## Contrasting transient and steady-state rivers crossing active normal faults: new field observations from the Central Apennines, Italy

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## ABSTRACT

We present detailed data on channel morphology, valley width and grain size for three bedrock rivers crossing active normal faults which differ in their rate, history and spatial distribution of uplift. We evaluate the extent to which downstream changes in unit stream power correlate with footwall uplift, and use this information to identify which of the channels are likely to be undergoing a transient response to tectonics, and hence clarify the key geomorphic features associated with this signal. We demonstrate that rivers responding transiently to fault slip-rate increase are characterised by significant long-profile convexities (over-steepened reaches), a loss of hydraulic scaling, channel aspect ratios which are a strong non-linear function of slope, narrow valley widths, elevated coarsefraction grain-sizes and reduced downstream variability in channel planform geometry. We are also able to quantify the steady-state configurations of channels, that have adjusted to differing spatial uplift fields. The results challenge the application of steady-state paradigms to transient settings and show that assumptions of power-law width scaling are inappropriate for rivers, that have not reached topographic steady state, whatever exponent is used. We also evaluate the likely evolution of bedrock channels responding transiently to fault acceleration and show that the headwaters are vulnerable to beheading if the rate of over-steepened reach migration is low. We estimate that in this setting the response timescale to eliminate long-profile convexity for these channels is  $\sim 1$  Myr, and that typical hydraulic scaling is regained within 3 Myr.

## INTRODUCTION

### **Study motivation**

Bedrock streams in steep mountain catchments are one of the most important agents that control landscape evolution (Howard & Kerby, 1983; Howard *et al.*, 1994; Whipple & Tucker, 2002). In the shorter term, these channels set hillslope gradients and hence determine topographic relief (Tucker & Bras, 1998; Tucker & Whipple, 2002), and over longer timescales they control both the erosional unloading of mountain belts (Whipple & Tucker, 1999; Willett & Brandon, 2002), and the type, quantity, size, and distribution of eroded sediment exported either towards the ocean or to neighbouring basins (Milliman & Syvitski, 1992). Because the fluvial system is sensitive to tectonically imposed boundary conditions, channel adjustment to externally driven forcing can potentially offer insight into phenomena as diverse as landscape response times (Snyder *et al.*, 2000) and basin stratigraphy (Cowie *et al.*, 2006) and may allow rates of tectonic uplift to be estimated where direct structural or geodetic data are unavailable (Lavé & Avouac, 2001; Finlayson *et al.*, 2002; Kirby *et al.*, 2003; Wobus *et al.*, 2006).

Landscape evolution models offer the most viable way to improve our understanding of these issues, because they allow forward modelling of fluvial systems coupled to hillslope processes, over a range of timescales, and under a suite of varying boundary conditions (Tucker *et al.*, 2001a; Willgoose, 2005). However, to model river incision successfully, particularly in response to changes in boundary conditions, we require the correct treatment of channel geometry as well as the appropriate erosion law, as both of these govern erosive power in any river system. Existing landscape evolution models are only as a good as the algorithms they employ and there remains considerable debate over two fundamental issues: (a) which fluvial incision laws to use within the models, e.g. 'detachment-limited' vs.

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'transport-limited' or various 'hybrid models' (see Whipple, 2004 for a review), and (b) How best to parameterise the downstream evolution of river morphology in upland areas, because fluvial incision at any point is a function of local channel geometry, grain-size and valley form (Pazzaglia et al., 1998; Duvall et al., 2004; Finnegan et al., 2005). In this paper, we address both of these challenges using a unique field study that characterises the hydraulic geometry and sediment calibre of three rivers in the Central Apennines of Italy, crossing active normal faults that differ in terms of their spatial distribution of uplift and also in terms of their temporal history of slip. We evaluate how these channels have adjusted to their tectonic setting, and the implications this has for understanding fluvial form in rivers undergoing a transient response to tectonics, compared with channels that have reached topographic steady state (i.e. where channel incision rate equals the tectonic uplift rate).

### **Background and paper aims**

Whipple & Tucker (2002) argued that to discriminate between competing fluvial incision laws, we need to examine rivers undergoing a transient response to a change in boundary conditions, because at topographic steady state, many different erosion laws can produce similar looking landscapes. In particular, they demonstrated that catchments responding to an increase in uplift rate relative to original base-level develop diagnostic morphologies depending on the erosion law chosen: for example, detachment-limited and hybrid rivers are predicted to develop a 'knickpoint' or convex reach in response to an increase in uplift rate, whereas the long profiles of purely transport limited channels tend to respond diffusively to identical conditions (Tucker & Whipple, 2002). This work led to a number of studies attempting to model transient river response to tectonic forcing, in the hope of obtaining definitive evidence for favouring one or more erosion laws (e.g. Snyder et al., 2003; Tomkin et al., 2003; Van der Beek & Bishop, 2003), to assess landscape response time (e.g. Snyder et al., 2000; Baldwin et al., 2003) or to model diagnostic geomorphic signals of transience in the landscape (e.g. Snyder et al., 2003; Bishop et al., 2005). So far these attempts have met with only limited success: Van der Beek & Bishop (2003) found it difficult to definitively fit any one erosion model to the Lachlan catchment, SE Australia, although in part this is because their data may not actually resolve enough information about the transient response. Snyder et al. (2003) evaluate channel response to tectonic forcing in the Mendocino triple junction region, but again do not definitively identify transient conditions. Baldwin et al. (2003) consider implications of a range of stream-power models for post-orogenic decay in mountain belts, and show, in theory, that the effects of tectonic uplift can persist in fluvially mediated landscapes over very long periods. However, they do not actually seek to identify modern day transient landscapes. Bishop et al. (2005) do identify 'knickpoints' in rivers draining the eastern coast of Scotland which they interpret as a transient response to post-glacial rebound of the coastline in the last 18 ka, but they have poor control on the timing and mode of knickpoint generation, and their interpretation rests on assumptions of topographic steady state. Published estimates of landscape response time also vary by several orders of magnitude (Merrits & Vincent, 1989; Snyder *et al.*, 2000).

A key feature of the above studies is that they use traditional hydraulic scaling relations (Leopold & Maddock, 1953) to evaluate the evolution of channel width, W, and depths, H, on a point by point basis downstream. The key assumption is that channel geometry can be described by a power law dependence on upstream drainage area, A, or river discharge, Q, giving equations such as

$$W = K_1 A^b \tag{1}$$

$$H = K_2 A^c \tag{2}$$

where  $b \sim 0.5$  and  $c \sim 0.35$  (Knighton, 1998). Although Eqns (1) and (2) were derived from data sets characterising lowland alluvial rivers, Montgomery & Gran (2001) argued that for mountain rivers in tectonically quiescent areas of uniform terrain, similar relationships may apply, resulting in the widespread adoption of these equations in landscape evolution models (albeit with varying values for exponents b and c). However, by using such relationships to study river response to tectonic forcing, the implicit assumption is that hydraulic geometry is insensitive to transient conditions. Conversely, valley and channel adjustment are accepted to be key ways in which rivers respond to spatial changes in boundary conditions because channel shape fundamentally controls the distribution of energy expenditure and frictional stresses, which are closely correlated with erosive force (Turowski et al., 2006). For example, several studies document narrowing and/or steepening in response to both harder lithologies (e.g. Pazzaglia et al., 1998), and higher uplift rate (e.g. Duvall et al., 2004; Whittaker et al., 2007) while Lavé & Avouac (2001) show that flood-plain widths also narrow in areas of high uplift rate. Additionally, Harbor (1998), documents changes in channel planform and grain size as the Sevier river crosses a zone of transverse uplift in southern Utah. In these examples, empirical relationships, such as Eqns (1) and (2), are violated locally. Some authors (e.g. Kirby et al., 2003) argue that, for systems in topographic steady state, simple empirical relationships are valid although the value of the exponent b Eqn. (1) may vary (see also Duvall et al., 2004; Wobus et al., 2006). However, channel adjustment has been shown numerically to occur as a dynamic response to temporal variations in climatic and tectonic conditions acting on bedrock rivers (Stark, 2006; Wobus et al., 2006). Thus, hydraulic scaling relationships may be inappropriate for characterising the transient response of fluvial systems and by implementing them in landscape evolution models, we may miss a crucial aspect of the system's adjustment to external perturbation.

The above studies raise two key issues: firstly, what criteria can we use to detect, unambiguously, transient responses in a fluvial system? And secondly, to what extent are widely used hydraulic scaling empiricisms, above, applicable for channels undergoing a transient response to tectonics? We explicitly tackle these outstanding questions using a unique dataset of three rivers crossing currently active normal faults in the Central Apennines of Italy ('Background: structural and tectonic setting'), where earlier studies (e.g. Whittaker et al., 2007) have already demonstrated that at least one river in the area is likely to be undergoing a transient response to tectonics. Here, we build on previous work by comparing and contrasting the morphology of rivers crossing both back-tilting normal faults and uniformly uplifting horsts that differ in terms of their temporal history of slip accumulation ('Data collection and methods'). In 'Study rivers', we present detailed field observations of channel geometry and sediment calibre in the three channels, to identify how the study rivers are responding to their differing tectonic settings. We then consider how channel aspect ratios evolve downstream in areas of active tectonics and evaluate the extent to which typical hydraulic scaling assumptions are valid for rivers perturbed by normal faults ('Data analysis'). We also assess which class of erosion laws is most appropriate for describing the long-term incision characteristics of the three rivers in question, and by evaluating downstream changes in Shields stress, we argue that all three channels must be close to the detachment-limited end member. With these observations in mind, we then consider explanations for the three channels' differing behaviour ('Discussion - explanations for differing channel behaviour'). By comparing the distribution of unit stream power in each of the channels with our reconstructions of the tectonic uplift field and base-level history experienced by each river, we evaluate which of the channels are likely to be in topographic steady state, and which are likely to be undergoing a transient response to tectonics. Finally, we assess how transient landscapes progress towards steady state, and estimate the response timescale of bedrock rivers in the area by contrasting channels that have been perturbed by tectonics at different times in the past. The results enable us to characterise, for the first time, the diagnostic field criteria of a transient river response to tectonics, and provide unique insights into the way in which the river system transmits tectonic signals to the landscape.

# BACKGROUND: STRUCTURAL AND TECTONIC SETTING

The central Italian Apennines initially developed as a north-east verging imbricate fold and thrust belt during the Miocene along the margins of the Adriatic microplate, in response to south-east retrograde motion of the Adriatic trench (Cavinato & De Celles, 1999). Compression largely ceased by the early Pliocene (Centamore & Nisio, 2003), and since  $\sim 3$  Ma extensional deformation has migrated eastward behind the thrust front (Lavecchia *et al.*,

1994; D'Agostino et al., 2001), producing a 150-km-long network of high-angle normal faults (Fig. la) that accommodates stretching of  $\sim 6 \,\mathrm{mm \, year^{-1}}$  across central Italy (Tozer et al., 2002; Hunstad et al., 2003; Roberts & Michetti, 2004). The faults uplift limestones of Jurassic to Palaeocene age, while the downthrown hangingwalls expose Miocene turbidite flysch. (Fig. 1b) (Accordi et al., 1986). The Apennines emerged above sea level by the Pliocene (Centamore & Nisio, 2003) and the remnants of the low relief land surfaces created then occur locally on the footwall blocks of normal faults (Galadini et al., 2003). These faults lie on the back of a long-wavelength topographic bulge interpreted to have formed either in response to corner flow above the Adriatic slab (Cavinato & De Celles, 1999) or mantle upwelling (D'Agostino et al., 2001). The combined uplift and extension has resulted in the formation of numerous half-graben basins which are now filled with continental deposits dating from the Late Pliocene onwards, considered penecontemporaneous with the onset of extension across the Apennines (Cavinato, 1993; Cavinato et al., 2002).

The area has continuing seismicity, and most of the normal faults are still active (Fig. lc) (Lavecchia et al., 1994; Roberts & Michetti, 2004), with fault scarps offsetting hillslopes that correspond to late glacial surfaces in the region (Giraudi & Frezzotti, 1997; Roberts et al., 2004). This extensional fault array is one the best constrained in terms of variation in displacement and slip rate, both between faults and along individual fault segments (Roberts & Michetti, 2004). Total displacements for the faults have been calculated from offset of geological horizons, and current throw rates have been calculated from scarp profiling of the offset of the late glacial surface. The size of this offset decreases away from the fault centres, indicating a spatial decline in displacement rate towards the fault tips (Morewood & Roberts, 2002; Roberts & Michetti, 2004; Roberts et al., 2004). Throw rate data derived from structural mapping agree well with data gained from current geodetic observations (Hunstad et al., 2003), trench sites across active fault strands (e.g. Michetti et al., 1996; Pantosti et al., 1996 and references therein), seismic surveys (Cavinato et al., 2002) and recent fault surface exposure dating using cosmogenic nuclides (Palumbo et al., 2004).

There is strong evidence that some of these faults have undergone temporal variation in slip rates. Cowie & Roberts (2001) show that those near the centre of the array, such as the Fiamignano fault (F, Figs lc and 2) have current throw rates which are large for their (relatively small) total displacements, and imply a basin initiation age which is too young compared with the known age of basin fill sediments; consequently throw rates on central fault segments must have increased. In contrast, faults nearer the edge of the array, such as the Leonessa and South Cassino segments (L, SC, Fig. lc), have throw rates that are consistent with their total displacement and consequently have not undergone any throw rate acceleration (Fig. 2). The acceleration has been explained as a result of fault interaction



**Fig. 1.** (a) Location map for study area (b) geological map of the Central Apennines (c) map showing active, inactive and studied normal faults. Boxes (A), (B) and (C) correspond to the three tectonic settings outlined in 'Background: structural and tectonic setting', and also to the detailed locality maps shown in Fig. 3. L, Leonessa fault; *R*, Rieti fault; *F*, Fiamignano fault; *S*, Sella di Corno fault; SC, South Cassino fault; FC, Fucino fault.



Fig. 2. Temporal accumulation of throw for three faults considered in this study. The Fiamignano fault, near the centre of the array is under-displaced for its present-day slip rate, whereas the Leonessa and Cassino faults have total throws which are consistent with constant slip rate of  $\sim 0.30-0.35$  mm year<sup>-1</sup> for 3 Ma.

(Cowie & Roberts, 2001). A synthesis of modelling and empirical data strongly suggest the throw rate acceleration occurred at ~0.75 Ma (Roberts & Michetti, 2004). This interpretation is supported by seismic evidence and borehole data from the centrally located Fucino basin (FC, Fig Ic) which show much thicker sediment sequences dipping towards the active fault from the mid-Pleistocene onwards, compared with that during late Pliocene–early Pleistocene times (Cavinato *et al.*, 2002).

We use this uniquely well-constrained data set to characterise how perennial rivers respond to variations in both spatial and temporal uplift rates on faults in three differing tectonic settings (shown on Fig. 1, and illustrated in detail in Fig. 3):

(A) Horst (uniform) uplift, with constant throw rate: Fig. lc - Rieti (R) and Leonessa (L) faults. We focus on the Fosso Tascino channel (Fig. 3a), crossing the Leonessa fault.

(B) Back-tilted fault block with constant throw rate: Fig. lc, South Cassino (SC) fault. We focus on the Valleluce river (Fig. 3b)

(C) Back-tilted fault block with increased throw rate: Fig. lc, Fiamignano (F) and Sella di Corno faults (S). We focus on the Rio Torto (Fig. 3c), which crosses the Fiamignano fault.



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Appendix).

represents the location of structurally perched late Pliocene sediments, Y is the location of mid-Pleistocene conglomerates and Z shows the location of the internally drained Rascino Plain (see

## DATA COLLECTION AND METHODS

To document hydraulic adjustment to the imposed tectonic boundary conditions we measured the following field variables:

- (i) high-flow (bankfull) channel width,  $W_{\rm b}$ ,
- (ii) maximum channel depth, H,
- (iii) local channel slope, S,
- (iv) valley width,  $W_{\rm v}$

Additionally, we measured rock mass strength, sediment calibre and field evidence for hanging-wall valley incision in the three study areas to evaluate whether changing lithology, varying channel roughness or external (i.e. nontectonic) control on base levels might also explain the signals seen. A 20 m DEM, validated by field survey, was used to extract river long profiles. Data characterising the tectonic boundary conditions for the three cases are shown in Table 1 and combine results from Roberts & Michetti (2004) with new measurements to better constrain present-day throw rates.

Hydraulic geometry was measured using a hand-held laser range-finder (precision - 1 cm) so errors associated with  $W_{\rm b}$  and H are largely associated with selecting the stage to measure: defining such parameters must be with respect to a reference; typically this is the bankful stage, where the river channel tops out into the over-bank (Leopold & Maddock, 1953; Knighton, 1998). Although the definition of such a stage remains a subject of debate (e.g. Copeland et al., 2000), widths and depths associated with formative conditions are readily estimated from the limits of active abrasion, vegetation boundaries, highest levels of bleaching on boulders and water-washed surfaces and the remains of high stage flood debris. This approach is typically used to define bedrock channel geometry (e.g. Montgomery & Gran, 2001; Snyder et al., 2003) and based on this precedent, we assume such measurements reflect active conditions in the channel. Moreover, the frequency of measurement (every 300 m downstream and substantially smaller intervals in many instances) means we are confident of having gauged a constant reference frame downstream. In gorges with no recognisable over-bank, we have measured the high-flow stage, as deduced from the same field indicators listed above. Channel slope measurements are reach-representative and typically cover a distance of 20-30 m as appropriate. Variation associated with hitting the target positioned downstream gives an empirically determined error of  $\pm 0.2^{\circ}$ . Valley widths were measured at a reference height of 2 m above the river; this was above bankful depth in almost all cases. Where H was > 2 m,  $W_v$  was measured at 0.5 m above this level. Rock resistance to erosion was evaluated using the Selby mass strength index (Selby, 1980). This represents a semi-quantitative assessment of rock hardness; geometry, orientation and size of joints/bedding; and the degree of weathering/groundwater saturation. Index values range from 0 to 100 with soils corresponding to values <25. In particular, the Selby index accommodates relative differences in intact rock strength and hardness (cf. Sklar & Dietrich, 2001), and structural constraints on bedrock resistance to erosion. This is important because intact rock strength alone is a poor indicator of erodibility in heavily jointed lithologies (Whipple *et al.*, 2000a). Coarse-fraction grain-size on the channel bed was estimated by Wolman point counting of the major and minor axes of 100–300 individual, randomly selected particles >1mm in size, mantling the channel (Wolman, 1954). The median value,  $D_{50}$ , and the  $D_{84}$  of the individual particles was taken to yield a representative measure of sediment calibre at each locality. Ancillary measurements indicate  $D_{50}$  estimates typically fluctuate by  $< \pm 0.5$  mm with increasing number of measurements in excess of 100 grains.

## **STUDY RIVERS**

In this section, we combine the uplift and base-level histories for the three rivers with detailed observations of channel form, geometry and grain-size as a function of downstream distance. Because there are good reasons for believing that rivers responding to active tectonics may not demonstrate typical hydraulic scaling (see 'Background and paper aims') we present the data on linear, rather than log scales, and return to the applicability of power-law scaling relationships for tectonically perturbed rivers in 'Do channel widths scale with drainage area for rivers crossing active faults?'.

### Case A – Horst uplift (Fosso Tascino, Leonessa fault)

The Fosso Tascino is the trunk stream of a 45 km<sup>2</sup> catchment, draining the uplifted horst block between the Leonessa and Rieti faults, which dip in opposite directions (Fig. 3a). Both have similar maximum throw rates of 0.35 mm year  $^{-1}$ . The river cuts across the Leonessa fault 500 m SE of Leonessa village. Total throw on the fault here is  $\sim 1000$  m, and where it intersects the river, the current throw rate is approximately  $0.3 \,\mathrm{mm \, year^{-1}}$ . This rate is consistent with both the total throw and the 3 Ma initiation age of faulting in this area (Fig. 2), indicating constant throw rate through time. The river displays a concave up profile (concavity,  $\theta = 0.42$ , where  $S \sim A^{-\theta}$ , Fig. 4). It exhibits a mixed cobble-gravel bed with occasional exposure of channel-floor bedrock in the upper part of the catchment, and wide open reaches which are largely alluviated in the lower part of the catchment near the fault. Geological surveying indicates the drainage is mono-lithologic Mesozoic limestone and in situ assessment of Selby rock mass strength yielded no substantial differences downstream, with average values of  $\sim$ 61; consequently there is little difference in rock resistance to erosion within the footwall. The hanging wall basin is filled with Plio-Pleistocene sediments, which are  $\sim$ 380 m thick (Michetti & Serva, 1990).

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Fault	Site	UTM (X)	UTM (Y)	Slip vector (deg.)	Plunge (deg.)	Throw pre-rift strata (m)	Scarp height (m)	Throw rate (mm/year)	Source
Leonessa	1	332 371	471 4421	53	58	1000	6.5	0.36	Roberts & Michetti (2004)
	2					100	<2?	< 0.1	Roberts & Michetti (2004)
Rieti	1	$332\ 000$	$4\ 695\ 000$	310	59	< 500	∧ .c	< 0.27	Roberts & Michetti (2004)
	2	328705	4 701 991	266	82	1000	L~	0.38	Roberts & Michetti (2004)
	3	323500	4711000	205	46	< 500	∧ v	< 0.27	Roberts & Michetti (2004)
South Cassino	1	400599	4 605 376	152	59	100	<2	< 0.1	Roberts & Michetti (2004)
	2	402901	4 602 754	170	45		5.1	0.28	This study
	3	406 414	4 596 916	181	45	1200	ιΩ	0.28	This study
	4	409548	4593686	225	53	1100	9	0.33	Roberts & Michetti (2004)
	ŝ	412 218	4 590 832	248	41	1100	4	0.22	Roberts & Michetti (2004)
	9	416 659	4 589 252	277	52	700	<2?	0.1	Roberts & Michetti (2004)
Fiamignano	1	337173	4690531	139	35	400	4	0.2	Roberts & Michetti (2004)
1	2a	$345\ 000$	4682000	232	51		16.5	0.92	Roberts & Michetti (2004)
	2b	345103	4 681 198	220	78	1700	$16.5 \pm 4$	0.9 + 0.1 = 1.1	This study
	3	345 556	4 681 219	220	70		18	1	This study
	4	350 175	4681481	180	52		17	0.94	This study
	IJ.	350 285	4 679 462	230	60		13.5	0.75	This study
	9	355 500	$4\ 674\ 500$	262	67	200	</td <td>&lt; 0.15</td> <td>Roberts &amp; Michetti (2004)</td>	< 0.15	Roberts & Michetti (2004)
South di Corno	1	354 574	4 706 216	130	65	1000	<2	< 0.1	Roberts & Michetti (2004)
	2	340759	4695166	223	57	400	9	0.33	Roberts & Michetti (2004)
Fault localities for thi	s study were o	btained from hand-l	held GPS and are acc	curate to <10 m. Cur	rent throw rates v	vere obtained by measurir	ng vertical off-sets of	the late glacial (18 ka) pala	eo-slope, consistent with the
methodology of Kobe	rts & Micheth	1 (2004) from which t	he remainder of the	tectonic data are sour	ced.				



**Fig. 4.** Long profiles of channel elevation against stream-wise downstream distance for rivers crossing Leonessa, Fiamignano and South Cassino faults, shown in Fig. 3.

These are presently incised by  $\sim$ 50 m from the upper surface, which is mid-Pleistocene in age (Michetti & Serva, 1990; Cavinato, 1993), indicating 50 m of base-level fall since 0.75 Ma. The rate of this base-level fall is not known precisely, but the succession of terraces inset within the valley of the Fosso Tascino (Michetti & Serva, 1990) and the lack of convex reaches on the Fosso Tascino, or on any other channels in the Leonessa basin, argue for alternating periods of incision with aggradational interludes.

Raw data for the channel geometry are shown in Fig. 5. Bankful channel widths increase downstream from <2 m in the headwaters to  $> 20 \,\mathrm{m}$  where the river crosses the fault. Channel slope is high in the headwaters and declines downstream as would be expected in graded, equilibrium channels (S typically < 0.05 beyond 4 km downstream). There is therefore no steepening in local stream-wise gradient as the river nears the fault. The ratio of channel width to valley width,  $W_{\rm b}/W_{\rm w}$  gives us a measure of the extent to which erosion is concentrated within the valley (Pazzaglia *et al.*, 1998). Here,  $W_b/W_v$  is highly variable, with no readily discernable trend with increasing distance downstream. Generally, the river flows through a valley which is approximately  $3 \times$  the width of the river itself and, significantly, there is no appreciable valley narrowing or gorge formation as the river nears the fault. There is also negligible correlation between slope S and  $W_{\rm b}/W_{\rm y}$  (correlation coefficient = -0.03). Importantly, for each of these measures, there is little evidence of the river systematically adjusting its form as a function of distance from the Leonessa fault, despite this being a zone of active uplift. In fact, the variability in channel form over small downstream distances is the most noticeable feature.

# Case B – tilted fault block with constant throw rate (Valleluce river, South Cassino fault)

The Valleluce river is a catchment of  $\sim 20 \text{ km}^2$ , crossing the South Cassino fault. This normal fault has a maximum throw of 1200 m, but where the river crosses the fault, the



Fig. 5. (a) High flow channel width,  $W_b$  (b) local channel slope, S (c) and ratio of  $W_b$ , to valley width,  $W_v$ , against downstream distance for the Fosso Tascino, Leonessa. Dashed lines with arrow heads indicate the smoothed trend of the data.

throw is  $\sim$ 950 m and the current throw rate is estimated to be 0.25-0.3 mm year<sup>-1</sup> similar to the Rieti and Leonessa faults above (Fig. 3b). This rate has been approximately constant since fault initiation (Fig. 2). Because the normal fault back-tilts to the NE, the uplift rate decays perpendicularly away from the fault into the distal footwall (Roberts & Michetti, 2004). The river thus flows towards the locus of maximum uplift as it approaches the fault rather than crossing a uniformly uplifting block as in case A. The hangingwall contains Miocene flysch with thick Pliocene-Recent cover, and the footwall contains uplifted Mesozoic limestone, with average Selby mass strength values of 60-65. However, one well-consolidated unit has Selby values of  $\sim$ 70 and there are also zones of carbonate cataclasite, where Selby strength falls to  $\sim 40$ ; these zones are highlighted in the channel geometry data in Fig. 6. The



**Fig. 6.** (a) High flow channel width,  $W_{\rm b}$ , (b) local channel slope, S, (c) and ratio of  $W_{\rm b}$ , to valley width,  $W_{\rm v}$ , against downstream distance for the Valleluce river, South Cassino. White background represents Selby rock mass strength values of 65; grey bars represent zones of weak cataclasite (Selby value ~40), stipple represents well-consolidated sandy limestone (Selby value ~70). Dashed lines with arrow heads indicate the smoothed trend of the data, excluding lithological outliers.

long profile (Fig. 4) shows that the river has a concave-up long profile (concavity = 0.51), with no prominent convex reaches. There is no evidence for any incision in the proximal hanging wall, indicating that externally driven baselevel falls have not exerted a significant control on the development of the catchment.

Channel width increases slowly towards the fault from < 2.5 m in the upper parts of the catchment to  $\sim 10$  m near the fault (Fig. 6a). Widths are higher beyond the fault, but this is attributable to the fact there is a large confluence near the village of South Elia with other rivers draining

the hangingwall of the fault. Measured local slopes decrease downstream, with the exception of a high gradient reach at 6 km downstream which appears to correspond to an area of increased rock mass strength (stipple in Fig. 6b). The ratio  $W_{\rm b}/W_{\rm w}$  although with some variability, increases systematically towards the fault (Fig. 6c), meaning that incision is being focussed in a narrower zone as the river approaches the zone of maximum uplift. Moreover, there is a moderate positive correlation between areas of high slope and lowered valley width (correlation coefficient = 0.44) showing that steeper reaches are associated with areas where the river is more tightly confined between the valley walls. This is particularly noticeable in the vicinity of the fault, despite the general trend of decreasing slopes with increasing downstream distance (Fig. 6b). Overall, channel planform shows many similarities to case A. The key differences are the relative constriction of the valley as the river approaches the fault (Fig. 6c), and the correlation of high local channel slopes with low values for  $W_{\rm b}/W_{\rm v}$ (Fig. 6b).

# Case C – tilted fault block, increased throw rate (Rio Torto, Fiamignano fault)

The RioTorto is the major river draining the footwall of the Fiamignano fault, with a catchment area  $> 65 \text{ km}^2$ . The normal fault is 25 km long, trends to the SE and dips SW; it has a displacement of  $\sim 1800 \,\mathrm{m}$  at its centre near the Fiamignano village, and is here estimated to have a throw rate  $\sim 1.1 \,\mathrm{mm \, year^{-1}}$  (Fig. 3c). The fault uplifts Mesozoic limestone of relatively uniform competence and juxtaposes it against Miocene flysch in the hanging wall (Fig 7a). The Rio Torto's headwaters lie near the tip of the Sella di Corno fault, and it then flows towards the Fiamignano fault, crossing SE of Fiamignano village, where the throw rate is  $\sim 0.9$  mm year<sup>-1</sup>. The upper parts of the Rio Torto are downthrown in the hanging wall of Sella di Corno fault, a 24-km-long segment with a total displacement of  $\sim 1000 \,\mathrm{m}$  and a maximum throw rate  $\sim 0.3 \,\mathrm{mm \, year}^{-1}$ (Roberts & Michetti, 2004). In addition, the Fiamignano fault underwent a throw rate acceleration from  $\sim 0.3 \,\mathrm{mm \, year^{-1}}$  at 0.75 Ma, to  $\sim 1 \,\mathrm{mm \, year^{-1}}$  (Fig. 2; Roberts & Michetti, 2004).

At present there is no significant accumulation of Pleistocene sediments on the hanging wall side of the Fiamignano fault, with which to constrain the baselevel history in the vicinity of the Rio Torto. However, there are conglomerates and lacustrine deposits of late-Pliocene age (1.8 Ma) preserved as a fault-bounded sliver,  $\leq 100$  m thick, within the proximal footwall (location X, Fig. 3c). These deposits are structurally perched at ~1000 m elevation whereas the elevation of the Rio Torto where it emerges onto the hanging wall is ~720 m. From these observations we infer that the amount of incision since 1.8 Ma must be in the range 100–280 m, depending on when these deposits were entrained within the fault zone. Mid-Pleistocene deposits near the village of South Pietro (location Y, Fig. 3c) crop out at elevations of up to ~770 m.



**Fig. 7.** (a) Selby rock mass strength and long profile in the Rio Torto. Inset shows plan view of catchment showing position of the major tributary, the Vallone Stretta. Numbers correspond to location of channel photos shown in (b). Photos show channel morphology downstream in the Rio Torto: (1) the headwaters, where local channel slopes are low, the valley wide and the channel is alluviated, (2) downstream of the slope break, where the channel forms a deep gorge, which cuts across the fault, (3) the hangingwall of the fault, where the river shallows and alluviates.

These deposits are the lateral equivalent of the classic Villafranchian sequence in the Citta Ducale gorge (also shown in Fig. 3c) (Accordi *et al.*, 1986; Cavinato, 1993). If these sediments extended as far as the Rio Torto at Fiamignano, it implies maximum aggradation of  $\sim$ 50 m between the Late Pliocene and the Mid-Pleistocene, and subsequent removal of this material from 0.75 Ma.

Unlike the other rivers (cases A and B), the Rio Torto channel has a prominent convexity in the long profile, which starts directly upstream of the fault and covers a vertical distance of >400 m in <5 km (Figs 4 and 7a). This

convexity cannot be attributed to lithology alone because there is no change in rock type or Selby mass strength until the river crosses into the hanging wall basin (Fig. 7a). There are also striking downstream changes in channel type within the Rio Torto as it flows towards the throw rate maximum where the river crosses the Fiamignano fault. In the headwaters (i.e. above the convex reach), the channel is shallow, partially alluviated and flows through a wide, open valley (photo 1, Fig. 7b). Downstream of the break in slope, in the convex reach, the channel forms a narrow gorge, with steep side slopes, and exposures of limestone bedrock



**Fig. 8.** (a) High flow channel width,  $W_b$ , (b) local channel slope, S, (c) and ratio of  $W_b$ , to valley width,  $W_w$  against downstream distance for the Rio Torto, Fiamignano. The black line in (b) represents a reference slope of 0.05. Dashed lines with arrow heads indicate the smoothed trend of the data.

in the base (photo 2, Fig. 7b). Once into the hanging wall, the river widens and alluviates, producing a channel morphology similar to that in the headwaters (photo 3, Fig. 7b).

Concomitantly with the morphological changes described above, there are significant variations in downstream channel geometry as the Rio Torto approaches the fault (Fig. 8). High flow channel width rises to  $\sim 10$  m within the first 3 km of the headwaters, but then remains approximately constant downstream towards the fault, despite the joining of a major tributary at  $\sim 8.5$  km downstream (Figs 7a and 8a). Channel widths widen markedly again as the river crosses from the uplifted footwall block to the hanging wall basin. Local channel slopes are gener-



**Fig. 9.** Grain-size for the rivers as a function of distance to the active fault. Closed circles;  $D_{50}$ ; open circles  $D_{84}$  for the Valleluce river, Cassino. Closed triangles;  $D_{50}$ ; open triangles;  $D_{84}$ , for the Fosso Tascino, Leonessa. Closed Diamonds;  $D_{50}$ , open diamonds  $D_{84}$  for the Rio Torto, Fiamignano. The dashed line indicates the  $D_{84}$  grain-size trend for the Rio Torto. All other data sets show approximately constant grain-size downstream.

ally low in the headwaters and before the convex reach (most values < 0.05), whereas slopes are generally > 0.05 $(3^{\circ})$  between the break in slope at 6 km and the fault. Maximum slopes here can reach > 0.3, and minimum documented slopes increase all the way to the fault. Slopes decline rapidly to values < 0.04 on crossing into the hangingwall. The variation in channel slope is positively correlated with the ratio of high-flow width to valley width,  $W_{\rm b}$ /  $W_{\rm v}$  (correlation coefficient = 0.5). Low channel slopes occur where the Rio Torto flows through wide open valleys in the upper part of the catchment, but the increase in slope in the convex reach is immediately matched by narrowing of the valley, forming a deeply incised gorge where  $W_{\rm b} \sim W_{\rm v}$  This focuses fluvial erosion into a corridor < 10 m wide through the footwall, and permits incision directly into bedrock as the river approaches the fault.  $W_{\rm b}$ /  $W_{\rm v}$  falls to very low values as the river enters the hangingwall basin. The correlation between S and  $W_{\rm b}/W_{\rm y}$  suggests that channel steepening is directly linked to incision and gorge formation near the fault. Additionally, Whittaker et al. (2007) show that this signal is transmitted to the hillslopes throughout the over-steepened reach, giving hillslope gradients  $> 30^\circ$ , wherever local channel slopes are high and valley widths low.

These data indicate that the Rio Torto shows systematic changes in key hydraulic geometry variables as the river approaches the fault, in contrast to the Fosso Tascino and Valleluce rivers above. These geomorphological signals are dramatic, and are clearly evidenced by the fact that it



**Fig. 10.** Channel aspect ratio as a function of local channel slope: open circles, Fosso Tascino, Leonessa. Grey circles, Valleluce river, Cassino. Black circles, Rio Torto, Fiamignano. Line shows best-fit power-law for Fiamignano data.

would be easy to predict the likely position of the Fiamignano fault at 10.5 km downstream using the channel data in Fig. 8 alone. This suggests that rivers crossing faults develop diagnostic signals in some circumstances, but apparently not in others, and we discuss the causes of this phenomenon in 'Discussion – explanations for differing channel behaviour'.

#### Grain size

Figure 9 shows sediment calibre (Wolman, 1954), for each of the three rivers, against downstream distance, L, normalised by distance to the fault,  $L_f$ . While the Rio Torto, Fiamignano, has the coarsest median grainsize (twice as large as the Valleluce river, Cassino), it is noticeable that for all three channels, D<sub>50</sub> does not vary greatly as the rivers flow towards the active faults. On average,  $D_{50} \sim 3.5$  cm for the Rio Torto, ~1.9 cm for the Fosso Tascino, near the Leonessa fault, and  $\sim$ l cm for the Valleluce river crossing the Cassino fault. However D<sub>84</sub> responds differently: while the rivers crossing the constant slip-rate faults (Fosso Tascino and Valleluce) maintain constant coarse fraction grain-size within the footwall of the fault,  $D_{84}$  increases in the Rio Torto from  $\sim 6 \,\mathrm{cm}$  near the start of the convex reach to  $\sim$ 9.5 cm near the fault. It decreases again to  $\sim$ 2 cm once the river enters the hanging wall of the fault. This therefore means that the spread in sediment size-distribution increases downstream in the Rio Torto as the river flows through the incised gorge upstream of the fault. Because the hillslopes in the Rio Torto are directly coupled to the incised channel, and there are a number of landslides directly entering the channel in the gorge, we interpret the increase in  $D_{84}$ , but not  $D_{50}$ , to represent an increase in coarse sediment input sourced directly from the neighbouring hillslopes. This is an additional component to the finer material sourced from upstream in the case of the Rio Torto, whereas coarse landslide-derived debris does not appear to be a significant input in either of the channels crossing the Cassino or Leonessa faults.

### DATA ANALYSIS

# How does channel aspect ratio vary in areas of active tectonics?

It has recently been hypothesised, with support from simple hydrological and erosional models, that the channel aspect ratio,  $W_{\rm b}/H$ , is constant downstream in bedrock rivers (Finnegan et al., 2005). However, if channel narrowing is a ubiquitous way in which rivers respond to steeper slopes (Turowski et al., 2006; Whittaker et al., 2007), then for rectangular channels (e.g. in gorges) for aspect ratio to remain constant, channel depth would have to also decrease by the same amount, which would consequently require flow velocity to increase by the square of the difference in order to maintain constant discharge. Figure 10 shows  $W_{\rm b}/H$  as function of local channel slope for the three channels considered. Most striking is the data for the Rio Torto, Fiamignano: here, we see a strong non-linear dependence of aspect ratio on slope, with high slopes > 0.1 typically correlated with low-aspect ratios ( $W_{\rm b}/H < 6$ ). This implies a deepening and a narrowing of the channel in the steep gorge as the river approaches the fault, as this is the zone of maximum slope (Fig 8b.) The relationship can be empirically fitted with a power law, giving  $W_{\rm b}/H$ ,  $\sim S^{-0.34}$ , and underlines the significant effect that active faulting has on hydraulic geometry in this setting, by controlling local channel slopes. In contrast, the signals for the constant slip-rate faults are much less clear. The Fosso Tascino, crossing the Leonessa fault, exhibits a much wider scatter in  $W_{\rm b}/H$ : average slopes in the catchment are considerably lower, < 0.05 and this is associated with  $6 < W_{\rm b}/$ H < 14. Despite this variability, there is a trend towards lower aspect ratio at higher slopes, as shown by the two data points taken in the steep headwaters where S > 0.3. For the Valleluce river, crossing the Cassino fault, recorded slopes do not exceed 0.13, and average  $W_{\rm b}/H \sim 5$ , but again there is a trend towards lower aspect ratios as local channel slope increases. All of these data suggest that there is an underlying tendency for channel aspect ratio to lower in areas of higher slope, as one might expect in the headwaters of the channel. However, in areas of tectonic activity, as shown here, slopes can be high even at relatively large drainage areas  $(>10 \text{ km}^2)$ , and in the Rio Torto case C,  $W_{\rm b}/H$  is much more tightly constrained as a function of slope, indicating that local channel gradient changes are transmitted directly to channel aspect ratio. That is to say, a functional dependence of channel aspect ratio on local slope is not necessarily a product of a transient response to tectonics, but is most clearly seen under transient conditions, because high channel slopes are more abundant downstream. Consequently, the results suggest that in tectonically perturbed areas, it is not justified to assume a constant aspect ratio. For the Rio Torto, width varies by a factor of  $\sim 10$  downstream, whereas depth varies by a factor of  $\sim$ 2, so most of the signal lies in  $W_{\rm b}$  variation. This study suggests that understanding the evolution of downstream channel width is therefore vital to understanding river response to tectonics.



**Fig. 11.** Highflow channel width as function of downstream distance (a, b, c) and drainage area (d, e, f) for the Fosso Tascino, Valleluce and Rio Torto rivers, respectively. Open circles depict mean measured widths, with error bars showing 1 SD for (a–c). The black line gives width predictions for each catchment according to  $W \sim A^b$ , as deduced in (b, d, f).

# Do channel widths scale with drainage area for rivers crossing active faults?

As we show in 'Background and paper aims', knowledge of channel width is required to predict fluvial erosivity at any point downstream in a channel and hydraulic scaling (Leopold & Maddock, 1953) is therefore used in most models to constrain  $W_b$  Eqn. (1). Typically, the scaling exponent b = 0.5, although some studies make use of field data which indicate that b values may differ from this as uplift rate increases (e.g. Duvall *et al.*, 2004). However, the raw data presented in 'Study rivers', and the discussion of aspect ratio, above, raise the issue of whether it is reasonable to assume that channel dimensions, such as width, can be expressed in terms of a simple power-law dependency on discharge, regardless of exponent used. We therefore assess the applicability of the  $W \sim Q^b$  paradigm by comparing predictions of high flow channel width yielded from best-fit power-law scaling relationships deduced using the real field measurements with downstream evolution of measured widths on the scale of the uplifted footwall block itself, binned in ~500-m intervals (Fig. 11). Drainage areas are obtained from a 20-m resolution DEM.

For the Fosso Tascino, Leonessa, non-linear regression of measured  $W_{\rm b}$  and A yields  $W_{\rm b} \sim A^{0.55 \pm 0.05}$ ,  $r^2 = 0.8$ (Fig. 11d). When width predictions from this relationship are compared with measured  $W_{\rm b}$  values as they evolve downstream in the uplifted horst, we find that the hydraulic scaling approach does give a reasonable fit to measured values (Fig. 11a). Similar conclusions can be drawn for the Valleluce river crossing the South Cassino fault: here, we obtain  $W_{\rm b} \sim A^{0.51 \pm 0.03}$ ,  $r^2 = 0.97$  (Fig. 11e) and when we compare width predictions made from this relationship with the real data again we see a good agreement between the deduced hydraulic scaling relationship and the distribution of real channel widths (Fig. 11b): i.e. predictions from hydraulic scaling lie within the error bars in  $W_{\rm h}$  considered over distances of  $\leq 500$  m. In these cases powerlaw scaling is adequate to describe the evolution of channel planform downstream; indeed we produce values very similar to the expected exponent of b = 0.5 (Montgomery & Gran, 2001).

However, a very different picture emerges when we use the same method on the Rio Torto crossing the Fiamignano fault. Regression of  $W_{\rm b}$  taken over  $\sim$ 3 orders of magnitude gives  $W_{\rm b} \sim A^{0.45 \pm 0.04}$ ,  $r^2 = 0.9$  (Fig. 11f). This is a somewhat lower b-value, but is similar to those documented in other studies for rivers in tectonically active areas (e.g. Duvall et al., 2004). However, when we look in detail at the predictions of this relationship with the actual downstream evolution of channel widths, it becomes immediately apparent that this scaling analysis is not effective for describing downstream channel evolution in the Rio Torto catchment. In particular, at a major tributary where the drainage area doubles at 8.5 km downstream, there is no immediate increase in channel width in the gorge: instead it remains at <10 m and even narrows slightly until the river crosses the fault. Beyond this, channel widths recover to nearer the predicted values of  $\sim$ 18 m. Near the fault the channel is  $\sim$ 3 $\times$  narrower than empirical predictions of width from traditional hydraulic scaling might imply. Consequently, in the zone of maximum uplift just upstream of the fault,  $W_{\rm b}$  is clearly decoupled from drainage area, and hence the hydraulic scaling paradigm is at its least effective in the region where the river is most sensitive to tectonics.

This analysis shows that even if one does calculate a *b* exponent from field data, much information about downstream evolution of channel widths can be lost. The problem is that for a single drainage area of  $\sim 7 \times 10^8 \text{ m}^2$ , widths range from < 5 m to > 20 m, as shown in Fig. 11f. However, what cannot be deduced from Fig. 11f, but which is clearly apparent in Fig. 11c is that this variation is actually *systematic* downstream, despite there being little change in drainage area. This means that a single *b* exponent, regardless of magnitude, cannot realistically describe channel evolution downstream. The measured width values are informative, and tell us about how the river is responding to fault-induced uplift, whereas the predicted widths mask this signal in the gorge. This loss of scaling (Fig. 11c) is due to the strong non-linear dependence that aspect ratio has on channel gradient, and shows that local slopes may be as important as discharge or drainage area in determining *W*. We therefore argue that power-law predictions of channel width must be used with caution in tectonically disturbed areas, *even* if they are generated from real field data (cf. Duvall *et al.*, 2004), and in 'From transient landscape to topographic steady state' we evaluate the timescale over which such a loss of hydraulic scaling may be regained within the fluvial network.

### What role does grain-size play in governing process and form in channels shaped by active tectonics?

A crucial aspect of the fluvial system is the sediment that the channel carries. Detachment-limited models of erosion (cf. Howard & Kerby, 1983) parameterise bedrock erosion as a function of bed shear stress, and do not explicitly include sediment flux from upstream. The assumption in this case is that the transport capacity of the flow is  $\gg$  sediment supply. Alternatively, if sediment supply is in excess of the river's capacity to transport, then the river is said to be transport-limited, and incision can be modelled as being proportional to the downstream divergence of sediment flux (Tucker & Whipple, 2002). Systems governed by these end-members respond differently to transient forcing, because they are underlain by very different mathematics (Whipple & Tucker, 2002). In general, transport limited systems tend to respond diffusively, whereas systems close to the detachment limited end-member show a more 'wave-like' response, with convexities in long profiles common. Differing channel responses could therefore be explained by differing long-term erosional dynamics. Unfortunately, although often attempted, predicting the dominant process from channel observations alone is non-trivial; for example a channel with 100% alluvial cover could be scouring bedrock at high stage if the sediment is merely a thin, ephemeral veneer; moreover, the propensity of sediment to act as tools or cover within a river depends on the distribution frequency of high flow events throughout the year, and the peakedness of the storm hydrograph (Knighton, 1998; Sklar & Dietrich, 2001).

We do not aim to quantitatively test erosion models here, but rather to assess whether it is likely that the three channels could be transport-limited, or whether detachment-limited processes may govern long-term erosion rates. Initial observations of bed characteristics for the Fosso Tascino channel (case A, near Leonessa), show exposures of some bedrock in the headwaters of the channel, but downstream of this the bed is covered by an alluvial veneer of > 0.2 m thickness. In contrast, in the Rio Torto (case C), bedrock exposure is approximately 10–20% in the headwaters of the river, but downstream of the slope increase at L = 6 km, typical exposure is >50%, with some reaches exposing considerably more than this. Alluvium, where present, never appears to be more than 0.5 m thick. Bedrock in the base of the channel is polished, showing signs of abrasive wear, and jointed horizons show evidence for plucking. The Valleluce river at Cassino (case B) has small exposures of bedrock, usually <20% of the channel bed, and never more than 50% at any site. Again there is evidence for scour in the base of the channel. Although these observations suggest that the Valleluce and Rio Torto channels (cases B and C) are strongly under-supplied (closer to the detachment limited end-member) they cannot provide a definitive answer alone. However, we can assess this issue quantitatively by looking at the Shields stress,  $\tau^*$  (Mueller & Pitlick, 2005), which represents the ratio of the basal shear stress ( $\tau_b = \rho_w gRS$ ) to the excess sediment density and size, and is evaluated as:

$$\tau^* = \frac{\rho_{\rm w} RS}{(\rho_{\rm s} - \rho_{\rm w}) D_{50}} \tag{3}$$

where R is the hydraulic radius of the channel, S is the local channel slope,  $\rho_s$  is the density of the sediment  $({\sim}2650\,kg\,m^{-3})$  and  $\rho_w$  is the density of water  $(\sim 1000 \text{ kg m}^{-3})$ . Importantly, for *transport-limited* gravelbed rivers, where the predominant transport mechanism is bed-load saltation rather than suspension, channels configure themselves so as to maintain a critical dimensionless shear stress downstream,  $\tau_{cr}^*$ , which has typically been measured to lie in the range 0.047-0.06 (Dade & Friend, 1998; Dade, 2000) Data also suggest that in these cases, Shields stresses do not exceed critical values by more than 20% (Mueller & Pitlick, 2005) In other words, for gravel bed rivers, grain-size helps to control channel slope. However, these relationships have been derived for alluvial rivers, and channels, which are not transport limited are not constrained into any such range.

Figure 12 shows Shields stress as a function of normalised fault distance, L/Lf for the (a) the Fosso Tascino, Leonessa, (b) the Valleluce river, Cassino and (c) the Rio Torto, Fiamignano. We calculate values using  $D_{50}$  (black circles) and also using  $D_{84}$  (open circles) to assess whether the coarsest grain-sizes in the channel are also likely to be mobilised at high flow. Because  $D_{50}$  and  $D_{84}$  values do not vary greatly downstream in each of the three rivers (Fig. 9) we take measured grain-sizes to be representative of sediment calibre both up- and downstream from the local measuring site. (The one exception is  $D_{84}$  for the RioTorto, where we extrapolate the grain-size trend for 0.7 < L/Lf < 1.2.)

It is immediately apparent for all of the channels that values for  $\tau^*$  are considerably in excess of the typical threshold for self-formed gravel bed rivers: and lie in the range  $0.5 < \tau^* < 8$ . This is between nine and 80 times the typical gravel bed threshold, and for transport-limited alluvial systems has *only* been documented for lowland rivers, with very fine grain sizes, where the dominant mode of entrainment is suspension (Dade & Friend, 1998). The high values obtained are the result of relatively small and homogeneous gravel supply in the Fosso Tassino and Valle-



Fig. 12. Shields stress against downstream distance for (a) Fosso Tascino, Leonessa, (b) Valleluce river, Cassino, and (c) Rio Torto, Fiamignano. Closed circles show values calculated using catchment  $D_{50}$  values, and open circles show values calculated using measurements of  $D_{84}$ . The grey dotted line indicates the threshold value for critical shear stress in gravel-bed (i.e. transport limited) rivers.

luce rivers, and by the high slopes seen upstream of the active fault in the Rio Torto, near Fiamignano. Consequently, most of the sediment will be moving at bankful flow and this would be the case even if we had under-estimated  $D_{50}$  by an order of magnitude. Moreover, the data suggest that the increase in coarse grain-size fraction found in the Rio Torto is unlikely to result in the break-down of hydraulic scaling, documented, because of large, immobile clasts blocking the channel as suggested by Wohl (2004):

Wolman  $D_{84}$  is only two to four times larger than  $D_{50}$ values, and it is highly improbable that we have under estimated this value by the amount required to produce  $\tau^* = 0.06$ . Indeed, it is only in the headwaters of the Rio Torto, where the river lies in the hangingwall of the Sella di Corno fault, that we see Shields stresses approaching typical 'transport-limited' values previously documented for gravel bed rivers in alluvial settings. It is, therefore, unlikely that sediment size is the dominant control on local channel slopes; instead in the case of the RioTorto, Shields stress is closely correlated with slope, suggesting that the channel does not respond to steepening by increasing grain size in order to maintain constant critical Shields stresses (cf. Harbor, 1998). These data, in combination with more qualitative observations of channel process and bedrock exposure, allow us to reject the idea that these channels are transport limited, and we therefore suggest that they are all likely to be under supplied to a greater or lesser extent. Furthermore, the data suggest that the loss of hydraulic scaling in seen in the Rio Torto gorge, upstream of the Fiamignano fault, cannot be explained in terms of a grain-size or roughness effect (cf. Wohl, 2004)

## DISCUSSION – EXPLANATIONS FOR DIFFERING CHANNEL BEHAVIOUR

The data sets presented above are the first to compare channel geometries developed in response to active tectonic forcing where we know the spatial and temporal boundary conditions explicitly. The three rivers, although located in the same region, flowing over very similar lithologies, and all crossing active faults, demonstrate contrasting fluvial responses to the imposed tectonic regimes they face. In particular, the Rio Torto exhibits a convex reach above the fault, and shows systematic deviations in hydraulic geometry, valley width and aspect ratio with proximity to the fault, elevated coarse fraction grainsize, a break-down in width scaling and a reduced variability in channel planform over short length-scales. The three rivers all appear to approach the detachment limited endmember, and the fact that these features are not generally seen on the other two channels suggests that neither can the downstream presence of an active fault, nor the structural style of a back-tilted fault-block be a sufficient explanation for the differences. We, therefore, need a more sophisticated interpretation if we are to account for why some rivers crossing faults have convex reaches above them and others do not. It is true that the Fiamignano fault has a slip rate, which is  $\sim 3 \times$  higher than either the Leonessa or Cassino cases, so it could be suggested that the Rio Torto cannot keep up with the higher uplift rate on the fault. We doubt this explanation is correct because there is no bedrock scarp preserved in the channel where the river crosses the Fiamignano fault, suggesting that at that point at least, the rate of fluvial incision is at least equal to the rate of uplift in the channel.

However, convex river profiles have been modelled to develop as a *transient* response to a *change* in uplift rate for channels approaching the detachment limited end-member model for river incision. (Tucker & Whipple, 2002; Whipple & Tucker, 2002). This exactly describes the Rio Torto situation, where a bedrock channel, which is not limited by its ability to transport sediment, crosses an active fault, which underwent a three-fold increase in slip rate at 0.75 Ma. We, therefore, make the interpretation that the development of a convex reach and the systematic deviations in channel form are a transient response of the fluvial system draining the footwall of the Fiamignano fault to the documented slip rate increase (Cowie & Roberts, 2001).

### **Defining landscape state**

To demonstrate this interpretation, we need to show both that the long profile and channel geometry of the Rio Torto are indeed transient forms that will evolve away from their current configuration over time, and that the response seen is controlled by fault acceleration, and not, for example, by regional base-level fall. To address the first issue, we need to be clear about what we mean by the terms 'equilibrium', 'steady state' and 'transient response' with respect to rivers. Hydrologists talk about rivers being in equilibrium if they have reached an optimal state by obeying energy considerations such as constant energy dissipation per unit area of channel, and minimised global energy dissipation across the network (Rinaldo et al., 1992; Rodriguez-Iturbe et al., 1992). Deviations from this norm could be thought of as a disequilibrium condition from which rivers may (over a presumably long timescale) attempt to recover.

In contrast, the issue of equilibrium for tectonic geomorphologists is more often cast in the language of topographic steady state (e.g. Lavé & Avouac, 2001; Tomkin et al., 2003). For example, a river crossing a zone of active uplift could adjust itself so that its incisional capability matches the range of rock uplift rates at all points. Such a river would then have reached a topographic equilibrium or steady state (sensu Willett & Brandon, 2002). However, assuming that the ability of a channel to incise is a function of energy expenditure on the bed (Finlayson et al., 2002; Finnegan et al., 2005) such a river would not be in hydraulic energy equilibrium. A third class of fluvial systems, meeting neither of these conditions would be in disequilibrium with respect to both, and might be expected to show transient behaviour. These three sets of conditions (energy equilibrium, topographic steady state, and transient response) could clearly produce rivers systems with very different geomorphic characteristics. We therefore, explicitly test whether the three rivers crossing faults in this study have achieved (i) energy dissipation equilibrium, (ii) topographic steady state or (iii) neither. We then investigate whether the differing temporal history of slip on the three faults is indeed the best explanation for the varying fluvial responses seen.

## Do the rivers have constant dissipation of energy downstream?

To calculate energy dissipation (Watts) per unit channel area  $(Wm^{-2})$ , we use the unit stream power,  $\omega$ , expressed as

$$\omega = \rho g Q S / W_b \tag{4}$$

where is  $\rho$  is the density of water, g is the acceleration due to gravity. Unit stream power is commonly used as an incision rate proxy for channels at (or near) the detachmentlimited end-member, and has been used to track variations in erosivity in both quiescent and tectonically active areas (Dadson et al., 2003; Duvall et al., 2004). We use the measured width data ('Study rivers') to calculate unit stream power. To derive discharge estimates for each river, we apply Manning's equation (Manning, 1891) to channel crosssections measured near the faults, allowing us to predict fluid velocity and hence a characteristic discharge at the fault. We scale this estimate for Q with drainage area to calculate discharge both up and downstream of the fault. This assumes that A is proportional to Q, which is reasonable for catchments of limited area (cf. Solyom & Tucker, 2004). We obtain discharges at the fault of 100, 110,  $60 \,\mathrm{m^3 \, s^{-1}}$ , for the Fosso Tascino, Rio Torto and Valleluce channels, respectively. These values represent storm runoff rates on the order of  $10 \text{ mm h}^{-1}$ , and are comparable with flood discharges measured on gauged rivers in the Italian Apennines with similar drainage areas (e.g. Ratto et al., 2003).

The three rivers considered show remarkably different energy expenditure patterns (Fig. 13). While they have similar values in the headwaters, the Rio Torto (Fig 13c) has unit stream powers  $> 20000 \text{ Wm}^{-2}$  as the fault is approached, and shows a very large increase between 8 and 10.5 km downstream. This channel is evidently not distributing its potential energy uniformly downstream. The increase in unit stream power is driven by high channel slopes between 6 and 10.5 km downstream, and by restricted channel widths, particularly beyond 8km downstream, where widths remain low despite a large increase in drainage area. Therefore, the increase in stream power is a direct result of the loss of hydraulic scaling, and the convex long profile. In contrast, the Fosso Tascino, crossing the Leonessa fault (Fig. 13a) shows hardly any increase, with unit stream powers varying only from 1000<  $\omega < 3000 \,\mathrm{Wm}^{-2}$  downstream; there is no marked change in these values as the fault is neared and the distribution could be adequately modelled as being approximately constant downstream. This is consistent with the fact that  $W\!\sim\!\!A^{0.5}$  (Fig. 11) and  $S\sim\!\!A^{-0.5}$  (Fig. 4) for this river, which would produce constant values of  $\omega$  if these relationships alone were substituted in Eqn. (4).

The Valleluce river, Cassino (Fig 13b) shows considerable scatter, but on average there is an increase in stream powers from  $< 2000 \text{ Wm}^{-2}$  in the headwaters to values which plateau around  $4000-6000 \text{ Wm}^{-2}$  near the fault. Energy expenditure falls again in the hangingwall. However, the river also has a concave up profile and apparently



Fig. 13. Unit stream power against downstream distance for (a) Fosso Tascino, Leonessa, (b) Valleluce river, Cassino and (c) Rio Torto, Fiamignano. Coloured bands illustrate pattern of stream power distribution and include >90% of all values; the black lines are the uplift fields for each fault as deduced in 'Discussion – explanations for differing channel behaviour'. (b) The open diamonds represent stream power normalised by valley width (i.e.  $QS/W_v$ ) and the dotted lines delimit all values for this ratio. The anomalously high stream powers around 6 km downstream on the Valleluce river correlate with the band of resistant calcareous sandstone, also shown on Fig. 6. Stream powers associated with this lithological feature are plotted as smaller symbols and are discounted when considering catchment-wide stream power trends.

good width scaling. This means that the emergence of elevated stream powers in the proximal footwall must be related to unevenly distributed residuals in width or local channel slope, despite the apparently good scaling overall. To test this, we consider the downstream distribution of the ratios  $W_b/W_{\text{predicted}}$ ,  $S/S_{\text{predicted}}$  (Fig. 14), where  $W_{\text{predicted}}$  are width predictions from non-linear regression of  $W_b$  and A (Fig. 11), and  $S_{\text{predicted}}$  is predicted



Fig. 14. Normalised channel widths  $(W/W_{\text{predicted}})$  and local slopes  $(S/S_{\text{predicted}})$  against downstream distance for the Valleluce river. If predicted widths and slopes are a good descriptor of field data, normalised values should cluster around 1 (grey bar).

channel slope, also derived from regression of S and A. The data indicate that while  $W_{\rm b}/W_{\rm predicted}$  cluster around a value of 1, showing that the scaling relationship in Fig lle is a good approximation,  $S/S_{\text{predicted}}$  is low in the headwaters, suggesting that power-law scaling over-predicts slopes here. S/S<sub>predicted</sub> gradually increases downstream, meaning that the channel is steeper near the fault than slope predictions would suggest. Moreover, S/S<sub>predicted</sub> is highly correlated with unit stream power (correlation coefficient = 0.91). Consequently, it is variation in local channel gradient, and not high-flow widths which enables the Valleluce river to increase its stream power in the vicinity of the Cassino fault; we note that these local slope changes are superimposed upon a river profile with a typical concavity overall (0.51) (cf. Kirby et al., 2003). Smallscale changes in slope, where the rate of change of drainage area is low can thus be an important way in which channels adjust to fault-induced uplift. These adjustments might easily be missed on log-log plots of slope and area but are clearly visible on a linear plot of slope vs. downstream distance. The observations suggest that only the Fosso Tascino (Fig. 13a) approaches energy dissipation equilibrium as defined above, and if stream power is taken to be a proxy for erosion rate (cf. Dadson et al., 2003) only this channel would have an appoximately constant erosion rate downstream.

# Are these catchments in topographic steady state?

To investigate the extent of topographic steady state we need to (a) reconstruct the distribution of uplift in the footwall blocks drained by the rivers, and (b) take account of any externally driven base-level changes as summarised in 'Study rivers'. An indication of whether any of the channels have reached topographic steady state can be assessed by comparing the distribution and wavelength of unit stream power with the calculated tectonic uplift field.

One issue is the division of uplift between the footwall and the downthrown hanging wall for each of the faults. Estimates of this range from ratio of 1:6 (Stein & Barrientos, 1985) to values of 1:1 for 'domino' blocks in the Basin and Range (Anders et al., 1993). However, as rivers crossing faults detect only the relative difference in uplift rate (i.e. of the footwall to the hangingwall) as the fault is crossed (Tucker & Whipple, 2002), the precise distribution is not important for our purposes. Indeed, the absolute base-level change experienced by the river as it crosses the fault must be the difference in tectonic uplift rate (footwall to hangingwall), minus any sediment aggradation or alternatively, *plus* any incision in the half-graben basin bounded by the fault. In the following sections, we apportion the uplift field equally between the hangingwall and the footwall, and we explicitly account for documented sediment aggradation and incision in the hangingwall ('Study rivers'), allowing us to reconstruct the absolute base-level changes affecting the catchments. Anders et al. (1993) also showed that a flexural model for footwall uplift is indistinguishable from that of a rigid tilted block when the fault spacing is  $\leq$  3 times the flexural wavelength. For central Italy where fault spacing is on average < 12 km and the flexural wavelength is  $\sim 10 \text{ km}$  (cf. D'Agostino & McKenzie, 1999) the tilted block model is thus an adequate model to reconstruct a footwall uplift profile. We therefore use linear extrapolation to calculate the distribution of footwall uplift from the fault to the fulcrum of the normal fault. This is consistent with seismic profiles across the Apennines (e.g. Cavinato et al., 2002).

#### Case A - Fosso Tascino (Leonessa fault)

Because the river incises the central section of an uplifting horst this is the simplest uplift field to constrain. To first order, the river is experiencing a spatially and temporally uniform tectonic uplift rate of  $\sim 0.3$  mm year<sup>-1</sup> (Fig. 3a). The hangingwall has also undergone aggradation of  $\sim$  320 m since fault initiation (  $\sim$  3 Ma) followed by up to 50 m of incision since 0.75 Ma ('Case A - Horst uplift'). Combining this information (Fig. 15a), it implies that where the Fosso Tascino crosses the Leoessa fault it has experienced a relative uplift rate difference of  $\sim 0.25 \text{ mm year}^{-1}$  until 0.75 Ma. If the incision since 0.75 Ma has taken place uniformly since, and has only affected the hanging wall, a maximum estimate of the relative uplift rate difference seen by the river of  $\sim 0.4 \,\mathrm{mm \, year^{-1}}$ for the period from mid-Pleistocene to present can be generated.

In Fig. 13a, we compare the tectonic uplift field with the distribution of stream power from the headwaters to beyond the fault. Within error, energy expenditure is constant downstream, implying a constant incision rate assuming the river lies near the detachment-limited end-member ('Do the rivers have constant dissipation of energy downstream?'). Moreover, as the river also experiences a constant tectonic uplift rate and has the typical



Fig. 15. Base-level history for (a) Fosso Tascino, Leonessa and (b) Rio Torto, Fiamignano. The channels feel the effect of the throw on the fault,  $\pm$  any sediment aggradation/incision in the hanging wall.

morphology of an equilibrium channel, then these observations together suggest that the river has reached topographic steady state, where the rate of uplift balances the rate of incision, despite the 50 m base-level fall in the last 0.75 Ma.

There are two ways of reconciling this apparent topographic steady state with the documented base-level fall: firstly, that the timescale of response to relatively small (i.e. 50 m) base-level falls is rapid and has already been transmitted through the system. Alternatively, it could be argued that these river systems are relatively insensitive to small changes in the base-level of the hangingwall: i.e. that relative uplift rate perturbations of less than 2 times are not sufficient to force significant catchment steepening or narrowing. Moreover, any residual knickpoint produced from this base-level fall must be < 50 m tall (which is what one would obtain from an instantaneous base-level drop of this magnitude at the river outlet; in reality the process would likely have been more gradual). As such, a knickzone would also degrade as it migrated upstream, it would, therefore, be difficult to identify unambiguously, and would be unlikely to impact significantly on catchment-wide estimates of unit stream power.

#### Transient response of rivers crossing active normal faults

#### Case B - Valleluce river (South Cassino fault)

As in the Leonessa example, above, we model the river as having experienced a constant  $\sim 0.3 \,\mathrm{mm \, year^{-1}}$  vertical uplift rate at the fault. However, for this back-tilting case the uplift decays away to the NE, in a direction perpendicular to the fault. We assume the fulcrum of the fault is positioned at 6 km into the footwall as this is half the typical fault spacing in the Southern part of the array. There is little evidence for any incision in the Cassino hangingwall since the Pliocene; instead base-level has remained the aggredational hangingwall plain that leads out to the sea ('Study rivers'). If sedimentation filled all the available hangingwall accommodation space, then the river would only be affected by the footwall uplift signal. In fact, the elevation difference between the hanging wall and the footwall observed today is generally > 600 m (Fig 3b), while there is good evidence that this area was a marine planation surface in the early Pliocene (Galadini et al., 2003). Hence, the Valleluce river is likely to have experienced a constant relative uplift of at least  $0.2 \,\mathrm{mm \, year^{-1}}$  since the initiation of faulting.

Stream powers in the Valleluce river (Fig. 13b) suggest that the channel cannot be in energy equilibrium: instead, incisional capability apparently increases towards the zone of maximum relative uplift rate near the fault. In general, the wavelength and pattern of tectonic uplift is similar to the stream power distribution along the river so the river appears to have reached topographic steady state. Nevertheless, it is noticeable that near the fault the stream power signal, although elevated on average, is quite diffuse, with individual values covering a range of 3000-7000 Wm<sup>-2</sup> in the 2 km upstream of the fault. However, we also documented a progressive decrease in valley width near the fault (Fig. 6c, where  $W_{\rm b}/W_{\rm v}$  increases from ~0.3 in the headwaters of the channel to  $\sim 0.7$  near the fault). This means that fluvial erosion processes will be concentrated, over long timescales, in a narrower zone near the fault than in the headwaters of the channel. If we normalise unit stream powers by this ratio [i.e.  $(\omega W_{\rm b})/W_{\rm w} = QS/W_{\rm y}$ ] as suggested by Pazzaglia et al. (1998), then we do find the increasing stream power more clearly mirrors the uplift distribution in the hanging wall (open diamonds, Fig. 13b) i.e. we see that 'valley width' stream powers decrease by  $\sim$ 50% from the fault to a point 4 km upstream, and this is mirrored by the uplift profile which declines from  $\sim$ 0.14 to  $\sim$ 0.07 mm year  $^{-1}$  over a similar distance. In particular, the range of stream powers values near the fault spans approximately 2000 Wm<sup>-2</sup>, which is half that of the values calculated with bank-full width measurements. In other words, valley narrowing helps the river to keep pace with fault uplift.

These data therefore indicate that uplift on the Cassino fault is likely balanced by long-term incision in the footwall (i.e. topographic steady state), and that valley width adjustments are also a key component of the way in which rivers adapt to tectonic forcing to maintain topographic steady state. Note that neither  $W_v$  changes nor

the higher local slopes (Fig. 14) would be resolved using traditional hydraulic scaling approaches to predict incision rate.

### Case C - Rio Torto (Fiamignano fault)

Because the Fiamignano fault is bounded to the NE by the Sella di Corno fault, its footwall constitutes the hanging wall of this latter fault. We again model the footwall as being a rotating, rigid block (cf. Anders et al., 1993). Fault spacing in this area is only 7-8 km, so if they had similar displacements and throw rates, we would expect the point of zero uplift, i.e. the fulcrum, to be at 3.5-4 km into the footwall of the Fiamignano fault. In reality, the Fiamignano footwall is bounded by the tip of the Sella di Corno fault (where throw and throw rates are lower), so we estimate that the fulcrum to be  $\sim$ 5 km into the footwall. We therefore permit the uplift field across the footwall block to decline linearly to zero over 5 km in a direction perpendicular to the fault-strike, and use this to calculate the relative tectonic uplift field as a function of downstream distance in the Rio Torto, as shown in Fig. 13c.

As we have seen, the Rio Torto shows dramatic increases in unit stream power within the gorge developed near the fault and clearly does not dissipate energy evenly downstream. More importantly, it is unlikely to be in steady state because the wavelength and magnitude of the stream power increase does not match footwall uplift rate: average  $\omega$  increases by >5 times in a downstream distance of 2.5 km, and over an order of magnitude from the headwaters. In contrast, footwall uplift only decreases by 20% in the 2.5 km upstream of the fault assuming linear decrease in uplift rate. To get the uplift field to decline by a factor > 5 over 2.5 km, would require unrealistically low values of elastic thickness, i.e. < 1 km, much lower than values that have been estimated for this area (  $\sim 4 \text{ km}$ , D'Agostino & McKenzie, 1999). Consequently, we infer that the Rio Torto is exhibiting a transient response, because uplift is not balanced by incision at all points along the channel. Incorporating valley width does not make much difference to the stream powers achieved in the gorge because  $W_{\rm b} = W_{\rm w}$  but it does significantly reduce stream powers downstream of the fault where valley widths are very high. Thus, incorporating valley widths only serves to enhance the disparity between the uplift field on the fault and the distribution of stream power along the river.

Is this transient response due to slip acceleration on the Fiamignano fault at 0.75 Ma or it could be due to (regional) base-level fall in the hangingwall of the fault? As argued in 'Case C- tilted fault block, increased throw rate (Rio torto, Fiamignano fault'), between the initiation of faulting at 3 Ma and the late Pliocene, approximately 100 m of sediment aggraded near the exit of the Rio Torto from the gorge. This sediment was then incised by 100–280 m, and probably in-filled subsequently by up to 50 m of Villafranchian sediment by the Mid-Pleistocene. These sediments have been stripped away since then. Figure 15b sum-

marises the cumulative effect of these base-level changes: we assume that all the incision took place in the hangingwall (which would maximise the rate difference at the fault) and use an average estimate of incision for between 1.8 and 0.75 Ma of 200 m. For the period of 3-0.8 Ma, we can fit an increase in relative uplift rate of <2 times, but given the assumptions made in calculating both base-level and throw on the fault through time, it is hard to argue that this is materially different from using a constant relative uplift rate of 0.35 mm year<sup>-1</sup> for this period. From 0.75 Ma to present, the relative rate seen by the channel largely tracks the total (tectonic) accumulation of throw. The following observations therefore suggest that the transient response above is due to tectonics, and not due to externally controlled base-level change:

- Although the acceleration in throw rate coincides with incision of mid-Pleistocene hangingwall sediments, this would only enhance the signature by approximately 20%.
- (2) Other rivers entering the hangingwall basin that do not cross the Fiamignano fault do not show over-steepened reaches, despite the same base-level history.
- (3) The rate of base-level fall seen by the Rio Torto before 0.75 Ma appears to be virtually the same as in the Fosso Tascino, and this has not resulted in a significant oversteepened reach.
- (4) Even if the total base-level change due to externally driven hangingwall incision were preserved in the long-profile of the river, the over-steepened reach would have an elevation difference considerably less than the 400 m observed, demonstrating that tectonics is the dominant control.

We, therefore, feel confident in asserting that the Fiamignano fault (a) is not in energy equilibrium (b) has not reached topographic steady state and (c) is undergoing a transient response to fault acceleration at 0.75 Ma.

# From transient landscape to topographic steady state

By comparing field observations between the studied catchments we can gain new insights into the processes and timescales by which transient landscapes evolve towards topographic steady state. We use the Rio Torto as an exemplar to quantify the propagation of topographic steady state and by comparison with the fluvial geometries evolved in the Fosso Tascino and Valleluce river ('Geomorphic transition to topographic steady state; response timescales') we draw some generic conclusions as to mechanisms by which topographic steady state is achieved within the landscape.

Figure 16a explicitly compares the tectonic uplift field on the Fiamignano footwall with the distribution of stream power, while Fig. 16b shows the current river long profile (labelled 1). We calibrated the uplift rate values to the unit stream power using the Valleluce river, Cassino (case B), where topographic steady state is achieved with

 $\sim 0.3$  mm year  $^{-1} = 5$  kW m $^{-2}$ . We used these values to infer that  $\sim 15 \,\mathrm{kWm^{-2}}$  equates to  $\sim 1 \,\mathrm{mm\,year^{-1}}$ , which is the uplift rate seen by the Rio Torto. We also allow for the  $1.5 \times$  increase in coarse sediment calibre (dotted line) immediately upstream of the fault to give a peak of  $\sim$ 25 kWm<sup>-2</sup> and we assume an incision rate at the Fiamignano fault of 1mm year<sup>-1</sup> as there is no scarp preserved in the channel at this point. The stippled zone between these lines in Fig. 16a indicates the stream powers that we infer from this calibration to be required to achieve topographic steady state in the gorge. These peak values coincide with the position of the gorge near the fault (zone D, Fig. 16a and b), where the channel has steepened and narrowed to match the increased rate of slip on the fault. According to this calibration, we predict therefore that in zone (C), between 6.5 and 8 km upstream, erosion rates are a little less than relative uplift rates although the river is beginning to adjust to the acceleration signal. Contrastingly, in zone (B) unit stream powers are considerably lower than at equivalent points in the Valleluce channel – only 300-1500 Wm<sup>-2</sup>. This strongly suggests that erosion rates are not sufficient to balance uplift in this portion of the channel, and indicates that the channel elevation is actually increasing here. The top of the catchment [zone (A)] is being back tilted in the hanging wall tip of the Sella di Corno fault, so it is likely that the uppermost part of the river is being actively downthrown, especially considering the lack of aggradation observed in the upper catchment (Figs 7b and 16a).

In Fig. 16c, we normalise the tectonic uplift rate (U) by the incision rate (E) using the calibrated values outlined above. Near the fault, estimates of the ratio (U/E) must lie near 1 i.e. topographic steady state, consistent with there being no scarp preserved in the channel. At distances  $< 8 \,\mathrm{km}$  from the channel head, U/E values rise, peaking at  $\sim 5 \text{ km}$  upstream of the fault with values of  $U|E \sim 5$ . The maximum U/E value lies just upstream of the slope break in river long profile (labelled 1 in Fig. 16b) where channel gradients are very low as a result of the tectonic back-tilt, but the uplift rate is relatively high. This is the locality where the channel is most vulnerable to defeat, i.e. where U|E is a maximum. In the Fiamignano case this danger is enhanced as U/E falls to negative values upstream, because the upper catchment of the Rio Torto is being actively down-thrown into the Sella Di Corno fault.

#### Propagation of topographic steady state

In 'Case C – Rio Torto (Fiamignano fault)', we interpreted the disparity between stream power and uplift pattern as a transient response initiated in response to fault acceleration. The transient response is characterised by a wave of incision that migrates upstream over time (cf. Tucker & Whipple, 2002; Whipple & Tucker 2002). Given that zones (A) and (B), above the break in slope in the long profile, are characterised by lower channel gradients, wide valleys and low stream powers, it is reasonable to conclude that they have not yet felt the effects of this incisional wave. This in-





Fig. 16. (a) Average unit stream powers in the Rio Torto, scaled to uplift. For the Valleluce river, Cassino,  $\sim 5000 \,\mathrm{Wm^{-2}}$  is needed at the fault to match the  $0.3 \text{ mm year}^{-1}$  uplift on the back-titled footwall. If we assume incision rate is linearly proportional to stream power, and hence uplift rate, and scale this to Fiamignano, we obtain 1 mm year  $^{-1} = 15$  kWm  $^{-2}$  at the fault. However coarse sediment input from landsliding also increases  $D_{84}$  by a factor of 1.5 in the lower gorge so greater energy expenditure is required to overcome enhanced channel roughness (Wohl, 2004). We scale stream powers by 1.5 to take account of this (dotted line). The area between these lines indicate stream powers which are likely required to achieve topographic steady state and these predictions agree well with average unit stream powers of 15- $30 \,\mathrm{kWm^{-2}}$  which are developed in the lower gorge. (b) Schematic diagram of the Fiamignano and Sella di Corno faults showing present day river profile (black line, 1), predicted profile 0.2 Myr into the future (dashed line, 2) and potential long profile after loss of headwaters (dots, 3) (c) Uplift rate/stream power-driven erosion rate for the present day (black line) and predictions for 0.2 Myr (dashed line) and 0.3 Myr (dotted line) in the future.

terpretation is also consistent with recent modelling results by Cowie *et al.* (2006). Our field observations enable us to address the following question: how long will it take for the headwaters to detect the effects of the increase in uplift rate? The wavelength of the stream power spike is 2.5 km (Fig. 16a), indicating topographic steady state has propagated this distance upstream (Fig. 16c), but the break in slope in channel gradient on the present-day long profile is  $\sim$ 4.5 km back from the fault (Fig. 16b, profile 1) so the geomorphic expression of fault acceleration propagates  $\sim 1.5 \times$  faster than the zone of steady state, assuming an initiation age of 0.75 Ma. If the top of the convex reach represents the total distance travelled by the incisional wave, this gives a rate of propagation of  $\sim$ 6 mm year <sup>-1</sup> upstream (4.5 km/0.75 Ma). At this rate the wave will take an additional 1 Myr to travel the remaining 6 km to the catchment headwaters, assuming constant velocity [a minimum time estimate as incision wave velocity is a function of drainage area (Tucker & Whipple, 2002)]. As full topographic steady state is achieved  $1.5 \times$  more slowly than the first geomorphic expression of fault acceleration the total response time would be  $\sim$ 2.25 Ma. However, as the elevation difference over the first 3.5 km downstream in the Rio Torto is only 80 m, and the upper catchment is actively back-tilting (see uplift field in Figs 13c and 16a), the headwaters are much more likely to be beheaded before the wave of incision arrives (due to back-tilting on the fault forming an interior drainage), and we calculate that this could take place in 200-300 vr (see Appendix A for derivation). The river is beheaded where U/Epeaks, just upstream of the break in slope (profile 2, Fig. 16b and c) at  $\sim$ 4 km downstream. The result is a foreshortened profile (e.g. shown schematically in profile 3, Fig. 16b) which will then decay to topographic steady state within a further 100 yr. This serves as a field demonstration of the fact that rivers in this tectonic setting, whose erosion processes lie towards the detachment limited end of the spectrum, are vulnerable to loss of the upper part of the catchment during a transient response unless the ate of propagation of the migratory wave upstream is rapid, as proposed by recent modelling work (Cowie et al., 2006).

#### Geomorphic transition to topographic steady state; response timescales

The development of topographic steady state (e.g. Valleluce river) from transient conditions (e.g. Rio Torto) involves a range of geomorphic adjustments, which do not necessarily have the same response timescale. In the Rio Torto, channel slopes have steepened in response to fault throwrate increase, and this steepening has propagated upstream, which in turn has led to reduced channel widths, reduced valley widths and low W<sub>b</sub>/H aspect ratios. The analysis above, and field observations in this paper give fundamental insights into how these perturbations evolve towards topographic steady state. Firstly, loss of the upper headwaters, such as has been witnessed in the Rio Torto, can shorten detachment-limited channels significantly [by  $\sim 40\%$  at the rate of knickpoint migration documented here (Appendix A)], eliminating the major convexities in long profiles, and allowing a more typical concavity to be regained relatively quickly. This process acts to limit response timescales for rivers in this tectonic setting so that it could occur within 0.4 Myr, giving a total response time of this process to slip-rate acceleration of 1.1-1.2 Myr. Secondly, as the Valleluce river has typical downstream width scaling, although incising across a constant slip-rate fault initiated at 3 Ma, then hydraulic geometry must also recover over this period: this process is aided substantially by loss of the headwaters, because catchment drainage areas are reduced, so that channel widths at the fault are no longer substantially lower than predicted by Eqn. (1). Channel widths in the new headwaters will narrow as the upstream drainage area is now low. The response timescale for this process is therefore between 1.2 and 3 Myr (age of fault inception). Finally, channel steepening and narrowing in response to fault slip-rate increase is also followed by decreased valley widths, which allows incision to be focussed into a narrow zone in the proximal footwall (Pazzaglia et al., 1998; Whittaker et al., 2007). For the simple example of block uplift (Fosso Tascino, case A), valley widths do appear to have relaxed, giving a maximum response timescale of 3 Myr (age of fault inception in this area). However, in the Valleluce river (tilted block, case B) these reduced valley widths near the fault are retained as part of the steady-state landscape, and help the river to balance the higher rate of uplift in the proximal footwall (Fig. 13b). Consequently, altered valley geometry can persist for several million years following a transient response to tectonics, as an alternative to significant long-profile concavity or channel width variations, when the tectonic uplift field has a non-uniform spatial distribution.

## CONCLUSIONS: IDENTIFYING TRANSIENT RESPONSES IN LANDSCAPE

The data presented above enable us to characterise for the first time the response of channels to tectonic forcing where the boundary conditions are known explicitly. By considering the hydraulic geometry, grain-size and uplift history we show that rivers near the detachment limited erosional end-member, and crossing active faults in the central Apennines of Italy have reached three different configurations that reflect differences in the space-time pattern of relative uplift:

(a) Equilibrium energy expenditure and topographic steady state for a channel incising an uplifting horst, and crossing a normal fault that has been slipping at a constant rate since 3 Ma (Fig. 17a).

(b) Topographic steady state but uneven downstream energy expenditure for a river crossing a back-tilting normal fault, with a constant slip rate since 3 Ma (Fig. 17b).

(c) A transient form where the river is neither in energy equilibrium nor topographic steady state, caused by fault acceleration after a linkage event at 0.75 Ma (Fig. 17c).

The three channels, shown schematically in Fig. 17, are characterised by disparate geomorphic signatures, and their form cannot be explained by appeal to differing lithology, erosion process, or hangingwall incision/aggradation. We are able to identify new diagnostic features of



Fig. 17. Schematic diagram showing landscape evolved during (a) energy equilibrium and topographic steady state on a horst block (e.g. Leonessa and Rieti faults (b) topographic steady state on a single footwall block (e.g. South Cassino fault) and (c) a transient response to fault acceleration (e.g. Fiamignano).

the transient fluvial response in addition to the oft-cited development of long-profile convexities, which do not correlate with changes in rock mass strength. In particular, rivers undergoing a transient response to fault acceleration display channel steepening and gorge formation near the fault, a breakdown in hydraulic scaling, and a reduced variability in channel planform over small length-scales which peaks near the fault. Additionally, we document a strong coupling of channel form to valley sides, which is linked to the input of coarse grain-sizes directly to the channel, and a strong non-linear dependence of channel aspect ratio on slope. The data indicate that the response timescale to fault acceleration is  $\sim$ 1 Myr to re-equilibrate local channel slopes, and <3 Myr to attain good hydraulic scaling. Transient conditions can thus persist for long periods in the landscape. Moreover, we show that a major risk for systems approaching the detachment limited endmember, and perturbed by normal-fault acceleration, is

Table 2. Summary tablbrium, (b) in topograph:	e outlining the differing characteristics and geometries evo ic steady state and (c) undergoing a transient response to tec	olved for rivers crossing active normal faults which are (a) in t ctonics	topographic steady state and hydraulic energy equili-
Feature	(a) Bedrock channel in topographic steady state and energy equilibrium	(b) Bedrock channel in topographic steady state	(c) Bedrock channel under-going a transient response to increased uplift rates
Definition	Uplift rate = erosion rate at all points downstream; even expenditure of energy downstream	Uplift = erosion rate at all points downstream, but energy expenditure not necessarily constant downstream	Uplift does not equal erosion rate, unequal energy expenditure
Long profile	Concave-up; $\theta = 0.5$	Concave-up $\theta$ potentially < 0.5 if uplift rate increases downstream, $\theta$ > if uplift rate decreases downstream	Large convexities present
Channel width	Scales with drainage area/discharge; $b = 0.5$	Scales with drainage area; $b = 0.5$	Width decoupled from drainage area/discharge; narrows in high slope zones near the fault
Channel slope	Decreases downstream: $S \sim A^{-0.5}$	Decreases downstream, but local channel slopes may be higher in zone increased uplift rates	Increases towards area of active uplift (e.g. fault)
Valley width	Uncorrelated with slope; tendency to increase with downstream distance	Narrows in areas of active uplift; $W_{\rm b}/W_{\rm v}$ weakly correlated with slope	Narrows in area of uplift; $W_{ m b}/W_{ m v}$ positively correlated with slope
Grain size	Constant; or declines downstream	Constant or declines downstream	Coupled to hillslope input $-D_{84}$ increases in zone of maximum incision
Aspect ratio	Slopes are generally low, so little variation in aspect ratio noted: can be thought of as constant	Slight dependency on local channel slope	Strong, non-linear dependence on slope: $W_{ m b}/M_{ m b}/M_{ m b}$
Hydraulic scaling	Good	Generally good	Poor
Unit stream power	Constant downstream	Increases downstream on same wavelength as uplift field	Wavelength of stream power response does not match uplift field
Channel morphology	Wide open valleys; partly alluviated or alluviated rivers	Valley widths narrow towards the fault	Presence of highly incised gorges; landslides directly feed channel, steep hillslope angles, Incision directly into bedrock

#### Transient response of rivers crossing active normal faults

that they are vulnerable to the loss of their headwaters if they are back-tilted before the over-steepened reach propagates upstream to the headwaters. This acts as a significant negative feedback on the response time of the fluvial network by physically shortening the active channel. For rivers crossing active faults which appear to have reached steady state, we also show that narrowing of valley widths in zones of higher uplift rate is a key way in which rivers maintain topographic steady state, even for those which exhibit good hydraulic scaling.

The characteristics identified in this study have important implications for anyone seeking to understand the transient response of channels to tectonics. These data challenge the current generation of fluvial algorithms in landscape evolution models by demonstrating that steady-state assumptions of hydraulic scaling and constant aspect ratio cannot be used if we are to successfully model channel response to transient conditions, because narrowing in response to tectonically driven steepening is an intrinsic way that channels adjust to changing boundary conditions. Moreover, we show that calculated scaling exponents from log–log plots, even when derived from field surveys, are likely to be misleading and local slope is shown to be as important a predictor of channel width as drainage area.

Because the three scenarios shown in Fig. 17 do differ significantly in terms of their geomorphic signatures, this study also provides key field criteria for workers attempting to identify transient signals in landscapes where the tectonic regime is less well constrained and we summarise these key differences in Table 2. Consequently, this study provides an important step towards being able to quantify tectonic forcing from landscape response, and while this goal remains an outstanding challenge facing workers in the field of fluvial geomorphology, we stress the value of detailed field data in achieving this aim.

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## **APPENDIX A**

Below we outline a simple numerical calculation to assess the time and position at which the Rio Torto is likely to be defeated by uplift on the fault. The change in long profile over time in the upper part of the catchment, in response to the tectonic setting can be expressed as:

$$Z_{f(L)}^{t} = Z_{f(L)}^{t=0} + (U_{f(L)} - E_{f(L)})t$$
(A.1)

where  $Z_{f(L)}^{t=0}$  is the current long profile,  $U_{f(L)}$  is the distribution of uplift rates as a function of downstream length,

 $E_{f(L)}$  is the distribution of erosion as a function of downstream distance, t is the time period considered, and  $Z_{f(L)}^{t}$  is the long profile at that time. Clearly, at any point in time,  $t_x$  where the following condition applies,

$$\frac{\mathrm{d}Z_{f(L)}^{\prime_{\mathrm{x}}}}{\mathrm{d}L} = 0 \tag{A.2}$$

the channel starts to be defeated and begins to form an internally drained basin. This equation can be solved to find the downstream length,  $L_x$ , at which the defeat occurs. However, the over-steepened reach is also migrating upstream, from its present position downstream at 6 km. Therefore the position of the break in slope,  $L_k$  at any time,  $t_k$ , is given by:

$$L_{\rm k} = 6000 - V t_{\rm k} \tag{A.3}$$

where V is the migration rate of the 'knickpoint' upstream, which we estimate in this case to be  $6 \text{ mm year}^{-1}$  ('From transient landscape to topographic steady state'). For simplicity we keep V constant; in reality the migration rate of the over-steepened reach will decline as the upstream drainage area falls, so these calculations are conservative estimates. We also consider that for cases where  $L > L_k$ the river is capable of adjusting so as to keep pace with fault uplift. Consequently, the question is whether there is a solution of Eqn A.2 for  $L_x < L_k$  and  $t_x = t_k$  We can solve Eqn A.1 using numerical iteration from our DEM extracted long profile and estimated uplift function on the Fiamignano fault, shown in Fig. 13. For each time step, we test whether Eqn. (A.2) is satisfied. First, we assess the simple case where erosion in the upper part of the catchment can be neglected, as shown in Fig. A.l. In this instance that the river starts to be defeated at 4 km downstream in only 100 kyr, in which time the migrating wave of incision, as indicated by the top of the convex reach, has only travelled 600 m upstream (i.e.  $L_k = 5.4$  km). By 300 kyr, the upper catchment elevation gradient has disappeared forming a



**Fig. A.1.** Long profile evolution of the upper catchment of the Rio Torto, neglecting fluvial erosion. t = 0 is the present day long profile, and graph shows the calculated profile for 100 ka time-steps. The star represents the position of the top of the oversteepened reach at each time, taken to be the upstream extent of the effect of the migrating wave of incision.

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**Fig. A.2.** Long profile evolution of the upper catchment of the Rio Torto, including fluvial erosion scaled to current unit stream power values. t = 0 is the present day long profile, and we show predicted profiles for 200 and 365 ka into the future. The star represents the position of the top of the over-steepened reach at each time, taken to be the upstream extent of the effect of the migrating wave of incision.

substantial internal basin, and the top of the over-steepened reach is still at a distance of 4.3 km downstream, confirming that the upper catchment is likely to be defeated.

Of course, the above calculation does not include fluvial erosion; however we can include this by using erosion rates, scaled to unit stream powers, for the upper part of the catchment: We calculate a downstream increase from  $\sim 500 \text{ Wm}^{-2}$  to  $2000 \text{ Wm}^{-2}$  between 1.5 and 6 km downstream (Fig. 13) and if we use the calibration of  $15 \text{ kWm}^{-2} = 1 \text{ mm year}^{-1}$ , ('From transient landscape to topographic steady state') this would give an erosion rate

increase from 0.03 to 0.12 mm year<sup>-1</sup> over this distance. We make the simplifying assumption that the distribution of stream power does not change through time, upstream of the break in slope; in reality erosion rates will decline as the catchment is back-tilted and the river gradient lowered, so the results below are a maximum estimate for the response time. In this instance, (Fig. A.2) we predict the formation of a small internally drained basin within 200 kyrs, bounded by a lip at 4.1 km downstream, by which time  $L_{\underline{k}} \leq 4.8$  km. By 365 kyr, the elevation gradient of the upper catchment is already lost before the top of the oversteepened reach arrives, producing a fore-shortened long profile.

These simple calculations demonstrate that the headwaters of the Rio Torto are very likely to be defeated in the time period of 200–300 kyr and in this case the top of the over-steepened reach would be expected to arrive in the new headwaters in <400 kyr, giving a total response time to the slip rate increase of ~1.1 Myr. We note that the migration rate of the over-steepened reach would have to be approximately twice as fast (~12 mm year<sup>-1</sup>) to ensure the survival of the headwaters.

These results underline the propensity for detachment limited systems to be beheaded unless knickpoint migration rates are rapid as argued by Cowie *et al.* (2006). Moreover, we note that the top of the Vallone Stretta, the main tributary to the Rio Torto, does indeed have a small internally drained basin (the Rascino plain) sitting just beyond the present headwaters of the channel (Z on Fig. 3c); This plain is separated from the current channel by a lip of just 10 m and we interpret this to represent the old headwaters, which have now been defeated, presumably because the fluvial erosion rate on the tributary was insufficient to keep pace with down-throw on the Sella di Corno fault.