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Degrees of separation: Hillslope-channel coupling and the limits of palaeohydrological reconstruction

John Wainwright

Department of Geography, University of Sheffield, Winter Street, Sheffield, S10 2TN, UK

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Abstract

Qualitative analysis of hillslope-channel coupling conditions suggests that the internal configuration of a catchment has a strong influence on the transfer of water and sediment through the fluvial system. Consequently, similar climatic inputs into catchments with otherwise similar characteristics can result in very divergent responses. A cellular modelling approach has been used to evaluate the potential magnitude of this effect, and to investigate the implications for palaeohydrological investigations. Simple catchments with coupled and progressively uncoupled conditions (marked by the presence of one and two floodplain levels) were subjected to simulated climate change using a 740 ka proxy climate record. Dynamic flow and sediment routing were simulated, and vegetation and soils allowed to evolve in parallel. The results suggest that catchment configuration has a very strong effect producing a complex system response based on the output of sediment from the catchment and that there are no simple relationships between climate and catchment output. The complex response itself also evolves through time as the catchment dynamically reorganizes, implying that system trajectory and historical contingency are also significant factors. Although events may be captured in the sedimentary record, these events will often relate to pulses of sediment working their way through the system, rather than being directly related to climate changes or variability. The complex, non-equilibrium behaviour of the system suggests that climate proxies derived from even high-resolution palaeohydrological records should be treated with extreme caution.

Keywords: Catchment hydrology; Climate proxies; Geomorphic modelling; Complex response; Landform evolution; Equifinality; Divergence

1. Introduction

In his overview of the place of palaeohydrology in scientific hydrology, Baker (1998, p. 3) notes that "the major function of pal[a]eohydrology in regard to applied/engineering and basic/geophysical hydrology is the supplying of data". These data may be used in terms of planning long-term landscape impacts or in the testing of hydrological models. Despite the logical limitations of both approaches (Baker, 1998; see also the discussion on validation in Mulligan and Wainwright, 2003), they continue to be applied within major projects. Baker (1998) discussed the limitations of the Past Global Changes (PAGES) project, but similar attempts to quantify past hydrological events is still a central feature of collaborative research efforts such as the Global Continental Palaeohydrology Commission of ICSU (Gregory and Benito, 2003; GLOCOPH, 2004). Furthermore, the general literature supports the argument for the development of palaeohydrological proxies from the fluvial record (e.g. Bradley, 1999; Lowe and Walker, 1996) despite the known complexity of the fluvial system (Knighton, 1998; Robert, 2003).

The development of a palaeohydrological proxy requires the production of a transfer function to relate some aspect of channel morphology or sediment characteristics to past flow régime. Relationships derived from modern channels assumed to be at equilibrium have generally been used to retrodict past flow conditions (e.g. Schumm, 1965, 1969). However, above and beyond the problems associated with defining whether a modern channel–not to mention a

E-mail address: j.wainwright@sheffield.ac.uk.

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palaeochannel–is at equilibrium (e.g. Bracken and Wainwright, submitted for publication), Dury (1985) demonstrated that there is a significant degree of uncertainty introduced by the power-law form of the transfer functions used in such an approach. The short timeframe of observation of modern channels and uncertainties in the climatic and flow regimes that generated them compounds this uncertainty. Uncertainties may be an order of magnitude or more, but this wide range may be considered acceptable given the lack of other quantitative information. Furthermore, the complex response of catchment systems has been recognized for more than 30 years (Schumm, 1973). This concept notes that the behaviour of the catchment system to a specific climatic impulse will depend on the configuration of the catchment and its trajectory of development. Wainwright (1996) attributed difficulties in reconstructing flooding characteristics from contemporary process observations following the Ouvèze floods of 1992 in SE France to uncertainties in the characteristics and dynamics of hillslope-channel linkages. Church (2002) demonstrated how the coupling between hillslopes and channels creates spatial thresholds in the hydrological and sedimentological characteristics of the channel system. The palaeohydrological implication of these thresholds is that detailed information about the catchment setting is required in order to reconstruct the climatic trigger of events or past flows. The nature of hillslope-channel coupling within a catchment system and the evolution of the coupling through time are thus important controls on the complex response of the catchment. A failure to account for these linkages will therefore significantly affect the quality of interpretations that can be made, specifically in terms of producing sound proxy information, even before consideration of the preservation potential (Lewin and Macklin, 2003) can be made.

This paper aims to discuss the limitations of developing potential transfer functions for palaeohydrological reconstructions of climate records in two linked ways. First, the effects of hillslope-channel coupling will be assessed in a qualitative sense, to evaluate the ways in which catchment structure may affect the transmission of water and sediment. Secondly, a model that is capable of representing the dynamics of differently coupled catchment systems will be used to evaluate the potential response of catchments over long time periods. In this way, the potential information that may be recoverable from relatively simple catchment systems will be evaluated and broader implications for palaeohydrological investigations will be drawn.

2. Qualitative aspects of hillslope-channel coupling

In the simplest case, the hillslope element of the catchment is directly coupled to the channel (Fig. 1a). Such a configuration is most common in headwater basins. Hydrographs from the hillslope into the channel will rise and fall sharply, and sediment production, assuming supply is not limiting, will relate directly to the flows creating them.

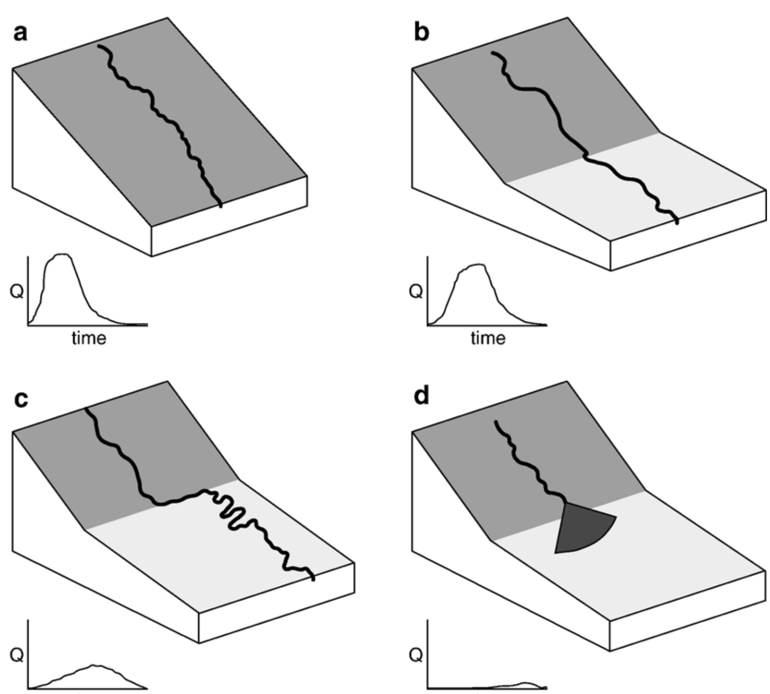

Fig. 1. Effects of different hillslope-channel coupling configurations on the water and sediment reaching the channel system: (a) directly coupled hillslope and channel; (b) hillslope decoupled from channel by floodplain; (c) decoupled case with more complex pathways; (d) hillslope completely decoupled from channel with fan deposition on floodplain.

A simple transfer function may be developed to relate rainfall to runoff and sediment yield, which can then be used to retrodict the climate. Even assuming that the transfer function is valid over long time periods (which is unlikely due to infiltration rate changes as soil and vegetation change and the topography of the hillslope changes through time) such a situation is of very limited palaeohydrological interest, because of the low preservation potential of the sediments or channel structures derived from them. In an ideal situation, records may be developed on a decadal to century timescale if such a configuration feeds into a small reservoir, pond or lake.

Further downstream, it is more common that sediment temporarily stored within the system forms floodplains which affect the ways in which water and sediment are transferred from the hillslopes to the channel. Where the floodplain has a high infiltration capacity – for example due to a very coarse texture or a long period of antecedent dry conditions – flows from hillslopes or first-order basins may be totally absorbed (Fig. 1b). The impact on the channel hydrograph will therefore be negligible, or at best much deferred and diminished due to the buffering effects of subsurface flows, at least in cases where the lateral valley gradient is insufficiently steep to permit the development of significant cross-floodplain drainage networks. Sediments are likely to be deposited as smaller or larger fans on the edge of the floodplain and add to the storage on the floodplain. Floodplains with lower infiltration capacities may allow a more direct transfer of water into the channel, although the decrease in stream power at the break of slope may often cause the localized deposition of sediments being removed from the hillslopes (Fig. 1c). Depending on the local topography, vegetation and presence of smaller channels due to past overbank flows, the water may travel more or less directly into the channel system. Where the flow is more indirect, the opportunity time for infiltration (transmission loss) is likely to reduce the total water transferred and diffuse its arrival time (Fig. 1d). These simple examples serve to illustrate how hillslope-channel coupling creates a divergence of system behaviour. The water and sediments arriving in the channel system in particular events will be closely controlled by the detail of the hillslope and floodplain configurations.

Interactions between the channel and adjacent hillslope or floodplain elements will also complicate the picture (Fig. 2). A channel that can erode laterally compared to one that is constrained within its banks (perhaps due to vegetation cover) may produce a more pulsed sediment-discharge régime — for example due to the initial release and pulsing of sediment through the system. Where coupling is direct, or where bank collapse occurs on a large scale, the channel may become blocked with sediment, which is again released after a time lag; alternatively, a sediment pulse may be generated from channel diversion and undercutting of the opposite bank or hillslope.

From a perspective of resulting channel form, there will thus be conditions of equifinality from the internal organization of the catchment. Proxies based on channel form will thus suffer from non-unique interpretations. In terms of looking at the response to climatic inputs, there will

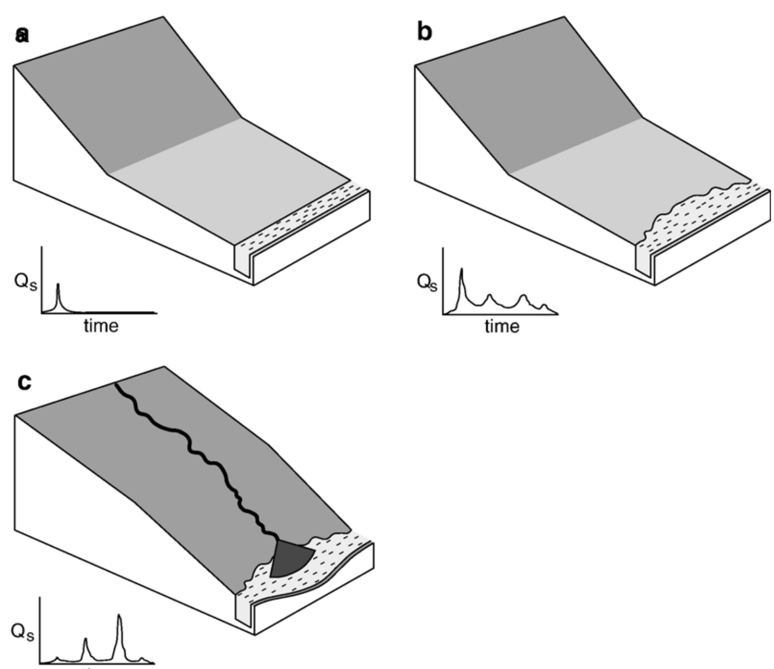

Fig. 2. Effects of different interactions between the channel and adjacent floodplain or hillslope elements on sediment response to the same climatic event: (a) channel interacting only marginally with the floodplain; (b) channel coupled directly to floodplain and causing lateral erosion; (c) directly coupled hillslope and channel with temporary blockage of channel due to fan development or bank collapse.

be a system divergence. Transfer functions to reconstruct these climate inputs will therefore have non-unique solutions. Given that the dynamics of hillslope-channel coupling are a fundamental but poorly understood element of catchment behaviour, there is a need to know the potential magnitude of the effects of coupling. Because of the complex response of the catchment system, it is difficult to evaluate the impacts of coupling simply beyond localized examples on an event basis of the simple style noted above. A modelling framework is the most effective method of evaluating complex systems of this nature, although the approach taken will be to use simple geometries to demonstrate that complex response need not be a function of complex spatial organization of a catchment.

3. Modelling catchment response

3.1. Model description

In order to model the long-term dynamics of hillslope-channel coupling within a catchment system, it is necessary to understand the interactions between flow, sediment transport and topography within a process-based framework. It is necessary to account for changes in climate, which will be dealt with here as input parameters, and the consequent changes in vegetation cover. Weathering is also included in order to simulate the long-term generation of soil materials on hillslopes so that supply limitation of sediment is not unrealistically produced. The simulations here will represent timescales of several hundreds of thousands of years to evaluate whether responses can be defined in relation to longer term feedbacks, as well as on annual cycles. There is a necessary trade-off in the temporal and spatial resolution of the features modelled. An annual time-step has thus been employed with a DEM cell size of 100 m. While the latter is too coarse to represent the detailed evolution of the floodplain features described above, it is sufficient to evaluate the impact of the broad-scale catchment configuration and thus produce a conservative estimate of the impacts of hillslope-channel coupling. A cell-based approach has been used to simulate the dynamic interactions and allow the self-organization of processes within the catchment, rather than to use a global solution to flow and sediment-transport equations, the numerical approximation of which would linearize the behaviour of the system (Chase, 1992; Coulthard et al., 2002; Favis-Mortlock et al., 2000; Mulligan and Wainwright, 2003; Murray and Paola, 1997).

Runoff production is simulated by estimating the annual runoff coefficient at each cell within the model domain. The runoff coefficient (RC [%]) is considered as a function of grain size (stone content of soil, G [%]), vegetation cover (V_c [%]) and soil moisture (θ [m m^{-1}]):

$$\mathrm{RC} = 100 \cdot (1 - 0.0085G) \cdot e^{-0.025V_c} \tanh\left(0.6 + \frac{\theta}{0.2}\right) \quad (1)$$

and $V_c = V/76.5$, where V is vegetation biomass [kg] (Thornes, 1988). Flow is calculated from the annual rainfall and then routed across the topography using the D8 algorithm (Quinn et al., 1991). This approach again provides a conservative estimate of the impacts of hillslope-channel coupling, in that it indirectly simulates the loss of water across floodplains through the soil–moisture effect, rather than by enhanced transmission losses.

Both diffuse and concentrated water erosion are simulated. Diffuse erosion by unconcentrated overland flow (E_D [mm]) is estimated using a modified Musgrave approach:

$$E_D = k_1 q^2 S^{1.67} \frac{(110 - G)^{0.877}}{61.7} e^{-0.07V_c} \quad (2)$$

where k_1 is diffuse erosion erodibility calculated using the USLE approach (Mitchell and Bubenzer, 1980; see Zhang et al., 2002 for details on this approach), q is runoff depth [mm] and S is surface slope [m m^{-1}]. Concentrated erosion by channelized flow (E_C [mm]) is simulated using a modified stream-power approach:

$$E_C = k_2 q_*^2 S e^{-0.07V_c} \quad (3)$$

where k_2 is concentrated erosion erodibility and q_* is runoff depth [mm] above the threshold for initiation of concentrated erosion, if the depth is above the threshold, or zero otherwise. Diffuse erosion is limited by the total soil thickness, whereas concentrated erosion is not considered to be so limited, so that bedrock incision is represented. Transport and deposition is simulated using a simplified version of the transport-distance approach (e.g. Wainwright et al., 1999) so that the spatial scaling of erosion is appropriately considered (Wainwright et al., 2001; Parsons et al., 2004). Although this approach with an annual time step does not permit the impacts of seasonality of climate and vegetation change to be simulated, as in the modelling approaches of Tucker and Slingerland (1997) or Coulthard et al. (2002), Band (1985) and Wainwright (1994) have demonstrated that in the longer term, this annual approach can produce equivalent average results in terms of overall landscape evolution.

Vegetation cover is a significant control of both flow and erosion in the model. Growth or die-off of vegetation biomass is simulated using a logistic growth model:

$$\frac{\mathrm{d}V}{\mathrm{d}t} = r^{\pm} V \left(1 - \frac{V}{V_{\max}}\right) \quad (4)$$

where t is time [a], $r^{\pm}$ is the rate of vegetation growth or die-off [a^{-1}], estimated as a function of mean annual temperature (T [°C]) and temperature change (ΔT [°C]):

$$r^{\pm} = \Delta T \left(1 \pm \tanh\left[\frac{T - 12}{2.5}\right]\right), \quad (5)$$

the sign in parentheses being the same as the sign of the temperature change, and $V_{\max}$ [kg] is the vegetation carrying capacity, estimated as a simple function of annual precipitation (P [mm]):

$$V_{\max} = 2500 + 5P. \quad (6)$$

The limitations of a simplified approach such as this compared to more complex physiological models is discussed in detail in Osborne (2003).

3.2. Climate reconstruction

In order to simulate realistic palaeoclimatic changes, data from the EPICA core from Dome C, Antarctica have been used (Augustin et al., 2004). The published 739,973-year sequence of δD data has been linearly interpolated for simplicity and converted to a relative temperature change dataset by comparison with the relationship between δD and temperature at the shorter Vostok core (Petit et al., 1999, 2001). This relationship results in relative temperature changes ranging from −9.68 °C to +4.83 °C. Relative magnitudes of precipitation change through time were estimated by comparison with reconstructions made from long pollen cores in Europe (Guiot et al., 1989), with timings and relative amount a simple linear function of the relative temperature change. The relative precipitation thus defined varies from 25% to 137% of the modern value. This approach is very simplistic in relation to the actual interrelationship between temperature and precipitation changes, but has been used in order to minimize the sources of variability in the modelling experiments.

3.3. Scenarios

To test the effect of hillslope-channel coupling on the response of the model, three simple scenarios were used in a simple, hypothetical catchment. The purpose of these scenarios is to evaluate the ways in which relatively minor changes in topographic configuration might affect the catchment response. Catchment response was defined in terms of water and sediment outflow past the lowest edge of the channelled part of the DEM to reflect the conditions at a representative cross section that might be investigated as part of a palaeohydrological study. Alternatively, in the optimal case of the catchment feeding into a lake which preserves the sediment produced as a series of laminae, it would allow the testing of the idea of laminar thickness providing a climate proxy from the system.

In the first scenario, the catchment is directly coupled to a linear channel, which is burnt into the DEM of the central portion of the lower part of the catchment (Fig. 3). The catchment is 5 km wide by 10 km long, has an average hillslope angle of 10° and channel slope of 1°. The second scenario has a decoupled configuration by inserting a floodplain into the middle 500 m of each side of the central channel. The floodplain slopes at an angle of 1° towards the channel and at 0.95° downstream, producing an increasingly incised system towards the outflow. Adding this floodplain changes 15% of the cells in the DEM by an average of 0.8 m. The third scenario further decouples the hillslopes from the channel by introducing a second floodplain into the middle 200 m of each side of the central channel. The second

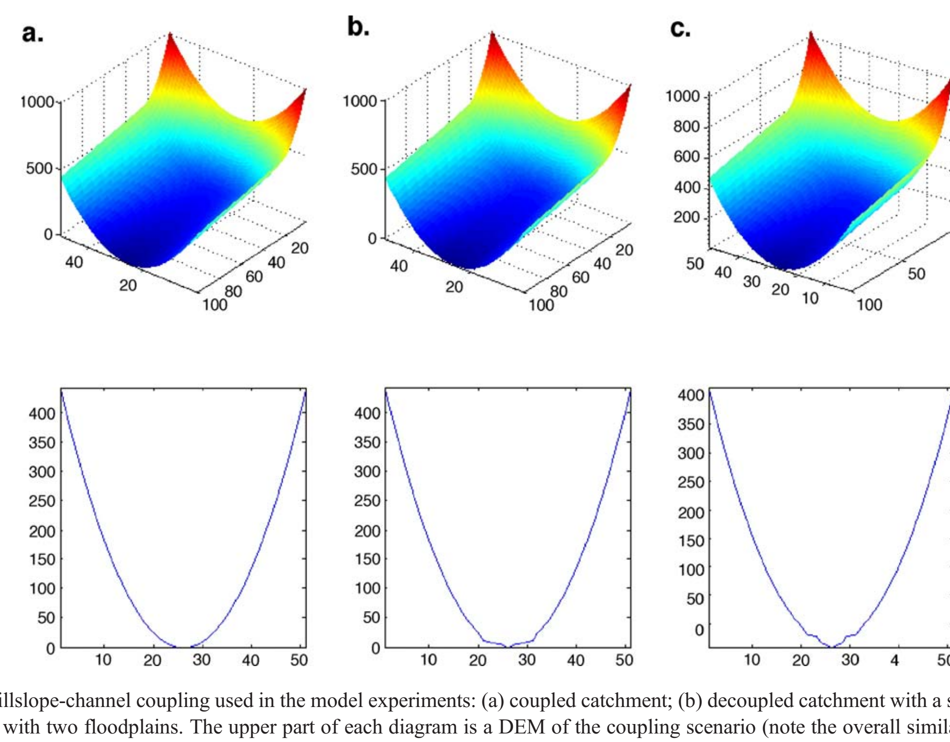

Fig. 3. Scenarios of hillslope-channel coupling used in the model experiments: (a) coupled catchment; (b) decoupled catchment with a single floodplain; and (c) decoupled catchment with two floodplains. The upper part of each diagram is a DEM of the coupling scenario (note the overall similarity of each of the three catchment forms), while the initial form of the outflow cross sections are shown for each case in the lower part.

floodplain has a channel-ward slope of 1° and a downstream slope of 1.07°. It makes up 7% of the DEM cells, with an average elevation change of 5.6 m in these cells compared to the single floodplain scenario.

In all cases, the model was run for 500 simulated years using the initial climate and vegetation conditions before the variable climate dataset was applied in order to allow it to adjust to the initial conditions and prevent them from having an undue impact on the results obtained. The 500-year start-up simulation allowed all simulations to start from conditions where transient variability in sediment transport had stabilized. The catchment size and DEM resolution were constrained by the need to carry out multiple model runs within a reasonable time frame. Typical run times using a fast PC (2.5 GHz Pentium with 512 Mb RAM) were of the order of 60–72 h to simulate the c.740 ka of the climate record used using an annual time step.

Climate scenarios were run with a modern baseline temperature of 12 °C with 1200 mm annual precipitation, 15 °C with 800 mm and 20 °C with 500 mm to represent humid, subhumid and Mediterranean-type conditions, respectively. In one set of model experiments, the temperature and precipitation changes were assumed to be synchronous (Fig. 4a and b). A second set of climate scenarios was also applied in which the precipitation change led the temperature change by 5 ka, representing the effects of feedbacks in climate change (for example, in terms of an ice-albedo effect).

The vegetation-biomass changes resulting from the climate scenarios are presented in Fig. 4c. The 20 °C–500 mm baseline scenario has the most constant biomass, while the two colder baseline scenarios often coincide after 400,000 simulated years (i.e. from ca. 340 ka bp). The 20 °C–500 mm baseline scenario produces the most dynamic changes, again especially after 400 ka of simulation, although the recent (equivalent to Holocene) values are relatively low (the EPICA data produce a Holocene that is apparently relatively cold when compared to Vostok and other ice-core records). The vegetation model is relatively insensitive to the asynchronous precipitation and temperature changes, because of the dominant effect of the temperature change on the simulated growth rate. The principal difference is that the asynchronous change scenarios produce slightly more peaky responses in the warmer, later parts of the simulations. Although no quantitative data are available for testing these changes, they are in broad agreement with changes in total arboreal pollen or vegetation associations as seen in a number of long European pollen cores (Mommersteeg et al., 1995; Pons and Reille, 1988; Pons et al., 1992; Tzedakis, 1994; Watts et al., 1996) and thus, there is a reasonable qualitative support for the vegetation changes that occur within the model experiments. However, the poor representation of vegetation in specific periods within the overall simulations does not affect the subsequent analysis, as direct predictions are not attempted. Analysis is restricted to relationships within the closed modelling system.

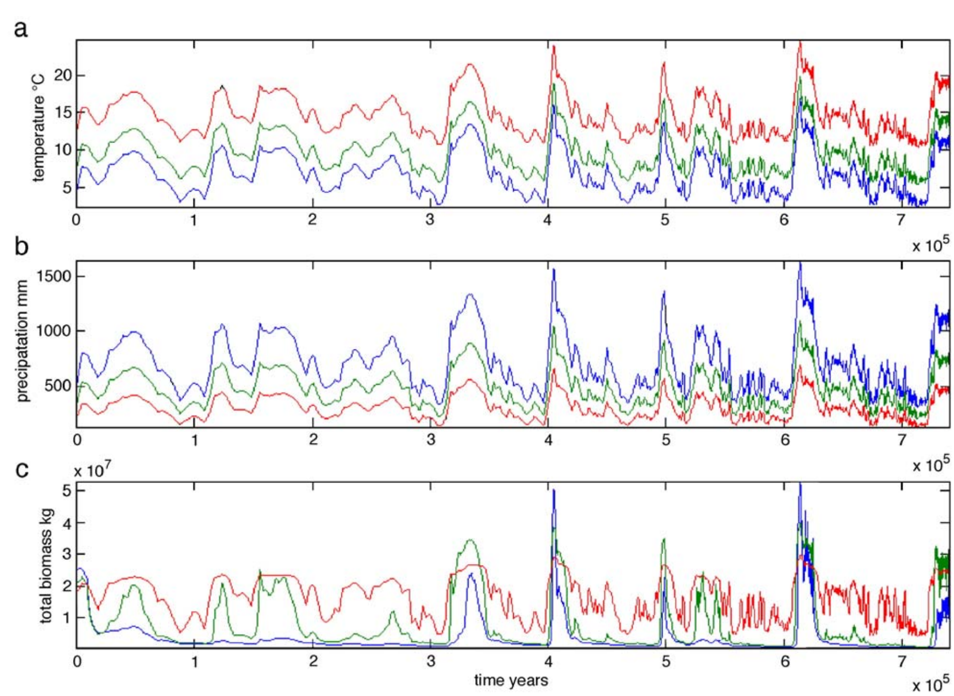

Fig. 4. Climate and vegetation scenarios used in the model experiments: (a) temperature changes relative to a modern baseline of 12, 15 or 20 °C; (b) annual precipitation change relative to a modern baseline of 1200, 800 or 500 mm, assuming synchronous precipitation and temperature changes; and (c) total simulated vegetation biomass over the DEM resulting from the climate scenarios with 12 °C–1200 mm, 15 °C–800 mm and 20 °C–500 mm baseline conditions.

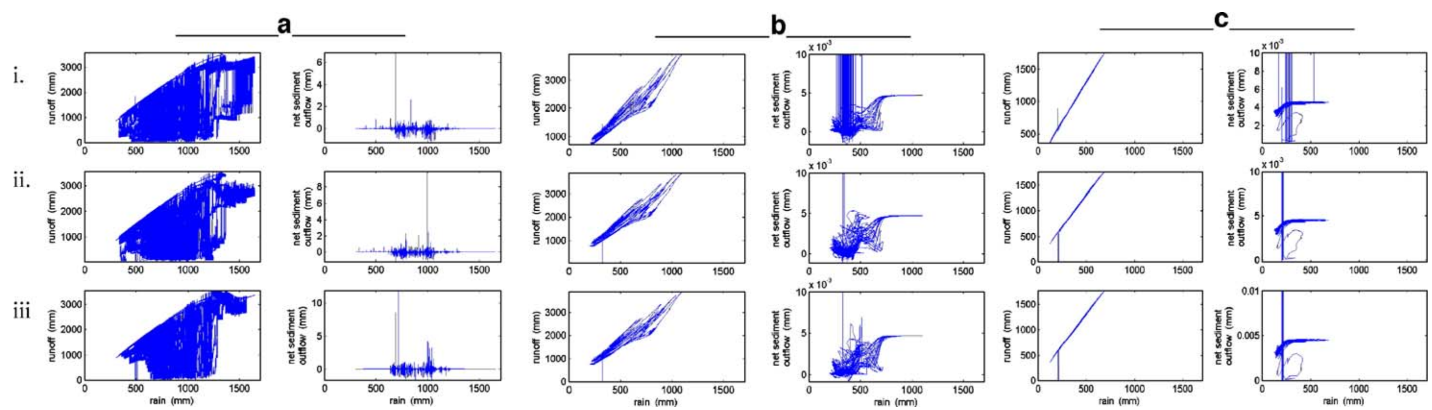

Fig. 5. Relationship between rainfall input and channel outputs (hydrograph and net sediment outflows) for the synchronous climate-change scenarios for all three coupling scenarios: (a) 12 °C – 1200 mm baseline climate conditions; (b) 15 °C – 800 mm baseline climate conditions; (c) 12 °C – 1200 mm baseline climate conditions; (i) initially coupled catchment; (ii) single floodplain catchment; (iii) double floodplain catchment.

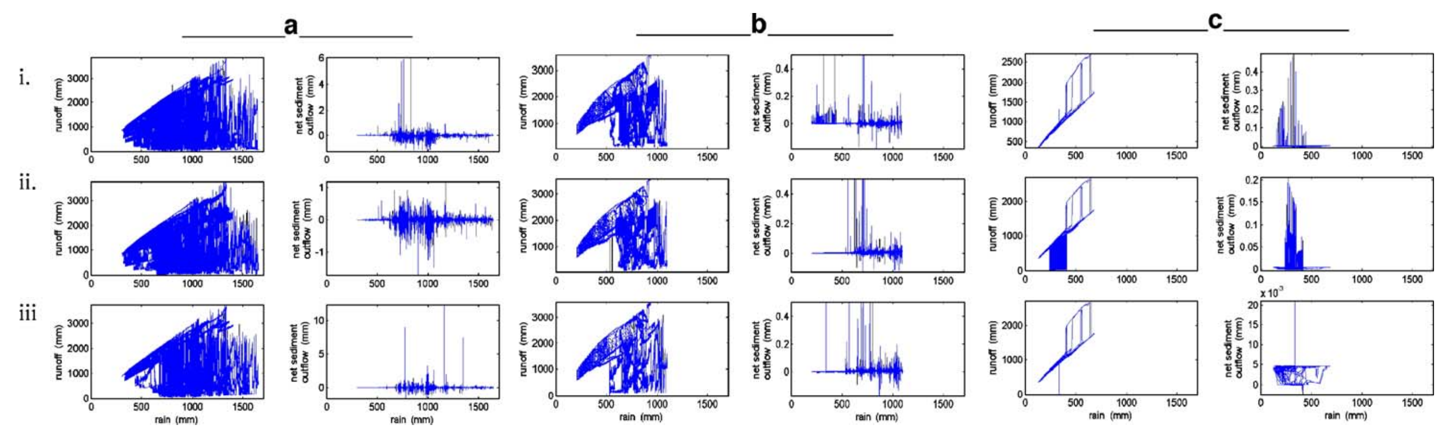

Fig. 6. Relationship between rainfall input and channel outputs (hydrograph and net sediment outflows) for the climate-change scenarios where precipitation changes lead temperature changes by 5 ka for all three coupling scenarios: (a) 12 °C – 1200 mm baseline climate conditions; (b) 15 °C – 800 mm baseline climate conditions; (c) 12 °C – 1200 mm baseline climate conditions; (i) initially coupled catchment; (ii) single floodplain catchment; (iii) double floodplain catchment.

4. Results

Because all calculations are made on a cell-by-cell basis, the net sediment outflow is the balance between erosion and deposition in the edge cells representing the channel. Negative values represent phases of aggradation while positive values represent downcutting of the channel, and as such represent the sedimentation characteristics of the catchment system at any point in time. The hydrological response of the system can be defined as the water outflow from these same edge cells. The catchment runoff and net sediment-outflow results are plotted as phase-space diagrams in Fig. 5 for the synchronous climate-change scenarios. There is a marked dominance of relatively linear relationships between rainfall input and catchment outflow, although only in the case of the 20 °C–500 mm scenario is there a straightforward relationship between the rainfall and runoff. With increasing humidity, the effects of vegetation become more important in controlling the complex relationships, and the historical sequence of catchment behaviour is increasingly important. There is a complex response in the net sediment outflow in all cases, with clear hysteresis in the 20 °C–500 mm and 15 °C–800 mm scenarios. In the 12 °C–1200 mm scenario, the sediment response is highly complex, with significant net aggradation or erosion possible across a wide range of rainfall rates. When the precipitation-lead scenarios are compared (Fig. 6), it can be seen that there is a complex set of relationships in all cases, with strong hysteresis in the rainfall–runoff relationships and increasing scatter with increasing humidity in the rainfall–net sediment outflow relationships. These results suggest that straightforward relationships derived from palaeohydrological data based on sedimentation are unlikely to provide the basis for good climate proxies. These results further support the complex response of catchment systems to climate change as found by the modelling analyses of Tucker and Slingerland (1997) and Coulthard et al. (2002, 2005; Coulthard and Macklin, 2001).

Time series of the catchment outflows for the synchronous climate-change scenarios for all three coupling scenarios are presented in Fig. 7. In the upper of each pair of figures, the hydrographs are plotted for each climate scenario, while the lower of the pair plots the net sediment outflow. In all cases, the hydrographs are relatively insensitive to the coupling scenario, being mainly controlled by the precipitation and vegetation changes. This result is controlled by the limited simulation of re-infiltration of water on floodplain elements in the current infiltration and flow-routing algorithm. By contrast, net sediment outflow varies significantly through the simulations in a more

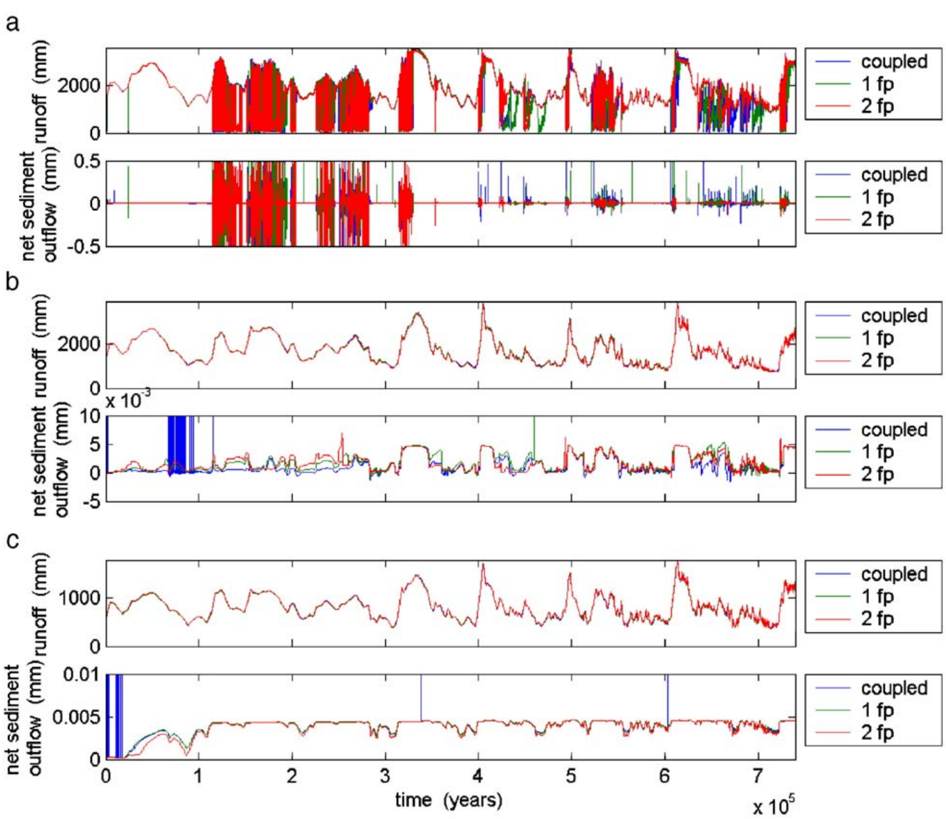

Fig. 7. Time series of model results for the synchronous climate-change scenarios for all three coupling scenarios: (a) hydrographs for 12 °C–1200 mm baseline conditions; (b) net sediment outflows (negative values reflect aggradation, positive values downcutting) for 12 °C–1200 mm baseline conditions; (c) hydrographs for 15 °C–800 mm baseline conditions; (d) net sediment outflows for 15 °C–800 mm baseline conditions; (e) hydrographs for 20 °C–500 mm baseline conditions; and (f) net sediment outflows for 20 °C–500 mm baseline conditions.

complex way. Peaks in the outflow occur particularly during periods of major, rapid climate change, although at different times for the different baseline climates. Similar links between sedimentation episodes and rapid climate change have also been observed in the sedimentary record (e.g. Maddy, 2002). In a number of cases, major oscillations occur, which relate to the generation of pulses of sediment within the catchment that are flushed through the system in a periodic way (e.g. as a sediment "slug"; Nicholas et al., 1995). The potential for channel aggradation is more marked in the cooler, wetter climate scenarios, and the 20 °C–500 mm scenario simulates no aggradation. The drier the climate, the less sensitive the output is to the coupling condition also. In the 15 °C–800 mm scenario, it can be seen that the initial coupling conditions are more important in the early part of the simulation, and that there is a convergence in behaviour after about 260 ka of simulation. This result is likely a consequence of the fact that as the catchments evolve, phases of incision and subsequent downcutting of the channel create decoupled conditions in all of the scenarios.

Fig. 8 presents the time series of the results for the climate-change scenarios where precipitation change precedes temperature change by 5 ka. The asynchronous climate changes tend to produce more dynamic and different runoff changes when compared to the synchronous time series. The variability in the sequences is also enhanced and even the 20 °C–500 mm scenario simulates aggradation. There is more simulated aggradation overall, especially in the later parts of the simulations. The decoupled scenario with two floodplains is the most dynamically changing, followed by the single floodplain case. This ordering is more marked than in the synchronous climate-change case, and probably reflects the dominance of channel processes where the hillslopes are decoupled from the channels. Perhaps because of this dynamism, there is also less convergence between the different coupled conditions through time.

The complex response of the simulated catchments is further demonstrated by considering the sediment delivery to the catchment outlet. Under conditions of synchronous climate change (Fig. 9), it can be seen that the total catchment erosion tends to be out of phase with periods when there is major activity–either in terms of aggradation or downcutting–at the catchment outlet. There seems to be a pattern of storage of sediment within the catchment that is then flushed out of the system during periods of more rapid climate change. The evolution of slopes is very punctual, and dominated by activity in colder and transitional periods. One possibly counterintuitive result is that the drier climate scenario produces less slope erosion overall, although the process understanding on which this interpretation is based may reflect the dominance of anthropic intervention producing accelerated erosion in such environments during the Holocene (e.g. Wainwright and Thornes, 2003), which is not simulated in this version of the model. Under conditions of asynchronous climate change (Fig. 10), it can be noted that the hillslope response is more peaked. In the drier

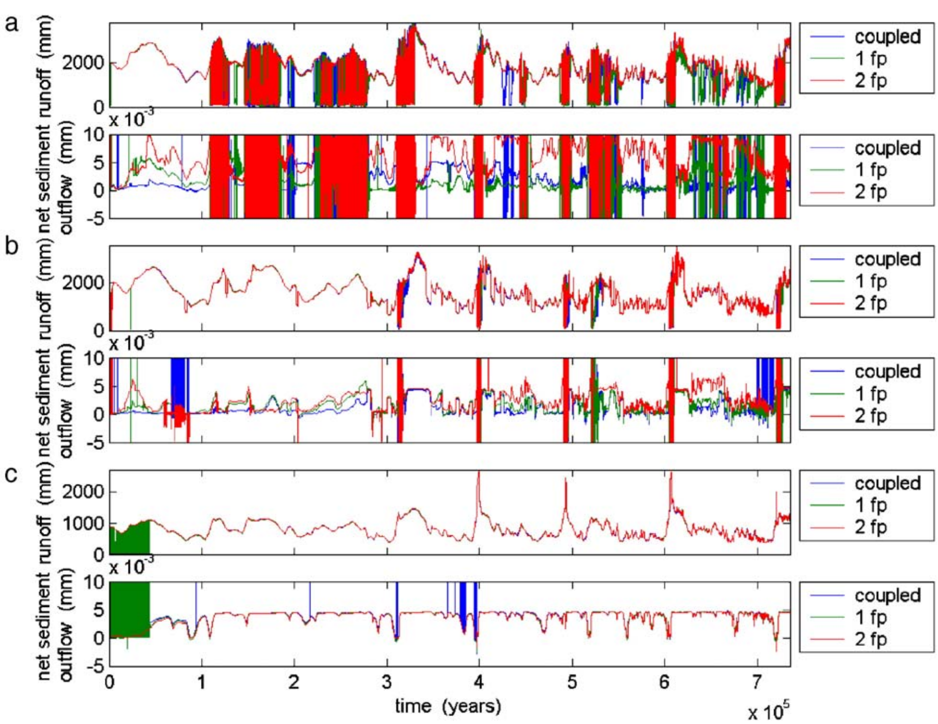

Fig. 8. Time series of model results for the climate-change scenarios where precipitation changes lead temperature changes by 5 ka for all three coupling scenarios: (a) hydrographs for 12 °C–1200 mm baseline conditions; (b) net sediment outflows (negative values reflect aggradation, positive values downcutting) for 12 °C–1200 mm baseline conditions; (c) hydrographs for 15 °C–800 mm baseline conditions; (d) net sediment outflows for 15 °C–800 mm baseline conditions; (e) hydrographs for 20 °C–500 mm baseline conditions; and (f) net sediment outflows for 20 °C–500 mm baseline conditions.

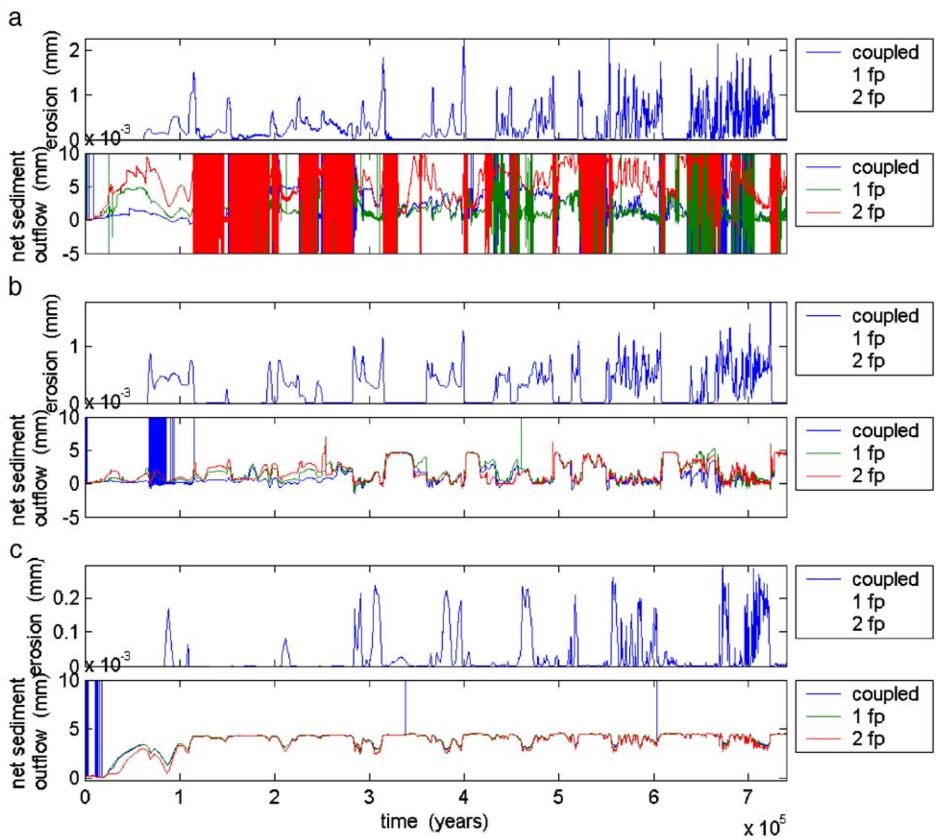

Fig. 9. Comparison of total catchment erosion and net sediment outflows for the synchronous climate-change scenarios for all three coupling scenarios: (a) total catchment erosion for 12 °C–1200 mm baseline conditions; (b) net sediment outflows (negative values reflect aggradation, positive values downcutting) for 12 °C–1200 mm baseline conditions; (c) total catchment erosion for 15 °C–800 mm baseline conditions; (d) net sediment outflows for 15 °C–800 mm baseline conditions; (e) total catchment erosion for 20 °C–500 mm baseline conditions; and (f) net sediment outflows for 20 °C–500 mm baseline conditions.

scenario, the responses tend still to be out of phase, while under the wetter scenarios, there is some coincidence of the accelerated responses both on the hillslopes and in the channel, particularly during periods where the climate is changing rapidly. In short, the complexity of the response of the system is enhanced under more complicated climate-change conditions.

5. Discussion and conclusions

The coupling of hillslopes, floodplains and channels within a catchment system produces equifinality in the form of the fluvial system and a divergence of system response to the same climatic input. Thus, in order to understand the meaning of palaeohydrological data at any point, it is necessary to have a clear idea of the catchment configuration upstream of that point. Following Church (2002), it may be feasible to consider the configuration in terms of broad thresholds of behaviour, in order to increase the confidence in interpretations made, given that it is unlikely that such detail will be reconstructable in all but the most recent (i.e. historical) catchments. Progressive evolution of catchment systems and destruction of such contextual information mean that confidence in interpretations decreases with age of deposits. The quality of any palaeohydrological proxy is likely to decline rapidly in increasingly old conditions unless a channel is strongly incising (Lewin and Macklin, 2003). However, incision will tend to lead to increasingly decoupled conditions with the consequences already outlined.

The complex response of fluvial systems has been analyzed from field data over the last 25 years. The modelling experiments carried out here suggest the complex response is a function at least of the extent of hillslope-channel coupling within the catchment system, storage within the floodplain and of the exact nature of climate change. The response changes dramatically even with relatively small changes in the catchment — the coupling changes represented in these scenarios represent changes to 15% and 7% of the catchment area, respectively, as one and then a second floodplain are added into the system. The modelling approach here has also been relatively conservative in terms of reflecting responses to coupling. Thus, even in the “best-case” scenario for evaluating the impacts of coupling, it can be seen that these impacts are substantial, non-linear and not necessarily predictable. Because the nature of the impacts is also a function of

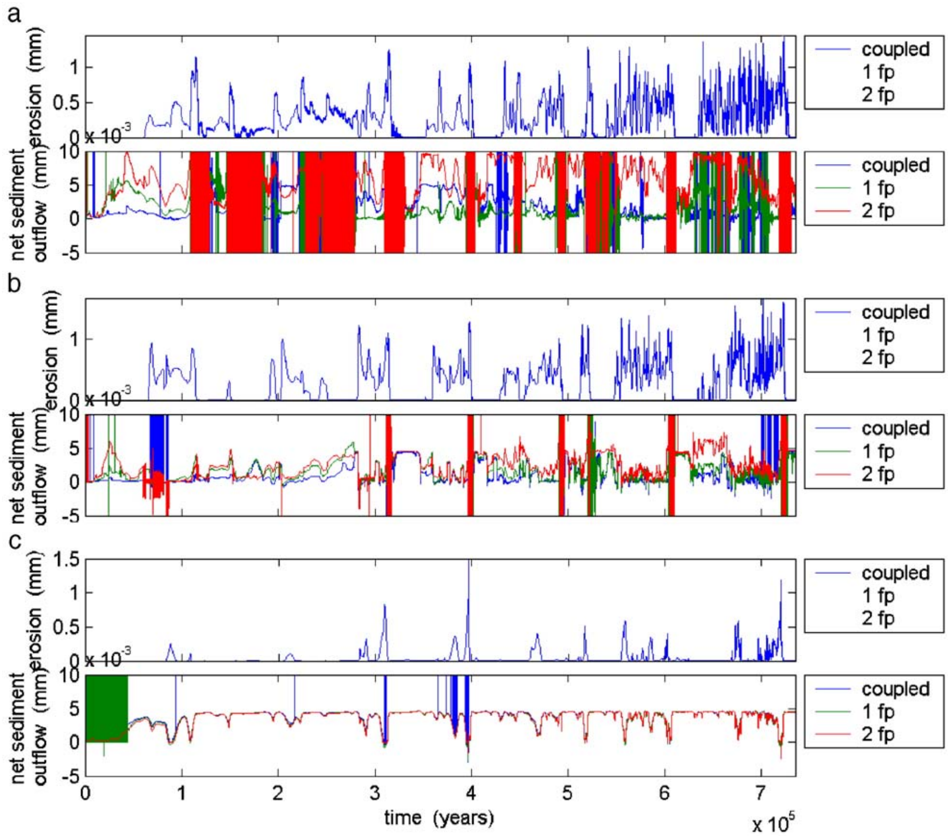

Fig. 10. Comparison of total catchment erosion and net sediment outflows for the climate-change scenarios where precipitation changes lead temperature changes by 5 ka for all three coupling scenarios: (a) total catchment erosion for 12 °C–1200 mm baseline conditions; (b) net sediment outflows (negative values reflect aggradation, positive values downcutting) for 12 °C–1200 mm baseline conditions; (c) total catchment erosion for 15 °C–800 mm baseline conditions; (d) net sediment outflows for 15 °C–800 mm baseline conditions; (e) total catchment erosion for 20 °C–500 mm baseline conditions; and (f) net sediment outflows for 20 °C–500 mm baseline conditions.

the detail of the climatic change, there is an inherent circularity in trying to generate climate proxies from palaeohydrological data. Thus, the "supplying of data" rôle of palaeohydrology is limited and the reconstruction of climate-change parameters from sediment data is likely to be extremely limited.

Baker's (1998) suggestion of the use of palaeohydrological data as semiotic indicators of the qualitative behaviour of catchments may be more supportable. However, caution is still required because the complex response means that even qualitative assessments are difficult and many changes may have counter-intuitive causes as a result of a focus on normative process-based studies based on limited recent information. Unfortunately, the use of palaeohydrological data to try to extend these studies is limited because of the circularity of argument that would be engendered in the linkage between cause and effect. Further caution is required because complex response and the sensitivity to initial conditions (in the sense of the initial coupling characteristics of the catchments) demonstrated by many of the model experiments here suggest that it is fundamental to be able to specify the past trajectory of a system in order to interpret further change. Again, this implication involves a significant degree of circularity, although iterative approaches (such as the abductive approaches suggested by Baker, 1998) may allow this circle to be broken.

Many of the model experiments suggest that "events" might be recorded in the fluvial record, but these events are not necessarily those that would be anticipated, and vary according to initial conditions and system trajectory. Phases of rapid change rather than of stable conditions seem to be most frequently represented. This result is in direct contrast with much interpretation of palaeohydrological records, which often talks of phases of wetter or drier conditions. Complex response – whether modelled or interpreted from field data–implies that the conceptual underpinnings of such statements need serious re-evaluation. This particular model suggests the response is more complex if temperature and precipitation changes are not synchronous. Asynchronous changes affect the magnitude, timing and occurrence of phases of enhanced depositional activity. Thus, what is recorded in the fluvial record will mean different things depending on the nature of the climate change, but again this is exactly what research programmes to develop palaeohydrological proxies are trying to evaluate. Complex response implies that such interpretations are not possible without recourse to external information about the nature of the climate driving the changes. Simple palaeohydrological proxies are therefore likely to provide reconstructions that are simplistic at best, and incorrect in many cases. It is possible that approaches based on reconstructions of the phase space of the system may have advantages, as the form of the response of the system seems to be a function of internal form and external dynamics (Figs. 5 and 6), but this approach is limited by the low preservation potential of fluvial systems.

The model experiments carried out in this study are intended as a heuristic device. The results are not intended to be predictive, and they should not be used to interpret specific cases. Their value lies in allowing us to extend assessments of complex interrelationships between the behaviour of a number of variables within a complex system. Such assessments are only rarely possible from field data, and often only in relatively simple contexts. This statement is not intended to decry the value of field data, only to suggest that when dealing with such complex systems, interpretations must be based on sound principles of how such systems work. The model here is “wrong” in the sense that it only simulates part of the system interactions, but it has been simplified deliberately in a way that will reduce the impact of hillslope-channel coupling in order to demonstrate that even in conservative evaluations, the impact is likely to be significant. It also suffers from using some of the same normative understandings of process criticised above. However, the generation of complex responses from simplifications of the process relationships again only serves to emphasize the impact on the system behaviour and output. The exact form of the catchment-evolution model used is in fact unimportant. For example, the model of Coulthard et al. (2002) when applied to actual catchments produces complex evolution results. In a number of cases, these are not incompatible with observed reconstructions of the behaviour of the River Swale system (Coulthard and Macklin, 2001). The models of Willgoose et al. (1991) and Rodriguez Iturbe and Rinaldo (1996) show complex behaviour and sensitivity to initial conditions that would also support the general conclusions being made here.

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