# Land–ocean sediment transfer in palaeotimes, and implications for present-day natural fluvial fluxes

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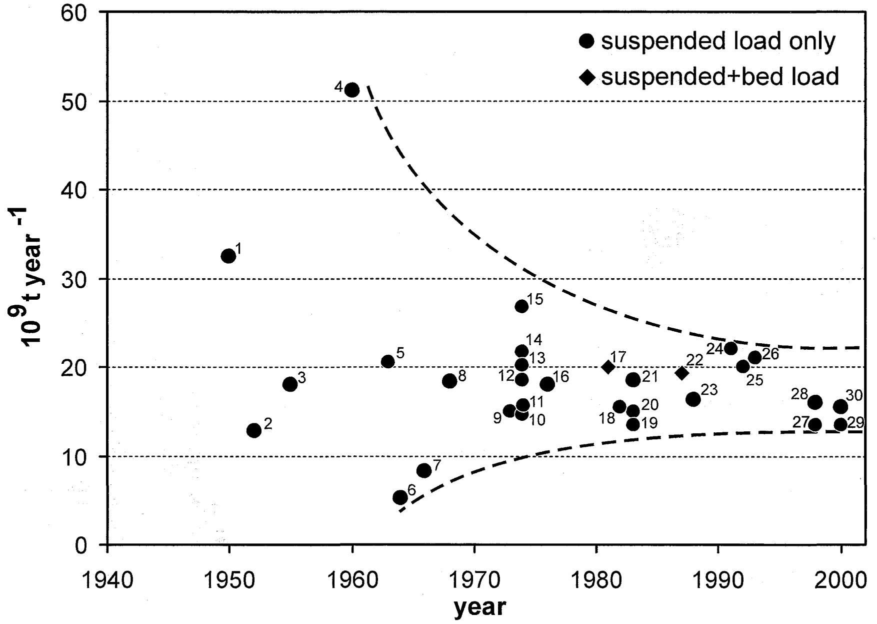
**Abstract** Detrital fluxes to the ocean during the period from the late Jurassic to the Pliocene ranged between 2.7–5.2 Gt year$^{-1}$, but increased up to 9.6–15.5 Gt year$^{-1}$ in the Pliocene–Holocene. Of the Holocene flux, 11–12 Gt year$^{-1}$ was carried by rivers. This estimate is suggested as the natural component of present-day fluvial sediment flux to the ocean. The contemporary flux is 1.1- to 2-fold higher, due to human disturbance of the land. Small basins make a major contribution to this increase, while many large rivers demonstrate stable fluxes. This is due to sediment trapping within the fluvial system, with sediment residence time increasing with increasing basin size. Over long time spans sediment delivery ratios (*SDR*) are less dependent on basin size. At a global land-ocean scale, the *SDR* is estimated to be in the range 5–20% for timescales ranging from decades to centuries but its value increases to 77% for the last 150 million years.

**Key words** anthropogenic; Cenozoic; denudation; global sediment flux; Holocene; Pleistocene; sediment delivery ratio

## INTRODUCTION

Interest in fluvial sediment fluxes to the world's oceans began in the early 1950s, and peaked in the 1970s (Fig. 1). Initial estimates of the magnitude of the fluvial sediment flux were based on limited data sets, and as a result, the various flux estimates differed by as much as an order of magnitude. However, current estimates, developed over the last two decades, tend to fall within the range $13.5 \times 10^9$ to $22 \times 10^9$ t year$^{-1}$. Despite increasing reservoir trapping, the contemporary sediment flux to the oceans is still believed to have increased relative to pre-human conditions. The term natural, or pristine has been used to refer to the sediment discharge that would have occurred in the absence of anthropogenic disturbance. It is supposed relatively stable due to the stability of the natural global water balance. The observed global sediment discharge to the oceans may be presented as the sum of an essentially constant natural component and a changing anthropogenic component. To distinguish between the natural and anthropogenic components researchers have employed different approaches including, for example, assessment of the land area subject to accelerated erosion (Bondarev, 1974), using an index of chemical alteration of suspended sediment (McLennan, 1993), making comparisons of sediment yields from basins subjected to different degrees of anthropogenic disturbance (Dedkov & Mozzherin, 2000), etc. The natural flux is thus estimated to lie within the range $6.0 \times 10^9$ to $13.5 \times 10^9$ t year$^{-1}$ , i.e. with more than a 2-fold range (Table 1). According to different authors, anthropogenic disturbance has raised the global sediment flux by factors of 1.7 (McLennan, 1993), 1.9 (Bondarev, 1974), 2.0 (Milliman & Syvitski, 1992), 2.6 (Dedkov & Mozzherin, 2000), and 3.0 (Harrison, 1994).




Sources of data: 1 – Kuenen (1950), 2 – Lopatin (1952), 3 – Gilluli (1955), 4 – Fournier (1960)*, 5 – Schumm (1963)*, 6 – Corbel (1964), 7 - Mackenzie & Garrels (1966)*, 8 – Holeman (1968)*, 9 – Sundborg (1973)*, 10 – Bondarev (1974), 11 – Alexeev& Lisitzina (1974), 12 – Lisitzin (1974), 13 – Gudzon (1974 – cit. in Safyanov, 1978), 14 – L'vovich (1974), 15 – Jansen & Painter (1974)*, 16 – Goldberg (1976)*, 17 – Holland (1981), 18 – Meybeck (1982 – cit. in Meybeck, 1988), 19 – Milliman & Meade (1983), 20 – Walling & Webb (1983), 21 – Gordeev (1983), 22 – Vassil'ev (1987), 23 – Meybeck (1988), 24 – Lisitzin (1991), 25 – Milliman & Syvitski (1992), 26 – McLennan (1993), 27 – Stallard (1998), 28 – Ludwig & Probst (1998), 29 – Harrison (2000), 30 – Dedkov & Mozzherin (2000).
*cit. in Walling & Webb (1996).

**Fig. 1** Estimates of the global land–ocean river sediment flux produced since the middle of the 20th century.

**Table 1** Natural and anthropogenic proportions of the global sediment flux.

| Reference | Natural ($10^9$ t year$^{-1}$) | (%) | Anthropogenic ($10^9$ t year$^{-1}$) | (%) |
|---|---|---|---|---|
| Harrison, 1994 | 6 | 33 | 12 | 67 |
| Dedkov & Mozzherin, 2000 | 6 | 38 | 9.5 | 62 |
| Bondarev, 1974 | 7.9 | 54 | 6.7 | 46 |
| Milliman & Syvitski, 1992 | <10 | 50 | >10 | 50 |
| McLennan, 1993 | 12.6 | 60 | 8.4 | 40 |

The object of this paper is to evaluate river sediment discharges over geological time on the basis of the sedimentary mass stored in the oceans. Data from the remote past can help to determine typical rates of sediment removal from the land over the Earth's history. Data from the near past could serve as an approximation for the present pristine sediment flux. Within the context of the sediment delivery concept (Walling, 1983), it is also interesting to consider what proportion of the denudation products remained on the land and how much was transferred to the oceans over the geological past. The background of geological history is seen as widening the scope of current understanding of contemporary sediment fluxes.

## GLOBAL SEDIMENT FLUX SINCE THE LATE JURASSIC

To calculate sediment fluxes, the volumes of sedimentary rocks on unglaciated land (i.e. Antarctica and Greenland excluded) and oceanic bottom given in Ronov (1993) were used. The mass of clastic marine rocks from a given epoch divided by the duration of this epoch gives the rate at which the products of mechanical denudation were being removed from the continents to the ocean. Glacial marine rocks are excluded to make the results closer to the value of river sediment discharge. Rock masses were corrected for subsequent destruction using an exponential function for observed (remaining, or survival) sedimentation rates given in (Tardy *et al.*, 1989). The obtained rates (Fig. 2) represent mechanical denudation of the continents (glacial denudation excluded), with the major part of it provided by fluvial action. Denudation rates were minimal at the boundary between the Cretaceous and the Cenozoic, which is known as a time of continental planation. In the Cenozoic, erosion rates have increased 3.5-fold, up to $9.6 \times 10^9$ t year$^{-1}$ in the Pliocene, as a response to increasing continental relief during the Alpine orogeny.

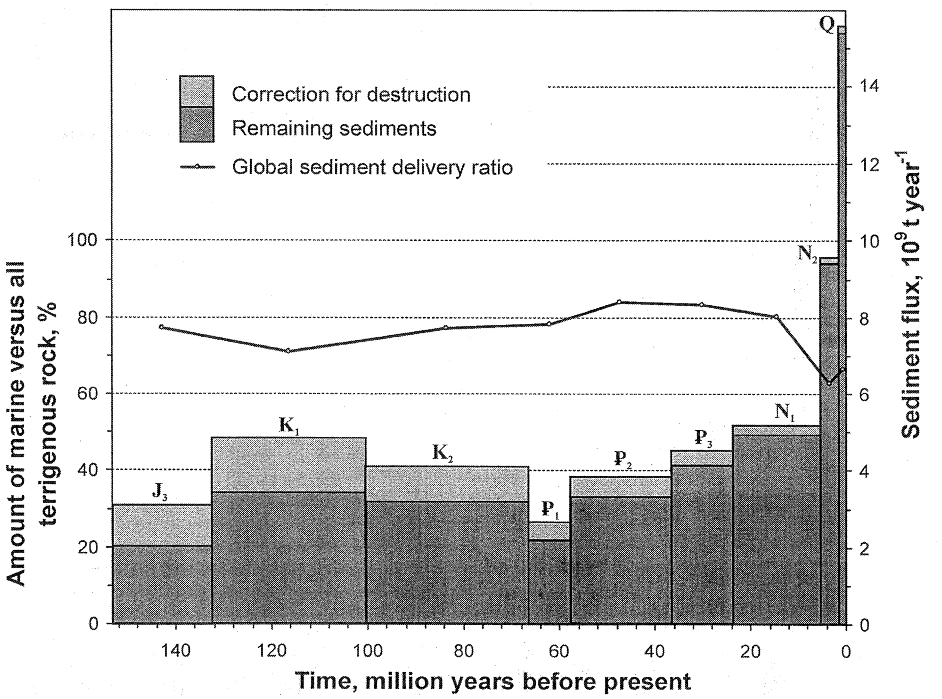



**Fig. 2** Rates of non-glacial detrital sediment input to the ocean and the global sediment delivery ratio since the late Jurassic (calculated from data in Ronov (1993) and Hay (1994)).

**Table 2** Change of sediment flux during the Late Cenozoic.

| Region | Ratio of sediment flux* | | | Source |
|---|---|---|---|---|
| | Q to $N_2$ | $N_2$ to $N_1$ | Q to $N_1$ | |
| European Alps | 1.5 | 1.7 | 2.6 | Kuhlemann *et al.* (2002) |
| European Alps | 3 | 6 | 18 | Hay *et al.* (1992—in Kuhleman *et al.*, 2002) |
| The Meuse River basin (western Europe) | 1.8 | – | – | Van Balen *et al.* (2000) |
| The Mississippi River basin | 3.9 | 1.5 | 5.7 | Hay *et al.* (1989) |

*Symbols for geological periods (absolute age in millions of years before present given in brackets): $N_1$, Miocene (23.7–5.3); $N_2$, Pliocene (5.3–1.6); Q, Pleistocene (1.6–0).

For the Pleistocene (the last 1.6 million years) the only data on global accumulation were reported by Hay (1994). His $24.7 \times 10^{21}$ g of noncarbonate marine deposits accumulated during the Quaternary represent an average land–ocean sediment flux of $15.4 \times 10^9$ t year$^{-1}$. Evidence from different regions confirms that in the Pleistocene sediment production from the continents continued to rise (Table 2). The lowest of the documented ratios (Table 2) gives a 1.5-fold increase in Pleistocene rates over those of the Pliocene and thus a lower estimate of the average sediment output in the Pleistocene of $14.5 \times 10^9$ t year$^{-1}$ which is still close to that calculated from Hay (1994).

There is less information on the Holocene sediment flux than that of the Pleistocene. In middle and high latitudes, sediment yield increased during cold epochs due to glacier action, although the area subjected to river action diminished considerably. In southeastern Asia the summer monsoon was weak during cold epochs (Colin *et al.*, 1999), thus both water and sediment discharges could be expected to decrease. For instance, the Yellow River at that time was periodically dry (Xia *et al.*, 1993). Taking into account the contemporary sediment contribution of the Asian monsoon region, the global sediment flux in the Holocene may be estimated to be close to that in the Pleistocene. This is confirmed by marine sedimentation rates reported by Hay (1994). According to Hay's data (his Tables 1.4 and 1.8 corrected for arithmetical errors), accumulation rates of noncarbonate materials in the deep sea decreased 4- to 5-fold in the Holocene compared to the Pleistocene average ($2.8 \times 10^9$ *vs* $12.9 \times 10^9$ t year$^{-1}$) because of high sea level after deglaciation. On the other hand, the same reason is responsible for the nearly two-orders of magnitude increase in accumulation on the shelves. The shelf accumulation comprises noncarbonate Holocene sediments on so-called detrital and carbonate shelves—$10.3 \times 10^9$ and $0.3 \times 10^9$ t year$^{-1}$, respectively. Holocene accumulation rates in marginal seas are highly variable but Hay (1994) suggests that typical rates were 20–25% of those in the Pleistocene, i.e. $0.5–0.6 \times 10^9$ t year$^{-1}$. Overall, Holocene marine accumulation of noncarbonate sediments, which can be regarded as detrital flux from the land, represents some $14 \times 10^9$ t year$^{-1}$—almost the same as the Pleistocene average given above.

This approximation comprises contributions from different agents of mechanical denudation. Non-fluvial sources may be taken into account by considering their contemporary levels. The aeolian dust deposition over the ocean area has been assessed as (in $10^9$ t year$^{-1}$): 0.06 (Garrels & Mackenzie, 1971), 0.07–0.35 (Goldberg, 1971), 0.5–0.85 (Prospero, 1981), 0.8 (Vassil'ev & Pomortsev, 1988), 0.9 (Duce *et al.*, 1991), 1.6 (Lisitzin, 1974), 2.0 (Bondarev, 1974). Direct glacial input has been estimated at ($10^9$ t year$^{-1}$): 0.3–0.4 (Losev, 1989), 1.5 (Lisitzin, 1974), 1.9 (Makkaveev, 1982). Abrasion of the ocean coast may add ($10^9$ t year$^{-1}$): 0.15 (Gilluly, 1955), 0.2 (from Kuenen's (1950) 0.12 km$^3$ year$^{-1}$), 0.3–0.4 (Losev, 1989), 0.7–1.0 (Bondarev, 1974), 1.0 (Safyanov, 1978). Taking the averages in each case (i.e. $0.9 \times 10^9$, $1.2 \times 10^9$ and $0.5 \times 10^9$ t year$^{-1}$, respectively) and subtracting them from the total Holocene flux of $14 \times 10^9$ t year$^{-1}$, one may estimate the Holocene fluvial sediment flux to the oceans at $(11–12) \times 10^9$ t year$^{-1}$. This range may represent the most probable estimate of the natural fluvial flux, i.e. the global river sediment discharge to the ocean that would occur under undisturbed conditions. It is close to the value of natural sediment flux suggested by McLennan (1993), but higher than that suggested by other workers (Table 1).

## CHANGES OF RIVER SEDIMENT FLUXES DUE TO HUMAN ACTIVITY

Since the above estimate of the natural sediment flux lies towards the bottom of the range of estimates of contemporary sediment flux proposed during the last two decades (Fig. 1), human activity is likely to have increased global sediment input to the oceans to some extent. Data on sedimentation mass balances provide clearer estimates for some large river basins. The two major contributors of sediment to the world's oceans are of primary interest, namely the Yellow (Huang He) River and the Ganges-Brahmaputra system.

The sediment discharge of the Yellow River is suggested to exhibit the same behaviour as sedimentation rates in the lower Yellow River region, according to Xu (1998): between 12 000 (beginning of the Holocene) and 6000 years ago the sediment flux is taken as constant ($S_0$), and during the next 4000 years it doubled by 2.2 ka BP (2 $S_0$). After that sediment discharge is supposed to be proportional to population in the middle Yellow River region. Three models have been explored, namely a linear, exponential and logarithmic-law increase of sediment discharge as a function of population increase since 200 BC, the two former giving almost similar results. The change of sediment discharge during the Holocene is calculated by means of optimization of the variable $S_0$, assuming a contemporary sediment discharge $S_m = 1.2 \times 10^9$ t year$^{-1}$, to derive the total Yellow River-derived Holocene marine sediment mass of $3000 \times 10^9$ t given in Milliman *et al.* (1987).

During the Holocene the Yellow River changed its course several times, flowing for some 30% of the time into the Gulf of Bohai, and 70% to the south of the Shandong Peninsula. However, some $1330 \times 10^9$ t (45%) of the river-derived marine deposits may be attributed to the northern position of the river mouth and only $1680 \times 10^9$ t (55%) belongs to the southern position (based on Table 2 in Milliman *et al.* (1987)). The closest distribution of marine sediments is provided by the linear erosion–population model. According to this scenario 2200 years ago at the beginning of sensible land disturbance, sediment discharge was as much as $0.35 \times 10^9$ t year$^{-1}$. This is only two times less than the mean sediment discharge since 1855, which may be estimated at between 0.6 to $0.75 \times 10^9$ t year$^{-1}$ on the basis of data presented by Shi *et al.* (2003).

Less evident is the increase in sediment load at the Ganges-Brahmaputra mouth caused by human activity. Kuehl *et al.* (1997) report that the subaqueous part of the Ganges-Brahmaputra delta accumulated some $1.97 \times 10^{12}$ t of sediment during the last 7000 $^{14}C$ years (8000 calendar years). The sediment amount leaving the shelf have been assessed from the core data stratigraphy in the Deep Bengal Fan presented by Colin *et al.* (1999), the dimensions of the fan measured on the map from Einsele *et al.* (1996). The total sediment volume over the subaqueous delta and Deep Bengal Fan provides an estimate of the mean sediment delivery to the sea of $440 \times 10^6$ t year$^{-1}$ during the last 8000 years. This seems to be an underestimate, since the cores of Colin *et al.* (1999) were located at the eastern edge of the Bengal Fan while a considerable portion of the sediment is transported via the so called Swatch of No Ground Canyon to the west. Michels *et al.* (1998) found that at a decadal and century scale the subaqueous part of the delta accumulates 20% of total load and 35–50% is absorbed by the canyon, so the ratio is 1 : 1.75–2.5 in contrast to the ratio of 2 : 1, which follows from the above estimate. If the ratio found by Michels *et al.* (1998) is spread over the Holocene, the sediment amount in the Bengal Fan would be 3.5–4 times as much as suggested above, and the river sediment input to the Bengal Bay would be calculated as $(0.60–0.75) \times 10^9$ t year$^{-1}$.

Recent estimates of the modern suspended load of the joint Ganges-Brahmaputra system are $1.06 \times 10^9$ (Milliman & Syvitski, 1992) or $1.14 \times 10^9$ t year$^{-1}$ (Delft Hydraulics/DHI, 1994—cit. in Michels *et al.*, 1998), with an average of $1.1 \times 10^9$ t year$^{-1}$. Flood plain accumulation over the subaerial delta is estimated from 30–45% (Michels *et al.*, 1998) to 40–80% (Milliman & Syvitski, 1992) of the total Ganges-Brahmaputra suspended load. To make it comparable to the ocean sedimentation rate one must also account for sediment losses. If the whole range of loss estimates is taken, the sediment delivery to the sea ranges from $0.22 \times 10^9$ to $0.77 \times 10^9$ t year$^{-1}$. When this range is compared with the lower and higher estimates derived from the Holocene marine sedimentation rates of $0.44 \times 10^9$ and $0.75 \times 10^9$ t year$^{-1}$, respectively, the existing sedimentation data provide the possibility of both an increase or decrease of the contemporary sediment contribution to the sea relative to the average Holocene rate.

Other large Asian rivers give less evidence of human-induced increases in sediment flux, for example, the Yangtze (Changjiang) River. An analysis of delta development undertaken by Chen (1996) reveals that rapid deltaic progradation during the last 2000 years is not due to anthropogenic activity, as had been previously thought, but results from river long profile adjustment after deep incision during the last glaciation. Chen (1996) also affirms that there is no evidence of any systematic increase of sediment discharge during the last 2000 years, or a significant difference in sediment discharge between the two periods before and after 2000 years before the present. The Mekong River, the tenth-ranked river for sediment flux (Milliman & Meade, 1983) also shows no evidence of a dramatic increase in sediment discharge due to human activity (Ta *et al.*, 2002). On the basis of sediment volume analysis they estimated the average sediment discharge of the Mekong River for the last 3000 years to be $(144 \pm 36) \times 10^6$ t year$^{-1}$ which is only 10% less than at present. In the case of the seventh-ranked Mississippi River the present-day sediment load of $210 \times 10^6$ t year$^{-1}$ (Milliman & Meade, 1983) is almost half that for the second half of the 19th century reconstructed by Kesel *et al.* (1992), due to sediment trapping in reservoirs.

Data from big rivers reveal that, with the exception of the Yellow River, large basins seem to contribute little to the anthropogenic sediment flux increase. In contrast, sediment discharge from rather small basins is found to respond strongly to land clearing and other human impact. In addition to numerous examples from inland areas (cf. Dearing & Jones, 2003), strong human impacts on near-shore sedimentation have long been reported from many coastal regions supplied with sediment by small and medium-sized rivers draining mountainous or hilly terrain. Wellington Harbour, New Zealand (35 km$^2$) may serve as an example. Data from several cores taken within 2 km of the coast reveal a significant increase in sedimentation rates since the end of the 19th century (Goff, 1997). Marine sedimentation before European colonization in the middle of the 19th century occurred at rates of 0.3–3.0

mm year$^{-1}$ at a millennial to century scale, while the average rates for the 20th century are 10–50 mm year$^{-1}$, with increases in individual cores being in the range of 10- to 100-fold.

## THE LAND–OCEAN SEDIMENT DELIVERY RATIO

The amounts of marine and continental detrital rock accumulated during a certain epoch not only reflect the rate of mechanical denudation but also the efficiency of transportation processes, namely the proportion of denudation products that reached the ocean and the proportion retained on the land. The marine component of all detrital rock may be regarded as a global sediment delivery ratio (*SDR*) from the land to the ocean during the given epoch, the majority being accomplished by fluvial action. According to Ronov (1993), during the last 150 million years some 67 220 × 10$^3$ km$^3$ of continental and 220 430 × 10$^3$ km$^3$ of marine clastic sedimentary rocks have been stored all over the Earth (to be exact, these are the volumes that have survived of what had ever accumulated). So the global *SDR* for this timescale is 77%. In the same way the global *SDR* values may be calculated for each geological epoch (Fig. 2). As no age-dependent trend is detected, information losses probably exercise little influence on the *SDR* values and their changes in time reflect the conditions for sediment transport from the continents to the oceans. Sediment delivery was about or higher than average when the sediment flux was low corresponding to planation of continental relief. In contrast, increased relief leads to increasing buffering capacity (term suggested by Walling, 1999): mountain building is accompanied by foredeep subsidence, which creates potential storage, and accounts for much sediment loss.

In the Pliocene the global *SDR* value decreased to its minimal value of 63%, and a slightly higher value is proposed for the Pleistocene (Table 3). The global *SDR* decreased in the Plio-Pleistocene, probably due to the increasing extent of inland areas with no discharge to the ocean. In any case this is much higher than modern *SDR*s obtained at a decadal or century scale. In small basins they are highly variable but still show a trend for rapid decrease with increasing basin area (Fig. 3). Both globally and over vast regions, values of

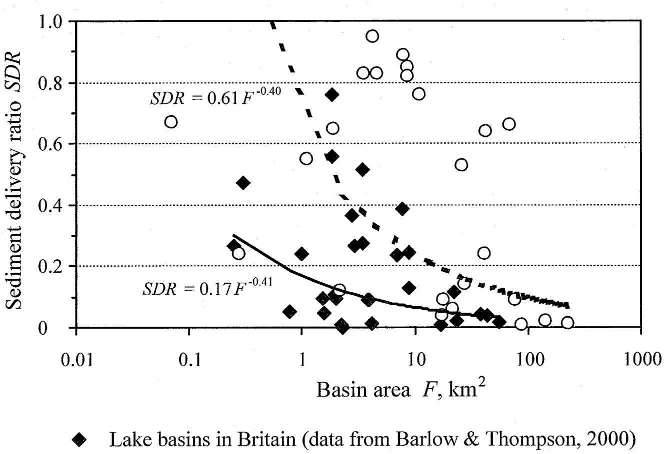



**Fig. 3** Sediment delivery ratios in small basins at a decadal to century time scale.

**Table 3** The land–ocean sediment delivery ratio (*SDR*) at different temporal scales (calculated from original data of indicated authors).

| Region | Period, years | *SDR* | Source |
|---|---|---|---|
| Global land* | last $153 \times 10^6$ | 77% | This paper |
| Himalayas | last $20 \times 10^6$ | 74% | Eincele *et al.*, 1996 |
| European Alps | last $5.5 \times 10^6$ | 90–95% | Kuhleman *et al.*, 2002 |
| European Alps | last $5.5 \times 10^6$ | 90% | Hay *et al.*, 1992 (in Kuhleman *et al.*, 2002) |
| Global land* | $(5.3 \div 1.6) \times 10^6$ (the Pliocene) | 63% | This paper |
| Global land* | last $1.6 \times 10^6$ (the Pleistocene) | 65–69% | Hay, 1994 |
| Meuse River Basin (Western Europe) | last $0.65 \times 10^6$ | 83% | Van Balen *et al.*, 2000 |
| United States | $n \times (10 \div 100)$ | 4–5% | Holeman, 1980 |
| Global land* | $n \times (10 \div 100)$ | 13–20% | Walling, 1999, Golosov, 2001 |
| Global land* | $n \times (10 \div 100)$ | 5–10% | Dearing & Jones, 2003 |

* Antarctica and Greenland not included.

modern *SDR* range, according to different authors, between 5 and 20%, i.e. about an order of magnitude less than that for geological epochs (Table 3). Essential differences between the modern and geological-scale *SDR*s result from repeated redeposition and remobilization of each sediment particle on its way from the initial source to the basin outlet. At any given time the river sediment load comprises only partly new erosion products, the remainder being supplied from different sediment sinks within the basin, namely fans, floodplains, etc. The ratio between these sediment sources is still unknown in large rivers and scarcely known in small rivers.

The delivery ratio seems to be a function of time, with the residence time of different sinks playing a key role. This subject is also poorly investigated, although it appears that the longer the residence times, the slower the sediment movement through the fluvial system and the lower the *SDR* value will be for small time spans. At time scales greater than the basin residence times, the *SDR* value only reflects long-term storage such as active tectonic depressions. This is seen from the fact that modern *SDR*s have essentially higher values in mountain areas than on plains (Walling, 1983) while over geological time scales both mountain and plain basins evidence high delivery ratios (Table 3, Himalayas and Meuse River basin).

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## REFERENCES

Alexeev, V. V. & Lisitzina, K. N. (1974) *Stok vzveshennykh nanosov* (Suspended Sediment Flux). Gidrometeoizdat Publ., Leningrad, Russia (in Russian).

Barlow D. N. & Thompson R. (2000) Holocene sediment erosion in Britain as calculated from lake-basin studies. In: *Tracers in Geomorphology* (ed. by I. D. L. Foster), 455–472. Wiley, Chichester, UK.

Bondarev, L. G. (1974) *Vechnoe Dvizhenie: Planetarnoye Peremeschenie Veschestva i Chelovek* (*Perpetual Motion: Planetary Matter Moving and the Man*). Mysl' Publ., Moscow, Russia (in Russian).

Chen, X. (1996) An integrated study of sediment discharge from the Changjiang River, China, and the delta development since the mid-Holocene. *J. Coast. Res.* **12**(1), 26–37.

Colin, C., Turpin, L., Bertaux, J., Desprairies, A. & Kissel, C. (1999) Erosional history of the Himalayan and Burman ranges during the last two glacial–interglacial cycles. *Earth and Planetary Science Letters* **171**, 647–660.

Corbel, J. (1964) L'erosion terrestre, étude quantitative (methods, techniques, resultats). *Ann. Geogr.* **73**(398), 385–412.

Dearing, J. D. & Jones, R. T. (2003) Coupling temporal and spatial dimensions of global sediment flux through lake and marine sediment records. *Global and Planetary Change* **39**, 147–168.

Dedkov, A. P. & Mozzherin,V. I. (2000) Globalniy stok nanosov v okean: prirodnaya i antropogennaya sostavlyayuschie (Global river sediment discharge into the ocean: Natural and anthropogenic components), 15–23. Erosionnye i Ruslovye Processy. Vypusk 3. Moscow University Publ., Moscow, Russia (in Russian).

Duce, R. A., Liss, P. A., Merrill, J. T. *et al.* (1991) The atmospheric input of trace species to the world ocean. *Global Biogeochem. Cycles* **5**, 193–259.

Einsele, G., Ratschbacher, L. & Wetzel, A. (1996) The Himalaya-Bengal fan denudation-accumulation system during the past 20 Ma. *J. Geol.* **104**, 163–184.

Garrels, R. M. & Mackenzie, F. T. (1971) *Evolution of Sedimentary Rocks*. Norton, New York, USA.

Gilluli, J. (1955) Geologic contrasts between continents and ocean basins. *Geol. Soc. Am. Spec. Pap.* **62**, 7–18.

Goff, J. R. (1997) A chronology of natural and anthropogenic influences on coastal sedimentation, New Zealand. *Mar. Geol.* **138**, 105–117.

Goldberg, E. R. (1971) *Atmospheric Transport. Impingement of Man on the Oceans*. Wiley, New York, USA.

Golosov, V. N. (2001) Pereraspredelenie nanosov v verkhnikh zvenyakh fluvial'noy seti zemledel'cheskykh regionov: teoria voprosa i opyt regional'nogo analysa (Sediment redistribution in upper courses of the fluvial net in agricultural regions: Theory and regional analysis on the plains of the temperate belt), 94–119. Erosia Pochv i Ruslovye Processy. Vypusk **13**. Moscow University Publ., Moscow, Russia, (in Russian).

Golosov, V. N. (2002) Temporal–spatial variations in the sediment delivery ratio of small drainage basins: the Russian Plain example. In: *The Structure, Function and Management Implications of Fluvial Sedimentary Systems* (ed. by F. J. Dyer, M. C. Thoms & J. M. Olley), (Proc. Alice-Spring Sypm. August 2002), 345–353. IAHS Publ. 276. IAHS Press, Wallingford, UK.

Gordeev, V. V. (1983) *Rechnye Vynosy v Mirovoy Ocean i ikh Khimicheskiye Characteristiki* (*River Inputs to the Oceans and their Chemical Characteristics*). Nauka Publ., Moscow, Russia (in Russian).

Harrison, C. G. A. (1994) Rates of continental erosion and mountain building. *Geol Rundsch* **83**, 431–447.

Harrison, C. G. A. (2000) What factors control mechanical erosion rates? *Int. J. Earth Sci.* **88**, 752–763.

Hay, W. W., Shaw, C. A. & Wold, C. N. (1989). Mass-balanced paleogeographic reconstructions. *Geol. Rundschau* **78**, 207–242.

Hay, W. W. (1994) Pleistocene—Holocene fluxes are not the Earth's norm. In: *Material Fluxes on the Surface of the Earth* (ed. by W. W. Hay) 15–27. National Academy Press, Washington, DC, USA.

Holeman, J. N. (1980) Erosion rates in the U.S. estimated by the Soil Conservation Service's inventory. *Am. Geophys. Union Trans.* **61**, 954.

Holland, H. D. (1981). River transport to the oceans. The Sea. In: *The Ocean Lithosphere*, vol. 7 (ed. by C. Emiliani), 763–800. Wiley, New York, USA.

Kesel, R. H., Yodis, E. G. & McCraw, D. J. (1992) An approximation of the sediment budget of the lower Mississippi river prior to major human modification. *Earth Surf. Processes Landf.* **17**, 711–722.

Kuehl, S. A., Levy, B. M., Moore, W. S. & Allison, M. A. (1997) Subaqueous delta of the Ganges-Brahmaputra river system. *Mar. Geol.* **144**, 81–96.

Kuhlemann, J., Frisch, W., Székely, B., Dunkl, I. & Kázmér, M. (2002) Post-collisional sediment budget history of the Alps: tectonic versus climatic control. *Int. J. Earth Sci.* (*Geol. Rundsch.*) **91**, 818–837.

Kuenen, P. H. (1950) *Marine Geology*. Wiley, Chichester, UK.

Lisitzin, A. P. (1974) *Osadkoobrazovanie v Okeanakh* (*Oceanic Sedimentation*). Nauka Publ., Moscow, Russia (in Russian).

Lisitzin, A. P. (1991) *Processy Terrigennoy Sedimentacii v Moryakh I Oceanakh* (Terrigenous Sedimentation in Seas and Oceans). Nauka Publ., Moscow, Russia (in Russian).

Lopatin, G. V. (1952) *Nanosy Rek SSSR: Obrazovanie i Perenos* (Sediments in the USSR Rivers: Origin and Transport). Geografgiz Publ., Moscow, USSR (in Russian)

Losev, K. S. (1989) *Voda* (The Water). Gidrometeoizdat Publ., Leningrad, Russia (in Russian).

L'vovich, M. I. (1974) *Mirovye Vodnye Resursy i Ikh Buduschee* (World Water Resources and Their Future). Mysl' Publ., Moscow, Russia (in Russian).

Ludwig, W. & Probst, J. -L. (1998) River sediment discharge to the oceans: present-day controls and global budgets. *Am. J. Sci.* **298**, 265–295.

Makkaveev, N. I. (1982) Denudacionnaya sostavlyauschaya balansa veschestva v sisyeme ocean—susha I yeye rol' v formirovanii peneplenov (Denudation as a component in the land-ocean mass balance and its role in peneplain formation). *Vodnye resursy* **3**, 147–155 (in Russian).

McLennan, S. M. (1993) Weathering and global denudation. *J. Geol.* **101**, 295–303.

Meybeck, M. (1988) How to establish and use world budgets of riverine materials. In: *Physical and Chemical Weathering in Geochemical Cycles* (ed. by A. Lerman), 247–272. Kluwer, Dordrecht, The Netherlands.

Michels, K. H., Kudrass H. R., Hübscher, C., Suckow A. & Wiedicke, M. (1998) The submarine delta of the Ganges–Brahmaputra: cyclone-dominated sedimentation patterns. *Mar. Geol.* **149**, 133–154.

Milliman, J. D. & Meade, R. H. (1983) World-wide delivery of river sediment to the oceans. *J. Geol.* **91**, 11–21.

Milliman, J. D. & Syvitski, J. P. M. (1992) Geomorphic/tectonic control of sediment discharge to the ocean: the importance of small mountainous rivers. *J. Geol.* **100**, 525–544.

Milliman, J. D., Qin, Y. S. & Ren, M. E. Y. (1987) Man's influence on the erosion and transport of sediment by Asian rivers: the Yellow River (Huanghe) example. *J. Geol.* **95**, 751–762.

Prospero, J. M. (1981) Eolian transport to the world ocean. In: *The Sea. The Ocean Lithosphere*, vol. 7 (ed. by C. Emiliani), 801–874. Wiley, New York, USA.

Ronov, A. B. (1993) *Stratisfera ili Osadochnaya Obolochke Zemli: Kolichestvennoye Issledovanie* (*Stratisphere, or the Sedimentary Sphere of the Earth: Quantitative Research*). Nauka Publ., Moscow, Russia (in Russian).

Safyanov, G. A. (1978) *Beregovaya Zona Oceana v 20-m veke* (The Oceanic Coastal Zone in the 20th Century). Mysl' Publ., Moscow, Russia (in Russian).

Stallard, R. (1998) Terrestrial sedimentation and the carbon cycle: coupling weathering and erosion to carbon burial. *Global Biogeochemical Cycles* **12**, 231–257.

Ta, T. K. O., Nguyen, V. L., Tateishi, M., Kobayashi, I., Tanabe, S. & Saito, Y. (2002) Holocene delta evolution and sediment discharge of the Mekong River, southern Vietnam. *Quatern. Sci. Rev.* **21**, 1807–1819.

Tardy, Y., N'Kounkou, R. & Probst, J.-L. (1989) The global water cycle and continental erosion during Phanerozoic time. *Am. J. Sci.* **289**, 455–483.

Van Balen, R. T., Houtgast, R. F., Van der Wateren, F. M., Vandenberghe, J. & Bogaart, P. W. (2000) Sediment budget and tectonic evolution of the Meuse catchment in the Ardennes and the Roer Valley Rift System. *Global Planetary Change* **27**, 113–129.

Vassil'ev, V. P. (1987) Tverdiy stok v Mirovoy ocean (Solid flux into the World Ocean). *Litologia i Poleznye Iskopaemye* **6**, 19–28 (in Russian).

Vassil'ev, V. P. & Pomortsev, O. A. (1988) Aridnye zony v oceane: ikh granitsy I vklad eolovogo materiala v oceanskuyu sedimentaciyu (Arid zones in the ocean: Their borders and the aeolian contribution to the oceanic sedimentation). *Litologia I Poleznye Iskopaemye* **6**, 3–16 (in Russian).

Walling, D. E. (1983) The sediment delivery problem. *J. Hydrol.* **65**, 209–237.

Walling, D. E. (1999) Linking land use, erosion and sediment yields in river basins. *Hydrobiologia* **410**, 223–240.

Walling, D. E. & Webb, B. W. (1983) Patterns of sediment load. In: *Background to Palaeohydrology* (ed. by K. J. Gregory), 69–100. Wiley, Chichester, UK.

Walling, D. E. & Webb, B. W. (1996) Erosion and sediment yield: a global overview. In: *Erosion and Sediment Yield: Global and Regional Perspectives* (ed. by D. Walling & B. Webb) (Proc. Exeter Symp., July 1996), 3–19. IAHS Publ. 236. IAHS Press, Wallingford, UK.

Xia, D. X., Wu, S. Y. & Yu, Z. (1993) Changes of the Yellow River since the last glacial age. *Mar. Geol. Quatern. Geol.* **13**, 83–88.

Xu, J. (1998) Naturally and anthropogenically accelerated sedimentation in the lower Yellow River, China, over the past 13 000 years. *Geogr. Ann.* **80A**, 67–78.