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A New Hydrological Method for Estimating the River Bed and Drainage Basin Components of Erosion and Suspended Sediment Fluxes in River Basins

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Abstract

This paper uses the results of river suspended sediment flux (SSF) analysis to propose a new hydrological method for quantitatively estimating the river bed and drainage basin (sheet erosion, rill and gully erosion) components of total erosion intensity in river basins. The suggested method is based on the establishment of the functional power connection between mean monthly water discharges (WD, Q_i) and suspended sediment fluxes (r_i) calculated for the low-water-discharge phases of a river's hydrological regime in various (on mean annual water discharges) years: $r_i = a \times Q_i^{\mu}$ (where a, μ are some empirical coefficients), and further extrapolation of this connection for other phases of the hydrological regime. Thus, the extrapolation allows us to calculate (in a long-term annual SSF) the proportions of sediments originating in river beds and drainage basins. The proposed method is tested using a long-term (not less than 10 years) series of observations for WD and SSF of 124 chiefly small and midsize rivers of the East-European plain, the Urals, the Eastern Carpathians, the Ciscaucasia and the Caucasus, and Central Asian mountains, containing data on the mean monthly values of WD and SSF. The paper also compares the method with other methods for estimating the components of erosion intensity and SSF.

Keywords: *Erosion, Drainage basin, River bed, Suspended sediment flux, Suspended sediment structure, Northern Eurasia.*

1. Introduction

Given the influence of various forms of atmospheric precipitation on the Earth's surface, as well as their effect on its mechanical denudation (mainly erosion), the aggregate discharge of eroded products by rivers out of their basins (so-called sediment flux) can be divided into *river bed* and *drainage basin* (washout by rainwater and melted water (snowmelt runoff and glacial runoff) within a catchment area components. Quantitative estimation of these two components of the rivers' SSF as one of the relative indicators of erosion rate in the river basins [1-2] is useful in solving a number of theoretical (geomorphologic, hydrological, geoecological and others) and applied problems. In particular, it facilitates the identification of areas where either river bed or drainage basin erosion is prevalent with general direction and intensity of their relief evolution in terms of valleys' deepening or basin surface leveling [3]. Existing methods for the structural separation of suspended sediment fluxes (more broadly – erosion) in river basins can be divided into the following main groups.

1) Hydrological approach, which analyzes the laws of SSF in connection with WD and other hydrological indicators (partitioning of the chronological graphs of river muddiness; thermal analysis of sediment fluxes; establishing the multi-factor connection between sediment flues and climatic elements and another natural factors and characteristics such as rocks lithology, topography, soil erodibility, vegetable cover, river basin area, etc [4-6; etc]). One such specific method was elaborated by Shcheglova [7], who, based on earlier ideas of Polyakov [8], Lopatin [9] and others, provided a quantitative assessment of the share of river bed sediments in the long-term annual SSF of some rivers in Central Asia. According to Shcheglova [7], the river bed SSF includes the sediment fluxes in low-water period (without rain floods). This method relates the dependency (the set of curves) of washout muddiness on the specific WD of rivers to different levels of river beds' stability. With low WD-values these curves are determined by the values of river muddiness in low-water periods; with high WD-values they are supplemented with muddiness values corresponding to flood periods when basin-surface washout temporarily stops due to a number of reasons (snow cover, relative climatic dryness, etc). The identification of such periods demands thorough analysis with hydro-meteorological information.

2) The partitioning of erosion products, using data obtained by analyzing the mineralogical and granulometric composition of suspended sediments, river bed and flood plain deposits [10, 11]. This

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method is considered to be most efficient if the mineralogical composition of rocks noticeably differs in various parts of a river basin. The method has not yet been broadly used due to the absence of reviews of rock compositions (especially alluvial ones) constituting entire river basins.
3) Determining sediment structures by comparing rivers with various environmental (landscape) conditions affecting erosion processes in their basins. This approach, as used, for example, by Dedkov and Mozzherin [1], is known for the low accuracy of its quantitative assessments.
4) The analysis of materials from stationary research on the surface of river catchments (observations of gullies, sheet and rill erosion, landslides, creep processes, etc.), in order to partition the share of basin-origin products [12–16, etc.]. The accuracy of calculation in this case depends on the research methods extrapolation of results to unstudied parts of basins and the completeness of the account of constituents of basin mechanical denudation.
5) The hydrological-cartographic approach, which is similar to the previous one. It estimates the mass of basin-origin sediments based on the analysis of maps showing the land disturbance by erosion in a river basin, and then compares this mass with total river sediment flux. One representative work is the research by Golosov [17] who determined the erosion structure in the Oka River basin using maps showing lands erosion danger and the density of gullies in the European part of Russia.

In spite of the multitude of methods and their positive features, practically all of them have one common drawback: at present they cannot be used to make a regional review due to differences in both methodology and availability information on various regions of the planet. One of the methods (hydrological group) applicable for large-region reviews is the simple method of quantitatively estimating river bed erosion by Dedkov and Mozzherin, suggested by them as the development of ideas of Polyakov [8] and Shcheglova [7]. It is based on the analysis of the dependence (linear relation (!) by Dedkov and Mozzherin) of SSF on mean monthly WD in low-water periods in years with various low-water discharges. The resulting graph depicting the relation between these parameters shows sediment flux corresponding to mean annual WD and suggest that it is solely determined by river bed erosion. By subtracting the long-term mean annual sediment flux of river bed origin from the total long-term mean sediment flux, we can find the volume of sediments characterizing the erosion by melted and rain water on a basin surface. Unlike the muddiness partition graph, this method is less labour-intensive and does not require a high level of details in muddiness samples, which makes it more promising in large-region researches. However, the method has one serious drawback: in most cases its proponents extrapolate the relation between WD and SSF in low-water periods up to the value of mean annual WD, which is incorrect for rivers with highly uneven WD within a year. It would be more accurate, in our opinion, to extrapolate this relation not up to the mean annual (a fortiori long-term mean annual!) values of water discharge, but, at least, up to the mean monthly values of WD of each analyzed year, as the following inequality clearly shows:

$$f((Q_1+\ldots+Q_n)/n)^{\mu} \neq f((Q_1^{\mu}+\ldots+Q_n^{\mu})/n) \qquad (1)$$

where $Q_{1\ (\ldots n)}$ is the mean WD for a certain month, n is the number of months in the long-term series of WD observations, μ is empirical power indicator of the relation between WD and the river-bed-origin sediment flux (see further equation (2) and (3)). After extrapolating the relation between low WD and sediment fluxes up to mean monthly values of WD, the differences in assessments using the method of Dedkov and Mozzherin will increase because intra-annual (inter-monthly) WD unevenness is higher and the power relation indicator μ diverges more strongly from 1. Proceeding from the above-stated overview, this article aims to improve the method of Dedkov and Mozzherin, by showing the limits of its accuracy and offering a more accurate methodological baser for research in large regions.

2. Data and Study Area

The data used to test the proposed method come from the long-term (10 years or more) monitoring of the Hydro-Meteorological Service of the former USSR for WD and SSF of 124 chiefly small and midsize rivers of the East-European plain, the Urals, the Eastern Carpathians, the Ciscaucasia and the Caucasus, and Central Asian mountains (Fig. 1). Data exist on the mean monthly values of WD and SSF in these rivers, which are characterized by an absence of large water reservoirs in their basins. The selection of certain rivers in Central Asia was determined by the possibility of comparing the results offered here with earlier work on the same rivers [7]. For the convenience of the analysis and presentation of results, the river basins were divided into three altitudinal groups: plain basins (where average altitudes are range from 0 to 500 m), low-mountain (from 500 to 2 000 m) and middle-mountain (from 2 000 to 3 500 m) ones. This basins' division is provisional as the structure of erosion within their limits is determined by a large set of factors that are very changeable even within the same altitudinal group, in particular river bed processes. The river basins under analysis in present study are marked by differences in drainage areas, the slopes of their surfaces and streams, as well as the degree of economic (mainly agricultural) transformation of their natural landscapes (chiefly vegetation cover). These factors affect the overall intensity of mechanical denudation (including erosion) and the values of river sediment fluxes in river basins (Table 1).

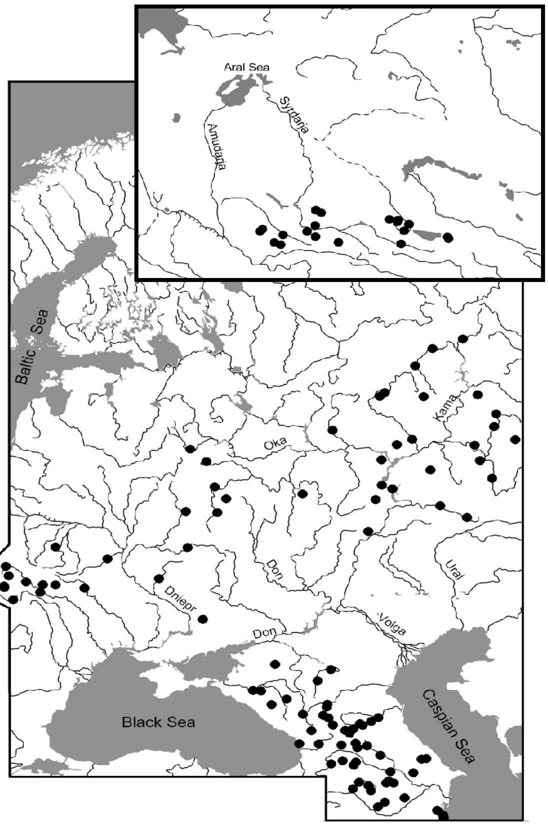

Fig. 1. Location maps of hydrological stations on rivers under analysis.

Table 1. Some average characteristics of river basins under analysis.

Characteristics	Altitudinal groups of river basins			
	plain (0–500 m)*	low-mountain (500–2 000 m)	middle-mountain (Caucasus) (2 000–3 500 m)	middle-mountain (Central Asia) (2 000–3 500 m)
N, units	40/4	35/10	28/20	21/20
T, years	16.9 ± 1.7 [1)]	15.9 ± 2.0	17.8 ± 2.5	20.6 ± 2.6
F, km^2	17 461 ± 8 446	4 206 ± 2 764	969 ± 370	2 262 ± 1 113
H, m a.m.s.l.	232 ± 24	1 121 ± 158	2 277 ± 89	2 782 ± 173
α, ‰	34.7 ± 15.1 [2)]	282.7 ± 47.6 [3)]	336.2 ± 57.2 [4)]	410.0 ± 79.0 [5)]
I, ‰	0.56 ± 0.33	20.4 ± 8.2	36.2 ± 10.2	31.0 ± 7.4
$\alpha : I$	17.5 : 1.0	9.3 : 1.0	7.9 : 1.0	10.9 : 1.0
$\tilde{A}$, points	2.3 ± 0.2	2.0 ± 0.3	1.3 ± 0.2	1.05 ± 0.09
M, l/(km^2 a sec)	6.2 ± 2.7	14.1 ± 3.0	25.9 ± 7.5	16.5 ± 3.9
R, t/(km^2 a year)	39.5 ± 26.7 (12.2 ± 5.3)	304.6 ± 112.9 (130.8 ± 80.4)	331.6 ± 129.7 (243.8 ± 120.5)	223.3 ± 83.8 (234.3 ± 83.1)

* – above mean sea level

N: is a quantity of river basins (general/including the basins of I category of land use (see below)); *T*: is an average duration of observations over WD and SSF of rivers under analysis; *F*: is an average river basin area; *H*: is an average altitude of river basins, α is average river basin slope, *I*: is an average river bed slope; $\tilde{A}$: is an average point (category) of economic (mainly agricultural) land use in river basins (point 1.0 (I category) is for weakly anthropogenically transformed basins where residual forest area (*L*) is more than 70%, a tilled (cultivated) area (*P*) is not more 30% (mainly in the steppe regions); point 2.0 (II category) is for basins with medium level of anthropogenic transformation of basin natural landscapes (*L* for forest zones and *P* for steppe zones are from 30% to 70%); point 3.0 (III category) is for basins with great transformation of natural landscapes (*L* for forest zones is less than 30%, *P* for steppe zones is more than 70%); *M*: is a long-term mean annual specific runoff; *R*: is a long-term mean annual specific SSF (the mean specific SSF for basins with I category are in the parenthesis).

Note: [1)] – here and farther the intervals of all mean characteristics are given with 95% confidence level, [2)] – on 5 basins from 40, [3)] – on 18 basins from 35, [4)] – on 20 basins from 28, [5)] – on 8 basins from 21.

3. Research Method

The division of suspended sediment fluxes into river bed and drainage basin components was carried out in several stages.

1. Graphical representation (here and further with the use of the Microsoft Excel software package) of the connection between mean monthly WD (Q_i, m^3/sec) and SSF (R_i, kg/sec) for the whole monitoring period for every river under analysis. This connection is described as the following power equation:

$$\check{R}_i = A_{er} \times I \times Q_i^m \quad (2)$$

where: $\check{R}_i$ is the theoretical (calculated) mean suspended sediment flux in calendar i-month with the mean water discharge Q_i, A_{er} is a complex erosion coefficient that depends on WD irregularity, the type of rocks composing the river bed and catchment area, the volume and grain size of the sediments delivered during river bed erosion, as well as from tributaries and catchment area. I is the slope of the river stream above the hydrological station, m is an empirical power indicator of the connection between Q_i and R_i. A_{er} and I may vary significantly from one river to another depending on the geological and geomorphic structure of their basins and river beds, as well as the environmental conditions within a catchment area [18].

2. From the graphical representation of the relationship between Q_i and R_i, only WD-values that correspond, first of all, to the winter low-water period in years with various low-water discharges were selected for further analysis. It is these water discharge values that mainly determine the river bed erosion when R_i-values are characterized by the complete absence of basin-origin sediments, including the products of sheet washout and linear erosion at interfluves, when their surface is "preserved" by a snow cover. Such months occur, though rarely, during the warm seasons. A good example of this is the abnormally hot and dry summer of 2010 during the majority of which the sediment fluxes of the East-European plain's rivers, for example, was composed almost completely of its river bed component. Q_i and R_i values for most low-water summer months were also selected for analysis. Water and sediment discharges, rejected at that stage, correspond to months with high-water (flood) regime, and the share of basin-origin sediments is, as a rule, very significant in them. This is a well-known fact, therefore, demands little comment.

3. From the graphical representation of the relationship between WD and SSF for winter low-water or similar summer water discharges, only those cases were selected which belong to a certain range of minimal R_i, values that, in turn, correspond to the comparatively "pure" low-water river bed erosion. There are two reasons for the selection of this range for analysis rather than only minimal SSF-values. First, with the same mean monthly WD-values the various volumes of sediments carried out by a river can be determined by the different mean daily distribution of water discharges during these months. Second, the uneven lithological structure of river beds and flood plains at various parts of the river length determines, to a certain extent, the various tempos of river bed transformations in different years, which is reflected in different volumes of sediments in equal or similar water discharges. The latter is especially true for plain rivers, whose river beds, flood plain and terrace complexes are composed of a wide range of rocks (from pebbly to silty) that produce different volumes of sediments during lateral erosion and change the ratio between suspended and bed-load components. It is quite obvious that even this group of minimal mean monthly SSF-values does not entirely consist of river bed washout products, as there is always a chance that they will be diluted by drainage-basin sediments during winter thaws and more frequent summer rains, which is obviously not registered in their mean monthly values. Outside any analyzed monitoring series even lower R_i-values (at the same Q_i) can be noted, maximally reflecting the intensity of low-water-discharge river bed erosion. However, the further analysis includes only the available results of regime observations at hydrological stations.

The scheme for the step-by-step selection of correlated values of WD and SSF is presented in Fig. 2.

4. The obtained equation of the connection between mean monthly WD- and SSF-values, conditionally corresponding to the model of "pure" river bed low-water erosion in the system "river WD – river bed erosion – river bed SSF" forms the basis for dividing the long-term annual SSF-values into drainage basin and river bed components. This equation has the following general view:

$$r_i = A_{er} \times I \times Q_i^{\mu} \quad (3)$$

where: r_i is the theoretical (calculated) mean monthly river bed sediment flux (for other coefficients see equations (1) and (2)). The power character of the equation most correctly reflects the interconnection of the water and sediment discharges, and also their mutual temporal variability. It should be noted that at the start only connection equations were considered suitable for analysis when the approximation values of power trends (coefficient of determination) – R^2 where larger than 0.5, thereby giving linear correlation coefficients between WD and SSF that are not less than 0.7 (high and very high correlation). Such selection increases the reliability of the results.

Figure 3 graphically presents the linear correlation between the power indicators m in equation (2) and μ in equation (3). For the majority of analyzed plain rivers (88%) and low-mountain rivers (75%) $m > \mu$; for Caucasian middle-mountains this is the case for only two-third of rivers. The situation is different in the middle-mountain belt of Central Asia, where $m < \mu$ for the majority of analyzed basins (86%). μ-values grow with the increase of low-water discharges and rivers' slopes, with the increase of the grain size of river-bed-forming sediments, and have certain limits of changes in each region.

5. Based on equation (3), we can estimate the probable values of river bed sediments for every month (r_i) of every analyzed year (t_i) (Fig. 4). In those rare months when the theoretical (calculated) value of sediment fluxes exceeds its actual mean monthly volume (R_i) (i.e. $r_i > R_i$) by value Δr_i, it is assumed that the resulting restored river bed sediment flux is equalized ("reduced" to) that mean monthly volume ($r_i = R_i$) (Fig. 4). The mean monthly SSF values that have been selected to build equation (3) are not divided into river bed and drainage basin components, because it is basically assumed that for them $r_i = R_i$. Thus, two series of mean monthly values of river bed sediments are created:

(i) a series calculated without taking into account cases when r_i exceeds the actual sediment flux R_i: all r_i-values are marked as $\check{r}_i$ (a series of $\check{r}_i$-values);

(ii) a series which takes into account cases when r_i exceeds the actual sediment flux R_i, by equating r_i to R_i on the condition that $r_i > R_i$ (a series of r_i-values).

It is the latter series of r_i-values that was used for analysis. Accordingly, the long-term mean river bed sediment value is underestimated by value $dr_{av} = (\Sigma r_i/\Sigma\check{r}_i)\times100\%$. However, this underestimation is small: dr_{av} for all 124 rivers does not exceed, on average, 3.5% (2.8% for plain rivers, 6.2% for low-mountain rivers, 2.6% for middle-mountain Caucasian rivers and 1.6% for Central Asian ones). Besides, the stated underestimation is compensated by the overestimation of low-water sediment flues that we associate with river bed washout, as basin-origin sediments can be included into it on certain days of analyzed months during likely rains and winter thaws.

6. The theoretical mean monthly values of river bed sediments (r_i) are averaged for every year ($r_{av}(t_j)$) and for the whole observation period (r_{av}) for each river. They are correlated, accordingly, with the mean annual ($R_{av}(t_j)$) and long-term mean (R_{av}) values of actual sediment flux (Fig. 4). The obtained value δr ($\delta r = (r_{av}/R_{av})\times100\%$)) characterizes the share of river bed component in the long-term mean total sediment flux.

This technique of estimating the share of sediments, formed at river bed washouts, has two main drawbacks. First, even taking into account the power character of the WD and SSF connection (see equation (3)), it is still impossible to reliably determine the inherent stream erodibility during the high-water periods (floods). This is because it is assumed that the stated low-water connection exists during these phases of the river's hydrological regime as well. It is obvious, however, that the erosion coefficient A_{er} significantly changes from low-water to high-water periods, depending, among other things, on the mass and grain size of the sediments washed into the rivers. It is well-known that basin-origin sediments, whose character is determined by a number of environmental factors within a catchment area, are basically not affected by the functioning of the system "river WD – river bed erosion – river bed SSF". Denudation material entering rivers during slope-gravity processes on actively eroding riversides represents one partial exception. Inherent stream erodibility is also significantly influenced by the transition from low-water to high-water periods in terms of hydrodynamics, the character of interaction with banks (including flood plains), etc. It is during large water discharges that the most intensive river bed transformation takes place.

The second main disadvantage of proposed technique is that the calculated structure of sediments passing through a hydrological station does not reflect the absolute ratio of sediments entering a river from its catchment area (as the result of erosion and the accumulation of basin erosion (denudation) products) and the sediments formed during river beds washouts upstream from a hydrological station. This happens because the accumulation of these two groups of sediments along the river bed during their transportation is not measured. This is especially true for rivers with gently sloped channels. From this it follows that the value of a SSF structure that varies along the river is rigidly connected with a particular hydrological station and is therefore cannot be reliably extrapolated to the whole river basin upstream from the hydrological station in question.

4. Testing and Discussion

The results from applying this method may be formulated as the following basic propositions:

1. Preliminary estimation shows that in the entire set of the analyzed basins of plain and mountain rivers, the share of river bed erosion and SSF (δr) does not exceed, on average, 21% with the lowest value of 8.5% identified in the middle-mountain group of Central Asian basins (Table 2). The δr-values of basins with landscapes that are more or less natural or only weakly transformed by economic (agricultural) activity (I category of land use (see Table 1)) exceed the average δr-values in all basins of corresponding altitudinal groups (Table 2). This excess is especially notable in plain river basins. It is interesting to compare for results on middle-mountain rivers in the Caucasus (δr = 17.2±7.1%) and Central Asia (δr = 8.5±2.2%). The small share of river bed sediments in Central Asia's middle-mountains can be explained by the relatively high climatic (environmental) aridity of this region in comparison with the Caucasus. This dryness means that the mechanical denudation of the basin slopes, which are poorly protected by vegetation cover, has a greater effect on formation of annual river sediment fluxes. It is note worthy that the overall tempo of mechanical denudation in natural landscapes is similar in these two mountain regions, judging by the SSF-values (Table 1).

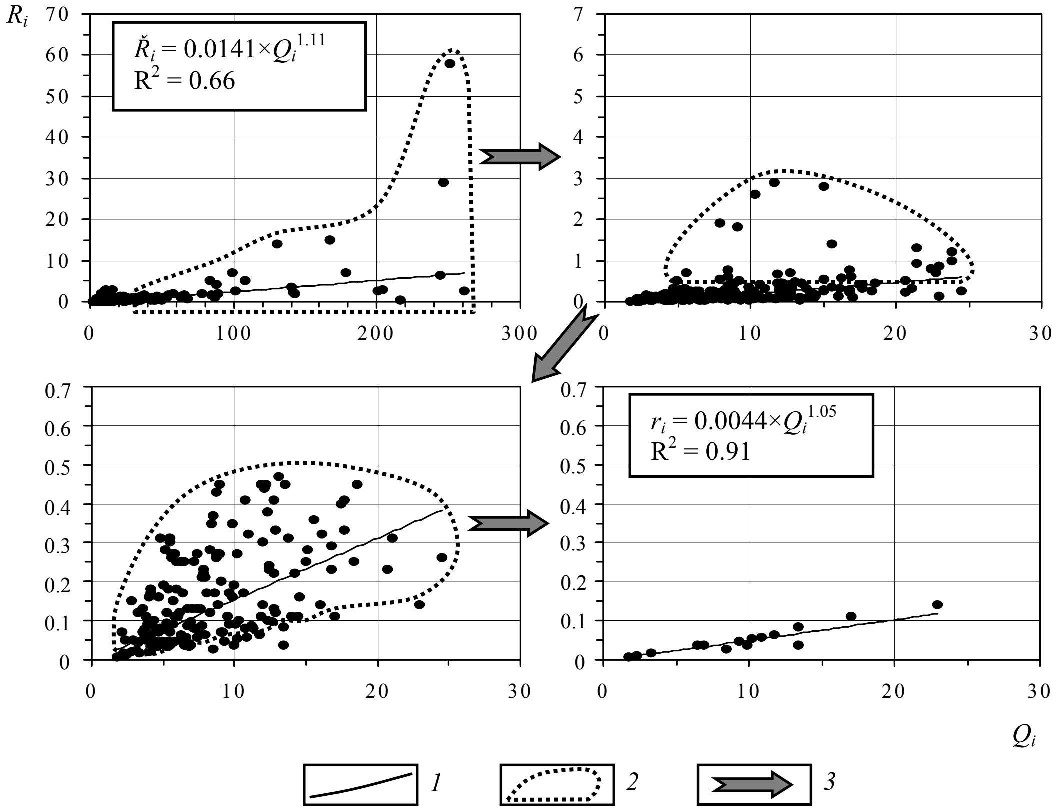

Fig. 2. The principled scheme for the step-by-step selection of correlated mean monthly values of WD (Q_i, m^3/sec) and SSF (R_i, kg/sec) to construct the power equation of their relationship (equation (3)) on example of the Teterev River /Makalevychi (Ukraine); *1* – the power trend, *2* – the graphic removal field of correlated Q_i and R_i values, *3* – the sequence of step-by-step selection; R^2 – determination coefficient for the power trend.

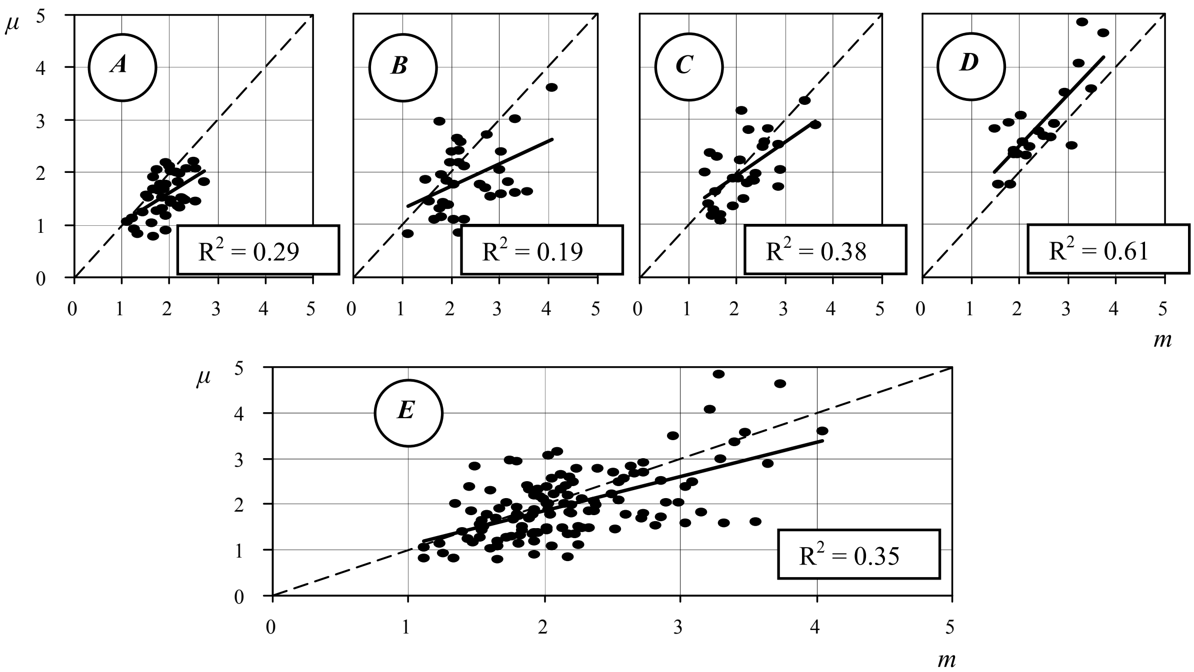

Fig. 3. The linear correlation between power indicators m in the equation (2) and μ in the equation (3) on altitudinal groups of river basins under analysis

Altitudinal groups: A – plain, *B* – low-mountain, *C* – middle-mountain (Caucasus), *D* – middle-mountain (Central Asia), *E* – all analyzed basins; R^2 – determination coefficient of the linear trend.

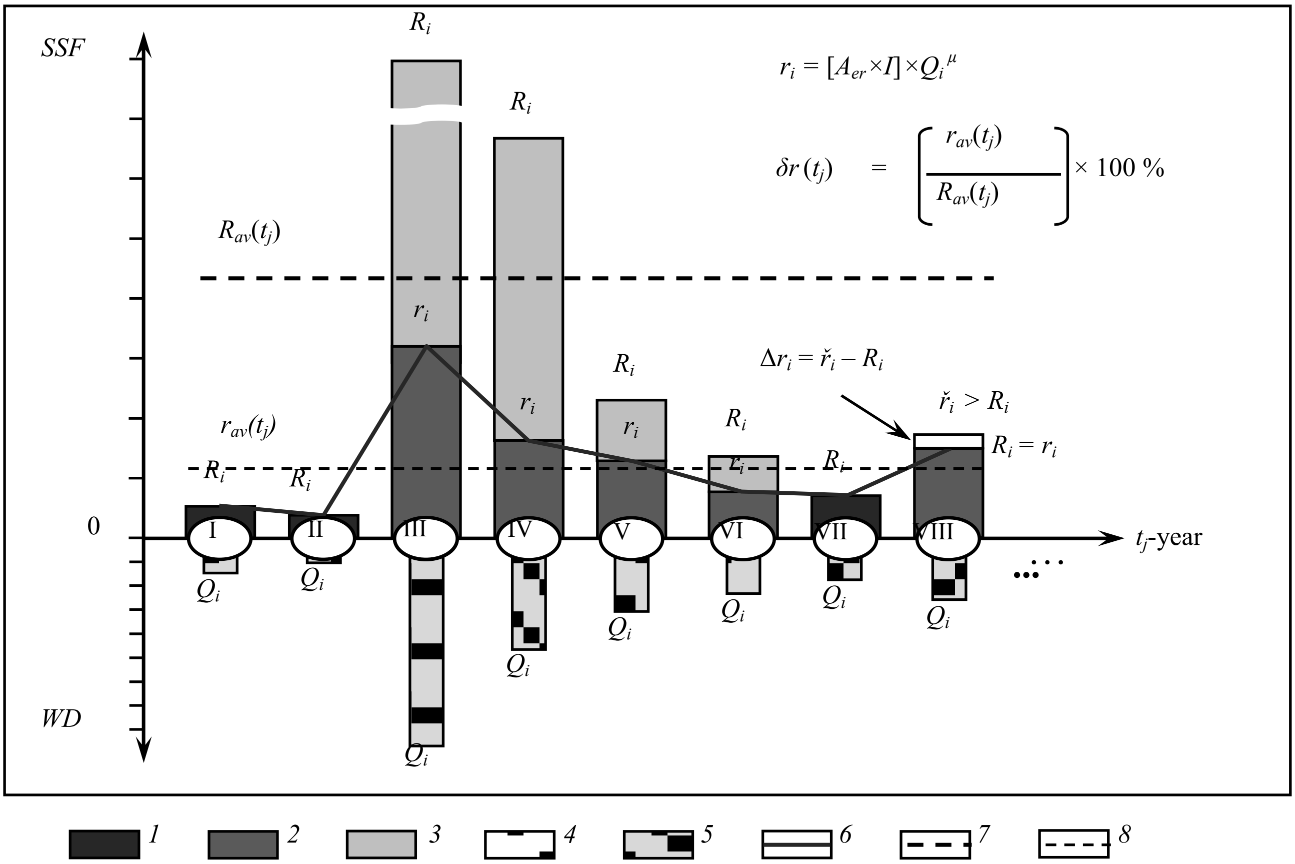

Fig. 4. The principled scheme for structural separation of SSF within a calendar year
1 – an undivided mean actual monthly SSF (R_i), *2* – a calculated mean monthly river bed SSF (r_i), *3* – a drainage basin component of actual mean monthly SSF ($R_i - r_i$), *4* – a difference (Δr_i) between calculated mean monthly river bed SSF (r_i) and actual mean monthly SSF (R_i), considering that $r_i > R_i$, *5* – a mean monthly WD (Q_i), *6* – a number of r_i-values for t_j-year, *7* – an actual SSF averaged for t_j-year ($R_{av}(t_j)$), *8* – a calculated river bed SSF averaged for t_j-year ($r_{av}(t_j)$); I, II, Ш … – the months of calendar t_j-year
Note: assuming that, in this case, in I, II, III and VII months the actual SSF (R_i) hypothetically equas to river-bed SSF (r_i) (i.e. $r_i = R_i$), these R_i-values are taken to create the equation (3).

Table 2. Some average river bed SSF characteristics of analyzed rivers.

Characteristics	Altitudinal groups of river basins			
	plain	low-mountain	middle-mountain (Caucasus)	middle-mountain (Central Asia)
μ	1.57 (1.92)	1.85 (2.00)	2.04 (2.66)	2.90 (2.93)
δr, %	20.7 ± 6.4 (62.3 ± 9.5)	16.1 ± 4.3 (24.8 ± 9.4)	17.2 ± 7.1 (18.2 ± 8.5)	8.5 ± 2.2 (8.7 ± 2.3)
λr, %	11.2 ± 3.1 (29.9 ± 0.1)	9.3 ± 2.6 (17.8 ± 7.6)	10.1 ± 4.0 (11.2 ± 5.2)	2.2 ± 0.5 (2.2 ± 0.5)
η, %	63.9 (47.7)	74.4 (72.3)	62.6 (63.3)	29.5 (29.0)

On river basins with I category of land use (see Table 1) is in parenthesis

What explains the reasons for such close δr-values in such different (by their morphometric characteristics) river basins grouped into the above-mentioned altitudinal categories? Let us answer this question with the example of two diverse groups – the middle-mountain and plain ones. On the one hand, the intensive basin erosion and significant values of the drainage basin SSF component in middle-mountain basins are promoted by the large slopes of their surfaces and highest specific WD, which, as a consequence, we see in the high overall specific SSF as well. The rivers are also filled by significant sediments mass due to gravity processes on the steep and high slopes of their valleys. The landslide masses entering the river beds of many mountain rivers in humid years greatly increase water muddiness. This is also true for plain rivers. On the other hand, this group of basins is also marked by the highest slopes of river beds as well as by a low ratio between the slopes of basin surfaces and river beds (Table 1). This fact actually increases the relative role of river bed (mainly deep) erosion in the overall intensity of erosion in the river basins of middle-mountain regions. At the same time, the river-bed-forming alluvium of the analyzed rivers is predominantly pebbly and pebbly-boulder, which gives, in general, a comparatively small (in relation to the river's potential erodibility) mass of suspended sediments during river bed deformations. Suspended sediments of such rivers practically do not participate in formation of the river-bed-forming alluvium. Besides, during the low-water periods the pebbly-boulder sediments of mountain rivers form an erosion "pavement" of the largest material, which ensures river bed stabilization during that phase of their hydrological regime (Fig. 5). However, during high-water periods, when the debris forming the erosion "pavement" begins to move, a lot of pebbly-sandy material, which earlier lay under the "pavement", enters the stream and greatly increases muddiness [18].

Within the highest analyzed altitudinal group, δr-values vary much more significantly depending on the overall intensity of denudation and annual SSF-values, the composition of river-bed-forming alluvium, the level of economic (agricultural) transformation of natural landscapes in river basins, and a number of other reasons (Fig. 6).

A different scenario is presented by plain rivers. The comparatively high ratio between the slopes of surfaces and river beds within their basins (Table 1) decreases the role of river-bed erosion (and of δr-value) in the overall intensity of erosion in the basins. Yet, the fact that the river beds (flood plains) of Eastern-European plain rivers are predominantly composed of sandy, sandy-silty, silty and silty-organic alluvium [19] explains high river bed (mainly lateral) erosion and increases the δr-value. Rivers with sandy-silty alluvium are characterized by an absolute predominance of suspended component in sediments, which settles after high water becomes lower [18].

The larger average area of the basins (and, accordingly, length of the rivers) of the analyzed plain rivers in comparison with the middle-mountain basins is another reason for their higher δr. It is well-known that the bulk of basin-origin sediments originates mainly in the upper parts (links) of a fluvial network (small rivers), becoming more and more settled while moving towards large the streams of basins [20]. Consequently, other conditions being equal, δr-values also increase in the same direction. In addition, during the movement of sediments, their constituent fractions become smaller and more differentiated along the river by their grain size and shape, petrographic and mineralogical composition, etc. [18]. As the river length increases, the probability of further movement of the thinned material in suspended form by river bed erosion also increases, consequently, increasing the δr-value. This pattern was marked, for example, in the basin of the Amur River [21]. It is less clear due to the geomorphic and lithological heterogeneity of basins, which is highest in mountain regions (characterized by the alternation of mountain ridges slit by rivers with pebbly-boulder river beds and inter-range depressions with significant presence of thinned material in alluvium, the washout of huge mountain-morainal, proluvial and colluvial deposits [22], and where tributaries have different sediment compositions.

It should be added that human activity which changes natural landscapes is one more factor influencing the relative importance of the river bed and drainage basin components of erosion and sediment fluxes in river basins. Let us compare, for example, the composition of SSF in two rivers of the forest zone of the Eastern-European plain, which have approximately equal drainage areas (F), but different residual percentages of forest lands in their basins (L): the Kama/Volosnitskoye (L = 77%, F = 9 750 km^2, I = 0.3 ‰, M =7.2 l/(km^2 a sec), R = 18.1 t/(km^2 a year)) and the Upa/Orlovo (L = 8%, F = 8 210 km^2, I = 0.21 ‰, M =4.9 l/(km^2 a sec), R = 19.7 t/(km^2 a year)). In case of the Kama River the relatively large share of the river bed component (δr = 75.1%) is determined by its smaller area (in its basin) of erosion-hazardous lands (including lands cultivated for crops) that could deposit large masses of denudated (eroded) material into the river network through melted and rain runoff. There is reason to suppose that the δr-value would be even higher if the upper Kama's fluvial bed was not composed of predominantly sandy (and probably rougher) river-bed-forming sediments, a large part of which moves as bed load. As for the Upa River, the significantly (in comparison with the Kama River) smaller river bed component (δr = 9.6%) largely results from the great number of basin sediments in the river as the erosion-hazardous areas in its basin are very large (cultivated lands are 55% of F).

2. Estimation of the river bed component of SSF based on the method presented in this paper (δr) differ more substantially from estimations obtained by Dedkov and Mozzherin when the power indicator μ in equation (3) diverges more significantly from 1. Out of all altitudinal groups, the low-mountain basins show the smallest differences between estimations (Table 2).The long-term average share of river bed SSF is denoted in the method of Dedkov and Mozherin by λr, which is calculated for each river by equations (5) and (6):

$$r_{av} = A_{er} \times I \times Q_{av}^{\mu} \qquad (5)$$

$$\lambda r = (r_{av}/R_{av}) \times 100\ \% \qquad (6)$$

where: r_{av} is the long-term mean annual river bed SSF, Q_{av} is the long-term mean annual WD of a river, R_{av} is the long-term mean annual SSF of a river (for other coefficients see equations (2) and (3)).

Fig. 5. The pebbly-boulder fluvial bed with riverside "pavement" of the Cherek Balkarskiy River (Kabardino-Balkarian high-mountain reserve, Northern Caucasus) (photo by author, August, 2009).

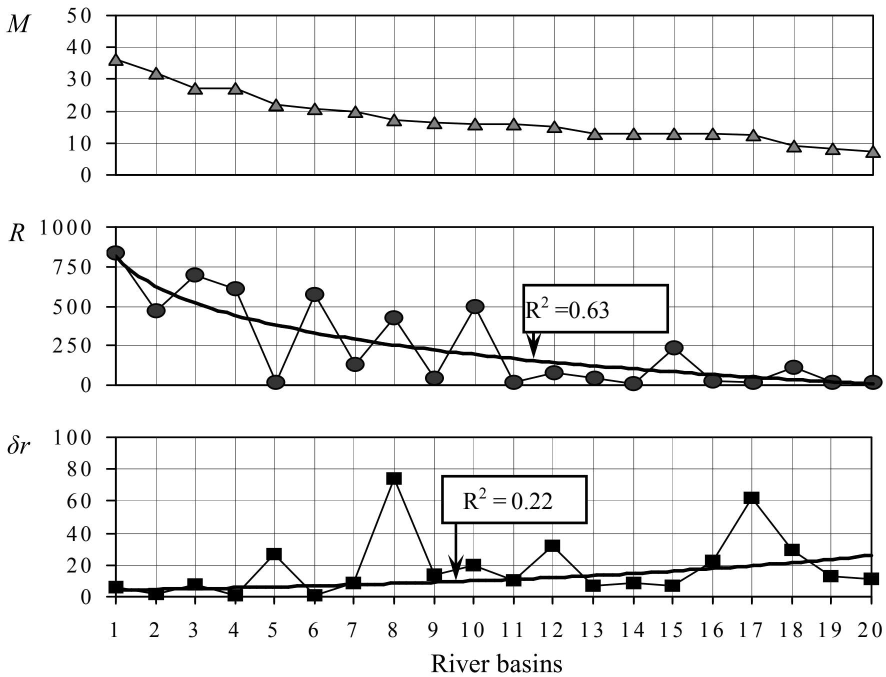

Fig. 6. The increase of river bed component of SSF (δr, %) against the reduction of water discharges (M, l/s×km 2) and suspended sediment fluxes (R, t/km^2×year) in the middle-mountain river basins of the Caucasus characterizing by weakly anthropogenic transformation of natural landscapes in them (R^2 – determination coefficient of the power trend).

River basins: 1 – Cherek Balkarskiy/Babugent, 2 – Chyornaya Aragvi/mouth, 3 – Ardon/Taminsk, 4 – Bol'shaya Liahvi/Kekhvi, 5 – Zivlanchay/Kyadamysh, 6 – Gizel'-Don/Dargavs, 7 – Kuban'/Kosta Hetagurov's village, 8 – Kusarchay/Kuzun, 9 – Dastafurchay/Karagullar, 10 – Baksan/Zayukovo, 11 – Geranchay/Yuhary, 12 – Ktsiya-Khrami/Eddikalisa, 13 – Gyandjachay/Zurnabad, 14 – Argichi/Getashen Verin, 15 – Nakhichevanchay/Karababa, 16 – Vorotan/Angekhakot, 17 – Marmarik/Agavnadzor, 18 – Malka/Kamenniy most, 19 – Paravani/Hertvisi, 20 – Gavaraget/Noraduz.

The correlation of λr and δr ($\eta = (\lambda r/\delta r) \times 100\%$) shows the limited accuracy of Dedkov and Mozzherin's method with large power indicators of μ. As we can see in Fig. 7, for these two close methods, the largest differences in η-estimations for a unit of μ-changing are characteristic for plain rivers and the smallest differences are for low-mountain rivers. At $\mu > 2$ the difference in estimations for the whole set of analyzed rivers is expected to be more than twofold (Fig. 7(*E*)).

It is quite obvious that the smallest differences are shown by the very few rivers for which $\mu \approx 1$. The acceptable divergence between δr and λr estimations (up to ± 25%, i.e. within the limits of average accuracy of SSF measurements) is acceptable, for all 124 rivers, with $\mu \approx 0.84 \div 1.25$ (for plain rivers $\mu \approx 0.82 \div 1.29$, for low-mountain rivers – 0.69÷1.34, for middle-mountain rivers of Caucasus – 0.87÷1.76, for middle-mountain rivers of Central Asia $\mu \approx 1.26 \div 1.63$).

The differences in estimations are large in the plain regions with semi-arid and arid climatic conditions, where the unevenness of river WD is most pronounced.

3. There is a generally high correlation between the estimations of SSF composition for twenty-one rivers in the middle-mountains of Central Asia obtained, using the method proposed in this paper and the estimations resulting from Shcheglova's method. Figure 8 gives such a comparison of the values of SSF river bed component of these rivers.

All values, except those for four rivers (this fact needs further scrupulous study), are closely correlated (the linear correlation coefficient of the estimations equals 0.86). Yet, there is a difference in the results of two methods concerning the river bed component share in sediments fluxes, which is easily explained: when streams are overloaded with basin-origin sediments, it is methodically difficult to determine the river bed component due to its relatively small value.

For all analyzed Central Asian rivers the correlation coefficient between δr and ξr (Fig. 8) is 0.72.

4. Comparing our estimations with Golosov's results [17] for a number of rivers in the Oka River basin, we notice their generally poor correlation (Table 3). This is especially notable for the estimation of the river bed component of erosion for the whole Oka River basin upstream from Kaluga City. Comparing this basin with the Upa River basin, which was mentioned above and forms part of the Oka River basin, we see, that with equal specific WD and SSF and equal shares of most erosion-hazardous lands (tilled lands) in these basins (55% in each), we can hardly expect similar percentages of river bed sediments (4.4% by Golosov) because the area of erosion-inactive lands (forests) differs three times between these basins. In relation to this, a question arises:

how can the equal specific SSF of these two rivers be formed when erosion in Upa's catchment area is even greater (25% higher, judging by the ratio of gullies density in these basins) and, probably, wider (due to smaller forest-coverage) than in the entire Oka catchment area? If we consider the erosion potential on other lands (not cultivated and not forest-covered territories) in the two rivers' basins to be more or less equal, then the difference in the basin-origin specific SSF of the Upa and the Oka can be completely (or significantly) compensated by sediments from the stronger river bed erosion (growth of δr) of the whole Oka River basin.

One reasons for the underestimation of the river bed SSF component of this river up to 4.4% (by Golosov) is, in our opinion, the technical difficulties of calculating sediment transportation (supply coefficient) from the eroded lands of a catchment area into streams with the increase in their area and, together with, the complication of intra-basin landscape structure. At the same time, we admit that the relatively high river bed SSF component in the Oka River, as calculated by method proposed here, may be partly due to the underwater mining of sand-gravel deposits along the bed of the river [23].

5. Conclusion

In spite of the above-mentioned drawbacks, the proposed method can be quite successfully applied to estimate of the structure of erosion and river suspended sediment flux within large regions of the Earth for which proper hydrological information exists (for example, the Northern Eurasia within the boundaries of the former USSR). This permits the zoning of their territories not only on the basis of the ratio of river bed and drainage basin components, but also by their intra-basin intensity, itself expressed both in area values (for basin erosion, t/km^2 a year)) and linear values (for river-bed erosion, t/km a year).The method also facilitates the geographical analysis of the factor stipulation of spatial variability for river bed and drainage-basin components of SSF.

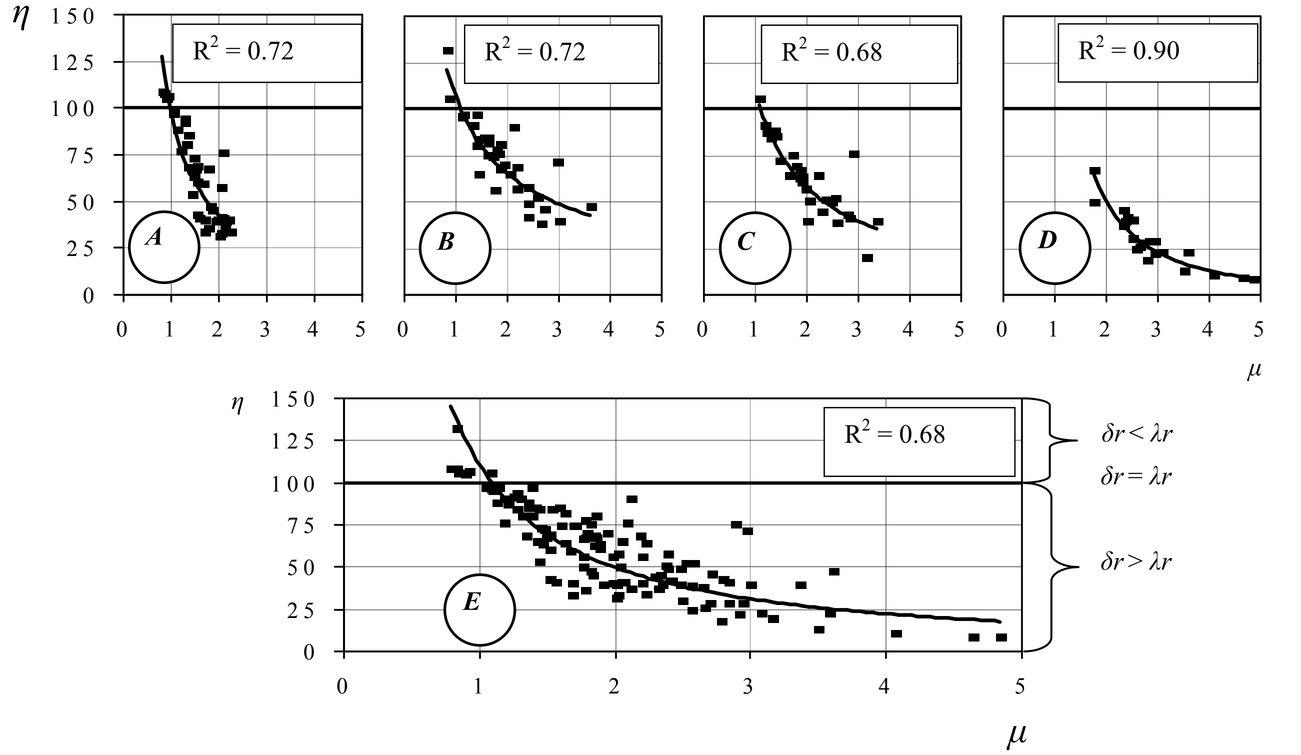

Fig. 7. Ratio of λr and δr assessments for river bed SSF component (η = ($\lambda r/\delta r$) × 100%)) depending on the change of power indicator μ in equation (3) on the altitudinal groups of analyzed river basins (*A* – the plain rivers; *B* – the low-mountain rivers; *C* – the middle-mountain rivers (the Caucasus); *D* – the middle-mountain rivers (the Central Asia); *E* – the all rivers); R^2 – determination coefficient for the power trend.

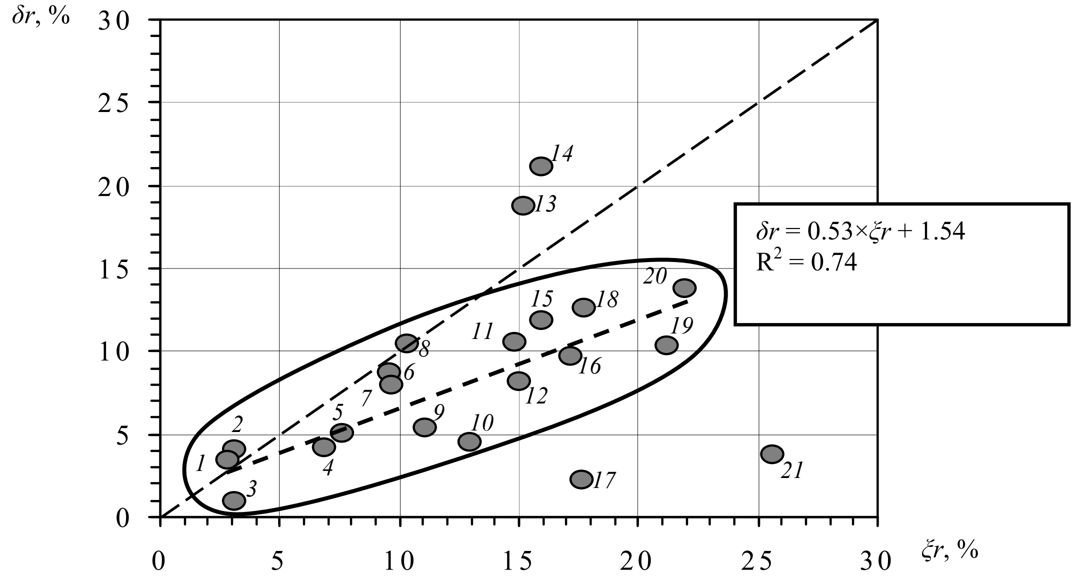

Fig. 8. A comparison of assessments of river bed SSF component calculated by the method proposed in the paper (δr) and by Shcheglova's method (ξr) [7] for the middle-mountain rivers of the Central Asia

Rivers: 1 – Tyup/Sartalgoi, *2* – Zeravshan/mouth of the Fandarya River, *3* – Karatag/Karatag, *4* – Issykata/Yuryevskoye, *5* – Aksu/Hazarnau, *6* – Naryn/Naryn, *7* – Ala Archa/mouth of the Kashkasu River, *8* – Chon-Kemin/mouth, *9* – Tentyaksai/Charvak, *10* – Djirlagan/Sovetskoye, *11* – Isfayramsai/θch Kurgan, *12* – Angren/Turk (above dam), *13* – Hodjabakirgan/Andarkhon, *14* – Kassansai/Uryukty, *15* – Yakkabagdarya/Tatar, *16* – Chatkal/Charvak, *17* – Ugam/Hodjikent, *18* – Isfara/Tash Kurgan, *19* – Varzob/Dagana, *20* – Pskem/Charvak, *21* – Chu/Kochkorka.

Table 3. Comparison of the assessments of river bed SSF component for some rivers in the Oka River basin (Russia).

River/ Hydrological station	*F*	*L*	*P*	*M*	*R*	*G*	River bed SSF, %	
							δr	by Golosov [17]
Moskva/ ?	500	46	25	?	?	12	–	2.3
Moskva/ Barsuki	755	?	?	6.9	11.6	?	22.6	–
Vad/ Avdalovo	1 930	37	35	3.6	8.4	37	29.4	5.1
Zusha/ Mtsensk	6 000	7	37	4.8	57.2	68	9.1	11.7
Upa/ Orlovo	8 210	8	55	4.9	19.7	60	9.6	4.4
Oka/ Kaluga	54 900	23	55	4.9	19.6	46	42.4	4.4

F: is a river basin area (km^2); L: is a residual forest area in river basin (%); P: is a tilled (cultivated) area (%); M: is a long-term mean annual runoff (L/(km^2 a sec)); R: is a long-term mean annual specific SSF (t/(km^2 a year)); G: is a gullies density in river basin (units/km^2).

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