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How plausible are high-frequency sediment supply-driven cycles in the stratigraphic record?

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Abstract

This paper is an attempt to quantify the plausible time scales of clastic sediment supply variations at the entrance of sedimentary basins. Our approach is based on the sedimentary system concept, which simplifies natural systems by dividing them into three zones of dominant processes: the erosion, the transfer, and the sedimentation subsystems. We examine recent results from geomorphology, which show that frequent climate changes can induce high-frequency sediment flux variations at the outlet of the source area. We put forward the crucial role of the transfer subsystem, which conveys sediment from the erosion zone to the basin. By applying a diffusive model to a number of worldwide rivers, we extend from large (>1000 km) to intermediate (>300 km) rivers the previous finding that the transfer subsystem acts as a buffer for short periods sediment pulses (tens to hundreds of kiloyears). This implies that high-frequency stratigraphic cycles in clastic accumulations fed by large drainage systems are unlikely to reflect sediment supply cycles of tens to hundreds of thousands of years of periodicities.

Keywords: Cycles; Sequences; Sediment supply; Sedimentary system; Geomorphology

1. Introduction

At first order, the stratigraphic record is made of sedimentation changes that encompass a large range of time scales (Einsele et al., 1991) from a few seconds (laminations) to several tens of million years or more (major global changes). The most commonly studied of those variations are basin scale repetitive packages of strata called sequences or cycles, with periods ranging from tens of thousands of years to several million years. In siliciclastic successions, such cycles can be recognized by tracking the movements of a stratigraphic indicator as, for instance, the gravel–sand transition or the shoreline in continental or marine deposits, respectively (Marr et al., 2000; Paola et al., 1992; Swenson et al., 2000). These movements indicate changes of the shape of the entire depositional system in the search for an equilibrium with changing boundary conditions. The goal of stratigraphy is to read this stratigraphic record of changing external factors. A crucial question then is: what is the origin of those cycles?

Since Sloss (1962) and the subsequent advances brought by sequence stratigraphy (e.g., Cross, 1988; Jervey, 1988; Posamentier et al., 1988; Schlager, 1993; Shanley and McCabe, 1994; Muto and Steel,

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1997, 2000; Blum and Törnqvist, 2000), it is now generally accepted that stratigraphic cycles are somehow governed by changes in the ratio between space available for sedimentation or accommodation (A) and sediment supply (S) to this space. In this way, all factors that can affect accommodation and/or sediment supply are virtually able to produce cycles in the stratigraphic record. Eustasy, or another base level in continental areas, basin tectonics, and sediment supply have all been claimed to be variable at all time scales and responsible for creating stratigraphic cycles.

There is broad agreement that climate-induced base-level oscillations with Milankovitch periodicities of tens to hundreds of thousands of years are responsible for creating high-frequency stratigraphic cycles (fourth- and higher-order cycles) with the same periodicities (e.g., Van Wagoner et al., 1990; Plint et al., 1992; Nystuen, 1998; Gale et al., 2002). Also, some recent studies (e.g., Perlmutter and Matthews, 1989; Weltje and de Boer, 1993; Weltje et al., 1996; Burns et al., 1997; Tiedemann and Franz, 1997; Perlmutter et al., 1998; Lopez-Blanco et al., 2000; Marzo and Steel, 2000; Van der Zwan, 2002) have suggested that the sediment flux to basins could vary with those periodicities due to climate changes or vertical movements (tectonics) in the source area, and should therefore have direct control over the high-resolution stratigraphic record. This idea is mainly influenced by the correlations found between various climatic (mean precipitation, total precipitation, temperature range, etc.) and geographic (drainage area, relief, maximum height, etc.) factors and the present-day sediment output at the mouth of rivers (e.g., Fournier, 1960; Milliman and Meade, 1983; Milliman and Syvitski, 1992; Pinet and Souriau, 1988; Summerfield and Hulton, 1994; Mulder and Syvitski, 1996; Hovius, 1998). Such correlations can be considered as erosion laws, but this requires the assumption that the system is at equilibrium with those factors.

In this paper, we put in question the high-frequency variability of sediment flux to basins (with tens to hundreds of thousands of years periods) and its link with climate and vertical movement changes in the source domain.

We first investigate how the sedimentary system concept, rooted in the earlier work of Schumm (1977), clarifies our questioning of the variability of sediment flux. In particular, this highlights the crucial role of the transfer zone, which conveys sediments from the source area to the basin. Then, we put constraints on the plausibility of high-frequency sediment flux variations to the basin by (1) examining the response times of the source area to climate and vertical movement changes, in light of recent results from geomorphology, and (2) analysing the first-order response time of some worldwide rivers to sediment input variations coming from the source area.

2. The sedimentary system

Following Schumm (1977) and Allen (1997), let us consider a sedimentary system (Fig. 1) as a closed domain at the lithosphere/atmosphere interface, composed of three subsystems each characterized by a dominant process: erosion, transfer (the balance between erosion and sedimentation), and sedimentation subsystems. This is valid at any space and time scales for which such distinct zones of dominant processes can be identified. Here we consider only

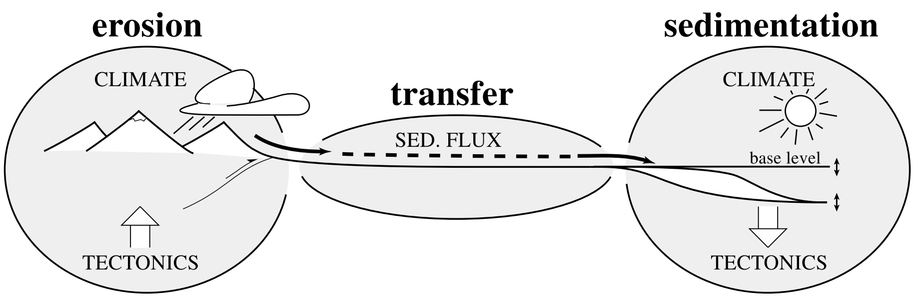

Fig. 1. Idealized cartoon of a mountain–river–delta sedimentary system showing the three elementary subsystems of erosion, transfer, and sedimentation.

macroscale sedimentary systems (e.g., schematic mountain–river–delta or catchment–fan systems) on geological time scales. This concept is aimed at distilling the first-order characteristics and dynamics of real systems from their natural complexity.

The erosion subsystem is composed schematically of hillslopes, which are the main sediment feeders, and channels, which incise and drain sediment downstream. It is controlled by vertical movements with respect to a reference level defined at its outlet, either due to tectonics or associated with base-level changes, and by climate.

The transfer subsystem is made up of rivers transporting the sedimentary flux leaving the erosion subsystem to the sedimentation subsystem. The length of the transfer subsystem varies from zero in catchment–fan systems to several thousands of kilometers in the largest current systems. At its upstream boundary, it is then subjected to sediment flux variations coming from the erosion subsystem, and to base-level changes at its downstream boundary.

The sedimentation subsystem (basin) stores the sediment flux in a variety of depositional environments whether continental and/or marine. It is subject, on one hand, to sediment flux variations at its boundary with the preceding subsystem (erosion or transfer) and, on the other hand, to base-level changes and basin tectonics, both of which modify the space available for sedimentation.

This raises a fundamental difference between accommodation modifying factors and sediment supply modifying factors: while base-level variations and basin tectonics apply directly to the sedimentation subsystem, the sediment flux appears to be a complex derivative of the effects of external forcing on the erosion zone and of their transmission by the transfer subsystem. Each subsystem has an intrinsic response time (or equilibrium time), which is the time needed to return to equilibrium after a change in boundary conditions (Paola et al., 1992; Beaumont et al., 2000). A system with stable boundary conditions is at equilibrium (with these conditions) when its shape does not evolve with time. If boundary conditions evolve slowly compared to the response time, the subsystem will respond in a quasi-equilibrium manner (i.e., at each time in equilibrium with its new conditions). On the contrary, if they vary rapidly compared to the response time, the response will not be in equilibrium with the forcing (i.e., out-of-phase with the forcing and of different amplitude). Therefore, sediment flux variations at the sedimentation subsystem entrance may not necessarily be tied in a straightforward way to allogenic changes in the erosion zone, depending on the response times of the erosion and transfer subsystems compared with the periodicities of allogenic forcing.

In the following, we do not address the response of the transfer subsystem to base-level changes (see, e.g., Paola et al., 1991; Burns et al., 1997) since the production and transport of sediments may be mostly upstream-controlled (Blum and Törnqvist, 2000). The understanding of the variability of sediment flux to sedimentary basins in relation with upstream allogenic controls therefore requires determining (1) the response time of the erosion zone to climate changes and vertical movements (tectonics/base level) and (2) the response time of the transfer subsystem to sediment flux variations coming from the erosion subsystem.

3. Constraints from geomorphology

3.1. The erosion subsystem

A number of recent works in geomorphology address the response of erosion to climate and tectonics.

A first important qualitative result, although not explicitly stated, is that the response to vertical movements will always take longer than the response to climate because the tectonic signal must propagate up the drainage network whereas climate can impact the entire drainage basin at once (Whipple, personal communication; e.g., Fernandes and Dietrich, 1997).

Secondly, the response times have been assessed by the field evidence of landscape adjustment to new conditions, and computed by the calibration of models on natural cases. The minimum response times to tectonics are on the order of 100 ka in small (10 km) catchment–fan systems (Allen and Densmore, 2000) and coastal drainage basins (3–10 km; Snyder et al., 2000), and best comprised between 0.25 and 2.5 Ma in drainage basins of medium size (20–40 km; Whipple, 2001). This suggests minimum response times to tectonics ranging from hundreds of thousands of years to 1 Ma, or more in larger drainage basins (Tucker and Slingerland, 1996; Whipple, 2001).

The response to climate in terms of sediment flux is different because there is an initial near-immediate response (i.e., on the order of 1 ka to several thousands of years, when erosivity is increased). This mostly reflects a release of the sediments stored in hillslope soils that affects the entire drainage area (Fernandes and Dietrich, 1997; Tucker and Slingerland, 1997). However, the ability of hillslopes to produce significant regolith depends on the dominant climate. For instance, Bull (1991) notes that during the last 130 ka, only three aggradation events are recorded in the arid San Gabriel Mountains and Mojave Desert, whereas the marine record yields 11 highstand terraces for the same interval. This suggests that initial response times can reach several tens of thousands of years in particular environments. In addition to the initial response time, over the long term, sustained climate shifts (greater than hundreds of thousands of years) can modify channel long profiles with the same response times as for responses to tectonic changes (i.e., on the order of hundreds of thousands of years or more) (Whipple, 2001).

In conclusion, the erosion subsystem will filter high-frequency tectonic events (<100 ka) and take more than 100 ka to adjust to sustained climate shifts and tectonic disturbances. By contrast, because of the initial response due to the interaction between channels and hillslopes, the erosion subsystem has the potential to respond immediately to frequent climate changes, and therefore to produce high-frequency sediment flux oscillations at its outlet (Tucker and Slingerland, 1997).

3.2. The transfer subsystem

In the transfer subsystem, rivers convey sediments from the upstream source area down to the sedimentation subsystem. Following Allen and Densmore (2000), one can consider, at first order, its behaviour as diffusive (Paola et al., 1992; Humphrey and Heller, 1995; Dade and Friend, 1998; Métivier, 1999; Métivier and Gaudemer, 1999). In that case, the response time T of the transfer subsystem is of the form:

$$T = L^2/K \tag{1}$$

with L as the length of the subsystem and K as its coefficient of diffusivity. The larger the transfer subsystem, the longer its response time, and the more diffusive it is, the shorter its response time. Note that we do not investigate here the fate of sediment waves supplied kinematically to channels, which can quickly deliver sediment downstream. These are short-term events for which the transfer subsystem may no longer be considered as a simple diffusive entity (see, e.g., Cui et al., submitted for publication; Lisle et al., 1997).

In natural systems, Dade and Friend (1998) have calculated river diffusivities by using the water flux per unit width and a sediment mobility parameter, which embodies the effects of bedload and suspended load in transport. They find response times of 65, 85, 21, 2.4, 74, and 5.5 ka for the Mississippi, Brahmaputra, Indus, Savannah, North Platte, and Cheyenne rivers, respectively (i.e., in the range of 1 ka to tens of thousands of years). Although the physical ground of this calculation is attractive, it is difficult to apply it to other rivers in the world because accurate data as bedload proportion and median grain size are usually not available. Also, those response times do not reflect the buffering effect of large Asian rivers for high-frequency sediment input variations, as evidenced by the correlation between currently measured sediment flux at their mouth and the average filling rates of their marine depocentres over the last 2 Ma, despite strong climatic variations (Métivier et al., 1999).

Métivier (1999) and Métivier and Gaudemer (1999) show that the diffusivity coefficient of a river approaching equilibrium conditions scales with its output sediment flux Qst, width W, and mean slope $\langle \partial z/\partial x \rangle$:

$$K = \frac{\mathrm{Qst}}{W\left\langle \dfrac{\partial z}{\partial x} \right\rangle} \tag{2}$$

With this relation and relation (1), they derive first-order response times in the range of 10^5–10^6 years for some large Asian river floodplains, which explains their strong buffering action for high-frequency sediment input variations (Métivier, 1999; Métivier and Gaudemer, 1999).

We apply relations (1) and (2) to the dataset of Hovius (1998) to further investigate the magnitude of the buffer effect for a greater variety of intermediate and large drainage basins worldwide ($>2.5 \times 10^4$ km^2; Table 1).

Table 1
Response times for 93 rivers of the Hovius (1998) database

River [units]	Stream length,[a] Lr [km]	Drainage area,[a] A [km^2]	Maximum height,[a] H_{max} [m]	Sediment flux,[a] Qst (with sediment density = 2700 kg/m^3) [10^6 m^3/a]	Mean slope,[a] S ($S=H_{max}$/Lr) [10^{-3}]	Minimum estimated width, W ($W=cA^b$ with $c=0.001$ and $b=0.5$) [m]	Diffusivity coefficient, K (K = Qst/WS) [10^6 m^2/a]	Response time, T ($T=L^2/K$ with Qst = suspended and dissolved loads) [ka]	Response time,[b] Tb (Tb = T/2 with Qstb = 2Qst) [ka]
Nile	6670	2,715,000	5110	53	0.8	1648	42	1060	530
Amazon	6299	6,150,000	6768	508.5	1.1	2480	190.8	208	104
Mississippi	5985	3,344,000	4400	194.4	0.7	1829	144.6	248	124
Ob	5570	2,500,000	4506	24.4	0.8	1581	19.1	1623	812
Yenisey	5550	2,580,000	3492	28.9	0.6	1606	28.6	1078	539
Yangtze	5520	1,940,000	6800	261.5	1.2	1393	152.4	200 (170–330)[c]	100
Yellow (Huang He)	4670	980,000	5500	52.6	1.2	990	45.1	483 (970)	242
Mekong	4500	810,000	6000	81.5	1.3	900	67.9	298	149
Parana	4500	2,600,000	6720	62.2	1.5	1612	25.8	784	392
Amur	4416	1,855,000	2499	26.7	0.6	1362	34.6	564	282
Lena	4400	2,430,000	2579	37	0.6	1559	40.5	478	239
Zaire	4370	3,700,000	4507	25.5	1	1924	12.8	1487	743
Mackenzie	4240	1,448,000	3955	70	0.9	1203	62.4	288	144
Niger	4160	1,112,700	2918	15.6	0.7	1055	21	823	412
Kolyma	3513	647,000	3147	2.2	0.9	804	3.1	4002	2001
Murray	3490	910,000	2239	14.4	0.6	954	23.6	516	258
Volga	3350	1,350,000	1638	38.1	0.5	1162	67.1	167	84
Indus	3180	960,000	8611	107.8	2.7	980	40.6	249 (440)	124
Salween	3060	325,000	6070	37	2	570	32.8	286	143
St. Lawrence	3060	1,185,000	1917	23.3	0.6	1089	34.2	274	137
Yukon	3000	855,000	6194	34.8	2.1	925	18.2	494	247
Rio Grande	2870	670,000	4295	11.9	1.5	819	9.7	851	426
Danube	2860	815,000	3087	48.1	1.1	903	49.4	166	83
Brahmaputra	2840	610,000	7736	215.2	2.7	781	101.1	80 (90)	40
Sao Francisco	2800	640,000	1800	2.2	0.6	800	4.3	1814	907
Shatt al Arab	2760	1,050,000	4168	44.8	1.5	1025	29	263	132
Orinoco	2740	945,000	5493	70	2	972	35.9	209	105
Zambezi	2660	1,400,000	2606	23.3	1	1183	20.1	352	176
Amudar'ya	2620	309,000	7459	44.8	2.8	556	28.3	242	121
Ganges	2510	980,000	8848	221.9	3.5	990	63.6	99 (470)	50
Ural	2430	237,000	1000	2.2	0.4	487	11.1	532	266
Colorado (Cal)	2333	640,000	4730	61.1	2	800	37.7	144	72
Irrawaddy	2300	410,000	5881	130.4	2.6	640	79.6	66	33
Syrdar'ya	2210	219,000	5880	8.9	2.7	468	7.1	684	342
Dnepr	2200	504,000	325	4.9	0.1	710	46.3	105	52
Xi Jiang	2129	464,000	2500	78.5	1.2	681	98.2	46	23
Columbia	1950	670,000	3748	18.5	1.9	819	11.8	323	162
Don	1870	422,000	367	7.4	0.2	650	58.1	60	30
Orange	1860	1,020,000	3482	38.1	1.9	1010	20.2	171	86
Pechora	1810	322,000	1894	4.9	1	567	8.2	401	200
Indigirka	1726	360,000	3147	5.9	1.8	600	5.4	550	275
Limpopo	1600	440,000	2322	12.2	1.5	663	12.7	202	101
Volta	1600	394,000	500	8.1	0.3	628	41.5	62	31
Magdalena	1530	260,000	5493	91.9	3.6	510	50.2	47	23

(continued on next page)

Table 1 (*continued*)

River [units]	Stream length,[a] Lr [km]	Drainage area,[a] A [km^2]	Maximum height,[a] H_{max} [m]	Sediment flux,[a] Qst (with sediment density = 2700 kg/m^3) [10^6 m^3/a]	Mean slope,[a] S ($S=H_{max}$/Lr) [10^{-3}]	Minimum estimated width, W ($W=cA^b$ with $c=0.001$ and $b=0.5$) [m]	Diffusivity coefficient, K (K=Qst/WS) [10^6 m^2/a]	Response time, T ($T=L^2/K$ with Qst = suspended and dissolved loads) [ka]	Response time,[b] Tb (Tb = T/2 with Qstb = 2Qst) [ka]
Godavari	1500	287,000	1300	63	0.9	536	135.6	17	8
Colorado (Tex)	1450	100,000	1440	6.7	1	316	21.2	99	50
Senegal	1430	441,000	1000	1.1	0.7	664	2.3	884	442
Brazos	1400	114,000	950	12.6	0.7	338	55	36	18
Chari	1400	880,000	3071	2.6	2.2	938	1.3	1556	778
Rufiji	1400	178,000	2959	6.3	2.1	422	7.1	278	139
Kura	1360	188,000	4480	15.2	3.3	434	10.6	174	87
Rhein	1360	225,000	4158	6.6	3.1	474	4.5	409	204
Dnestr	1350	72,100	2058	2.4	1.5	269	5.9	310	155
Liao He	1350	170,000	2029	15.2	1.5	412	24.5	74	37
Krishna	1290	256,000	1892	24.1	1.5	506	32.4	51	26
Chao Phraya	1200	160,000	2300	5.2	1.9	400	6.8	213 (1400)	106
Red (Song Koi)	1200	120,000	3000	45.6	2.5	346	52.6	27	14
Kizil Irmak	1151	75,800	3916	8.5	3.4	275	9.1	146	73
Elbe	1110	148,000	1603	0.3	1.4	385	0.6	2200	1100
Fraser	1110	220,000	4043	11.5	3.6	469	6.7	183	92
Loire	1110	120,000	1885	0.6	1.7	346	0.9	1305	652
Kuskokwim	1080	116,000	6194	2.8	5.7	341	1.4	820	410
Mobile	1064	57,000	1360	2.3	1.3	239	7.6	148	74
Vistula	1014	198,000	2499	5.7	2.5	445	5.2	196	98
Rio Colorado (Arg)	1000	65,000	6960	2.6	7	255	1.4	694	347
Rio Grande Santiago	960	125,000	4577	0.4	4.8	354	0.2	4194	2097
Ebro	930	86,800	3404	7.8	3.7	295	7.2	120	60
Meuse	925	29,000	692	0.3	0.7	170	2	420	210
Oder	909	112,000	1603	2.6	1.8	335	4.5	185	92
Apalachicola	880	51,800	1458	0.4	1.7	228	1.1	674	337
Jana	872	238,000	3000	1.5	3.4	488	0.9	861	431
Sanaga	860	135,000	2000	2.2	2.3	367	2.6	289	145
Mahanadi	858	133,000	1027	22.2	1.2	365	50.9	14	7
Sepik	825	81,000	4500	29.6	5.5	285	19.1	36	18
Rhone	810	99,000	4810	22.2	5.9	315	11.9	55	28
Seine	780	78,600	902	4.9	1.2	280	15	41	20
Fly	744	64,400	3993	25.9	5.4	254	19	29	15
Susquehanna	733	72,500	950	0.7	1.3	269	1.9	281	141
Rio Negro (Arg)	729	130,000	4800	4.8	6.6	361	2	262	131
Weser	724	46,000	1142	0.1	1.6	214	0.4	1451	725
Tana	720	91,000	5200	11.9	7.2	302	5.4	95	48
Po	691	75,000	4810	10.4	7	274	5.4	88	44
Burdekin	680	131,000	1277	1.1	1.9	362	1.6	283	141
Colville	662	60,900	2320	2.6	3.5	247	3	146	73
Garonne	650	86,000	3308	0.8	5.1	293	0.5	774	387
Haiho	650	50,800	2870	30	4.4	225	30.1	14	7
Terek	623	43,200	5642	10	9.1	208	5.3	73	37
Sacramento	610	73,000	3187	9.3	5.2	270	6.6	57	28
Kemijoki	600	37,800	807	0.1	1.3	194	0.2	1695	847

Table 1 (*continued*)

River [units]	Stream length,[a] Lr [km]	Drainage area,[a] A [km^2]	Maximum height,[a] H_{max} [m]	Sediment flux,[a] Qst (with sediment density = 2700 kg/m^3) [10^6 m^3/a]	Mean slope,[a] S ($S = H_{max}$/Lr) [10^{-3}]	Minimum estimated width, W ($W = cA^b$ with $c = 0.001$ and $b = 0.5$) [m]	Diffusivity coefficient, K (K = Qst/WS) [10^6 m^2/a]	Response time, T ($T = L^2/K$ with Qst = suspended and dissolved loads) [ka]	Response time,[b] Tb (Tb = T/2 with Qstb = 2Qst) [ka]
San Joaquin	560	80,100	4420	0.7	7.9	283	0.3	946	473
Delaware	518	22,900	1360	0.6	2.6	151	1.6	171	86
Susitna	454	50,300	6190	9.3	13.6	224	3	68	34
Copper	360	61,800	5952	25.9	16.5	249	6.3	21	10

[a] All the rivers in the Hovius (1998) database are used, less the Mahakam, Ord, Sevemaya Dvina, and Uruguay rivers for which the lengths are not given.

[b] Tb is the response time computed with a sediment flux two times larger than given in Hovius (1998) in order to take into account a maximum of 50% bedload contribution.

[c] For comparison, the values computed by Métivier (1999) for some large Asian rivers are given between parentheses.

The total sediment flux output Qst of these rivers is given by the sum of the total annual suspended and solute loads (Table 1). Taking into account the bedload contribution, usually considered to be on the order of 10% (Hovius, 1998), and trying to avoid sediment flux underestimations, we increase sediment flux Qst by a factor of two (Qstb). This is based on Paleogene sediment volume measurements in the North Sea basin, which have shown an average proportion of up to 50% sand (Liu and Galloway, 1997).

The estimate of the river width W at its mouth is based on the classic hydraulic geometry relation (Leopold and Maddock, 1953):

$$W = cA^b$$

where A is the drainage area, and c and b are two positive coefficients. We use $b = 0.5$ to respect the classic square root relationship between width and discharge for alluvial channels (Leopold and Maddock, 1953; Knighton, 1998). Although the coefficient c is naturally specific for each river, we use the same $c = 0.001$ for all rivers because it is a minimum coefficient observed on natural alluvial reaches (Montgomery and Gran, 2001), and in the hope that this would therefore provide a minimum width (Table 1).

The mean river gradient $\langle \partial z / \partial x \rangle$ (S in Table 1) is calculated by dividing the maximum elevation H_{max} in the drainage basin by the stream length Lr.

The obtained response times (Table 1, Fig. 2) range between a maximum T calculated with only suspended and dissolved loads, and a minimum Tb calculated with a sediment flux at river mouths two times larger to account for 50% of bedload sediment transport.

The comparison of our results with the response times of Métivier and Gaudemer (1999) (Table 1) shows that even our maximum response times may be large underestimations of real ones, and may therefore strongly minimize the buffering action of rivers. This is mainly due to the underestimation of river widths by about one order of magnitude compared with certain real values (Penn, 2001). In the minimum case (Tb), 58% of those rivers have response times of more than 100 ka, 78% of more than 40 ka, and 91% of more than 20 ka. Therefore, even intermediate rivers, compared to large Asian rivers, can have a strong buffering effect for high-frequency sediment input disturbances.

By analogy with the skin distance in heat diffusion problems (e.g., Turcotte and Schubert, 1982), we have plotted (Fig. 2) a "buffer distance" Bd for sediment flux oscillations with periods of 20, 40, and 100 ka (Fig. 2B–D respectively):

$$\mathrm{Bd} = \sqrt{\frac{K\lambda}{\pi}}$$

This buffer distance is the distance over which sediment flux disturbances of period λ are lessened by one

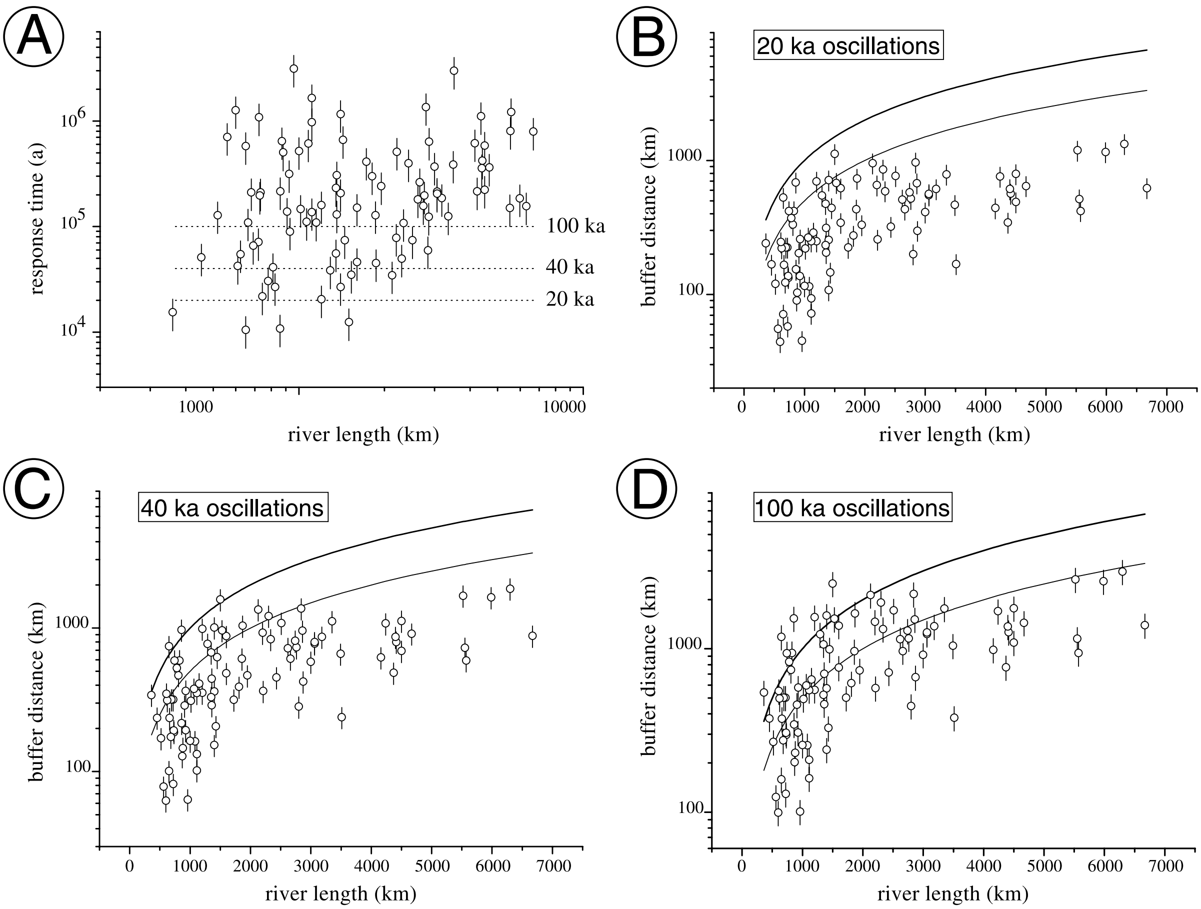

Fig. 2. Response times and buffer distances of some modern rivers (data from Hovius, 1998) as a function of their length, considering a diffusive behaviour for the transfer of sediments by rivers. For each point, the vertical line represents the uncertainty associated with a more or less 50% of bedload contribution to the total sediment load. (A) Response times versus river lengths: a majority of response times reach values of more than 100 ka. (B–D) Buffer distances versus river lengths for 20, 40, and 100 ka sediment flux disturbances, respectively. The bold and tight curves represent river and half-river lengths, respectively.

third of their initial amplitude. After the distance Bd, the disturbances are phase-delayed by 1 rad (i.e., by about 0.16 λ) (Turcotte and Schubert, 1982). For most rivers, sediment flux disturbances of 20 and 40 ka periodicities are attenuated by one third of their initial amplitude over distances of less than half their length (Fig. 2B and C). This is less well defined for 100 ka disturbances (Fig. 2D), although it remains valid for a majority of rivers.

Note that a drawback of this model is that it assumes a constant diffusivity with time. In particular, the influence of water discharge variations (due to climate change), which are expected to accompany sediment input variations to the transfer subsystem, is not accounted for by this model. This means that the model only investigates the response time of the transfer subsystem to sediment input variations.

Taking into account that we have computed minimum estimates, it appears that not only large but also intermediate rivers (>300 km) can act as a strong buffer for high-frequency ($\leq$ 100 ka) sediment input variations to the transfer subsystem. For most natural rivers and for disturbances with periodicities between 20 and 100 ka, the buffer effect induces a significant signal attenuation over a distance of less than half the river length.

4. Discussion

There is no debate as to whether the sediment input is a fundamental variable in controlling the stratigraphic record (e.g., Galloway, 1989; Lawrence, 1993; Schlager, 1993). The debate is about the time scales of this control. By focusing on the depositional area, stratigraphers have often assumed that sediment flux was comparable to relative sea level in terms of variability (i.e., that sediment flux variations to the basin were directly tied to climate or tectonic disturbances in the source area). In the light of the sedimentary system concept, however, it appears that (1) the sediment flux is a derivative of tectonic and climatic changes in the source area and (2) then has to be transported from its production zone to the deposition zone, which is unlikely to be instantaneous. This last point is usually neglected (e.g., Perlmutter et al., 1998; Van der Zwan, 2002). The sedimentary system concept therefore puts forward that the first-order controls on the time scales of variation of the sediment flux to the basin are the response times of the erosion and transfer subsystems.

The analysis provided here by using a simple diffusive model for fluvial entities shows that intermediate and large transfer subsystems (>300 km) will buffer high-frequency ($\leq$ 100 ka) sediment input disturbances coming from the erosion subsystem. This is in agreement with the buffering action evidenced for large Asian floodplains facing potential high-frequency climate-induced sediment flux variations during the last 2 Ma (Métivier, 1999; Métivier and Gaudemer, 1999). The transfer subsystem therefore plays a crucial role in the final stratigraphic record of allogenic forcings.

A strong implication is that stratigraphic studies interested in clastic successions should always be aware of the first-order dimensions of the erosion and transfer subsystems in order to assess the plausibility of high-frequency sediment flux variations. In sedimentary systems with short (perhaps less than 300 km) to negligible transfer subsystems, such as catchment–fan systems, high-frequency variations of sediment flux to the sedimentation subsystem can occur in equilibrium with climate changes in the source area. In such systems, if the influences of basin factors, which combine with the sediment flux to yield the final stratigraphy, can be unraveled, the stratigraphic record may therefore provide valuable information about short-term climatic and tectonic changes in the source zone. In detrital accumulations fed by way of intermediate to large transfer subsystems, as in the case of large deltas for example, high-frequency (100 ka) sediment flux oscillations may not occur in equilibrium with allogenic changes in the source area. Therefore, high-frequency stratigraphic cycles cannot be an equilibrium response to such allogenic changes. In these accumulations, only over the long term (i.e., of more than hundreds of thousands of years) can sediment supply variations in equilibrium with climate or tectonic changes in the source area be detected, as evidenced in several studies (e.g., Sloss, 1979; Raymo et al., 1988; Hay et al., 1988; Galloway and Williams, 1991; Nott and Roberts, 1996; Liu and Galloway, 1997; Peizhen et al., 2001), and have a possible influence on the stratigraphic record. The high-resolution stratigraphic record of basins fed by intermediate to large transfer subsystems can provide information about high-frequency variations of basin factors as eustasy or basin tectonics, but not about high-frequency climatic or tectonic changes in the upstream zones. Note that we do not argue here that high-frequency sediment supply variations at the outlet of the transfer subsystem do not occur. We only put forward that they will not be in equilibrium with the forcing if the transfer length is intermediate to large. In this way, our conclusions do not preclude rich stratigraphic responses of alluvial basins to rapidly changing sediment supply or diffusivity (Paola et al., 1992).

A weakness of our analysis is that it is based on modern river data and on assumptions, such as the square root relationship between drainage area and river width, which may be different in ancient sedimentary systems (Paola, 2000). Also, the problem of approximating the transfer of sediments by linear diffusion should be further addressed.

Therefore, our conclusions underscore the need for future research on the behaviour of the transfer subsystems and of the sedimentary system in general.

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