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Stability of output fluxes of large rivers in South and East Asia during the last 2 million years: implications on floodplain processes

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ABSTRACT

We compare the present-day sediment discharge (solid phase) of some of the largest rivers in Asia to the average discharge deduced from the mass accumulated in several sedimentary basins during the Quaternary. There is a very good correlation, especially for the largest rivers: the Ganges–Brahmaputra, the Changjiang, the Huanghe and, to a lesser extent, the Indus and the Zhujiang. This suggests that present-day average discharge at the outlet has remained constant throughout the Quaternary at least for very large rivers (drainage area of the order of 10^5–10^6 km^2). This, in turn, suggests either that continental denudation of large Asian catchments has remained on average constant, implying a strong tectonic control on erosion during the Quaternary, or that the river network has the ability to buffer changes in hillslope erosion or in sea-level in order to conserve the total discharge at the outlet. We show how this buffering capacity relies on the characteristic reaction time-scale of Asian alluvial plains (of the order of 10^{5-6} years), that is, much higher than the time-scales of the Quaternary climate oscillations (of the order of 10^4 years). A short-term perturbation originating in hillslopes will be diluted by the floodplain. At the outlet the signal should have a longer time span and a smaller amplitude. In the same manner, an alluvial plain should not instantaneously react to a 10^4-year sea-level drop because of its inertia. Along with long-term tectonic control we infer this buffering to be the main cause for the average constancy of sediment yield of large Asian rivers during the Quaternary.

INTRODUCTION

The purpose of this paper is to look at the correlation possibly existing between long-term and short-term mass fluxes carried to the sea by large Asian rivers and to discuss the possible implications of such a correlation. Because rivers are the main carriers of erosion products from the continents to the oceans, measuring their sedimentary load provides invaluable constraints on the rate of continental denudation. There is, however, a problem in that these measurements cover a very short time-span, at most a few decades, while erosion processes operating in a basin have very different time constants. Some are almost instantaneous, like avalanches (velocity of the order of 10^0 to 10^2 m s^{-1}) while others, like soil creep (velocity of the order of 10^{-10} to 10^{-12} m s^{-1}), may be too slow to be observed (Goudie, 1995). This raises the following question: how representative of long-term average denudation rates are present-day measurements of solid loads in rivers? And what information does this bear on upstream processes and transport capacity of the rivers?

To address the relevance of long-term denudation rates deduced from sediment accumulation, Granger *et al.* (1996) dated sands in rivers using cosmogenic isotopes. They then deduced average denudation rates on the scale of 10^4 years and compared them to sediment accumulation. Their results show a good fit, indicating the relevance of using average denudation rates on that time scale. Lavé (1997) derived fluvial incision rates in the Himalayas from the study of strath terraces. He extrapolated his results to the entire range and compared the results to sediment volumes accumulated in the Ganga plain and the Bengal fan (Métivier, 1996; Métivier *et al.*, 1999). He showed a good average correlation between the two results, suggesting that short-term fluvial incision rates are comparable to long-term rates deduced from sedimentary volumes. Recently, Burbank *et al.* (1998), conducting a similar study in the San Gabriel mountains, found also good agreement between rates of denudation at different time-scales. This correlation, however, does not appear to be unique. Kirchner *et al.* (1998), comparing sediment yield and denudation rates at the 10^4-year scale, found that, for small to median size catchments (less than 5×10^4 km^2), the suspended load was systematically lower than average denudation rates. They attribute it to the fact that their record is too short to incorporate extreme erosion events like exceptional landslides.

To tackle this important problem of present-day and long-term average fluxes of sediment, we have compared the present-day solid loads, Q_p, of several large rivers of Asia (catchment area greater than 10^5 km^2) with the filling rate, Q_f, of the sedimentary basins fed by the same rivers averaged over the last two million years (Fig. 1). This time span approximately corresponds to the Quaternary Era for which data are available from drill logging in almost every Cenozoic basin of Asia (Métivier *et al.*, 1999). We show that the correlation we obtain suggests that rates of mass transport to the sea by large Asian rivers has remained, on average, a constant. We furthermore propose this constancy to be explained either by sustained erosion rates in the catchments, that would mainly be controlled by high tectonic uplift rates, or by buffering of variations of the sediment influx from the catchment by the river floodplain. We illustrate this latter possibility by considering a 10^4-year sea-level drop and by looking at the possible effects on the catchments and on erosion fluxes.

DATA ANALYSIS

The average filling rates (Q_f in Table 1) for the last two million years were derived from a study of mass accumulation in the sedimentary basins of Asia (Métivier *et al.*, 1999). Local depths and thickness of sediments measured on isopach maps, cross-sections and drill-logs, either published or provided by TOTAL-CFP company, were interpolated, then integrated over the entire depositional areas in order to provide volumes of sediments deposited in the basins of Asia for several periods covering the Cenozoic. The volume estimates are fraught with an uncertainty of the order of 35–40%. A complete description of the method and associated errors is given in Métivier & Gaudemer (1997) and Métivier *et al.* (1998, 1999). It is at present impossible to achieve a time resolution smaller than about 2 million years because of the lack of reliable dating of recent coarse-grained clastic sediments for all Asian basins.

Suspended load values (Q_p in Table 1) were gathered from several published sources (Milliman & Meade, 1983; Milliman & Syvitski, 1992; Summerfield & Hulton, 1994; Meybeck & Ragu, 1995). Sampling rates are usually of the order of one sample each day over a period on order of 1–10 years. Sampling techniques vary between precise sampling – at different depths on several vertical profiles across the river section as, for example, in the case for the Indus river – and one point sampling (Fournier, 1960). In the last case the results are extrapolated to the entire river bed assuming homogeneous turbulent mixing of the suspended load (Fournier, 1960). The data given are generally in good agreement from one publication to the other. When they differ markedly, we selected the most recently published value.

These reports are incomplete however. There is, for instance, no value for the Salween river that drains part of SE Tibet and Burmese mountains before flowing into the Andaman sea. Similarly, there is no value for several albeit small (drainage area $A \leq 10^5$ km^2) rivers that drain the Makran range in southern Iran and Pakistan and bring sediments to the Indus fan in the Arabian Sea. For the Huang He (Yellow River), we take the value of 100 million tons per year (hereafter t year^{-1}) given by Milliman *et al.* (1987), which is corrected for the anthropogenic activity, namely agriculture on the loess plateau in the last 2000 years (Wang *et al.*, 1998).

Another source of uncertainty is the lack of reliable estimates of bed load transport as no precise measurement technique is available at present. Clearly, bed load can be important in the sediment budget as has been shown by more or less empirical relationships between bed load transport and discharge or stream power (Ashmore, 1988). In the absence of any applicable relationship or measurements we will assume a large uncertainty similar to the Q_f uncertainty (i.e. 40%).

We did not include solute load transport for two reasons. First, the sediments deposited in the basins of Asia are almost exclusively composed of clastics transported as solids by the river systems (Métivier, 1996). Second, apart from the Changjiang, the contribution of solute transport appears to be largely negligible and in any case remains within the 40% error bars we assume (Summerfield & Hulton, 1994).

Finally, a very important point is that the Quaternary infill of the basins we study is not only composed of the fine portion of sediments that leave a catchment, the so-called washload made of particles smaller than 0.2 mm in size. One observes the presence of clays, sands, gravels and even conglomerates in large amounts in all the basins studied (Métivier, 1996; Wang *et al.*, 1998; Métivier *et al.*, 1999), and especially in their proximal and thickest parts where coarse material can account for most of the sediment pile. Therefore in the rivers we study the coarse fraction of erosion, that is the size fraction produced at the outcrops and stored in the floodplains and alluvial fans, reaches, in some proportion, the sedimentary basins through the fluvial system. Analysis of these river basins should therefore allow us to discuss floodplain processes that involve both the suspended and bedload components of sediment transport.

RESULTS AND DISCUSSION

Equality of the present-day and Quaternary average fluxes

Comparing the average discharge deduced from basin filling, Q_f, with the corresponding present-day discharge,

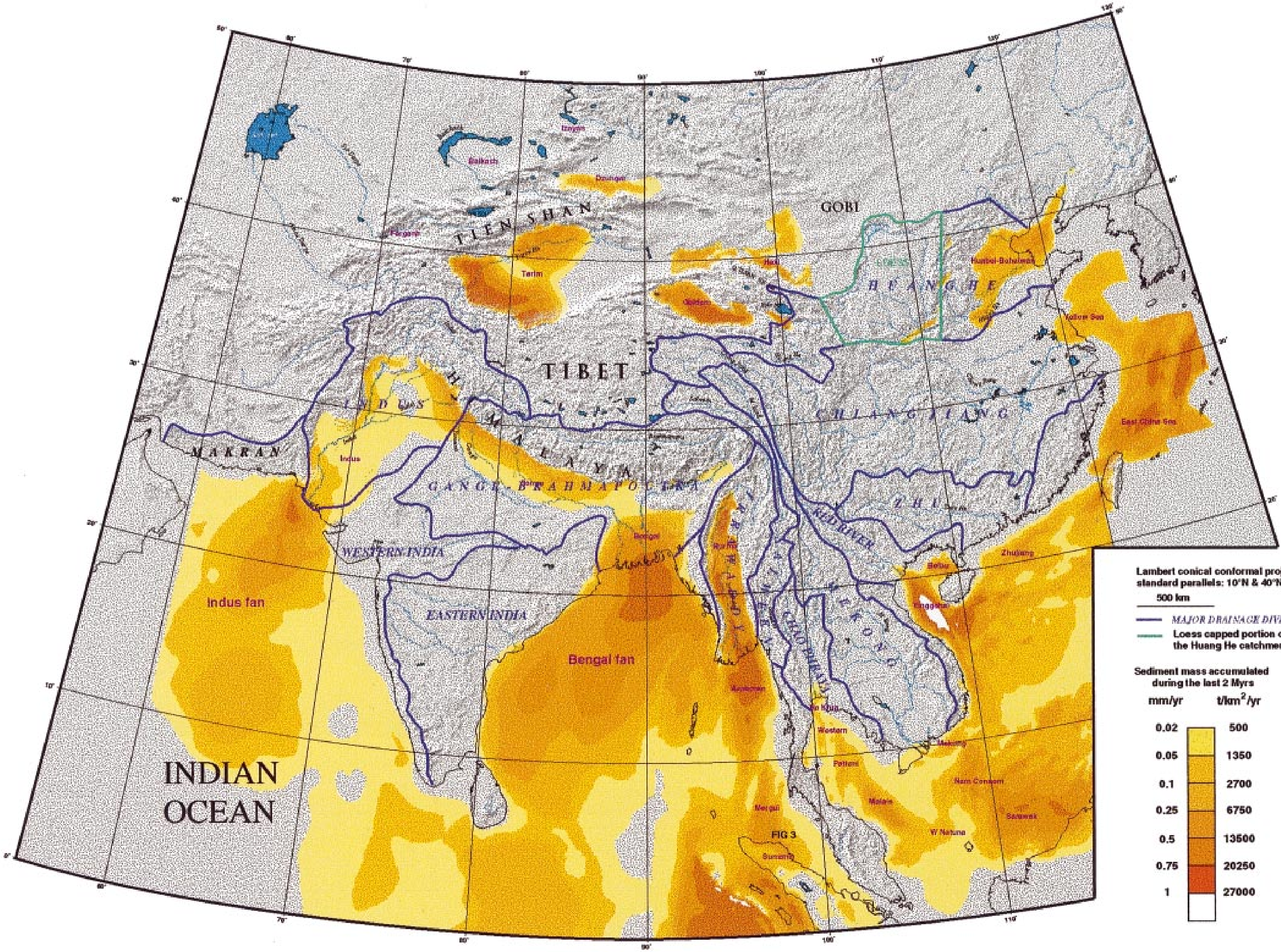

Fig. 1. Average accumulation rates $q_f(x, y)$ (t km^{-2} yr^{-1}) in Asian basins during the last 2 million years (Métivier, 1996; Métivier *et al.*, 1999). Regional rates (over area S) were derived from this database following $Q_f = \iint_S q_f(x, y)\mathrm{d}x\,\mathrm{d}y$.

River	Q_p (10^6 t yr^{-1})	Q_f (10^6 t yr^{-1})	Deposition area
Ganges	520		
Brahmaputra	540		
Godavari	170		
Krishna	64		
Mahanadl	60		
Damodar	28		
Brahmani	20		
Total Ganges	1402	1285	Bengal fan
Changjiang	480	419	East China Sea + Okinawa
Irrawaddy	260	250	Burma + Andaman Sea
Indus	250		
Narmada	125		
Mahi	10		
Total Indus	385	475	Indus Fan
Mekong	150	75	Mekong + Nam Consom
Honghe	130	188	Yinggehai
Huanghe	100	94	Hubei + Bohai + Subei
Zhujiang	69	52	Zhujiang
Chao Phrya	11	15	Gulf of Thailand

Table 1. Comparison between the present-day solid load, Q_p, of major rivers in Asia and the average filling rate, Q_f, of corresponding sedimentary basins. t yr^{-1}: metric tons per year.

Q_p, measured in the rivers that feed the basin, three cases are possible: (1) the fluvial sediment discharge is equal to the average sedimentary basin filling rate, (2) the present-day discharge is greater than the average discharge needed to fill the depression and (3) the present-day sediment discharge is not sufficient to account for the Quaternary infill of the basin.

Because mass is conserved, we can sum the discharges of several rivers and compare the resulting value with the mass accumulated in a basin fed by these river systems. This is the case for the Ganges–Brahmaputra and the Krishna–Godavari river systems that together flow into the Bay of Bengal. Likewise, we summed the measurements available for the Indus river and the rivers that drain the western Ghats forming the submarine Indus fan in the Indian ocean (see Table 1).

The overall correlation is good (Fig. 2) which means that present-day Asian river systems discharge their sediment at the average Quaternary rate deduced from basin infilling. Most important is the excellent correlation obtained for three major river systems of Asia: the Ganges–Brahmaputra (and to a much lesser extent the Godavari), the Changjiang (Blue river), the Huang He (Yellow River). There is also a good agreement (within the error bars) for the Indus combined with the Narmada river that drains part of Western India, and the Zhujiang rivers.

This striking correlation concerns rivers that have drainage areas that together extend over 30 degrees of latitude, under variable climatic conditions. It therefore suggests that these very large rivers (with catchments of the order of 10^6 km^2), have, on average, maintained the present-day output fluxes throughout the last two million years, a very important and constraining boundary condition that, as we shall argue, may have implications on the upstream mechanics of the floodplain.

A clear inconsistency between the average discharge and the present-day discharge is observed for the Mekong river, while no conclusion can be drawn for the Red River (Hong He) and the Chao Phrya (Thailand) for which the correlation is poor, though still within the error bars (Fig. 2). Also, in the case of the Irrawaddy no conclusion can be drawn because of the absence of any reliable data concerning the carrying capacity of the Salween river (which also carries its load to the Andaman Sea).

In the case of the Mekong river, the present-day solid load is much larger than the average filling rate. If we still assume, as suggested by the results obtained for the other large fluvial systems of Asia, that the present-day load of the Mekong river holds over the last two million years, then the river must have discharged its load elsewhere. Looking for possible sediment troughs, the Malay basin appears as a good candidate. It is fed by no river at present and is bordered by the Quaternary Mekong floodplain and delta, as shown by the stratigraphic evidence (Métivier *et al.*, 1999, and references therein). Bank erosion of several metres per year is quite common in alluvial floodplains (Goudie, 1995). Since the Mekong delta is very flat (slope of the order of 10^{-4}), a shift of 100 km could occur in a short time (less than 100 kyr). In this case the combined mass of sediments accumulated in the Malay, Mekong and Namconsom basins during the Quaternary should be equal to the present-day mass flux of the river. This is the case within the error bars, as can be seen from Table 2.

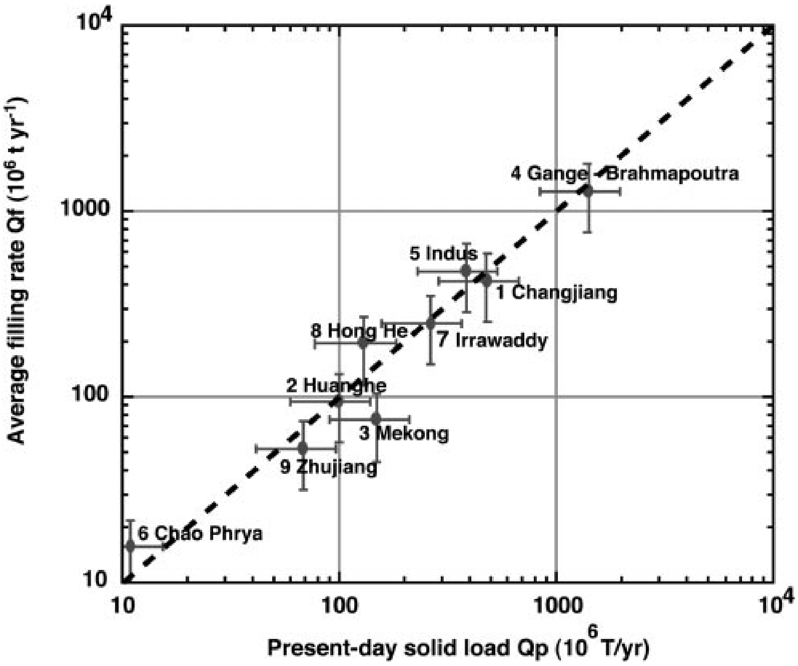

Fig. 2. Plot of average present-day load Q_p of great rivers in Asia versus average filling rate Q_f of corresponding sedimentary basins during the last 2 Myr (data in Table 1). Values are given in t yr^{-1}. 1, Changjiang; 2, Huanghe; 3, Mekong; 4, Ganges–Brahmaputra and East-Indian rivers; 5, Indus and west-Indian rivers; 6, Chao Phyra; 7, Irrawaddy; 8, Hong He; 9, Zhujiang.

Table 2. Comparison between the present-day solid load, Q_p, of the Mekong river and the average filling rate, Q_f, of different sedimentary basins

River	Q_p (10^6 t yr^{-1})	Q_f (10^6 t yr^{-1})	Deposition area
1 Mekong	150 ± 60	75 ± 30	Mekong and Nam Consom
2 Mekong	150 ± 60	125 ± 50	Mekong, Nam Consom and Malay

DISCUSSION

Robustness of the correlation

One could argue that the correlation we show is 'pure chance' and linked to the fact that the interglacial present-day climatic conditions represent some kind of average of the 2-million-year period. Both measures of average denudation rates over 10^4-year periods derived from cosmogenic dating of sediments (Granger *et al.*, 1996) and measures of short-term present-day fluxes in catchments of the San Gabriel mountains (California), and the Himalayas (Burbank *et al.*, 1998) contradict this inference. Cosmogenic dating of sands shows that denudation rates on time-scales of the order of the last glacial cycle are coherent with the volumes of sediment stored at the outlet of small drainages of the American west (Granger *et al.*, 1996); furthermore, Lavé (1997) and Burbank *et al.* (1998) also show that despite changes in climatic conditions, long-term denudation rates are closely correlated to present-day fluvial incision rates.

Only in the case of the Huang He is the anthropogenic influence known (Milliman *et al.*, 1987) and taken into account. In this case it leads to a reduction of the sediment yield by approximately an order of magnitude. The Huang He flows from the high Tibetan plateau into the Ordos grabens and then into Hubei and the China Sea. During more than half of its course it flows through highly erodible loess-capped landscapes (Fig. 1). Settlements, agriculture and deforestation, all along the river course in these loess regions, has led to a dramatic increase of the sediment yield of the river (Milliman *et al.*, 1987; Wang *et al.*, 1998). It can safely be considered an exception because nowhere else in Asia does one encounter the same geological conditions where highly erodible and unstable material is available in large amounts (thicknesses of loess deposits commonly reach 100 m and much more on the loess plateau), over large areas. The Indus, Ganges and Brahmaputra rivers, for example, have their upstream catchments flowing on Himalayan rocks that are much less susceptible to anthropogenic influence. Furthermore, as stated before, orders of magnitude of the average sediment yield during the Quaternary are comparable to the 10^4-year average erosion rates deduced from terrace incision of Himalayan rivers (Lavé, 1997) when anthropogenic factors were negligible. Thus, although human influence should be taken into account, we think that it does not change the orders of magnitude discussed here. As a reminder, had the Huang He not been corrected for anthropogenic influence it would clearly stand out from the correlation line. Hence we think that 'profoundly' anthropogenized rivers should be expected to have much higher sediment yields.

We may therefore reasonably suggest that large river sediment fluxes in Asia have remained, on average, constant throughout the Quaternary. Possible causes for this are (1) very active tectonic uplift in Asia may be a dominant control on erosion rates or (2) large drainage areas and floodplains have some strong smoothing effect that averages spatial and temporal variability inside the catchments.

Tectonic control

Our correlation may reflect the effect of tectonic uplift rates in Asia. All the rivers we studied have a significant portion of their catchments undergoing active uplift at present. Furthermore, these areas of active uplift correspond to the highest parts of the catchment where denudation rates are expected to be the highest. Average uplift rates are of the order of 1–10 mm yr^{-1} (Lavé, 1997; Tapponnier *et al.*, 1990; Avouac *et al.*, 1993; Jackson & Bilham, 1994; Meyer *et al.*, 1998). Burbank *et al.* (1996) have shown that continuous and important incision of the Indus river (discharge up to 3000 $m^3\ s^{-1}$), may have been controlled by tectonics of the Nanga Parbat Haramosh massifs. They also show that incision rates may have kept pace with very rapid uplift rates (up to 12 mm yr^{-1}) during 0.5 and 1 Myr.

Figure 3 shows the Huang He river as it crosses the Tiangjin Shan range, an actively uplifting mountain range in north-western China. Uplift in this range is controlled by movement along the left lateral Haiyuan fault that accommodates eastward movement of NE Tibet (Gaudemer *et al.*, 1995; Lasserre *et al.*, 1999). Inside the range the river has a narrow course leaving strips of cut terraces inside meanders as remnants of its past elevation during the Quaternary. The absence of paired terraces attests to the impossibility of the river to widen its bed during high-stage and wetter periods because of rapid uplift. South of the uplifting mountain (south of the SPOT images shown on Fig. 3), large (kilometre-wide), paired terraces can be followed for several tens of kilometres. Just a few kilometres north of the thrust front that bounds the range the rivers starts to braid. This shows that as the river is left free to use its stream power it naturally expands its width. It represents a very nice example of the control active tectonics may exert on a large river such as the Huang He (present-day maximum discharge of the order of 1500 $m^3\ s^{-1}$ at Lanzhou $\sim$100 km to the south). According to neotectonic studies uplift there is of the order of $\sim$1–2 mm yr^{-1} (Gaudemer *et al.*, 1995). We therefore infer that most of the catchments that have uplift rates of the order of a few millimetres per year or higher may directly control incision of large rivers, and even more of their tributaries. Other examples of such control may be found upstream in the catchment of the Huang He (Van der Woerd, 1998), that attest to such a tectonic control on the Yellow River incision and erosion rates in this area of NE Tibet.

In conclusion, it appears that large-scale rapid uplift of Asian catchments may be one major reason for the constancy of the sediment yield carried by large rivers to the ocean.

Smoothing effects of the floodplain

Another possible explanation for the constancy of Quaternary mass fluxes to the sea may be found in the ability of alluvial plains, as they leave the tectonically active highlands of Asia, to exert a control on the sediment discharge to the sea. What are then the implications on the coupling between floodplain storage and climate conditions during glacial and interglacial periods? The alluvial floodplain is the place where a river temporarily stores sediment when stream power is not sufficient to carry away the erosion products coming from the slopes. This can be due either to a rise in the flux of eroded material while the stream power remains constant or to a reduction of the stream power. Paired cut terraces attest to this variation in sediment storage of the river. We suggest that if climate changes exert some control on hillslope processes upstream then conservation of sediment fluxes at the outlets implies that the river floodplain adjusts proportionally to these changes. Thus, if hillslope erosion is reduced, then the river may incise its alluvial plain in order to keep the sediment discharge constant at the outlet. The floodplain reservoir is then emptied. On the contrary, if hillslope processes are more vigorous, then the river will use this supplementary load to recharge its alluvial plain. The floodplain reservoir thereby increases in volume. The river system could therefore be defined as a buffer that smoothes climate-controlled variations of the hillslope fluxes. This point has been addressed by Humphrey & Heller (1995) who have shown how a floodplain could react to a perturbation in the catchment by oscillations of the river bed, thereby causing cyclic erosion and sedimentation.

Climate change plays a role through different mechanisms among which are changes in the volume of liquid water and type and amount of precipitation. Changes in the volume of liquid water partly control sea-level variations (Chappel & Shackelton, 1986) and glacier extent. Changes in precipitation form (rainfall or snowfall) induce changes in kinetic energy available for splash erosion of soils (e.g. Ellison, 1944; Al-Durrah & Bradford, 1982), whereas change in precipitation height controls runoff and discharge variability (Rodriguez-Iturbe & Rinaldo, 1997).

Syvistki & Morehead (1999) have suggested that despite changes in elevation, catchment size and precipitation, the average sediment yield carried by the Eel river (northern California) during the Last Glacial Maximum, 18 kyr BP (Chappel & Shackelton, 1986; Bigg, 1996) to the sea was of the same order of magnitude as the present-day fluxes (differences $\leq$20%). It is therefore not straightforward to decide whether or not change in climate conditions will exert a simple linear influence on sediment yields at the outlet of a catchment, although mechanical abrasion, rain splash and weathering are, at least in part, controlled by climate (e.g. Howard *et al.*, 1994; Goudie, 1995).

Hereafter we show that alluvial floodplains to the first

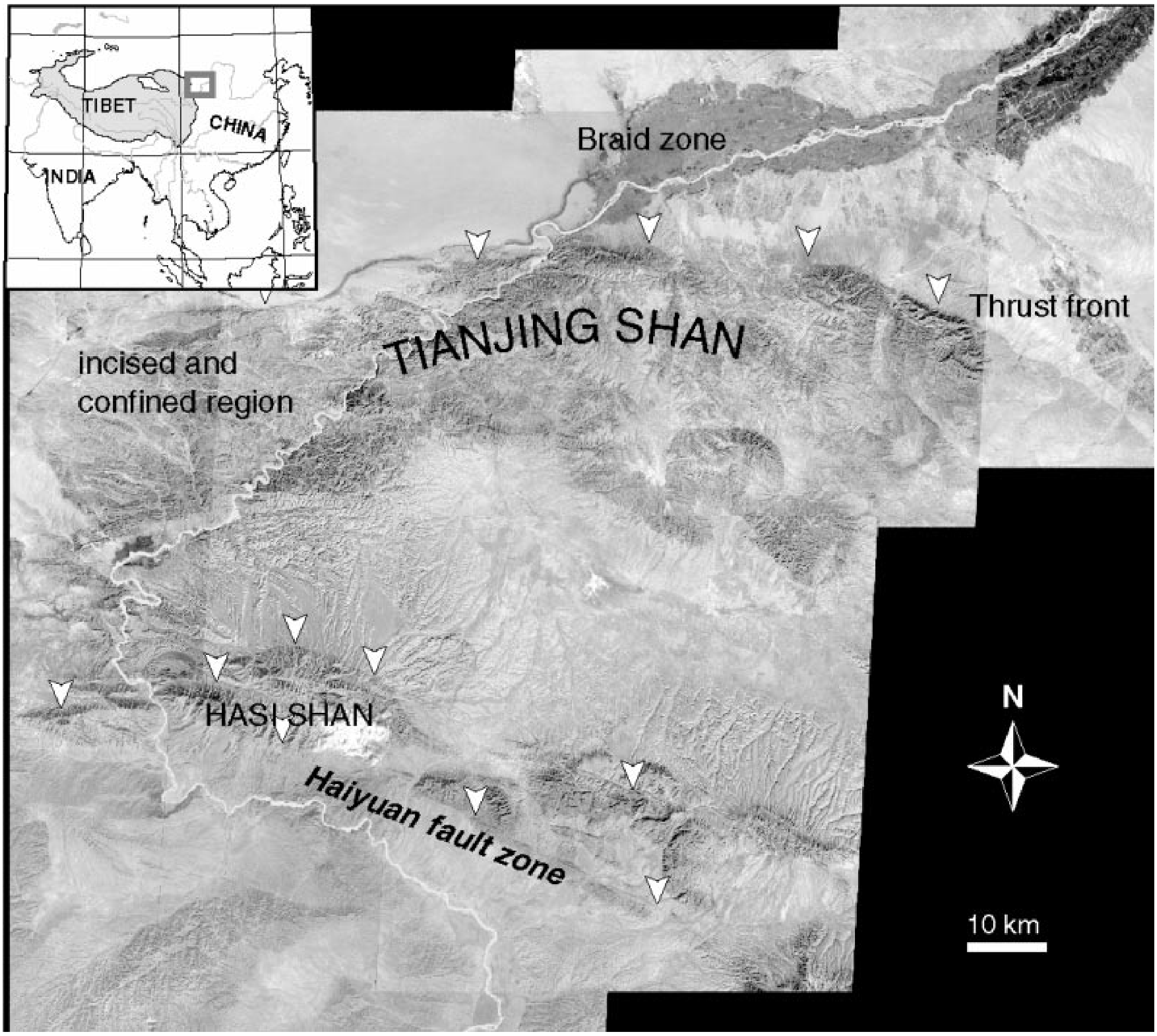

Fig. 3. Mosaic of SPOT satellite images (pixel resolution of 10 or 20 m depending on the type of image). White arrows outline active fault traces. Note sharp contrast of the river configuration to the south (incised), and north of the Tiangjing Shan thrust front (braided stream). All the glossy, grey and incised portions of the image are capped with loess.

order may be considered as diffusive systems, and have characteristic time-scales on average much larger than climate-induced variations. This order of magnitude difference in characteristic time-scales induces long-term buffering of short-term signals and reduces the amplitude of changes either in hillslope erosion or in sea-level change.

Characteristic time scale of a large alluvial floodplain: its influence on diffusive like buffering processes. It has long been shown that floodplains can to the first order be approximated as diffusive-like systems, that is systems where mass diffusion or transport is driven by bed slope (Soni, 1981; Paola *et al.*, 1992; Humphrey & Heller, 1995; Graf & Altinakar, 1996; Dade & Friend, 1998). Simplification of hydrodynamic and mass conservation equations leads to

$$Q_f \propto -vw\frac{\partial z}{\partial s} \qquad (1)$$

where Q_f is the sediment yield, s (m) defines the distance along the river bed, w (m) defines the floodplain width and v ($m^2\,s^{-1}$) defines the mass diffusivity of the alluvial system. When the river approaches steady conditions, the mass diffusivity of the system scales with the average plain gradient, width, and sediment yield, as

$$v \sim \frac{Q_f}{w\langle \partial Z/\partial s\rangle} \qquad (2)$$

where $\langle\ \rangle$ denotes spatial average (Humphrey & Heller, 1995). The characteristic reaction time (τ_r) of a floodplain scales as

$$\tau_r \sim \frac{L^2}{\nu} \sim \frac{L^2 w \langle \partial Z/\partial s \rangle}{Q_f} \sim \frac{LwH}{Q_f} \quad (3)$$

where H is the elevation of the floodplain at its upper end and $\langle \partial Z/\partial s \rangle \sim H/L$ (if the floodplain does not end at the sea then H stands for the maximum relief between the upstream and downstream ends of the plain). For the Asian rivers orders of magnitude commonly are $L \sim 10^6$, $w \sim 10^5$ m, $\langle \partial Z/\partial s \rangle \sim 10^{-3}$ to 10^{-4} ($H \sim 1\text{–}2 \times 10^2$ m), and $Q_f \sim 10^{7\text{–}8}$ m^3 yr^{-1}. Thus the characteristic reaction time scale is of the order of

$$\tau_r \sim 10^5\text{–}10^6 \text{ years} \quad (4)$$

Table 3 gives estimates of the reaction time-scales of Asian floodplains that confirm the estimate of eqn 4 ($\tau_r \sim 4.5 \times 10^5$ years on average), with one exception. The Zhujiang seems to have a very small reaction time because of the reduced size of its floodplains (the Zhujiang has virtually no floodplain compared to other large Asian rivers). In the case of this river, diffusive buffering cannot realistically be the cause of the stability of sediment discharge. For the majority of the rivers we studied however, this reaction time-scale is large. This means that a signal like an erosion flux, defined by its amplitude A and period Z_{cl}, is transmitted at the upstream end of an alluvial plain and reaches the other end of the system (namely the sea), it is transformed as a diluted signal with amplitude A' on order of $A' \sim A(Z_{cl}/\tau_r)$ and period $\tau' \sim \tau_r$ (see Fig. 4). Hence diffusion has smoothed out the original perturbation by the time it reaches the sea. Therefore, climate-induced pulses (with time-scale of the order of 10^4 years like the Last Glacial Maximum 18 kyr BP, will have their amplitude divided by a factor of 10–100 at the other end of the floodplain and their duration multiplied by the same factor. Only events larger than at least 10^5 years will have a chance to pass through the floodplain buffer unaffected and therefore induce a significant change in sediment yield at the outlet. This kind of diffusive behaviour can be exemplified using analytical (Métivier, 1999) or numerical solutions of the diffusion equation (the 'tintinnabulations' of Humphrey & Heller, 1995).

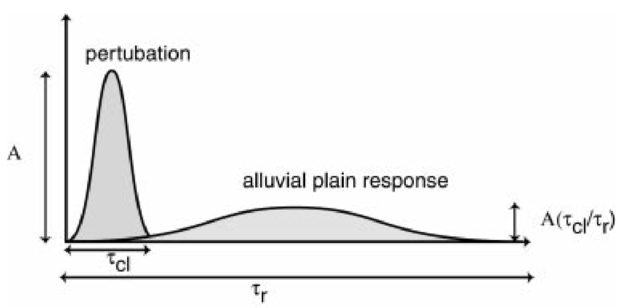

Fig. 4. Sketch showing the transformation from a pertubation of amplitude A and period τ_{cl} at the upstream end of an alluvial plain to pertubation filtered and buffered by the alluvial plain (new amplitude and period).

The buffering efficiency relies on this reaction time that scales with the size of the floodplain. Hence the larger the catchment and its alluvial plain the more efficient the buffer. This may help explain why the Eel river that has a small catchment (10^3 km^2) can experience a 20% variation in sediment yield during the Last Glacial Maximum (Syvistki & Morehead, 1999). We illustrate this showing how an alluvial plain like that of the Brahmaputra can react to a change in sea-level, averaging approximately 100 m during 10 kyr, which is the order of the sea-level change that occurred during the Last Glacial Maximum (Chappel & Shackelton, 1986).

Floodplain reaction to a 100-m sea-level drop. Sea-level changes have been recorded in quite a precise manner through the Quaternary (Chappel & Shackelton, 1986). Variations recorded can be as high as 150 m (at about 18 and 140 kyr BP). As an example of climatic perturbation to a system that has achieved steady state previously, we consider two extreme cases of a large sea-level fall (of the order of 100 m during 10 kyr). In the first case the slope of the river and subaqueous delta/continental shelf remains constant whereas in the second case we assume that an abrupt change in slope occurs at the river mouth. For each case we discuss the order of magnitude of the consequences of such a baselevel drop in terms of floodplain and catchment reaction.

Case 1: constant slope. Change in sea-level may clearly affect the surface of a catchment if the slope of the continental shelf is small (Syvistki & Morehead, 1999). As an extreme example, a slope of 10^{-3} to 10^{-4} combined with a sea-level fall of 100 m as happened during the Quaternary should lead to the expansion of the floodplain by $10^{5\text{–}6}$ m seaward. Does this change in floodplain surface affect the mass fluxes of sediments as is suggested by Mulder & Syvitski (1996) and Syvistki & Morehead (1999)? These authors have proposed, on the basis of a statistical analysis of present-day river data, that sediment yield is related to elevation (H) and drainage area (A) according to

$$Q_p \propto H^{3/2} A^{1/2}. \quad (5)$$

Looking at the possible effects of sea-level changes on sediment yield, these authors show that changes in sediment fluxes (for sea-level variations of the order of 25–200 m), are on average less than 30% (within the precision of the estimate). Furthermore these authors do not consider possible delays in the reaction of the catchment due to diffusive type inertia of the floodplain as discussed above. We therefore think that such a direct influence is not straightforward and this even more because no change in slope occurs. For a constant floodplain width, unchanged slopes ($\langle \partial Z/\partial s \rangle \sim 0$) should not trigger deep changes in rates of mass transport to the sea whatever the change in surface of the entire catchment including the floodplain. It should even more lead to a rise of the reaction time according to eqn 3.

The extension of the surface of the catchment therefore concerns a steady floodplain which is not the primary source of the erosion but only a temporary storage place. We do not question the first-order statistical validity of eqn 5 at a given time span but if one looks at variations of the sediment yield with respect to time variations in the catchment according to eqn 5 one finds that

$$\left.\frac{\partial Q_{\mathrm{p}}}{\partial t}\right|_{H} \propto \frac{H^{3/2}}{\sqrt{A}} \frac{\partial A}{\partial t}. \tag{6}$$

Hence, for the same variation pattern ($\partial A/\partial t$), the larger the catchment the smaller the time variation in sediment yield. Therefore large catchments act together with large flood plains to level the total flux of mass eroded and carried to the sea.

Case 2: abruptly varying slope. If the slope of the shelf differs from the slope of the alluvial plain then a change in sea-level should induce the formation of a knickpoint. Regressive erosion towards the upstream end of the floodplain would then tend to achieve a new equilibrium state.

Let us assume that $\Delta h_{\mathrm{sea}} \sim 100$ m change in sea-level occurs. Let us also consider a catchment for which the continental shelf is narrow and where an abrupt change of slope occurs at or near the river mouth. In order to maximize the magnitudes we derive, we furthermore assume that the slope of the talus is nearly vertical (Fig. 5), and use characteristic values from one of the largest systems in Asia that has one of the shortest timescales: the Brahmaputra. With the condition of varying slope being followed, a change in sea-level induces the formation of a knickpoint at the river mouth. In order to accommodate this perturbation of its profile the river will incise its floodplain in order to smooth the gradient and go back towards some new equilibrium profile.

For a large floodplain like that of the Bramaputra's (length $L \sim 0.9 \times 10^6$ m, width $w \sim 10^5$ m), the change in slope required to reach a new first order linear profile is $\tan \alpha \sim \Delta h_{\mathrm{sea}}/L \sim 10^{-4}$. The volume of material available for erosion in the floodplain V_{err} can then be approximated by

$$V_{\mathrm{err}} \sim \frac{\tan(\alpha) L^2 w}{2} \sim \frac{\Delta h_{\mathrm{sea}} L w}{2} \sim 4.5 \times 10^{12}\ \mathrm{m}^3. \tag{7}$$

Averaged to the time of the climatic oscillation $\tau_{\mathrm{cl}} \sim 10^4$ years, we get an average volume of sediments that can be released from the floodplain, on order of

$$\frac{\Delta h_{\mathrm{sea}} L w}{2\tau_{\mathrm{cl}}} \sim 4.5 \times 10^8\ \mathrm{m}^3\ \mathrm{yr}^{-1}. \tag{8}$$

Taking compaction into account (correction factor ≥ 0.6 for near-surface sediments, see Métivier *et al.*, 1999) this leads to a yearly solid volume of the order of 2.7×10^8 $\mathrm{m}^3\ \mathrm{yr}^{-1}$ to be compared to the present-day $2 \times 10^8\ \mathrm{m}^3\ \mathrm{yr}^{-1}$. Assuming that the river can erode all of this during the duration of the sea-level drop, the average nondimensional erosion flux $\hat{Q}_{\mathrm{av}}$ that the river can discharge (scaled to the average Quaternary flux), to accommodate the sea-level change is then

$$\hat{Q}_{\mathrm{av}} \sim \frac{\Delta h_{\mathrm{sea}} L w}{3.3 Z_{\mathrm{cl}} Q_{\mathrm{f}}} \sim 1.35. \tag{9}$$

Therefore, floodplain erosion can alone account for more than 150% of the Quaternary averaged mass flux carried to the sea.

The problem in the previous calculation is that we assumed a rapid alluvial plain reaction whereas we have

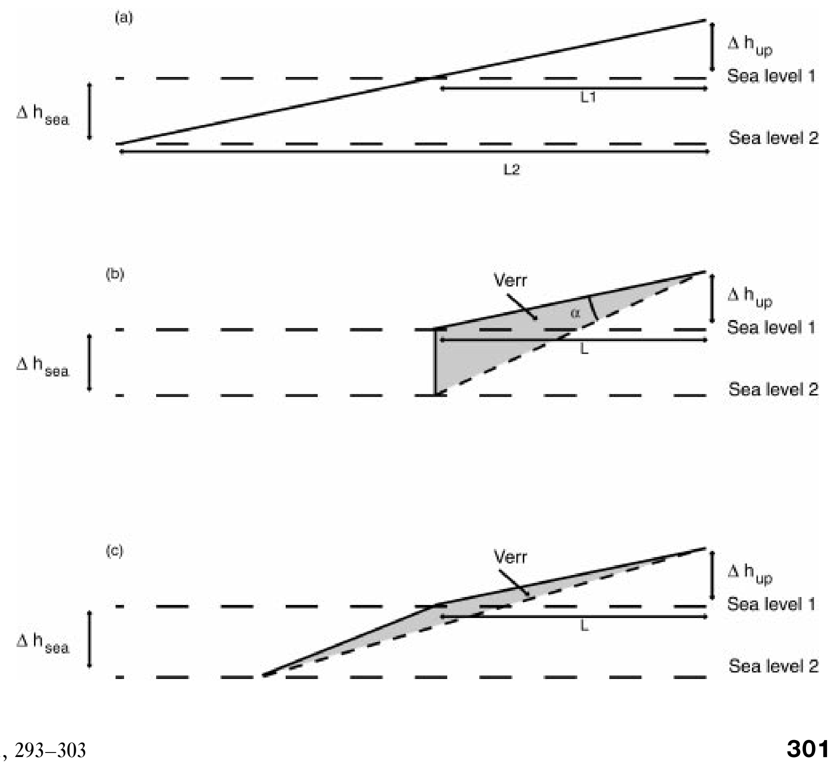

Fig. 5. Geometries of the floodplain and continental shlef studied in the text. (a) Slope remains constant when passing from the floodplain to the shelf; (b) the slope break idealizes the absence of a shelf; (c) more realistic geometry with varying slope between the floodplain and the continental shelf.

River	Orders of magnitude of the floodplain				
	Length (m)	Width (m)	Maximum relief (m)	Mass flux ($m^3\ yr^{-1}$)	Reaction time (yr)
Bramaputra	9×10^5	10^5	2×10^2	2×10^8	9×10^4 (8.5×10^4)
Chao Phraya	4×10^5	10^5	2×10^2	5.7×10^6	1.4×10^6
Chiang Jiang	10^6	1–2×10^5	3×10^2	1.8×10^8	1.7–3.3×10^5
Ganges	1.5×10^6	2×10^5	3×10^2	1.9×10^8	4.7×10^5
Hong He	3×10^5	0.5–1×10^5	3×10^2	6.9×10^7	0.6–1.3×10^5
Huang He	6×10^5	2×10^5	3×10^2	3.7×10^7	9.7×10^5
Indus	10^6	2×10^5	2×10^2	9×10^7	4.4×10^5
Mekong	7×10^5	2.5×10^5	3×10^2	5.5×10^7	9.5×10^5
Zhujiang	10^5	10^4	10^2	1.9×10^7	5.2×10^3

Table 3. Estimate of the reaction time of large alluvial plains in Asia. Numbers given are orders of magnitude as the shapes of the floodplains may vary significantly along the river's course. Values in parentheses for the Brahmaputra come from Dade & Friend (1998) and are given for comparison. Hypsography from Defense Mapping Agency (1992).

shown that the characteristic reaction time for an alluvial plain the size of the Brahmaputra is of the order of 0.9×10^5 years (Table 3; Dade & Friend (1998)). Therefore, the mass flux that will really be eroded from the floodplain to accommodate this sea-level variation will scale with the ratio of the characteristic climate time-scale to the characteristic alluvial plain time-scale as

$$\hat{Q}_{er} \sim \frac{\tau_{cl}}{\tau_r}\hat{Q}_{av} \sim \frac{\Delta h_{sea} L w}{3.3\tau_r Q_f} \sim 0.15. \qquad (10)$$

The river is not left the time to level all of its floodplain during such a climate oscillation and we only expect a change in the fluxes of the order of 15% to occur at the river mouth.

This explains why available mass may well exceed that being carried to the sea. The example case we treated is an extreme one. A 100-m sea-level drop is an extreme event and the Brahmaputra floodplain does not end abruptly by a vertical slope (Fig. 5c). Therefore, the volumes available to erosion should be lower. Nevertheless, the orders of magnitude hold and show that floodplain processes can clearly be regarded as of first order importance in understanding how a drainage system smoothes climate-induced pertubations.

SUMMARY

The comparison between the present-day discharge of the greatest rivers of Asia and the mass accumulated in the sedimentary basins they feed suggests that river systems in Asia with catchments of the order of 10^6 km^2 seem to have remained stable in tectonically active regions during the last two million years. This suggests either that incision rates are mainly controlled by uplift rates in rapidly uplifting mountain ranges or that large catchments may, to the first order, smooth climate-induced changes in erosion through diffusive buffering by their alluvial plains.

We have shown that the time-scales characterizing the alluvial plain reaction are larger, on average, by more than an order of magnitude compared with time-scales of Quaternary climate changes. We suggest this difference to be responsible for the smoothing of any perturbation (change in erosion rates upstream or change in sea-level at the outlet), that may affect large rivers on the Asian continent. As the characteristic time-scale of a floodplain reaction scales with its size (Eq. 3), we may safely propose that the larger the catchment or the alluvial plain the more stable the mass fluxes.

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