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Lithology and the evolution of bedrock rivers in post-orogenic settings: Constraints from the high elevation passive continental margin of SE Australia

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Abstract

Understanding the role of lithological variation in the evolution of topography remains a fundamental issue, especially in the neglected post-orogenic terrains. Such settings represent the major part of the Earth's surface and recent modelling suggests that a range of interactions can account for the presence of residual topography for hundreds of million years, thereby explaining the great antiquity of landscapes in such settings. Field data from the inland flank of the SE Australian high-elevation continental margin suggest that resistant lithologies act to retard or even preclude the headward transmission of base-level fall driven by the isostatic response to regional denudation. Rejuvenation, be it episodic or continuous, is 'caught up' on these resistant lithologies, meaning in effect that the bedrock channels and hillslopes upstream of these 'stalled' knickpoints have become detached from the base-level changes downstream of the knickpoints. Until these knickpoints are breached, therefore, catchment relief must increase over time, a landscape evolution scenario that has been most notably suggested by Crickmay and Twidale. The role of resistant lithologies indicates that detachment-limited conditions are a key to the longevity of some post-orogenic landscapes, whereas the general importance of transport-limited conditions in the evolution of post-orogenic landscapes remains to be evaluated in field settings. Non-steady-state landscapes may lie at the heart of widespread, slowly evolving post-orogenic settings, such as high-elevation passive continental margins, meaning that non-steady landscapes, with increasing relief through time, are the 'rule' rather than the exception.

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Bedrock channels are the ‘backbone’ of the landscape because they set much of the relief structure of tectonically active landscapes and dictate relationships between relief, elevation and denudation (Howard *et al*. 1994; Whipple & Tucker 1999; Hovius 2000). Bedrock channel evolution is thus the key to understanding landscape history and sediment flux: bedrock trunk channels provide the baselevel to which the whole drainage net and hillslope processes operate (Wohl & Merritt 2001). The last two decades have seen a blossoming of research on bedrock channels, reviving interest in, and building on, the fundamental research of Hack (1957) and Flint (1974) (Tinkler & Wohl 1998). This recent research has addressed a wide range of aspects of bedrock river morphology and processes, including: the distinction between debris-flow bedrock channels and fluvial bedrock channels (e.g., Stock & Dietrich 2003); the stream-power rule for bedrock river incision and its application in the numerical modelling of landscape evolution (e.g., Howard & Kerby 1983; Howard *et al*. 1994; Whipple & Tucker 1999; Tucker & Whipple 2002; van der Beek & Bishop 2003); refinement of the stream-power rule to take explicit account of sediment (e.g., Sklar & Dietrich 1998, 2001; Turowski *et al*. 2007); bedrock river long profile morphology and development (e.g., Bishop *et al*. 1985; Merritts *et al*. 1994; Whipple *et al*. 2000; Roe *et al*. 2002; Kirby *et al*. 2003; Kobor & Roering 2004; Brocard & van der Beek 2006; Goldrick & Bishop 2007); bedrock river knickpoint processes (Holland & Pickup 1976; Gardner 1983; Hayakawa & Matsukura 2003; Bishop *et al*. 2005; Crosby & Whipple 2006; Frankel et al. 2007); and bedrock river incision rates and processes (e.g., Bishop 1985; Burbank *et al*. 1996; Sklar & Dietrich 1998; Hancock *et al*. 1998; Hartshorn *et al*. 2002). The development of cosmogenic nuclide analysis (e.g., Bierman 1994; Bierman & Nichols 2004), with its capability of determining the exposure ages and/or erosion rates of bedrock surfaces, has been a fundamentally important breakthrough in studying bedrock channels (e.g., Burbank *et al*. 1996; Hancock *et al*. 1998; Brocard *et al*. 2003; Reusser *et al*. 2006).

Much of this research has had a focus (explicit or implicit) on tectonically active areas, where high rates of rock uplift, high rates of seismicity, and high (to extreme) rates of precipitation ‘drive’ bedrock incision (e.g., Burbank *et al*. 1996; Hovius 2000; Hartshorn *et al*. 2002; Dadson *et al*. 2003, 2004). Likewise, steady-state rivers, in which rock uplift is matched by river incision and/or catchment lowering, underpin numerical modelling of bedrock rivers (e.g., Whipple & Tucker 1999) and physical modelling of eroding landscapes (e.g., Bonnet & Crave 2003). Less attention has been paid in recent work to post-orogenic terrains, such as passive continental margins, that are not experiencing ongoing tectonically-driven rock uplift (Bishop 2007). The focus on steady-state tectonically active terrains and the relative neglect of post-orogenic terrains probably reflect the fewer complexities associated with the conceptualisation and numerical modelling of denudation of steady-state terrains (e.g., Whipple & Tucker 1999), and the impressive and eye-catching excitement of tectonically active terrains, where bedrock channel lowering may be physically measured in just one year (e.g., Hartshorn *et al*. 2002). On the other hand, the focus on orogenic belts has drawn attention away from understanding bedrock river processes and long-term landscape evolution in non-orogenic and post-orogenic areas, which constitute by far the majority of the Earth's sub-aerial surface area and which were the focus of research when geomorphology was first emerging as a discipline (Davis 1899, 1902).

Passive margin highland belts are perhaps the single post-orogenic terrain to have continued to receive widespread attention in the last two decades, especially in terms of low-temperature thermochronology (Moore *et al*. 1986; Dumitru *et al*. 1991; Brown *et al*. 2000a, 2000b; Persano *et al*. 2002, 2005), landscape evolution (e.g., Bishop 1985, 1986; Gilchrist & Summerfield 1990; Pazzaglia & Gardner 1994, 2000; Gunnell & Fleitout 1998, 2000; Cockburn *et al*. 2002; Matmon *et al*. 2002; Campanile *et al*. 2008), bedrock river evolution (e.g., Bishop *et al*. 1985; Young & McDougall 1993; Goldrick & Bishop 1995), and numerical modelling (e.g., Kooi & Beaumont 1994; Tucker & Slingerland 1994; van der Beek & Braun 1998; van der Beek *et al*. 1999, 2001). Baldwin *et al*. (2003) used numerical modelling to explore the factors responsible for landscape persistence and the timescales of post-orogenic decay of topography. Their work addresses the formerly-widespread viewpoint that landscapes cannot be much older than the Tertiary and are probably no older than the Pleistocene (e.g., Thornbury 1969), a viewpoint which is demonstrably incorrect in many post-orogenic terrains (Young 1983; Twidale 1998). Baldwin *et al*. (2003) showed that a 'standard' detachment-limited river incision numerical model predicts that orogenic topography will decay to 1% of the original topography at the channel head within only about 1-10 Myr of the cessation of orogenic activity. The addition of isostatic compensation (i.e., rock uplift in response to denudation) increases the decay time to 10-30 Myr, and a switch to transport-limited conditions extends this to 36-90 Myr. Finally, if a critical shear stress for erosion is introduced, along with stochastic variability of flood discharge, the timescale of post-orogenic decay of topography extends to hundreds of millions of years.

Baldwin *et al*. (2003) also identified, but did not assess, other factors that could control the rate of post-orogenic topographic decay, including the slowing of landscape evolution by resistant lithologies. In this mechanism, resistant lithologies control the relaxation time of the long profile in (re-)attaining some form of equilibrium (steady-state) long profile after perturbation. We focus on this issue here and ask the question: What role does lithology play in bedrock river morphology and landscape evolution in post-orogenic terrains? We examine this issue in the context of the post-orogenic setting of the Lachlan River catchment on the high elevation passive continental of southeastern Australia, by analysing the long profile morphology of the right- and left-bank tributaries of the upper Lachlan River, one of the major streams that drains the inland flank of the SE Australian highlands. Our long profile analysis is based on the DS form of the equilibrium bedrock river long profile (Goldrick & Bishop 2007).

Background: DS plots and long profile analysis and projection

The DS form of the equilibrium long profile plots the logarithm of a reach's slope against the logarithm of that reach's downstream distance; it is a slope-distance equivalent (hence 'DS') of the slope-area plot (e.g., Whipple & Tucker 1999) and takes the following form (Goldrick & Bishop 2007):

$$S = kL^{-\lambda} \quad \text{or} \quad \ln S = \gamma - \lambda \ln L \qquad 1$$

where *S* is channel slope, *L* is downstream distance, λ is a constant, *k* is a constant equal to $\frac{RI_{grade}}{il}$, and γ is equal to ln *k*. *R* is some measure of lithological resistance to erosion, I_{grade} is the equilibrium rate of channel incision, *i* is a constant that describes the proportion of stream power that is expended in incision, and *l* is a constant.

A key feature of the DS form of the long profile is that, in principle, transient channel steepening, such as a knickpoint propagating in response to base-level fall, can be distinguished from equilibrium channel steepening in response to a more resistant lithology. The latter (equilibrium steepening on more resistant lithology) is indicated on the DS plot by a parallel shift in the plot, whereas a disequilibrium knickpoint plots as disordered outliers on the DS plot (Goldrick & Bishop 2007).

Long-profile disequilibrium is commonly generated by a drop in base-level and is resolved over time by the upstream passage of a wave of incision (a knickpoint), either as a retreating step that maintains the knickpoint's height and form, or as a step that rotate backwards by 'inclination' or 'replacement' and diffuses away (Gardner 1983; Bishop *et al*. 2005; Crosby & Whipple 2006; Frankel *et al*. 2007). At any time between the initiation of a persistent, retreating knickpoint and its arrival at the stream's headwaters the stream can be thought of as consisting of three reaches: the upstream reach, which is yet to be affected by the rejuvenation and may therefore remain graded to the previous base-level; the over-steepened reach comprising the knickpoint; and the downstream reach that is graded to the new base-level. In order to compare present and past base-levels (and thereby to quantify, for example, the amount of trunk stream incision as a result of base-level driven by surface uplift), it is necessary to reconstruct the pre-rejuvenation profile using the upstream (unrejuvenated) reach and to project that reconstructed profile downstream.

With the exception of theory-free curve fitting (e.g. Jones 1924; Hovius 2000), long profile reconstructions and projections used to compare present and ancient profiles and to identify changes in base-level have been based on *a priori* models of the equilibrium long profile. We here use the DS form for such projections, re-arranging Goldrick & Bishop's (2007) equation 2 to yield:

$$\ln(H_0 - H) = \ln(\frac{k}{1-\lambda}) + (1-\lambda)\ln L \qquad 2$$

It follows from equation 2 that the value of λ can be determined from the slope of a least-squares linear regression of the log of downstream distance versus the log of the fall (H_0 - H). Once the value of λ has been determined, the value of k can be determined from the intercept (b) of the same regression:

$$b = \ln\frac{k}{1-\lambda} \quad \text{or} \quad k = (1\text{-}\lambda)e^b \qquad 3$$

The difficulty in solving for λ and k comes from the need to know the value of H_0, which is not the same as the elevation of the divide (Goldrick & Bishop 2007). This difficulty can be overcome by solving equations 2 and 3 iteratively. Estimates of H_0 can be substituted into equation 2 to yield values of λ which are used in turn to derive values of k. These can then be substituted into equation 2 to give a mathematical description of the long profile. The goodness of fit of each description so generated can be evaluated by comparison with the observed profile using the criterion of the standard error of the estimate of elevation, se_H (Goldrick & Bishop 2007). This procedure is repeated and the best estimate of H_0 determined by converging on that value which yields the best fit between the modelled and observed long profile.

The downstream projection of the equilibrium reach above a knickpoint to estimate the amount of base-level fall is illustrated by a hypothetical example in Fig. 1, which shows a stream with a knickpoint at A and a profile reconstruction and projection based on the DS model of the graded reach upstream of A. The magnitude of the 40 m base-level fall that generated the knickpoint at A can be estimated by projecting the reconstructed equilibrium profile above A to the stream's base-level (Goldrick & Bishop 1995, 2007). The difference between the elevation predicted by the reconstructed profile and the actual elevation at any point along the stream is here called H^*, and the value of this difference at base-level at the downstream limit of the stream is designated H^*_D. H^*_D is an estimate of the magnitude of stream incision in response to the base-level fall. The DS model prediction of the amount of incision/base-level fall at the downstream limit of the channel in Fig. 1 is 35 m whereas the 'true' value is 40 m. The discrepancy of 5 m between the actual base-level drop and the value of H^*_D given by the DS model is mainly a result of the variability resulting from the standard error of 2 m associated with the elevation data. In consequence the calculated values of λ and k are 0.89 and 48, respectively, rather than the "true" values of 0.9 and 50.

The confidence interval for H^*_D is given by:

$$t\sqrt{\frac{\sum_{j=1}^{n}\left(H_j-\hat{H}_j\right)^2}{n-2}\left(\frac{1}{n}+\frac{\left(L_j-\overline{L}\right)^2}{\sum_{j=1}^{n}L_j^{\,2}-n\overline{L}^2}\right)} \tag{4}$$

where *j* is the point for which the confidence interval is to be calculated; *n* is the total number of points; H_j and $\hat{H}_j$ are the observed and predicted elevations at *j,* respectively; L_j is the downstream distance of *j*; and, $\overline{L}$ is the mean value of *L* (after Ebdon 1985 p.117). A 95% confidence interval for the example in the preceding paragraph gives $H^*_D = 34 \pm 7$ m, a range which encompasses the actual base-level drop of 40 m. This range is large, relative to the size of the base-level drop, an outcome that emphasises the sensitivity of long profile projections to errors in the input data, especially when a projection is made over long distances.

The projection of reconstructed profiles to the downstream limit of the stream enables the estimation of total base-level fall to which the stream has responded by incision and knickpoint propagation. Profile projection can also be used to quantify the height of a knickpoint, for which a projection is made from the (graded) section above the knickpoint to the upstream limit of the (graded) section downstream of the knickpoint. In the case of the stream in Fig. 1, projection to the upstream limit of the graded reach below A estimates the height of the knickpoint to be 41 ± 2 m.

Denudational isostatic rebound of the Lachlan catchment and long profile adjustment

The Lachlan River catchment drains the inland (western) side of Australia's southeastern Highlands on the Tasman passive continental margin (Fig. 2). The Lachlan is bounded to the east by the continental divide, to the west by the Cenozoic intracratonic Murray Basin, and to the north and the south by catchments of the Macquarie and Murrumbidgee Rivers, respectively. Downstream of Mandagery Creek, the Lachlan flows across a low gradient interior depositional lowlands before terminating in a very low gradient inland swamp (O'Brien & Burne 1994). The Murray-Darling system has its base level in the Southern Ocean some 1000 km downstream of Mandagery Creek, the downstream limit of the study area here (Fig. 2). The distance across the low gradient, swampy, depositional lowlands to the base-level in the Southern Ocean means that Cenozoic eustatic sea-level fluctuations are unlikely to have rejuvenated the Lachlan long profile.

Basaltic lavas were erupted into the Lachlan catchment at various times in the Tertiary, providing data on the drainage net at the times of eruption (Bishop *et al*. 1985; Bishop 1986). K-Ar ages on the basalts, coupled with reconstructions of the valleys into which the basalts flowed, demonstrate that much of the relief and many of the drainage systems of the SE Highlands, including the drainage network of the Lachlan, were established by the Early to Middle Tertiary

(Wellman & McDougall 1974; Bishop *et al.* 1985; Bishop 1986; Young & McDougall 1993). These data notwithstanding, the history of the SE Highlands in New South Wales remains somewhat controversial in detail (Bishop & Goldrick 2000). It has been widely accepted that highlands surface uplift was related to Tasman margin extension and rifting around 100-90 Ma (Veevers 1984; Wellman 1987; Ollier & Pain 1994; O'Sullivan *et al*. 1995, 1996), which may have 'rejuvenated' a pre-existing topography (e.g., Persano *et al*. 2005). Few find evidence for active Neogene uplift of this portion of the highlands, notable exceptions being Wellman (1979a 1987), van der Beek *et al.* (1995) (but see van der Beek *et al*., 1999), and Tomkins & Hesse (2004) for the catchment to the north. Recent interpretations of widespread neotectonics in Australia (see Quigley *et al.* this volume) do not seem relevant to the Lachlan, however, where the known rates of catchment-wide Neogene denudation of the Lachlan (Bishop 1985) and the isostatic equilibrium of the highlands (Wellman 1979b) mean that denudational isostasy can account for the observed Neogene rock uplift in the Lachlan (Bishop & Brown 1992, 1993).

The strongest evidence for, and constraint upon, rock uplift along the Lachlan's highlands margin, whatever the mechanism for that uplift is, comes from the 12 Myr old Boorowa basalt (informal name), at the inland edge of the highlands (van der Beek & Bishop 2003) (Figs 2 and 3). This basalt overlies fluvial gravels at an elevation 65 m above the modern Lachlan River, indicating a long-term incision rate of about 5 m Ma^{-1}. This incision must approximate the amount of relative base-level fall at the highlands margin, which forms base-level for the bedrock Lachlan catchment studied here. In the absence of relative base-level fall at the highlands margin, incision rates at the highlands margin should be close to zero. Indeed, where the base-level is the margin between an eroding highlands and a depositional basin, as it is in the case of the Lachlan, one might expect an absence of surface uplift to be accompanied by aggradation of the base level and a gradual up-catchment shift in the margin by back-filling of the highlands valley. In fact, the Lachlan upstream of the Boorowa basalt is characterised by an incised bedrock gorge with terraces up to 35m above the modern river (Bishop & Brown 1992; van der Beek & Bishop 2003). As noted above, the known rates of denudation of the Lachlan catchment (Bishop 1985) mean that such base-level fall is most reasonably interpreted as the result of isostatic response to catchment-wide denudation (Bishop & Brown 1992, 1993).

The upper Lachlan

Bedrock in the Lachlan catchment consists of meridional belts of Palaeozoic granites and quartz-rich metasediments and silicic volcanics (Fig. 3). The only post-Palaeozoic rocks are the thin Tertiary basaltic lavas noted above and sediments scattered throughout the catchment and sporadic patches of Quaternary alluvium along drainage lines. Even in the areas of apparently more continuous alluvium, the channel bed appears to be everywhere formed in bedrock.

Sediment that locally blankets the channel bed reflects catchment disturbance and sediment mobilisation over the last century or so, following the introduction of sheep farming by European settlers. Thus, the entire drainage net is essentially formed in bedrock.

The narrow flows of Early Miocene Bevendale Basalt and Wheeo Basalt lie adjacent to the modern streams in the upper Lachlan (Fig. 4) (Bishop 1984). These flows thus provide a maximum time limit for stream evolution and have been used several times for the study of evolution of bedrock river long profiles (Bishop *et al*. 1985; Stock & Montgomery 1999; van der Beek & Bishop 2003) . The flows are thin valley-fill basalts, preserving the ancient valleys into which they flowed (Fig. 4). Field relationships, such as the uniform elevation of the upper surfaces of the highest flow remnants, well below the interfluves, and the lack of any basalt remnants at higher elevations, demonstrate that the lavas did not fill their valleys and overtop the interfluves (Bishop *et al.* 1985; Bishop 1986). Post-basaltic stream incision has resulted in relief inversion so that flows now persist as hilltop remnants and ridges that discontinuously preserve the E Miocene valleys.

The basement geology of the upper Lachlan consists of Ordovician quartz-rich slate and greywacke of the Adaminaby Group surrounding gneissic granites of the Wyangala Batholith (Fig. 4); all of these lithologies show meridional foliation. A well-defined band of locally high relief roughly coincides with the western parts of the granite, which is drained by the upper Lachlan's eastern (right-bank) tributaries, whereas the Ordovician metasediments drained by the lower-relief western (left-bank) tributaries are associated with lower relief due to their lower resistance to erosion. A prominent ridge of resistant contact metamorphic hornfels extends for a considerable length along the western contact of the Ordovician sediments with the granite (Fig. 4). The modern Lachlan Rivers lies a few kilometres to the west of this ridge, with right-bank tributaries flowing westward from the granite through the ridge; basalt remnants demonstrate that the same situation pertained in the E Miocene.

Goldrick & Bishop's (2007) preliminary DS analysis of the right- and left-bank long profiles showed that the right-bank steepened reaches on and about the resistant hornfels and granite lithologies are disequilibria: the DS profiles of the right-bank tributaries exhibit two outliers associated with the resistant hornfels and the granite, with these DS outliers separating essentially linear DS reaches that we interpreted to be in equilibrium (graded). The left-bank tributaries are much less perturbed by disequilibria and generally grade smoothly to the trunk stream. We now examine this asymmetry further, using the DS plots.

The upper Lachlan's right-bank (eastern) tributaries

The right-bank tributaries rise at the continental divide, descending the eastern flank of the upper Lachlan's valley to the Lachlan; in doing so, they cut through the hornfels ridge and across the

Bevendale Basalt. The formerly continuous valley-filling basalt thus now crops out as a series of hill-top cappings isolated from each other by the tributaries.

We digitised the ‘blue lines’ on the New South Wales Central Mapping Authority’s 1:50,000 topographic sheets (20 m contour interval) to generate long profiles and DS plots of the perennial streams of the Lachlan’s highlands drainage net (Goldrick & Bishop 2007). As we noted above, the right-bank tributaries generally exhibit linear (equilibrium) DS profiles separated by knickpoints. Streams 59 and 60 are useful exemplars of these long profile characteristics; they are twin lateral streams to the major east-west remnant of the lava that flowed to the trunk Lachlan from the continental divide in the E Miocene (Bishop 1986) (Figs 5 and 6).

Stream 59 rises on the Adaminaby Group metasediments and flows westward across the Wyangala Batholith below that east-west basalt cap. It cuts through the hornfels ridge and the line of the basalt before joining the trunk stream. The DS plot shows that, with the exception of a small discontinuity, stream 59 is graded where it flows across the Adaminaby Group sediments (the standard error of the estimate of elevation, se_H = 1.0 m). Downstream, on the Wyangala Batholith, the DS plot is marked by high variability and several peaks or knickpoints, including one where the stream crosses the hornfels ridge. The long profile (Fig. 6a) shows that a downstream projection based on the upstream graded reach projects approximately to the base of the basalt where the stream cuts through it. At that point, the projection yields a value of H* equal to 60 ± 10 m.

Stream 60 is almost the mirror image of Stream 59, rising on the Adaminaby Group metasediments and flowing westward across the granite, through the hornfels ridge, between basalt remnants and then joining the trunk Lachlan. The DS plot (Fig. 6b) shows a low-variability upstream reach that appears graded but the value of se_H is high at 2.8 m. Downstream of that graded reach there is high variability in the DS plot including two well defined knickpoints, with the second of those located immediately upstream of the hornfels ridge. These knickpoints are clearly evident on the long profile (Fig. 6b). Downstream projection of the graded reach is graded approximately to the basalt, with a value of H* equal to 83 ± 13 m where it crosses the line of the basalt.

The profiles of all the eastern right-bank tributaries are broadly similar to Streams 59 and 60, being characterised by equilibrium reaches (low variability on the DS plot) interspersed with high-variability disequilibrium reaches. All the profiles show knickpoints near their junction with the trunk stream. The values of H* associated with these knickpoints are in broad agreement ranging from 60 ± 10 to 83 ± 13 m, save for the anomalously low, and currently inexplicable, H* = 16 ± 4 m for Stream 57 (Table 1a). Streams 57 and 58 also have knickpoints further upstream for which the values of H* at their confluences with the trunk are 122 and 143 ± 10 m, respectively (no uncertainty is given for the projection of the upstream graded reach in Stream 57 because it is based on only three points) (Table 1b).

The spacing of the elevation contours of the base of the basalt shows that the north-south portion paralleling the Lachlan preserves the E Miocene trunk and the east-west portion preserves an E Miocene tributary, with a notably pronounced steepening upstream of the confluence with the trunk (Fig. 4). The sub-basaltic elevation data are sufficiently complete to allow the construction of a DS plot and long profile for the E Miocene channel down which it flowed. This projection assumes that the headwaters of the basalt-filled channel must have been at least as far east as the easternmost outcrop of the valley-filling basalt (asterisk in Fig. 5) and that the headwaters can have been no further east than the modern continental divide, because the position of the continental divide has been stable since pre-Miocene times (Bishop 1986).

The Miocene basalt-filled channel's DS plot (Fig. 7) shows a marked terminal knickpoint and a steep, but nonetheless linear reach upstream of that. The best fit for a reconstruction of the long profile based on that linear upstream reach is achieved when the Miocene divide is presumed to be at the eastern limit of the valley-filling basalt (asterisk in Fig. 5), but this is a rather poor fit with se_H equal to 7.5 m. Projecting this profile yields a value of H* equal to 103 ± 30 m where the tributary basalt flow meets the Miocene trunk Lachlan at the elbow in the flow adjacent to the modern Lachlan. In short, and notwithstanding the uncertainties, the Miocene tributary, like most of the modern tributaries, was characterised by considerable disequilibrium and a steep fall to the Miocene trunk.

The knickpoint steepenings, both modern and Miocene, are broadly spatially associated with granite and/or hornfels, but there is no clear and precise relationship between steepening and geology in general, nor between the terminal knickpoints and the hornfels ridge in particular. Several streams have graded reaches which cross the geological boundaries, while others have knickpoints on both the granite and the metasediments. Only in the case of Stream 59 is the terminal knickpoint coincident with the hornfels ridge. The terminal knickpoints of Streams 60 and 58 are upstream of the hornfels ridge on the granite, and the terminal knickpoints of Stream 57 and the Miocene tributary are downstream of the hornfels ridge on the metasediments. The clearest relationship that does emerge from the long profiles analysis is that those streams that flow across the path of the Bevendale Basalt have an upstream reach that is graded approximately to the base of the basalt at the point where they cross the basalt.

The upper Lachlan's left-bank (western) tributaries

Streams 67 and 69 are tributaries of Stream 68 (Fig. 4). The headwaters of Stream 67 flow across granitic rocks of the Wyangala Batholith while downstream reaches flow across metasedimentary rocks of the Adaminaby Group. The headwaters of Stream 68 are also on the granite and in its lower reaches it flows across Adaminaby Group metasediments. Stream 69 flows entirely across metasedimentary rocks of the Adaminaby Group. The long profiles and DS plots of Streams 67 and 68 (Fig. 8) show that both streams are graded throughout most of their lengths except for the presence of a relatively small, but well defined, knickpoint on each stream. The values of H* at the confluence of Streams 67 and 68 are 30 ± 2 m (se_H = 0.52 m) and 25 ± 14 m (se_H = 1.16 m), respectively. Despite the very large confidence interval associated with the projection of Stream 68, due to the long distance over which it has been projected, the values of H* are very similar suggesting that the knickpoints are the result of base level change rather than lithological variation (cf. Goldrick & Bishop 1995, 2007). The value of H* for Stream 68 upstream of the junction with the Lachlan is 26 ± 17 m, about 50 m less than typical values for the eastern tributaries, and the projected profile lies approximately 40 m below the basalt.

The DS plot of Stream 69 shows that the stream is not far from graded (se_H = 2.6 m) but there is a suggestion of a discontinuity which does not appear to be associated with a lithological change (Fig. 9a). A projection of the upstream reach of Stream 69 (se_H = 1.27 m) to the confluence with Stream 68 returns a value of H* = 16 ± 5 m (Fig. 9a). This is similar to the value of H* for Stream 68 itself at this point (25 ± 14 m) so that there is a suggestion that the discontinuity along Stream 69 has a similar baselevel-fall origin to the knickpoints on Streams 67 and 68, but the steepening on the Stream 69 is more a diffuse 'knickzone' than a discrete knickpoint.

In summary: the western tributaries of the upper Lachlan River show some evidence of disequilibrium steepening unrelated to lithology, but the degree of steepening along these streams and the corresponding values of H* (Table 2) are much less than those in the eastern (right-bank) tributaries. Several of the tributaries have comparable values of H* at their junctions lending support to the proposal that this steepening is due to a relative fall in base level.

Synthesis

The eastern (right-bank) tributaries are characterised by marked steepening, especially in their distal reaches, whereas the western (left-bank) tributaries have much smoother profiles. The southern tributaries are not treated in detail here, but they are transitional between these two groups with Streams 61 to 63 similar in form to the eastern tributaries and Streams 64 to 66 similar in form to the western tributaries. The obvious and simplest explanation for the pronounced east-west asymmetry in the tributary long profiles is the lithological influences exerted on the eastern tributary long profiles by the granite/hornfels combination and the basalt. The western tributaries flow mainly across Adaminaby Group rocks of moderately low resistance whereas the eastern tributaries flow across a variety of rock types including the more resistant granites of the Wyangala Batholith and the hornfels ridge. The eastern tributaries also all cut through the Bevendale Basalt upstream of their junctions with the trunk stream. Finally, and most importantly, all the eastern tributaries have a reach that is graded approximately to the base of the Bevendale Basalt where the tributaries cross the line of the basalt.

A major perturbation was introduced into the drainage net by the Bevendale Basalt flowing down the east-west tributary and then northward along the Miocene trunk Lachlan. The fact that all the eastern tributaries have a reach that is graded approximately to the base of the Bevendale Basalt suggests the evolutionary history presented schematically in Fig. 10. After the eruption, a new trunk stream formed on the western edge of the valley basalt so that the western tributaries maintained an unimpeded path to that new trunk whereas the eastern tributaries flowed across the newly emplaced (and resistant) valley basalt to join the trunk.

The basalt in the trunk would no doubt have introduced some disequilibrium into the profiles of the western streams (flattening them) but in the absence of resistant rocks it is likely that the new trunk and the western tributaries re-attained equilibrium early after the eruption. Disruption of the eastern tributaries must have been considerably greater. The re-establishment of a graded landscape east of the trunk stream would have been inhibited by the greater resistance of the granite, hornfels and the basalt. The influence of the latter was particularly important, because the eastern tributaries' low gradients across the top of the basalt, relative to the basalt's lithological resistance, would have led to a differential between the rate at which the trunk and its western tributaries were incising and that at which the eastern tributaries were incising into the basalt. If the attainment of grade in detachment-limited, post-orogenic settings, such as this, is a 'bottom-up' process (Bishop 2007) then the basalt would have acted as a local base level with the upstream reaches of the eastern tributaries becoming graded to it. Differential incision rates would have resulted in a steepening of the gradients between the new trunk stream and the basalt (Fig. 10 Post-eruption 2).

Over time, the trunk would have continued to incise at a faster rate than the tributaries on, and upstream of, the basalt so that the height differential between the two would have increased as would have the gradients across the basalt. Throughout this stage (Fig. 10 Post-eruption 3), the basalt would have acted as a temporary local base level for the eastern tributaries. Eventually these tributaries would have cut through the basalt removing the temporary base level provided by the basalt and triggering the upstream migration of a knickpoint (Fig. 10 Post-eruption 4). This stage, which represents the present state of the upper Lachlan drainage net, is characterised by a downstream disequilibrium reach and an upstream reach which is graded approximately to the level of the basalt remnant at the time when it was breached (i.e., the base of the flow).

This model offers the most likely explanation for the gross morphology of the tributaries of the upper Lachlan catchment with the western tributaries essentially graded to the modern trunk with only minor disequilibrium, and the eastern tributaries characterised by marked downstream disequilibrium steepening and an upstream reach that is graded approximately to the base of the basalt. The data confirm that resistant lithologies can act a temporary local base levels and retard the re-establishment of equilibrium, influencing relaxation times of streams so that resistant reaches can temporarily isolate upstream reaches from the effects of downstream perturbations.

Numerical simulation of the influence of lithology on stream evolution

This interpretation of the role of the basalt in delaying the headward transmission in the right-bank tributaries of base-level fall in the trunk was tested using a 1D finite difference stream evolution simulation based on the DS form, in which the amount of incision at each iteration is determined by the variables k, λ, L, i and S (see eq. 1 above, and Goldrick & Bishop (2007)). The modelled long profile consists of 100 reaches: reaches 1 to 72 and 89 to 100 have a lithological resistance corresponding to R = 50, and reaches 73 to 88 have a doubled lithological resistance of R = 100. Each point, x, in the long profile is lowered according to the formula:

$$incision = \frac{il}{R} L_x^{\lambda} \left(\frac{H_x - H_{x+1}}{L_{x+1} - L_x} \right) \qquad 2$$

where H_x, H_{x+1}, L_x and L_{x+1} are the elevations and downstream distances, respectively of x and x+1. At each iteration the amount of incision is calculated for each reach before the elevations are adjusted.

The development of the profile in the “post-eruption” period is recorded in Fig. 11. Fig. 11a and b depict the immediate post-eruption profiles of the tributary, which can be divided into three reaches. The reach downstream of the basalt is graded to the trunk stream which acts as a base level and this reach continues to incise at the same rate as the trunk, at the long-term denudation rate driven by denudational isostatic rebound. The reach flowing across the basalt is in disequilibrium because its gradient is low relative to the erosional resistance of the basalt, and it therefore incises at a slower rate than the trunk and the downstream reach. The reach upstream of the basalt is also no longer in equilibrium because the basalt is acting as a temporary local base level that is lowering at a slower rate than the base level to which this reach was previously graded.

After 25000 iterations (Fig. 11c and d), a clear knickpoint, indicated by a peak on the DS plot, has formed at the downstream edge of the basalt as a result of the rate differential between incision on the country rock and on the basalt. This incision rate differential results in a decrease in the gradients immediately upstream of the basalt, as indicated by low values on the DS plot and the flattening of the long profile immediately upstream of the basalt. After 50000 iterations the knickpoint has eroded through the downstream edge of the basalt and has become more pronounced (Fig. 11e and f). At the same time gradients upstream have decreased as an increasingly extensive reach has become graded to the upstream edge of the basalt.

75000 iterations see an even more pronounced knickpoint and the low gradient reach has become more extensive. This pattern continues until, at 125000 iterations, the stream is at the point of breaching the basalt (Fig. 11i and j). After the basalt has been breached, the knickpoint begins to migrate more rapidly upstream and declines ('lies back') so that the peak on the DS plot is less marked and knickpoint becomes laterally more extensive. That is, the knickpoint evolves from a discrete knickpoint into a more diffuse knick-zone. The form of the stream after 150000 iterations (Fig. 11k and l) is similar to the modern forms of the eastern tributaries of the upper Lachlan with an upstream reach graded approximately to the base of the basalt, and a knickpoint between this reach and the trunk stream.

The simulation is consistent with the hypothesis that basalt (or any other resistant rock) acting as a temporary local base level can retard the development of the upstream reaches, induce the formation of a knickpoint, and produce the type of profile that is typical of the eastern tributaries.

Discussion

The 'engine' for landscape evolution in the post-orogenic setting of the Lachlan River is denudational isostatic rebound: denudational unloading of the uplands, at rates of ~5 m Myr^{-1} (Bishop 1985), triggers the isostatic response of rock uplift, which in turn triggers drainage net rejuvenation at the drainage net's base-level (in this case the inland edge of the highlands, where the Lachlan flows onto the sedimentary fill of the continental interior) (Bishop & Brown 1992). For reasons that are not yet fully clear, but which likely reflect discontinuous movement on the highland-edge fault(s) on which the denudational isostatic rebound is accommodated, this rejuvenation is evidently discontinuous. Each denudational isostatic uplift 'event' presumably triggers a steepened reach where the Lachlan leaves the highlands around Cowra, and this steepened reach propagates as a knickpoint, headwards along the trunk stream and progressively through the drainage net. The contrast in the long profiles of the upper Lachlan's eastern (right-bank) and western (left-bank) tributaries shows that resistant lithologies act to slow knickpoint propagation, whereas non-resistant lithologies allow rejuvenation to be readily propagated headwards and for graded (equilibrium) long profiles to be re-established (e.g. Fig. 9). There are no indications as to the magnitude of each of these rock uplift events, but if the altitudinal separation of each of the terraces in the bedrock reach upstream of the highlands edge (Bishop & Brown 1992) is indicative of the magnitude of the uplift, then each is very substantial. That said, there may be a climatic component to the generation of these terraces and, in any event, the key issues are the triggering of knickpoints at the highlands edge by rock uplift of unknown magnitude as a result of denudational rebound, the propagation of those knickpoints through the drainage net, and the retardation of these knickpoints by resistant lithologies. The values of H^* in the upper Lachlan right-bank tributaries that flow across resistant lithologies integrate all the rebound events for a given time interval.

This lithological retardation of knickpoint propagation is independent of the eruption of basalt into the upper Lachlan's drainage net, as confirmed by the knickpoints in the long profile into which the basalt was extruded 21 Ma (Fig. 7). In other words, in the absence of the basalt, the lithological resistance of the granite and hornfels still acts to retard knickpoint propagation. That conclusion is confirmed by the fact that streams 57 and 58 also have upstream knickpoints for which the graded reaches above the knickpoints have values of H^* of 122 and 143 $\pm$ 10 m at their confluences with the trunk (see above). Those graded reaches project to elevations above the basalt and so must pre-date that basalt. As shown above (Fig. 7), projection to the trunk stream of the long profile of the basalt-filled east-west tributary yields $H^* = 103 \pm 30$ m and so the graded reaches with $H^* = 122$ and 143 m in streams 57 and 58 are likely to reflect long profile equilibria prior to 21 Ma, given that they project to a base-level in the trunk that is 20-40 m above that to which the basalt-filled E Miocene tributary long profile projects.

The perturbation caused by the basalt in the trunk stream is, in one sense, a special case, but it has general implications in enabling us (i) to use the instantaneous 'injection' of a resistant lithology into the long profile to assess the role of resistant lithologies in retarding knickpoint propagation, and (ii) to place that retardation in a dated time-frame. The basalts have clearly retarded the propagation of post-basaltic rejuvenation into the right-bank tributaries, with the implication that the reaches below the lowermost knickpoint in those right-bank tributaries are still adjusting to post-basaltic rejuvenation (unlike the left-bank tributaries which have long ago accommodated the earlier post-basaltic rejuvenation). That is, H^* in the left-bank tributaries is much less than the 50-80 m of post-basaltic incision along the trunk (which approximates the right-bank H^* values of 60-80 m – Table 1). An important implication of that interpretation is that the left-bank tributaries must be incising at approximately the same rate as the trunk Lachlan whereas the right-bank tributaries must have a range of incision rates along their lengths: between the trunk and the right-bank tributaries' first knickpoint, incision rates must range from the Lachlan's rate in the tributaries' downstream reaches, to (much?) higher rates in the knickpoint reaches, where the delayed rejuvenation is still being accommodated. In the graded reaches of the right-bank tributaries above those downstream-most knickpoints, incision rates are set by the channels' discharges, sediment fluxes and gradients and not by the trunk stream's rate of incision. In each case, the knickpoint in effect disconnects the graded reach and the trunk Lachlan.

More generally, the data reported here demonstrate the central role that resistant lithologies play in slowing landscape evolution in post-orogenic settings in which landscape response to rock uplift is via bottom-up processes of headward propagating rejuvenation (cf. Baldwin *et al*. 2003). The knickpoints on resistant lithologies in the Lachlan's eastern (right-bank) tributaries mean that the upper catchments are not lowering at the same rate as the trunk stream and the left-bank tributaries, and that the relief of the catchment must be increasing. This conclusion is confirmed by the fact that the headwaters of the modern streams rise on the E Miocene basalts whereas these modern streams have incised below the basalts (Fig. 4) (Bishop *et al*. 1985). We note that the presence of the basalts demonstrates the reality of increasing relief and places a time-scale on that relief increase, but it is not a pre-requisite for that relief increase.

The numerical modelling of Baldwin *et al*. (2003) highlights the role of transport-limited conditions in extending the 'life' of post-orogenic terrains to tens and hundreds of millions of years. Those transport-limited conditions mean that river beds consist of a layer of sediment covering the bedrock of the channel bed, shielding it from erosion (cf. Sklar & Dietrich 1998, 2001). In such transport-limited situations, rivers rarely 'interact' with the substrate and, in effect, do not 'feel' lithological variation (e.g. Brocard & van der Beek 2006). We do not envisage that situation to be the case in the upper Lachlan, which is characterised by low to very low rates of denudation and sediment flux (cf. Bishop 1985). The knickpoint reaches are not mantled by sediment and so their low rates of erosion are due to a combination of factors including their lithological resistance and the low rates of sediment flux for abrading the bed. In other words, detachment-limited conditions must also be incorporated, and may be a central element, in explanations of the longevity of post-orogenic terrains.

We envisage that the knickpoints are slowed and caught on the steeply dipping hornfels and jointed and foliated granites of the upper Lachlan as has been documented in field and laboratory observations by Frankel *et al.* (2007). In their experiments, a knickpoint on steeply dipping resistant lithologies evolved by a combination of parallel retreat on the knickpoint face and vertical channel incision on the knickpoint 'top' (though it is unclear whether this channel incision on the knickpoint top is the commonly observed draw-down effect on knickpoints – Haviv *et al*. (2006)). Whatever its precise nature, that vertical channel incision must limit the height that knickpoints can attain, and, in any event, knickpoints cannot continue to grow indefinitely because knickpoints of extreme heights will not be stable. The Lachlan data show that knickpoints eventually cut through resistant lithologies and continue their headward propagation. Breaching of the basalts reflects the finite vertical thickness of the basalt (Fig. 10) but the presence of more than one knickpoint in some of the right-bank tributaries (e.g., streams 57 and 58), with the graded reaches above the upper knickpoints projecting to elevations above the basalt-filled E Miocene channel, confirms that knickpoints do propagate through the hornfels and into the granites.

Bishop *et al*. (1985) and Young & McDougall (1993) observed that the passage of the whole of the Neogene (and almost certainly longer) has not resulted in anything that approaches planation of the post-orogenic landscape of SE Australia. Quite the opposite, the data here show increasing relief, much as hypothesised by Crickmay (1974, 1975) and Twidale (1976, 1991), who has been a major champion of Crickmay's "hypothesis of unequal activity". This hypothesis proposes that fluvial erosional energy is progressively concentrated in large river valley bottoms and that lower erosion rates in the smaller rivers of the upland areas mean that denudation of upland areas slows (cf. Bishop 2009). Incision of major channels mean that they become de-coupled from slopes. That is, relief amplitude must increase as rivers continue to incise and upstream areas erode more slowly (Crickmay 1975; Twidale 1976). Formalising this approach in a model of landscape evolution involving increased and increasing relief amplitude (and going beyond the relative sizes of trunk and tributary streams proposed by Crickmay as the key issue), Twidale (1991) highlighted the role of resistant lithologies, structure and different groundwater conditions throughout a drainage basin as the central factors responsible for increasing relief. As Twidale argued: "water … is concentrated in and near major channels, for, once a master stream develops, not only surface water but subsurface drainage too gravitates towards it … [U]plift induces stream incision and water-table lowering, leaving high plains and plateaux perched and dry" (Twidale 1998 p.663). The retardation of knickpoints on resistant lithologies adds a further dimension to that argument, as does the lack of discharge and sediment supply to incise the bedrock channels and these knickpoints.

Conclusion

The post-orogenic landscape of SE Australia enables the clarification of several major issues in long-term landscape evolution, not least because of its excellent evidence of landscape evolution as recorded by Cenozoic valley-filling basalts. These basalts provide chronologically rigorous evidence of Cenozoic landscape character, long-profile morphology and rates of landscape evolution, as well as the opportunity to assess responses to the perturbation of a major resistant lithology being introduced into the drainage network. The long profile morphology of the Lachlan River drainage net demonstrates that ongoing rock uplift driven by denudational isostatic rebound is propagated headwards through the drainage net from the inland edge of the bedrock highlands. That rock uplift 'signal' is transmitted more rapidly through those parts of the drainage net formed on less resistant lithologies, such as regionally metamorphosed meta-sandstones and phyllites, and more slowly on more resistant granites and contact metamorphic hornfels. The basalts themselves also retard landscape response to rock uplift, but the slowing of such responses occurs on the granites and contact metamorphic rocks, whether the basalts are present or not.

Four general conclusions may be drawn from the study reported here:

1. Denudational isostatic rebound is an important and fundamental mechanism for prolonging the timescale for the postorogenic decay of topography (Bishop & Brown 1992; Baldwin *et al*. 2003).
2. Resistant lithologies, and the delay that they exert on the transmission of the signal of rock uplift triggered by denudational isostatic rebound, are further important factors in prolonging the timescale of postorogenic decay of topography; this second group of factors has hitherto not been evaluated rigorously.
3. The role of resistant lithologies indicates that detachment-limited conditions are a key to the longevity of (at least some) post-orogenic landscapes. The general importance of transport-limited conditions, as proposed by Baldwin et al. (2003), remains to be evaluated in field settings.
4. The data demonstrate that the appropriate model for the evolution of landscapes such as that described here is one of spatially unequal activity and increasing relief, as Crickmay (1974 1975) and Twidale (1991) have emphasised.

The delays in the propagation of knickpoints may persist for considerable periods, reinforced by very low stream power that reflects low discharges in these often semi-arid, intra-continental interiors and their low gradients, as well as low fluxes of sediment to act as erosional 'tools'. In other words, non-steady-state landscapes may lie at the heart of widespread, slowly evolving post-orogenic settings, such as high-elevation passive continental margins, meaning that non-steady landscapes, with increasing relief through time, are the 'rule' rather than the exception on the Earth's surface.

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Figure captions

Fig. 1. (a) Long profile of a hypothetical stream (lower line) with $\lambda = 0.9$ and $k = 50$ for all reaches and where the standard error of elevation is 2 m and normally distributed; the reach from D = 800 to D = 720 is a diffuse knickpoint of 40 m fall generated by the headward propagation of a base-level fall of 40m. The plot shows the downstream projection of the equilibrium profile upstream of A in accordance with the DS model (upper line); the method of projection is presented by Goldrick & Bishop (2007). The DS-based estimate of the amount of incision in response to the 40 m base-level fall is 35 m (i.e., $H^* = 35$ m). (b) DS plot of this stream, showing the knickpoint at A.

Fig. 2. The Lachlan River's uplands (bedrock) drainage net. Box gives the area in Fig. 4.

Fig. 3. Geology of the Lachlan's uplands (bedrock) catchment. The Tertiary basaltic volcanics upstream of Cowra at the confluence of the Boorowa and Lachlan Rivers is the 12 Myr old Boorowa basalt. Box indicates the area in Fig. 4.

Fig. 4. The drainage network, geology and sub-basaltic contours of the upper Lachlan catchment. Sub-basaltic contours (20 m contour interval) have been taken from Fig. 8 of Bishop *et al*. (1985) and their elevations decrease westwards and northwards. These contour lines have not been labelled in order to avoid clutter and because it is their relative spacing and orientation rather than their absolute elevations that are important; likewise, we number (rather than name) the streams to avoid clutter and multiple stream names. The Bevendale Basalt is the southern- and western-most linear flow remnant (flowing westwards and then northwards adjacent to the Lachlan trunk stream [stream number 61]), and the Wheeo Basalt is the parallel linear flow, to the north and northeast of the Bevendale Basalt (Bishop 1984). The broad tabular basalt in the east, at the continental drainage divide, is the Divide Basalt. LGF = Lake George Fault.

Fig. 5. Detail of drainage network, geology and sub-basaltic contours in the vicinity of Streams 59 and 60. The asterisk marks the western limit of the possible location of the drainage divide for the basalt-filled east-west tributary; the eastern limit for the divide location is the catchment boundary in the east (the continental drainage divide). The text gives more detail.

Fig. 6. The DS plot (top) and long profile (bottom) of Stream 59 (left) and Stream 60 (right). Solid diamonds: data points used for profile reconstruction; broken lines: upper and lower confidence limits of projections of reconstructed profiles; squares: elevations of base of basalt (from Fig. 5); thin vertical lines above squares: estimated thickness of the basalt flow; thin vertical lines below

squares: uncertainty of elevation for base of the basalt. Shading and two-letter labels indicate underlying geology (Ad: Adaminaby Group (Palaeozoic quartz-rich metasediments); Wy: Wyangala batholith (Palaeozoic granitic rocks); thick vertical line: hornfels ridge).

Fig. 7. (a) DS plot and (b) long profile of the sub-basaltic elevations of the east-west (tributary) E Miocene lava flow, assuming that the Miocene divide was at the asterisk in Fig. 5.

Fig. 8. The DS plot (top) and long profile (bottom) of (a) Stream 67 and (b) Stream 68. Symbols as in Fig. 6.

Fig. 9. The DS plot (top) and long profile (bottom) of (a) Stream 69 and (b) Stream 72. Symbols as in Fig. 6.

Fig. 10. Schematic of hypothesised post-eruption evolution of tributaries of the upper Lachlan. After the eruption, the trunk stream is re-established on the left side of its basalt-filled valley. Trunk-stream incision generates knickpoints where right-bank tributaries cross the basalt; left-bank tributaries do not have to incise basalt. Long-profile projections show that the right-bank tributaries above the KP are graded to the base of the basalt (Fig. 6), whereas the western tributaries did not have to negotiate the basalt and so do not exhibit such grading (Figs 8 and 9, especially Fig. 8b). In the top three cartoons an earlier KP is also shown propagating up the right-bank tributary.

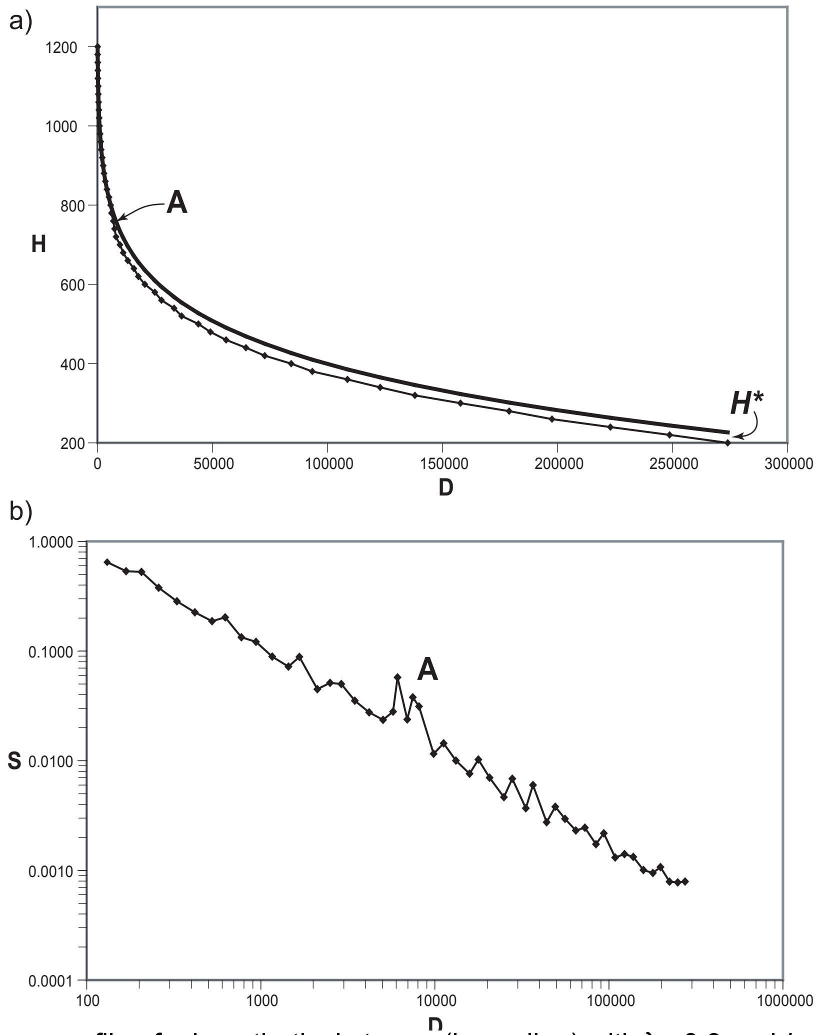
Fig. 11. Numerical simulation of the evolution of the DS plot and long profile of an initially graded stream profile after the emplacement of a segment of resistant lithology. The broken line on the long profile is the original profile and the closely-spaced vertical lines indicate the extent and thickness of the resistant lithology.

Table 1. *Values of H* for long profile projections to the trunk Lachlan River of the graded reaches upstream of the downstream-most knickpoint in the upper Lachlan's eastern (right-bank) tributaries (see Figure 4 for location of streams)*

Projected stream	**H* (m) at confluence with Lachlan**
(a)	
56	63 ± 4
57	16 ± 4
58	80 ± 5
59	60 ± 10
60	83 ± 13
(b)	
57 (upstream)	122
58 (upstream)	143 ± 10
Miocene trib.	103 ± 30

Table 2. *Values of H* for long profile projections of upstream graded reaches of the western (left-bank) tributaries to the upper Lachlan River; and the confluences to which they have been projected*

Projected Stream	H* (m) at confluence	Confluence
67	30 ± 2	67 & 68
68	26 ± 17	67 & 68
68	25 ± 14	68 & 69
69	16 ± 5	68 & 69
73	21 ± 0	73 & 61



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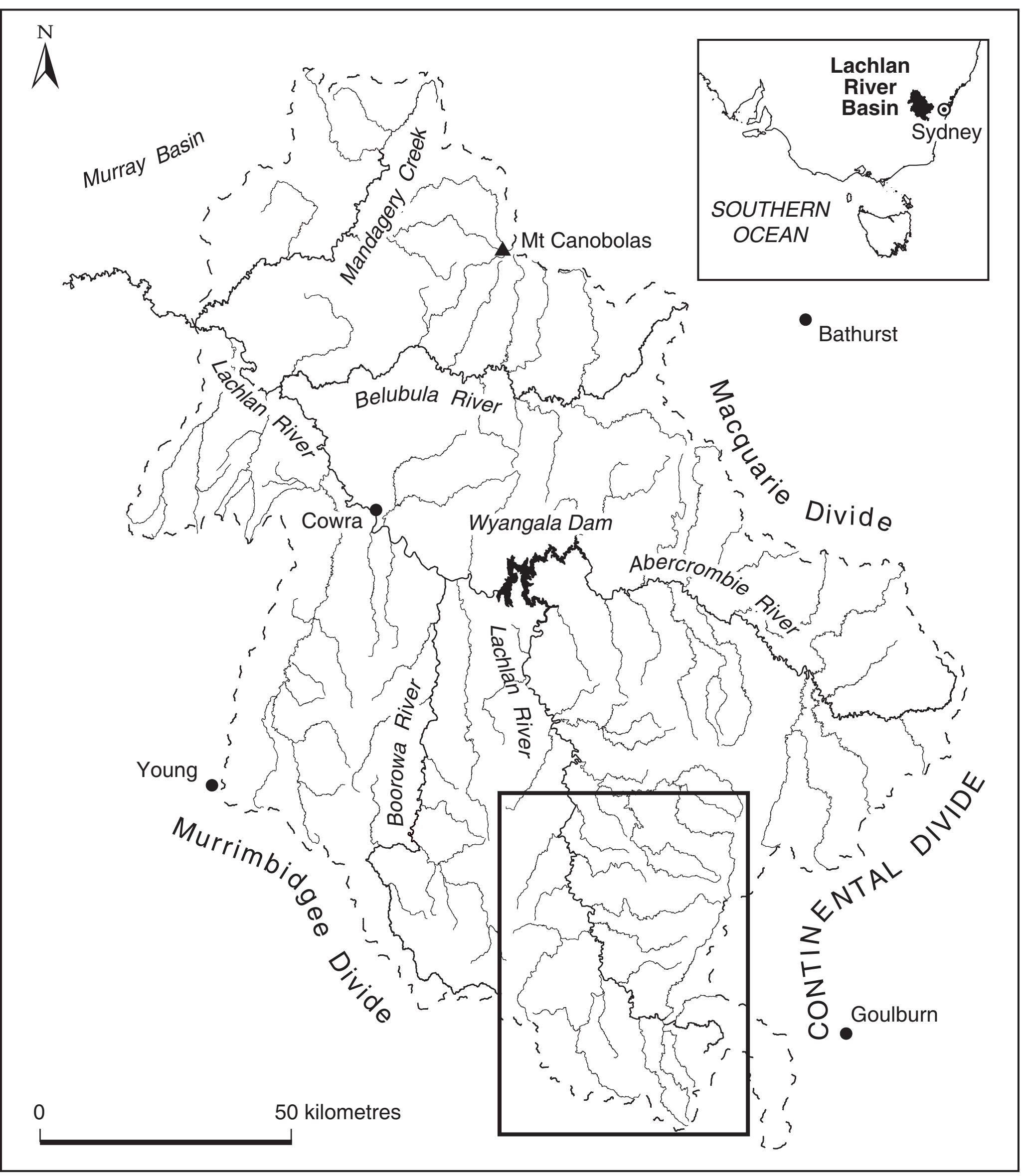

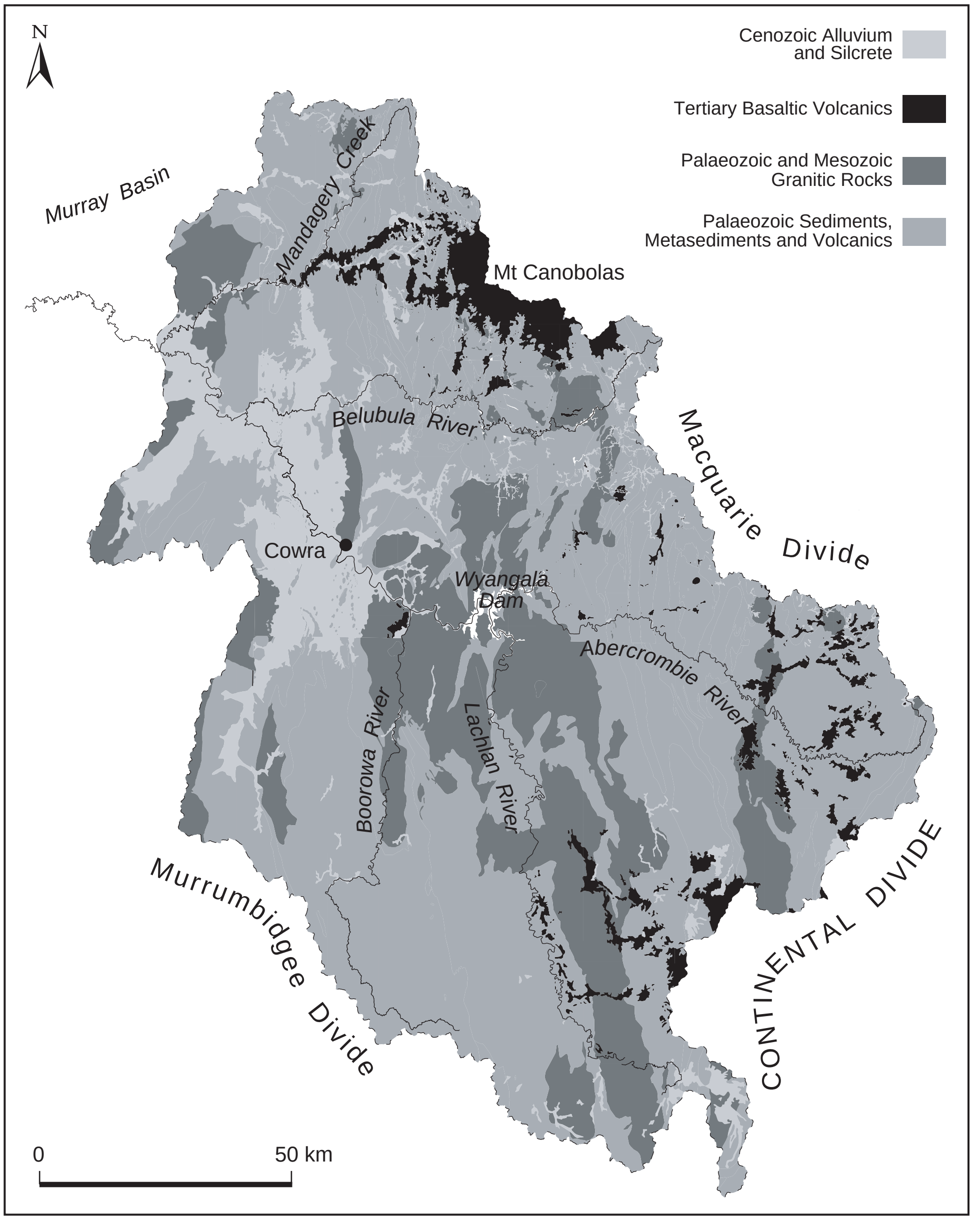

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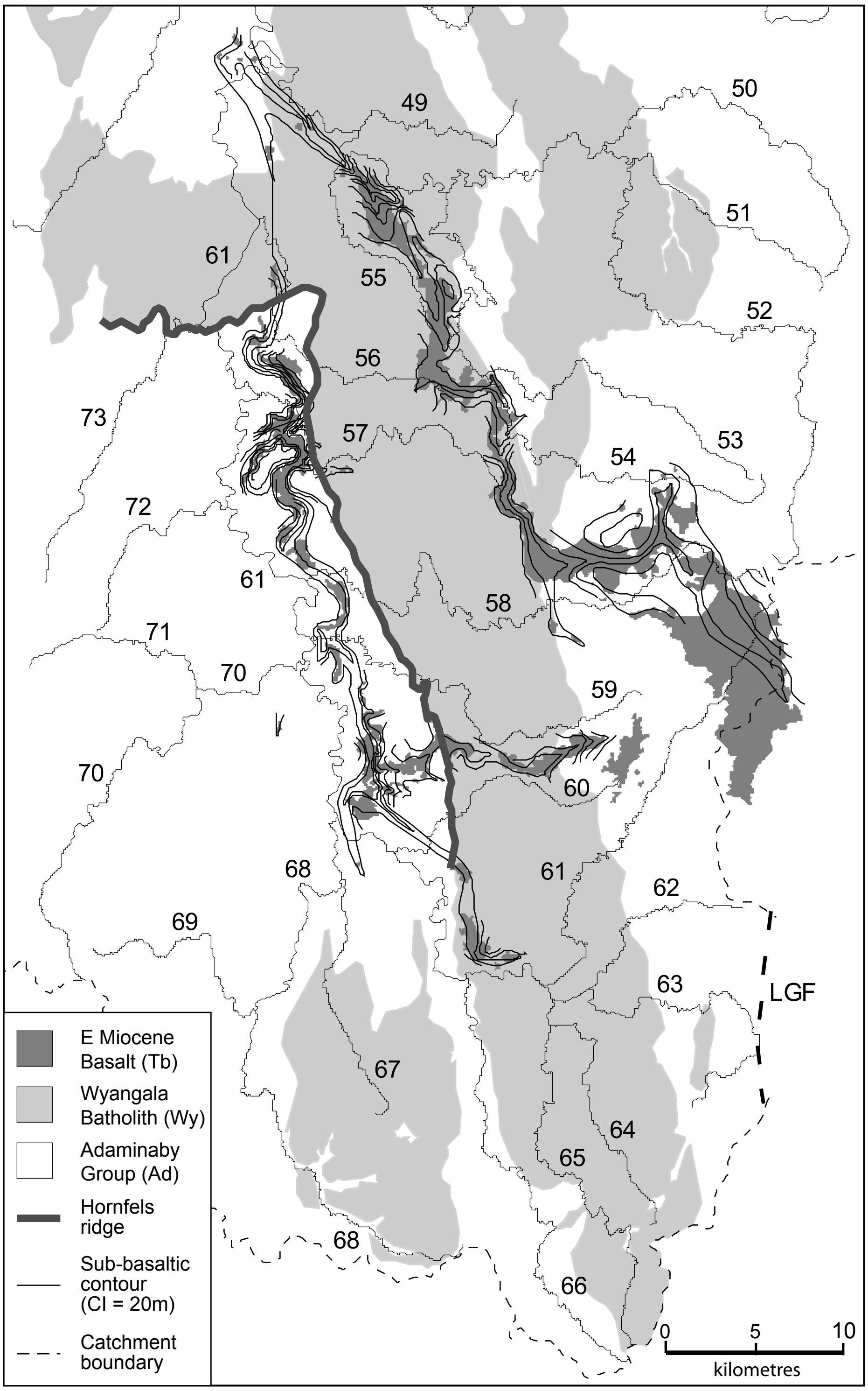

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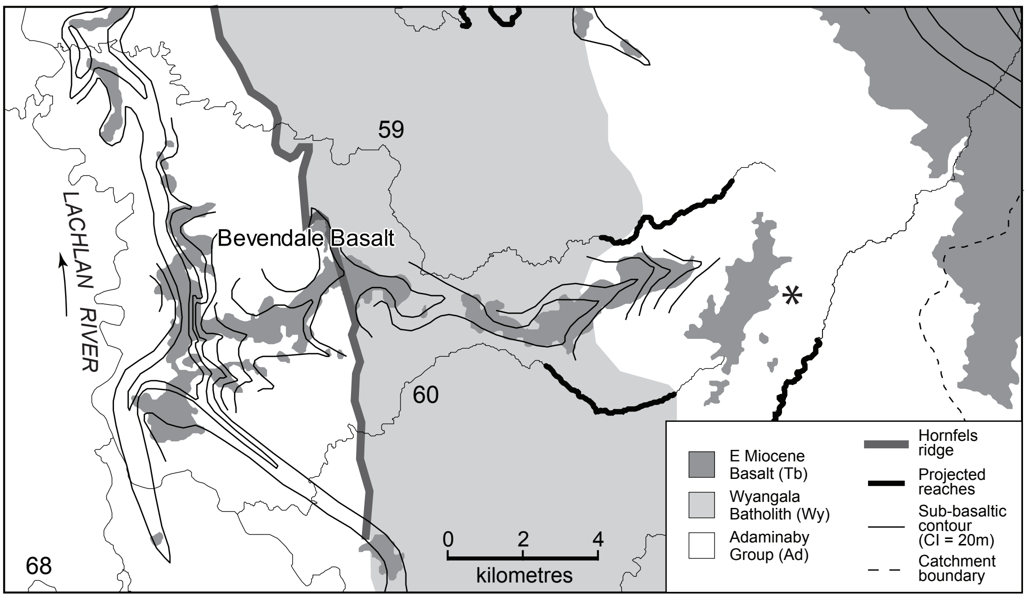

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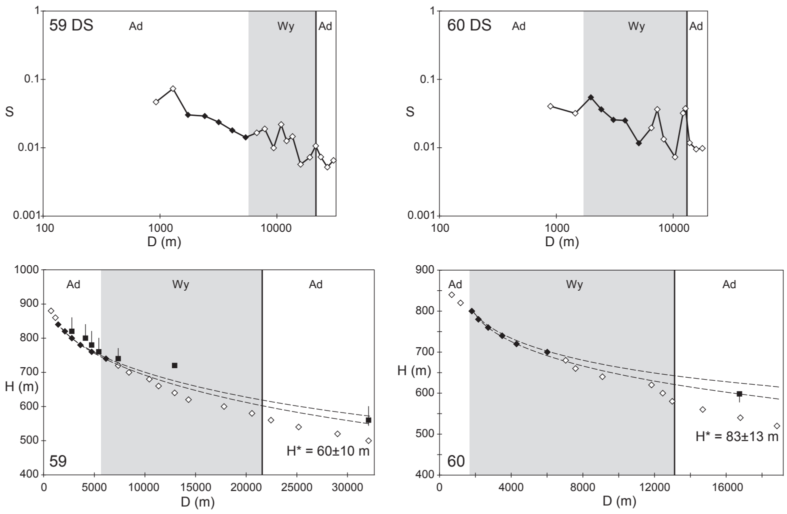

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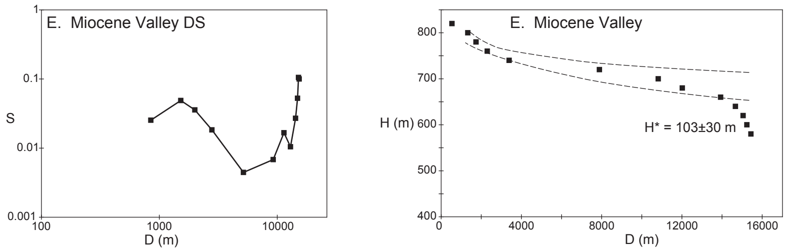

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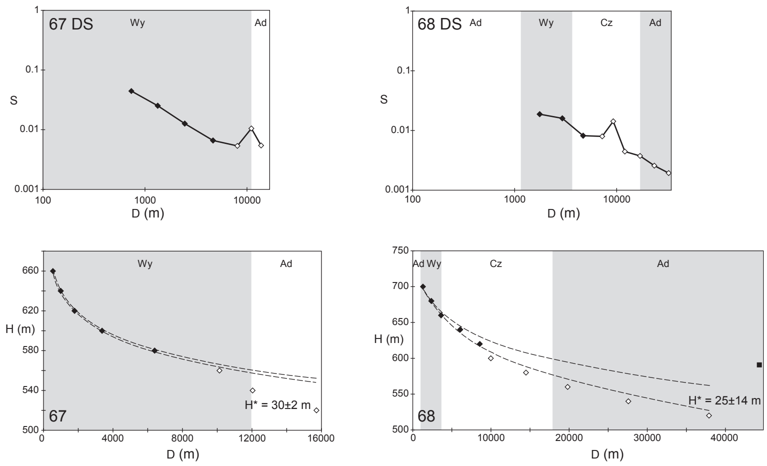

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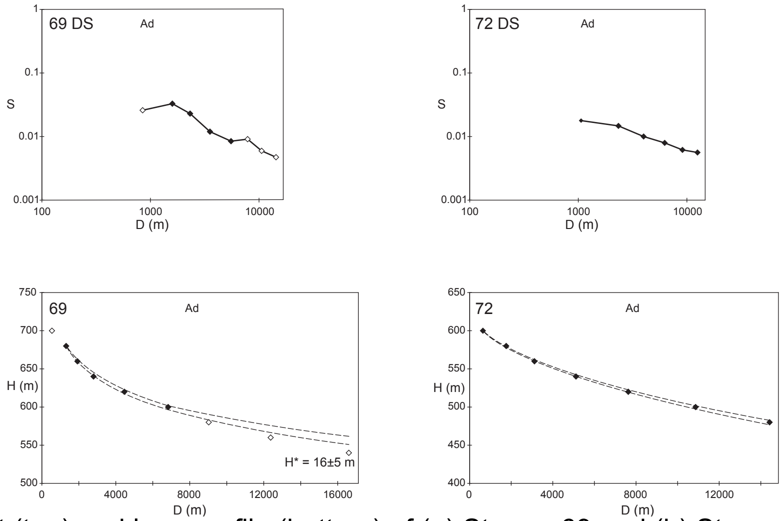

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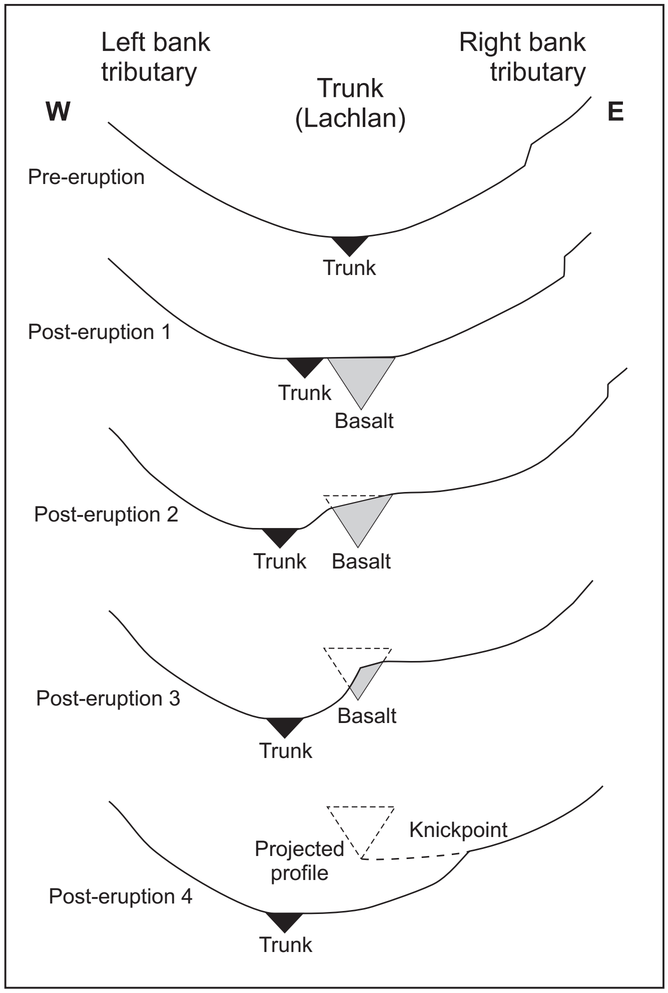

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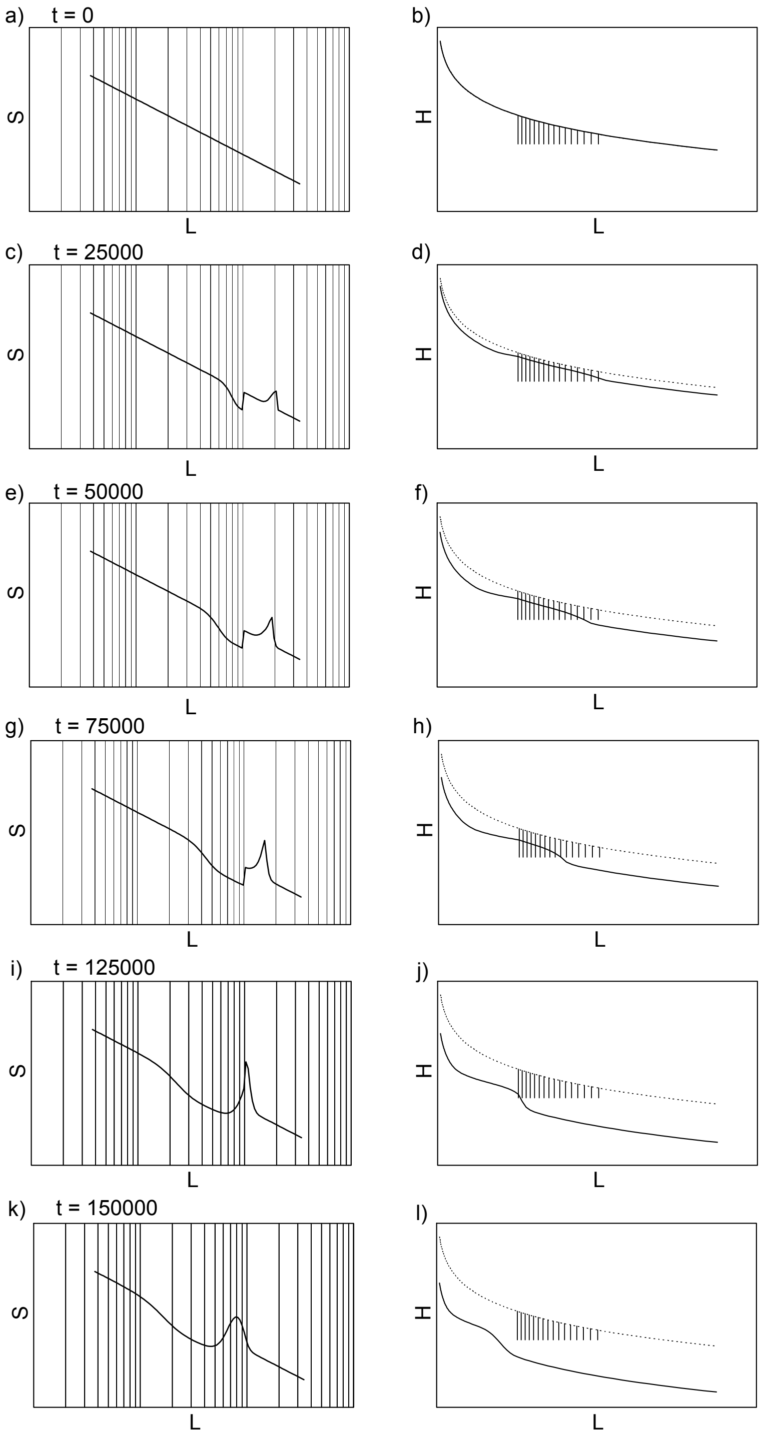

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