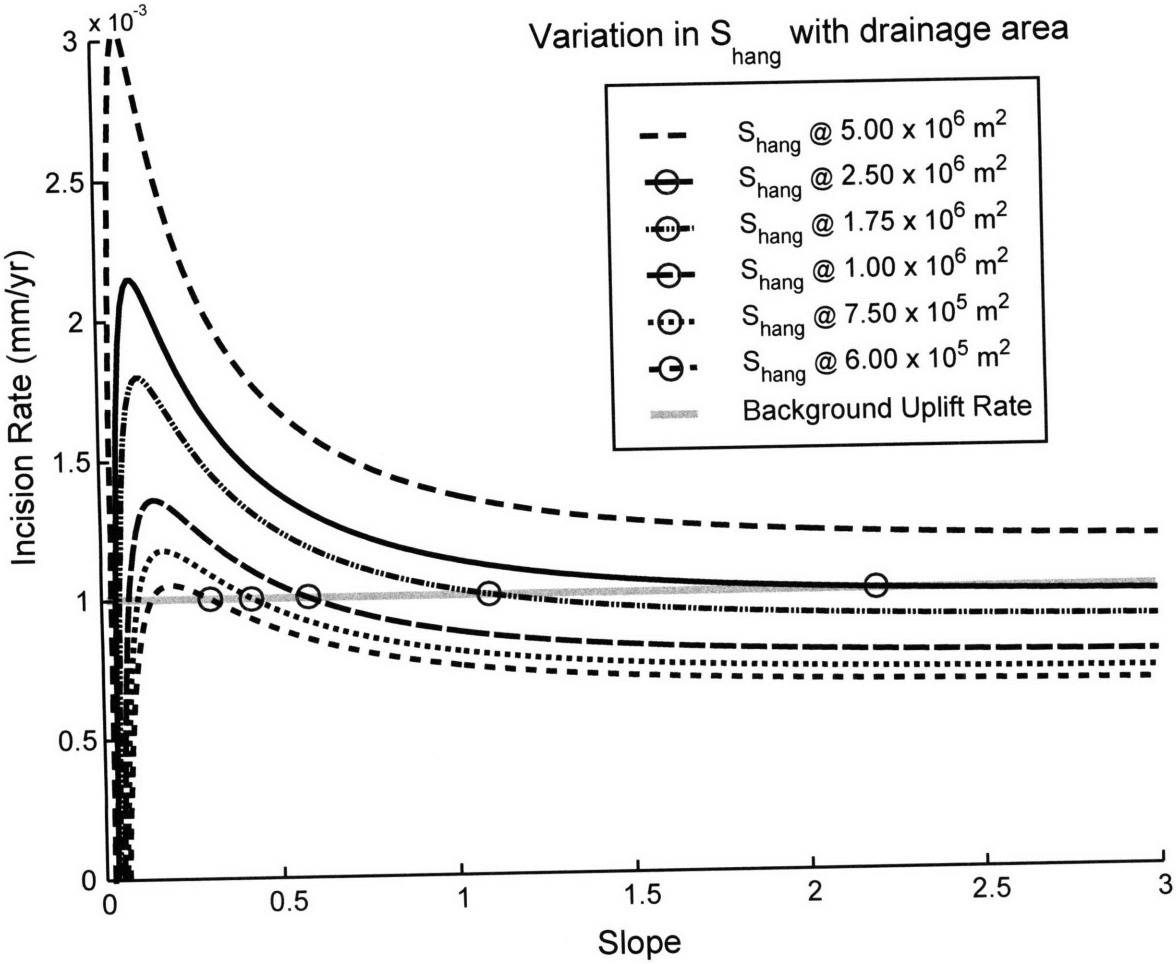

Variation in S hang with drainage area
S hang @ 5.00 x 10^6 m^2
S hang @ 2.50 x 10^6 m^2
S hang @ 1.75 x 10^6 m^2
S hang @ 1.00 x 10^6 m^2
S hang @ 7.50 x 10^5 m^2
S hang @ 6.00 x 10^5 m^2
Background Uplift Rate
x 10^-3
Incision Rate (mm/yr)
3
2.5
2
1.5
1
0.5
0
0
0.5
1
1.5
2
2.5
3
Slope

Figure 3

Figure 4

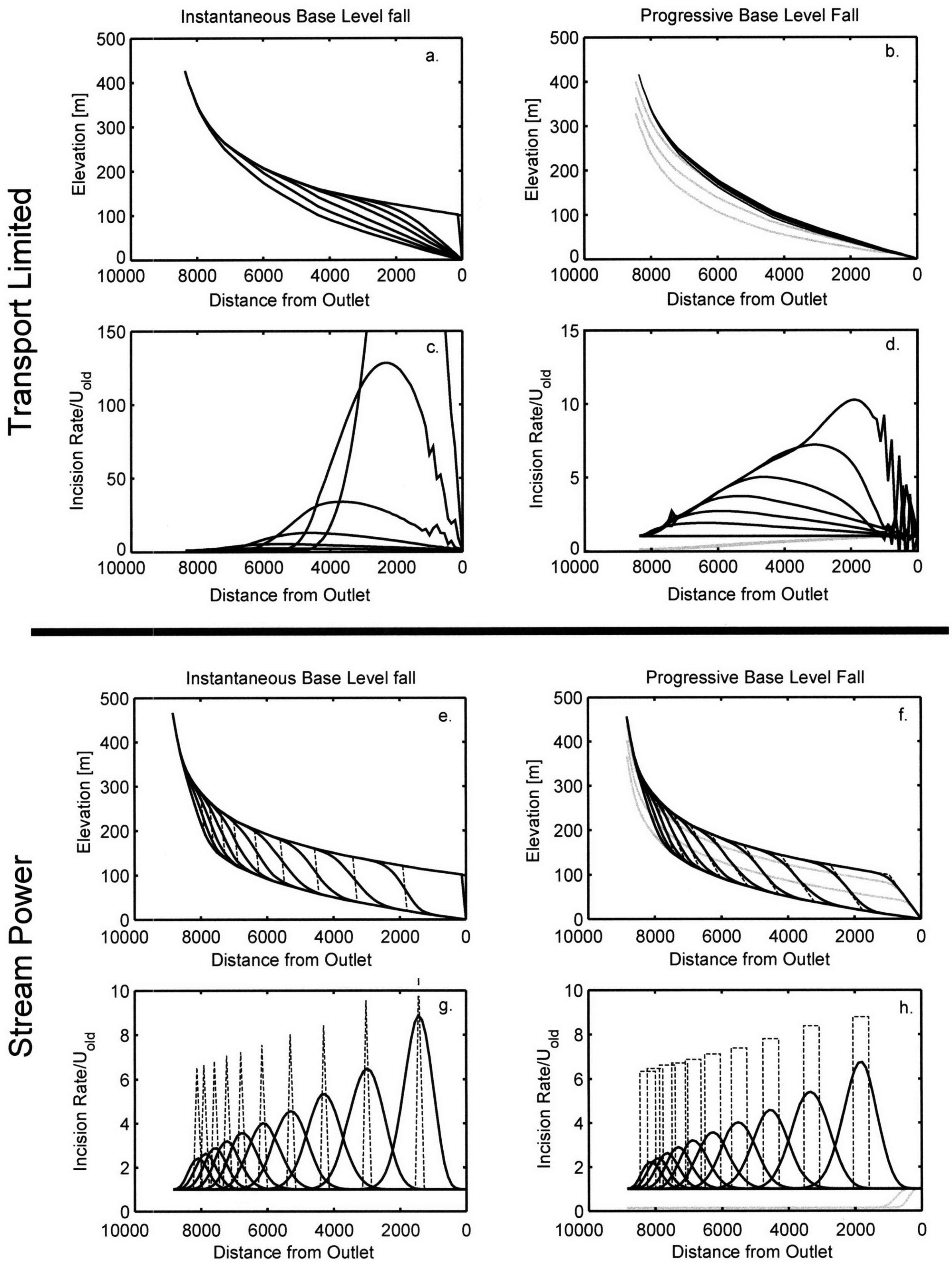

Transport Limited
Stream Power
Instantaneous Base Level fall
Progressive Base Level Fall
a.
b.
c.
d.
e.
f.
g.
h.
Elevation [m]
Incision Rate/U_old
Distance from Outlet
500
400
300
200
100
0
150
50
15
10
5
8
6
4
2
10000
8000
6000
4000
2000

Figure 5

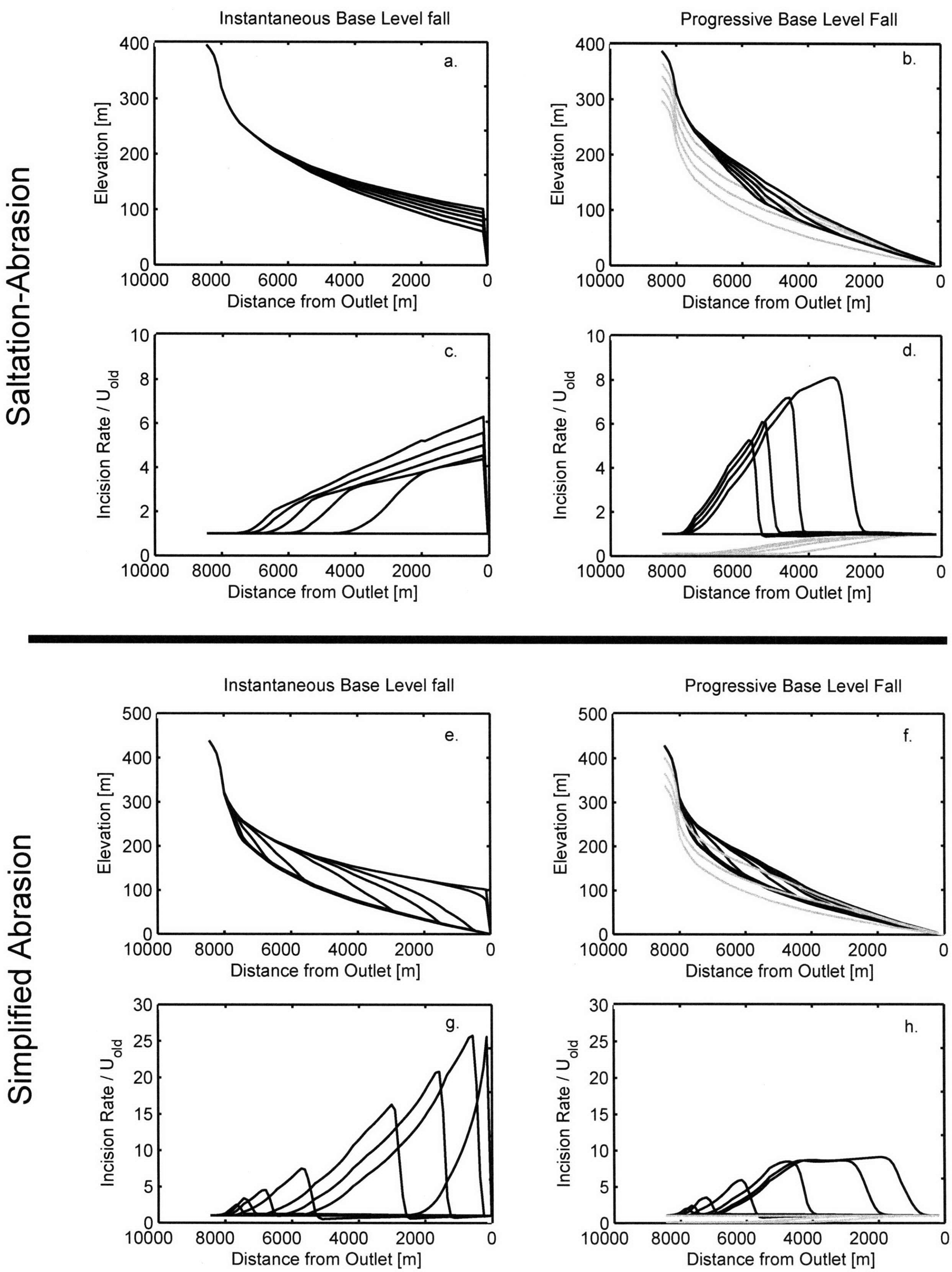
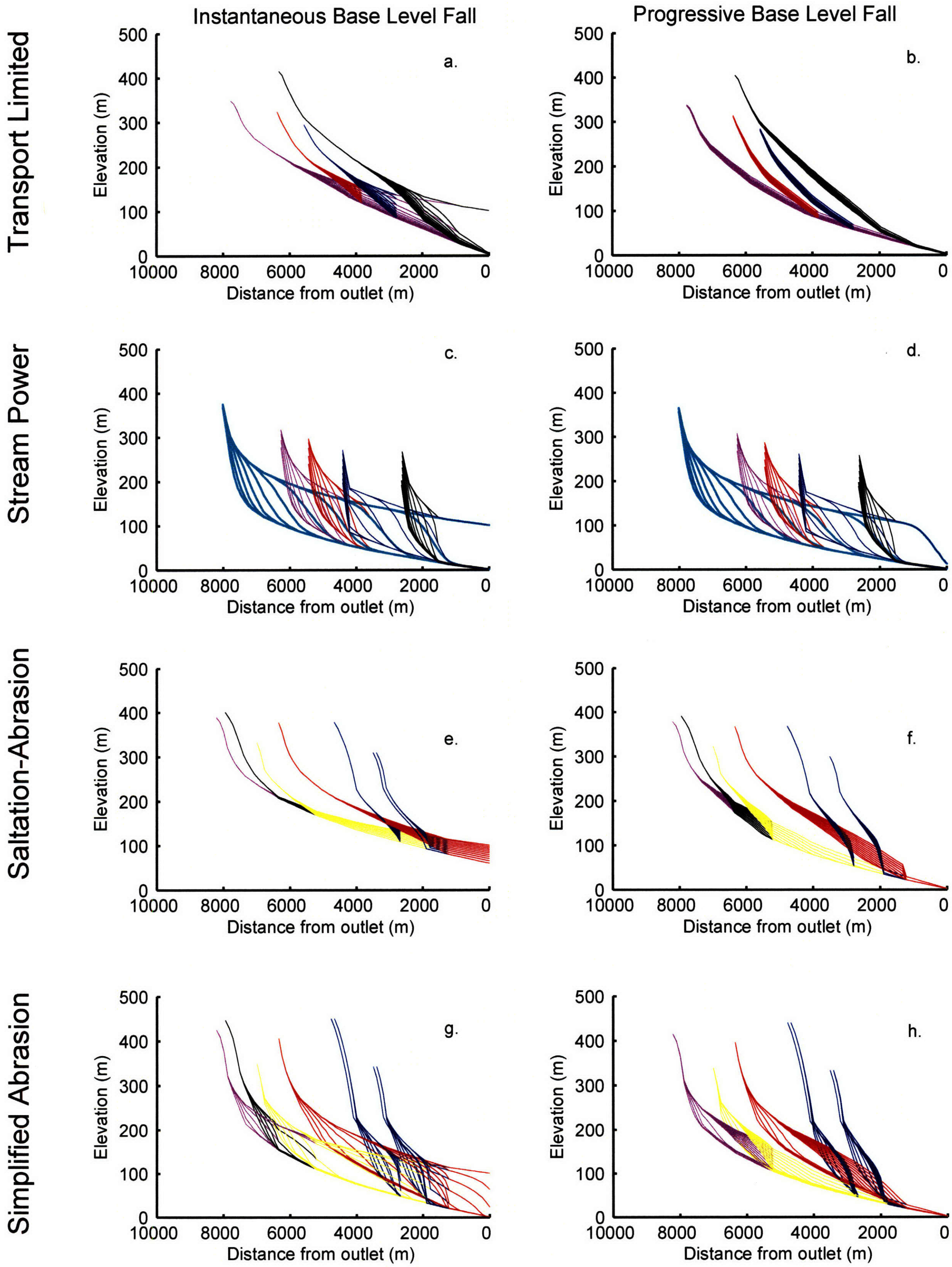

Saltation-Abrasion
Instantaneous Base Level fall
Progressive Base Level Fall
a.
b.
c.
d.
Elevation [m]
Incision Rate / U_{old}
Distance from Outlet [m]
400
300
200
100
0
10
8
6
4
2
10000
8000
6000
4000
2000
Simplified Abrasion
Instantaneous Base Level fall
Progressive Base Level Fall
e.
f.
g.
h.
500
30
25
20
15
5

Figure 6



Instantaneous Base Level Fall
Progressive Base Level Fall
Transport Limited
Stream Power
Saltation-Abrasion
Simplified Abrasion
a.
b.
c.
d.
e.
f.
g.
h.
Elevation (m)
Distance from outlet (m)
500
400
300
200
100
0
10000
8000
6000
4000
2000
0

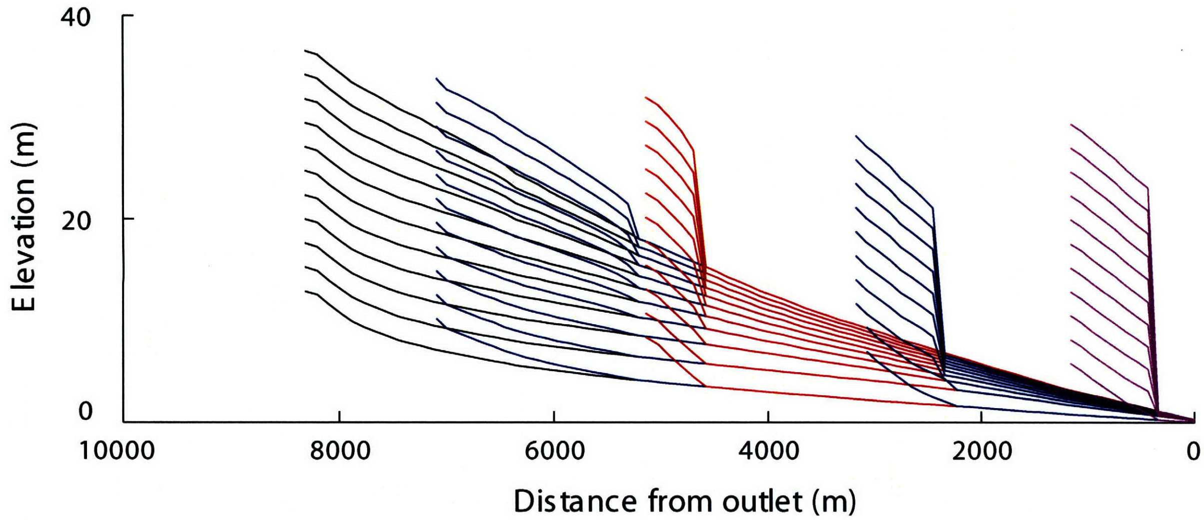

40
20
0
Elevation (m)
10000
8000
6000
4000
2000
0
Distance from outlet (m)

Figure 7

Changes in Channel Morphology and Bed State in Response to River Incision – Deterministic Trends or Inherent Variability?

Benjamin T. Crosby[1,2]
Kelin X Whipple[1,3]

1 Department of Earth, Atmospheric and Planetary Sciences, Massachusetts Institute of Technology, Cambridge, Massachusetts 02139
2 Now at: Department of Geosciences, Idaho State University, Pocatello, ID
3 Now at: School of Earth and Space Exploration, Arizona State University, Tempe, AZ
Primary Contact: Benjamin Crosby at crosby@alum.mit.edu

Abstract

We examine how channel gradient, bankfull width, mean grainsize and percent bed cover respond following a discrete period of channel incision. Detailed field surveys were collected across prominent knickpoints in 27 tributary and trunk streams of the Waipaoa River catchment on the North Island of New Zealand. These knickpoints mark the upstream extent of a 50 to 100 meter magnitude incision signal that initiated ~18,000 years ago. Prominent knickpoints related to this period of incision were located in 236 different channels in the Waipaoa River catchment. The 27 channels we selected for detailed investigation represent a broad range of drainage areas, knickpoint forms and positions within the basin. Most are situated within the same lithologic unit, the early-Miocene Tolaga Group, a clay-rich mudstone highly susceptible to decomposition by shrink-swell weathering processes. Our aggregate analysis of all the survey data reveal that in reaches upstream of the knickpoints, gradients typically increase as the knickpoint is approached while bankfull widths remain relatively stable. Mean grainsize appears to steadily increase within ~300 m upstream of the knickpoint and persists for another ~400 m downstream of the knickpoint. Though the mean percent bed exposure erratically fluctuates between ~40 and ~80% bed exposure over the surveyed reach, there is a well defined increase and decrease in bed exposure between ~150m upstream and ~150 m downstream of the knickpoint lip. Downstream of the knickpoint, an initial pronounced increase in channel gradient decays to background values with increasing distance downstream of the knickpoint. Bankfull widths in the same reach increase abruptly at the knickpoint and remain high in the downstream direction. Interestingly, in these reaches affected by the passage of a transient pulse of incision, Slope-Area and Width-Area scaling are not only both well defined, but regressions through data upstream or downstream of the knickpoint are not statistically separable. This suggests that though the knickpoint defines the upstream limit of profile adjustment to the pulse of incision, channel slope and width are well adjusted to the current discharge and sediment flux conditions. As well, it suggests that gradient and width can adjust fast enough to keep pace with the upstream migration of the knickpoint. Future work must to explore the sources of variability in knickpoint form and address why knickpoints do not behave in a deterministic manner.

1. Introduction

During the transient response to a tectonic, climatic or base-level disturbance, river channels adjust their bed state and morphologic parameters such as elevation, slope or width until the channels reestablish equilibrium with the new conditions. For disturbances that initiate incision, the headward extent of this upstream propagating fluvial adjustment is often described as a knickpoint. Though many interpretations of the term exist, we define a knickpoint (Figure 1a) as the boundary in the fluvial system that separates the upstream, relict topography (adjusted to the previous condition), from the downstream reach that has already begun to adjust to the new conditions (Crosby and Whipple, in press). The evolution of the form of the knickpoint as it propagates upstream can be considered as a consequence of a competition between three different erosion rates (Figure 1a): the incision rate at the top of the knickpoint, the retreat rate at the knickpoint face and the incision rate at the base of the knickpoint. Variations in the relative magnitudes of these three rates determines how the form of the knickpoint evolves as the signal propagates upstream (Figures 1b-1g). At present, it is unclear whether knickpoints retain form as they propagate through fluvial systems (Seidl and Dietrich, 1992; Whipple and Tucker, 1999) or whether they are temporary features that may initiate locally but are not maintained as they retreat upstream (Crosby and Whipple, in press).

Some field settings provide evidence for parallel retreat (Figure 1b), where the knickpoint retains its form as it propagate upstream, lowering the channel step-wise from the initial to final condition (Gilbert, 1896; Haviv et al., in press; Seidl and Dietrich, 1992). In this scenario, the distribution of knickpoints within the fluvial network at any point in time simply provides a snapshot of the continuous upstream propagating wave of adjustment. The only field setting where the parallel retreat mechanism is unequivocally demonstrated is where an erosion resistant cap-rock in a horizontally stratified package of units overlies softer, more erodable materials (Berlin and Anderson, submitted; Gilbert, 1896; Haviv et al., in press). In field settings without cap-rocks, others suggest that steep negative steps in large drainage area channels are inherently unstable, and therefore knickpoints only develop in response to local structural or lithologic inhomogeneities (Gardner, 1983; Miller, 1991). However, even in non-stratified, relatively homogeneous substrates, discrete knickpoints are frequently located in tributaries just upstream of the confluences with the trunk stream (Crosby et al., submitted; Snyder et al., 1999; Wobus et al., 2006a). Knickpoints at tributary junctions do not require that discrete knickpoints ever existed in the mainstem. It is plausible that because of the disparity between the tributary's and trunk's capacity to keep pace with mainstem incision, that tributaries may over-steepen at their junction beyond a threshold slope where fluvial processes are no longer effective (Crosby et al., submitted; Wobus et al., in press). The frequency of knickpoint observations at tributary junctions may also follow as a consequence of the theoretical expectation that knickpoint retreat rates are drainage area dependent (Bishop et al., 2005; Crosby and Whipple, in press; Rosenbloom and Anderson, 1994; Stock and Montgomery, 1999; Weissel and Seidl, 1998; Whipple and Tucker, 1999) and that knickpoints could simply spend much of their lifetime at or near tributary junctions, where there is an immediate decrease in drainage area. If that were the case, one would expect to find evidence for continued fluvially dominated incision processes near tributary knickpoints.

In order to distinguish between these explanations for knickpoint evolution and try to constrain the processes active at and near steep steps in river channels, others have undertaken a diverse suite of laboratory experiments. Independent of whether a cohesive (Gardner, 1983; Holland and Pickup, 1976) or non-cohesive (Brush and Wolman, 1960) substrate was used, flume studies of channel response to instantaneous base level fall do not to produce knickpoints that retain their form as they propagate upstream. In most cases, the discrete step decreases in slope as it propagates upstream (e.g. Figure 1e) until it can not be distinguished within the channel. The only laboratory scenario that partially retains a significant portion of the discrete knickpoint as the adjustment signal propagated upstream (e.g. Figure 1f) was for the special case of a horizontally stratified, erosion resistant cap-rock over a less cohesive substrate (Holland and Pickup, 1976). In other experiments that study the behavior of fluvial networks (Hasbargen and Paola, 2000; Schumm et al., 1987) or a suite of divergent channels, knickpoints were observed to propagate upstream but it is unclear whether the observations of these small scale features in noncohesive sediment can be appropriately scaled-up to explain the processes active at knickpoints observed in the field (Hancock et al., 1998; Montgomery, 2004; Sklar and Dietrich, 2004; Stock et al., 2005; Whipple et al., 2000). We recognize that field studies are limited by slow rates of knickpoint evolution, substrate variability and unsteady forcing. As well, we recognize that experimental observations are limited by the difficulty of scaling both substrate rheology and erosion processes up to large magnitude features observed in the field. Though our understanding of the incision processes in bedrock river channels is increasing (Weissel and Seidl, 1997 and references therein), we recognize that the extreme slopes and the limited interaction of sediment or water with the bed in knickpoint reaches will limit the applicability of many of these existing relations. Instead, understanding erosion processes near knickpoints will require consideration of non-fluvial processes such as those that depend on mass wasting, physical and chemical weathering or rock mass stability (Crosby and Whipple, in press). In this work, we examine variations in channel morphometric parameters upstream of, immediately at and downstream of knickpoints in order to provide new insight into knickpoint erosion processes and ultimately to shed light onto what determines the temporal evolution of knickpoint form. In this project, we surveyed 27 channels from the Waipaoa River catchment that contain knickpoints (Figure 2). After introducing the field area and our survey methods, we then present the measured variations in channel gradient, bankfull width, mean sediment grainsize, and percent bed exposure adjacent to knickpoints. Following the presentation of this data, we discuss how of the observed variations in morphometric parameters relate to knickpoint incision processes and knickpoints' temporal evolution.

Individual channels were selected in order to represent a broad range of drainage areas in a number of different sub-catchments. Care was used to assure that all knickpoints were a consequence of the same incision event. All knickpoints surveyed are within relatively uniform lithologic units and none are retained on particularly erosion-resistant contacts or ribs of rock. No plunge pools, sapping features or cap-rocks are observed to dominate incision processes in the field area. In this portion of the paper, we first introduce the Waipaoa River field area and the pulse of incision that initiated the observed knickpoints and then present the survey methods used to collect the data

2. Field Area

We focus our analysis on the Waipaoa River catchment because the channel network contains an extensive suite of knickpoints (Figure 2), formed as a consequence of an incomplete transient response to a large magnitude episode of rapid incision initiated ~18 ka (Eden et al., 2001; Litchfield and Berryman, 2005).

2.1 Geologic Setting

The Waipaoa River basin drains 2150 km^2 of the east flank of the North Island of New Zealand (Figure 2). It is etched into Cretaceous and Cenozoic forearc sediments accreted above the west dipping Hikurangi subduction margin. Arc volcanism inboard of this subduction margin produced the extensive tephra deposits preserved on inactive fluvial terraces and hillslopes through much of the Waipaoa River basin. These deposits constrain the timing of the initiation of rapid incision (Mazengarb et al., 1991), which consequently resulted in the development of the knickpoints we address in this study.

Latest Cretaceous to Pliocene shelf and slope sediments were structurally deformed during the late-Miocene to recent accretion process (Black, 1980; Mazengarb and Speden, 2000). Though there are blocks of coherent and consistent stratigraphy, extensive faults dissect the basin into different structural domains, each with a relatively characteristic dip direction. Horizontal units are rare and thus there were no cap-rock knickpoints observed. The bedrock is dominated by clay-rich siltstone and mudstone units, interbedded with sandstone and limestone lenses (Brown, 1995; Ota et al., 1992). Due to different structural histories (e.g. sheared argillite vs. undeformed sandstone) and different chemical compositions (e.g. clay-rich mudstones vs. calcite cemented sandstones) the bedrock units within the Waipaoa drainage basin exhibit significant variability in strength. Most channels surveyed for this analysis are within the early-Miocene Tolaga group, composed of relatively uniform clay-rich siltstones. This unit's high clay content makes it particularly sensitive to decomposition by slaking or shrink-swell weathering processes. (Taylor and Smith, 1986)

Published estimates of rock uplift rates in the northeastern corner of the North Island of New Zealand range between 0.5 and 4 mm/yr depending on whether the record comes from uplifted shorelines (Berryman et al., 2000; Litchfield and Berryman, 2006), river terrace sequences (Yoshikawa, 1988) or summit accordance along the crest of the Raukumara Range (Berryman et al., 2000; Litchfield and Berryman, 2006). Incision rates derived from strath terrace sequences along the Waipaoa River and its tributaries suggest rock uplift ranges between 0.5 and 1.1 mm/yr (Berryman et al., 2000; Berryman et al., in preparation; Eden et al., 2001) and are likely to be the most reliable for the study area.

2.2 Initiation of Incision

At present, the Waipaoa River drainage basin is adjusting to a large magnitude (up to 120 meters) pulse of incision initiated ~18,000 years ago (Litchfield and Berryman, 2005). This adjustment is manifest in numerous knickpoints, incised gorges, unstable hillslopes and strath terraces. These transient landforms help distinguish the mobile boundary between the portion of the landscape that is adjusting to the new base level from regions that retain their relict, pre-disturbance form (Figure 2).

The broad valleys and extensive terraces with low downstream gradients observed within the relict portions of the landscape result from a period of extensive aggradation during the last glacial maximum (LGM), ~30-18 ka (Chappell, 2001; Shackelton, 1987). Toward the end of that period, around the time of the initiation of incision, sea level was at its absolute lowest (Berryman et al., 2000; Litchfield and Berryman, 2005). The aggradational surface that we observed today in the relict portion of the landscape was graded to that LGM base level, some ~130m below present sea level (Brown, 1995). During sea level rise, the relict surface and any associated knickpoints in the southern third of the Waipaoa River basin were buried by sediments consequent to the shore line transgression (McGlone, 2001, 2002; McLea, 1990; Newnham et al., 1999).

Though not cold enough to produce glaciers, the cooler temperatures during the LGM had a profound effect on the local landscape; vegetation was dominated by scrub brush and grasses, sediment production increased and precipitation decreased (Litchfield and Berryman, 2005). In the Waipaoa River catchment, thick fill deposits on strath terraces and inactive alluvial fans near tributary junctions demonstrate the extent of this aggradation. Observed fill thicknesses vary widely within the drainage basin, ranging from 5 to over 45 meters. These sediment-mantled conditions during the LGM were not restricted to the Waipaoa River drainage basin but are documented along the entire East Coast of New Zealand (2000).

Multiple hypotheses have been presented to explain the initiation of incision at the end of the LGM. Berryman et al. (2005) and Litchfield and Berryman (McGlone, 2001, 2002; McLea, 1990) suggest that incision into the aggraded catchment occurred as a consequence of two related factors. First, warmer and wetter climate established a more cohesive conifer forest cover (Chappell, 2001; Shackelton, 1987), providing greater soil cohesion and reducing the sediment supplied from the hillslopes and physical weathering processes. Second, increased precipitation increased stream discharge, raising transport capacity and the channels' erosive potential. With these changes, channels were capable of incising into and transporting out the previously aggraded sediment, eventually cutting further into the underlying bedrock, and creating the incised landscape we observe today.

We suggest a complementary hypothesis for enhancing the magnitude the incision. This hypothesis takes advantage of the fact that during the LGM, the sea level lowstand was > 100 m below the present mean sea level (MSL) (Foster and Carter, 1997; Orpin, 2004). Using present-day bathymetric data (Crosby and Whipple, in press; e.g. Snyder et al., 2002), we find this depth to be just above the shelf-break. If sea level dropped over this abrupt break in slope, the accelerated erosion over the steep reach could have initiated a wave of incision could have propagated upstream into the channel network, thus enhancing the climatically-driven incision signal (2005). Litchfield and Berryman (Foster and Carter, 1997) note that LGM fluvial deposits observed seismically on the continental shelf (Kuehl et al., 2003) suggest that incision related to sea level fall would have been limited to the middle or outer shelf. Whether there are fluvial channels incised into these deposits that could have transmitted a signal of incision upstream remains a subject of active research in the NSF-MARGINS program (Litchfield and Berryman, 2006).

A third hypothesis suggests that the initial mobilization of the aggraded sediment was a consequence of climate change as presented in the first hypothesis, but the large magnitude of bedrock incision specifically observed in the Waipaoa River and a limited number of other catchments along the east coast was a local consequence of accelerated local rock uplift within the accretionary wedge due to deep-seated subduction processes (Collot et al., 2001; Eberhart-Phillips and Chadwick, 2002; Lewis and Pettinga, 1993). This assertion builds off the observation that subducting seamounts destabilize the trench slope and perturb the force balance within the accretionary wedge, suggesting that oceanic plate topography and rheology strongly influence the time evolution of wedge deformation (Litchfield et al., submitted). Lateral variation in uplift rate (manifest as fluvial incision rates) along the East Coast of New Zealand have been attributed local seamount subduction and/or sediment underplating within the accretionary wedge (Crosby and Whipple, in press).

2.3 Present Catchment Morphology

The present morphology of the Waipaoa River catchment reflects the partial adjustment of the hillslopes and river network to a large magnitude pulse of incision. The well exposed and laterally extensive relict remnants of the ~18 ka aggradational surface provide an excellent reference from which to measure the progress of the transient response. In previous work, we recognized that >50% of the aerial extent of the Waipaoa River catchment is within the relict region (Figure 2) and that most knickpoints are located in low drainage area tributaries, just upstream of their junction with the mainstem, and at roughly the same elevation as the local relict surface (Crosby and Whipple, in preparation). As discussed in Section 2.1.2, the relict surface is only observed in the northern two-thirds of the catchment. In that region, in the reaches downstream of the knickpoints, the total incision below the top of the aggradational surface ranges between 30 and 120 meters. This measurement is the sum of the height of the bedrock strath above present river level and the thickness of the overlying aggradational fill deposit. In some trunk streams of the Waipaoa River, the incision is entirely within extremely thick aggradational deposits while in other trunk stream reaches, the aggradational deposits are less than 5 meters thick and the incision is almost entirely through bedrock. This variability in observed fill thickness is a consequence of both local sediment sources such as alluvial fan deposits at tributary mouths and a consequence of whether or not the post-aggradational incision initiated while the channel was above the buried pre-aggradation paleochannel (Berryman et al., in preparation; Crosby and Whipple, in preparation).

Within the trunk streams of the Waipaoa River, flights of strath terraces and meander cut-offs provide evidence, at least within in some reaches, that the incision in large drainage area channels was not instantaneous. Whether the incision was gradual over a protracted period of time or was accomplished by episodic, small magnitude incision events, the terrace record refutes the possibility that the observed single-step knickpoints found in tributaries of the Waipaoa were formed by a the passage of a similar knickpoint up the trunk streams. In the Waihuka Stream, this is particularly apparent (DeRose et al., 1998; Marden et al., 2005). In most low drainage area tributaries that contain knickpoints, there is no evidence for progressive or episodic incision. This lack of evidence may be a consequence of lack or preservation of these features or could suggest that in these reaches that base level fall was instantaneous, as would be anticipated for incision driven by a headward propagating knickpoint.

All trunk streams in the Waipaoa River have already begun responding to the large magnitude pulse of incision. At drainage areas less than ~ 5×10^7 m^2, there is partial or scarce sediment cover observed on the bed of trunk stream channels. At drainage areas greater than ~ 5×10^7 m^2, trunk streams are low gradient (between 0.003 and 0.01) and have the form of typical gravel-bedded alluvial channels (see section 2.1.4). Tributaries to the trunk streams typically have drainage areas less than ~ 5×10^6 m^2. Knickpoints are typically found in these tributaries and can be loosely described as ~10-50 meter drops composed of one or two near-vertical negative steps in the channel profile. Though the total incision below the W1 surface can reach ~120 m, the fluvial relief in the knickpoint-reach is accounted for in both the steep step and the steep fluvial reach downstream of the knickpoint. As the characteristics of the knickpoint and the reaches up and downstream of this landform are the focus of this study, we reserve further discussion until Section 3.

2.4 Anthropogenic Impacts

Deforestation in the Waipaoa River was initiated following Maori settlement ~700 years ago (Jones, 1988) and was completed following European settlement beginning in ~1820 (Pullar, 1962). Livestock grazing also was initiated at the time of European settlement and large scale instabilities in hillslopes developed almost immediately thereafter. These instabilities resulted in large increases in sediment delivery to the mainstem channels, inducing rapid aggradation. In some reaches of the Waipaoa mainstem, there are anecdotal records of over 10 meters of aggradation. Recent implementation of commercial forestry practices and other tree planting campaigns have helped slow sediment delivery and the headward extension of gullies, but some large features remain active despite efforts to stabilize those hillslopes (Howard and Kerby, 1983; Whipple and Tucker, 1999; Wobus et al., 2006b). Though our surveys are outside of the most impacted regions and upstream of the influence of the aggraded mainstem, we recognize that observations of sediment grainsize and percent bed cover may be substantially influenced by historic and current land use practices.

3. Methods

Surveys of channels that contain knickpoints were collected over 5 field seasons (between January 2002 and February 2005) in the Waipaoa River catchment. To facilitate efficient access to all portions of the basin and to assure a safe and dry office for the electronic survey equipment, a central camp was established at Te Hau Station just east of Whatatutu village, near the confluence of the Waingaromia Stream and the Waipaoa River.

3.1 Survey Tools

Field reconnaissance to locate suitable regions for survey was accomplished using 1:50,000 topographic map sheets and ~1:25,000 color stereo aerial photographs flown by Air-Maps New Zealand between 1997 and 1999. We also used a digital elevation model to examine channel profiles but found that the step-like elevation artifacts at contour crossings limited the usability of channel profile data. Channel geometry data was collected in the field using two techniques. In some channels, an inclinometer and laser rangefinder (or measuring tape) was the most efficient method. In other channels we utilized an in-house assembled digital surveying tool. This tool was composed of a laser range finder (with integrated compass and inclinometer) connected to a GPS-enabled handheld computer running ArcPad GIS software. Within ArcPad, we ran a custom channel survey utility which provided a series of forms for data entry and plotted the result in real-time on digital base maps.

3.2 Survey Techniques

Along-channel survey station locations were selected in an effort to provide the longest shot along a straight, consistent reach of channel. Stations were also placed at notable features such as tributary junctions, debris flow dams or at abrupt changes in any of the measured channel parameters. The methods for collecting the various data are provided below. Channel longitudinal profile data were collected at every station but the parameters describing the channel condition were usually collected at every other or every third station depending on the distances between stations. Channel profile data (e.g. Figure 3a and 3c) were collected relative to the water-line elevation rather than the thalweg elevation. Channel drainage areas were extracted from the 20 m posting DEM (e.g. Figure 3b). Extreme care was taken to recognize and correct for any discrepancies between in the observed and calculated stream network. Bankfull width was measured as the distance between the high-flow marks on either bank, without going onto a floodplain (e.g. Figure 3d). Bankfull width was well constrained in most cases in these deeply incised channels. Valley width was measured in the valley bottoms as the distance between the points where the channel walls rise abruptly up from the valley bottom. Mean and maximum grainsize were determined by visual estimation (e.g. Figure 3e). We were aware of the potential bias for larger grainsizes using visual estimates and calibrated our observations using both periodic point-counts and independent visual estimates by the two surveyors. Percent bed and bank exposure was also measured (e.g. Figure 3f) using visual estimation within the same region where we estimated grainsize. Other observations of incision process or channel character were photographed and noted in field books.

Some channels were surveyed multiple times using the different survey tools and then compared to the elevation data from topographic maps and the DEM. Both survey techniques were within the error of elevations we estimated using 20 m contour topographic maps. We found that the laser rangefinder's accuracy in measuring profile information and ability to shoot through trees was vastly improved by frequent compass calibration and when a reflective target was used.

4. Channel Morphology and Bed State

In this section, we present and discuss survey data hand-collected from 27 channels that contain knickpoints. For each measured parameter (slope, width, grainsize and bed exposure), we discuss the presence or absence of trends in the three reach-types: upstream of the knickpoint, at the knickpoint and downstream of the knickpoint. For each reach, we discuss the mean parameter values measured in individual channels as well as the trends present in the aggregated data of all 27 channels. We do not provide a discussion of downstream trends in individual reaches. Though this analysis was done, we find that the change in drainage area in a single reach is relatively small compared to the variation in the measured parameters (e.g. slope or width) and there are no discernable trends. But this does not rule out trends with distance where the area is roughly constant. Because the parameters we are analyzing are influenced by the multiple variables that we are plotting them against, we must be careful about the discernable trends that we identify. For example, plotting width against slope may not reveal a correlation between these two variables but instead reflect the fact that width can be correlated with area and that area is correlated with slope.

4.1 Channel Gradient

Measurements of channel gradient provide significant insight into the capacity for a channel to transport sediment and erode its bed. In alluvial channels, the steady-state slope is suggested to be 'self-formed' as a consequence of the channel's necessity to transport the upstream-imposed sediment load (Willgoose et al., 1991). In bedrock channels, the steady-state slope is suggested to be a function of the rock uplift rate, the erodability of the substrate, the erosion mechanism and the upstream drainage area (Howard and Kerby, 1983; Whipple and Tucker, 1999). We are confident that uplift rate, substrate erodability and drainage area do not vary significantly in the relatively short reaches upstream and downstream of knickpoints. This allows us to assume that the channel gradient within the studied reaches is largely a consequence of changing incision processes and the adjustment to the pulse of incision. We use a linear least squares regression through the measured slopes upstream and downstream of the knickpoint to determine a single characteristic slope within these two reaches. In these measurements we left a small (~100m) buffer around the knickpoint so that the local influence of the knickpoint would not impact the slope measurements. Due to the moderate sinuosity of the channels, the measured gradient is significantly lower than that for the valley centerline or the aggradational terrace surface. Comparisons of our surveyed channel gradients to those measured from the topographic contour maps reveal that our findings are consistent with the topographic maps, within the error of the contour elevations (±10 meters, NZ Lands Office). Using Flint's law (1974), we also examine the variation in the parameters defining the power law relationship between channel slope and drainage area, $S = k_s A^{-\theta}$, where S is the local channel gradient, A is the upstream drainage area, k_s is the channel steepness index and θ is the channel concavity.

4.1.1 Channel Gradient Upstream of the Knickpoint

By aggregating and bin-averaging (at ~60 meter intervals) the gradient data for all 27 channels (Figures 4a and 5a), we are able to observe along-stream trends not apparent in individual channels. Plotting gradient data with respect to distance downstream of the knickpoint (Figure 4a), we find that the variability in channel gradients increases as the knickpoint is approached from upstream. We also find that the bin-average gradient also persistently increase from 0.0126 to 0.0569 in the first kilometer upstream of the knickpoint. (Figure 4a). Though our visual observations and impression in the field suggested that there was no discernible steepening, this measured change in average gradient upstream of the knickpoint lip suggests that incision is somewhat enhanced upstream of the knickpoint lip. The fact that the increasing gradient persists over such a large distance upstream of the knickpoint suggests that this response is more related to enhanced mobilization of sediment, rather than a local effect responding to the acceleration of the flow as it becomes super-critical at the free overflow. This is supported by the fact that there was no evidence for discrete notching at any of the knickpoint lips. It has been suggested that at free overflows, as sediment just upstream of the knickpoint lip is mobilized, the bare bedrock bed's decreases the effective roughness, thus initiating a response that propagates upstream. Upstream patches of sediment are mobilized by the progressive removal of downstream sediment. In the discussion section we expand on whether this hypothesis is supported by our other observations.

By aggregating all the gradient data for reaches upstream of the knickpoints, we are able to span almost two and a half orders of magnitude in drainage area, ranging from 2 x 10^5 m^2 to 5 x 10^7 m^2 (Figure 5a, red points). Plotting the variation in gradient with respect to drainage area, we find that the suite of measured gradients provide a surprisingly well defined relation between gradient and area (Figure 5a, red line). The coefficients for the stream power law discussed above are $k_s = 55.57 \pm 1.10$ and $\theta = 0.517\pm$ 0.074 (errors are 2Ω). These values are very close to the values measured in non-knickpoint reaches in other portions of the basin using DEM-extracted, contour-only stream profile analysis. Because both reaches have a concavity is close to 0.5, we can compare the measured steepnesses in the two reaches.

Examining the mean channel gradient upstream of the knickpoint in each individual channel (Figure 6a, open circles) we find that most (17 of 27) channels have a lower mean gradient in the upstream reach than in the reach downstream of the knickpoint. When error bars reflecting the standard error of the mean are provided for each point, only 13 or the 27 channels have gradients that can be considered statistically different (without overlapping error bars). Of those 13, 10 still have lower gradients in the upstream reach. This result is consistent with the observed lower slopes in the upstream reaches in Figure 4a.

4.1.2 Channel Gradient at the Knickpoint

Knickpoint gradients observed in the 27 surveyed streams range from ~15 degrees up to ~85 degrees at the steepest headwalls. The highest gradients are not recorded in the survey data as it was too difficult or dangerous to survey directly over the knickpoint lip. Instead an inclinometer was used at the base of the knickpoint to approximate the mean slope of the headwall. In more than half the cases, knickpoint were composed of multiple large steep steps, distributed over than no more than 50m of horizontal channel distance. The upstream extent of our knickpoint reach is defined as the lip of the most upstream large, steep step. The downstream extent of the knickpoint reach is defined by where the decrease in channel elevation is no longer primarily on near vertical steps > 0.5 meter in height. The gradient values presented in the survey data (green dots in Figures 4a and 5a) often reflect the averaging of multiple steep steps and do not adequately reflect the fact that the knickpoint reaches are composed multiple steep steps, culminating at a large, very steep headwall. As visible in Figure 4a, knickpoint reaches can persist for considerable distances downstream (up to 300m), especially in channels where the magnitude of the local base level fall is greater than 100m. As the gradients and processes observed within the knickpoint reach are so steep that the assumption of steady and uniform flow are violated and our surveys to not accurately capture the extreme slopes in these reaches, we do not perform analyses of concavity or steepness within the knickpoint reach.

4.1.3 Channel Gradient Downstream of the Knickpoint

Immediately downstream of the knickpoint reach, channel gradients remain high (~0.2) but by ~300 meters downstream of the knickpoint lip (Figure 4a, blue points), the gradients decrease gradually toward an average value of 0.0372. The variability in gradients also decreases with increasing distance from the knickpoint (Figure 4a). None of the surveys in this data set extend into the aggraded reaches downstream of the knickpoint. These trends in channel gradient are most visible when viewing the log-bin averaged points (drainage area range was divided into 50 log-distributed bins). We interpret the higher slopes observed in the reach downstream of the knickpoint as a consequence of incomplete adjustment to the base level fall signal. This incomplete response is apparent in oversteepened hillslopes shedding coarse debris into the channel that are ubiquitous along these sections of the surveyed streams.

A power-law regression through the gradient data downstream of the knickpoint from all 27 surveyed channels (Figure 5a, blue line) confirms that channel gradients are characteristically higher in the downstream reach than in the upstream reach ($k_s = 89.61 \pm 1.06$, 2Ω). The rate of change in slope ($\theta = 0.508 \pm 0.071$, 2Ω) is very similar to that observed in the upstream reach, suggesting that the two reaches are equally adjusted (or unadjusted) to the disturbance. This is supported by the fact that the regression through both the upstream and downstream gradient data (Figure 5a, black line) has a very similar concavity to the two separate reaches, ($\theta = 0.500 \pm 0.054$, 2Ω) thus, their slopes in log-log space are parallel.

As discussed in section 3.2.1, the mean gradients for each individual channel reach are presented in Figure 6a. 17 of the 27 surveyed channels are steeper in their downstream reaches and if you only consider the data where the two means are statistically unique (non overlapping standard errors of the means), then only 10 of 13 channels are steeper in the downstream reach. We will expand on the reasons for the steeper downstream slopes in the downstream reaches in the discussion section.

4.2 Bankfull Width

We utilize a definition for bankfull width (e.g. Montgomery and Gran, 2001; e.g. Snyder et al., 2000) where the measurement is collected between the two distinct edges of the active floodplain, perpendicular to the downstream direction of the channel. In the incised channels of the Waipaoa, flood plains are scarce so we primarily used evidence for scour and vegetation disturbance to determine the high-water width (e.g. Hack, 1957). In an alluvial channel, where the bed and banks are assumed to be composed of non-cohesive materials, the channel's bankfull width is argued to be "self-formed." This implies that the channel's form (slope, depth, width, etc.) is adjusted to the channel's necessity to transport out the upstream-imposed sediment load. Under these conditions numerous workers (Suzuki, 1982; Wohl and Ikeda, 1998; Wohl and Merritt, 2001) have recognized a power-law scaling between the bankfull width and upstream drainage area, $W = k_w A^c$, where W is the bankfull width, k_w is a dimensional constant, A is the upstream drainage area and c is a non-dimensional constant.

In contrast to alluvial channels, the beds and banks of bedrock streams are less susceptible to rapid modification by the flow. Some argue that, as in an alluvial channel, the width of a bedrock channel is self determined, but that sediment and substrate properties exert significantly more influence on the stable width (Montgomery and Gran, 2001; Snyder et al., 2003; Whipple, 2004). Though bedrock and alluvial channel widths are anticipated to be functions of different variables, it is surprising that the width-area scaling for alluvial gravel-bedded channels often cannot be distinguished from the width-area scaling of well adjusted, moderately erodable bedrock channels (Finnegan et al., 2005; Montgomery, 2004). Others have discussed how local adjustments in channel width often correspond to changes in channel slope, substrate erodability, or rock uplift rate (Hack, 1957; Montgomery and Gran, 2001; Whipple, 2004). These observations suggest that in the reaches downstream of knickpoints, we would expect a decrease in channel width as a consequence of elevated incision rates or channel confinement in gorges below the knickpoint.

4.2.1 Channel Width Upstream of the Knickpoint

Above the knickpoint, within the relict portion of the landscape, the channels have incised through the alluvial fill and are etched 0.5-1.5 meters into the underlying bedrock. This incision is not restricted to the reach just upstream of the knickpoint but persists for 0.5 – 1.0 kilometer upstream. Because these channels are both surprisingly flat-bottomed and confined within the incised alluvial fill, there is little change in the width of individual channels within the surveyed reach. The bin-averaged aggregated data for the upstream widths of all surveyed channels present relatively consistent values within 500 meters of the knickpoint (Figure 4b). Within this reach, bin-averaged (at a ~60 meter interval) widths range between 3.15 m and 3.71 m. Though widths appear to be increasing in the upstream direction at distances greater than ~750m, this is largely the consequence of the decreasing number of channels surveyed that far upstream. The variability in the measured widths is large when the data for all channels is complied into a single data set and plotted against distance from knickpoint. In any particular channel, there was little variability in width. This is evident in Figure 6b where the gray vertical error bars are for the width data are considerably smaller than those for the slope or percent bed exposure data. This low variability in any single channel is a consequence of the incised, confined nature of the streams. Because the channel is confined within bedrock banks, the stream has little lateral mobility.

In Figure 5b (the red line) we demonstrate the well defined power-law relation between width and drainage area in the reaches upstream of the knickpoint in the Waipaoa. The coefficient for the power-law k_w is 0.02266 ± 0.615 and the exponent on drainage area, c, is 0.335 ± 0.041. These are in good agreement with previously observed values (e.g. Finnegan et al., 2005) but the coefficient is considerably smaller and the error is considerably large.

4.2.2 Channel Width at the Knickpoint

Channel widths at the knickpoint and within the knickpoint reach are variable and show no consistent trend. Within the knickpoint reach, the adjacent banks are oversteepened and highly unstable and the resulting channel widths reflect whether recent failures have or have not occurred. In knickpoints with large headwall waterfalls, we perceive a widening of the valley just downstream of the falls. This could be simply a consequence of the lack of large, established vegetation in that location relative to further downstream or a it could be a consequence of enhanced shrink-swell weathering of the adjacent valley walls where spray is intermittently directed. As shrink-swell weathering is highly effective form of regolith generation in the Waipaoa River catchment, this process could dominate the failure and retreat of steep knickpoint faces where there is little interaction between the sediment/flow and the face of the knickpoint. Further surveying of these features is necessary to understand this observation more completely.

4.2.3 Channel Width Downstream of the Knickpoint

Downstream of the knickpoint, the channels are deeply incised into bedrock and confined by high, steep bedrock walls ~20-100 meters high with gradients often greater than 1. In some locations where large landslide deposits fill the valley floor, temporary floodplains were noted and the channel width was measured appropriately. In Figure 4b, the compiled width data downstream of the knickpoint reveal that high-flow widths are consistently higher than those observed in the reaches upstream of the knickpoint. In the ~750 meters downstream of the knickpoint, log-bin averaged widths range between 4.12 m and 5.24 m. This observation is surprising considering that (1) the downstream reaches are the most incised portion of the channel and (2) the downstream reaches range over the same breadth of drainage areas as do the upstream measurements. The consistency in the elevated widths in the downstream direction suggest that the observation is not just a local effect near the knickpoint, but could be a response to increased sediment flux due to the oversteepened banks. As discussed above, the behavior of the width values at distances greater than 750 m downstream of the knickpoint reflect a significantly smaller sub-sample of the 24 analyzed channels and are not considered in this analysis.

Plotting all of the surveyed downstream width data against drainage area (Figure 5b, blue line), we find that, as was the case in the upstream reach, the data are well described by a power law function where the k_w value is 0.01298 ± 0.640 and the c value is 0.379 ± 0.043. The regressions of the upstream and downstream reaches have a very similar form and are very close to the regression thought both datasets (Figure 5b, black line).

Comparisons between the mean widths of individual channels are plotted in Figure 6b. 18 of the 24 channels have larger mean widths in the downstream reach than in the upstream reach. If considering only the channels that have statistically unique means (non-overlapping standard error of the mean error bars), we find that 13 of 15 channels are steeper in the downstream reach. It is valuable to note that in the width plot, the upstream and downstream mean areas are significantly offset, but that the lines between the two data points are steeper than the trend of the regression through the data. If they were parallel, the difference in the channel widths could be explained as a consequence of increasing drainage area downstream.

The anticipated relation between width and slope suggests that as slopes become steeper, channels narrow. To test this expectation, we plotted width data from all three reaches and all 27 channels against local slope (Figure 7a). We find that at the extreme slopes (high and low), there are only a few data points constraining the log-bin averages, but in the core of the data, there is a recognizable trend toward increasing widths with slope. In the steepest reaches, the slope remains constant at the value of the mean slope. This agrees with our observation that there is no notching or pronounced slot in the knickpoint or within the downstream incised reaches. We will return to this observation in the discussion section.

4.3 Sediment Grainsize

Frequent visual estimates of mean grainsize (an approximation of D_{50}) were collected within surveyed reaches. These estimates were made with the intention of distinguishing whether the sediment upstream of the knickpoint was characteristically larger or smaller than that found downstream of the knickpoint. Based on preliminary observations, we hypothesized that mean grainsize would coarsen downstream of the knickpoint as a consequence of abundant material shed locally off the over-steepened gorge walls. Because each channel survey is relatively short, we did not anticipate observable downstream fining. Given that the Waipaoa River catchment has experienced intense modification in the last few hundred years (deforestation and livestock grazing), we are also cautious of whether the observed grainsize trends are demonstrative of steady channel processes or are consequent to recent land use practices.

Plotting all of the mean grainsize data relative to the distance downstream of the knickpoint (Figure 4c), we observe no distinct difference between upstream and downstream reaches. Instead, the bin-averaged grainsizes increase roughly monotonically over the distance between ~320 m upstream of the knickpoint to ~420 m downstream of the knickpoint. Over that distance, mean grainsize increases from 22 mm to 77 mm. There are no abrupt changes downstream of the knickpoint. At distances greater than ~320 m upstream of the knickpoint, the grainsizes appear to be decreasing in the downstream direction, but this is likely a consequence of the decreasing number of channels represented with increasing upstream distance.

Plotting all of the mean grainsize data relative to increasing drainage area, we find no downstream trends that would be indicative of downstream fining. Instead grainsize increases and decreases with no discernable trend in the downstream direction. There are no distinct gravel or sand fractions visible in the data.

In order to compare upstream and downstream reaches on a channel-by-channel basis, we plot the mean values in each reach for each channel in Figure 6c. In agreement with Figure 4c, 14 of the 21 channels show an increase in grainsize between the upstream and downstream reaches. If we only consider reach pairs whose standard errors or the mean are not overlapping, we find the 12 of the 17 surveyed channels steepen between their upstream and downstream reaches. Plotting grainsize against slope (Figure 7b), we find that at extremely high or low slopes, the log-bin averaged mean grainsize decreases significantly. The decreasing grainsize may be a consequence that during the last large event all coarse material was flushed out, and during low flow conditions that the survey was completed under, the only sediment moving in the channel may have been the fine material. The decrease in grainsize in lower slope reaches could be a consequence that there is a negative correlation between channel slope and stability of the adjacent hillslope. In low slope reaches (upstream of the knickpoint or a significant distance downstream of the knickpoint (see Figure 4a)), the channel tends to have less steep and thus more stable banks, resulting in less coarse sediment being shed into the channel.

We attribute the observed trends (or lack thereof) in the mean grainsize data to a suite of factors. In the surveyed channels, the bedrock lithology is dominated by a clay-rich mudstone that is highly susceptible to aggressive shrink-swell weathering processes. As material is shed into the channel by mass failures along the banks of the channel or by tributaries, the sediment is rapidly broken down by shrink-swell weathering from cobble to fine gravel over half dozen wet-dry cycles. We found that most mudstone cobbles exposed to a few wet and dry cycles were easily broken into fine-gravel under the weight of our feet. Remobilized terrace deposits or infrequent interbeds of sandy or less clay rich material do provide more robust clasts, but in our surveys of sediment grainsize we did not differentiate between different lithologies that certainly have different downstream fining behaviors. In the narrow gorges that are below the sediment transport threshold most of the time, it is also difficult to differentiate between which sediments are locally derived deposits and which are derived from further upstream. When a storm event does occur, grainsizes and bed morphology are radically altered. In channels that were repeat surveyed, the locations of and mean grainsizes of the point and mid-channel bars were radically different and are largely composed of the highly angular material shed into the channel during the storm event. This suggests that in these highly confined, flashy channels, storm events create discharges that are above the threshold to mobilize most of the bed sediment and thus, the deposits and bed material that we observe during our low-flow survey season do not represent long term, steady grainsize conditions. Recent deforestation and grazing practices may have decreased the hydraulic lag time during a storm event, subsequently changing flow dynamics and subsequently the observed sediment grainsize distributions. Further comparative studies between the hydraulic response of channels within pristine catchments and those in modified catchments are necessary to determine if there have been significant changes in discharge characteristics as a consequence of land use.

4.4 Percent Bed Cover

The exposure of the bedrock floor of a channel increases its susceptibility to erosion by abrasion and weathering by wet/dry cycling. Recent work has demonstrated that when channel beds typically covered with a consistent blanket of alluvium are swept clear, incision rates are enhanced (Crosby and Whipple, in preparation). In the tributaries of the Waipaoa River catchment we anticipate bed cover to reflect the different hydraulic and sediment delivery conditions present above, at and below the knickpoint. Plotting the percent bed exposure against the distance downstream of the knickpoint (Figure 4d), we find that through the entire length of the surveyed channel the mean bed exposure is around 50 %. Examining the individual measurements in Figure 4d, we also find that the % bed exposure is highly variable along the channel's entire length. This suggests that even in locations where the % exposure is very low or even zero, the bed cover is not anticipated to be thick enough to remain intact during a large discharge event. There is a subtle but robust peak in the percent bed exposure between ~175 m upstream of the knickpoint and ~125 m downstream of the knickpoint (Figure 4d). To confirm this peak, we decreased the bin size for the bin averaging and found that the monotonic increase and subsequent decrease in bed exposure remained consistent even with bin sizes reduced from 60 m to 15 m. Though this confirms that there is a consistent local increase in bed exposure adjacent the knickpoint, outside of this local region, for the 15 m bins, the bin averaged data are highly variable and show no consistent trends. This suggests that flow acceleration and the increased slope near the knickpoint as well as the extreme slopes downstream of the knickpoint keep the bed more consistently exposed. Further downstream, the materials shed off the oversteepened hillslopes cannot be transported as effectively as in the steep reach that persists for ~125 m downstream of the knickpoint. Though not surveyed, we observed that ~1-2 km upstream and downstream of the surveyed reaches the bed cover becomes consistently 100%. If we had surveyed these more distal reaches, we would observe the upstream and downstream extremes in Figure 4d to go consistently toward zero, thus making the plot bell-shaped with the peak centered on the knickpoint.

Plotting bed exposure against drainage area (Figure 5d), we find no consistent trends. The same is true when plotting the reach-average for each of the individual channels (Figure 6d). We find that half the channel beds (12 of 24) are more exposed upstream and the other half are more exposed downstream.

Considering only the channels whose reaches have standard error's that do not overlap, we find the same result; 7 of 14 channel beds are more exposed upstream and an equal number are more exposed downstream.

Plotting percent exposure against slope (figure 7c), we confirm that for slopes greater than 0.01, the exposure consistently increases with increasing slope. At lower slopes there is high variability in percent exposure. This observed relation between slope and bed exposure may be particularly well expressed in the Waipaoa channels because the sediment is so easily broken down via weathering processes. In regions with more resilient sediment, we may expect instead of an increase in bed exposure to find a coarsening of mean grainsize with increasing slope as only the larger clasts would be immobile on the steeper slopes.

5. Discussion

The analysis of channel survey data presented above outlines the relations (and lack thereof) observed among the parameters measured in 27 channels that contain knickpoints. The analysis also sheds light on the necessity for further work to measure longer reaches and collect additional types of data discussed below. The analysis of aggregated data above assumes that all knickpoints were responding to the same disturbance in the same manner, thus facilitating lumping of observations and their subsequent analysis as a single dataset. Under this assumption, variability observed in the data is considered a consequence of local noise such as a harder substrate or a recent landslide. It is equally plausible that there are multiple knickpoint types, each responding to different base level fall histories and thus possessing their own type-specific morphology, erosion processes and bed state (grainsize, percent exposure, etc). By identifying and separating out the different types, the measured parameters may have provided more consistent or definitive results. In the end, we decide to lump rather than split the knickpoint survey data because we found each knickpoint's unique morphology or downstream trend could be explained by a particular circumstance. These circumstances could be either internally (within the local tributary) or externally determined (a consequence of boundary conditions with the tributary's trunk stream). When trying to split, we found nearly as many knickpoint types as there were knickpoints and thus, we focus this analysis on establishing the consistencies between all channels. It appears that the channels' morphologic and bedstate responses following base level fall are not a deterministic response but rather a probabilistic one where highly variable internal and external forcings introduce the observed scatter in the data. We dedicate another manuscript to elucidating the internal and external sources of the knickpoint variability (2004).

We also found that channel surveys examining downstream trends in parameters with respect to drainage area require the aggregation of multiple channels because the variation in the measured parameter is often significantly larger than the changes in drainage area. This is evident in Figure 6, where the error bars representing the standard error of the mean are much greater in the vertical direction (the Measured

Parameter) than in the horizontal direction (Drainage Area). Aggregate channel data broadens the range of areas and facilitates observations of downstream trends not visible in smaller datasets.

5.1 Incision Process Observations and Discussion

5.1.1 Incision Processes Upstream of the Knickpoint

The low gradient reaches upstream of the knickpoint are dominated by abrasion and weathering processes. Cross-sectional profiles across upstream reaches are typically flat bottomed with abrupt, steep sided banks, well approximated by a trapezoid. Beds are frequently extremely planar, with few steps or over-deepened pools. Potholes and fluted features were observed within these planar beds, suggesting abrasion by fine bedload or suspended sediment. Though abrasion features sculpt the floor of the channel, the most effective erosion mechanism appears to be shrink-swell degradation of portions of the bed that rise above the low-water line. Throughout the upstream reach, if local scour created a depression in the bed so that the low discharge could be routed through the deep portion, the above water bedrock was often fractured and exfoliated to a depth of ~1cm. Bedrock that remains continuously submerged retains its structural integrity and has no fractures. Following the work of Montgomery (2005) and Stock et al. (Mason et al., 2004; Pederson et al., 2004), we suggest that the channel bed self-regulates to maintain a planar surface. As local incision occurs, other portions of the channel are elevated above waterline and susceptible to rapid degradation by weathering processes. As the high points are worn down, the channel tends back toward a planar morphology.

The gradual increase in gradient as the knickpoint is approached from upstream suggests that the knickpoint exerts a non-local control over channel gradient (Figure 4a). Earlier we discuss whether this could be a consequence of enhanced sediment mobility due to the free overflow and the development of an upstream propagating decrease in effective bed roughness. This hypothesis is not well supported by observed highly variable bed exposure upstream of the knickpoint (Figure 4d), but is supported by visual observations that at significant distances upstream of the knickpoint (~ 2 km) bedrock exposure drops to near-zero. That said, observations of bed cover are likely highly variable depending on the time since the last large storm event and both natural (landslide activity) and anthropogenic (livestock grazing or forestry) disturbances. These issues could be resolved by extending future surveys further upstream.

5.1.2 Incision Processes at the Knickpoint

The steep nature of the knickpoint face prevents the effective interaction between sediment or water and the rock. Impact of water and sediment on the bare bedrock bases of most of the knickpoints appears to be an effective vertical incision mechanism right at the toe of the knickpoint. Of the 27 channels surveyed, only three could be considered to have shallow plunge pools at their base. In those cases, the pools do not under-cut the face.

In a study of knickpoints in a weathering dominated lithology (e.g. Sklar and Dietrich, 2004), buttress-like protruding knickpoint faces are suggested to form where active discharge keeps the knickpoint face wet and inhibits weathering by wetting and drying. The flanks of the knickpoint buttress are not kept continually wet and thus erode faster than the face of the falls. In the Waipaoa, the plan-form profiles of the knickpoints are either linear or amphitheatre shaped and in no location were the knickpoints observed protruding from the adjacent canyon wall. The multiple processes that likely drive the retreat of the knickpoint face all depend on the preparation of the rock material through shrink-swell weathering processes. The most fundamental source of episodic wetness is sue to spray from the waterfalls that remain active throughout the year. During the dry season, the knickpoint faces are only partially kept wet and the fringes of the falls are often composed of shattered rock that is frequently observed collapsing. Though there are no active springs, permeability contrasts or undercut faces on the Waipaoa knickpoints, there is some evidence that seepage or groundwater discharge from and near the knickpoint face. Episodic flow through these fractures would result in shrinking and swelling of the opposing faces and overtime loosen blocks toward the open face. Sapping-driven amphitheater development could also be responsible for the perceived widening of the incised valley near the knickpoint face discussed in section 3.3.2.

Some knickpoints are composed of multiple drops, seldom separated by more than 20 meters along-stream distance. We propose that the downstream steps could have been the previous position of the knickpoint, and that after experiencing collapse of only the upper portion of the face, that portion of the step retreated upstream. Our observations from the Waipaoa River catchment favor the interpretation that most large magnitude, near-vertical knickpoints are no longer retreating at a rate determined by fluvial processes. Instead, the retreat of the face is largely a consequence of weathering-mediated block failure. Fluvial processes are responsible for the transport of the collapsed material, but its degeneration, and preparation for transport is largely mediated by shrink-swell processes (Stock et al., 2005and references herein; Taylor and Smith, 1986). More study is required to constrain the erosional efficiency of falling water and sediment at the base of the falls and how that process contributes to maintaining the steep face of the knickpoint.

5.1.3 Incision Processes Downstream of the Knickpoint

Downstream of the knickpoint, channels are steep but not step-dominated as is the case in the knickpoint reach. Surveyed channel slopes are always considerably steeper than those measured on the terrace treads directly above (e.g. Figure 1a). This difference in slope is expected considering (1) the terrace slope was established during an aggradational period and (2), the modern channel reflects the incomplete adjustment of the channel downstream of the knickpoint. In the parallel retreat case (Figure 1b), where the initial equilibrium slope is lowered to another equilibrium slope as the knickpoint retreats upstream, we would expect the downstream reach to share the same slope as the terrace tread. This scenario is highly unlikely because the downstream reach lacks adjusted hillslopes and has not reestablished equilibrium sediment fluxes from local or upstream sources. In sediment flux dependent incision rules (Crosby et al., submitted), channels cannot reestablish equilibrium form until sediment flux from the hillslopes to the channels has also equilibrated. In the Waipaoa, many of the reaches downstream of the knickpoint are in narrow gorges with terraces on either side. The terraces act as buffers, capturing and storing modern sediment shed off hillslopes. If small tributaries entering the reach downstream of the knickpoint remain hung over the lip of the incised gorge, they fail to communicate the signal of incision into the adjacent hillslopes . The most effective mechanisms observed in the Waipaoa for hillslope reestablishment of connection with their incised trunk stream are (1), small progressive slope failures of the gorge walls and (2), large, catastrophic, deep seated mass movements that deliver entire hillslopes down toward the gorge. Though the large mass movements can block channels and initiate dams, the weak and weathering-susceptible nature of the bedrock prevent these features from persisting long enough to leave a long term signature in the channel profile (Korup, 2004).

Though the same active weathering processes apply in the downstream reaches, the steeper slopes and higher sediment fluxes have an influence on the observed incision processes. Downstream reaches are less consistently flat bottomed that the reaches upstream of the knickpoint. There are frequent small steps and irregularities in the bed where plucking processes are observed. It is likely that the fractures bounding the loose block are mediated by weathering processes, but its ultimate removal and downstream transport are fluvially driven. In most channels, the bedrock banks are littered with locally exfoliated chips. During high flow events, this material is swept away, leaving a high-water mark and exposing the next fresh surface for shrink-swell weathering.

5.1.4 Improvements for Future Observations of Incision Processes near Knickpoints

Future work will benefit from including more extensive cross-sectional profiling of the downstream reach to quantify the degree of relaxation of the gorge walls with distance downstream from the knickpoint. These measurements may help constrain our interpretations regarding the mobility of the knickpoint and the formation of broad amphitheatre heads at locations of knickpoint stagnation. It will also be valuable to collect cross sectional data upstream of the knickpoint to determine how far the incision upstream of the knickpoint persists. Measurements of channel roughness, or the rate of change in local slope, could provide further insight into the differences between upstream and downstream reaches. As the knickpoint faces are actively eroding, it would also be valuable to collect data monitoring their horizontal retreat.

6. Conclusions

Our field observations suggest that knickpoint headwalls in the Waipaoa River catchment are actively retreating by weathering facilitated block failure. The presence of a free overflow upstream of the knickpoint and the steep slopes downstream of the headwall prevent sediment accumulation and keep the bed exposed to abrasion and weathering dominated erosion processes. Further upstream of the knickpoint, the well adjusted, flat-bottomed nature of the channels suggests that these reaches are stable and equilibrated to the upstream, internal forcing. The low sediment fluxes assumed present in this subdued, relict portion of the landscape inhibit the incision of the knickpoint lip. As well, the active retreat of the knickpoint face removes any record of local incision at the lip. The adjustment of the oversteepened reach downstream of the collapsing face is driven by fluvial plucking, abrasion, and removal of material prepared through shrink-swell weathering processes. Though oversteepened gorge walls downstream of the knickpoint increase the supply of coarse sediment to the channel, the rapid destruction of these mudstones through shrink-swell weathering limits its viability as a tool for abrasion of the channel bed. It was surprising to find well defined Slope-Area and Width-Area relations in these transient portions of the landscape, and even more surprising to find the relations so similar between the upstream and downstream reaches. The trends in mean grainsize and % bed exposure when plotted relative to slope suggest that these parameter should be further examined in further studies of channels that do and do not contain knickpoints. Though our analysis groups together multiple morphologically and genetically distinct knickpoints, the consistent trends recognized in the data suggests similarities in response that are common for all knickpoints.

Reference

Berlin, M. M., and Anderson, R. S., submitted, Roan Plateau Waterfalls record Landscape Response to Late Cenozoic Incision of the Colorado River: JGR-Earth Surface.

Berryman, K., Marden, M., Eden, D., Mazengarb, C., Ota, Y., and Moriya, I., 2000, Tectonic and paleoclimatic significance of Quaternary river terraces of the Waipaoa River, east coast, North Island, New Zealand: New Zealand Journal of Geology and Geophysics, v. 43, no. 2, p. 229-245.

Berryman, K., Marden, M., Palmer, A. S., Wilson, K., Mazengarb, C., and Litchfield, N., in preparation, The post-glacial downcutting history in the Waihuka tributary of the Waipaoa River, Gisborne District, New Zealand, and implications for tectonics and landscape evolution.

Bishop, P., Hoey, T. B., Jansen, J. D., and Artza, I. L., 2005, Knickpoint recession rate and catchment area: the case of uplifted rivers in Eastern Scotland: Earth Surface Processes and Landforms, v. 30, p. 767-778.

Black, R. D., 1980, Upper Cretaceous and Tertiary geology of Mangatu State Forest, Raukumara Peninsula, New Zealand: New Zealand Journal of Geology and Geophysics, v. 23, no. 3, p. 293-312.

Brown, L. J., 1995, Holocene shoreline depositional processes at Poverty Bay, a tectonically active area, northeastern North Island, New Zealand: Quaternary International, v. 26, p. 21-33.

Brush, L. M., Jr., and Wolman, M. G., 1960, Knickpoint behavior in noncohesive material--A laboratory study: Geological Society of America Bulletin, v. 71, no. 1, p. 59-73.

Chappell, J., 2001, Sea Level Change Through the Last Glacial Cycle: Science, v. 292, p. 679-686.

Collot, J., Lewis, K., Lamarche, G., and Lallemand, S., 2001, The giant Ruatoria debris avalanche on the northern Hikurangi margin, New Zealand; results of oblique seamount subduction: Journal of Geophysical Research, Solid Earth, v. 106, no. B9, p. 19,271-19,297.

Crosby, B. T., and Whipple, K. X., in preparation, Internal and External Sources of Knickpoint Variability: Undecided Journal.

-, in press, Knickpoint Initiation and Distribution within Fluvial Networks: 236 waterfalls in the Waipaoa River, North Island, New Zealand: Geomorphology.

Crosby, B. T., Whipple, K. X., Gasparini, N. M., and Wobus, C. W., submitted, Formation of Fluvial Hanging Valleys: Theory and Simulation: Journal of Geophysical Research – Earth Surface.

DeRose, R. C., Gomez, B., Marden, M., and Trustrum, N. A., 1998, Gully erosion in Mangatu Forest, New Zealand, estimated from digital elevation models: Earth Surface Processes and Landforms, v. 23, no. 11, p. 1045-1053.

Eberhart-Phillips, D., and Chadwick, M., 2002, Three-dimensional attenuation model of the shallow Hikurangi subduction zone in the Raukumara Peninsula, New Zealand: Journal of Geophysical Research, Solid Earth, v. 107, no. B2, 2033, p. doi: 10.1029/2000JB000046.

Eden, D. N., Palmer, A. S., Cronin, S. J., Marden, M., and Berryman, K. R., 2001, Dating the culmination of river aggradation at the end of the last glaciation using distal tephra compositions, eastern North Island, New Zealand: Geomorphology, v. 38, p. 133-151.

Finnegan, N. J., Roe, G., Montgomery, D. R., and Hallet, B., 2005, Controls on the channel width of rivers: Implications for modeling fluvial incision of bedrock: Geology, v. 33, no. 3, p. p. 229–232; doi: 10.1130/G21171.1.

Flint, J. J., 1974, Stream gradient as a function of order, magnitude, and discharge: Water Resources Research, v. 10, p. 969-973.

Foster, G., and Carter, L., 1997, Mud sedimentation on the continental shelf at an accretionary margin - Poverty Bay, New Zealand: New Zealand Journal of Geology and Geophysics, v. 40, no. 2, p. 157-173.

Gardner, T. W., 1983, Experimental study of knickpoint and longitudinal profile evolution in cohesive, homogeneous material: Geological Society of America Bulletin, v. 94, no. 5, p. 664-672.

Gilbert, G. K., 1896, Niagara Falls and their history, National Geographic Society: The Physiography of the United States: New York, The American Book Co., p. 203-236.

Hack, J. T., 1957, Studies of longitudinal stream profiles in Virginia and Maryland, U. S. Geological Survey Professional Paper 294-B, p. 45-97.

Hancock, G. S., Anderson, R. S., and Whipple, K. X., 1998, Beyond power: Bedrock river incision process and form, *in* Tinkler, K., and Wohl, E. E., eds., Rivers Over Rock: Fluvial Processes in Bedrock Channels: Washington, D. C., AGU Press, p. 35-60.
Hasbargen, L. E., and Paola, C., 2000, Landscape instability in an experimental drainage basin: Geology, v. 28, no. 12, p. 1067-1070.
Haviv, I., Enzel, Y., Whipple, K. X., Zilberman, E., Stone, J., Matmon, A., and Fifield, K. L., in press, Amplified erosion above waterfalls and oversteepened bedrock reaches: Journal of Geophysical Research – Earth Surface.
Holland, W. N., and Pickup, G., 1976, Flume study of knickpoint development in stratified sediment: Geological Society of America Bulletin, v. 87, no. 1, p. 76-82.
Howard, A. D., and Kerby, G., 1983, Channel changes in badlands: Geological Society of America Bulletin, v. 94, no. 6, p. 739-752.
Jones, K. L., 1988, Horticulture and settlement chronology of the Waipaoa River catchment, east coast, North Island, New Zealand: New Zealand Journal of Archaeology, v. 10, p. 19-51.
Korup, O., 2004, Geomorphic characteristics of New Zealand landslide dams: Engineering Geology, v. 73, no. (1-2), p. 13-35.
Kuehl, S., Carter, L., Gomez, B., and Trustrum, N., 2003, Holistic Approach Offers Potential to Quantify Mass Fluxes Across Continental Margins: EOS TRANSACTIONS AGU, v. 84, no. 38, p. 379, 388.
Lewis, K. B., and Pettinga, J. R., 1993, The emerging, imbricate frontal wedge of the Hikurangi margin, *in* Ballance, P. F., ed., Sedimentary Basins of the World: South Pacific Sedimentary Basins: Amsterdam, Elsevier Science Publishers B.V., p. 225-250.
Litchfield, N. J., and Berryman, K. R., 2005, Correlation of fluvial terraces within the Hikurangi Margin, New Zealand: implications of climate and baselevel controls: Geomorphology, v. 68, p. 291-313.
-, 2006, Relations between postglacial fluvial incision rates and uplift rates in the North Island, New Zealand: Journal of Geophysical Research – Earth Surface, v. 111, p. F02007, doi: 10.1029/2005JF000374.
Litchfield, N. J., Ellis, S., Berryman, K. R., and Nicol, A., submitted, Subduction related uplift in the Hikurangi Margin, New Zealand: Journal of Geophysical Research – Earth Surface.
Marden, M., Arnold, G., Gomez, B., and Rowan, D., 2005, Pre- and post-reforestation gully development in Mangatu Forest, East Coast, North Island, New Zealand: River Research and Applications, v. 21, no. 7, p. 757-771.
Mason, L. J., Pederson, D. T., Goble, R. J., Voorhies, M. R., and Anonymous, 2004, Salt weathering of waterfall escarpments along the Niobrara River near Valentine, NE: Abstracts with Programs - Geological Society of America, v. 36, no. 5, p. 231.
Mazengarb, C., Francis, D. A., and Moore, P. R., 1991, Sheet Y16 - Tauwhareparae, Geological map of New Zealand, scale 1:50,000: Department of Scientific and Industrial Research.
Mazengarb, C., and Speden, I. G., 2000, Geology of the Raukumara Area: Institute of Geological and Nuclear Sciences 1:250,000 geological map 6: Lower Hutt, New Zealand, Institute of Geological and Nuclear Sciences Limited, 60 p.
McGlone, M. S., 2001, A late Quaternary pollen record from marine core P69, southeastern North Island, New Zealand: New Zealand Journal of Geology and Geophysics, v. 44, no. 1, p. 69-77.
-, 2002, A Holocene and latest Pleistocene pollen record from Lake Poukawa, Hawke's Bay, New Zealand: Global and Planetary Change, v. 33, no. 3-4, p. 283-299.
McLea, W. L., 1990, Palynology of Pohehe Swamp, Northwest Wairarapa, New Zealand; a study of climatic and vegetation changes during the last 41,000 years: Journal of the Royal Society of New Zealand, v. 20, no. 2, p. 205-220.
Miller, J. R., 1991, The Influence of Bedrock Geology on Knickpoint Development and Channel-Bed Degradation Along Downcutting Streams in South-Central Indiana: Journal of Geology, v. 99, no. 4, p. 591-605.
Montgomery, D. R., 2004, Observations on the role of lithology in strath terrace formation and bedrock channel width: American Journal of Science, v. 304, no. 5, p. 454-476.
Montgomery, D. R., and Gran, K. B., 2001, Downstream variations in the width of bedrock channels: Water Resources Research, v. 37, no. 6, p. 1841-1846.
Newnham, R. M., Lowe, D. J., and Williams, P. W., 1999, Quaternary environmental change in New Zealand: a review: Progress in Physical Geography, v. 23, p. 567-610.

Orpin, A. R., 2004, Holocene sediment deposition on the Poverty-slope margin by the muddy Waipaoa River, East Coast New Zealand: Marine Geology, v. 209, p. 69-90.
Ota, Y., Hull, A. G., Iso, N., Ikeda, Y., Moriya, I., and Yoshikawa, T., 1992, Holocene marine terraces on the northeast coast of North Island, New Zealand and their tectonic significance: New Zealand Journal of Geology and Geophysics, v. 35, p. 273-288.
Pederson, D. T., Mason, L. J., Goble, R. J., Voorhies, M. R., and Anonymous, 2004, Freeze thaw erosion of waterfall escarpments along the Niobrara River near Valentine, NE: Abstracts with Programs - Geological Society of America, v. 36, no. 5, p. 231.
Pullar, W. A., 1962, Soils and agricultrue on the Gisborne Plains: New Zealand Department of Scientific and Industrial Research Soil Bureau Bulletin, v. 20, no. 92.
Rosenbloom, N. A., and Anderson, R. S., 1994, Hillslope and channel evolution in a marine terraced landscape, Santa Cruz, California: Journal of Geophysical Research, Solid Earth, v. 99, no. B7, p. 14,013-14,029.
Schumm, S. A., Mosley, M. P., and Weaver, W. E., 1987, Experimental fluvial geomorphology, p. 413.
Seidl, M. A., and Dietrich, W. E., 1992, The problem of channel erosion into bedrock: Catena Supplement: Functional geomorphology; Landform Analysis and Models, v. 23, p. 101-124.
Shackelton, N. J., 1987, Oxygen isotopes, ice volumes and sealevel: Quaternary Science Reviews, v. 6, p. 183-190.
Sklar, L. S., and Dietrich, W. E., 2004, A mechanistic model for river incision into bedrock by saltating bed load: Water Resources Research, v. 40, p. no.6, 21.
Snyder, N. P., Whipple, K. X., Tucker, G. E., and Merritts, D. J., 2000, Landscape response to tectonic forcing; digital elevation model analysis of stream profiles in the Mendocino triple junction region, Northern California: Geological Society of America Bulletin, v. 112, no. 8, p. 1250-1263.
-, 2002, Interactions between onshore bedrock-channel incision and nearshore wave-base erosion forced by eustasy and tectonics: Basin Research, v. 14, no. 2, p. 105-127.
-, 2003, Channel response to tectonic forcing: field analysis of stream morphology and hydrology in the Mendocino triple junction region, northern California: Geomorphology, v. 53, no. 1-2, p. 97-127.
Snyder, N. P., Whipple, K. X., Tucker, G. E., Merritts, D. J., and Anonymous, 1999, Evidence for an equilibrium between main-trunk channel incision and tectonic uplift; Mendocino triple junction region, Northern California: Abstracts with Programs - Geological Society of America, v. 31, no. 7, p. 444-445.
Stock, J. D., and Montgomery, D. R., 1999, Geologic constraints on bedrock river incision using the stream power law: Journal of Geophysical Research-Solid Earth, v. 104, no. B3, p. 4983-4993.
Stock, J. D., Montgomery, D. R., Collins, B. D., Dietrich, W. E., and Sklar, L. S., 2005, Field measurements of incision rates following bedrock exposure; implications for process controls on the long profiles of valleys cut by rivers and debris flows: Geological Society of America Bulletin, v. 117, no. 1-2, p. 174-194.
Suzuki, T., 1982, Rate of lateral planation by Iwaki River, Japan: Transactions, Japanese Geomorphological Union, v. 3, no. 1, p. 1-24.
Taylor, R. K., and Smith, T. J., 1986, The engineering geology of clay minerals: Swelling, shrinking and mudrock breakdown: Clay Minerals, v. 21, p. 235-260.
Weissel, J. K., and Seidl, M. A., 1997, Influence of rock strength properties on escarpment retreat across passive continental margins: Geology, v. 25, no. 7, p. 631-634.
-, 1998, Inland Propagation of Erosional Escarpments and River Profile Evolution Across the Southeast Australian Passive Continental Margin, *in* Tinkler, K. J., and Wohl, E. E., eds., Rivers Over Rock: Fluvial Processes in Bedrock Channels: Washington, D. C., American Geophysical Union, p. 189-206.
Whipple, K. X., 2004, Bedrock Rivers and the Geomorphology of Active Orogens: Annual Review of Earth and Planetary Sciences, v. 32, p. 151-185.
Whipple, K. X., Hancock, G. S., and Anderson, R. S., 2000, River incision into bedrock: Mechanics and relative efficacy of plucking, abrasion, and cavitation: Geological Society of America Bulletin, v. 112, no. 3, p. 490-503.
Whipple, K. X., and Tucker, G. E., 1999, Dynamics of the stream-power river incision model; implications for height limits of mountain ranges, landscape response timescales, and research needs: Journal of Geophysical Research, Solid Earth, v. 104, no. B8, p. 17,661-17,674.
Willgoose, G., Bras, R. L., and Rodriguez-Iturbe, I., 1991, A coupled channel network growth and hillslope evolution model. 1. Theory: Water Resources Research, v. 27, no. 7, p. 1671-1684.

Wobus, C. W., Crosby, B. T., and Whipple, K. X., 2006a, Hanging valleys in fluvial systems: Controls on occurrence and implications for landscape evolution: Journal of Geophysical Research – Earth Surface, v. Vol. 111, no. No. F2, p. F02017, 10.1029/2005JF000406.
-, in press, Hanging valleys in fluvial systems: Controls on occurrence and implications for landscape evolution: Journal of Geophysical Research – Earth Surface.
Wobus, C. W., Whipple, K. X., Kirby, E., Snyder, N. P., Johnson, J., Spyropolou, K., Crosby, B. T., and Sheehan, D., 2006b, Tectonics from topography: Procedures, promise and pitfalls, *in* Willett, S. D., Hovius, N., Brandon, M.T., and Fisher, D. M., ed., Tectonics, Climate and Landscape Evolution: Geological Society of America Special Paper 398: Penrose Conference Series, p. 55-74.
Wohl, E. E., and Ikeda, H., 1998, Patterns of bedrock channel erosion on the Boso Peninsula, Japan: Journal of Geology, v. 106, no. 3, p. 331-345.
Wohl, E. E., and Merritt, D. M., 2001, Bedrock channel morphology: Geological Society of America Bulletin, v. 113, no. 9, p. 1205-1212.
Yoshikawa, T., 1988, Pattern and rate of tectonic movement and late Quaternary geomorphic development in the Raukumara Peninsula, northeastern North Island, New Zealand: Bulletin of the Department of Geography, University of Tokyo, v. 20, p. 1-28.

Figures

Figure 1: In this schematic diagram, we provide a generalized profile of a knickpoint from the Waipaoa River catchment (1a) and define the locations of the relict, knickpoint and adjusting portions of the channel. We suggest that the evolution of the form of the long profile as the knickpoint retreats upstream depends on the relative magnitudes of three incision vectors: v1, vertical incision above the knickpoint, v2, horizontal erosion of the knickpoint face and v3, vertical incision at the base of the knickpoint. Given an initial simplified channel profile (thick black line in b-g), different combinations of vector magnitudes result in a diverse set of transient and final knickpoint morphologies (b,c,d,e,f,g). Note that in some scenarios, though their transient evolutions are distinctly different, the final knickpoint profiles are similar.

Figure 2: This map of the upper Waipaoa River catchment, located on the North Island of New Zealand, shows the locations of mapped knickpoints (white triangles), the upstream extent of the landscape adjustment to the base level fall signal, and the 27 surveyed channels (numbered thick black lines). The dense cluster of surveyed channels in the western portion of the basin is composed of tributaries to the Waihuka Stream. Note that all surveyed channels cross the boundary between the adjusting downstream region and the upstream relict region.

Figure 3: Representative channel survey data from the Ruahine Stream (#11 in Figure 2). Channel profile and bed-state (a, c, d, e, f) were hand surveyed using a laser-rangefinder compass, GPS and real-time GIS data acquisition software. Drainage areas (b) were calculated at and upstream of tributary junctions using a DEM accumulation array. The vertical gray line indicates the upstream extent of the knickpoint reach for reference in all plots.

Figure 4: Point-wise data from all 27 surveyed channels plotted with respect to their distance downstream of the upstream extent of the knickpoint. Negative values are upstream of the knickpoint. Red, green and blue dots represent data from upstream, knickpoint and downstream reaches, respectively. The large black dots represent the ~60 m bin-averages of the upstream and downstream data. Note that upstream of the knickpoint, mean slopes increase as the knickpoint is approached (a) though mean widths remain relatively steady (b). Downstream of the knickpoint, mean slopes are initially high, and decrease with distance downstream (a), but mean widths remain consistently higher than in the upstream reach (b). As discussed in the text, though the mean slopes and widths plotted here are influenced by the increasing density of low-drainage area channel surveys near the knickpoint, the observed response of the channel's gradient and width are robust. Mean grain size of bed sediment and percent bed exposure (c and d respectively) do not demonstrate a clear dependence on the distance from the knickpoint, besides the abrupt increase in bed exposure near the knickpoint between ~-175 m and ~125 m.

Figure 5: Point-wise data from all 27 surveyed channels plotted with respect to their drainage area. Red, green and blue dots represent data from upstream, knickpoint and downstream reaches, respectively. Note that these colors are well distributed across the range of drainage areas. The large black dots represent the 50 log-distributed bin-averages of the upstream and downstream data. Note the anticipated power law relation between gradient and area (a) and width and area (b). Of particular interest is the consistency between these power law relations for the upstream and downstream reaches. Mean grainsize of bed sediment and percent bed exposure do not demonstrate a clear dependence on drainage area (c and d respectively).

Figure 6: Four plots comparing mean channel parameters upstream and downstream of their knickpoint. Closed circles indicate the mean value the parameter downstream of the knickpoint while open circles describe the mean values for the upstream reach. Lines connecting open and closed circles help establish whether a particular parameter is larger above or below the knickpoint. Gray lines indicate the standard error of the mean for each point. Note that the gradients and widths (plots a and b) most often have lower mean values in upstream reaches than downstream reaches. Also note that the power law dependence of gradient and width on drainage area observed in Figures 5a and 5b is retained in these representative reaches. More channels are coarser in the reach downstream of the knickpoint, (c), though this is not as well pronounced as with gradient and width. There is an equal split between channels that have a greater bed exposure downstream of the knickpoint compared to those upstream of the knickpoint (d).

Figure 7: Three plots demonstrating how bankfull width, mean grainsize and percent bed exposed vary with gradient. Width increases slightly with decreasing gradient in the central cloud of data, yet the relation is not well constrained (a). Mean grainsize increases with gradient, but at high gradients, the grainsize decreases, likely as a consequence of the only material present is brought in at low flows (b). The percent bed exposure increases with gradient, as expected, but it is unclear why bed exposure increases at lower gradients. This could reflect a sampling from only a few lower gradient channels.

Figure 1

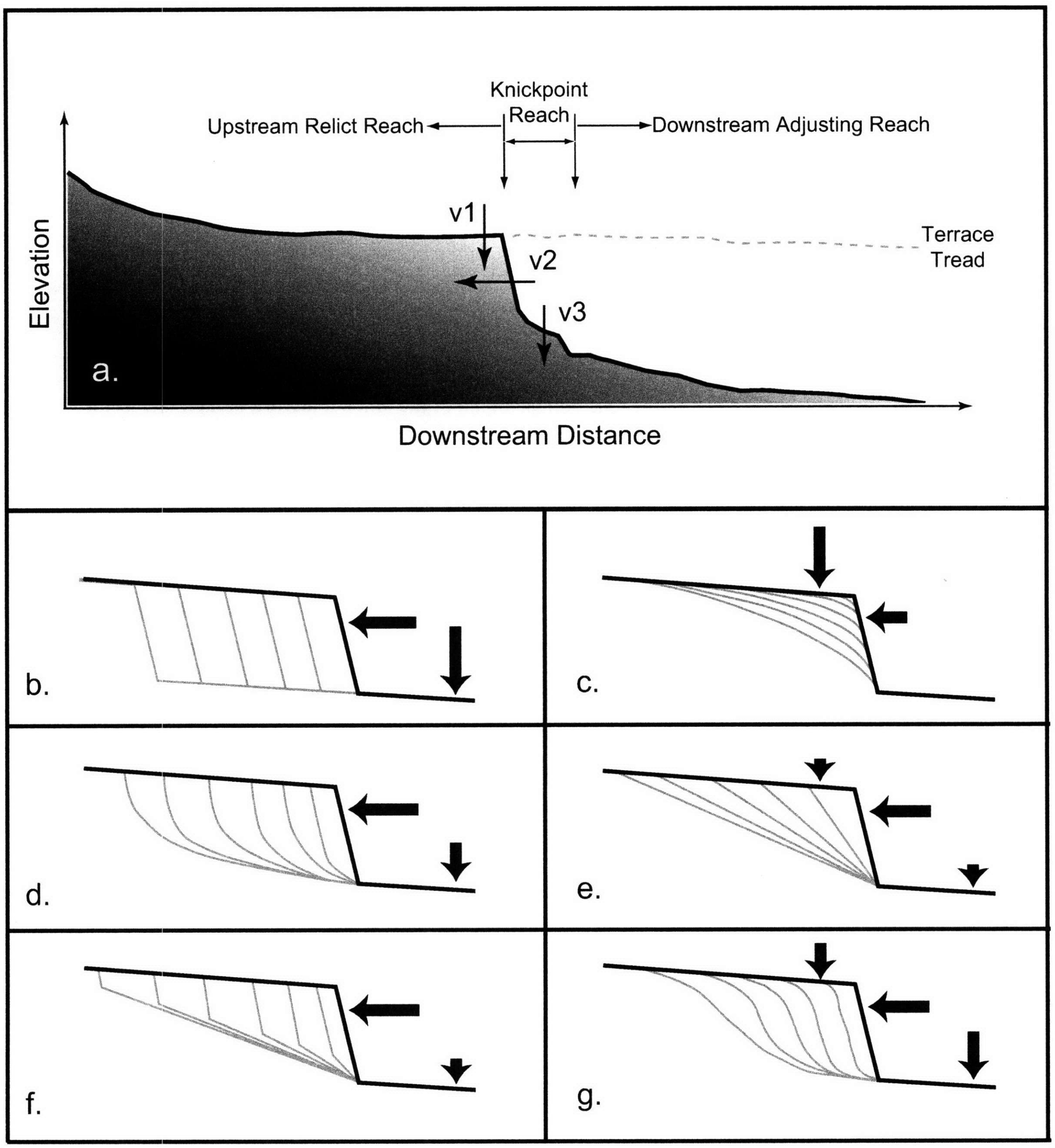

Knickpoint
Reach
Upstream Relict Reach
Downstream Adjusting Reach
v1
v2
v3
Terrace
Tread
Elevation
a.
Downstream Distance
b.
c.
d.
e.
f.
g.

Figure 2

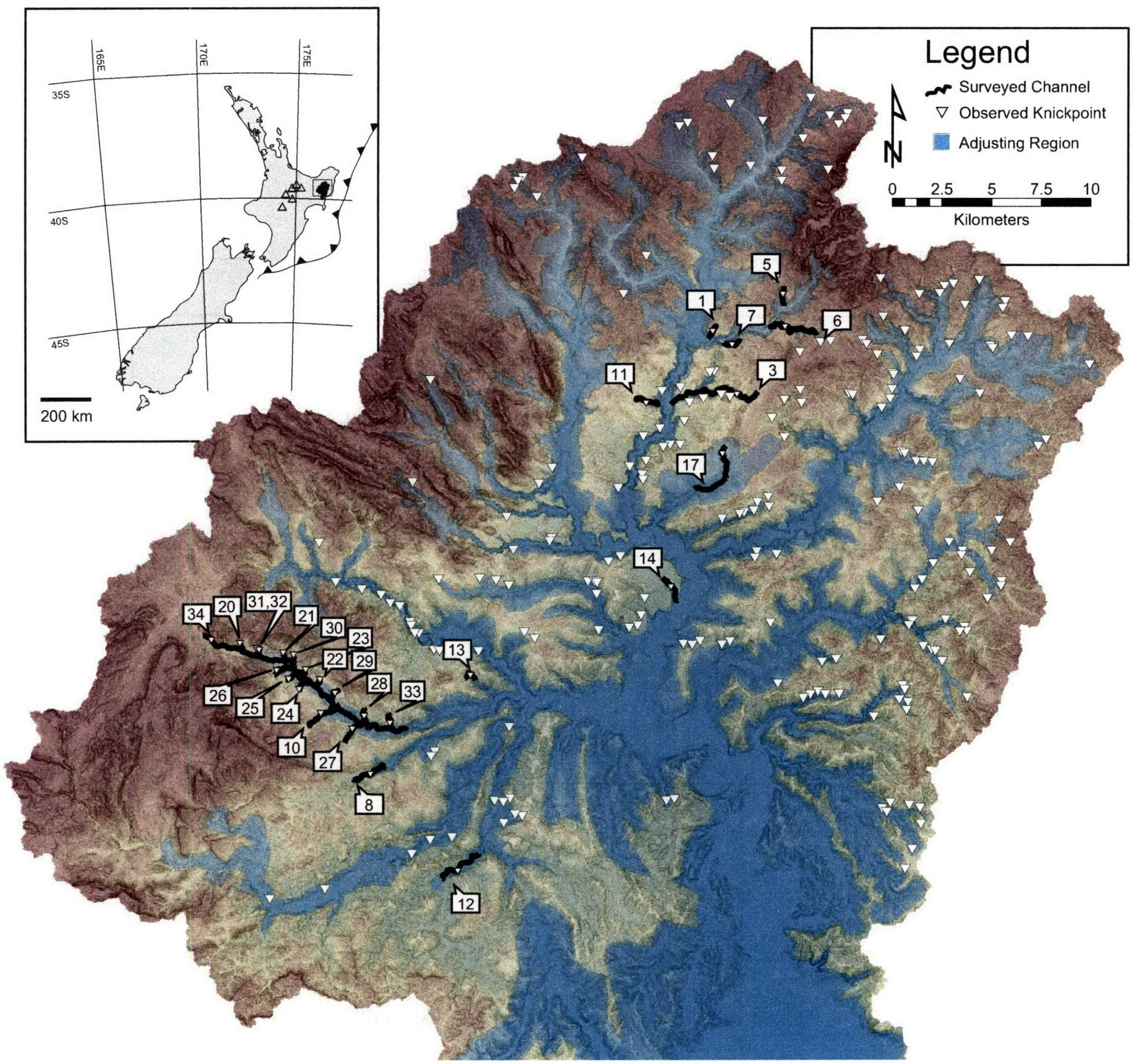

Legend
Surveyed Channel
Observed Knickpoint
Adjusting Region
N
0 2.5 5 7.5 10
Kilometers
165E
170E
175E
35S
40S
45S
200 km
1
3
5
6
7
8
10
11
12
13
14
17
20
21
22
23
24
25
26
27
28
29
30
31,32
33
34

Figure 3

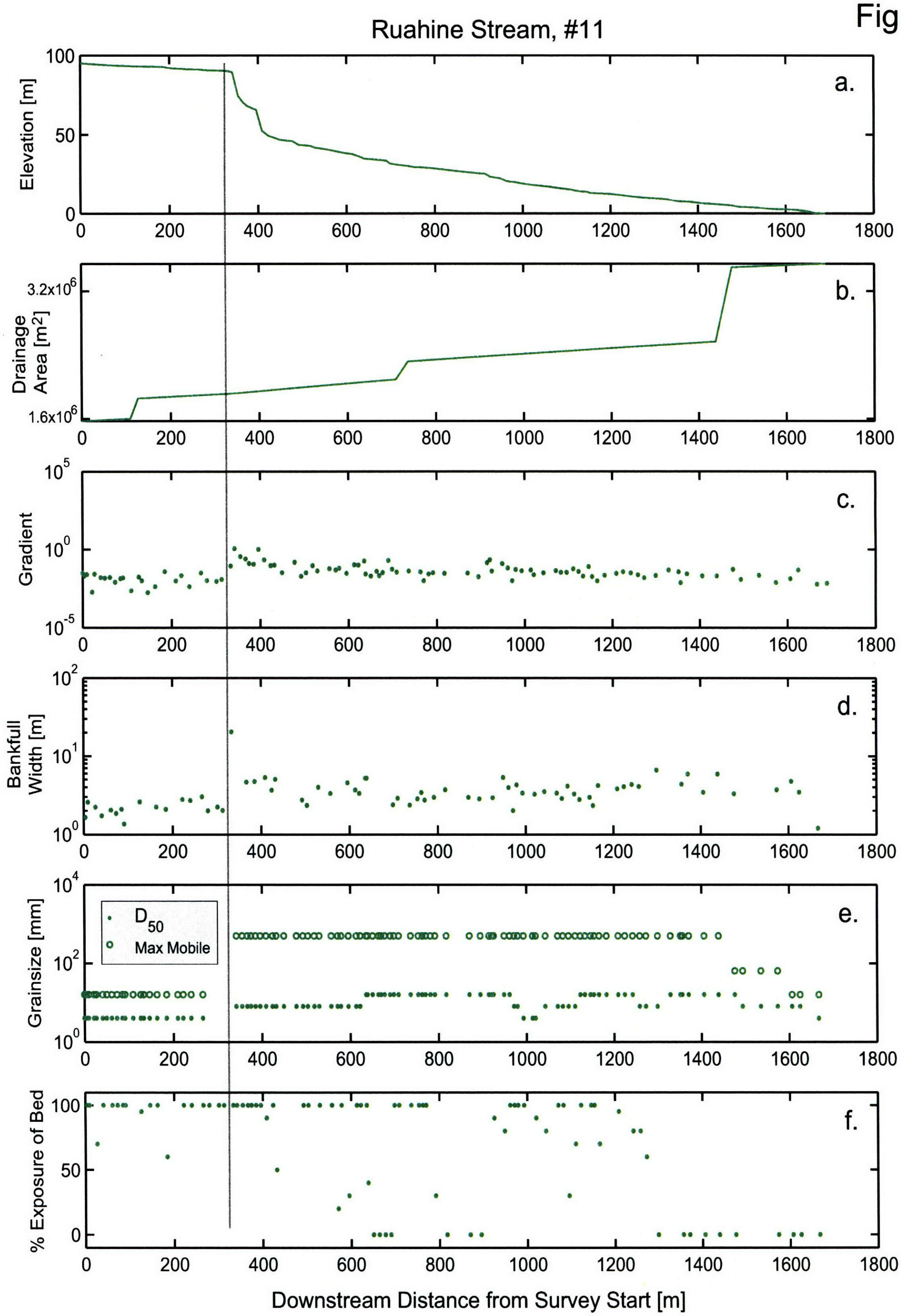

Ruahine Stream, #11
Elevation [m]
a.
Drainage Area [m2]
b.
3.2x10^6
1.6x10^6
Gradient
c.
Bankfull Width [m]
d.
Grainsize [mm]
e.
D50
Max Mobile
% Exposure of Bed
f.
Downstream Distance from Survey Start [m]

Figure 4

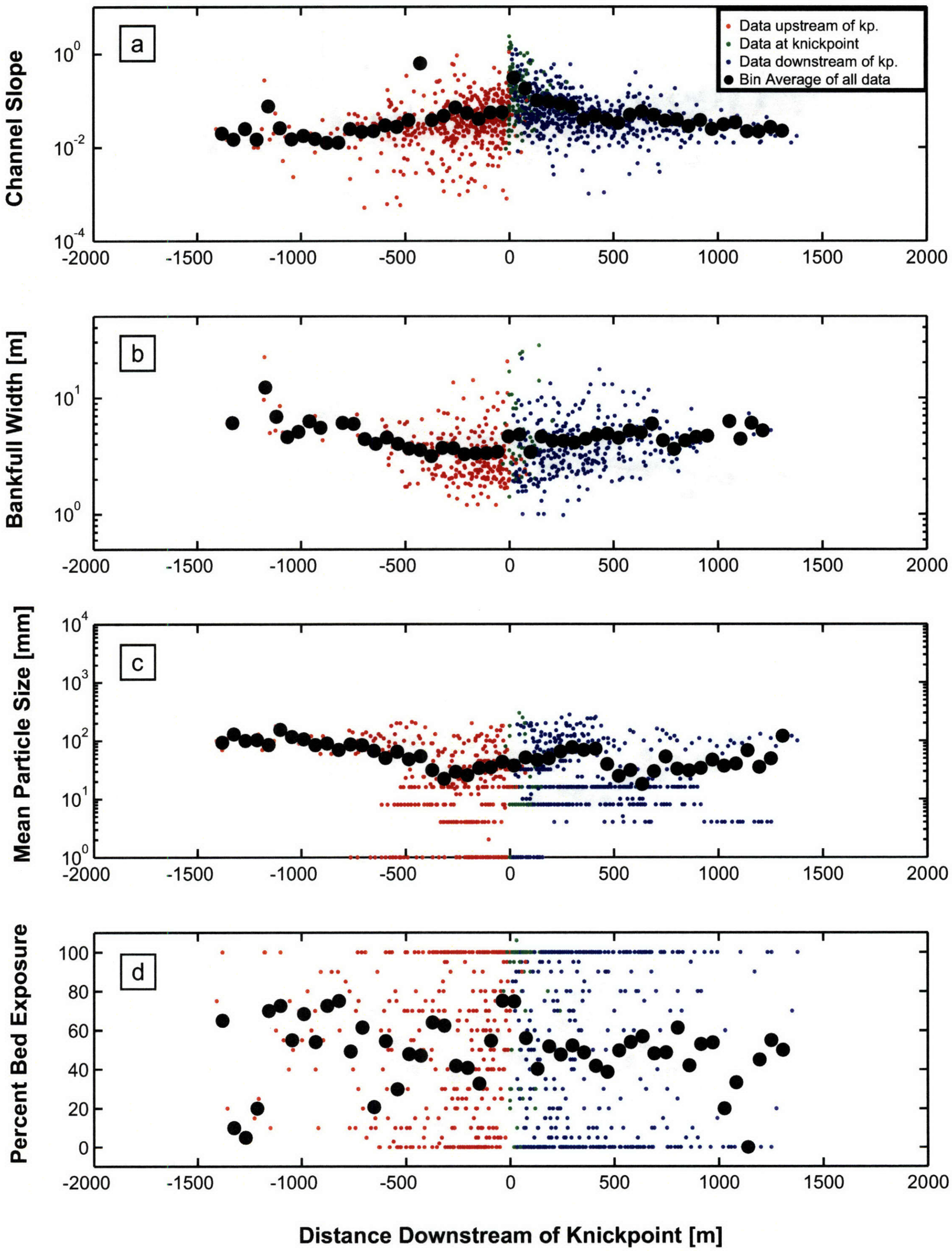

a
Channel Slope
10^0
10^{-2}
10^{-4}
Data upstream of kp.
Data at knickpoint
Data downstream of kp.
Bin Average of all data
b
Bankfull Width [m]
10^1
10^0
c
Mean Particle Size [mm]
10^4
10^3
10^2
10^1
10^0
d
Percent Bed Exposure
100
80
60
40
20
0
-2000
-1500
-1000
-500
0
500
1000
1500
2000
Distance Downstream of Knickpoint [m]

Figure 5

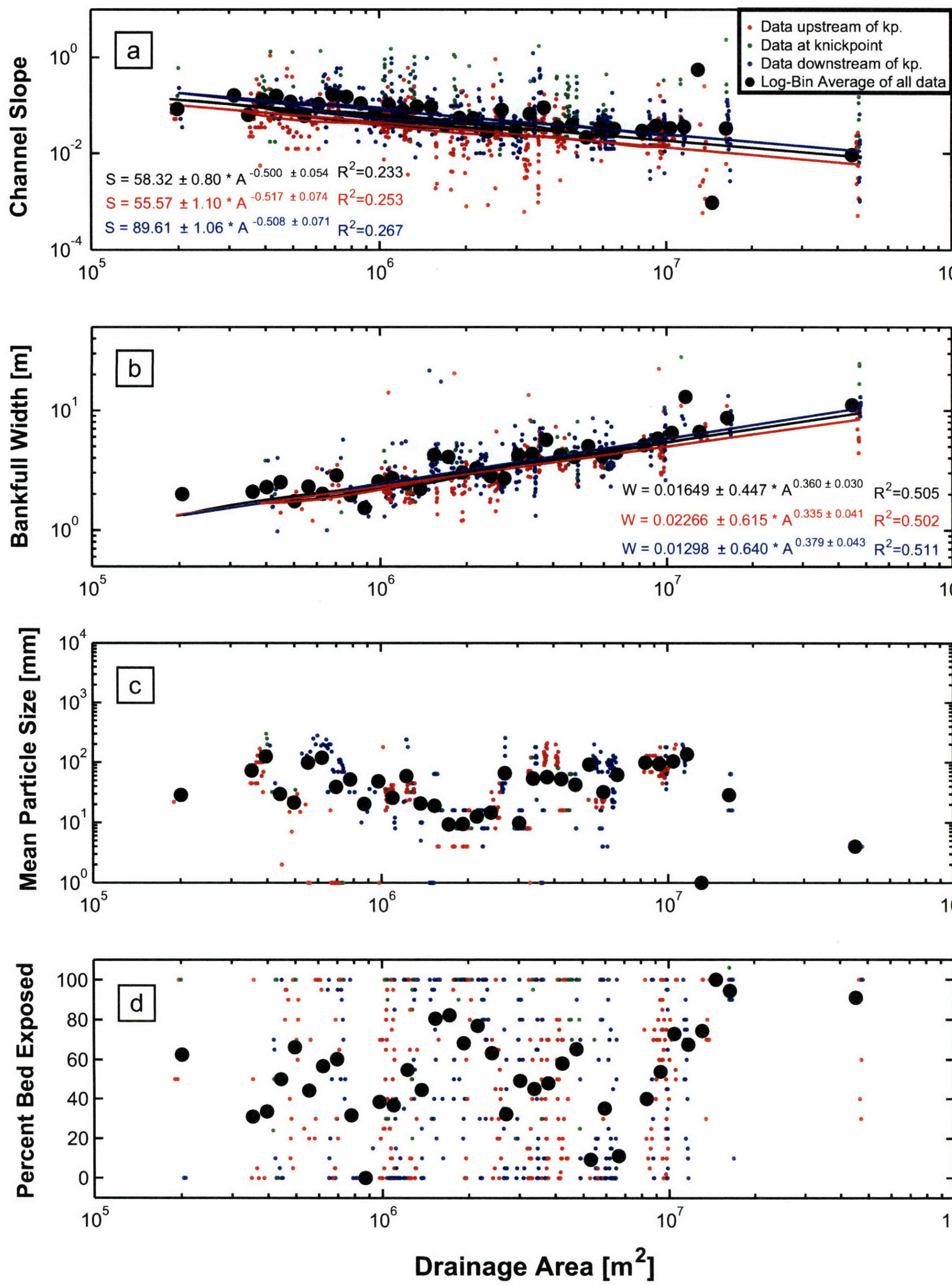

a
Data upstream of kp.
Data at knickpoint
Data downstream of kp.
Log-Bin Average of all data
S = 58.32 ± 0.80 * A -0.500 ± 0.054 R2=0.233
S = 55.57 ± 1.10 * A -0.517 ± 0.074 R2=0.253
S = 89.61 ± 1.06 * A -0.508 ± 0.071 R2=0.267
Channel Slope
b
W = 0.01649 ± 0.447 * A 0.360 ± 0.030 R2=0.505
W = 0.02266 ± 0.615 * A 0.335 ± 0.041 R2=0.502
W = 0.01298 ± 0.640 * A 0.379 ± 0.043 R2=0.511
Bankfull Width [m]
c
Mean Particle Size [mm]
d
Percent Bed Exposed
Drainage Area [m2]

Figure 6

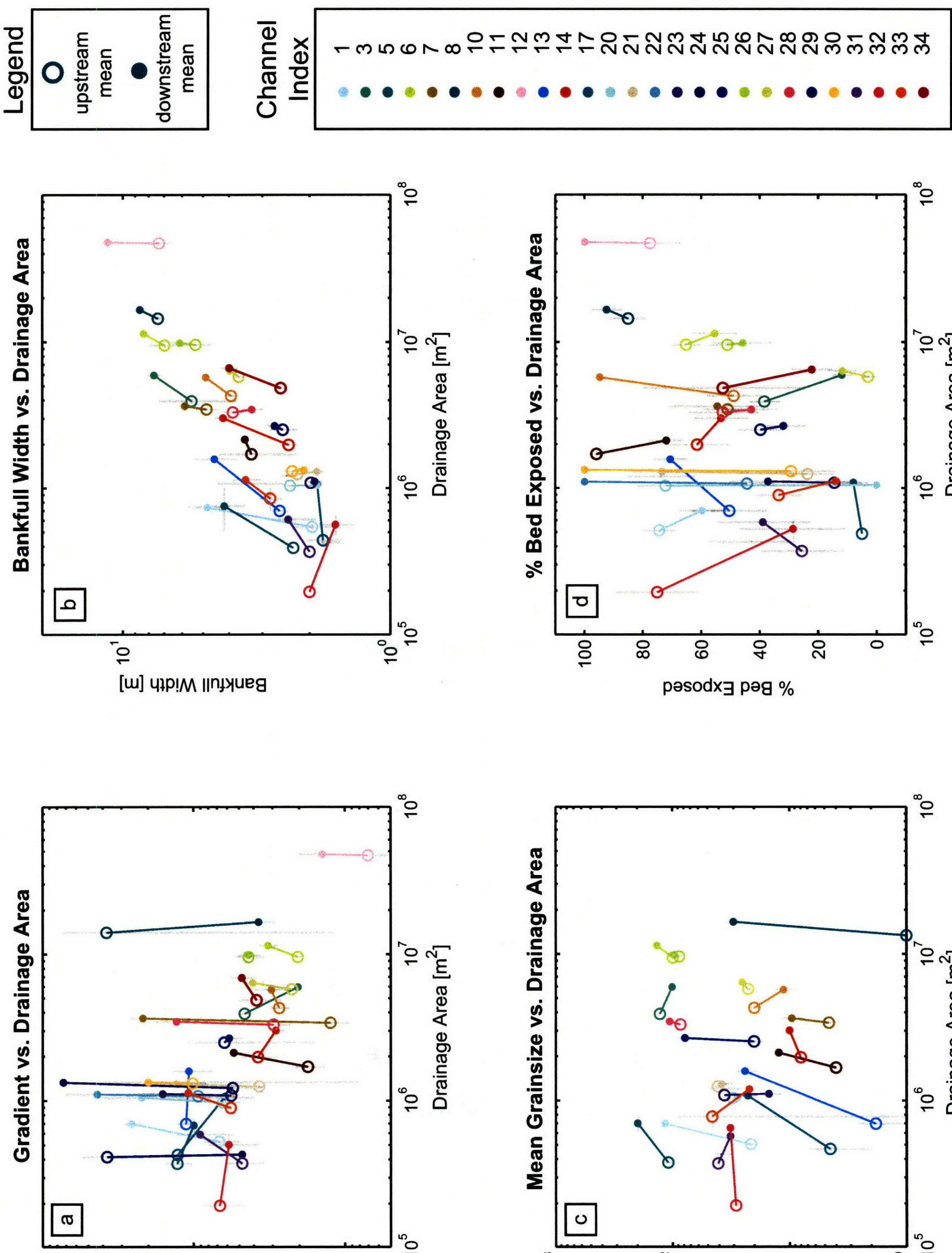
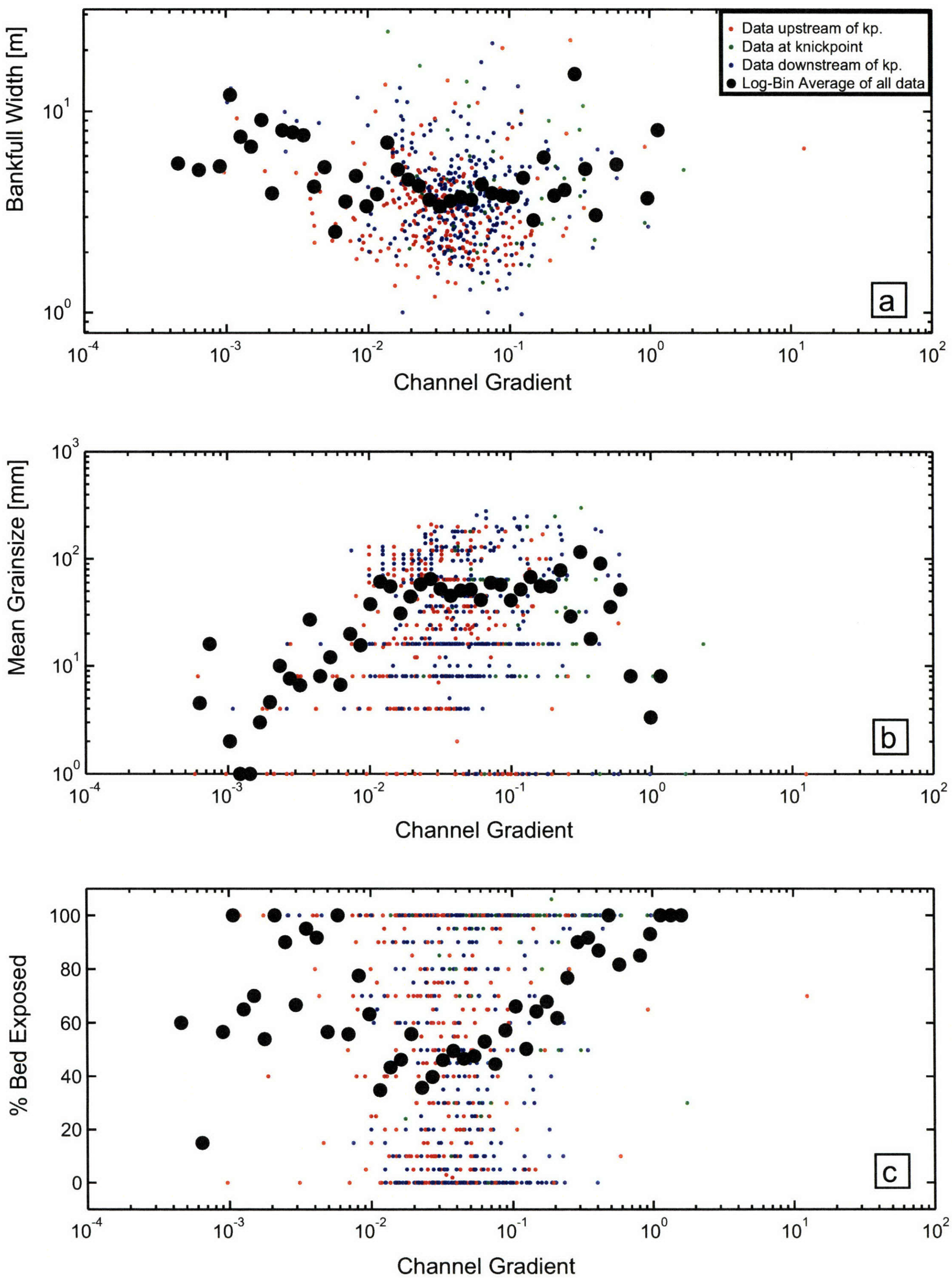

Gradient vs. Drainage Area
Channel Gradient
Drainage Area [m²]
a
Bankfull Width vs. Drainage Area
Bankfull Width [m]
b
Mean Grainsize vs. Drainage Area
Mean Grainsize [D50, mm]
c
% Bed Exposed vs. Drainage Area
% Bed Exposed
d
Legend
upstream mean
downstream mean
Channel Index
1 3 5 6 7 8 10 11 12 13 14 17 20 21 22 23 24 25 26 27 28 29 30 31 32 33 34

Figure 7



Data upstream of kp.
Data at knickpoint
Data downstream of kp.
Log-Bin Average of all data
Bankfull Width [m]
Channel Gradient
a
Mean Grainsize [mm]
Channel Gradient
b
% Bed Exposed
Channel Gradient
c

Knickpoint Initiation at Tributary Junctions:

a field-based comparison of trunk and tributary response to incision

Benjamin T. Crosby[1,2]

Kelin X Whipple[1,3]

1 Department of Earth, Atmospheric and Planetary Sciences, Massachusetts Institute of Technology, Cambridge, Massachusetts 02139

2 Now at: Department of Geosciences, Idaho State University, Pocatello, ID

3 Now at: School of Earth and Space Exploration, Arizona State University, Tempe, AZ

Primary Contact: Benjamin Crosby: crosbenj@isu.edu

Abstract

The distribution of knickpoints within fluvial basins has been explained previously as either the headward extent of an upstream-migrating incision signal or simply a consequence of where within the network thresholds in incision processes inhibit upstream propagation of the signal. In order to differentiate between these two models for knickpoint distribution, we executed a field survey of a trunk stream and 15 tributaries to the upper Waihuka Stream, a sub-catchment of the Waipaoa River on the North Island of New Zealand. In each tributary, a distinct knickpoint designates the present upstream extent of a transient pulse of incision that has lowered the mainstem Waihuka on the order of 40 meters in the last 18,000 years. The study catchment is an ideal location for this study because is relatively small (6.4×10^7 m^2 and ~17 km long), the drainage areas of the tributaries are similar (~1×10^6 m^2), substrate lithologies are relatively uniform, and the main stem contains a well exposed in record of progressive local base level lowering. Field observations reveal that the long profile forms of affected tributaries are a consequence the combined influence of downstream-sourced external forces such as the local base level fall history and upstream sourced internal forces such as local water and sediment discharge across the knickpoint. We anticipated that the range of internal and external influences on knickpoint form would be relatively small. To constrain the base level signal experienced at each tributary, we surveyed well-exposed and abundant terrace surfaces along the entire length of the mainstem Waihuka. The amount of vertical bedrock incision varies along stream due to along-stream variation in substrate erodibility and lateral inputs of coarse sediment from tributaries. Surveys of the tributaries reveal three basic knickpoint forms: (1) those hung directly at the tributary-mainstem junction; (2) those retreated upstream retaining their single-step form and (3) those that are composed of multiple smaller steps over an extended distance. We also found that hanging valley knickpoints are correlated with limited lateral incision in the mainstem and that multi-step knickpoints are correlated with the largest drainage area tributaries. Given the record of progressive mainstem lowering through time and the high frequency of hanging knickpoints, we suggest that knickpoints can form in place at tributary junctions regardless of whether a knickpoint existed in the mainstem. Also, variations in knickpoint form simply reflect diverse local internal and external influences active at each tributary junction. For the majority of cases, the diversity of internal and external influences inhibits the emergence of a deterministic knickpoint form.

1. Introduction

Records of bedrock channel incision are often utilized to infer recent changes in tectonic or climatic conditions. During transient adjustment to a pulse of channel incision, landforms such as strath or bedrock terraces and knickpoints help delineate the boundary between the downstream portion of the landscape that has begun to respond to the incision signal from the upper portion of the landscape that retains its relict, unadjusted form. When carefully analyzed, interpretations of strath terrace heights, extents, compositions and slopes can provide insight into the rates and patterns of the incision event, as well as provide some information regarding hydraulic and sediment supply conditions prior to the incision event (Merritts et al., 1994). The form and distribution of knickpoints within fluvial networks has been less successfully exploited in inferring past conditions because, unlike terraces, they are mobile erosional features that leave little record of previous location or form as they migrate upstream. Though less useful than terraces for studying temporal histories, knickpoints do offer the unique benefit of delineating the present upstream extent of the incision signal and their propagation is often considered the primary mechanism by which the incision signal extends throughout the basin, thus setting the fluvial portion of the basin's response time following disturbance (Bishop et al., 2005; Crosby and Whipple, in press; Harbor et al., 2005; Weissel and Seidl, 1998). Paired observations of modern, mainstem knickpoints and the fluvial terraces created by the upstream passage of that knickpoint have been frequently used as evidence for channel incision by knickpoint retreat (Garcia et al., 2004; Reneau, 2000; Reusser et al., 2004; Seidl and Dietrich, 1992; Stock and Montgomery, 1999; Zaprowski et al., 2001). The two major weaknesses in existing studies using terraces and knickpoints to study bedrock incision are (1), few acknowledge the along-stream variation in fill deposit thickness above the strath terrace surface and (2), most focus on the response of the mainstem channel and do not evaluate the communication of the incision signal into the tributaries. In this work, using data from the Waihuka tributary of the Waipaoa River in the North Island of New Zealand (Figure 1), we suggest that knickpoints need not have existed in the mainstem for knickpoints to be observed in tributaries just upstream of their mainstem junction..

1.1 Fill Thicknesses on Strath Terraces

Variations in alluvial fill thickness above the strath terrace surface are often difficult to quantify and thus are seldom documented. The tread, or upper surface of the overlying fill deposit, is the final surface occupied and shaped by the channel prior to abandonment by incision. Because of the accessibility and visibility of the tread surface, is often the one measured and correlated downstream using methods such as field surveying (GPS or traditional techniques), aerial photograph analysis or measurement using topographic data sets such as paper contour maps, or digital elevation data (via contour interpolation or LIDAR/SAR measurement). Elevations of the terrace tread surface are often plotted above the fluvial stream profile as a means of evaluating variation in incision along stream. Although it is widely recognized that care must be taken to distinguish between incision into bedrock and incision into alluvial fill, the causes and effects of varying fill thicknesses along stream profiles has not been studied in detail. Fill thickness is difficult to quantify and can be measured using three techniques: (1), physical measurement in outcrops where the terrace has been incised by trunk or tributary streams, (2) by geophysical methods such ground-penetrating radar or (3), by digging pits or using drilling techniques. Issues with limited outcrop exposure, equipment mobility and time expense frequently prevent quantification of fill thickness and limit studies of terraces to tread elevation only. If aggradation precedes incision, the assumption of constant fill thickness may be in error because the aggradational surface is easily modified by erosion post-abandonment. Channel slopes (dictated by local sediment flux, water discharge and base level conditions) at the time of strath development may be significantly different from those established during aggradation. If fill thickness varies systematically along-stream, then bedrock incision may also vary along stream. In this study we collected 106 local measurements of fill thickness exposed along the mainstem to constrain along-stream variation in bedrock incision (Figure 2).

1.2 Variability in Knickpoint Form and Function

The along-stream variation in bedrock incision rate within the trunk stream dictates the local base level fall signal experienced at each tributary junction. During a pulse of incision, as the trunk stream lowers, the incision signal is passed affects the tributary channels. If the base level fall in the mainstem is too rapid or the sediment flux from the tributary is to slow to respond to the base level fall, then it is possible for the tributary junction to oversteepen to the point that the knickpoint cannot viably retreat upstream by fluvial processes, thus stagnating the propagation of the incision signal and forming a hanging tributary (Crosby et al., submitted; Wobus et al., in press). Between the extremes of base level fall so slow that the tributary never forms a knickpoint and the case where hanging valleys are created, we hypothesize that the subsequent form and evolution of each mobile tributary knickpoint is a consequence of internal (within the tributary) and external (downstream of the tributary) influences. For two tributaries with identical internal and external influences, their knickpoints may share similar attributes. But in the more likely case where the local base level fall signal, the tributary drainage area or the degree of lateral migration of the mainstem differ between the two tributaries, the knickpoints will likely vary in form and function. This variability in knickpoint form is documented in a another study (Crosby and Whipple, in preparation) where morphological and bedstate parameters describing 27 channels that contain knickpoint are compared. We seek to understand whether the variability can be segregated into knickpoint 'types' that share relatively similar internal and external influences or whether the knickpoints and their influences are each unique.

In this study, we document variability in knickpoint form in 15 tributaries of the upper Waihuka Stream, a large tributary (6.4×10^7 m^2 and ~17 km long) to the Waipaoa River on the North Island of New Zealand (Figure 2). The Waipaoa River catchment provides an excellent location for studies of knickpoint distribution and form because it is actively responding to a pulse of incision initiated ~ 18,000 years ago (Berryman et al., 2000; Eden et al., 2001) and some 236 knickpoints consequent to that incision have been mapped and characterized (Crosby and Whipple, in press). By focusing on tributaries within a single sub-basin of the Waipaoa we attempt to reduce the variability in the base level fall signal that would be present in an all-basin analysis. The Waihuka sub-basin also is the site of a focused study using terraces and abandoned meanders to constrain the 18,000 year old incision history within a 2.2 km stretch of the Waihuka (Berryman et al., in preparation). By studying knickpoints in 15 tributaries to a single trunk stream with a well constrained incision history and well exposed mainstem fluvial terraces, we have been able to identify and characterize the internal and external influences the create the observed variability in knickpoint form.

This project utilizes a focused, experimental-style approach, attempting to control variables through careful site selection and collection of pertinent data only. In January of 2004, we completed one ~2 km survey of a tributary (# 10, Figure 2) and visually inspected 4 others in preparation for the following field season. During two weeks of work in February, 2005, we surveyed ~17 kilometers of the mainstem Waihuka and another 9 kilometers in 14 tributaries. Collected data was compiled onto a master longitudinal along-stream profile including all mainstem, tributary and terrace data (Figure 2). This study provides the analysis and interpretation of the mainstem incision history and its influence on the observed from of knickpoints in the surveyed tributaries. In the proceeding section we present the geologic setting of the field area, a discussion regarding the initiation of the pulse of incision 18,000 years ago and the methods used to collect and process the data.

2 Field area

2.1 Tectonic and Geologic Setting

The Waipaoa River catchment, located in the Northeast corner of the North Island of New Zealand, is etched into the accretionary sediments of the Hikurangi subduction zone. In this subduction zone the west dipping Pacific Plate is subducting under a continental fragment of the Australian plate, producing active arc magmatism in the Taupo Volcanic Zone which runs parallel to the trench, ~140 km east of the Waipaoa River catchment. The Pacific Plate in this region is characterized by abundant, large seamounts that destabilize the accretionary wedge, producing large mass failures of the continental slope (Collot et al., 2001; Lewis and Skinner, 1997). It is argued that seamount subduction and underplating of the accretionary wedge provide two viable mechanisms for the unsteady, non-uniform uplift histories recorded along the east coast of the North Island (Litchfield et al., submitted).

Though there are few observed active faults within the Waipaoa catchment, this may reflect the generally poor exposure and rapid erosional modification of landforms. There is frequent seismicity and anecdotal evidence for surface displacements on the order of half a meter in some locations. It is unclear whether these observations are consequence of surface-rupturing faults or scarps of large deep seated landslides in this characteristically unstable landscape. Bedrock units in the Waipaoa are dominated by Cretaceous to Pliocene accretionary wedge sediments (Black, 1980; Mazengarb and Speden, 2000). Though limestones, conglomerates and shelly clastic beds are locally observed, the most pervasive lithology in the Waihuka catchment (Figure 3) is the early Miocene Tolaga Group (eMt) which is composed of clay rich mudstones and siltstones interbedded with infrequent planar sandstone layers (Mazengarb and Speden, 2000). Other units observed in the study area include (in decreasing aerial extent): a more erosion resistant sandstone unit within the Tolaga Group (eMta); a softer calcareous mudstone known as the Weber Formation within the Mangatu Group (Ogw); a sandstone unit called the Wanstead Formation within the Mangatu Group (Egwg); and a marine sandstone within the Tolaga Group (Emts). Extensive vegetation coverage limits the certainty of the mapped contacts and many are estimated using landform characterization alone (personal communication, Mazengarb, 2000). During our field surveys, it was very difficult to discern differences between many of the rock types, though channels sourced in the eMTa did contain a larger percentage of coarse, resilient sandstone cobbles and boulders. The Otoko-Totangi normal fault in the eastern portion of the study catchment as well as the south vergent thrust in the southern portion of the catchment are not considered active (Mazengarb and Speden, 2000). We assume that these faults have been inactive since the initiation of incision because there is no observable deflection or deformation of terrace treads or DEM-collected stream profiles where they intersect the mapped faults. Though these local faults are apparently inactive, recent compilations of marine and fluvial terrace datasets suggest that the Waipaoa is experiencing rock uplift rates around 1 mm/year (Berryman, 1993; Berryman et al., 2000; Litchfield and Berryman, 2006; Ota et al., 1988).

2.2 Aggradation during the Last Glacial Maximum and the Subsequent Initiation of Incision

During the last glacial maximum (LGM), prior to the initiation of incision ~ 18,000 years ago, the Waipaoa River and all other major east-draining catchments on the North Island of New Zealand experienced network-wide aggradation (Litchfield and Berryman, 2005). This valley-filling event persisted between ~30ka and 18ka, burying bedrock-floored trunk streams and constructing alluvial fans at tributary junctions that extended upstream into these sub-basins. Throughout the Waipaoa catchment, we observe fill thicknesses between 5 and 45 meters, and within the study area fill thicknesses range between 3 and 20 meters. We will refer to this basin-wide aggradation surface as the "Waipaoa 1" (W1) terrace as defined in Berryman et al. (2000). We use this terrace to define the initial topographic condition prior to the initiation of the pulse of incision. The aggradation is attributed to colder and dryer climatic conditions (McGlone, 2001, 2002; McLea, 1990) that resulted in less vegetation to provide cohesion on unstable hillslopes and less water discharge to transport the sediment supplied from local hillslopes.

The timing of the initiation of incision is constrained by characterizing the stratigraphic accumulation of undisturbed tephra deposits on the abandoned W1 terrace surface (Berryman et al., 2000; Eden et al., 2001). The eruptive age of the lowest (or oldest) tephra establishes when stream incision into the aggraded sediments abandoned the W1 alluvial surface. Throughout the Waipaoa River catchment, the first tephra observed above the alluvial terrace deposits is the Rerewhakaaitu, erupted from the Okataina Volcanic Center ~17.6 ka (Lowe et al., 1999). The next two oldest tephras, the Okareka (21ka) and the Kawakawa (26.6ka) (Lowe et al., 1999) are only observed within active fluvial deposits in the W1 terrace, suggesting that the river channel had not yet incised as was continuing to deposit sediment (Berryman et al., in preparation; Eden et al., 2001). As this brackets the time of initiation of incision between 21 ka and 17.6 ka, we elect to refer to the initiation of incision event as occurring ~18 ka.

It is not certain why incision initiated at ~18 ka in the Waipaoa River catchment, but it was likely triggered by changes in climate at the end of the LGM and enhanced by a change in rock uplift rate. A detailed comparison of three proposed mechanisms are elaborated upon in a another publication (Crosby and Whipple, in preparation) and will be briefly summarized here. First, previous workers proposed that a warming and moistening climate decreased sediment supply (increasing vegetation density, and thus increasing hillslope stability) and increased the rivers' transport capacity (increased discharge) (Berryman et al., 2000; Eden et al., 2001; Litchfield and Berryman, 2005). Second, we have also suggest that during the LGM sea level dropped below the shelf-slope break, potentially initiating the transmission of a of a rapid base level fall signal through the channel network (Crosby and Whipple, in preparation, in press). The third potential driver for incision is related to subduction dynamics and attributes the large magnitude of the incision signal to either seamount subduction or underplating within the accretionary wedge (Litchfield et al., submitted). Any combination of these three mechanisms could be working in concert to create the observed incision response.

In the study presented here, the important information is not why incision initiated, but how incision progressed through time. Downstream of the Waihuka study catchment, near its confluence with the Waipaoa River, Eden et al. (2001) recognized that the Rerewhakaaitu was the oldest tephra preserved both on the top of a W1 terrace and on another terrace 15 m lower than the W1. This supports the expectation that that the initial pulse of incision through the aggraded alluvium was rapid, as anticipated. Recent work by Berryman et al. (in preparation) examines a high resolution record of river incision within our study reach in the Waihuka tributary (Figure 2), taking advantage of organic and tephra rich deposits preserved in a tiered sequence of terraces and cut-off meander loops along 2.1 km of the mainstem. Their efforts constrain the rate of progressive incision history in the Waihuka and provide local and temporal context for our examination of the tributaries' response to base level fall.

2.3 Current Morphology of the Waipaoa and Waihuka Catchments

Some 236 knickpoints define the fluvial boundary between the relict and adjusting portions of the Waipaoa River catchment (Crosby and Whipple, in press). Though more than 50% of the aerial extent of the basin retains its relict form, most knickpoints are located just upstream of tributary junctions and at small drainage areas (~1×10^6 m^2). These knickpoints can be generally described as being composed of three characteristic reaches (Crosby and Whipple, in preparation). The downstream reach is a steep, near-planar channel where the change in elevations is mostly accomplished along the bed of the channel but there are infrequent bedrock steps and shallow pools. The change in elevation in the knickpoint reach is largely accomplished over extremely steep steps that grow in height as the main head wall is approached. At the main headwall, the knickpoint is typically a near-vertical waterfall 10-30 m high without a plunge pool or undercut face. At a few knickpoints, evidence of seepage at the headcut was observed but it was not along any consistent permeability contrast and did not appear to discretely drive the failure of the headcut face. The dominant mechanism for knickpoint retreat does not appear to be fluvially driven, but rather weathering-mediated block failure at the knickpoint face. At and above the knickpoint lip, there is no evidence of discrete notching. We do find though that the in the 500 m reach upstream of the knickpoint the mean slope gradually increases toward the lip. The lack of discrete notching may be a consequence of the rapid retreat of the knickpoint face which would remove notches before they would have time to fully develop (Haviv et al., in press). Of all three knickpoint reaches, the reach upstream of the knickpoint is the lowest gradient and is typically incised through the W1 fill deposit and only shallowly into the underlying bedrock. Though the above description describes the mean knickpoint surveyed in the Waipaoa, there is considerable variability in that form. Understanding the source of that observed variability is one goal of this paper.

Extensive terraces are observed in every major tributary and alluvial fans extend up into some of the smaller tributaries. Terraces trace upstream and can be found to terminate at or slightly above knickpoint headwalls. Terrace deposits were well exposed, offering frequent opportunities to observe sediment character and measure both the height of the strath above the modern channel and fill thickness. The pre-abandonment alluvial deposits show no systematic upward fining or coarsening but rather alternate frequently between coarse and fine deposits up to the contact with the post-abandonment cover deposits. All alluvial terrace deposits are well-rounded and clast-supported with no evidence for debris flow or mudflow deposition.

The Waihuka tributary reflects many of the attributes observed throughout the entire Waipaoa River catchment. Abundant knickpoints of variable morphology are found in each tributary and well exposed terraces are abundant along the entire length of our study reach. The mainstem Waihuka is a bedrock channel with only infrequent transient patches of alluvium. At the downstream end of our ~17 km study reach, the drainage area is 6.4×10^7 m^2 and elevations within the study basin range between ~110 m above sea level (asl) and 922m asl.

3. Data Collection Methods

We focus on the Waihuka stream because of the excellent record of base level fall at the downstream end of our study reach (Berryman et al., in preparation) and because the landscape reflects many of the transient features observed throughout the rest of the Waipaoa River catchment. Reconnaissance planning for field surveys was accomplished using a DEM with 25m posting produced by Landcare Research NZ, a 1:50,000 topographic map produced by Land Information NZ, a 1:250,000 geologic map (Mazengarb and Speden, 2000) and color stereo aerial photographs flow in 1998 by Air-Maps NZ. Using stream profiles extracted from the DEM, we were able to identify the height and coarse character of the knickpoints and recognize that many were hung at or near the tributaries' junctions with the mainstem. Aerial photos helped us identify which terraces were associated with the W1 aggradation surface and which tributaries contained alluvial fans.

Streams were surveyed using a compass and inclinometer-equipped laser range finder attached to a GIS enabled hand held computer. When possible, horizontal positions of survey points were collected using an integrated consumer-grade GPS (±10m accuracy). As each survey point was collected, it was added to the database and plotted on the digital map in real time. The in-house survey software allowed the user to use the laser range finder to measure not only channel profile and width data, but also to measure the strath and tread elevations above the mainstem. All survey data was collected with elevations and downstream distances relative to the upstream extent of our mainstem survey (Figure 2). No absolute elevation data was collected using a GPS, but all relative differences in elevation in tributary and mainstem surveys were accurate when compared with elevation differences measured off the topographic map.

4. Results

4.1 Mainstem Waihuka Survey

Our survey of the mainstem Waihika begins above a knickpoint at an elevation of ~400 m and a drainage area of 4.85 x 10^6 m^2 and continues downstream for ~17,000 m to an elevation of ~100 m and a drainage area of 6.4 x 10^7 m^2. Measurements of the channel profile (distance, elevation, slope), channel width, percent bed exposure, upstream drainage area, terrace strath and tread height and tributary location were collected where available. We divide our presentation and interpretation of this data into three sections: mainstem morphology, mainstem longitudinal profile and terrace profile data

4.1.1 Mainstem Morphology

The mainstem Waihuka is characterized as a bedrock channel with intermittent bed coverage by thin sheets of sediment. There is no systematic change in percent bed exposure in the downstream direction (Figure 5c), and even in the reaches of lowest bed exposure, the sediment on the bed was not thick or coarse enough to remain immobile during moderate discharge events. The channel is trapezoidal shaped with a flat bottom and distinct banks. Valley width is typically wider than the bankfull width and there are both low, active floodplains and low terraces with straths surfaces typically < 5 m above the modern channel bed (Figure 2, T3 terraces). Bankfull width increases systematically in the downstream direction (Figure 5b) and the increase in width with drainage area can be well described by the power law function, $W = k_w A^c$, where W is the bankfull width, k_w is a dimensional constant, A is the upstream drainage area and c is a non-dimensional constant. For the Waihuka, k_w is equal to 0.0115 ± 0.796 and c is equal to 0.373 ± 0.0470 with a R^2 value of 0.645 (Figure 6d). Though the mainstem Waihuka study reach drainage areas ranges only a little over one order of magnitude, the power-law fit is very similar to that calculated from tributaries that contain knickpoints throughout the Waipaoa (Crosby and Whipple, in preparation) as well as that calculated for bedrock and gravel bed rivers from around the world (e.g. Hack, 1957; Montgomery and Gran, 2001; Snyder et al., 2003; Whipple, 2004).

The modern channel is moderately sinuous with a sinuosity of 1.45 over the entire study reach. Upstream of the main convex inflection in the profile at ~6,000 m the sinuosity is 1.36 and downstream of this inflection the sinuosity is 1.50. Downstream of the inflection in the modern profile we observe multiple cut-off meander loops between the W1 surface and the modern channel. Also, in numerous places (Figure 4) we saw semi-continuous minor strath levels or slip-off slopes spanning most of the total incision, proving that the 15-30 m high knickpoints found in the tributaries did not sweep up the mainstem – they had to form in place. These observations suggest that during the time of incision, the channel maintained a meandering form. We were not able to confirm whether any of the abandoned meanders were incised into bedrock or whether they were inset just within the W1 aggradational deposits. Strath terraces certainly provide evidence for progressive bedrock incision. The observations of relict meanders, slip-off slopes and the significantly wide valley bottoms in some stretches of the Waihuka suggest that incision was not purely vertical but also included a significant lateral component. This lateral mobility of the mainstem has important implications on the local base level fall experienced at each tributary junction. If the mainstem is both incising it's bed and laterally eroding toward the tributary junction, the tributary experiences an greater rate of base level fall. If the mainstem is eroding laterally away from the tributary junction, the base level fall signal experienced by the tributary is diminished.

4.1.2 Mainstem Longitudinal Profile

The longitudinal along-stream profile of the Waihuka has three distinct convexities. The uppermost is at the upstream limit of our survey where gradient decreases significantly upstream of the W1 knickpoint. Upstream of this point, the channel is incised ~8 m into fan deposits that merge with the W1 alluvial gravels downstream. Of those 8 meters, ~4 are within the underlying bedrock. Another small convexity at ~700 meters downstream is a consequence of coarse material provided by a large tributary. At 2500m another large tributary enters from the south. Downstream of both junctions, we recognize a significant increase in the mean grainsize, suggesting that in this bedrock channel, the coarse armoring of the bed by the tributary's sediment inhibits erosion, thus increasing the local slope (e.g. Howard, 1998). The most prominent inflection in the modern profile of the Waihuka is at ~6000 m downstream of the survey start and is also related to a tributary junction. In this case, the Parihohonu tributary (# 26), entering the channel from the south, constructed a large alluvial fan during the aggradational period, shifting the mainstem Waihuka laterally against its northern bank. Unlike the inflections upstream, which we attribute to localized armoring by coarse sediment, we suggest that the steepening observed near the Parihohonu fan is a consequence of the northward displacement the Waihuka relative to its paleochannel (Figure 4). In this location, when the aggradation in the Waihuka ended and incision into the W1 surface initiated, the mainstem channel had to incise entirely into bedrock, rather than incising into the unconsolidated alluvial deposits filling its paleochannel. Further evidence for this process and discussion of this phenomenon elsewhere along the Waihuka are discussed further in section 3.1.3.

Our survey data also allow us to examine the power law relation between slope and drainage area, $S = k_s A^{-\theta}$, where S is the local channel gradient, A is the upstream drainage area, k_s is the channel steepness index and θ is the channel concavity index. Analyzing the entire study reach, we find $k_s = 98.85 \pm 1.84$ and $\theta = 0.517 \pm 0.109$ (a pretty common concavity index in Waipaoa) with a R^2 value of 0.194. Because of the pronounced inflection in the profile at ~6000 m, we also analyzed reaches upstream and downstream of this inflection to determine whether they were significantly different. For the upper reach, we find $k_s = 271{,}437.26 \pm 4.55$ and $\theta = 1.013 \pm 0.281$ with a R^2 value of 0.241. Downstream, we find $k_s = 1386.72 \pm 4.43$ and $\theta = 0.664 \pm 0.254$ with a R^2 value of 0.118.

4.1.3 Terrace Profile

W1 terrace profiles from the Waihuka mainstem reveal along-valley variation in fill thickness and bedrock incision. In this section we present observations of that variation and discuss sources of that variability. Terrace data is analyzed along a valley center-line profile in order to remove the influence of the modern channel's sinuous form (Figure 4). Tread and strath elevation data surveyed along that sinuous course was projected perpendicularly from the observation location to the valley center-line (Figure 7).

Post-W1 bedrock incision and fill thicknesses are significantly different upstream and downstream of the inflection in the modern profile at the Parihohonu Fan (sourced from tributary #26). Upstream of the fan, total bedrock incision ranges between 0 and 20 meters and the fill thickness maintains a relatively stable value of ~5 meters (Figures 8b and 8c). At the upstream end of this upper reach of the Waihuka, terrace treads increase rapidly in elevation with the increasing slope of the channel. Some of this behavior can be attributed to the rapidly decreasing drainage area, but there is also a pronounced narrowing in the valley at this location. Fan deposits sourced from the headwaters of the mainstem Waihuka (tributary #34) line the banks of this extremely steep reach, producing the two highest terrace treads at ~500 down-valley in Figure 7. Downstream of this steep, fan-dominated reach, W1 terrace fill thickness become extremely regular. Though the down-valley terrace profile upstream of the Parihohonu Fan is slightly concave, the concavity of the modern channel profile is significantly higher. This results in a systematic increase then decrease in bedrock incision into the W1 strath surface in the down-valley direction (Figure 8b). Downstream of the main inflection in the profile, bedrock incision increases downstream with a few minor high incision rate locations where the channel could not reoccupy its paleochannel and instead had to incise entirely through bedrock. Just upstream of the fan, between 3750 and 4250 m down-valley, the valley widens significantly and there are numerous terrace surfaces. In this reach, we propose that during the time of aggradation and after the climatically driven increase in transport capacity, the growing height of the Parihohonu fan limited the passage of sediment from upstream, creating a depositional region upstream of the fan. At the upstream end of the Parihohonu fan (4500 m down-valley), the channel is flowing over bedrock, but the only fan deposits are observed in the banks (up to 22 meters thick) suggesting that no bedrock incision has occurred here yet. The main inflection in the modern profile thus defines the upstream extent of any upstream migrating incision signal. The incision upstream of the main inflection suggests that a process internal to the upper basin (not base level fall driven), such as a climate-driven change in transport capacity of sediment supply, was partially responsible for the observed upper-basin incision.

In the reach downstream of the Parihohonu fan and the inflection in the modern profile, the measured bedrock incision increases over a down-valley distance of ~ 2000 m from 0 m to a relatively steady 30 m of incision (Figure 8b). During our survey we discovered multiple explanations for down-valley variability in measured bedrock incision. As introduced above, tributary fans and other laterally sourced deposits can shift the position of a channel away from the position of its paleochannel. This is particularly important in aggraded channels where there is little lateral confinement. When incision began again, the channel either remobilized the aggraded sediment and reestablished its position in the paleochannel or shifted laterally from its previous course and subsequently incised a new channel in bedrock. Variations in fill thickness and bedrock incision below the W1 surface in the Waihuka mainstem reveal that this phenomenon was not uncommon. The two best examples are at the meanders created by the Parihohonu Fan (Tributary 26) and near the mainstem confluence with Tributary 25 (Figure 4). The incised meanders and tributary valleys provide a three-dimensional visualization of where the modern mainstem exits and rejoins the paleochannel. In the reaches that incise purely through bedrock, there are little or no fill deposits and the bedrock extends to the height of the W1 surface. In both cases, tributaries 26 and 25 enter the mainstem on the inside edge of the meander bend and dog-leg abruptly downstream before joining the mainstem (Figure 4). We find that this tributary behavior is also a consequence of tributaries reoccupying the mainstem paleochannels before joining the modern mainstem. In these two cases, the reoccupation has consequences on the observed knickpoint morphology as discussed ahead. Another explanation for the variation in observed bedrock incision is that modern or late-stage lateral incision can remove the record of the maximum fill thickness in the paleochannel. This results in overestimation of the total bedrock incision and an underestimation of the fill thickness.

To quantify the fill thicknesses measured along the Waihuka, we subtract the terrace tread elevation from the strath elevation. Because of the significant terrace heights above the mainstem Waipaoa, we were not always able to confirm a strath and a terrace tread elevation at each survey station. These unpaired measurements are not accounted for in the points plotted in Figure 8c. To work around this problem, we interpolated, by hand, the along stream variation in fill thickness using the unpaired measurements of strath and W1 fill heights shown in Figure 7. We find that between 5500 and 11000 meters down-valley the fill thickness has an average value of ~20 meters but increases slightly in the downstream direction. The variability in fill thickness is a consequence of multiple processes including the lateral displacement of the mainstem away from its paleochannel or the deposition of alluvial fans at tributary junctions with the mainstem.

We also collected tread and strath elevation data regarding two intermediate terraces between the W1 surface and the present river bed (Figure 4). The upper intermediate terrace (Figure 7, green marks) is infrequently observed along stream and is often found only where meander bends preserve a sequence of slip-off slopes. At these locations there are many intermediate strath terraces confirming the progressive incision of the mainstem and refuting the possibility that the a large, discrete knickpoint could have passed up the mainstem. Though more careful study of the chronology of these isolated surfaces could constrain incision rates along the trunk stream, the many surfaces do not appear to be correlative in the downstream direction and may reflect incision events initiated by local controls such as landslides or meander cutoffs.

The lower intermediate terrace is of consistent height and observed frequently along stream (Figure 7, yellow marks). The strath surface is typically 3-5 meters above the modern channel bed and capped with another 2-3 meters of alluvial fill. The lower intermediate terrace is observed at 42 locations over the entire length of the study reach and unlike the W1 or upper intermediate terrace, strath and fill heights were almost always well exposed. The consistency in the elevations of this terrace above and below the major inflection in the mainstem profile (at ~4500m) suggests that this most recent episode of incision was evenly distributed along the length of the study reach. This inference assumes that the observed lower intermediate terraces are of the same age.

4.2 Waihuka Tributary Surveys

We surveyed 15 tributaries to the Waihuka from their mainstem junction to upstream of the knickpoint in order to discern if there are robust explanations for the observed variability in longitudinal profiles. Compared to measurements of grainsize distribution, percent bed exposure or channel width, longitudinal profiles are the metric least susceptible to modification by modern land use practices. For this reason, we use the longitudinal profile to group the knickpoints into three 'types': (1) hanging valley knickpoints; (2) discrete, retreating knickpoints and (3) convex and multi-step knickpoints. For each tributary, we characterize the internal (tributary relief, drainage area, substrate properties) and external (mainstem base level fall rate, magnitude of mainstem bedrock incision, degree of lateral mainstem erosion) influences on the knickpoints form. We also discuss the relation between the observed knickpoints and the elevation of the strath and the thickness of the W1 fill at tributary junctions.

4.2.1 Type 1: Hanging Valley Knickpoints

We define hanging valley knickpoints as those whose longitudinal profile rises abruptly from the confluence and remains steep and step-dominated until cresting the upper lip of the knickpoint and entering the reach graded either to the W1 surface or the pre-aggradation channel elevation (approximately the height of the strath in the mainstem). Tributaries 20, 22, 24 and 30 contain hanging valley type knickpoints (Figure 9). We find that the projected profile of the upstream, relict reach intersects the mainstem around or just below the elevation of the local strath height. This suggests that the relict reach has experienced minor incision and retains a profile similar to its pre-aggradation form. Visual inspection and the steep, step dominated form of the knickpoints suggests that fluvial incision processes are not active, but that mass failure, weathering or seepage mediated block failure probably drives modification of the knickpoint form.

We were surprised to find that all the hanging valley knickpoints in the Waihuka join the mainstem in locations where there is very little evidence for lateral erosion during the last 18,000 years. This was surprising because if a channel enters the stream on the outside of a meander bend, the progressive lateral incision of the channel will oversteepen the tributary at its outlet. This leads us to suggest that low lateral mobility (especially if the lateral erosion moves the mainstem away from the tributary confluence) may favor the formation of hanging valley knickpoints. Two knickpoints are downstream of the Parihohonu fan (22, 24), one is in the bedrock, incised reach adjacent to the fan (30) and one (20) is significantly upstream of the fan (see Figure 2). Because the fill thickness and total bedrock incision in the mainstem is different at each tributary confluence and the tributaries are well spaced throughout the basin, we have little evidence to suggest that the rate of base level fall was similar in any of the channels.

The four hanging valley knickpoints have differing amounts of fluvial relief (junction to tip of blueline) and are not unique from the other knickpoint types (220 m , 201 m, 260 m and 328 m for Tributaries 20, 22, 30 and 24 respectively). All hanging valley tributaries are completely within eMt and only two (20 and 22) have significant remnant accumulations of W1 sediment in the valley upstream of the knickpoint. Drainage areas at the junction for the four channels are below the mean (2.60×10^6 m^2) of the 15 tributaries (1.05×10^6 m^2, 1.10×10^6 m^2, 1.33×10^6 m^2, 1.33×10^6 m^2 for Tributaries 20, 22, 30 and 24 respectively), but very close to the median (1.3×10^6 m^2).

4.2.2 Type 2: Discrete, Retreating Knickpoints

We define discrete, retreating knickpoints as those whose longitudinal profile contains three separable reaches: a downstream reach that graded to the mainstem and is not dominated by large, abrupt steps; a knickpoint reach composed of a single large step or discrete sequence of near vertical steps that comprise most of the total elevation gain; and the upstream reach, characterized as a low gradient channel graded either to the mainstem W1 surface or the pre-aggradation channel elevation (mainstem strath height). Tributaries 23, 28 and 29 contain discrete, retreating knickpoints (Tributary 10 may also belong in this group, but it also fits with Type 3) (Figure 10). We find that a downstream projection of the upper portion of the profile intersects the mainstem both above (Tributary 28) and below (Tributaries 23 and 29) the elevation of the local strath height. Again, the steep, step dominated form of the knickpoints seen in the hanging valley knickpoints suggests that fluvial incision processes are not active, but that weathering or seepage mediated block failure drives modification of the knickpoint form.

Compared to the hanging valley knickpoints, the major distinguishing external characteristic is that the tributary confluences of the retreating knickpoints occur in mainstem reaches where there has been significantly more lateral movement of the mainstem. We observe that the Type 2 knickpoints have retreated back relative to valley wall (Figure 4) and thus do not suggest that the extended reach downstream of the knickpoint was created by the progressive movement of the mainstem away from the tributary confluence. We instead suggest that the increased lateral mobility of the mainstem may be indicative of a slower or more episodic incision rate, thus providing more time for the tributary to partially respond to the small increments of lowering. This episodic or prolonged base level fall forcing is similar to that suggested for the convex or multi-step knickpoints, but it is unclear why in the discrete, retreating knickpoints the abrupt face of the knickpoint is retained.

Internal influences on knickpoint form fail to distinguishing between Type 1 and Type 2 knickpoints. Fluvial relief is comparable with the other knickpoint types (250 m , 292 m and 265 m for Tributaries 23, 28 and 29 respectively) and the drainage areas are much more variable than in the hanging valley knickpoints (1.12 x 10^6 m^2, 3.49 x 10^6 m^2 and 0.452 x 10^6 m^2) but still have a similar mean that closely matches the mode of all 15 channels.

Comparing the internal influences of substrate and geologic structures we find no obvious correlations. Tributary 23 taps an extensive relict fan deposit but the others flows entirely within the eMt unit with little of the local coarse sediment supply available in Tributary 23. The relict surface in Tributary 28 is anomalously high and is adjacent to the mapped Otoko-Totangi normal fault. We cannot attribute the high relict surface to movement along the fault for two reasons. First, the fault does not displace the W1 surface. If the relict basin owes its height to displacement along the fault, that displacement predates the formation of the W1 aggradational surface. Second, though both the orientation and displacement direction are uncertain, Mazengarb and Speden (2000) suggest movement along the fault down to the east. If this suggested geometry is correct, then we would expect the relict surface in Tributary 28 to be abnormally low, not high. We suggest instead that the high elevation of the relict surface is because Tributary 28 runs through the more erosion resistant, sandstone-dominated eMta lithology. The high relict surface in that tributary may pre-date the W1 aggradation, but further tephrachronology work on that upper relict surface could confirm this suggestion.

The network form and position of Tributary 29 suggests that stream piracy has divided two branches of a channel that drained a large fan on the north side of the Waihuka. It is unclear whether the reach we surveyed was the primary outlet for the former catchment. There is also some evidence that the extended low gradient reach at the downstream end of Tributary may be due to lateral migration of the mainstem away from the tributary junction.

4.2.3 Type 3: Convex and Multi-Step Knickpoints

We define convex or multi-step knickpoints as those whose longitudinal profile is composed of an extended, oversteepened downstream reach that contains some minor steps but the steps do not comprise a significant amount of the fluvial relief downstream of the relict reach. Tributaries 27, 10, 26, 33, 25, 21, 31 and 32 contain convex or multi-step knickpoints (Figures 11 and 12). This large group of knickpoints contains the most diverse collection of observed internal and external influences and cannot be fairly discussed as a single group. Instead, known influences will be highlighted in the streams that are affected.

Tributaries 25 and 26 enter the Waihuka on the inside bank of large meanders that have been previously discussed as locations where the mainstem channel was shifted laterally out of its pre-aggradation paleochannel. Both tributaries take abrupt right (east) bends before joining the mainstem (Figure 4). We find that this morphology is a consequence of the tributary channel establishing its course within in the more easily eroded pre-aggradation mainstem channel. By assuming the mainstem's old course, the length of each tributary increases, potentially decreasing the magnitude of the base level fall signal experienced by the tributary. The lower reaches of these two channels are composed of multiple discrete steps that do not culminate at a large W1 knickpoint (Figures 11c and 12a). We suggest that the large steps in the upper reaches of the channels are exhumed and recently modified older, pre-W1 features. These pre-W1 knickpoint features are observed in other tributaries such as 20 and 21 (Figures 9a and 9b). Both Tributary 25 and 26 have large drainage areas and large fluvial relief (2.71×10^6 m^2 and 366 m; 9.91×10^6 m^2 and 580 m respectively) and Tributary 26 also taps into the harder, more resistant lithology of the eMta. These internal influences could have also affected the development of the observed stepped knickpoint profile.

Because Tributary 21 enters the mainstem Waihuka on the inside of a meander bend, the tributary has lengthened and presently flows over the slip-off slopes of the meander before joining the mainstem. The profile, composed of three distinct steps (Figure 12b), is likely the consequence of its position on inside edge of a mainstem meander rather than internal influences (drainage area = 1.30×10^6 m^2, fluvial relief = 288, all within the dominant eMt lithlology). The fan observed extending southward out of Tributary 21 and its interaction with the opposing Parihohonu fan may have also played a role in the development of the meander that ultimately dictate the stepped knickpoint profile. In summary, Tributaries 25, 26 and 21 owe their from to lateral erosion of the trunk stream away from the tributaries' mainstem junctions.

Tributaries 27 and 10 (Figure 11a and 11b) contain extremely linear downstream and upstream reaches, separated by an extended knickpoint reach. Channel banks and adjacent hillslopes of the downstream reaches of both channels are highly unstable. Pods of clay-rich mélange identified in field may be below the resolution of the geologic map (Figure 3) but were observed in both bed and banks of these tributaries, downstream of the steepest upper reaches. The mélange is soft and deforms in a ductile manner as observed in other regions of the Waipaoa River catchment where the material is mapped as a clay-dominated mélange. While the upstream reach of Tributary 10 projects well above the mainstem strath terrace elevation, the upstream reach of Tributary 27 projects well below the mainstem strath elevation. Both Tributary 27 and 10 have identical upper-catchment geology (Figure 3) and similar drainage areas and fluvial relief (6.51×10^6 m^2 and 520 m; 6.13×10^6 m^2 and 427 m respectively) and their mainstem confluences are separated by only ~1.5 kilometers.

The knickpoint form in Tributary 33 also is highly dependent on the substrate properties. A ridge-scale mass movement sourced off the western side of the lower catchment significantly altered the form of the valley and the channel profile. Like in Tributaries 27 and 10, the landslide affected reach (between ~100 and 500m downstream in

Figure 11d) is close to linear and the substrate is extremely soft clay with little structural rigidity. All three tributaries are mapped as within the eMt lithology, suggesting that the weak character of the substrate is not an intrinsic attribute of the unit, but rather a consequence of near surface deformation. Like in Tributary10, the upstream reach of Tributary 33 projects to the top of the W1 fill deposit. In the steeper portion of the downstream reach, coherent bedrock is observed beneath the landslide material.

Tributaries 31 and 32 join together just upstream of their confluence with the mainstem Waihuka (Figure 4). Both profiles share the small step just upstream of the confluence. This step is around the same elevation of the low strath surfaces observed throughout the catchment (Figure 12c and 12d). Both profiles also have a higher knickpoint at roughly the same elevation as the strath terrace in the mainstem Waihuka. Though projections of the upstream reach to the confluence with the Waihuka are well below the strath surface, we attribute this to the downstream jog in the channel that extends the distance between the knickpoint and the mainstem (similar to if the channel was highly sinuous). These knickpoints could also have been grouped under the retreating knickpoints type, but because of their multi-step form I discuss them here. Tributaries 31 and 32 are completely within the eMt unit, have similar relief to the channels in both the hanging and retreated knickpoint group (320 m and 250 m, respectively) and share the small same drainage area at the confluence (0.653 x 10^6 m^2).

The multiple internal and external influences that create multi-step knickpoints suggest that this knickpoint form has no single common genetic influence, but rather reflects the multiple local conditions that either disturb the channel bed or complicate the junction of the mainstem and tributary. The main influences are: mainstem mobility, pre-existing bedrock topography in the mainstem/variable fill thickness and local lithologic effects (substrates susceptible to mass movement).

5. Discussion

The above observations and analysis of tributary profile form in the Waihuka catchment provides ample evidence to discuss which factors, internal and external to the tributary, exert the greatest influence on knickpoint form.

5.1 Aggradation Provides Greatest External Influence on Tributary Response

Our initial intention in this study was to select a discrete sub-catchment with a well constrained mainstem incision history so that we could interpret the variability in knickpoint profile form simply as a function of internal influences. Our results suggest that even though Berryman et al. (in preparation) provide an excellent time series of post-W1 incision in the lower reach of our study area, their time series may not adequately capture the variability in incision history occurring further upstream. The attributes of base level fall that we expect to exert influence on tributary response include the rate of base level fall (fast or slow lowering, the influence of lateral erosion), the pattern of base level fall (steady or episodic lowering), the magnitude of the base level fall signal (paleochannel position and fill thickness influence on total bedrock incison) and the consistency of base level fall (multiple cut-and-fill cycles or steady lowering). In the Waihuka, we attribute the along-stream variability in local base level fall history to three consequences of the aggradation that preceded the incision into the W1 surface.

First, aggradation and the construction of alluvial fans at tributary mouths supported lateral displacement of the mainstem channel away from the filled paleochannel so that when incision initiated ~18 ka, some reaches had to incise through bedrock while others lowered easily through unconsolidated alluvium. This variation in along-stream bed erodibility likely influenced the pace of base level fall experienced at local tributaries. In the locations where the channel must incise bedrock, the gorges are narrower and have been correlated with the location of hanging tributaries.

The second consequence of aggradation is the inherited legacy of sediment stored in alluvial fans, terraces and tributaries. The remobilization of these unconsolidated sediment sources provides both tools and, in excess, armor to the modern channel bed, potentially enhancing or inhibiting the channel's capacity to erode. Though sediment cannot be isolated as the only influence, where fans or tributaries demonstrably increase sediment delivery, we also find higher channel gradients, such as at the Parihohonu fan. Upstream of that fan, only a small portion of the base level fall signal is communicated into the upper reaches. Because the observed incision in the upper reaches exceeds the communicated base level fall signal, we suggest that some portion of that upper-basin incision is climatically driven and was temporarily facilitated by the availability of aggraded sediment filling the valleys.

The third consequence of aggradation is increased lateral mobility of the incising mainstem channel. Though the modern Waihuka is a bedrock floored channel, during aggradation and incision times, ample upstream sediment supply and alluvial substrate (in some locations) would favor high lateral mobility of the mainstem. This would help develop broad valleys and mobile meanders in some locations. As discussed above, lateral migration can significantly influence the base level fall signal at tributary junctions.

5.2 Internal Controls on Tributary Form.

Tributaries to the Waihuka are characterized by a relatively limited range of internal characteristics. Drainage areas range between 0.452×10^6 m^2 and 9.91×10^6 m^2 with a median value of 1.3×10^6 m^2. We find that smaller tributaries tend to have more discrete, single step knickpoints while 4 of the 5 largest tributaries have extended multi-step knickpoints. Because of the numerous competing influences on knickpoint from it is difficult to definitively conclude that knickpoint form is correlated with drainage area only. Because fluvial relief in tributaries scales with drainage area, we can make a similar statement that tributaries with greater fluvial relief tend to have less discrete knickpoint profile forms. Though the sample of tributary lithologies does not range over a wide spectrum of erodibilities, we found the tallest single-step knickpoint in the hardest lithology (Tributray 28 in eMta) and multi-stepped, diffused knickpoints in rocks weakened by mass movement. In tributaries where the lithology is more susceptible to mass movement, the knickpoint form can be erased or highly modified by failure of adjacent hillslopes. Sediment supply stored within tributary fans does not appear do dictate knickpoint profile form as there are examples of hanging valley knickpoints, knickpoint that retreated from the confluence and multi-step knickpoints that contain these fan features. What is likely more important is the frequency with which these deposits are eroded, remobilized and transported through the tributary. The erosion of fan deposits likely scales with the size of the tributary, suggesting that large drainage area tributaries may have multi-step knickpoints because of the combined sediment source and capacity to transport it through the channel. If increased sediment flux enhances abrasion and bedrock incision in knickpoint reaches where steep slopes prevent bed armoring by alluvial deposits, single step knickpoints may not be retained.

5.3 Implications for Future Studies of Landscape Response to Incision

Our analysis highlights how pre-aggradation paleotopography, variations in fill thickness and the lateral mobility of the mainstem channel all exert a significant influence on the tributary response to trunk stream incision. Localized observations of incision history along the mainstem may provide a high fidelity record of base level fall at a point, but are not necessarily applicable when applied to a larger network. At the larger scale, the relationship between trunk and tributary streams is variable in space and time as water and sediment fluxes intrinsically vary within the network and also in response to changes in vegetation and climate. We find that tributaries have influenced the long-term behavior of the mainstem through the formation of coarse alluvial fans and the lateral displacement of the mainstem away from its paleochannel, resulting in along-stream variation in bed erodibility. At the same time, the mainstem influences the development of knickpoints in tributaries by setting the rate and style of base level fall as well as controlling the lateral position of the tributary/trunk junction. This interaction between trunk and tributary increases the difficulty of discerning whether internal or external influences determine the form of knickpoints.

In this work, we test the applicability of a classification scheme for the profile forms of Waihuka knickpoints. Although the classification of knickpoints into three types elucidated some robust differences in the applicable internal and external influences, we more strongly suggest that each knickpoint uniquely reflects its own particular base level history and internal characteristics of substrate and sediment supply. We approached this discrete field site anticipating that we could utilize the near uniformity of the substrate, the similar tributary sizes and relief structure and the well exposed, straightforward record of incision to identify common knickpoint forms that reflect uniform forcing, but in looking closer we find further evidence for the lack of deterministic control on knickpoint form.

6. Conclusions

Variations in knickpoint form observed in 15 tributaries to the Waihuka Stream suggest that the complex interactions between the internal characteristics of the tributary and the external influence of the mainstem base level fall signal do not produce a well defined suite of knickpoint types. Instead, we find that knickpoints reflect the along-stream variability in the base level fall signal as recognized in the variations in fill thickness and strath elevation of terraces along the mainstem Waihuka. Though most knickpoints reflect this complex interaction between trunk and tributary we do find some consistent characteristics in the suggested knickpoint types. Tributaries with knickpoints hung just above the mainstem confluence only occur along reaches of the mainstem where there was very little lateral erosion. Also knickpoints composed of multiple, separate steps are observed in large drainage area tributaries where the mainstem contains a record of active lateral incision. Future examinations of the interaction between knickpoint form and records of mainstem incision will need to acknowledge the complicated feedback between tributary and trunk stream throughout the transmission of the transient signal

References

Berryman, K., 1993, Age, height, and deformation of Holocene marine terraces at Mahia Peninsula, Hikurangi subduction margin, New Zealand: Tectonics, v. 12, no. 6, p. 1347-1364.

Berryman, K., Marden, M., Eden, D., Mazengarb, C., Ota, Y., and Moriya, I., 2000, Tectonic and paleoclimatic significance of Quaternary river terraces of the Waipaoa River, east coast, North Island, New Zealand: New Zealand Journal of Geology and Geophysics, v. 43, no. 2, p. 229-245.

Berryman, K., Marden, M., Palmer, A. S., Wilson, K., Mazengarb, C., and Litchfield, N., in preparation, The post-glacial downcutting history in the Waihuka tributary of the Waipaoa River, Gisborne District, New Zealand, and implications for tectonics and landscape evolution.

Bishop, P., Hoey, T. B., Jansen, J. D., and Artza, I. L., 2005, Knickpoint recession rate and catchment area: the case of uplifted rivers in Eastern Scotland: Earth Surface Processes and Landforms, v. 30, p. 767-778.

Black, R. D., 1980, Upper Cretaceous and Tertiary geology of Mangatu State Forest, Raukumara Peninsula, New Zealand: New Zealand Journal of Geology and Geophysics, v. 23, no. 3, p. 293-312.

Collot, J., Lewis, K., Lamarche, G., and Lallemand, S., 2001, The giant Ruatoria debris avalanche on the northern Hikurangi margin, New Zealand; results of oblique seamount subduction: Journal of Geophysical Research, Solid Earth, v. 106, no. B9, p. 19,271-19,297.

Crosby, B. T., and Whipple, K. X., in preparation, Changes in Channel Morphology and Bed State in Response to River Incision -- Deterministic Trends or Inherent Variability? for submission to GSA Bulliten.

-, in press, Knickpoint Initiation and Distribution within Fluvial Networks: 236 waterfalls in the Waipaoa River, North Island, New Zealand: Geomorphology.

Crosby, B. T., Whipple, K. X., Gasparini, N. M., and Wobus, C. W., submitted, Formation of Fluvial Hanging Valleys: Theory and Simulation: Journal of Geophysical Research – Earth Surface.

Eden, D. N., Palmer, A. S., Cronin, S. J., Marden, M., and Berryman, K. R., 2001, Dating the culmination of river aggradation at the end of the last glaciation using distal tephra compositions, eastern North Island, New Zealand: Geomorphology, v. 38, p. 133-151.

Garcia, A. F., Zhu, Z., Ku, T. L., Chadwick, O. A., and Chacon Montero, J., 2004, An incision wave in the geologic record, Alpujarran Corridor, southern Spain (Almeria): Geomorphology, v. 60, no. 1-2, p. 37-72.

Hack, J. T., 1957, Studies of longitudinal stream profiles in Virginia and Maryland, U. S. Geological Survey Professional Paper 294-B, p. 45-97.

Harbor, D., Bacastow, A., Heath, A., and Rogers, J., 2005, Capturing variable knickpoint retreat in the Central Appalachians, USA: Geografia Fisica e Dinamica Quaternaria (Testo stampato), v. 28, no. 1, p. 23-36.

Haviv, I., Enzel, Y., Whipple, K. X., Zilberman, E., Stone, J., Matmon, A., and Fifield, K. L., in press, Amplified erosion above waterfalls and oversteepened bedrock reaches: Journal of Geophysical Research – Earth Surface.

Howard, A. D., 1998, Long profile development of bedrock channels: Interaction of weathering, mass wasting, bed erosion, and sediment transport, *in* Tinkler, K., and Wohl, E. E., eds., Rivers Over Rock: Fluvial Processes in Bedrock Channels: Washington, D. C., AGU Press, p. 297-319.

Lewis, K., and Skinner, D. N. B. e., 1997, The succession of seamount impacts and giant avalanches on the Hikurangi margin

Geological Society of New Zealand 1997 annual conference; programme and abstracts: Geological Society of New Zealand 1997 annual conference, v. 95A, p. 99.

Litchfield, N. J., and Berryman, K. R., 2005, Correlation of fluvial terraces within the Hikurangi Margin, New Zealand: implications of climate and baselevel controls: Geomorphology, v. 68, p. 291-313.

-, 2006, Relations between postglacial fluvial incision rates and uplift rates in the North Island, New Zealand: Journal of Geophysical Research – Earth Surface, v. 111, p. F02007, doi: 10.1029/2005JF000374.

Litchfield, N. J., Ellis, S., Berryman, K. R., and Nicol, A., submitted, Subduction related uplift in the Hikurangi Margin, New Zealand: Journal of Geophysical Research – Earth Surface.

Lowe, D. J., Newnham, R. M., and Ward, C. M., 1999, Stratigraphy and chronology of a 15 ka sequence of multi-sourced silicic tephras in a montane peat bog in eastern North Island, New Zealand: New Zealand Journal of Geology and Geophysics, v. 42, p. 565-579.

Mazengarb, C., and Speden, I. G., 2000, Geology of the Raukumara Area: Institute of Geological and Nuclear Sciences 1:250,000 geological map 6: Lower Hutt, New Zealand, Institute of Geological and Nuclear Sciences Limited, 60 p.

McGlone, M. S., 2001, A late Quaternary pollen record from marine core P69, southeastern North Island, New Zealand: New Zealand Journal of Geology and Geophysics, v. 44, no. 1, p. 69-77.

-, 2002, A Holocene and latest Pleistocene pollen record from Lake Poukawa, Hawke's Bay, New Zealand: Global and Planetary Change, v. 33, no. 3-4, p. 283-299.

McLea, W. L., 1990, Palynology of Pohehe Swamp, Northwest Wairarapa, New Zealand; a study of climatic and vegetation changes during the last 41,000 years: Journal of the Royal Society of New Zealand, v. 20, no. 2, p. 205-220.

Merritts, D. J., Vincent, K. R., and Wohl, E. E., 1994, Long river profiles, tectonism, and eustasy; a guide to interpreting fluvial terraces: Journal of Geophysical Research, B, Solid Earth, v. 99, no. 7, p. 14,031-14,050.

Montgomery, D. R., and Gran, K. B., 2001, Downstream variations in the width of bedrock channels: Water Resources Research, v. 37, no. 6, p. 1841-1846.

Ota, Y., Berryman, K. R., Hull, A. G., Miyauchi, T., Iso, N., Pirazzoli, P. A. e., and Scott, D. B. e., 1988, Age and height distribution of Holocene transgressive deposits in eastern North Island, New Zealand

Quaternary coastal changes: NATO Advanced Study Institutes Programme symposium on Late Quaternary sea-level correlation and applications | XIIth INQUA congress, v. 68, no. 2-4, p. 135-151.

Reneau, S. L., 2000, Stream incision and terrace development in Frijoles Canyon, Bandelier National Monument, New Mexico, and the influence of lithology and climate: Geomorphology, v. 32, no. 1-2, p. 171-193.

Reusser, L. J., Bierman, P. R., Pavich, M. J., Zen, E. a., Larsen, J., and Finkel, R., 2004, Rapid late Pleistocene incision of Atlantic passive-margin river gorges: Science, v. 305, no. 5683, p. 499-502.

Seidl, M. A., and Dietrich, W. E., 1992, The problem of channel erosion into bedrock: Catena Supplement: Functional geomorphology; Landform Analysis and Models, v. 23, p. 101-124.

Snyder, N. P., Whipple, K. X., Tucker, G. E., and Merritts, D. J., 2003, Channel response to tectonic forcing: field analysis of stream morphology and hydrology in the Mendocino triple junction region, northern California: Geomorphology, v. 53, no. 1-2, p. 97-127.

Stock, J. D., and Montgomery, D. R., 1999, Geologic constraints on bedrock river incision using the stream power law: Journal of Geophysical Research-Solid Earth, v. 104, no. B3, p. 4983-4993.

Weissel, J. K., and Seidl, M. A., 1998, Inland Propagation of Erosional Escarpments and River Profile Evolution Across the Southeast Australian Passive Continental Margin, *in* Tinkler, K. J., and Wohl, E. E., eds., Rivers Over Rock: Fluvial Processes in Bedrock Channels: Washington, D. C., American Geophysical Union, p. 189-206.

Whipple, K. X., 2004, Bedrock Rivers and the Geomorphology of Active Orogens: Annual Review of Earth and Planetary Sciences, v. 32, p. 151-185.

Wobus, C. W., Crosby, B. T., and Whipple, K. X., in press, Hanging valleys in fluvial systems: Controls on occurrence and implications for landscape evolution: Journal of Geophysical Research – Earth Surface.

Zaprowski, B. J., Evenson, E. B., Pazzaglia, F. J., and Epstein, J. B., 2001, Knickzone propagation in the Black Hills and northern High Plains: A different perspective on the late Cenozoic exhumation of the Laramide Rocky Mountains: Geology, v. 29, no. 6, p. 547-550.

Figures

Figure 1: Regional map showing the location of the Waipaoa Catchment in the Northeast corner of the North Island of New Zealand. The large central image shows the topography of the northern 2/3 of the Waipaoa catchment as well as the location of the Waihuka Stream and the 15 tributaries addressed in this study. The aqua region is the portion of the basin that has begun adjusting to the pulse of incision initiated ~18,000 years ago. The 236 knickpoints mapped within the Waipaoa River catchment demonstrate the fluvial extent of the transient incision signal.

Figure 2: Along-stream longitudinal profile showing survey data the mainstem Waihuka Stream and 15 tributaries. Note the convexity in the mainstem profile at ~ 5500 meters downstream. Knickpoint locations in the tributaries are marked with a red dot. Note that most knickpoints are near the elevation of the base of the W1 fill deposit. All tributary longitudinal profiles are rotated into the plane of the mainstem channel and plotted with the same vertical exaggeration as the mainstem. Map-view image in the upper right identifies individual channels and demonstrates the high density of survey stations (small blue dots) in the mainstem Waihuka and its tributaries. The three levels of terrace data collected along the mainstem Waihuka are plotted lightly only for reference in this figure and are addressed in Figure 7.

Figure 3: Geologic map of the Waihuka drainage basin [following (Mazengarb and Speden, 2000)]. Note that the mainstem is almost entirely within the eMt as are the majority of the tributaries. Note the high frequency of survey locations within mainstem and tributary channels. No structures are reported as active and there was no observed deflection of the mainstem channel or the 18 ka W1 terrace surface where the Otoko-Totangi fault crosses the channel.

Figure 4: This aerial photo montage shows the distribution of W1 terrace and fan surfaces as well as intermediate terraces between the W1 elevation and the modern channel. The large black arrows demonstrate how the two strips overlap. Tributary locations and center-line distances are also shown. Note the frequency of W1 fan deposits along the study reach and the large number of intermediate terraces in the reach studied by Berryman et al. (in preparation).

Figure 5: Survey data from the mainstem Waihuka presenting the along-stream longitudinal profile (a) and the elevation of the W1 terrace treads (+) and strath surfaces (.). Variations in bankfull width (b) and percent bed exposure (c) are also plotted relative to the along-stream distance. Note the steady increase in channel width and width variability in the downstream directions. Also note that though there is no coherent downstream trend in the percent bed exposure, the bed is partially exposed along most the surveyed length, suggesting that during high flow events most of the sediment is mobilized, exposing the bed to bed-load abrasion processes.

Figure 6: Variations in slope (c) and bankfull width (d) as a function of drainage area within the mainstem Waihuka. Profile and W1 terraces (a) and change in drainage area with distance downstream (b) are plotted for reference. Note that when the slope-area data is analyzed above and below the large inflection in the mainstem profile, both reaches

have significantly higher concavities, particularly in the upstream reach. It is worth noting that these relationships are well resolved even when using slightly more than an order of magnitude of change in drainage area.

Figure 7: Terrace data in the mainstem Waihuka plotted relative to the down-valley distance. This figure removes the influence of the sinuosity of the mainstem channel, plotting each terrace elevation relative to the valley center-line (shown as a red line on the network map in the upper right corner). Note the significant change in alluvial fill thickness and bedrock incision downstream of the large inflection in the mainstem profile at ~4500 m. Also note that though there is very little bedrock incision at the lip of the inflection, bedrock incision not only persists upstream but increases with distance upstream of the inflection. This persistence of incision above the limit of the base level fall signal suggests that the incision post-W1 aggradation is not purely base level fall driven but also represents a response to changing climate. The arrows that locate the 'displaced mainstem' show where the mainstem was shifted laterally northeastward away from its paleochannel and had to incise bedrock rather that reoccupy its old course. Note in those locations that terrace fill thickness decreases significantly.

Figure 8: Plot demonstrating the down-valley variation in post-W1 bedrock incision (b) and measured fill thickness (c). Again, we provide a plot of the modern profile and terrace elevations for reference (a). Note that bedrock incision upstream of the large inflection in the modern profile at ~4500 m increases in the upstream direction, reaching a maximum of ~20 meters of incision. This incision is not base level fall driven but is related to the increase in discharge and decrease in sediment supply following the last glacial maximum. The large magnitude of bedrock incision downstream of the inflection is attributed to the combined climatic and base level fall incision signal observed throughout the Waipaoa catchment. Because pairs of fill and strath elevation do not always represent the full fill thickness, we interpolate fill thickness using adjacent data and field observations (the dashed gray line in Figure 8c). This same interpolated fill thickness is used plot the fill region in Figure 7.

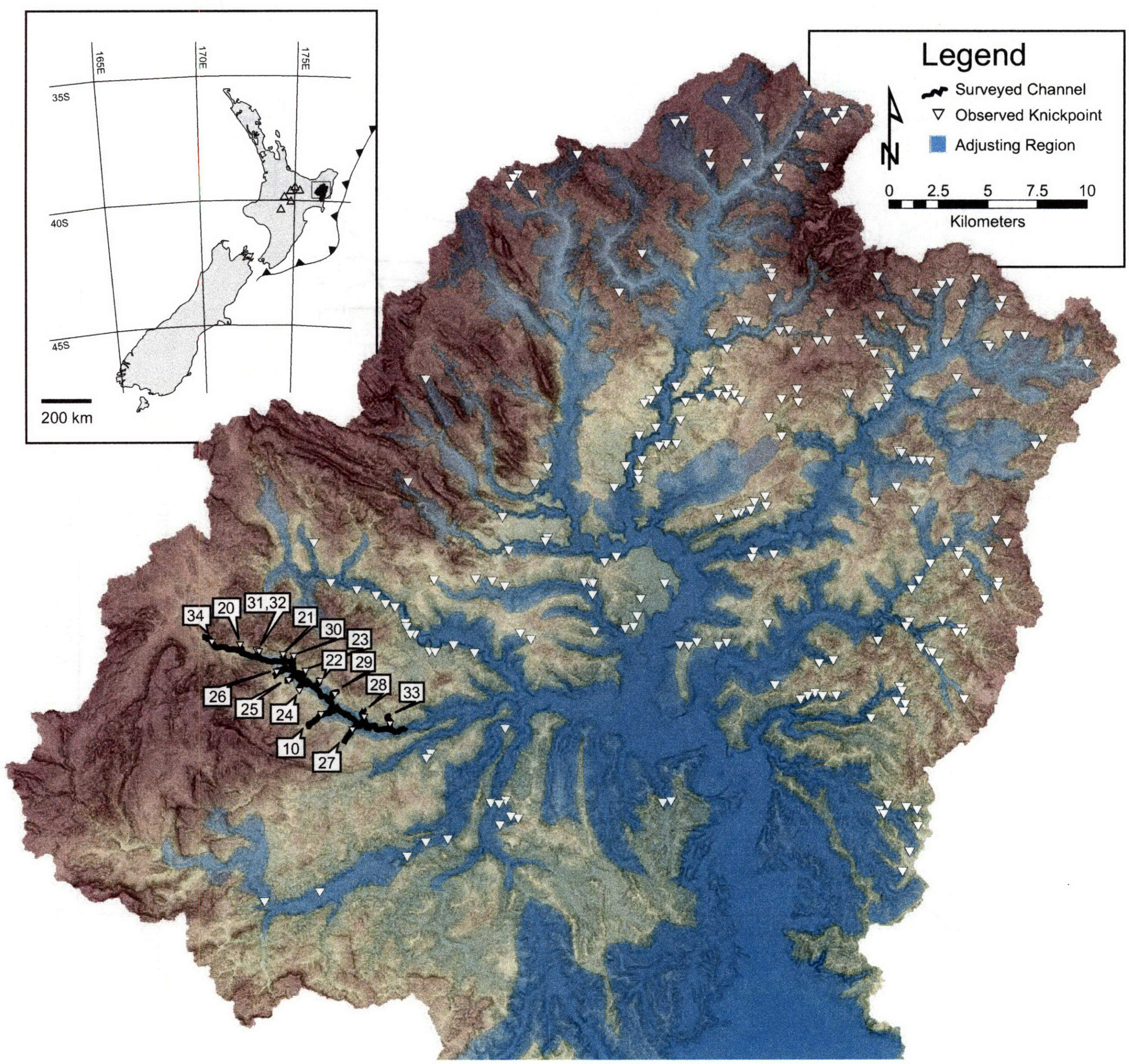
Figure 9: Tributary longitudinal profiles plotted relative to distance downstream of the hanging-valley-type knickpoint and elevation above the confluence. Note that all profiles are the same scale. W1 tread and strath elevations at the mainstem junction are plotted at the downstream end of the profile. Note that the reach upstream of the knickpoint projects close to the strath surface in the mainstem. The two upstream steps in the channel profile of tributaries 20 and 22 (9a and 9b respectively) to not appear to be related to incision into the W1 aggradation surface and likely predate that aggradation.

Figure 10: Tributary longitudinal profiles plotted relative to distance downstream of the discrete, retreating-type knickpoint and elevation above the confluence. Note that all profiles are the same scale. W1 tread and strath elevations at the junction with the mainstem are plotted at the downstream end of the profile. Though the length of the downstream, partially adjusted reach varies in each tributary, each terminate at a single, abrupt step in channel elevation.

Figure 11: Tributary longitudinal profiles plotted relative to distance downstream of a large magnitude convex or multi-step-type knickpoint and the elevation above the confluence. Note that all profiles are the same scale. W1 tread and strath elevations at the junction with the mainstem are plotted at the downstream end of the profile. Note the significant variability in knickpoint form between the four plotted profiles.

Figure 12: Tributary longitudinal profiles plotted relative to distance downstream of a small magnitude convex or multi-step-type knickpoint and the elevation above the confluence. Note that all profiles are the same scale. W1 tread and strath elevations at the junction with the mainstem are plotted at the downstream end of the profile. Note the significant variability in knickpoint form between the four plotted profiles.

Figure 1


Legend
Surveyed Channel
Observed Knickpoint
Adjusting Region
N
0 2.5 5 7.5 10
Kilometers
165E
170E
175E
35S
40S
45S
200 km
34
20
31,32
21
30
23
22
29
26
25
24
28
33
10
27

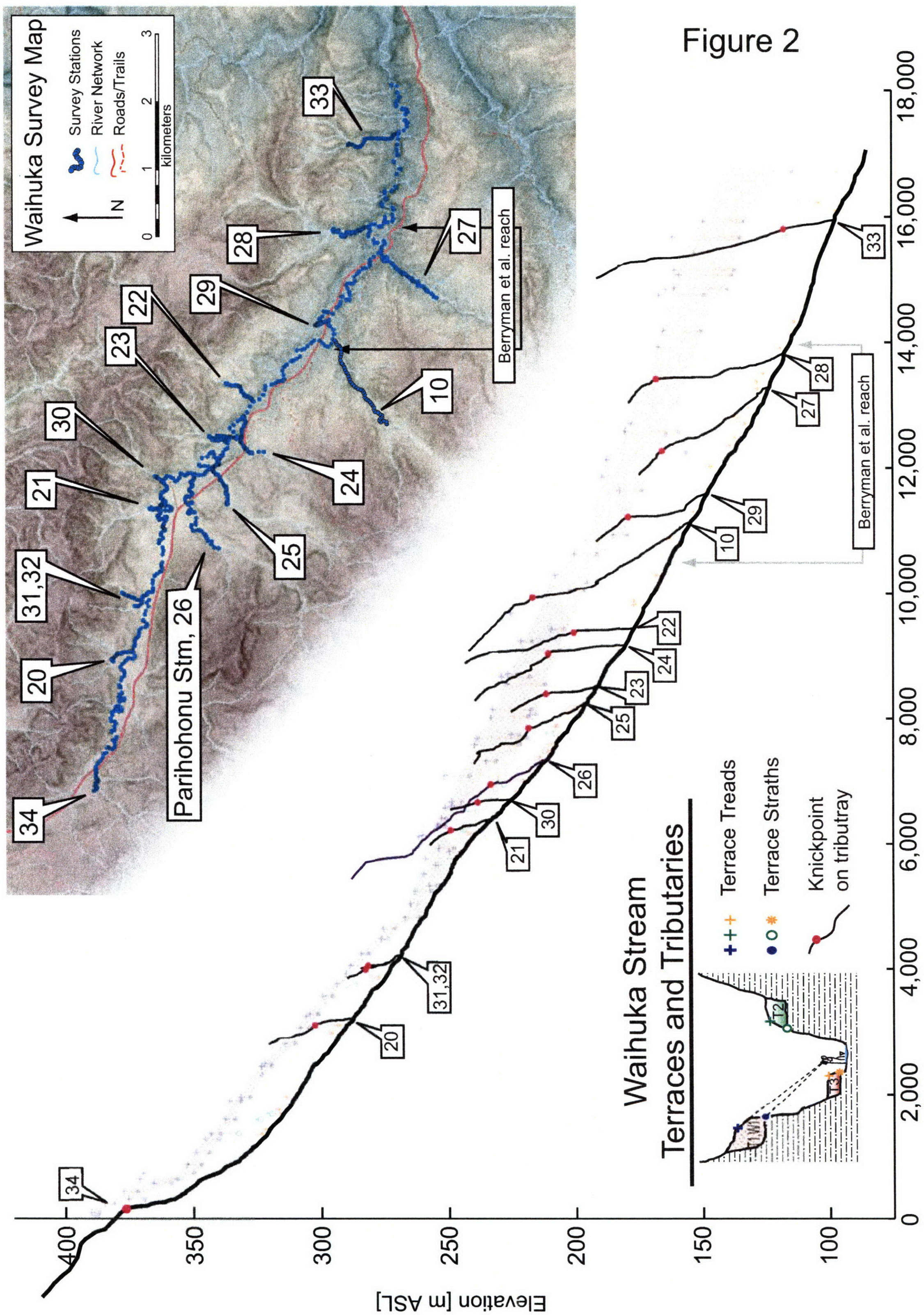
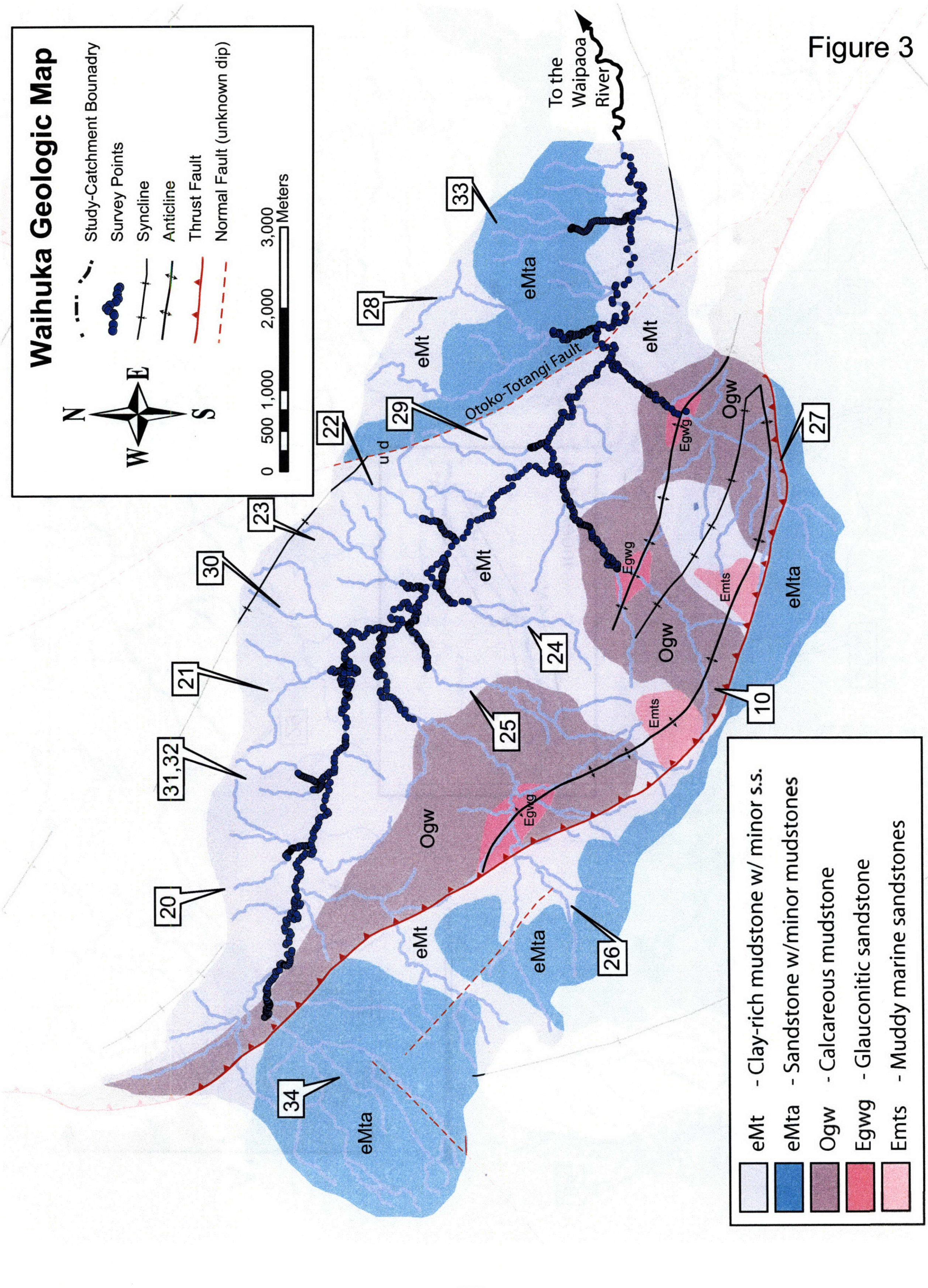

Waihuka Survey Map
N
Survey Stations
River Network
Roads/Trails
0 1 2 3
kilometers
Parihohonu Stm, 26
Berryman et al. reach
34
20
31,32
21
30
25
24
23
22
29
10
28
27
33
Waihuka Stream
Terraces and Tributaries
Terrace Treads
Terrace Straths
Knickpoint on tributray
T1
T2
T3
Elevation [m ASL]
Along-Stream Distance [m]
0
2,000
4,000
6,000
8,000
10,000
12,000
14,000
16,000
18,000
100
150
200
250
300
350
400

Figure 2


Waihuka Geologic Map
N E S W
Study-Catchment Bounadry
Survey Points
Syncline
Anticline
Thrust Fault
Normal Fault (unknown dip)
0 500 1,000 2,000 3,000 Meters
Figure 3
To the Waipaoa River
Otoko-Totangi Fault
u d
eMt
eMta
Ogw
Egwg
Emts
10
20
21
22
23
24
25
26
27
28
29
30
31,32
33
34
eMt - Clay-rich mudstone w/ minor s.s.
eMta - Sandstone w/minor mudstones
Ogw - Calcareous mudstone
Egwg - Glauconitic sandstone
Emts - Muddy marine sandstones

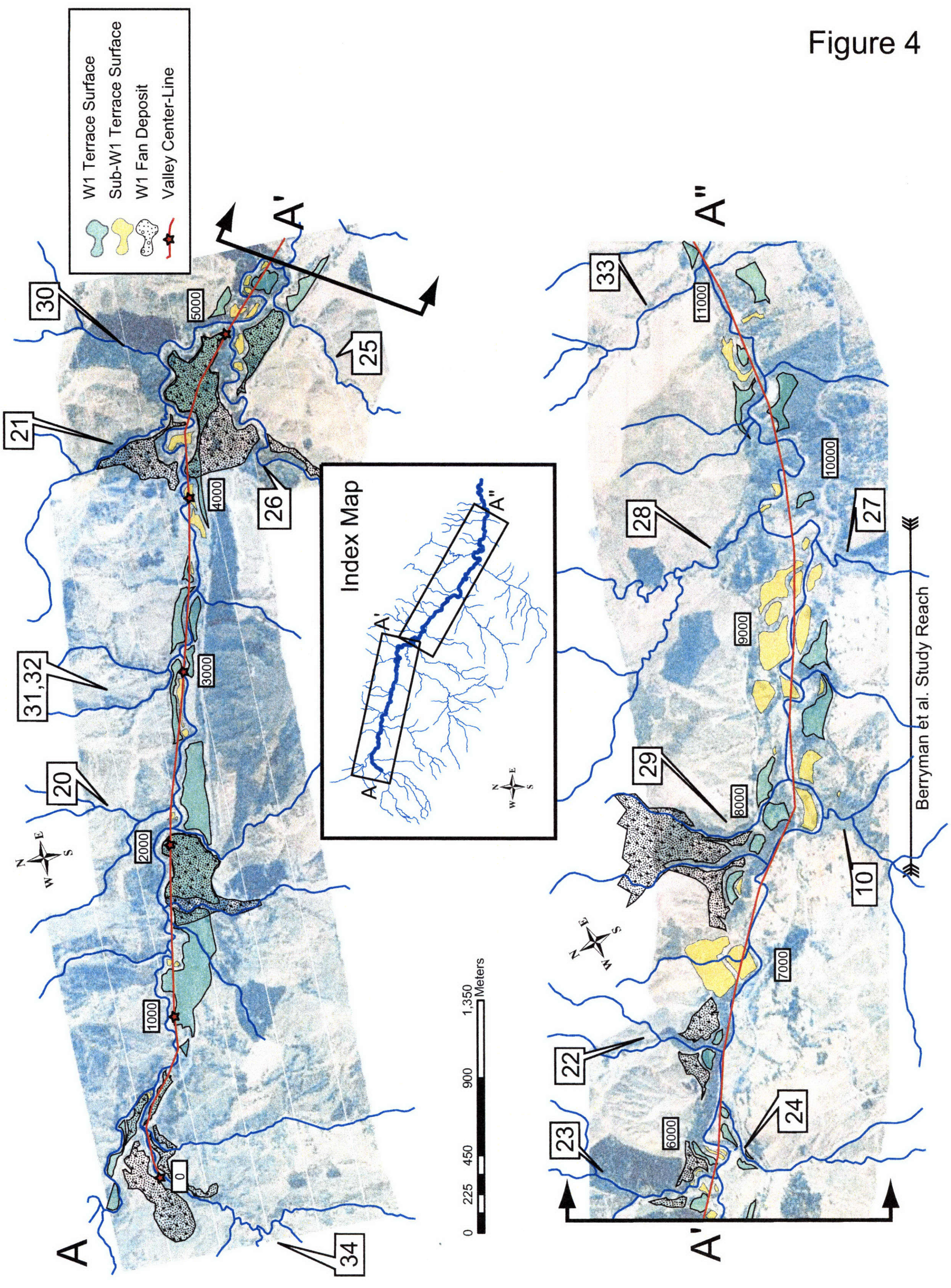

W1 Terrace Surface
Sub-W1 Terrace Surface
W1 Fan Deposit
Valley Center-Line
Index Map
A
A'
A"
Meters
0
225
450
900
1,350
Berryman et al. Study Reach
0
1000
2000
3000
4000
5000
6000
7000
8000
9000
10000
11000
34
20
31,32
21
30
26
25
23
24
22
29
10
28
27
33

Figure 4

Figure 5

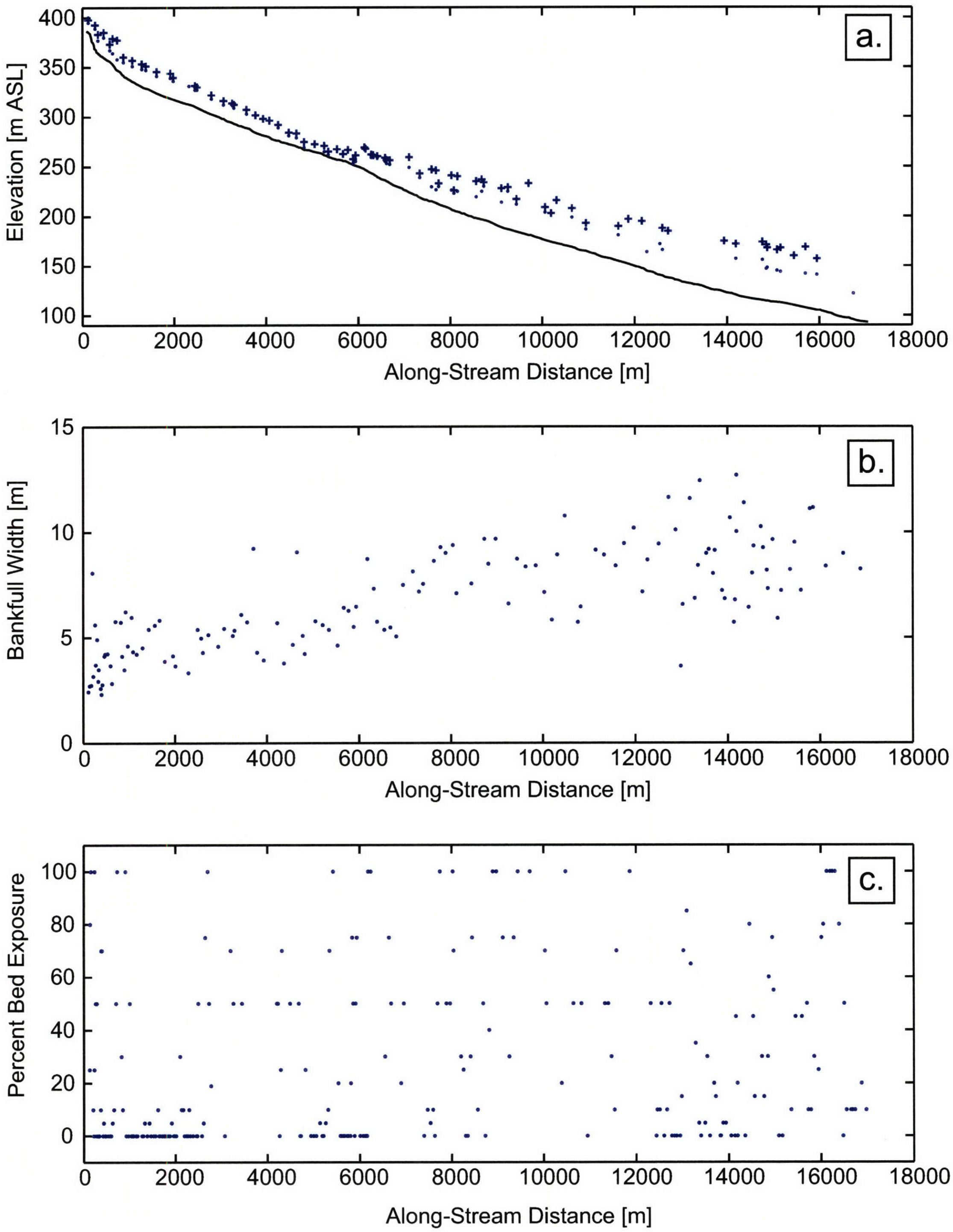

a.
Elevation [m ASL]
400
350
300
250
200
150
100
0
2000
4000
6000
8000
10000
12000
14000
16000
18000
Along-Stream Distance [m]
b.
Bankfull Width [m]
15
10
5
0
Along-Stream Distance [m]
c.
Percent Bed Exposure
100
80
60
40
20
0
Along-Stream Distance [m]

Figure 6

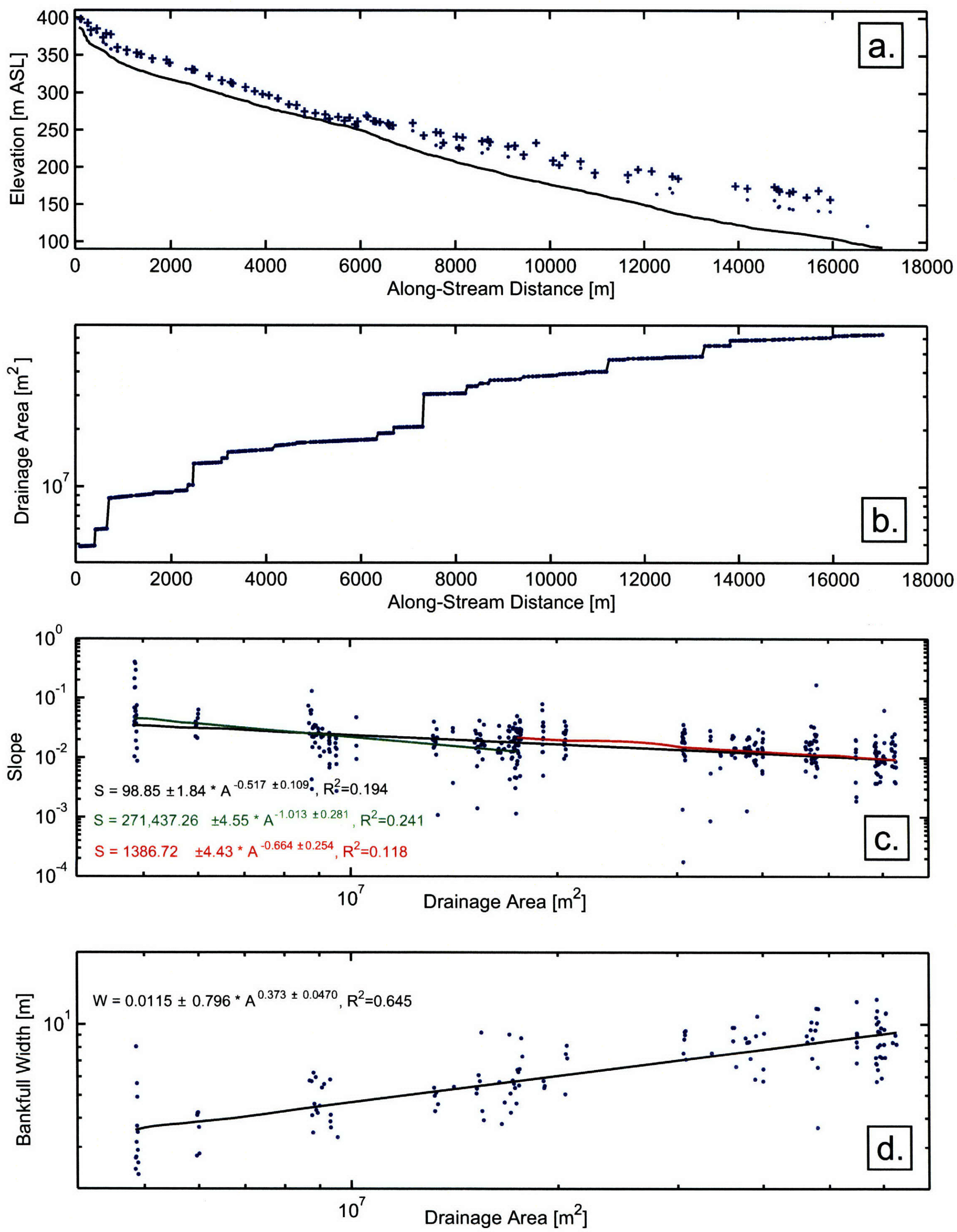

a.
b.
c.
d.
Elevation [m ASL]
Along-Stream Distance [m]
Drainage Area [m²]
Slope
Bankfull Width [m]
S = 98.85 ±1.84 * A^-0.517 ±0.109, R²=0.194
S = 271,437.26 ±4.55 * A^-1.013 ±0.281, R²=0.241
S = 1386.72 ±4.43 * A^-0.664 ±0.254, R²=0.118
W = 0.0115 ± 0.796 * A^0.373 ± 0.0470, R²=0.645

Figure 7

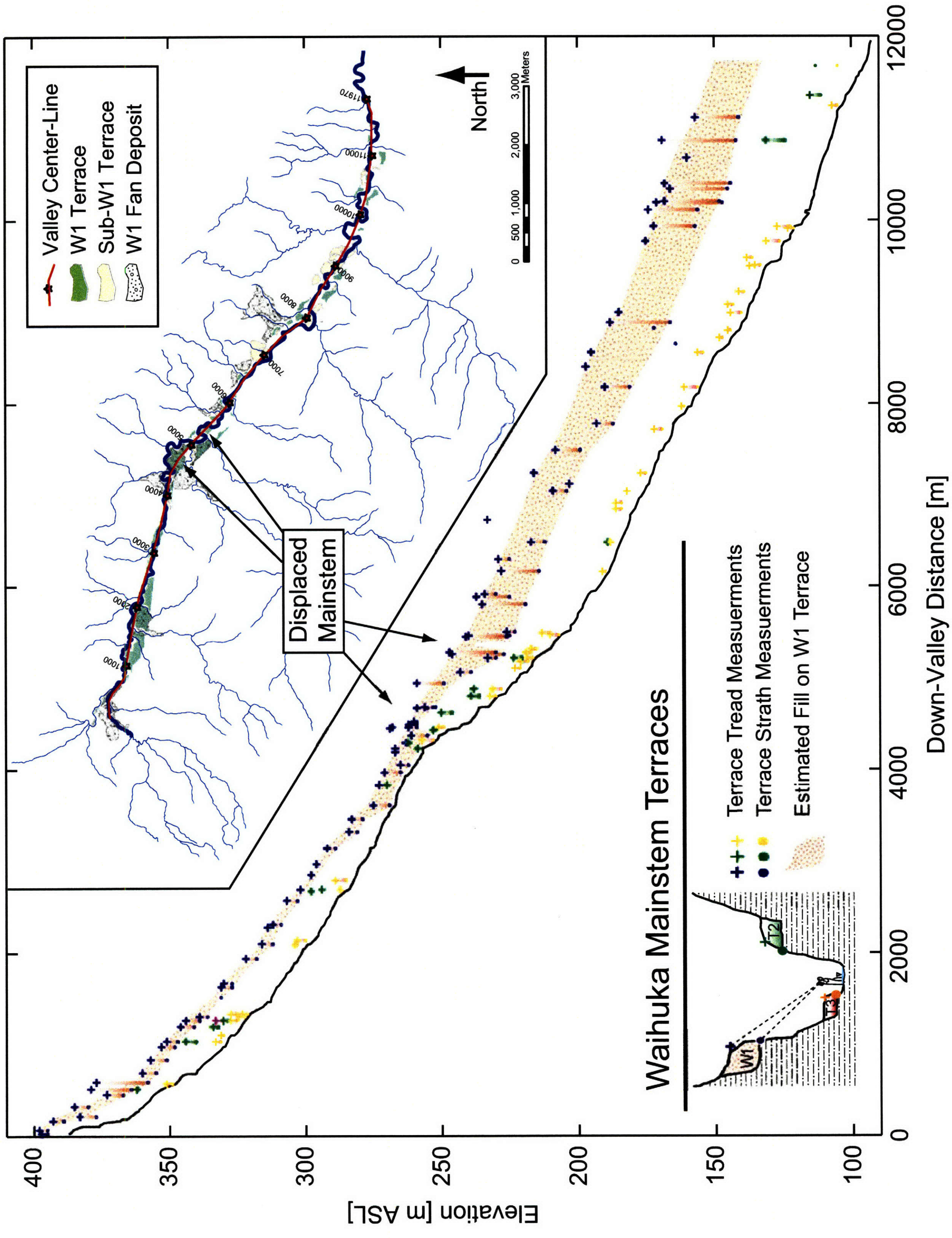

Waihuka Mainstem Terraces
Valley Center-Line
W1 Terrace
Sub-W1 Terrace
W1 Fan Deposit
North
0 500 1,000 2,000 3,000 Meters
Displaced Mainstem
Terrace Tread Measuerments
Terrace Strath Measuerments
Estimated Fill on W1 Terrace
W1
T2
T3
Elevation [m ASL]
Down-Valley Distance [m]
400
350
300
250
200
150
100
0
2000
4000
6000
8000
10000
12000

Figure 8

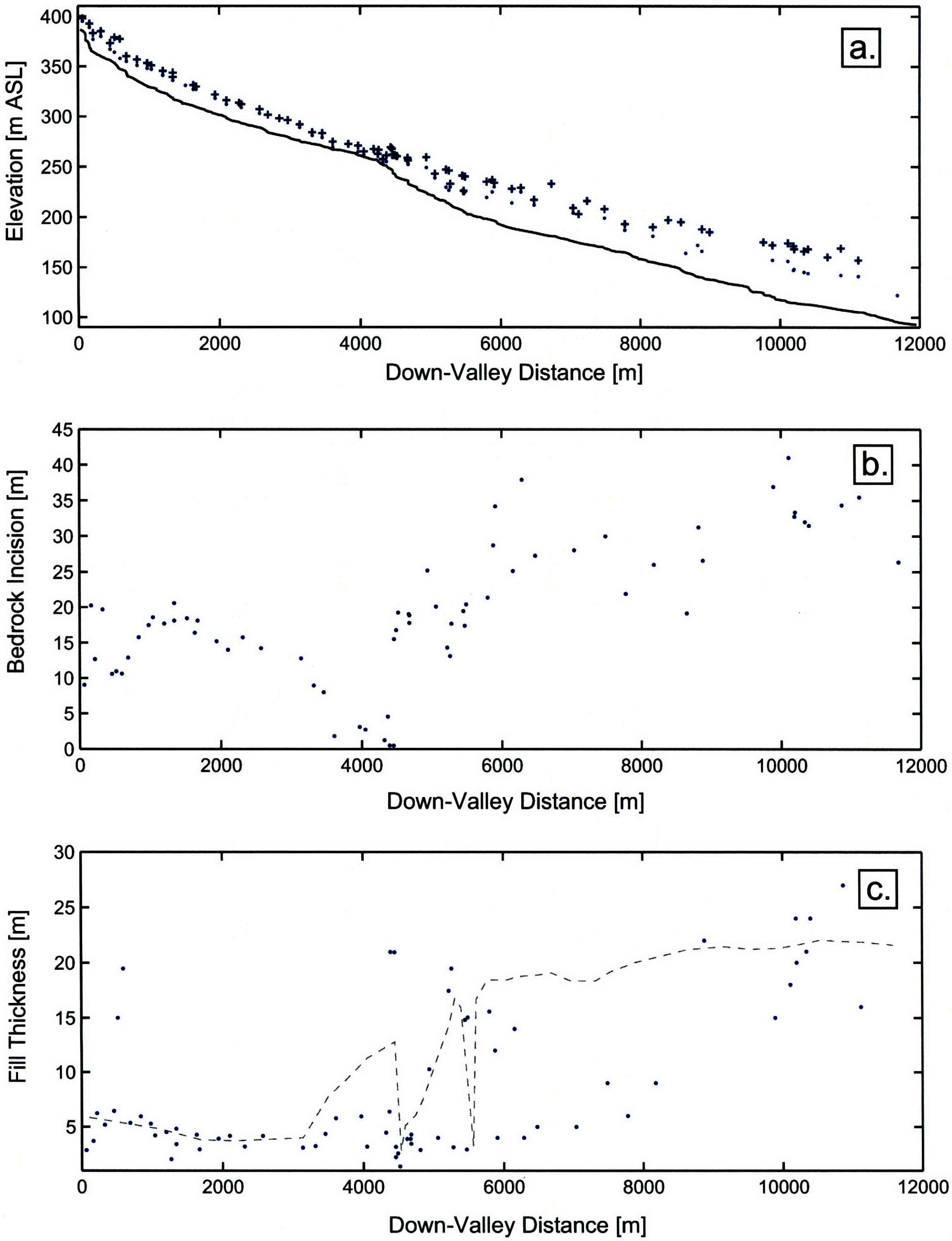

a.
Elevation [m ASL]
400
350
300
250
200
150
100
0
2000
4000
6000
8000
10000
12000
Down-Valley Distance [m]
b.
Bedrock Incision [m]
45
40
35
30
25
20
15
10
5
0
Down-Valley Distance [m]
c.
Fill Thickness [m]
30
25
20
15
10
5
Down-Valley Distance [m]

Figure 9

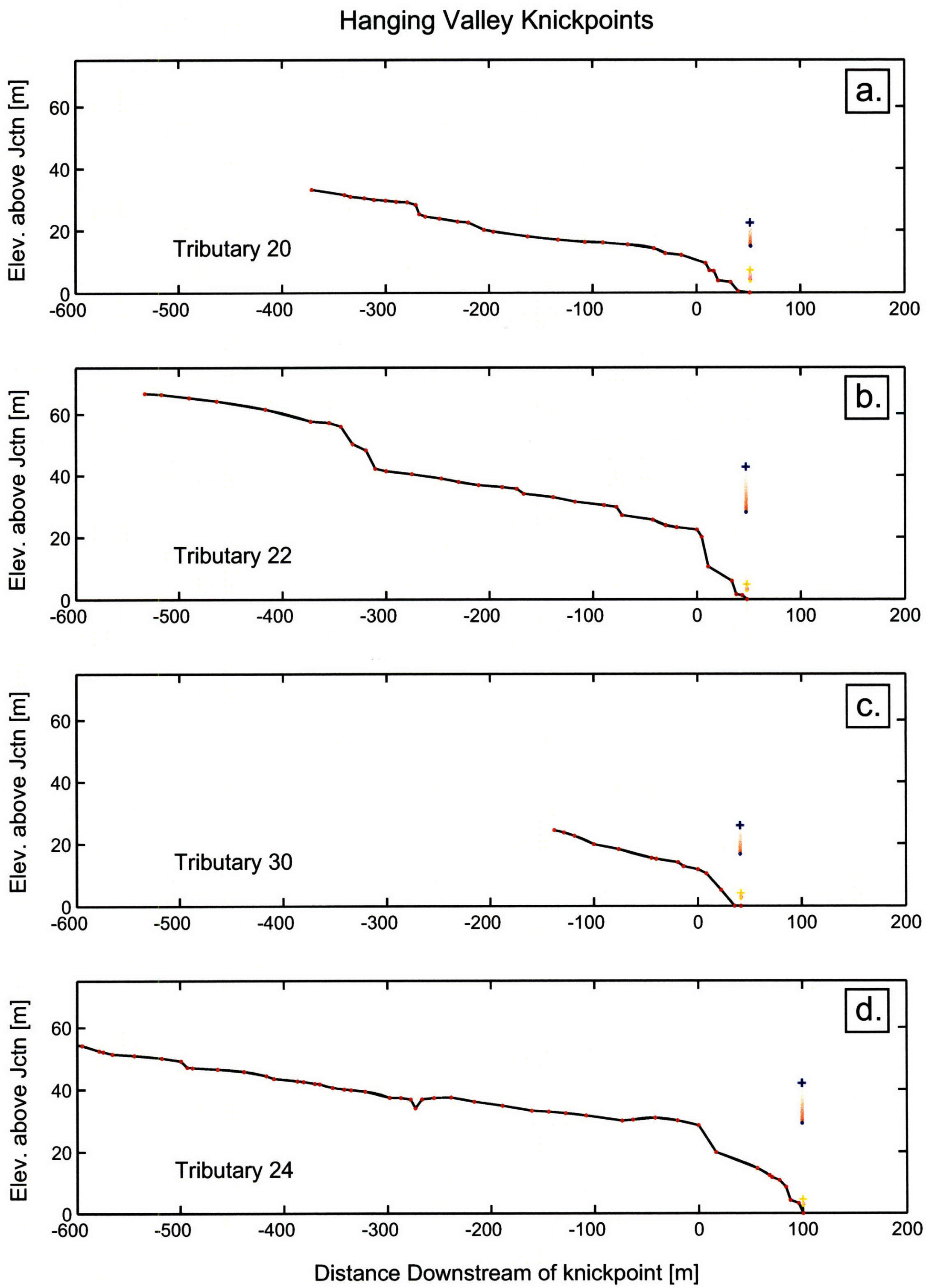

Hanging Valley Knickpoints
a.
Tributary 20
b.
Tributary 22
c.
Tributary 30
d.
Tributary 24
Elev. above Jctn [m]
0
20
40
60
-600
-500
-400
-300
-200
-100
0
100
200
Distance Downstream of knickpoint [m]

Figure 10

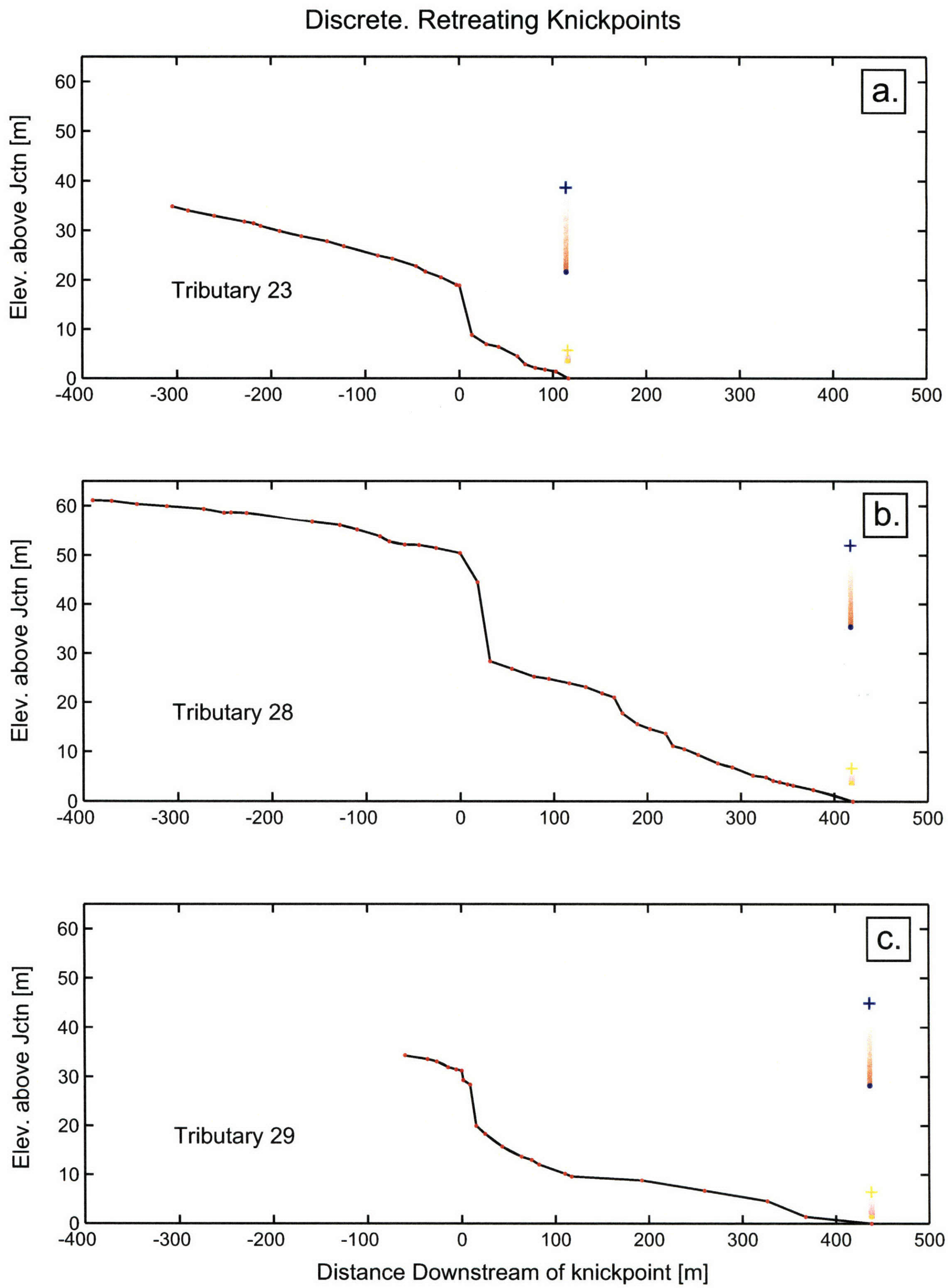

Discrete. Retreating Knickpoints
a.
Tributary 23
b.
Tributary 28
c.
Tributary 29
Elev. above Jctn [m]
Distance Downstream of knickpoint [m]
0
10
20
30
40
50
60
-400
-300
-200
-100
0
100
200
300
400
500

Figure 11

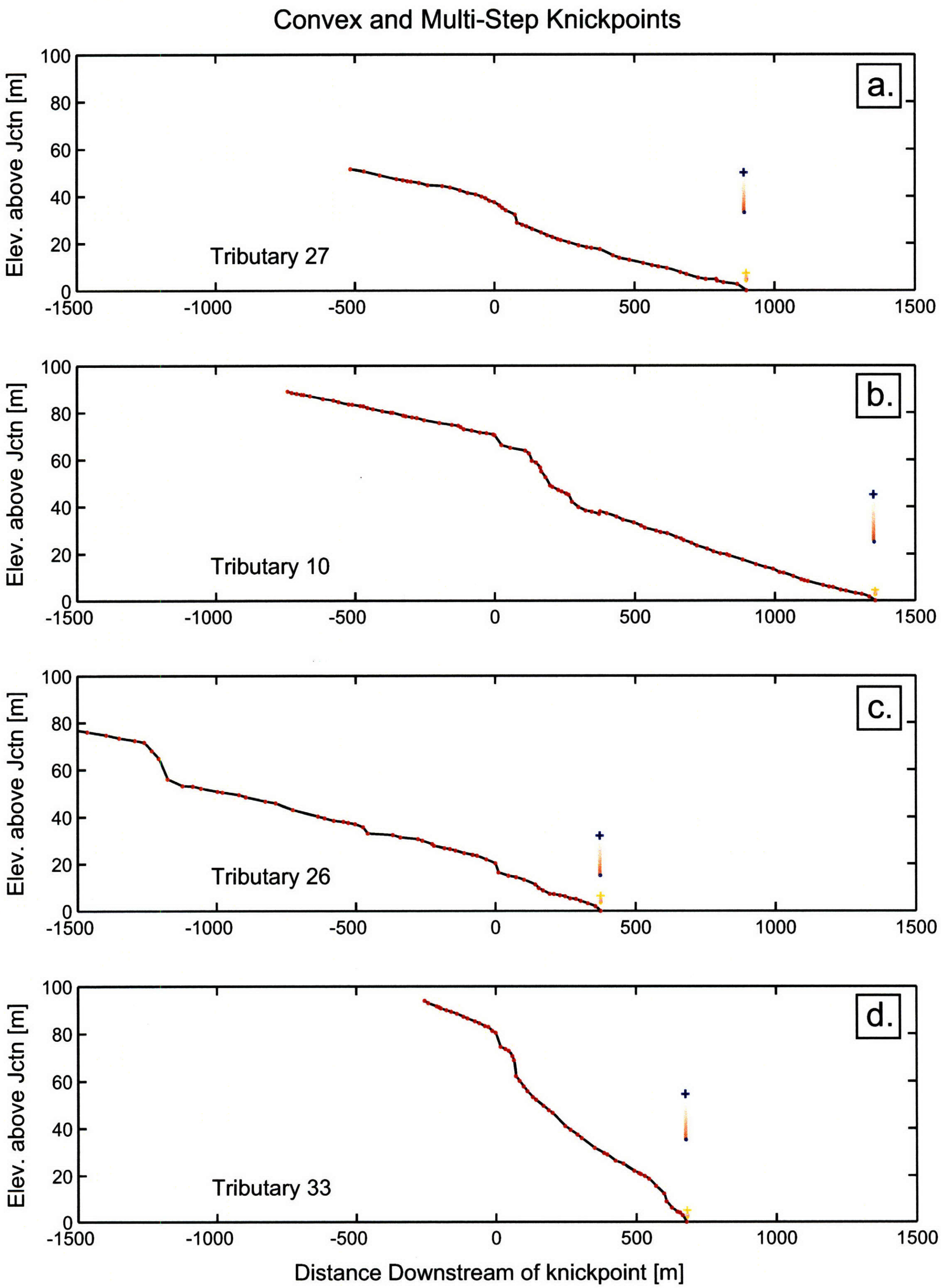
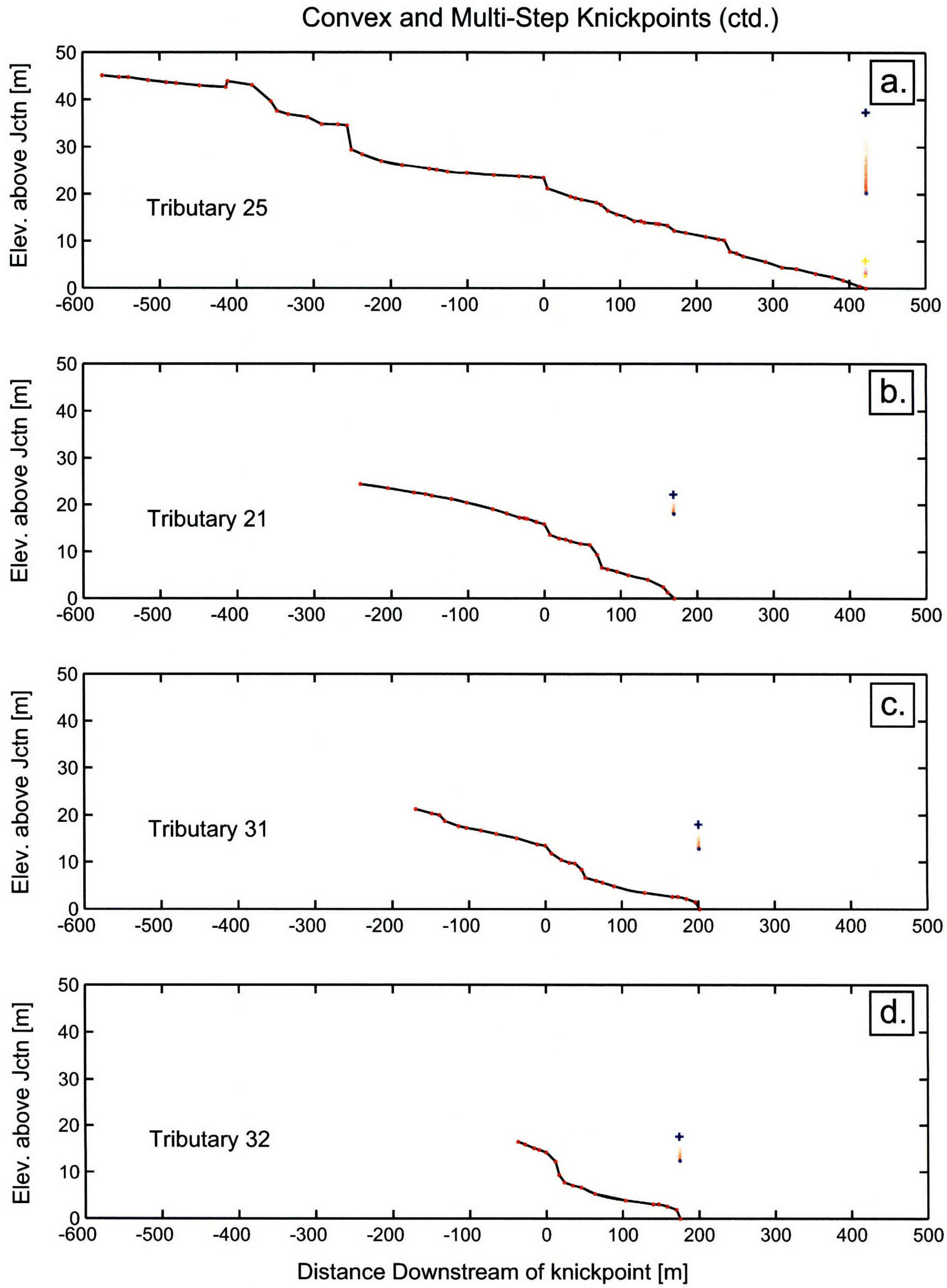

Convex and Multi-Step Knickpoints
a.
Tributary 27
b.
Tributary 10
c.
Tributary 26
d.
Tributary 33
Elev. above Jctn [m]
Distance Downstream of knickpoint [m]
100
80
60
40
20
0
-1500
-1000
-500
0
500
1000
1500

Figure 12


Convex and Multi-Step Knickpoints (ctd.)
a.
Tributary 25
b.
Tributary 21
c.
Tributary 31
d.
Tributary 32
Elev. above Jctn [m]
Distance Downstream of knickpoint [m]
50
40
30
20
10
0
-600
-500
-400
-300
-200
-100
0
100
200
300
400
500

Chapter 6: Thesis Synthesis

The preceding chapters provide significant new insight regarding the formation of knickpoints and their role in the transient response of bedrock river networks to incision. Observations of knickpoint form and distribution within the Waipaoa River catchment on the North Island of New Zealand provide a rich dataset from which we can evaluate different theoretical models for knickpoint initiation and retreat. In our field study, knickpoints define the upstream extent of a large magnitude (50-100 m) pulse of incision that initiated ~18,000 years ago. Observations, made using field techniques and digital topographic analysis demonstrate that though the knickpoints in the Waipaoa River are found distributed over a broad range of elevations, they are primarily found at relatively similar, small drainage areas and located within tributaries, just upstream of their junction with the trunk stream. Upstream of the knickpoints, the low relief, subdued relict topography supplies little sediment to the modern streams. Directly downstream of the knickpoint, where the channel has begun adjusting to the incision signal, the over-steepened gorge walls supply coarse debris to the channel. Though there is an ample supply, the clay-rich mudstone that dominates the bedrock in the Waipaoa River catchment is highly susceptible to shrink-swell weathering processes driven by cyclic wetting and drying of the rock. This results in the rapid decomposition of supplied sediment into small particles that are rapidly swept downstream and appear not to function as effective tools for bedrock abrasion. Though sculpted bedforms and pocked surfaces suggest that abrasion is the dominant incision process, at some locations we observed bedrock incision facilitated by weathering processes. At these locations, it appears that erosion is accomplished mainly by hydraulic plucking of blocks and exfoliated slabs weakened by shrink-swell weathering. Retreat of the knickpoint face is observed to be weathering dominated because the slopes of the knickpoint faces are near vertical and there is little opportunity for interaction between the falling water and the knickpoint face. Rather than retreating by abrasion of the waterfall face, most knickpoints appear to be retreating by gravitational toppling of blocks loosened by shrink-swell weathering processes. The wet-dry cycling driving these processes occurs as a consequence of both waterfall spray and weak groundwater seepage. Very few knickpoints have plunge pools at their base. Instead, falling water and sediment directly impact bedrock surfaces at the base of the falls, abrading the exposed rock.

In order to constrain variation in channel character across knickpoint reaches, surveys of channel profile, bankfull width, mean grainsize and percent bed exposure were completed in 27 tributaries to the Waipaoa. These surveys reveal that upstream reaches steepen slightly as the knickpoint is approached. Also, we were surprised to find that channel widths were, on average, wider in the reach downstream of the knickpoint. This was contrary to our expectation that channel widths in the incised downstream gorges would be narrower than in the upstream reaches. This widening could be a consequence of the increased channel slope in this downstream reach. Mean sediment grainsize showed no consistent trends in the downstream direction. Though percent bed exposure does notably increase within a 100 m reach upstream and downstream of the knickpoint lip, there was no consistent downstream trend. The large variation in the profile, morphology and bedstate within the surveyed knickpoint reaches suggest that the channel forms do not reveal a deterministic process but rather reflect the unique local conditions in each tributary.

In order to explore why knickpoints are frequently observed just upstream of tributary junctions and why the tributary response appears not to follow a deterministic behavior, we executed a focused study of trunk stream/tributary relations in the Waihuka River, a large tributary within the Waipaoa River catchment. We completed stream profile and terrace surveys of ~14 km of the Waihuka trunk and ~12 km of tributary streams. In these surveys we recognized that knickpoint form is a function of both internal (within the tributary) and external (determined by the trunk stream) factors. The most important internal control on knickpoint form is the stability of the local substrate and whether steep slopes (both in the channel and on the hillslopes) can be maintained during the pulse of incision. In weaker substrates, mass failures can destroy the over-steepened knickpoint present at the tributary junction, thus facilitating the fluvial propagation of the incision signal upstream. Where the substrate is less susceptible to failure, knickpoints are more frequently found close to their junction with the trunk stream. In the Waihuka River, external controls on knickpoint form include: the pace and magnitude of trunk stream incision and the amount and direction of lateral erosion by the trunk stream. It was particularly important to recognize that though flights of strath terraces observed along the Waihuka trunk stream provide a record of progressive incision, most tributaries have large, single-step knickpoints near their confluence with the mainstem. This reveals that the large knickpoints observed in the tributaries were not created by the upstream propagation of similar features in the trunk stream, but rather by the tributary stream's inability to keep pace with trunk stream incision.

These field observations suggest that tributary knickpoints are not created by main stem knickpoints, but rather as a consequence of a threshold in bedrock incision processes that occurs at the junction between the main stem and the tributary. To explore this threshold, we employ a two-dimensional numerical landscape evolution model (CHILD) to test how of four different expressions for river incision communicate signals of either gradual or instantaneous base level fall throughout a fluvial network. We find that permanent hanging valleys are predicted only if the incision rule predicts a non-monotonic relationship between incision rate and channel slope. In particular, we find that expressions that predict a convex hump-like relationship between incision rate and slope can predict hanging valleys. In an expression of this form, the incision rate increases with slope until a maximum incision rate. Beyond this slope, the efficiency of the incision mechanism decreases until the slope becomes so steep that the channel incision rate is lower than the background uplift rate. At this point, the tributary is unable to keep pace with the incision in the main stem. Though the Sklar and Dietrich model for abrasion by saltating bedload possesses this theoretical form, we suggest that this finding is not exclusive to the incision mechanism represented in their formulation. We suggest that any incision relation that predicts a humped relation between incision rate and slope predicts the formation of hanging valleys when mainstem incision outpaces the tributary response and creates oversteepened tributary junctions.

Through application of theoretical, numerical and field techniques, this thesis provides insight into the mechanisms behind knickpoint formation, the variation in knickpoint form and the distribution these features within fluvial networks. Further work is required to better understand the degradation of knickpoint forms and the eventual communication of the incision signal beyond the tributary junction and into the upper reaches of the fluvial network. The degradation and contemporaneous retreat of the knickpoint form is likely a weathering-mediated mass stability problem and will be extremely sensitive to the chemical and physical properties of the local substrate. To fully quantify the response time of river basins following an episode of channel incision, it will be necessary to constrain not only knickpoint formation and their subsequent degradation but also to understand the rates and mechanisms by which hillslopes respond to incision in the streams that they flank. As well, it will be valuable to examine how buffers between incised channels and hillslopes (such as fluvial terraces) extend the response time following incision. Through these improvements in our understanding of the transient response of landscapes to fluvial incision, we will be better prepared to address the unloading and deformation of mountain ranges and the consequent delivery of eroded materials to depositional basins.