

Movement of bed forms and sediment yield of rivers

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Abstract Factors influencing the characteristics of suspended load yield and bed load yield were studied. Estimation methods of bed load yield for unstudied rivers, which are mainly composed of sandy bed deposits, were substantiated. Correlation between suspended load yield and bed load yield for the rivers which flow in different natural conditions was analysed.

Key words bed load; channel relief; sediment yield; suspended load

INTRODUCTION

Transportation of water, bed and suspended (wash) load, chemical substances, biological substances and heat, is constantly taking place along rivers. Oscillations in matter streams along rivers and in time, cause many dangerous processes influencing the safety of population, economics, and conditions of the existence for aquatic biota. In particular, their trend and intensity mainly depends on sediment yield. Sediment yield (W) includes two components: suspended load W_R and bed load W_G , i.e. $W = W_R + W_G$. Suspended load is composed of small fractions (mud, clay, dust, sand). The bed load component also includes fractions of gravel, pebbles and boulders. In different natural conditions the correlation between the components of sediment yield is rather variable and affects the character of bed deformations.

FACTORS OF SUSPENDED LOAD VARIATION

Suspended load yield depends on the mean long-term discharge Q_o and sediment concentration S , i.e. $W_R = pSQ_o$ where p is the duration of a year (3.15×10^6 s). Sediment concentration depends on the zonal and azonal natural factors, anthropogenic impact on river basin and the channel. When the drainage area is the same the increase of water turbidity is observed from humid to arid areas, from mountains to plains. It causes complicated spatial changes of suspended load yield W_R , mean long-term discharge Q_o and sediment concentration in rivers S . In similar natural conditions ($S = const$) suspended load yield mainly depends on the stream order (N) where N is the function of the number of the tributaries (L) with a length of 10–15 km. When $L = 1, 2-3, 4-7, 8-15, 16-31, \dots 512-1023, 16384-32767$, then $N = 1, 2, 3, 4, 5 \dots 10 \dots 15$ (Alekseevskiy, 1998). The greater N is the larger the drainage basin and the values Q_o and W_R accordingly. This regularity is quite often broken, especially on large rivers crossing several natural zones.

Characteristics of suspended load yield undergo long-term, annual, seasonal and daily oscillations. It is a result of the specific rhythm of denudation products inflow from the river basin. Depending on the combination of the water balance components of the territory and

the variability of natural conditions of the drainage basin, more or less washing products come into the rivers that determine the variation of water runoff, and suspended load yield.

The analysis of spatial and temporal variability of suspended load shows that a large number of natural processes exist, which influence the value and conditions of mineral particles movement. For some period of time:

$$W_R = \sum_i^n W_{Ri} = W_{R1} + W_{R2} + \dots + W_{Rn} \quad (1)$$

where W_{Ri} is the impact of certain natural processes into the total suspended load yield. The genetic role of these processes is seen after transformation of the equation (1):

$$1 = \frac{W_{R1}}{W_R} + \frac{W_{R2}}{W_R} + \dots + \frac{W_{Rn}}{W_R} \quad (2)$$

When the expression of the type $W_{R\gamma}/W_R$ could be named as the coefficient of the genetic value $\alpha_\gamma = W_{R\gamma}/W_R$ and consequently for some period of time $\sum_i \alpha_i = 1$. The change of α_γ is

the result of the specific way of sediment origin and temporal averaging of W_R . When $\Delta t \rightarrow \infty$ all sources of sediment yield change for a certain section of the river can be taken into account using equation (1). The value of separate sources can depend on certain natural conditions. Differences in frequency of appearing of such conditions can lead to the situation when the period Δt is completed, γ process hasn't yet effected W_R , i.e. $\alpha_\gamma = 0$. In the river mouths sediment yield decreases up to zero and consequently the coefficient of genetic value of accumulation processes $\alpha_A = -1$. *Vice versa*, in the upper part of the slopes where the washing out of soil is the main source of sediment yield the coefficient $\alpha_c = 1$. If natural processes for Δt contribute the increase of W_R : $0 \leq \alpha_\gamma \leq 1$. In general case: $-1 \leq \alpha_\gamma \leq 1$.

FEATURES OF BED LOAD TRANSPORT

The movement of individual particles depends on the hydraulic characteristics of streams and particle size of alluvial deposits. The greater the stream velocity is the higher the probability of movement for sandy, gravel and other coarse particles on the bottom. When the stream velocity $v < v_H$ where v_H is the critical velocity, they are stable and bed load discharge $G = 0$. If $v > v_H$ then $G > 0$. Depending on the ratio between v and v_H bed load particles either roll on the bottom surface or saltate. Both forms can coexist in time and can form a cycle of the process when the smooth and bed form phases of bed relief change each other successively. The increase of v causes transition from the stable state of the particles in the surface layer of sediments to saltation and formation of the movable elements of bed relief. As a result the structure of bed load transport is changed since the total discharge $G = G_1 + G_2$ where G_1 is bed load caused by shifting of relatively small bed forms (of the length l_g) and G_2 is bed load due to particle movement in the form of larger elements of bed relief and saltation length $l > l_g$. The value $G_1 \rightarrow 0$ and $G \approx G_2$ in the process of washing out of the smaller bed form. The process is repeated (Goncharov, 1962) and when G_2 decreases larger bed forms are formed, resulting in the increase of G_3 and so on. As a result, a complicated relation between the total discharge and bed load discharge G_3 appears due to the form shifting being in close correspondence with hydraulic condition of the stream. The total

discharge of bed load is greater than G_i where i is the index of the bed relief form type. According to Rossinsky & Debolsky (1980) $G_i = 0.7G$. The coefficient depends on the mean diameter of bed deposition and the relative depth of the channel h/b_p , where b_p is the width of the stream.

Variation of the conditions of the flow and river deposits interaction causes the change of form parameters. The increase of water discharge and stream velocity leads to the increase of their length and height. When the levels become lower the new elements of bed relief do not always manage to reflect the whole cycle of the opposite changes. Small forms corresponding to hydraulic conditions of low water and large bed forms preserved from the previous high water regime can occur in bed relief at the same time (Fig. 1). Bed relief forms of different size move both actively and passively (Znamenskaya, 1968). The active movement is connected with particle transport at the distance $l = l_{gi}$. The passive movement is characterized by the movement of large forms by shifting of smaller forms along their surface. This creates a complicated structure of bed load transport in different water regime phases, when different kinds of bed relief elements move by active and/or a passive shifting mechanism. The longer the bed form, the lower the probability of its active shifting and the higher "the weight" of its passive shifting. In contrast, the smallest forms only migrate by active shifting. The forms of intermediate size undergo active and passive shifting from time to time.

A STRUCTURE OF BOTTOM RELIEF

Side bars (point bars) are the largest bed forms (A type) in river channel relief structure (Fig. 1). Their length is in proportion to the stream order, the channel width (Popov, 1965). The macroform is distinguished by the typical change of the bottom slope (μ): (i) $\mu = 0$ at the beginning of the macroform; (ii) $\mu < 0$ at the upper part of the form; (iii) $\mu = 0$ at its crest; $\mu > 0$ at the lower part of the form, $\mu = 0$ at the beginning of the next macroform. In a similar way, forms of smaller linear size could be distinguished. The method, based on the

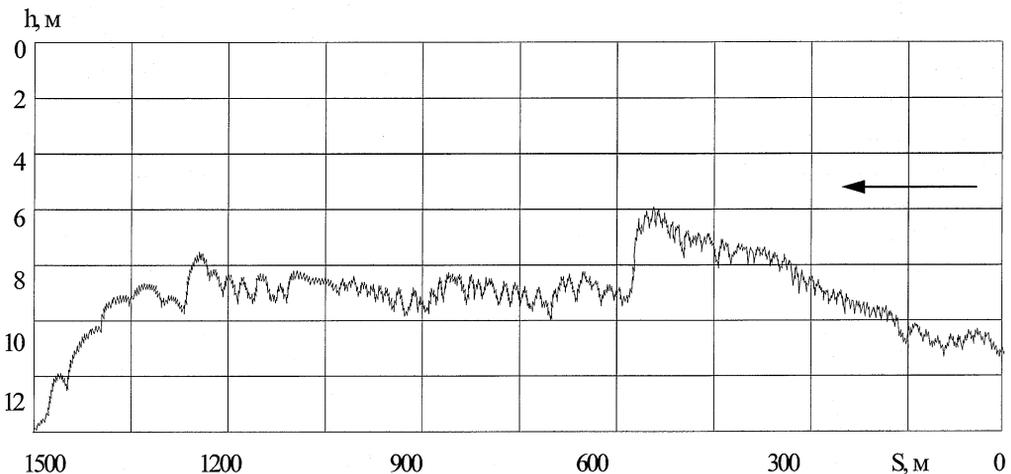


Fig. 1 Hierarchy of bed forms (S , the distance; h , depth of the stream).

double change of sign of bed slope along the form is repeated until the form of the smallest length is determined.

The forms of five types exist in the hierarchy of bottom forms despite the river size. To escape wrong interpretation of the terms we suggest that a letter index of the forms should be used: the largest form A coincides with the length of macroform (point bar), which depends on the channel width and the stream order N . One or two forms of Б type can exist on the surface of this macroform. Б type forms develop on the surface of Б type form and finally Г and Д forms are situated on their surface. The smallest Д-form type (microform) has the length $l_g \leq h$ where h is the stream depth.

The height of different type forms is a perpendicular to the form bottom from its crest (full form height h_g) or the distance from the bottom of the given type to the bottom of the next form (Fig. 2). For example, the proper height of A-form (h_{gs}) is equal to the thickness of the deposit layer lower the bottom of Б form type. Between $(h_{gs})_i$ and $(h_g)_i$ there is a connection:

$$(h_{gs})_i = \beta_i (h_g)_i \tag{3}$$

Where β_i is the adaptation coefficient for the i form type. It is accordingly equal to 0.55; 0.31 and 0/39 for A, Б and Б forms type. The proper and full height for Г and Д forms coincide, for them $\beta = 1$ since the shifting of the smallest bed forms is quick and does not effect the difference between the full and proper height of Г form type.

There is a certain correspondence between form parameters of every type and the stream order. The treatment of natural data shows a close statistical correlation ($\sigma > 0.75$) between $(h_g)_i$ and N , where N is the order of the river. The height of the relief forms of all types, which are in accordance with the hydraulic regime of the stream, is proportional to N (Table 1). The increase of N results in the nonlinear increase of the height of bed forms of one type in accordance with the equation:

$$(h_g)_{ij} = a_{ij} N^{bij} \tag{4}$$

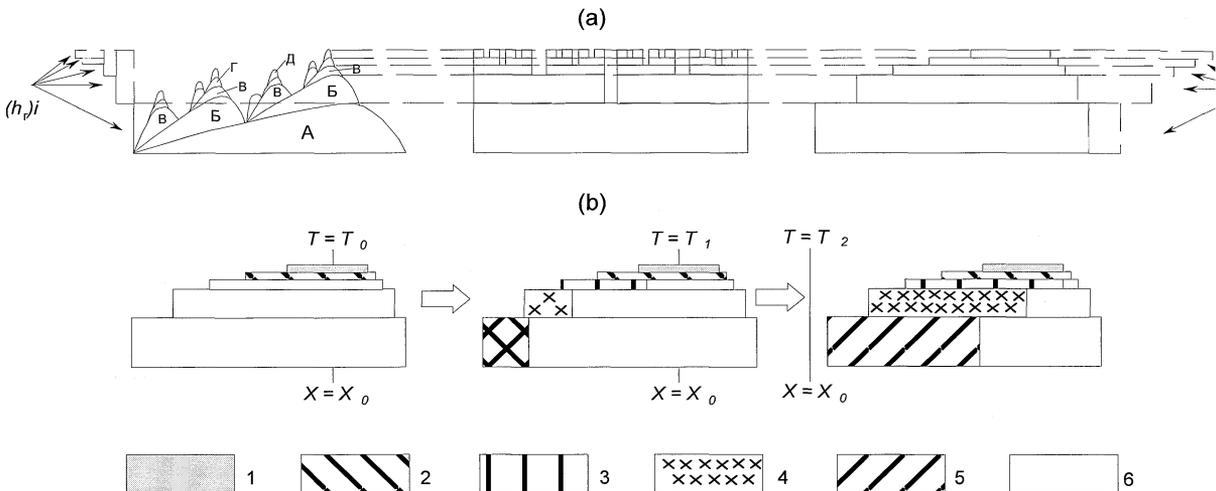


Fig. 2 The structure of relief (a) and the scheme of active (1,2,3,4,5) and passive (b) migration of bedforms. T_0, T_1, T_2, X_0 – time and spatial coordinates.

Table 1 Parameters of equation (4).

Water season	Parameter	Type of forms, <i>i</i>				
		A	Б	B	Г	Д
High water (<i>j</i> = 1)	<i>a</i>	0.046	0.018	0.032	0.023	0.005
	<i>ε</i>	1.95	2.0	1.5	1.4	1.9
Low water (<i>j</i> = 2)	<i>a</i>	0.046	0.018	0.008	0.005	0.003
	<i>ε</i>	1.95	2.0	2.0	2.0	2.0

where a_{ij} , b_{ij} are the empirical parameters, j is the index of water season.

The forms of bed relief move along the stream at some velocity C_g . Analysis of shifting of A, Б, B, Г and Д forms brings to the conclusion about the seasonal character of the relation between shifting velocity (C_g) and the river order N . The variability of relative shifting velocities of forms of different type depends on the stream order (Fig. 3). For one type of forms the shifting velocity increases, for other it decreases, when N increases. For A and B forms the dependence between the variables is of increasing character and for Б, Г and Д forms of decreasing character. In general case they look like:

$$\frac{C_{u_{ij}}}{v_j} = r_{ij} N^{p_{ij}} \tag{5}$$

The coefficients r_{ij} and p_{ij} characterize the influence of form type and water season phase on this form type mobility. In low water the greater part of large forms is moving passively due to active shifting of small forms. Observations on many rivers show that only Д and Г forms undergo such a shifting. The velocity of this process for the forms of both types is the same, hence:

$$C_{g_{\Gamma, \Delta}} = 600v_m N^{-2.0} \tag{6}$$

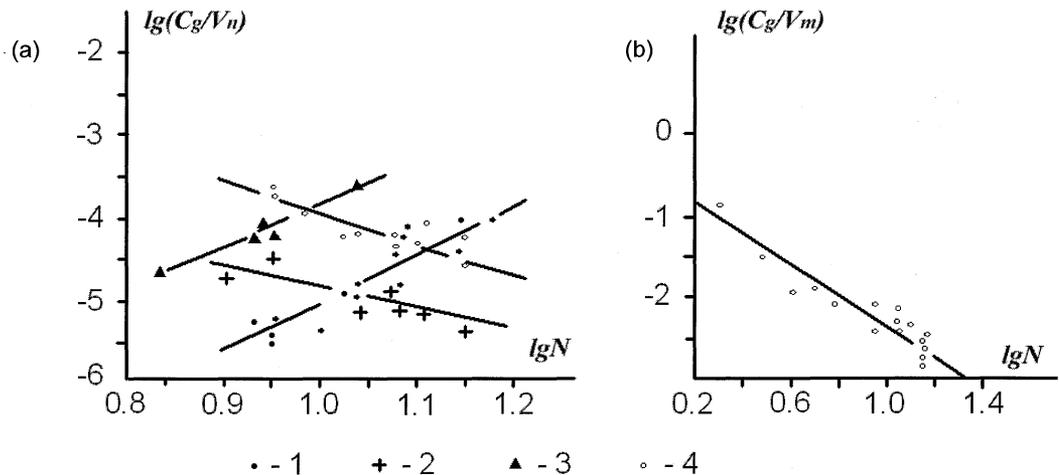


Fig. 3 The change of relative rate of different types of bed forms migration (A (1), Б (2), B (3), Г and Д (4)) for the period of high ($lg(C_g/V_n)$, (a) and low ($lg(C_g/V_m)$), (b) water.

The increase of the river order and the decrease of the mean velocity of the stream in low water is accompanied by the decrease of the $C_{g\Gamma, \Delta}$ (m day⁻¹) which is in agreement with the known quantitative conclusion: in large rivers the bottom forms move slower (Gryshanin, 1979). The dependence (6) shows a good approximation to natural data.

In high runoff the active movement of Γ and Δ forms is subjected to the similar in structure dependence:

$$C_{g\Gamma, \Delta} = 2.2 \cdot 10^4 \cdot v_n \cdot N^{-3.5} \quad (7)$$

where v_n (m s⁻¹) is the mean velocity of the stream in the period of high water. In this season larger forms shift more actively as well. To simplify the process it can be assumed that in the period of high runoff the whole structure of the bed relief is preserved and the "washing out" processes of small forms are included into the velocity value of their shifting. This hypothesis is proved by the observations on the Niger River (Sidorchuk, 1983) and the Kama River (Ivanov, 1984). Its application leads accordingly to obtaining the equations to calculate the velocity of active shifting (m day⁻¹) of B, Б and A forms:

$$C_{gB} = 3.5 \cdot 10^{-4} \cdot v_n \cdot N^{-4.6} \quad (8)$$

$$C_{gБ} = 3.1 \cdot 10^2 \cdot v_n \cdot N^{-2.6} \quad (9)$$

$$C_{gA} = 1.1 \cdot 10^{-6} \cdot v_n \cdot N^{-5.4} \quad (10)$$

The character of relation between the variables in equations (8)–(10) demonstrates various changes of different types of form mobility, when the stream order increases (Fig. 3a). The increase of v_n and N results in the decrease of the intensity of the active shifting of Γ , Δ and Б forms and in the increase of B and A forms of shifting velocity. The reason for such a change of the velocity of the active shifting of different types of bottom relief forms lies in the change of the effect of individual forms in bottom relief structure on the formation of the total bed load transport when moving from small rivers to large ones (Alekseevskiy, 1998).

BED LOAD YIELD CALCULATION

Bed load discharge G (kg day⁻¹) depends on the height of the form h_g (m) and its shifting velocity C_g (m day⁻¹) within the active width of the bed b_p (where $v > v_n$) when the forms of one type move actively. The value G depends on the density of river deposits (δ_0) and the parameters of the element bottom relief form $k = 0,6$, hence:

$$G_g = k\delta_0 b_p (C_g h_g) \quad (11)$$

Following the equation (11) some forms $n = b_p/b_g$ of the same width b_g , length and height move through the river cross section at C_g velocity. The greater the longitudinal and transverse velocity gradient the greater this schematization differs from the natural conditions of bed load transport. In particular, the probability of the active movement of the forms of other type, represented in the hierarchy of bed relief forms, becomes higher. At low water (Fig. 2, $T = T_0$) the movement of the active forms causes the passive dynamics of greater forms, which do not have their own development mechanism. It explains the coincidence of the G values, calculated according to the equation (11) using the parameters of the active and passive forms.

The situation changes radically during the flood, when the particles move at the distance $l > l_g$. Under this, the conditions for the active evolution of the forms of coordinated types are created. Their participation in the total transport of bed load needs the improvement of the equation (11), the estimation of the impact of active forms of different types on the total bottom deformation $|\Delta Z|$, caused by the shifting of the whole hierarchy forms for a rather long period of time. During the active movement of all forms ($T = T_1, T = T_2$) this impact is proportional to the proper height of forms (Fig. 2). Taking into account equation (1) the active layer of river deposit is the sum of individual values of $h_{gs,i}$:

$$|Z| = \sum_{i=1}^M \beta_i h_{gi} \tag{12}$$

here M is the number of forms of different type, crossing the calculated gauge for the time interval Δt . The volume of the transported material connected with i form type migration and individual deformation of the deposit surface ΔZ_i may be calculated according equation:

$$W_i = k \cdot b_{pa} \cdot l_{gi} \cdot |\Delta Z_i| \tag{13}$$

where k is the form coefficient, l_{gi} is the length of i type forms. Sediment discharge, taken as a mean value Δt , corresponds to the volume W_i :

$$G_i = k \cdot b_{pa} \cdot \sigma_0 \cdot C_{gi} \cdot |\Delta Z_i| \tag{14}$$

where the velocity of form shifting $C_{gi} = l_{gi} / \Delta t$. In case of the active development of the whole form complex the total bed load discharge may be determined from following equation:

$$G_i = k \cdot b_{pa} \cdot \sigma_0 \cdot \sum_i^M \beta h_g C_g)_i \tag{15}$$

Bed load discharge G characterizes the mean value of particle transport for T time, when the macroform (form A) is shifted at full length. At some period of time individual sediment discharge connected with the shifting of i type forms do not coincide with G since they make up only a part of the total discharge. Only for the time T the longitudinal coordination of the erosion and accumulation volumes and the sediment discharge, obtained using the form parameters, approaches its real value.

The mean annual and long-term volume of bed load yield may be calculated from equation:

$$W_G = W_{mg} + W_{ng} = (365 - T_n) \cdot \sum_{i=1}^2 G_i + T_n \cdot \sum_{i=1}^5 G_i \tag{16}$$

where T_n is the duration (in days) of high water. The individual values of bed load yield G_i (kg day^{-1}) are calculated according to the equation (11) using the dependences to calculate the height (Table 1) and the velocity of shifting of forms (equations 6–10).

RIVER RUNOFF AND BED LOAD YIELD CORRELATION

It is usually assumed that bed load discharge does not make up more than 1% of the suspended sediment discharge (when the mean diameter of the deposits is $d < 0.25$ mm) and

Table 2 Suspended sediment and bed load discharge ratio upstream and within the branched section of the Lena River near Yakutsk during the flood in 1989.

Data	Q ($\text{m}^3 \text{s}^{-1}$)	Upstream the branched section			Within branched section		
		R (kg s^{-1})	G (kg s^{-1})	$G/(R+G)$ (%)	$\sum R$ (kg s^{-1})	$\sum G$ (kg s^{-1})	$\sum G/(\sum R + \sum G)$ (%)
06 June	34 000	2400	1320	35.5	2400	500	17.2
23 June	24 600	1740	850	32.8	2060	470	18.6

in the case of prevailing in pebble and boulder sediments and when $d/h \geq 0.01$ the ratio R/G increases up to 5–10% (Karasev, 1975). On plain rivers (Kopaliani, 1985) the ratio of bed load discharge to river sediment discharge $R/G = \delta = R/(R+G)100$ as a rule makes up 1–3% (rarely 10%). In mountain streams the value varies from 15 to 70%. It depends on the natural conditions of R formation and on the water volume of the rivers Q (Chalov *et al.*, 2000). Bed load yield is maximum for the rivers and their parts, where the suspended yield is relatively small. On some rivers bed load yield predominates ($\delta = 90\%$) in river sediments. The smaller the river is the greater the bed load yield and the value δ .

The correlation between the suspended sediment and bed load discharge changes in the branched section of the Lena River (Table 2). According to the observations upstream the branched section at large water discharge bed load discharge makes up 32.8–35.5% of the total river sediment discharge. In the branched part of the river the structure of sediment transport changes under the water runoff distribution. In river branches the relative impact of bed load is not higher than 19% of the total river sediment discharge. The main part of bed load discharge during the flood enlarges the volume of river deposits. Specific discharge of bed load (G/B) in the arms of the branched section within different development tendency is the function of the Reynolds number: $G/B = n(Vh)^m$ where Vh is the mean velocity and the depth of the stream; $n \sim \nu^{-m}$. Here ν is the kinematic viscosity at water temperature 18°C , m is the empirical parameter. The corresponding values n for the developing, stable and disappearing (dying) arms in the middle course of the Ob River are equal to 2.75×10^{-4} ; 1.8×10^{-2} ; 2.9×10^{-2} , and the parameter m are 3.5; 1.26 and 1.0. The ratio $G/B = \text{const}$ is reached in these branches under different values of Vh . In the developing arms a greater increase of Vh is needed to move the mineral particles with the intensity of $0.03 \text{ kg s}^{-1} \text{ m}^{-1}$ in comparison with the stable and dying arms. When the value G decreases the value R and a part of coarse fraction of bed load discharge increases.

CONCLUSIONS

River sediment yield is an important component of the matter flux in different parts of the network. It is often considered to be practically equal to the suspended sediment yield. In a number of cases such a hypothesis is incorrect, it results in the errors in the determination of river sediment outflow into the reception reservoirs, the time of reservoir silting and the intensity of channel deformations. The study showed, that a part of the bed load yield could not be assumed as a constant value equal to 1–10% of the suspended sediment yield. It is changed in a big variety of values depending on the suspended sediment yield, the type and size of the river and other factors.

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