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The use of fluvial archives in reconstructing landscape evolution: the value of sedimentary and morphostratigraphical evidence

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Abstract

Evidence-based interpretations of fluvial evolution, and especially of river-terrace formation, have advanced significantly in recent decades, with a notable contribution made by activities of the Fluvial Archives Group. Well-dated river-terrace sequences provide frameworks for the understanding of landscape evolution, since they record valley-floor levels that were higher in the past, attributable, from their patterns of occurrence, to regional uplift. The role of climate fluctuation during the Quaternary is also paramount, since this has been an important driver of the varied fluvial activity that has given rise to the staircases of terraces that characterise the temperate latitudes. This approach is contrasted with a more theoretical methodology for using rivers as recorders of landscape evolution, again with an emphasis on uplift, based on the concept of the formation of knickpoints at particular base levels and their migration upstream. Although different timescales can be explored by the two methods, the concept of headward-migrating knickpoints implies a mechanism for incision that is difficult to reconcile with the formation of the broadly parallel river terraces that are observed in many systems. Knickpoints can frequently be observed to coincide with gorge reaches, where river valleys are constricted as a result of resistant bedrock and/or the effects of localised active crustal deformation. This raises the possibility that knickpoints have generally formed in response to factors of local geology rather than migrating from downstream.

Keywords: climatic fluctuation, knickpoint recession, Quaternary, river terraces, stream power law, uplift

Introduction

Amongst the research interests of Jef Vandenberghe (in whose festschrift this article appears) has been identification, from fluvial sedimentary archives, of the main influences on the activity of rivers at various Quaternary timescales. Vandenberghe has, indeed, placed considerable emphasis on the observation that controls on fluvial activity are different at different timescales. This discussion paper concerns the evidence for landscape evolution over Quaternary and longer timescales that can be derived from records of fluvial activity and, in particular, of fluvial incision (down-cutting), as recorded by river terrace sequences. Central to understanding such records is the role of climate change as a control on river terrace formation at Milankovitch (glacial–interglacial) timescales. This is a topic on which Jef has contributed on multiple occasions (Vandenberghe, 1993, 1995, 2002, 2003, 2008; Vandenberghe et al., 1994).

The development of theories concerning river terrace formation

The role of climate

The thesis that glacial-interglacial climatic fluctuation has been important in the generation of river terrace sequences dates back to before the true complexity of such fluctuation during the Quaternary was realised. Thus by the mid 20th Century a connection had been suggested between the different fluvial activities recorded in terrace records (particularly

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alternations between vertical incision and sedimentation) and climatic change (e.g. Zeuner, 1945; Bourdier, 1968, 1974; Wymer, 1968; Tyràček, 1983). Advocates of such ideas envisaged an influence of climate on fluvial processes that was independent of base-level change, although the latter is of course closely related to climatic fluctuation through the eustatic control of global sea level (cf. McCave, 1969; Törnqvist, 1998; Karner & Marra, 1998; Blum & Straffin, 2001). With the evidence for more numerous Quaternary climatic cycles that came from the analysis of deep oceanic sediments, the potential causative relationship between glacial-interglacial cyclicity and river terrace formation gained further adherents, amongst whom Jef Vandenberghe (1993, 1995, 2002) was prominent, along with (e.g.) Bull & Kneupfer (1987), Green & McGregor (1987), Bull (1991), Antoine (1994), Bridgland (1994, 2000, 2010), Maddy et al. (2000, 2001a), Starkel (2003) and Cordier et al., 2006, 2012). An important line of reasoning in disassociating base-level influences from the formation of aggradational river terraces was the recognition that the ubiquitous gravel deposits that characterise such fluvial sequences, even those in downstream reaches near to coasts, have generally been laid down during periods of cold climate (e.g. Rose & Allen, 1977; Green & McGregor, 1980; Gibbard, 1985; Vandenberghe, 1995, 2002; Macklin et al., 2002); at such times base levels would have been low and rivers should, if base-level were the main forcing factor, have been incising rather than in depositional mode.

Also of importance was the realisation that fluvial activity has differed greatly between periods of different climate. Thus during cold episodes rivers have typically adopted braided regimes, if only because sediment supply was plentiful, with minimal vegetation to stabilise valley sides and adjacent land surfaces (Vandenberghe, 1995, 2003; Maddy et al., 2001a). From sedimentological evidence it was realised that the major part of the gravel terrace deposits that characterise river valleys throughout the temperate latitudes had indeed been deposited on braid plains during cold episodes. During warmer periods rivers have tended to develop single-thread channels, often meandering, and the fact that most incision takes place under a single channel regime perhaps explains the earlier view that incision had taken place during interglacials (Zeuner, 1945; cf. Vandenberghe, 2002). This overly simplistic view was superseded by the recognition (Vandenberghe, 1995) that down-cutting has typically occurred during periods of climatic transition. Indeed, Vandenberghe (1995) was amongst the first to recognise that the primary geomorphological effects of river activity are best expressed at times of climatic change. Down-cutting at warm-cold climatic transitions seems to be the norm in mainland Europe, as noted by Vandenberghe (2008), although in Britain the preponderance of evidence is for incision at cold-warm transitions; this British record sometimes incorporates narrow deep channels, including interglacial deposits within their infill, with more widespread gravel sheets above (e.g., Bridgland, 1988, 2006).

Not only has the 'rejuvenation' between one terrace level and the next tended to occur during transitional periods within the glacial-interglacial cycle (cf. Bridgland & Maddy, 1995; Bridgland, 2000); Jef Vandenberghe and co-workers have also pioneered an approach that charted fluctuations of fluvial regime during periods of climatic fluctuation at sub-Milankovitch scale, particularly within the last climate cycle, for which the sedimentary and geomorphological records are generally excellent (Vandenberghe et al., 1994; Kasse et al., 1995, 2010; Mol et al., 2000). Vandenberghe (1995, 2008) also recognised, however, potential danger in uncritical attribution of terrace sequences to climatic fluctuation, seeking to reconcile aspects of the thinking of process geomorphologists and Quaternary geologists in the development of a non-linear fluvial response model. This seems highly apposite, since the process geomorphologists (e.g. Schumm, 1977, 1979; Hey, 1979) had developed their understanding of fluvial dynamics, demonstrating that major changes can be generated intrinsically, generally without consideration of the effects of Quaternary climate change, whereas the promoters of 'climatic geomorphology', such as Büdel (1977, 1982), Tricart (1972) and Tricart & Cailleux (1972), had sometimes used subjective interpretations to reinforce their arguments. In seeking to incorporate the best of both, Vandenberghe (2008) developed a 'modified non-linearity model' of river development, in which he emphasised the different styles of fluvial activity related to the different 'senses' of climatic fluctuation: warm-cold and cold-warm. Thus cold-warm transitions would see single-channel meandering rivers making relatively deep incisions but of limited lateral extent, whereas at warm-cold transitions the rivers would switch to braiding mode and would therefore effect relatively shallow down-cutting over a wide area. Vandenberghe considered that this difference might explain the rarity in the European record of evidence for cold-warm incision events, since the laterally more extensive braided fluvial incision at warm-cold transitions would tend to obliterate the evidence for the previous phase of down-cutting. This combination of different styles of down-cutting at the alternating climatic transitions can also explain the British preservation pattern, described above, in which the later, broader but shallower valley erosion is superimposed above the earlier narrow incision (cf. Bridgland, 2006).

The role of uplift

Other workers have emphasised the importance as a driving factor in river terrace formation of uplift, not seen just in plate-boundary areas or active fault zones, but regional (epeirogenic) uplift that appears to have occurred in many parts of the world (Veldkamp & Van den Berg, 1993; Antoine, 1994; Van den Berg, 1994; Bridgland, 1994, 2000; Maddy, 1997; Antoine et al., 2000; Westaway et al., 2003, 2006, 2009; Bridgland & Westaway, 2008a, b). The effects of climate change and the associated fluvial responses are thus integrated with Quaternary landscape development, with uplift typically resulting in the preservation (as river terraces) of valley floors from a succession of stages of Quaternary time. Knowledge of this type of record has been expanded greatly under the auspices of FLAG (the Fluvial Archives Group), to which Jef has been a consistent contributor since its inception in 1996 (Vandenberghe & Maddy, 2001; Vandenberghe & Vanacker, 2008; Vandenberghe et al., 2010). This expansion is seen particularly in the output from IGCP projects 449 and 518 (Sinha & Tandon, 2003; Bridgland et al., 2004, 2007a; Westaway et al., 2009) as well as in other FLAG compilations (Maddy et al., 2001b; Bridgland & Sirocko, 2002; Pastre et al, 2004; Stokes et al., 2012).

It can be concluded, from these accumulated empirical data, that in most areas rivers have formed deeply incised valleys in response to widespread regional uplift. For those sequences with the most extensive records, and which have the best dating control, this uplift can be shown to have accelerated in response to post-Tertiary climatic cooling and then again as a response to the greater severity of the cold-climatic cycles following the Mid-Pleistocene Revolution (MPR; e.g., Mudelsee & Schulz, 1997). The correlation between climatic change and the rate of uplift suggests coupling between surface and deep-Earth processes, with rivers contributing to an isostatically (mass-balance) driven mechanism, their role being to erode and transport material from uplifting areas and deposit it in areas of sediment accumulation, which may well be subsiding under the cumulative load of that sediment. This is a description of erosional/sedimentary isostasy, although some mechanism of positive feedback would be required for it to provide a satisfactory explanation of the scale of uplift that is observed in the IGCP data; indeed, flow of mobile lower-crustal material from subsiding to uplifting areas has been suggested as a mechanism for this feedback effect (Westaway, 2002a, b, c, 2007; Bridgland & Westaway, 2008a).

The importance of the FLAG/IGCP data is that it includes numerous examples of well-dated terrace sequences that can be used to constrain the progress of valley incision. An important finding is that comparable records of uplift are found throughout the world, wherever there is dynamic (post-Archaean) crust that is not loaded by sediment accumulation. The uplift is best recorded in the temperate regions, where glacial-interglacial climatic fluctuation has led to the formation of well-developed river terrace sequences (except where rivers are constricted in gorge reaches formed in resistant bedrock: see below). Similar crust in the tropics has also uplifted but, without the pronounced climatic fluctuation as a catalyst for fluvial activity, the terrace record is sparser (Bridgland & Westaway, 2008b; cf. Büdel, 1977, 1982). In the case of Archaean cratons and Early Proterozoic crustal provinces there has been no such progressive uplift, as is clearly indicated by the fluvial archives from rivers such as the Vaal (Helgren, 1978; Westaway et al., 2003) and the Dnieper (Matoshko et al., 2004; Bridgland & Westaway, 2008a). The uplift is independent of plate-tectonic processes, since it occurs in the interiors of plates and is too widespread to be explained by the long-distance transfer of compressional stresses from plate boundaries (contra Cloetingh, 1988; Japsen et al., 2010; Cloetingh & Burov, 2011; Pedoja et al., 2011), given that not all such boundaries involve crustal shortening. Indeed, important phases of accelerated uplift, such as that at ~0.8 Ma (i.e., following the MPR), do not correspond with known changes in plate motions.

An alternative methodology: the 'stream power law' approach

At the same time as contributors to FLAG were compiling the data discussed above, an extensive parallel literature has developed in which process-based explanations of the development of fluvial systems have been promoted (e.g., Rosenbloom & Anderson, 1994; Whipple & Tucker, 1999, 2002; Snyder et al., 2000; Kirby & Whipple, 2001; Lague & Davy, 2003; Crosby & Whipple, 2006; Roberts & White 2010). In this 'stream power law' approach, rivers are typically parameterised in terms of scaling relationships, such as the common assumption that discharge, Q, is proportional to downstream distance along a river, x, raised to some power m; the resulting equations are then solved to predict the development of rivers over time. This methodology generally includes no consideration of climate or, indeed, the Quaternary context of modern fluvial systems (i.e., that these systems have experienced a succession of climatic fluctuations over the past ~2 Ma). As a result, interpretations arising from this alternative approach are often entirely at odds with the empirical evidence such as has been researched by FLAG contributors.

A recent development has been the suggestion that uplift histories can be determined from the longitudinal (gradient) profiles of rivers traversing the region in question (e.g., Pritchard et al., 2009; Roberts & White, 2010; Hartley et al., 2011). This method has been developed on the basis of previous analyses of relations between geomorphology and fluvial incision, such as those by Whipple & Tucker (1999, 2002), Snyder et al. (2000), Kirby & Whipple (2001), and Lague & Davy (2003). Such analyses are based ultimately on the accumulation of empirical data pertaining to correlations between longitudinal gradients of rivers and other parameters compiled by Hack (1957). Fundamental to the method is the expectation that rivers have smoothly varying longitudinal gradients, reflecting the progressive downstream increase in discharge (since, for any particular river, the further downstream a reach is situated the larger will be the catchment area upstream). In this approach, localised increases in the downstream gradients of rivers, or 'knickpoints', are attributed to past changes in uplift rates, the idea being that these knickpoints originated at the downstream end of each river and propagated upstream. According

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to such theory, the distance of a knickpoint from the mouth of a river thus determines how long ago the corresponding increase in uplift rate took place (i.e., knickpoints are of increasing age with distance upstream). In addition, the detailed form of the longitudinal profile around the knickpoint is thought to indicate how much uplift was involved and the precise period of time during which this uplift occurred. It has thus been argued (Bishop, 2007) that the presence or absence of knickpoints along rivers can be used as a test for whether the landscape through which they flow is in steady state, with (rock) uplift in balance with erosion; the occurrence of knickpoints is thus seen by Bishop as an indication that a landscape is not in steady state. It should be noted that in this context a 'steady-state' landscape means a landscape in which the uplift rate balances the spatially-averaged erosion rate, not one in which uplift balances valley incision, as was suggested by Lewin & Gibbard (2009). Indeed, empirical evidence shows that rates of valley incision typically exceed spatially-averaged erosion, because interfluves are being eroded less rapidly than valley floors. In contrast, software systems that model the isostatic response to the non-steady-state conditions that have developed in response to Quaternary climate change (e.g., Westaway, 2002c, 2007) predict that uplift rates exceed spatially-averaged erosion rates. This prediction is consistent with the widely-used assumption that rates of valley incision, as observed from river terraces, have been equivalent to rates of Quaternary uplift (e.g., Maddy, 1997; Bridgland, 2000), which is thus indicative of non-steady-state landscape development during the Quaternary.

The 'stream power law' methodology harks back to a form of 'denudation chronology' that was a popular theme for research in the mid-20th Century and which had its original derivation in the classic geomorphological concept of the Davisian cycles of erosion (Davis, 1899). In its mid-20th Century version this approach involved the recognition of terraces and knickpoints as integral parts of fluvial archives. An influencial proponent was Zeuner (1945, 1959, 1961), who believed that the major fluvial down-cutting events (rejuvenations) occurred in response to falls in sea level, with propagation of the valley deepening upstream by means of knick-point migration. Accordingly terraces were reconstructed that converged upstream with the valley floor, grading into it immediately above a knickpoint, upstream from which the valley floor and the terrace coincided (see Fig. 1). The implication, seemingly, was that successively older knickpoints had propagated progressively further upstream. Examples of this approach include the reconstruction of terraces in the River Great Stour in Kent, SE England (Wooldridge & Kirkaldy, 1936) and in the Derwent, a left-bank tributary of the River Trent that drains the Peak District (Waters & Johnson, 1958; Fig. 1), as well as Zeuner's (1961) interpretation of the Thames. However, none of these schemes is compatible with more recent terrace mapping, such as by the British Geological Survey (which has invariably identified terraces that are broadly parallel with the valley floor), and they have generally been ignored or superseded (see Fig. 1).

Discussion

The effects of fluctuating climate

It has already been noted that the fluctuating climate of the Quaternary has been regarded as a significant influence on fluvial evolution by those seeking to explain terrace sequences from sedimentary archives, although advocates of the 'stream power law' methodology have largely ignored palaeo-climatic effects (see previous section). Nonetheless it can be argued that the continued climatic fluctuation would have been a considerable impediment for landscapes being in steady state during this most recent geological period, as has been demonstrated observationally from case studies in which rates of (rock) uplift have exceeded rates of erosion (e.g., Bridgland & Westaway, 2008a; Westaway et al., 2009). Such behaviour can be predicted by numerical models for the behaviour of continental crust in response to climate-driven changes in erosion rates (e.g., Westaway, 2002c) and this is so regardless of whether rivers in the studied regions have knickpoints or not (cf. Bishop, 2007; see previous section). Thus in regions where landscapes are demonstrably not in steady state there are many rivers that lack knickpoints. This suggests that rivers are generally capable of maintaining quasi-equilibrium longitudinal profiles in many settings within non-steady-state landscapes, although knickpoints can develop in situations where there is perturbation as a result of lithological constriction or active faulting, as in examples to be discussed later in this paper. The presence or absence of knickpoints is perhaps related to factors dictating channel-incision rates, which have given rise to a bipartite classification of modern river channels as 'detachment-limited' or 'transport-limited'. In the latter the capacity to incise is limited by the ability of the river to erode the bedrock, whereas in the former it is limited by ability to transport sediment derived from upstream and from the incision process (e.g., Howard, 1980); in a typical river the former situation will pass downsteam into the latter (Whipple & Tucker, 2002; Brocard & Van der Beek, 2006), reflective of the key influence of gradient. Although this distinction can be readily applied to present-day rivers, its value in reconstructing past landscape development is uncertain, primarily because of the marked changes to fluvial regimes that that have resulted from climate change. Indeed, empirical evidence from the study of terrace sediments suggests that climate change has resulted in phases of incision throughout the length of a river, interspersed with phases of fluvial aggradation (e.g., Veldkamp & Van den Berg, 1993; Antoine, 1994; Bridgland, 1994, 2000; Maddy, 1997; Antoine et al., 2000, 2007; Bridgland & Westaway, 2008b).

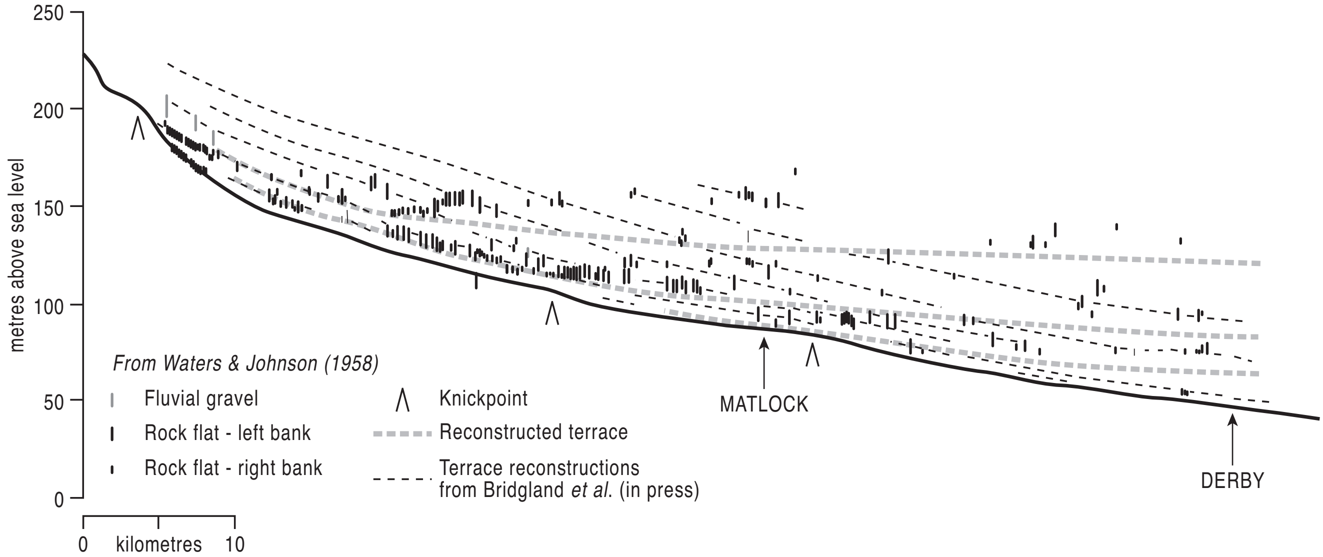

Fig. 1. Terraces of the River Derwent, according to Waters & Johnson (1958). Their scheme of largely erosional (bedrock) terraces envisaged subhorizontal terraces with gentle downstream gradients, converging in turn (upstream) with the much steeper modern valley floor. Their scheme thus implied that each point of convergence was a knick point that had propagated upstream only to that particular locality and that there had been no net valley deepening upstream of each convergence point since the time when the associated terrace deposits were emplaced. This scheme is impossible to reconcile with more recent mapping of the Trent terrace system (e.g., Brandon & Sumbler, 1988, 1991) and had never been incorporated in more recent work. An alternative scheme for the Derwent, recently proposed as part of a reappraisal of the Trent system (Bridgland et al., in press), is overprinted; it envisages terraces that are approximately parallel with the modern valley floor.

Whipple & Tucker (2002) discussed the potential effects of climate change, asserting that the upstream extent of a river that is detachment-limited might vary with respect to climate but otherwise giving the impression that climate change would not be expected to affect fluvial regimes greatly. Roberts & White (2010) noted that the effects of climate change 'might be important' for the development of rivers but stated that it was unclear to them how such effects might be parameterised for inclusion in their 'stream power law' calculations. They also argued that the average discharge by rivers governs the incision of their valleys over long timescales, presumably by a gradualist 'uniformitarian' extrapolation of modern rates of activity. This contrasts with the view espoused by Vandenberghe (1995), and supported by others (e.g., Bridgland & Maddy, 1995, 2002), that fluvial incision has been concentrated during occasional brief periods of extreme climatic conditions and/or climatic instability (such as transitions between cold and warm episodes during the Quaternary, when there might have been enhanced seasonal snowmelt), and, crucially, that the average behaviour of a given river at the present time does not represent these periods. Even for well-studied rivers in north-west Europe, estimation of the discharge at these times of peak palaeoflow remains in its infancy (e.g., Gupta et al., 2007; Toucanne et al., 2009; Westaway & Bridgland, 2010, 2011); Jef Vandenberghe has also contributed to this field (Busschers et al., 2011). In studies such as that by Roberts & White (2010) there is, in effect, an assumption that present-day fluvial discharge is representative not only of the Quaternary as a whole, with its multifarious fluctuation of climate, but also of timescales spanning tens of millions of years of pre-Quaternary time. In other words, there is an implicit assumption that the numerous climate changes that are known to have occurred during these intervals and have, indeed, been recognised as an overarching characteristic of the Quaternary, have had little effect on rivers. The work by Vandenberghe (1995, 2003, 2008) and others has clearly demonstrated the contrary to be the case: as noted above, during some parts of the climate cycle rivers have formed narrow, deep channels and during others they have formed broad floodplains with braided channel networks. The present-day valley floor level is thus unrepresentative of 'average' conditions.

Valley deepening in the absence of uplift

A minority of workers have disputed the concept of widespread regional uplift during the Quaternary, or have claimed that rates of fluvial incision are unrelated to uplift rates, the implication being that the flights of river terraces that occur across widespread regions distant from plate boundaries must have been formed as a response to climatic fluctuation alone (e.g., Hancock & Anderson, 2002; Gibbard & Lewin, 2009), with valley deepening occurring throughout the longitudinal profiles of rivers, or by the upstream propagation of climatically-controlled eustatic marine base level. If the latter had been the normal mechanism for valley deepening in a stable crustal setting, longitudinal reconstructions of river terraces should indicate downstream convergence, with longitudinal gradients inevitably become shallower with each phase of rejuvenation. Since it is known that global sea level has not lowered progressively during the Middle and Late Pleistocene, merely having fluctuated in response to global ice volume (Sidall et al., 2003; see above), valley deepening without rock uplift would have led to a declining downstream gradient, eventually becoming a constraint on the fluvial system. The fact that reconstructed river terraces are typically subparallel with one another is thus suggestive of progressive uplift, which is seen as a requirement for the formation of substantial terrace systems (cf. Westaway et al., 2002, 2009). Downstream convergence is seen in reconstructed terrace long profiles in unusual circumstances, namely where a river flows from an uplifting to a subsiding area, such as the Rhine in its course from Germany to the Netherlands (Brunnacker et al., 1982) and the Danube as it flows from the Buda Mountains to the Great Hungarian Plain (Gábris & Nador, 2007; Miklós & Neppel, 2010).

An important observation, and one often overlooked by process geomorphologists, is that offshore from the mouths of modern rivers there is generally a drowned extension of their valleys that was occupied during the last glacial, only ~15,000 years ago, when sea level was ~130 m lower than at present. If the crust both onshore and offshore has been stable in the interim period then any future decline in sea level would merely cause the rivers to extend onto the continental shelf and reoccupy these drowned valleys, perhaps involving the removal of any accumulated marine and estuarine sediment but requiring no formation of a knickpoint and providing no imperative for valley deepening (cf. D'Olier, 1975; Bridgland & D'Olier, 1995). Indeed, in coastal areas with wide continental shelves rivers (such as the Thames: Bridgland & D'Olier, 1995; Bridgland, 2002) can have terrace staircases, formed at times of low sea level, that continue offshore, implying that these offshore areas have also been uplifting. Only where the shelf is narrow might episodes of very low sea level, when this falls below the shelf edge, result in valley incision within the present onshore area (Bridgland, 2000, 2002).

Nonetheless, glacial low-sea-level episodes (lowstands) have been claimed to have resulted in incision; examples are provided by case studies of the Mississippi (Törnqvist & Blum, 1998) and the Rhine/Maas (Törnqvist, 1993). It has been noted, however, that incision in the Mississippi resulting from sea-level decline at lowstands is restricted to 'far downstream reaches', with little or no effect being registered in piedmont and upstream reaches (Blum et al., 2000; Blum & Straffin, 2001). This reflects the fact that, by the late 20th Century, process-based theory considered the upstream influence of sea-level fluctuation on fluvial incision to be limited (Leopold & Bull, 1979; Schumm, 1993). Observations in southern Britain and on the nearby coastal area of north-west Europe also suggest that large-scale fluctuation of sea level has had little effect on fluvial activity, except in the lowest reaches of valleys (Bridgland & D'Olier, 1995; Bridgland & Allen, 1996; Maddy et al., 2000; Lewis et al., 2004). This may be a direct consequence of the wide continental shelf around the UK, which means that lowstand coastlines were in areas now far offshore and many tens of kilometres from modern estuaries; for example, the lowstand Rhine-Thames-Seine river system had its mouth between the present locations of Cornwall and Brittany (cf. Maddy et al., 2000; Bridgland, 2002). On the other hand, the synthesis by Macklin et al. (2002) of a large body of dating evidence for phases of aggradation and incision in many Mediterranean rivers revealed no evidence of any effect of sea level; in the view of these authors, the observed phases of fluvial activity reflect short-timescale climate instabilities, some of which were demonstrably synchronous with Heinrich events (i.e., the brief periods of cold climate associated with disruption of the thermohaline circulation in the North Atlantic Ocean). The continental shelf is typically much narrower in the Mediterranean than offshore of north-west Europe, so a greater potential for preserving evidence for sea-level control would be expected.

Upstream propagation of knickpoints

The 'stream power law' methodology makes a theoretical prediction that, following a sudden fall in sea-level (or a sudden rise in the land relative to the sea surface), a knickpoint will develop in the lower reaches of a river and will subsequently propogate upstream, eventually reaching its headwaters. This idea was first proposed by Rosenbloom & Anderson (1994) and was subsequently developed by Whipple & Tucker (1999), Crosby & Whipple (2006) and others. This idea is now highly engrained in fluvial geomorphological literature; for example, Bishop (2007) reported it as though it were established fact, rather than a hypothetical prediction of theory. This theory has been conceived without recognition or accommodation of the instability of recent (i.e. Quaternary) sea level, which (as noted above) has fluctuated markedly in response to climatic fluctuation and its effect on global ice volume.

Another shortcoming of the 'stream power law' approach, as acknowledged previously (e.g., Westaway, 2004; Bishop, 2007), is that in most analyses the rates of upstream propagation of knickpoints is treated as a free parameter, the value of which is 'tuned' to fit observations, rather than being constrained independently. Indeed, a parallel literature has developed in which rates of upstream propagation of waterfalls have been measured observationally (e.g., Hayakawa & Matsukura, 2003; Hayakawa et al., 2008). Many of the studied waterfalls are in fact rapids, involving reaches of rivers with steep longitudinal gradients, rather than abrupt vertical drops. This literature is thus describing knickpoint recession and as such should be reconciled with the parallel 'stream power law' literature on the same topic.

Hayakawa & Matsukura (2003) derived an equation for the recession rate V of a waterfall, showing that this depends on the area of the waterfall face experiencing erosion, taken as W, the width of the channel, multiplied by H, the vertical drop:

$$V = k\,[(Q / (W \cdot H))\,\sqrt{(\rho / S)}]^n \qquad (1)$$

where Q is discharge, ρ is the density of water (1000 kg/m^3) and S the unconfined compressive strength of the bedrock. Hayakawa & Matsukura (2003) established this equation by measuring amounts of waterfall recession over known time-scales, the parameter S being determined in the field using a Schmidt hammer. Subsequent regression analysis determined values for the coefficients n and k of 0.73 and 99.7 m/a, respectively. In cases where values of S for local lithologies are unavailable, as for the examples discussed below, they can be estimated using standard values for particular lithologies from publications (e.g., Bienawski, 1974) or data books. In the case of a knickpoint, rather than a waterfall, the area experiencing erosion is W × H / s, where s is the longitudinal gradient of the river within the knickpoint. The equation obtained by Hayakawa & Matsukura (2003) thus becomes

$$V = k\,[(Q \cdot s / (W\,H))\,\sqrt{(\rho / S)}]^n \qquad (2)$$

This equation can be used to calculate rates of upstream propagation of knickpoints in a manner that is consistent with observational evidence, for comparison with rates that have simply been 'tuned' to fit observations.

As noted above, particularly influential for analysis and interpretation of knickpoint evolution has been the study by Crosby & Whipple (2006) of the ~10^5 km^2 catchment of the River Waipaoa in the North Island of New Zealand. They identified a widespread Late Pleistocene terrace within this catchment representing the starting-point for incision, throughout the main valley system, that they attributed to the lowering in global sea-level at the Last Glacial Maximum (LGM). They suggested that the resulting incision phase propagated upstream from the coastline by knickpoint regression, translating into numerous of the various dendritic affluents of this system and culminating in many of these a few km from their confluence with the main Waipaoa, a phenomenon attributed to lower stream power in these smaller rivers. Indeed, Crosby & Whipple (2006) documented a total of 236 knickpoints in the Waipaoa catchment, which were typically located 98±1% of the distance between the present-day coastline and the headwaters of each tributary. Of a subset of 161 subjected to detailed study, 78% were shown to be located in tributaries within 1 km of confluences with larger streams. Crosby & Whipple (2006) observed that the proportion of knickpoints located within erosion-resistant strata was high compared with a random distribution of knickpoints, although they thought this might indicate that knickpoint recession has slowed in these harder bedrock outcrops. It seems to the present authors equally likely that Crosby & Whipple (2006) have predominantly observed a 'hanging tributary' effect; tributaries tend to develop steep lower reaches as they grade to the level of larger parent streams that, with greater erosive power, have been able to incise their valleys more rapidly. In other parts of the world the lowering of sea level at the LGM merely served to extend rivers into the offshore area, as would a future fall in sea level in regions such as north-west Europe, where offshore submerged valley systems can be reoccupied with a minimal change in gradient (see above). Things might be different with the narrow continental shelf offshore from the Waipaoa, which (as Crosby & Whipple noted) is ~100 km from the Hikurangi subduction trench. However, their observation that terrace remnants occur within the incised reaches, something they are at pains to explain away, seem incompatible with the concept of upstream knickpoint migration as the principal mechanism for down-cutting and must raise doubts about the veracity of the entire hypothesis.

As Crosby & Whipple (2006, p. 35) acknowledged, "the very idea of knickpoint migration has been challenged and debated episodically for over 100 years". Although it is clear that knickpoints exist and that they can recede upstream for modest distances, such as by waterfall erosion (see above; cf. Hayakawa & Matsukura, 2003), the validity of the idea that they can migrate and maintain their height over long distances, rather than being removed as the gradient progressively evens out or the causative resistant strata are left behind, has yet to be convincingly demonstrated. In contrast, the increasing wealth of sedimentary evidence documenting the occupation by rivers of well-dated higher-level floodplains (now terraces) that run parallel with modern valley floors is strongly suggestive of an alternative mechanism in which down-cutting has generally been achieved by erosion throughout (at least the middle and lower reaches of) fluvial courses.

An attempt to test the results of the Crosby & Whipple (2006) analysis of the Waipaoa River (see above) against the predictions by Hayakawa & Matsukura (2003) proved difficult, as essential information (on discharge, channel width, and bedrock lithology) has not been reported in the former publication. Nonetheless, Carey et al. (2006) reported the mean discharge of the Waipaoa as 33 m^3/s. From published images, the typical width of the river in its lower reaches is ~50 m, whereas the typical knickpoint height was estimated by Crosby & Whipple (2006) as ~50 m, and in the unconsolidated alluvium along its lower reaches S can be estimated as ~1 MPa. Using equation 1, in its original form for a vertical knickpoint (as envisaged by Crosby & Whipple, 2006), the rate of knickpoint retreat in the lower reaches of the Waipaoa can be estimated at ~0.3 m/yr. Crosby & Whipple's (2006) analysis used a 'stream power law' type of approach (see above), with no specific calibration for hydrology or bedrock lithology, but assuming dependence of knickpoint propagation on catchment area. They envisaged that a knickpoint developed at the coastline at around 18 ka and propagated upstream through the lower

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reaches of the river, at the rapid rate of ~100 m/yr, thus reaching its headwaters (~35 km upstream of the present-day coastline) within a few hundred years. In contrast, using the Hayakawa & Matsukura (2003) calibration, in 18,000 years a knickpoint would have migrated only ~6 km upstream of the lowstand coastline; not only would it not have reached the headwaters of the system, it would probably not have reached inland of the modern coastline. Notwithstanding the uncertainties involved, including the fact that the hydrology of the Waipaoa river system has been dramatically modified by the effects of human settlement (e.g, Carey et al., 2006; Kettner et al., 2007), these calculations cast further doubt on the 'propagating knickpoint' hypothesis, suggesting instead that any latest Pleistocene or Holocene rejuvenation of the Waipaoa has instead occurred as a result of downcutting throughout the system.

Rationale for modelling based on 'stream power law'

The algebraic complexity of the 'stream power law' modelling gives the impression of sophistication, although the underlying assumptions about the behaviour of rivers are generally simplistic, and in some cases contrary to experience. For example, the notion that $Q \propto x^m$ is based on the assumption that downstream distance is a proxy for catchment area, this being regarded as a proxy for discharge (e.g., Hack, 1957; Weissel & Seidl, 1998). Studies of this type have nonetheless been used to predict complex development of river systems, such as to suggest that fluvial incision varies along rivers in a complex manner that bears no systematic relation to uplift. For example, Whipple & Tucker (2002) proposed that increases in the rate of uplift result in steepening downstream fluvial gradients, which would mean that the amount of incision would vary along the length of a river, even in cases where the same uplift history had been experienced throughout the catchment. Indeed, Whipple & Tucker (2002, fig. 7) discussed one case for which a doubling in uplift rate resulted in the prediction that the downstream gradient of the model river increased by ~60%. This was a theoretical prediction for a hypothetical river and was not tested against any empirical fluvial data. Nonetheless, this type of deduction raises doubts as to whether fluvial incision can be used as a general proxy for uplift, as is claimed as the basis for the uplift-incision modelling of river terrace sequences (e.g., Westaway et al., 2002, 2009).

The assumptions underpinning 'stream power law' analyses will be examined here, taking as an example the study by Roberts & White (2010), which applied this methodology to a number of river systems in Africa. First is the assumption, noted above, that discharge is proportional to x^m. This is simplistic, as many rivers will experience abrupt increases in discharge at tributary confluences, rather than the smooth variations predicted by this theoretical approximation. Some rivers, furthermore, pass through different climatic zones and, even where this is not the case, rainfall may be significantly less in downstream reaches, at low altitude, than in headwaters. Indeed, examples are known in which (as a result of loss to evaporation) discharge decreases downstream. This will be so for the endoreic river systems that die out in the interior of the Sahara Desert, which Roberts & White (2010) analysed; given that these rivers die out because their discharge decreases downstream to zero, it makes little sense to make an assumption of downstream increase in discharge in any analysis of their longitudinal profiles.

A second assumption is that all reaches of a given river have experienced the same uplift history; if this assumption were to be relaxed, the calculations would become intractable. However, evidence from many north-west European rivers indicates that uplift rates vary over quite small spatial scales. For instance, in south-east England post-MIS (Marine Isotope Stage) 12 uplift shows a significant eastward increase between the upper and middle reaches of the River Thames over distances of ~50 km (e.g., Westaway, 2011), whereas in western Germany the uplift on corresponding timescales shows a significant northward decrease between the middle and lower reaches of the Rhine over similar distances, tapering downstream towards a region (beneath the Netherlands) of subsidence and vertical accumulation of Quaternary fluvial sediment (e.g., Brunnacker et al., 1982; Ruegg, 1994). The course of the cold-climate offshore continuation of the Rhine beneath the southernmost North Sea and English Channel passes back into an area with terraces (it is therefore uplifting), over a distance of ~150 km (e.g., Bridgland & D'Olier, 1995; Busschers et al., 2007, 2008; Westaway & Bridgland, 2010). In western Asia, records from rivers such as the Ceyhan (Seyrek et al., 2008) and Orontes (Bridgland et al., 2012) show marked changes in uplift rate between different reaches, or even switching between uplift and subsidence, something also observed in Europe's two largest rivers, the Rhine (Meyer & Stets, 2002) and the Danube (see above; Miklós & Neppel, 2010).

A third assumption, that modern discharge can be used as a basis for analysis irrespective of the effects of past climate change is also questionable, as has already been noted.

Knickpoints in relation to bedrock type and geological structure

The idea that many knickpoints are associated with fluvial reaches through erosion-resistant bedrock is well established observationally (e.g., Hack, 1960, 1975; Brocard & Van der Beek, 2006). Since valleys cut through such resistant rocks are generally constricted laterally, often forming gorges, such reaches typically lack terraces (Bridgland & Westaway, 2008a, b), making it difficult to demonstrate that knickpoints in modern valley-floor long profiles are mirrored by equivalent knickpoints in terraces. Nonetheless the parallelism of terraces upstream and downstream of such reaches can provide an indication of this (e.g., between the upper and middle reaches of the River Orontes in western Syria: Bridgland et al., 2012).

It has been noted, however, that some reaches cut through erosion-resistant bedrock do not correspond with knickpoints, which has led to the suggestion (Brocard & Van der Beek, 2006) that knickpoints develop in such reaches only if incision is detachment-limited (see above). In contrast, Roberts & White (2010) have argued that there is not in general any correlation between knickpoints and bedrock lithology. In their view, however erosion-resistant a bedrock type might be, it will be erodable on timescales that are short compared with the millions of years required for upstream knickpoint migration, thus negating any likelihood of correspondence between short, steep reaches (i.e. knickpoints) and resistant bedrock. This theoretical position, however, ignores the plethora of field examples in which resistant bedrock lithology correlates with valley constriction and steepening of downstream gradient, thus coinciding with knickpoints (see above).

By way of example, a constricted reach of the River Tigris, coincident with a steepening of the downstream profile, is found at Diyarbakır, south-east Turkey (Bridgland et al., 2007b; Westaway et al., 2009; Fig. 2). The Tigris valley here, cut into Miocene clay bedrock, is flanked by fluvial terraces, many capped by basalt flows resulting from local Quaternary volcanism. The valley floor is typically several kilometres wide and its downstream gradient is ~0.8 m/km. However, for a distance of ~2 km in the southern outskirts of Diyarbakır, basalt flows (dated to ~1.1-1.2 Ma; Bridgland et al., 2007b; Westaway et al., 2009) cap fluvial terraces at levels of ~70 m above the modern river on both sides of the valley. Following the emplacement of these lava flows, the Tigris has cut a gap no more than a few hundred metres wide between basalt-capped terrace deposits. The narrow floodplain in this short (~2 km) confined reach thus has a steeper downstream gradient of ~2 m/km, forming a knickpoint in the longitudinal profile. It is likely, however, that this knickpoint has not migrated upstream from the coast to this location but instead has developed here because of the valley constriction. That is indeed apparent because the Tigris terraces further downstream (Westaway et al., 2009) show no evidence of the variation in longitudinal gradient that a migrating knickpoint would have caused. The knickpoint and valley constriction have persisted here long after the river incised below the basalt and any underlying Tigris terrace deposits into softer clay bedrock, suggesting that resistant rock capping the immediate valley sides can have a similar effect to resistant bedrock in the channel floor.

The evolution of this knickpoint can also be investigated using the Hayakawa & Matsukura (2003) approach. From the foregoing, H = ~4 m and s = ~2 m/km or 0.002. The width of the Tigris channel through this knickpoint is ~50 m (field observations by the authors) and for the local bedrock clay S can be estimated as ~2 MPa. The upstream catchment has an area of ~3000 km^2 and typical annual rainfall of ~300 mm. The mean discharge, Q, can thus be estimated as ~28 m^3/s. Using equation 2, V can thus be estimated as ~0.016 m/yr. Upstream propagation of the knickpoint can thus be calculated as only ~160 m per 100 ka climate cycle. Following the constriction of this reach of the Tigris valley, the river has repeatedly incised and then new terrace deposits have been emplaced at a succession of lower levels. The resulting repeated phases of downcutting have evidently recreated the knickpoint in the constricted reach of the Tigris valley; any subsequent upstream propagation prior to the next downcutting phase can therefore be expected to be no more than a few hundred metres.

In the wider reach of the Tigris valley further downstream, W is ~100 m and S can be estimated as ~1 MPa, for unconsolidated Quaternary fluvial deposits. With these parameter values, the rate of upstream propagation of a vertical knickpoint can be estimated, using equation 1, as ~1.1 m/yr, indicating that upstream propagation by ~110 km would have occurred during each 100 kyr climate cycle. It follows that, if landscape development in the Tigris catchment had been controlled by phases of rejuvenation caused by base-level change (as a result of phases of global sea-level fall) and associated upstream propagation of knickpoints (as envisaged for the Waipaoa by Crosby & Whipple, 2006), the expected result would be a succession of knickpoints every ~100 km along the Tigris valley. However, there is no such evidence, a

Fig. 2. View southward, looking obliquely downstream in the Tigris valley from the southern edge of the historic city centre of Diyarbakır (at Universal Transverse Mercator (UTM) coordinates FB 086 960). The multi-arched bridge, ~2 km from the viewpoint, is within the reach of the valley constricted by basalt flows (see text), in contrast to the much wider valley upstream of the constriction, seen in the left foreground.

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further indication that this conceptual approach to the modelling of fluvial systems is incorrect.

The other great river of Mesopotamia, the Euphrates, provides a second example. In south-eastern Turkey there is a significant knickpoint coincident with a constricted gorge reach of the Euphrates at the northern boundary of the Arabian Platform, where it adjoins the southern margin of eastern Anatolia. The bedrock along this constricted gorge reach is Mesozoic limestone (e.g., Altınlı & Erentöz, 1961), a similar lithology to that found along much of the Euphrates further downstream, where no knickpoint is present. In detailed recent analyses of the Euphrates system (Demir et al., 2007a, b, 2008, 2012) no evidence has been found that would support the upstream migration of the knickpoint to this location; it evidently formed in situ due to the rapid Late Cenozoic uplift of the region and the localised bedrock-controlled constriction. In this constricted reach, of ~40 km downstream extent, the longitudinal gradient of the Euphrates is ~3 m/km, in contrast to the ~0.5 m/km typical elsewhere in south-eastern Turkey (Demir et al., 2012). The mean discharge (Q) of the Euphrates here is ~1000 m^3/s (e.g., Ionides, 1937; Demir et al., 2008). The excess fall of the river in this reach, providing an estimate of H, is thus ~100 m, and with W, say, ~50 m (cf. Demir et al., 2009) and s ~0.003, and S taken as ~200 MPa (appropriate for highly lithified bedrock), then the application of equation 2 suggests a rate of upstream knickpoint migration of only ~5 mm/yr. This knickpoint has evidently formed in situ, presumably as a consequence of the much faster Late Cenozoic uplift of the Anatolian Plateau compared with the adjacent parts of the Arabian Platform (Demir et al., 2009), and has not migrated significantly upstream. If the uplift has been concentrated in the past ~3 million years, as Demir et al. (2009) inferred, then the knickpoint has migrated upstream by no more than ~15 km, so remains localised in essentially where it formed.

A further example of a major knickpoint on a large river is provided by the reach of the Colorado along the Grand Canyon, through the uplifting Colorado Plateau, as illustrated in Fig. 3. Along this ~300 km reach the Colorado falls by ~600 m, at a longitudinal gradient of ~2 m/km, roughly ten times greater than in the reach of the same river further upstream. The overall excess fall at this knickpoint, H, is thus ~540 m and with Q = 590 m^3/s prior to loss of flow due to damming (e.g., Cohen et al., 2001), s = 0.002, W = 100 m, and S = 300 MPa (the latter value appropriate for highly lithified basement rocks) then, from equation 2, the rate of upstream propagation of this knickpoint can be estimated as ~40 mm/yr. As for the Euphrates (see above), this rate is very low. The modern geometry of the River Colorado came into being in the latest Miocene, prior to which the Gulf of California (into which it now flows) did not exist. This arm of the sea began to form when the geometry of plate motions bordering western North America changed, such that the San Andreas Fault Zone and the modern plate boundary in the Gulf of California began to develop. For example, Oskin & Stock (2003a, b) established that plate motions became localised in the Gulf of California at ~6 Ma; Dorsey et al. (2007) reported the onset of marine conditions in the northern Gulf of California at ~6.3 Ma and the first appearance of sand from the incipient Colorado River at ~5.3 Ma. Furthermore, the well documented

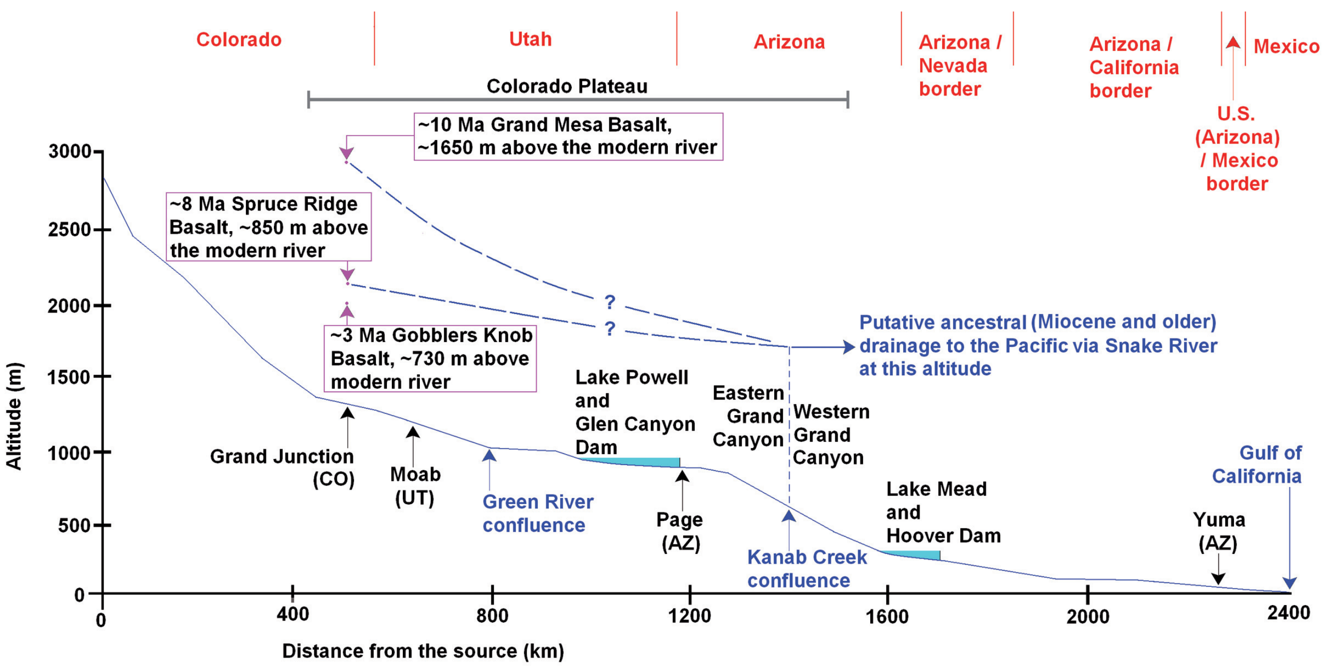

Fig. 3. Long profile of the Colorado River. Putative Miocene and older drainage of the upper reaches of the modern Colorado system to the Snake River is from Luccitta et al. (2011); disposition of dated basalt flows in the upper Colorado catchment is from Bridgland & Westaway (2008b), who listed the original sources of information. Modified from http://newyorkscienceteacher.com/sci/files/user-submitted/ColoradoProfile2001.pdf.

presence around the mouth of the Grand Canyon of sediments indicative of internally-draining basins, characteristic of the Basin and Range Province, provides strong evidence that prior to the latest Miocene or earliest Pliocene no fluvial drainage existed along the modern course of the lower Colorado (e.g., Longwell, 1936; Lucchitta, 1972). Lucchitta et al. (2011) have recently proposed that the pre-Latest Miocene Colorado flowed northwards for ~800 km to join the Snake River in southern Idaho and thus reach the Pacific Ocean (Fig. 3). Although there is no sedimentary evidence to support a former fluvial connection of this type, it could explain the similarities in fish faunas between the Snake and Colorado systems (e.g., Spencer et al., 2008).

The history of incision by the Colorado River, and uplift of the landscape through which it flows, was discussed by Bridgland & Westaway (2008b), based on a compilation of previously published evidence from the vicinity of Grand Junction, Colorado, some 1000 km upstream of the putative diversion point discussed above. In the Grand Junction reach the Colorado has formed an extensive terrace staircase, several of its terraces being precisely dated as a result of basalt cappings from local volcanism or tephra from larger eruptions such as that of Yellowstone at ~0.6 Ma. The Colorado in this region has experienced phases of fluvial downcutting linked to times of global climate change (notably circa 3 Ma and following the MPR), as established in many European rivers, suggesting that its Pliocene-Pleistocene incision history has been controlled by phases of climate-induced erosion (Bridgland & Westaway, 2008b). This is consistent with the earlier calculation of the rate of upstream propagation of the Grand Canyon knickpoint; if governed by the Hayakawa & Matsukura (2003) theory, with the parametrer values specified, this knickpoint will have propagated upstream by no more than ~240 km during the existence of the modern Colorado system, and so will not yet have reached the Grand Junction area. Indeed, as for the Tigris example, the repeated phases of incision every ~100,000 years will probably have resulted in the abandonment of former valley floors and the re-establishment of the knickpoint, so upstream propagation by no more than ~40 mm/yr × ~100 kyr or ~4 km can be expected. The Bridgland & Westaway (2008b) explanation of the uplift and fluvial incision in this region (as a consequence of erosional isostasy) is supported by local specialists (e.g., Lazear et al., 2010; Pederson et al., 2010), notwithstanding the alternative hypothesis that they are the product of movement of 'hot blobs' within the Earth's mantle (e.g., Moucha et al., 2009; Robert et al., 2011). The Grand Canyon knickpoint evidently owes its existence to the extreme constriction of the Colorado gorge at this point, as the river here is entrenched into highly erosion-resistant Palaeozoic and Precambrian bedrock. Although the longitudinal profile of the Colorado (Fig. 3) might superficially be explained in terms of a knickpoint that has migrated inland from the Gulf of California to the modern Grand Canyon region, the overall evidence indicates that such a hypothesis has no foundation.

Amos & Burbank (2007) have recently investigated the effect on river valley morphology of valley constriction due to active faulting in the South Island of New Zealand. They noted that a river might keep pace with localised uplift within upthrown fault-bounded blocks by developing narrower and deeper channels; however, if this adjustment fails to produce the necessary increase in stream power the river adapts by locally increasing its downstream gradient. The behaviour of rivers in such a situation (and, by analogy, in localities such as Diyarbakır, where valley constriction has been produced by bedrock lithology rather than by active faulting: Fig. 2) is thus quite complex, being dependent on thresholds in relation to the erosional power of the river and the magnitude of the valley constriction. An informative comparison can be made with the neighbouring River Euphrates in its south-east Turkey reach. Much larger than the Tigris, the Euphrates cuts through a number of active anticlines, evidently underlain by active reverse faults, which correspond with lateral constriction of its valley although without any measurable localised variation in downstream gradient (Demir et al., 2012). However, despite the lack of evidence of variations in downstream gradient, the localised deformation can be recognised by tilting and warping of the Euphrates terraces. This river thus appears to be in the mode (cf. Amos & Burbank, 2007) of valley constriction without a change of gradient, whereas the smaller Tigris at Diyarbakır (Fig. 2) is in the mode where valley constriction has also required a local increase in downstream gradient.

Another notable example of a knickpoint occurring where a river flows through a zone of localised crustal deformation is Tiger Leaping Gorge, in the valley of the River Yangtze, China (Westaway, 2009; Figs 4 & 5). Despite being one of the world's largest rivers, the Yangtze has evidently been unable to incise sufficiently rapidly to keep pace with the rapid localised uplift in this gorge reach, along which the river flows through the Yulong antiform, a zone of active crustal shortening. On the basis of thermochronology, the Yulong antiform has uplifted, and the gorge has become entrenched, by ~4 km in the past ~5 million years, since the present-day pattern of crustal deformation in this region began (see Westaway, 2009, and references therein). Others (Kong et al., 2009a, b, 2010, 2012) have subsequently argued that the rates of landscape development in this region are even faster, and would account for the present-day topography and gorge morphology in less than a million years. However, these rates are largely based on cosmogenic dating, which does not always produce reliable ages for fluvial deposits, as different assumptions about burial and erosion histories can result in very diffrent numerical ages (e.g., Hanks & Finkel, 2005), whereas thermochronology, as used by Westaway (2009), is considerably more robust. The very high rates of landscape development in this region that would follow from these cosmogenic dates have given credence to views that the Yangtze has experienced major diversions

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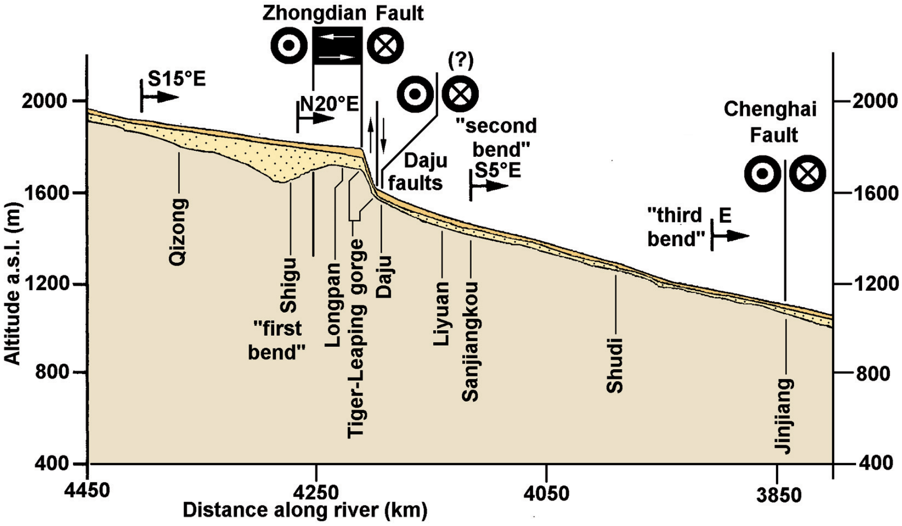

Fig. 4. Longitudinal profile of part of the upper Yangtze, showing variations in the thickness of fluvial sediment (stippled) below the modern channel bed and sites where the valley is offset by recent active faulting. Paired arrows and arrowhead and tail symbols provide standard indications of senses of transcurrent motion. Adapted from fig. 5-1 of Yang (2006). Note the marked variation in conditions at the upstream end of Tiger Leaping Gorge (see also Fig. 5), which would be inexplicable if there were no significant active slip on the Zhongdian Fault (cf. Burchfiel & Wang, 2003) and/or no significant active shortening in the western limb of the Yulong antiform (cf. Kong et al., 2010), the core of which coincides with this gorge. From fig. 4 of Westaway (2009).

Fig. 5a.

b.

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c.

Fig. 5. *Field photographs contrasting the constricted reach of the Yangtze in Tiger Leaping gorge and the reaches above and below this constriction, in which the river is responding to regional uplift (at an estimated rate of ~0.5 mm/a: Westaway, 2009) without the additional effect of localised active crustal deformation. a. View eastward from (UTM) PA 18808 20510, looking downstream and across the Daju basin from the outlet of Tiger Leaping Gorge. Several Yangtze terraces within the Daju basin are visible, as are terraces in the footwall of the Daju normal fault (facing away from the viewpoint) bounding the western margin of this basin, which (it is suggested) are offset counterparts of the highest terraces within the Daju basin (cf. Westaway, 2009). The fluvial terrace that is offset by the normal fault (near the centre of the view) is ~1900 m a.s.l in the footwall but ~1800 m a.s.l in the hanging wall, with the river at ~1600 m a.s.l. This is the normal fault that Kong et al. (2009b, 2010) envisaged as having a vertical slip rate of ~6 mm/a and as controlling the uplift of the Yulong mountain range. This interpretation is not favoured here, in part for the reasons discussed in the main text and in part because the ~50% variation in the height of the offset fluvial terrace is not consistent with the extremely high slip rate envisaged, notwithstanding the cosmogenic dating evidence that Kong et al. (2009b, 2010) presented in support of their argument; b. View eastward from PA 17901 19325, looking across the Yangtze valley just upstream of its outlet from Tiger Leaping Gorge, showing the highest of its terraces, ~300 m above the river (~1900 m against ~1600 m a.s.l; also visible in 'a', which locally forms a strath, cut into Palaeozoic marble. Buildings on the terrace surface provide an indication of scale; c. View NNE down the lower half of Tiger Leaping Gorge from PA 13418 13878, in the core of the Yulong antiform. Buildings in the middle distance provide an indication of scale. The cliff, mostly in Palaeozoic marble, in the middle distance (~5 km away) rises to a spot height at 3617 m a.s.l, the river below being ~1700 m a.s.l. The far skyline (~15 km away) marks the eastern margin of the Daju basin (cf. 'a'); d. View eastward across the Yangtze from PV 05813 90030, near Longpan, showing the two Yangtze terraces, ~35 and ~10 m above the present low-stage river level, the deposits of which (according to estimation by Westaway, 2009), may date from MIS 4 and MIS 2, repectively. The viewpoint is ~1800 m above sea level; the land surface in the background rises to the snow-capped summit of the Yulong or Jade Dragon mountain range at 5596 m a.s.l. The Yangtze, in the foreground, flows locally northward (i.e., from right to left); before turning abruptly eastward on entering Tiger Leaping Gorge, which cuts through the Yulong mountain range. Parts 'a' to 'c' are from Fig. S2 in the online supplement to Westaway (2009), which provides additional detail about this reach of the Yangtze.*

d.

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during the Quaternary, including flow reversal over thousands of kilometres of its length (e.g., Brookfield, 1998; Clark et al., 2004; Kong et al., 2012), so its present course is very young; such ideas were dismissed as speculation by Westaway (2009). Regardless of the precise chronology, localised crustal deformation in this area (Fig. 4) has created the present highly constricted morphology with an extremely steep local downstream gradient; as Fig. 4 illustrates, the river falls by ~200 m in ~20 km at a roughly uniform gradient of ~10 m/km.

Westaway (2009) inferred that development of the Yulong mountain range has resulted primarily from anticlinal folding, with only minimal contributions from vertical components of slip on faults at the ends of the gorge. The resulting uplift rate of ~4000 m / ~5 Myr or ~0.8 mm/yr along the range axis was thus envisaged as accommodated by continuous warping of the crust, so the river could readily adjust to it to maintain the observed uniform downstream gradient along Tiger Leaping Gorge (Fig. 4). In contrast, Kong et al. (2010) envisaged that the uplift is much faster and is occurring as a result of slip on the Daju normal fault at the downstream end of the gorge, depicted schematically in Fig. 4, at a vertical slip rate of ~6 mm/a, this rate resulting from metres of slip in occasional large earthquakes (say, 2 m of vertical slip every ~300 yr). If correct, this would require that the resulting knickpoint, produced by the vertical fault offset, must have time to propagate to the upstream end of the gorge before the next earthquake, otherwise the cumulative effect of many earthquakes would be to cause the river gradient to steepen towards the fault (as in the examples depicted by Amos & Burbank, 2007), rather than being uniform throughout the gorge (Fig. 4). This proposal can be tested using equation 2, assuming S = 200 MPa (appropriate for the Palaeozoic metamorphic rocks that crop out along Tiger Leaping Gorge), Q ~2000 m^3/s (estimated for a ~150,000 km^2 upstream catchment area with mean annual rainfall of ~400 mm), W ~20 m (from field observation), with H assumed to be 2 m at the exit of the gorge and s/H assumed to decrease linearly upstream from there to zero at the upstream end of the gorge. It can thus be estimated from Equation 1 that V is ~47 m/yr at the Daju fault scarp and, from Equation 2, that V will decrease upstream to ~1.2 m/yr after 100 m, ~0.4 m/yr after 500 m, ~0.2 m/yr after 1000 m, and so on. It will therefore take more than 500 yr for the 'knickpoint' generated by the fault offset to propagate 500 m upstream, longer than the assumed earthquake recurrence interval. Similar results will be obtained for any other earthquake recurrence history consistent with a 6 mm/a vertical slip rate. It is thus concluded that the mechanism and rate envisaged by Kong et al. (2010) for the formation of Tiger Leaping Gorge are inconsistent with the observed downstream gradient of the Yangtze. Although these are only crude calculations, this case study demonstrates the value of applying this quantitative theory to the fluvial geomorphology to test a controversial geological interpretation.

Knickpoint migration in 'deep time'

The 'stream power law' methodology, based on knickpoint analysis, offers the ability to explore the development of rivers back into 'deep time', on timescales that extend to before the formation of their oldest terrace deposits. For example, the rivers studied by Roberts & White (2010) in arid parts of Africa are hundreds of kilometres long and provide information, according to those authors, about past uplift events over several tens of millions of years. This is possible because the climate aridity of this region necessitates parameter values that result in relatively low calculated rates of upstream knickpoint migration (although continuous aridity over the period represented has not been established).

It is evident from foregoing examples that in large rivers knickpoints can develop along constricted gorge reaches and remain localised there for long periods of time, as is generally indicated by terrace evidence from upstream/downstream of the constricted reaches. Indeed, in the cases of the Yangtze and Colorado, the example knickpoints can be shown to have existed in approximately their present locations for ~5 million years. On the other hand, the validity of upstream knickpoint migration as a mechanism for effecting downcutting has yet to be established. It has been pointed out (see previous section, for the Tigris) that, if this has occurred, knickpoints related to successive climate cycles should be preserved along river valleys at well-defined intervals that relate to the succession of Pleistocene sea-level changes. These would represent the general mechanism for the long-timescale fluvial incision observed globally, but no such succession of knickpoints has been documented in any world river. The spacing of the succession of knickpoints would depend on the conditions in each river system. For example, if each Pleistocene sea-level fall is assumed to initiate a vertical knickpoint with H = 10 m that propagates upstream, then taking W = 200 m, s = 0.2 m/km (Fig. 3), and S = 1 MPa for the unlithified fluvial deposits forming the modern valley floor, the rate of upstream propagation of such a knickpoint can be estimated as ~3.3 m/yr. Each resulting knickpoint would therefore propagate upstream by ~330 km every 100,000 years, so (even allowing for uncertainties in the above-mentioned parameter values) several of them would be expected to be present in the ~800 km long reach of the Colorado between the downstream end of the Grand Canyon and the coast. The fact that none is observed casts obvious doubt on the hypothesis.

Conclusions

Two contrasting approaches have emerged in recent decades, both seeking to make use of data from river systems to determine the history of uplift and resultant landscape evolution. One, advocated by the authors, is grounded in the study of fluvial sedimentary records, especially river terrace staircases. This has the advantage of being thoroughly grounded in the record of Quaternary climatic fluctuation, which is seen as responsible for driving the changes in fluvial activity, partly by way of the influence of climate on vegetation (and therefore slope stability), that have given rise to aggradational river terraces in uplifting regions; Jef Vandenberghe has indeed frequently engaged in discussion of the complex behaviour of fluvial systems in relation to climate. In some areas, such as along the south coast of England, fluvial incision can indeed be shown to be a proxy for uplift, given its consistency with heights of marine terraces and karstic levels (e.g., Westaway et al., 2006; Westaway, 2010); however, in many other regions such direct tests are not possible, even though the principal conditions for validity of this assumption (the parallelism of river terraces and the comparable sedimentology, regarded as evidence for similarity in palaeohydrology; e.g., Westaway et al., 2002) may be readily observed.

The second methodology, here termed the 'stream power law approach', predicts different behaviour of fluvial systems, for example offering the possibility of identifying phases of uplift in the distant past from patterns of knickpoint distribution in fluvial long profiles. The former technique makes use of multi-proxy evidence to determine the dating of occupation by rivers of particular floodplain (now terrace) levels, whereas the latter is not strongly evidence based and has little scope for calibration with reference to geochronology. Its premise that knickpoints have formed at coasts and have migrated upstream over geological timescales is difficult to reconcile with available evidence, not only from terrace records but also from calculated rates of knickpoint retreat. A particular shortcoming of the second of these approaches is that it takes no account of the effects of climatic fluctuation, despite this being the overwhelmingly predominant characteristic of the most recent geological record. Indeed, this omission must raise considerable further doubt about the validity of the 'stream power law approach'. In this context it is worth quoting from an early contribution to the debate on climatic geomorphology, by Passarge (1972, p. 91; an English translation of a paper from 1926): "present-day surface landforms are for the most part not the result of present climates, but the product of Pleistocene processes". Therein he summarised the folly of a modelling approach that takes account only of modern-day conditions.

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