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Knickpoint retreat and transient bedrock channel morphology triggered by base-level fall in small bedrock river catchments: The case of the Isle of Jura, Scotland

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ABSTRACT

A sudden drop in river base-level can trigger a knickpoint that propagates throughout the fluvial network causing a transient state in the landscape. Knickpoint retreat has been confirmed in large fluvial settings (drainage areas >100 km^2) and field data suggest that the same applies to the case of small bedrock river catchments (drainage areas <100 km^2). Nevertheless, knickpoint recession on resistant lithologies with structure that potentially affects the retreat rate needs to be confirmed with field-based data. Moreover, it remains unclear whether small bedrock rivers can absorb base-level fall via knickpoint retreat. Here we evaluate the response of small bedrock rivers to base-level fall on the isle of Jura in western Scotland (UK), where rivers incise into dipping quartzite. The mapping of raised beach deposits and strath terraces, and the analysis of stream long profiles, were used to identify knickpoints that had been triggered by base-level fall. Our results indicate that the distance of knickpoint retreat scales to the drainage area in a power law function irrespective of structural setting. On the other hand, local channel slope and basin size influence the vertical distribution of knickpoints. As well, at low drainage areas (~4 km^2) rivers are unable to absorb the full amount of base-level fall and channel reach morphology downstream of the knickpoint tends towards convexity. The results obtained here confirm that knickpoint retreat is mostly controlled by stream discharge, as has been observed for other transient landscapes. Local controls, reflecting basin size and channel slope, have an effect on the vertical distribution of knickpoints; such controls are also related to the ability of rivers to absorb the base-level fall.

1. Introduction

Bedrock rivers play an important role in the evolution of mountainous landscapes by controlling the rate of long-term erosion (Burbank et al., 1996; Hancock et al., 1998; Whipple and Tucker, 1999; Whipple, 2004). An increase in the rate of rock-uplift or eustatic adjustment may produce a drop in base-level, resulting in the formation and propagation of knickpoints (Whipple and Tucker, 1999; Snyder et al., 2002; Bishop et al., 2005; Jansen et al., 2011). Base-level fall knickpoints have been observed in a range of tectonic settings (Loget and van den Driessche, 2009), such as convergent plate margins (e.g., Snyder et al., 2000; Duvall et al., 2004; Crosby and Whipple, 2006; Harkins et al., 2007; Reinhardt et al., 2007), mildly tectonically active landscapes (Berlin and Anderson, 2007; Goldrick and Bishop, 2007), passive margins (e.g. Miller, 1991; Anthony and Granger, 2007; Frankel et al., 2007), and landscapes affected by glacio-isostatic rebound (Bishop et al., 2005). From these studies, it is known that landscape response to base-level fall is influenced by (1) the rate of knickpoint retreat through the fluvial network, and (2) the response of hillslopes to the pulse of knickpoint retreat-driven incision. Rates of knickpoint retreat depend upon stream discharge, sediment supply and substrate erodibility (Loget and van den Driessche, 2009). The response of hillslopes is not well understood, but steepening and increased landslide activity are often observed in the wake of retreating knickpoints, albeit with significant lag-times (Reinhardt et al., 2007). Rapid landscape response to base-level fall requires close coupling between hillslopes and channels; hence steady-state landscapes generally feature high bedrock incision rates (Hovius et al., 2000; Stolar et al., 2007). However, most landscapes exhibit a mismatch between the response of rivers and hillslopes, leading to a prevalence of transience; that is, a non-steady state condition (Whipple, 2001; Reinhardt et al., 2007).

Landscape transience has been evaluated by estimating bedrock incision rates (e.g. Harkins et al., 2007; Whittaker et al., 2007), detecting changes in sediment size (e.g. Attal and Lavé, 2006; Whittaker et al., 2010), recording variations in channel width (e.g. Amos and Burbank, 2007; Finnegan et al., 2007; Whittaker et al., 2007), and examining

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the genesis of hanging valleys (e.g. Wobus et al., 2006a; Crosby et al., 2007; Goode and Burbank, 2009). For the case of knickpoint retreat, the response has been predicted theoretically (Howard et al., 1994; Rosenbloom and Anderson, 1994; Whipple and Tucker, 1999), and simulated in flume experiments (e.g., Holland and Pickup, 1976; Gardner, 1983; Frankel et al., 2007). Field-based data have also confirmed the knickpoint retreat (e.g. Hayakawa and Matsukura, 2003; Bishop et al., 2005; Crosby and Whipple, 2006; Jansen et al., 2011). Empirically-based models point to drainage area, a surrogate of discharge, as a key control on knickpoint retreat rate (e.g., Bishop et al., 2005; Crosby and Whipple, 2006); however, the existence of a minimum threshold drainage area at which knickpoint retreat stalls is yet to be confirmed (Crosby and Whipple, 2006).

Knickpoint retreat may be slowed by resistant lithologies (e.g., Miller, 1991; Bishop and Goldrick, 2010), which may also have the effect of steepening the knickpoint face (e.g., Frankel et al., 2007), perhaps leading to collapse and further headward retreat of the knickpoint (Haviv et al., 2010). Motivated by the need to understand transience in bedrock rivers, we studied base-level fall-induced knickpoints and channel morphology in 17 small bedrock rivers (drainage areas <100 km^2) on the Isle of Jura, western Scotland (UK). We ask: (1) Does knickpoint retreat scale to drainage area? Finding a positive answer to that, we ask three further questions: (2) What is the effect of geological structure on knickpoint retreat? (3) Does the vertical distribution of knickpoints scale to drainage area? and (4) Are the reaches downstream of knickpoints fully adjusted to the base-level fall that triggered the knickpoint? We use Jura as a natural laboratory to investigate these questions.

2. Field setting and base-level history

2.1. Geomorphology and geological structure

Jura is a 368 km^2 island located just off the west coast of Scotland (Fig. 1A). The topography is hilly and notably steep on the uplands, with peaks rising to 780 m a.s.l. and annual precipitation exceeding 2.2 m (Met Office, 2010). The Neoproterozoic Jura quartzite crops out across the entire island except for minor areas in the southeast; all rivers studied here flow across the quartzite. NW–SE trending dolerite dikes cut the quartzite along fractures and apparently control the orientation and valley development of some streams. The Jura quartzite was deposited in a tidal shelf environment and exhibits no clear horizons to subdivide it (Anderton, 1976). The quartzite dips to the east–southeast, usually at 20° to 40° with a slight variation of 10° to 15° at the southern corner of the island (Anderton, 1976). Jura is elongated along-strike, meaning that the east coast rivers flow down-dip (‘dip-slope streams’) whereas the western ones are ‘scarp-slope’ streams that incise across the dip. The marked dip of layers to the east provides an excellent opportunity to examine the influence of structure on stream incision (Fig. 1A,C). Measurements of catchment morphometry reveal area–length exponents ranging from ~1.0 to ~0.5 (Fig. 1B). None of the streams has major tributaries, a condition that obviates the need to consider the issue of abruptly decreasing catchment area at tributary confluences (cf. Crosby and Whipple, 2006) in the analysis of knickpoint retreat.

2.2. Base-level history

On the west coast of Jura a raised marine bench occurs at ~32–34 m a.s.l. (Dawson et al., 1999). The emergence of this bench was driven by glacio-isostatic rebound following decay of the British–Irish Ice Sheet ~16 ka (Shennan et al., 2002; Chiverrell and Thomas, 2010). There are no available data for the surface uplift of Jura but on Islay, an island just south-east of Jura (Fig. 1A), estimated uplift rates are ~9 to 15 mm yr^{-1} at ~13 to 10 ka decreasing to ~2 mm yr^{-1} to the present (Firth and Stewart, 2000). If correct, such rates may approximate those of Jura and confirm that west coast islands have experienced a slowing of glacio-isostatic rebound over time (Lambeck, 1993, 1995). The rebound of Jura is also evidenced by spectacular ‘staircases’ of raised gravel beaches on its higher-energy west coast, from ~35 m a.s.l. down to a prominent bench surface at ~20 m a.s.l. (Dawson, 1991, 1993). The bench is fronted by an uplifted marine cliff, below which are further beaches that slope down to the modern beach (Dawson, 1984). Relative sea-level curves for northern Britain (Lambeck, 1993, 1995; Shennan et al., 2002, 2005) indicate that the ~35 m raised beach deposits emerged ~13 ka, which is confirmed by our unpublished cosmogenic ^{10}Be exposure ages of ~13.6 ka on the ~35 m a.s.l. beach (Bishop et al., in prep.). These beach deposits are consistent with the elevation and age of the Perth Shoreline (east coast of Scotland), which began to be uplifted ~13 ka (Jardine, 1982). The relative sea-level curves indicate very rapid postglacial emergence, as is also confirmed by Bishop et al.'s (in prep.) cosmogenic ^{10}Be exposure ages of ~13.6 ka on the ~20 m bench surface on Jura's west coast. We treat the emergence of the beaches at 13.6 ka as an essentially instantaneous surface uplift (base-level fall) of ~15 m. The formation of the 20 m a.s.l. bench likely reflects a period of reasonably constant relative sea-level when glacioisostatic rebound and eustatic sea-level rise were approximately matched.

3. Methods

3.1. Model of bedrock channel incision

A stream power-based model was used to evaluate the presence of disequilibrium in the Jura streams; that is,

$$E = KA^mS^n \tag{1}$$

where E is the erosional lowering rate of the river bed, K is a dimensional coefficient that aims to capture rock erodibility, A is drainage area, used as a proxy for stream discharge, and S is local channel slope (Howard and Kerby, 1983; Whipple and Tucker, 1999). Two formulations of the ‘stream power rule’ have been used to explore knickpoint retreat. In one formulation, bedrock erosion is proportional to stream power ($m = n = 1$), and in the other, bedrock erosion is proportional to channel bed shear stress, in which case $m = 0.3$ and $n = 0.7$ (Howard et al., 1994). These two formulations are of interest here because they describe the behavior of a propagating knickpoint in steady-state streams. In the shear stress case ($m = 0.3$; $n = 0.7$), knickpoints rotate as they propagate headwards, diffusing away, whereas for the stream power case, where A exerts more influence ($m = n = 1$), knickpoints are broadly maintained as they propagate headwards (Fig. 2). In the extreme case of $n = 0$ (i.e., slope exerts no influence on knickpoint propagation, which is therefore a function only of A), knickpoint morphology is maintained during propagation (e.g., Rosenbloom and Anderson, 1994; Weissel and Seidl, 1998; Whipple and Tucker, 1999; Crosby and Whipple, 2006). In short, the maintenance of knickpoints as they propagate implies that A is the dominant controlling factor in that propagation. Thus, for knickpoints triggered by glacio-isostatic rebound in east coast streams in Scotland, Bishop et al. (2005) found that estimates of m and n are 0.6 and 0.2, respectively, confirming the importance of catchment area (discharge) in determining knickpoint retreat rates and morphology.

It has been observed empirically that in many rivers channel slope decreases exponentially as drainage area increases, following a power law function (Flint, 1974; Kirby and Whipple, 2001):

$$S = k_sA^{-\theta} \tag{2}$$

where k_s is the channel steepness, and θ the channel concavity. A parallel approach to that of the stream power rule substitutes the distance from the divide (L) for A, following the power relationship