Earth Surface Processes and Landforms
Earth Surf. Process. Landforms **30**, 767–778 (2005)
DOI: 10.1002/esp.1191

Knickpoint recession rate and catchment area: the case of uplifted rivers in Eastern Scotland

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Received 6 November 2003;
Revised 28 September 2004;
Accepted 25 October 2004

Abstract

Knickpoint behaviour is a key to understanding both the landscape responses to a base-level fall and the corresponding sediment fluxes from rejuvenated catchments, and must be accommodated in numerical models of large-scale landscape evolution. Knickpoint recession in streams draining to glacio-isostatically uplifted shorelines in eastern Scotland is used to assess whether knickpoint recession is a function of discharge (here represented by its surrogate, catchment area). Knickpoints are identified using DS plots (log slope versus log downstream distance). A statistically significant power relationship is found between distance of headward recession and catchment area. Such knickpoint recession data may be used to determine the values of m and n in the stream power law, $E = KA^mS^n$. The data have too many uncertainties, however, to judge definitively whether they are consistent with $m = n = 1$ (bedrock erosion is proportional to stream power and KPs should be maintained and propagate headwards) or $m = 0{\cdot}3$, $n = 0{\cdot}7$ (bedrock incision is proportional to shear stress and KPs do not propagate but degrade in place by rotation or replacement). Nonetheless, the E Scotland m and n values point to the dominance of catchment area (discharge) in determining knickpoint retreat rates and are therefore more consistent with the stream power law formulation in which bedrock erosion is proportional to stream power. Copyright © 2005 John Wiley & Sons, Ltd.

Keywords: knickpoint; stream power law; bedrock river; glacio-isostatic rebound; Eastern Scotland

Introduction

A knickpoint is a steep reach in a fluvial long profile, reflecting localized bed incision (Gardner, 1983). This steepening may be a response to more resistant lithology, in the way that Hack (1957, 1973) envisaged, or a response to an increase in shear stress or, in the alluvial case, a decrease in shear strength (ground cover) leading to the development of a propagating gully head. A knickpoint may also be a disequilibrium steepening in response to a relative fall in base level. Such disequilibrium knickpoints in bedrock rivers are commonly triggered by surface uplift and sea-level fall, and are our focus here.

The knickpoint (KP) provides the key communication link between the base level and the upstream catchment (Whipple and Tucker, 1999; Whipple *et al*., 2000). Bedrock channels therefore set much of the relief structure of tectonically active landscapes and dictate relationships between relief, elevation, and denudation (Howard *et al*., 1994). Thus, elucidating KP behaviour, especially in terms of the *rates* and *styles* of KP propagation, is fundamental to understanding catchment-wide responses to perturbation and landscape recovery times, as well as for elucidating sediment fluxes from catchments in response to base-level perturbations (see, e.g., Tucker and Slingerland, 1996; Paola, 2000).

Rates and styles of KP behaviour have been investigated using field observations (e.g., Young, 1985; Miller, 1991; Nott *et al*., 1996; Seidl *et al*., 1996; Weissel and Seidl, 1998; Hayakawa and Matsukura, 2003), physical ('sand-box') modelling of KP recession (e.g. Holland and Pickup, 1976; Gardner, 1983) and both large-scale numerical modelling of KP retreat (van der Beek *et al*., 2001) and more simple mathematical treatment of dated long-profile perturbations as diffusive phenomena (Begin, 1988). The *style* of KP recession is probably most commonly thought of as parallel retreat, but Gardner (1983) has highlighted other styles of KP behaviour. If the KP dies out by backwards rotation (that is, by Gardner's (1983) processes of inclination or replacement), the base-level perturbation is either accommodated in the KP's vicinity or diffuses away upstream, and the full magnitude of the base-level fall may not be communicated upstream. Headward propagation of the KP's full height by parallel retreat communicates the whole perturbation to the catchment, causing bed incision, steepened hillslopes and increased sediment flux. Field observations indicate that bedrock lithology and structure are important determinants of KP morphology and behaviour (e.g. Miller, 1991; Alexandrowicz, 1994) but neither undercutting nor a caprock is essential for knickpoint/waterfall maintenance and propagation (Young, 1985; Bishop and Goldrick, 1992).

The *rate* of KP recession is also a key determinant of landscape response to base-level perturbation (cf. Nott *et al.*, 1996; Whipple *et al.*, 2000). In the extreme case, such as in the KP which does not propagate upstream, there is no catchment-wide rejuvenation and no increase in sediment flux following the base-level perturbation. Hayakawa and Matsukura (2003) presented data on rates of propagation of waterfalls, arguing that recession rate is a function of discharge over the waterfall and the area of the waterfall face. It is intuitively reasonable that discharge, which Hayakawa and Matsukura (2003) approximated by the product of catchment area upstream of the waterfall and the catchment's annual precipitation, should be an important determinant of KP recession rate. This intuitive reasonableness finds expression in the 'stream power law', incorporating discharge and channel slope, which is widely used as a first-order representation of bedrock river evolution in, for example, numerical surface process models (SPMs) of landscape evolution (see, e.g., Howard and Kerby, 1983; Willgoose *et al.*, 1991a, 1991b; Howard *et al.*, 1994; Tucker and Slingerland, 1994; Whipple and Tucker, 1999; Kirby and Whipple, 2001). This stream power law does not incorporate detail of waterfall or KP morphology because it is assumed that catchment area is an adequate surrogate for discharge, which is itself assumed to be adequate as a first-order determinant of bedrock channel processes. The study presented here uses bedrock KP recession from uplifted postglacial shorelines in E Scotland to assess these assumptions by testing the hypothesis that the distance of KP recession (and hence the rate of KP recession, assuming that the shorelines have the same approximate age along their lengths) is a function of discharge as represented by catchment area.

Approach

Catchment area is a well established surrogate for discharge and is used in all of the stream-power-based numerical models referenced in the previous paragraph. The applicability of a stream-power-type law for KP recession was assessed by testing for a relationship between the distance of inland recession of KPs from the uplifted E Scotland coast and the catchment area draining to the coast across each KP. The E Scotland data were also used to derive values of the m and n parameters in the stream power formulation of fluvial incision. The data of Hayakawa and Matsukura (2003) on the controls on rates of KP recession were also re-examined to assess the relationship between rate of KP recession and catchment area.

Shorelines, knickpoints, and catchment areas

Catchment area and distance of KP recession were measured on streams that drain to the coastlines of Kincardineshire and Berwickshire, E Scotland (Figure 1). These shorelines are widely reported as post-glacially uplifted (see, e.g., Boulton *et al.*, 1991; Firth *et al.*, 1993; Lambeck, 1995). A prominent bench (or terrace in the terminology of Cullingford and Smith (1980)) of varying width occurs at 30–35 m ASL on the Kincardineshire coast and is the highest of up to eight Kincardineshire shorelines (Cullingford and Smith, 1980). This bench is marine and was formed during deglaciation when the decaying ice front was still close by. This deglaciation age is indicated by the nearby ice-contact fluvio-glacial landforms, as well as the fluvio-glacial outwash terraces that lie landward of the bench and pass laterally into the bench or are graded to slightly below the bench. The same relationships between fluvio-glacial outwash terraces and shorelines are found in the Fife area, to the south of Kincardineshire (Cullingford and Smith, 1966).

Sequences of uplifted shorelines or benches on the Berwickshire coast also culminate in a prominent bench at elevations similar to, or slightly lower than, those in Kincardineshire (Figure 2). However, the Berwickshire benches have received considerably less attention than those in Kincardineshire. Glacial till overlying the Berwickshire bench indicates a pre-last glacial maximum age for the bench and Rhind (1965) has argued that it is at least therefore of an interstadial, but otherwise indeterminate, age (see also Hall, 1989). The Berwickshire bench is interpreted here to be a marine shoreline on the bases of its consistent elevation and morphology along the coast, the way in which it cuts across bedrock structure, and its similarity to the Kincardineshire bench in gross field appearance. It almost certainly pre-dates the last glacial maximum but, for the moment, knickpoint retreat in Berwickshire is treated as having occurred since the last deglaciation (but see below).

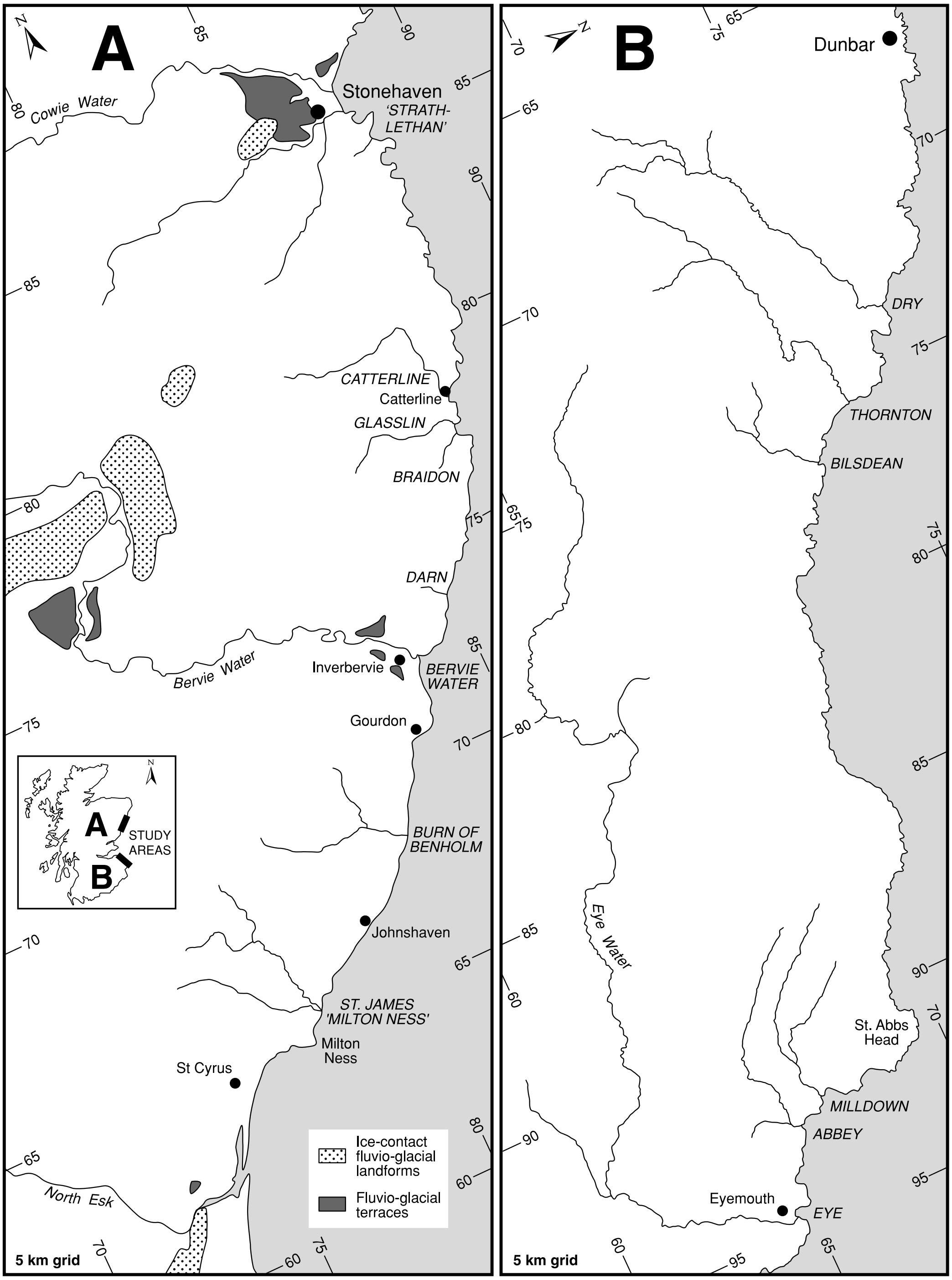

Figure 1. Maps of (A) Kincardineshire and (B) Berwickshire, showing in upper case italics the streams used in the analysis. The inset shows Scotland. See Table 1 for grid references for all streams used in this study.

Andersen's (1981) chronology dates the deglaciation of Kincardineshire and Berwickshire to between about 13 ka and 14 ka, with Berwickshire emerging from the ice a little earlier than Kincardineshire. These generalized deglaciation dates are essentially confirmed by Smith *et al.* (2000), who dated the early deglacial shoreline to 13·5 ka–12·5 ka, whereas Boulton *et al.* (1991) dated the deglaciation of east Fife to about 15 ka and Berwickshire a little earlier. There is clearly

Figure 2. The mouth of Abbey Burn in Berwickshire, highlighting the way in which the uplifted shoreline bench has been incised by Abbey Burn. This figure is available in colour online at www.interscience.wiley.com/journal/espl

an urgent need for more precise age controls on the deglaciation of these parts of the E Scotland coast but precise age(s) for these shorelines and deglaciation are not critical to this study, and we adopt a deglaciation age of 14 ka.

The Kincardineshire study area is dominated by Devonian Lower Old Red Sandstone sediments (Haughton and Bluck, 1984). The sediments consist mainly of conglomerates, which become finer to the south. Occasional dykes and sills intrude the sedimentary units and thicker volcanic formations of varied composition are also found. Dips are generally about 20° towards the SW to W, and the sequence is locally faulted. The Berwickshire study area is even less homogeneous geologically, with Lower and Upper Old Red Sandstones and Silurian and Carboniferous mixed lithologies present.

The prominent coastal benches at 30–35 m ASL in Kincardineshire and 25–35 m ASL in Berwickshire were identified in the two areas by field inspection and on UK Ordnance Survey maps at 1:25 000 scale (5 m contour interval), supplemented by the mapping of Cullingford and Smith (1980) for Kincardineshire. On several stretches of the study coastline, the bench was not present (probably due to retreat of these more exposed stretches of coastline); rivers that drained to these stretches of coast were not studied.

The long profiles of coastal streams that have incised into the Kincardineshire and Berwickshire benches were obtained by digitizing the blue line and the 5 m contour crossings on the 1:25 000 scale maps. The long profiles of all these streams exhibit steepening, as discrete steps in the middle and lower reaches of larger streams and as an overall steep lower reach in the smaller streams. These steepenings are interpreted as KPs and this is confirmed by DS plots. A DS plot is a plot of the logarithm of gradient of a river reach versus the logarithm of downstream distance of that reach (Goldrick, unpublished PhD thesis; see Bishop and Goldrick, 2000, for a summary). The DS form of the long profile is built on the simple and well known power relationship between channel discharge (Q) and downstream distance (L) (Howard and Kerby, 1983; Seidl and Dietrich, 1992):

$$Q = lL^b \tag{1}$$

where l and b are constants. Equation 1 is combined with the stream power law:

$$E = KA^m S^n \tag{2}$$

where E is fluvial incision, K is a dimensional coefficient of erosion, A is catchment area (a surrogate for channel discharge), S is channel gradient, and m and n are positive constants), to give the equilibrium (time-independent, steady-state) DS profile form:

$$\ln S = \gamma - \lambda \ln L \qquad (3)$$

where L is downstream distance, and γ and λ are constants.

An equilibrium, steady-state long profile gives a straight line on a DS plot (enabling assessment of whether a long profile is in equilibrium). A further key contribution of the DS form is its ability to distinguish transient channel steepening in response to base-level fall (i.e. disequilibrium steepening that propagates as a KP) from channel steepening that is a persistent (time-independent), steady-state (equilibrium) response to the channel substrate or discharge conditions. Equilibrium steepening in response to more resistant lithology is indicated on the DS plot by a parallel shift in the plot, whereas a disequilibrium KP plots as disordered outliers on the DS plot (see Figure A11.1 of Bishop and Goldrick, 2000). The widely used Hack (1973) SL long profile form is unable to distinguish equilibrium steepening from disequilibrium steepening in this way.

It is critical to this study that the disequilibrium outliers on the DS plots are indeed KPs that are propagating headward (and not, for example, bedrock structural features or remnants of sub-glacial erosion). The widespread preservation of the marine bench surface along the edges of the fluvial gorge downstream of the KP (Figure 2) is consistent with the KP having formed by headward retreat since deglaciation. Headward retreat is also confirmed more generally by the spatial distribution of KPs in trunk and tributaries of large streams (Figure 3).

Representative long profiles and DS plots for two of the 15 E Scotland coastal rivers are presented in Figure 4. For all streams in this study, the KP was initially identified on the DS plot as the first outlier (spike) upstream of the shore (circled on Figure 4). Several KPs are to be expected on the lower reaches of the E Scotland coastal streams because the vertical sequences of multiple Late Pleistocene and Holocene shorelines identified in E Scotland (e.g. Sissons *et al*. (1966) for the Firth of Forth area and the 'staircases' of benches (terraces) of Cullingford and Smith (1966, 1980) in East Fife and Kincardineshire, respectively) point to several discrete falls of base level (relative sea level). It is assumed here that the essentially continuous uplift of E Scotland with several base-level stillstands, corresponding to periods when sea level and the land were rising at the same rate, resulted in markedly disequilibrium lower reaches, representing several, perhaps-coalesced long profile disequilibria (KPs) on the smaller streams with lower stream power (e.g. Figure 4A). The 5 m contour interval is inadequate for distinguishing these multiple KPs in the smaller streams. In the case of larger streams with higher stream power, there may be separation of individual KPs (DS outliers), such as in Figure 4B, where the reach immediately inland of the shoreline is a cluster of DS outliers taken to indicate the most recent (Holocene) relative base-level fall (surface uplift).

The lip of a KP is at the upstream end of the reach represented by a DS spike, and the reach above this lip was projected downstream to the shore. Geometry demands that the elevation of this downstream projection at the shoreline should be the same as – or at least close to and certainly not lower than – the present elevation of the uplifted shorelines. If this was not the case, the next KP upstream was evaluated in the same way to identify the KP for which it was appropriate to measure the distance of recession from the shore (Figure 5).

Generally, the two to three long profile points (contour crossings) upstream of the knickpoint were projected linearly to the coast (Figure 5). A linear projection is only an approximation to the more correct non-linear projection (cf. Goldrick and Bishop, 1995; Goldrick, unpublished PhD thesis), but was felt to be justified in light of simplicity and the complexities and uncertainties associated with the fully correct DS projection (Goldrick, unpublished PhD thesis). The uncertainties in a linear projection increase with the distance of projection.

Catchment areas were measured on hard copy and digital 1:50 000 scale maps. The accuracy of the measurements was checked by comparing them with catchment areas given by the National Water Archive (Natural Environment Research Council) for Bervie Water and Eye Water, the two largest streams studied. The catchment areas and distances of KP recession for all the E Scotland streams are given in Table I.

Results and Discussion

There is a highly significant relationship between distance of KP recession and catchment area for both the Kincardineshire and Berwickshire data (Figure 6). Hayakawa and Matsukura (2003) argued that waterfall recession rate is a function of discharge over the waterfall and the area of the waterfall face. Treating their data as we have done here also shows highly significant relationships between recession rate and both discharge and catchment area (Figure 7). The relationship with discharge is only very marginally more significant than that with catchment area, reflecting the relative uniform average annual rainfall values in the study area of Hayakawa and Matsukura (2003). In other words, for areas of uniform precipitation, catchment area is adequate as a first-order control on KP recession rate. The fuller relationship of Hayakawa and Matsukura (2003) for KP recession as a function of catchment area, precipitation, width and height of the waterfall face, water density and unconfined compressive strength of the bedrock

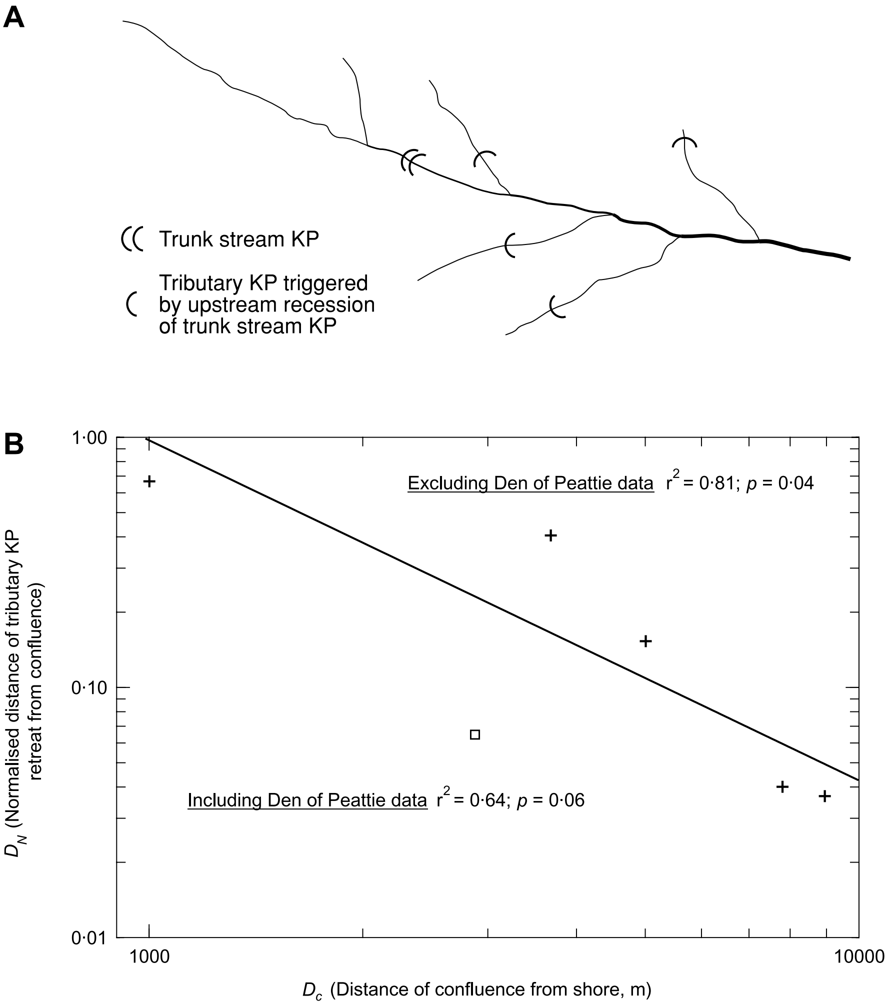

Figure 3. A. Diagrammatic representation of field relationships, highlighting the way in which the spatial distribution of tributary KPs is consistent with headward of propagation of KPs (cf. Seidl and Dietrich, 1992; Goldrick and Bishop, 1995). KPs in tributaries closer to the shoreline have propagated relatively further up their tributary, because these tributary KPs have been 'triggered' earlier by the KP propagating up the trunk stream. An inverse relationship between normalized distance of tributary KP recession and the distance of the tributary confluence from the shore is consistent with headward propagation of the KPs. B. Normalized distance, D_N, of tributary KP recession versus distance of tributary confluence from the shore, D_C, for the deglaciation KP in tributaries of Bervie Water, the largest catchment studied here: KPs in tributaries closer to the shoreline have propagated relatively further up their respective tributaries. We use normalized distance of tributary KP recession to account for differences in the sizes of tributary catchment areas. The anomalous Den of Peattie (□), which is not included for the fitting of the trend line, may reflect an extreme lithological influence.

of the waterfall certainly provides a fuller explanation than the simple relationships explored here ($r^2 = 91\%$ for the full relationship, $r^2 = 82\%$ for the relationship between R and A or AP, where P is mean annual precipitation) (Figure 7). However, for situations in which data are not available for the full parameter set, AP and, more simply, A provide adequate substitutes. It is also noteworthy from Figure 7 that the E Scotland data plot within, but a little lower than, the data of Hayakawa and Matsukura's (2003). These slightly lower R values in E Scotland compared to Japan probably reflect the higher rates of tectonic activity and more erodible lithologies of the Japanese field area (cf. Hayakawa and Matsukura, 2003), as well as precipitation differences and stochastic effects.

It is noteworthy that the Berwickshire and Kincardineshire data are essentially indistinguishable (Figure 6). This relationship suggests that there may be little difference in the timing of KP initiation in the two study areas, implying in turn that the age of the Berwickshire marine bench is close to that of the Kincardineshire (i.e. interstadial age rather

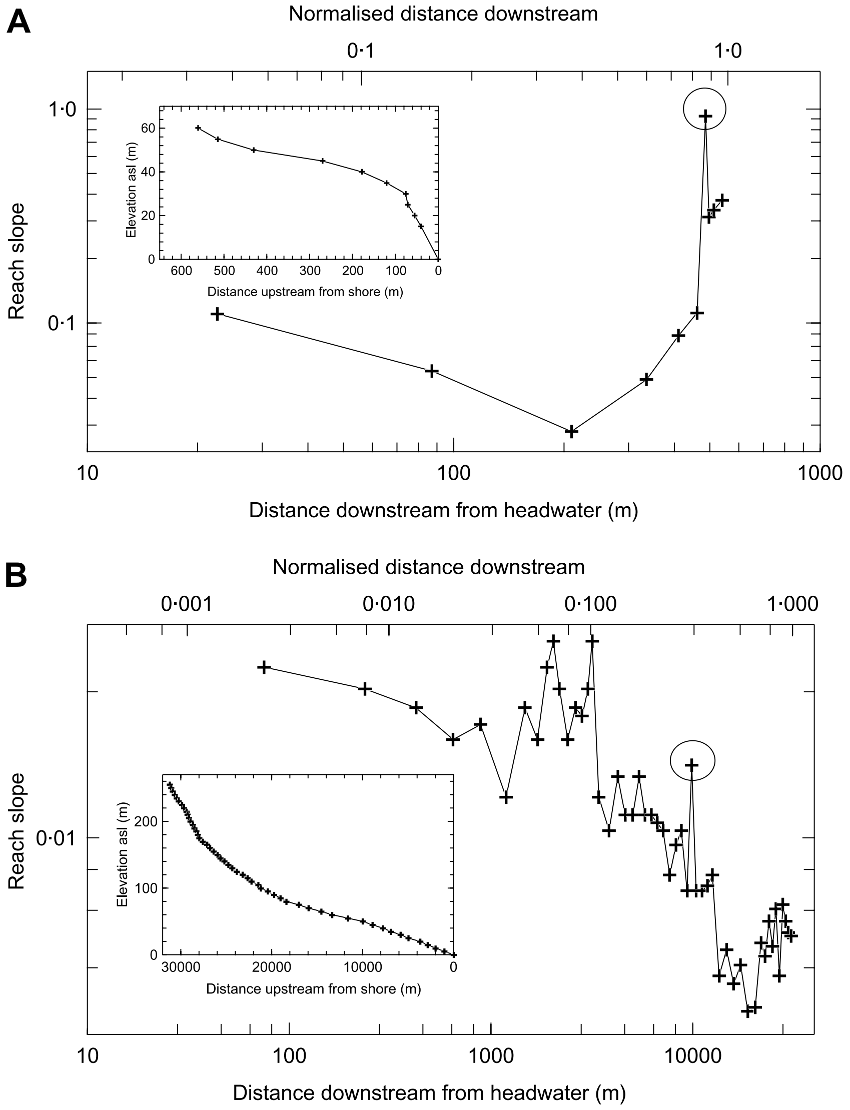

Figure 4. DS plot (with long profile in inset) for (A) 'Strathlethan Bay' (smallest stream in this study) and (B) Bervie Water (largest stream). The KPs used here are circled (see text).

than inter-glacial; cf. Rhind, 1965). The best fit to the combined data set from E Scotland is a highly significant power relationship with the equation $D = 50{\cdot}8A^{1{\cdot}24}$ (Figure 6). The residuals from this relationship are likely to be related to factors such as inheritance from the local pre-existing glacial morphology, inaccuracies associated with KP identification, and local structural and lithological influences. The lack of primary control of KP recession distance by lithology parallels the finding of Bishop and Cowell (1997) that lithology exerts, at most, only a secondary control on valley dimensions in river valleys in coastal E Australia. The role of lithology in long profile morphology and evolution (and landscape evolution in general) merits further investigation, not least because of the emphasis that Hack (1973) placed on the importance of lithology (cf. Miller, 1991; Alexandrowicz, 1994).

The relationships revealed by the data of Hayakawa and Matsukura (2003) and this study indicate that KP recession in numerical surface process models (SPMs) can safely be represented as a function of discharge or catchment area, at least for first-order understanding of long-term landscape evolution. Such a 'first order' approach is consistent with the comment by Kooi and Beaumont (1994) that 'as demonstrated in other sciences, a fruitful approach to such problems of [the range of] scale [that must be captured in SPMs] is to set aside (for the time being) the small-scale, short-timescale picture and to explore simple relationships cast in terms of large-scale, long-timescale quantities' (p. 12192).

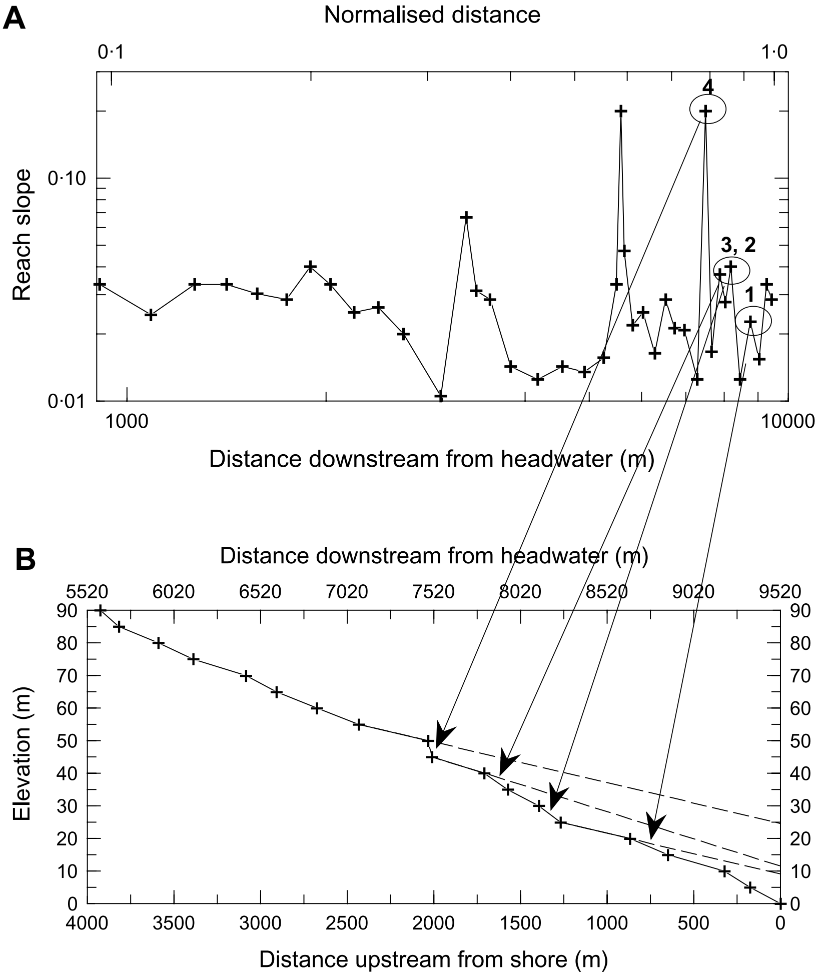

Figure 5. (A) DS plot and (B) long profile of the downstream reach of Thornton Burn, Berwickshire to illustrate the method of identifying KPs and the downstream projection to the shoreline of the reach above the KP. Only the downstream projection of KP 4 gives an elevation at the shoreline that is appropriate for the current elevation of the uplifted Berwickshire shoreline (20–25 m ASL). The downstream projections of the reaches above KPs 1 and 3 are too low in elevation at the shoreline, and the downstream projection from upstream of KP 2 is below present sea level at the shoreline (not illustrated).

This approach has recently been demonstrated to be adequate for the representation of fluvial processes in general in SPMs (van der Beek and Bishop, 2003). Thus, attempts to represent bedrock incision processes more accurately than the relatively simple, commonly used stream power formulation are welcome (e.g. Sklar and Dietrich, 1998), but the stream power formulation is adequate until these more sophisticated representations are realized.

A key requirement for using the stream power fluvial incision law and adequately representing the transient behaviour of knickpoints is appropriate parametrization of the law (Equation 2). The values of the m and n coefficients are critical in this regard and so are the subject of considerable research (e.g. Howard *et al.*, 1994; Hancock *et al.*, 1998; Whipple *et al.*, 2000; van der Beek and Bishop, 2003). For physical reasonableness, and depending on precisely how the incision law is formulated, m should take values of either 0·3–0·5 or 1, and n should lie between 0·7 and 1 (Howard *et al.*, 1994; van der Beek and Bishop, 2003). For $m = n = 1$, bedrock erosion is proportional to stream power and KPs are maintained and propagate headwards. Where bedrock incision is proportional to shear stress, $m = 0{\cdot}3$ and $n = 0{\cdot}7$, and KPs do not propagate but degrade away by rotation or replacement (cf. Gardner, 1983; Howard *et al.*, 1994).

Propagating KPs, such as are found on the E Scotland coast, would be expected to be associated with values of $m = n = 1$. Values for m and n may be derived from long profile data following Whipple and Tucker (1999). A regional

Table I. E Scotland streams used in this study. Where the name of a stream is unknown, a topographic name associated with the stream's mouth is given in inverted commas. The Den of Peattie (italics) plots as a marked outlier in the Bervie Water tributary data (Figure 5)

Area	Stream	Grid reference of mouth	Distance of KP recession from shore (m)	Catchment area (km^2)
Kincardineshire	'Strathlethan Bay'	8 830, 8 460	76	1·0
	Catterline Burn	8 695, 7 760	1 920	20·6
	Glasslin Burn	8 685, 7 740	290	4·2
	Braidon Burn	8 680, 7 720	220	2·2
	'Darn Bay'	8 490, 7 380	30	2·0
	Bervie Water	8 355, 7 240	21 500	128·3
	Burn of Benholm	8 135, 6 850	1 550	12·7
	Denfinella	7 770, 6 550	1 170	19·9
Berwickshire	Dry Burn	7 350, 7 600	1 700	18·9
	Thornton Burn	7 530, 7 420	2 035	17·3
	Bilsdean Burn	7 660, 7 280	225	6·5
	Milldown Burn	9 200, 6 630	660	8·3
	Abbey Burn	9 250, 6 550	1 255	6·5
	Eye Water	9 450, 6 450	18 000	127·0
			Distance of KP recession from confluence (m)	**Length of tributary (m)**
Bervie Water tributaries	Pitcarry Burn	8 300, 7 305	2 625	3 925
	Den of Peattie	*8 180, 7 380*	*300*	*4 625*
	Burn of Pitcarles	8 120, 7 425	970	2 385
	Hareden Burn	8 050, 7 430	225	1 465
	Woodburn	7 795, 7 470	155	3 855
	Mill of Garvock	7 730, 7 485	100	2 715

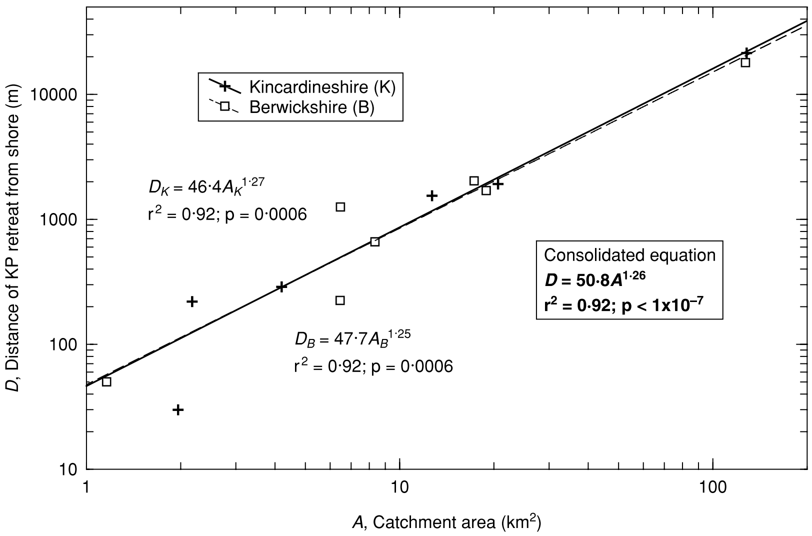

Figure 6. Distance of KP recession from the coast, D, versus catchment area, A, for deglaciation shoreline KPs in coastal streams in Kincardineshire (K) and Berwickshire (B), E Scotland. The box gives the best-fit relationship for the combined data sets.

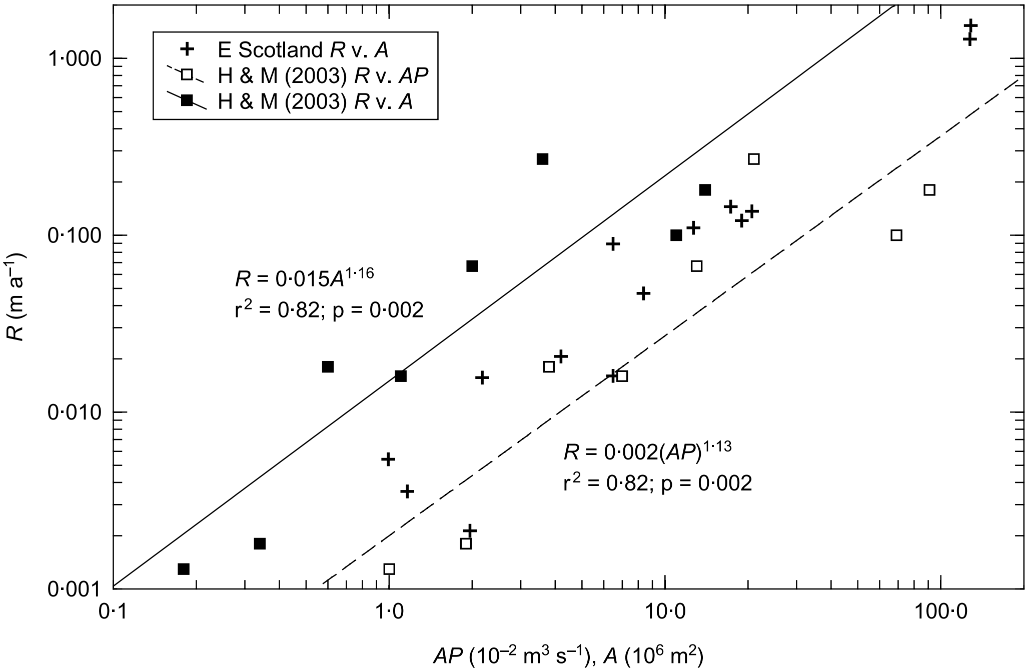

Figure 7. Plot of data from Hayakawa and Matsukura (H & M) (2003) for relationships between rate of KP recession, *R*, and discharge *AP* (the product of catchment area, *A*, and mean annual precipitation, *P*) and between rate of KP recession, *R*, and catchment area, *A*. Data for waterfall 1 (Soho Falls) of Hayakawa and Matsukura (2003) were not plotted because its plot of *R* versus *A* is approximately one order of magnitude higher than the trend of the remainder of the data of Hayakawa and Matsukura (2003). Soho Falls has the shortest period of KP recession in their data (50 years) and so is most likely not to have experienced the periods of tectonic quiescence that are interspersed with the events that trigger and drive KP recession. These periods of tectonic quiescence might be expected to be associated with slower rates of KP recession. Also included are this study's *R* versus *A* data for E Scotland.

catchment area (A)–stream length (x) relationship using all of the studied catchments yielded $A = 905x^{1\cdot10}$. This relation was entered into the kinematic wave celerity equation for the sudden base-level fall case (Whipple and Tucker, 1999, equation 25) along with measured headwater-to-knickpoint distances and mean reach slopes immediately upstream of the knickpoints (taken as the mean gradient over a reach with a 20 m fall). Multiple regression yielded values of $m = 0{\cdot}6$ and $n = 0{\cdot}2$. However, the data are very scattered, largely as a consequence of having been derived from 1:25 000 scale topographic maps with 5 m contour intervals, and the ranges for the estimates of m and n (±1 standard error) are 0·27–1·0 and 0–0·64, respectively. The data are thus inconclusive in terms of the values for m and n. They are nonetheless consistent with a conclusion that catchment area is likely to be more influential than slope in determining KP retreat rates. This finding suggests overall that discharge (or its surrogate, catchment area) is a fundamental control on KP retreat rate. More detailed field data from other studies highlight the importance, for a given discharge, of local channel slope and channel width in controlling bedrock channel incision rates (e.g. Hancock *et al.*, 1998). There is no necessary contradiction between these results: discharge provides a first-order control on KP retreat rate (or bedrock channel incision rate), which is modulated by the local flow field (especially via local channel width and/or local channel gradient). As in the study by Hancock *et al.* (1998) on bedrock channel incision, lithology appears not to exert a fundamental control of rate of retreat of KPs in the eastern Scotland streams.

The data presented here make it clear therefore that KP propagation following a base-level fall generated by, say, surface uplift will be greater and more rapid the larger the catchment, all other things being equal. Thus, in the critical matter of reading the signal of source area tectonic activity from the flux of sediments to sedimentary basins (including the volume, texture, and rate of supply of sediments), the sizes of possible source catchment areas must be borne in mind. All other things equal, larger catchments would be expected to have more rapid responses to base-level fall and to generate higher sediment volumes in response to base-level fall. As already noted, response time and sediment flux also depend in detail on stream gradient. Likewise, the degree of connectedness to the base level is an important element in understanding response times. In other words, steep high discharge streams that are well connected to base level will respond most rapidly to a rejuvenating base-level fall. On the other hand, there must be a minimum stream power (i.e. a minimum combination of catchment area (discharge) and gradient – minimum QS or AS) for which KP propagation is able to keep pace with uplift. Streams with QS (or AS) values lower than this threshold cannot accommodate uplift by propagation of a single KP. In these cases, more complex KP responses and behaviours may be triggered.

Acknowledgements

We thank the referees for very helpful comments that considerably improved an earlier version of this paper, and Mike Shand for assistance with the diagrams. JDJ is funded by the Royal Academy of Engineering (Anglo-Australian Postdoctoral Fellowship).

References

Alexandrowicz Z. 1994. Geologically controlled waterfall types in the Outer Carpathians. *Geomorphology* **9**: 155–165.

Andersen BG. 1981. Late Weichselian ice sheets in Eurasia and Greenland. In *The Last Great Ice Sheets*, Denton GH, Hughes TJ (eds). Wiley: New York; 1–65.

Begin ZB. 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.

Bishop P, Cowell P. 1997. Lithological and drainage network determinants of the character of drowned, embayed coastlines. *Journal of Geology* **105**: 685–699.

Bishop P, Goldrick G. 1992. Morphology, processes and evolution of two waterfalls near Cowra, N.S.W. *Australian Geographer* **23**: 116–121.

Bishop P, Goldrick G. 2000. Geomorphological evolution of the East Australian continental margin. In *Geomorphology and Global Tectonics*, Summerfield MA (ed.). Wiley: Chichester; 227–255 (esp. Appendix).

Boulton GS, Peacock JD, Sutherland DG. 1991. Quaternary. In *Geology of Scotland*, 3rd ed., Craig GY (ed.). Geological Society: London; 503–543.

Cullingford RA, Smith DE. 1966. Late-glacial shorelines in eastern Fife. *Transactions of the Institute of British Geographers* **39**: 31–51.

Cullingford RA, Smith DE. 1980. Late Devensian raised shorelines in Angus and Kincardineshire, Scotland. *Boreas* **9**: 21–38.

Firth CR, Smith DE, Cullingford RA. 1993. Late Devensian and Holocene glacio-isostatic uplift patterns in Scotland. *Quaternary Proceedings* **3**: 1–24.

Gardner TW. 1983. Experimental study of knickpoint and longitudinal evolution in cohesive, homogeneous material. *Geological Society of America Bulletin* **94**: 664–672.

Goldrick G, Bishop P. 1995. Distinguishing the roles of lithology and uplift in the steepening of bedrock river long profiles: An example from southeastern Australia. *Journal of Geology* **103**: 227–231.

Hack JT. 1957. Studies of longitudinal stream profiles in Virginia and Maryland. *United States Geological Survey Professional Paper* **294-B**: 45–97.

Hack JT. 1973. Stream profile analysis, and stream gradient index. *United States Geological Survey Journal of Research* **1**: 421–429.

Hall AM. 1989. Pre-Late Devensian coastal rock platforms around Dunbar. *Scottish Journal of Geology* **25**: 361–365.

Hancock GS, Anderson RS, Whipple KX. 1998. In *Rivers over Rock*, Tinkler KJ, Wohl EE (eds). *American Geophysical Union Geophysical Monograph* **107**: 35–60.

Haughton PDW, Bluck BJ. 1984. Diverse alluvial sequences from the Lower Old Red Sandstone of the Strathmore region, Scotland – implications for the relationship between Late Caledonian tectonics and sedimentation. In *Devonian of the World*, McMillan NJ, Embry AF, Glass DS (eds). *Canadian Society of Petroleum Geologists Memoir* **14**: 269–293.

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

Holland WN, Pickup G. 1976. Flume study of knickpoint development in stratified sediment. *Geological Society of America Bulletin* **87**: 76–82.

Howard AD, Dietrich WE, Seidl MA. 1994. Modeling fluvial erosion on regional to continental scales. *Journal of Geophysical Research* **99**: 13 971–13 986.

Howard AD, Kerby G. 1983. Channel changes in badlands. *Geological Society of America Bulletin* **94**: 739–752.

Kirby E, Whipple K. 2001. Quantifying differential rock-uplift via stream profile analysis. *Geology* **29**: 415–418.

Kooi H, Beaumont C. 1994. Escarpment evolution on high-elevation rifted margins: insights derived from a surface process model that combines diffusion, advection, and reaction. *Journal of Geophysical Research* **99**: 12 191–12 209.

Lambeck K. 1995. Late Devensian and Holocene shorelines of the British Isles and North Sea from models of glacio-hydro-isostatic rebound. *Journal of the Geological Society, London* **152**: 437–448.

Miller JR. 1991. The influence of bedrock geology on knickpoint development and channel-bed degradation along down-cutting streams in south-central Indiana. *Journal of Geology* **99**: 591–605.

Nott J, Young R, McDougall I. 1996. Wearing down, wearing back, and gorge extension in the long-term denudation of a highland mass: quantitative evidence from the Shoalhaven catchment, southeast Australia. *Journal of Geology* **104**: 224–232.

Paola CP. 2000. Quantitative models of sedimentary basin filling. *Sedimentology* **47** (Suppl. 1): 121–178.

Rhind DW. 1965. Evidence of sea level changes along the coast north of Berwick. *Proceedings of the Berwickshire Naturalists' Club* **37**: 10–15.

Seidl MA, Dietrich WE. 1992. The problem of channel erosion into bedrock. *Catena Supplement* **10**: 51–66.

Seidl MA, Weissel JK, Pratson LF. 1996. The kinematics and pattern of escarpment retreat across the rifted continental margin of SE Australia. *Basin Research* **12**: 301–316.

Sissons JB, Smith DE, Cullingford RA. 1966. Late-glacial and post-glacial shorelines in south-east Scotland. *Transactions of the Institute of British Geographers* **39**: 9–18.

Sklar L, Dietrich WE. 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*, Tinkler KJ, Wohl EE (eds). American Geophysical Union: Washington; 237–260.

Smith DE, Cullingford RA, Firth CR. 2000. Patterns of isostatic land uplift during the Holocene: evidence from mainland Scotland. *The Holocene* **10**: 87–103.

Tucker GE, Slingerland RL. 1994. Erosional dynamics, flexural isostasy, and long-lived escarpments: a numerical modeling study. *Journal of Geophysical Research* **99**: 12 229–12 243.

Tucker GE, Slingerland R. 1996. Predicting sediment flux from fold and thrust belts. *Basin Research* **8**: 329–349.

van der Beek P, Bishop P. 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, 10.1029/2002JB002125.

van der Beek PA, Pulford A, Braun J. 2001. Cenozoic landscape development in the Blue Mountains (SE Australia): lithological and tectonic controls on rifted margin morphology. *Journal of Geology* **109**: 35–56.

Weissel JK, Seidl MA. 1998. Inland propagation of erosional escarpments and river profile evolution across the southeast Australian passive continental margin. In *Rivers over Rock*, Tinkler K, Wohl E (eds). *AGU Geophysical Monograph* **107**: 189–206.

Whipple KX, Snyder NP, Dollenmayer K. 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 KX, Tucker GE. 1999. Dynamics of the stream-power incision model: implications for height limits of mountain ranges, landscape response timescales, and research needs. *Journal of Geophysical Research* **104**: 17 661–17 774.

Willgoose G, Bras RL, Rodriguez-Iturbe I. 1991a. A physically based coupled network growth and hillslope evolution model. 1. Theory. *Water Resources Research* **27**: 1671–1684.

Willgoose G, Bras RL, Rodriguez-Iturbe I. 1991b. A physically based coupled network growth and hillslope evolution model. 2. Applications. *Water Resources Research* **27**: 1685–1696.

Young RW. 1985. Waterfalls: form and process. *Zeitschrift für Geomorphologie N.F. Supplementband* **55**: 81–85.