Calculation of suspended sediment deposition intensity in grass-covered floodplains

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Abstract Flooded river valley meadows are very important for river ecology, as they entrap sediments and sediment-bound nitrogen or heavy metals. Nevertheless, the grass-covered floodplains, particularly in river deltas, are often separated from the rivers by dykes. When such systems are designed, it is necessary to model sediment deposition in these separated areas. However, the existing calculation methods and models are adapted for flow over the sandy bottom. In the meadow flows there are other boundary conditions more favourable for sediment deposition.

The focus of our studies was the processes of suspended sediment deposition in the flooded delta of the River Nemunas. Calibration results of the mathematical model with common sediment deposition formulae for riverbed flows did not correspond to the data of measurements. The differences were obvious. Consequently, it was necessary to study the peculiarities of flow and sediment motion under these conditions, as well as to work out formulae suitable for calculations.

Keywords Grassed flow bottom; sediment deposition

Introduction

Water quality of seas, gulfs and lagoons depends on the amount of fine sediments (especially clay and silt particles) brought in floodwater from the catchment's areas. These sediments usually contain large amounts of bioorganic matter and heavy metals (Deletic 2001). According to the existing considerations, the wash load containing fine particles with fall velocity $w < 1.6vi$ (v–depth-averaged velocity; i–water surface gradient) can be transported in significant quantities (Van Rijn 1993). Therefore sedimentation processes occurring in low-speed coastal waters and floodplain areas are of great importance. For example, due to sedimentation only about one-third of sediments transported by the Rhine reach the North Sea. The rest of the sediments deposit in lower reaches and floodplains of the river (Middelkoep and Haselen 1999). After periods of intensive flooding, the river can leave large amounts of clay and silt on its floodplain. The sedimentation process is especially intensive in inundated grassland meadows. However, particularly in deltas, these floodplains are often separated from the main river canal by dykes for intensive farming or other purposes. When these systems are designed, it is necessary to calculate the sediment deposition in these separated areas in order to estimate possible environmental damage (Wang *et al.* 2003; Jesper and Windolf 2003).

When studying the sediment deposition in the Nemunas delta it was estimated that sediment deposition in flooded grassland is much more intensive than in the flow over the sandy bottom, because in the case of flow in the meadows there are other better boundary

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conditions for sediment deposition. However, in the existing hydraulic mathematical models the same calculation methods are employed (Young *et al.* 1987; MIKE 21 1995; Roger and Lin 2003). Therefore the calculated hydro-ecological damage can be several times lower than the actual one.

To evaluate the influence of the sedimentation process on the reduction of pollution in the Curonian Lagoon and the Baltic Sea, suspended sediment deposition in flooded meadows of the River Nemunas delta was studied (Vaikasas and Rimkus 1996, 2003; Rimkus and Vaikasas 1999a, 1999b; Rimkus 1999, 2000a, b), and pollution of deposits with heavy metals and bioorganic matter was estimated (Gipiškis 1999). Field investigations on the sedimentation processes were performed during spring floods.

It was established that fine clay and silt particles deposited in the valley formed during the period 1950–1991 a 4–6 mm thick layer that contained about 50–60 t/ha of mud deposits with 250 t of potassium, 950 t of phosphorus, 38 000 t of calcium and even 147 000 t of organic matter saturated with nitrogen (Vaikasas and Rimkus 2003). The sedimentation differences between the results received from field measurements and those obtained by calculations showed that more exact calculation methods needed to be created. The ability of grass to entrap and retain sediments was the main reason for the differences. This feature has already been observed and investigated by other researchers. Grass cover was suggested through the creation of vegetation strips serving as barriers to retain sediments transported by runoff (Barfield *et al.* 1979; Thornton *et al.* 1997). The effect of water grass on the sedimentation process was investigated too (Pasche and Rouve 1984; Christiansen and Wiberg 1997; Carpena *et al.* 1999; Deletic 2001). However, the flow over grass-covered vegetative floodplains has been insufficiently studied and therefore a necessary investigation needed to be done.

In fact, we have previously observed the process of intensive sediment deposition on grass-covered channel slopes of canalised brooks of drainage systems. Grass-covered channel slopes were formed with silt and clay sediments; while on the bottom even sand particles were transported (Rimkus and Vaikasas 1999a).

Since the flow velocities were rather low in the valley of the Nemunas delta, mostly only the sedimentation process was observed in this area. The erosion process was absent; therefore, the subject of our study was the suspended sediment deposition process.

The specifics of suspended sediment deposition calculations in sandy riverbeds

Sediment deposition process in the riverbed has been studied much more thoroughly than in flooded valleys with meadows prevailing. At the beginning the recent calculation methods created for riverbeds and applied in the known mathematical models are discussed. In order to describe suspended sediment distribution in water flow, the two-dimensional advection–diffusion equation is used:

$$\frac{\partial \overline{C}}{\partial t} + u\frac{\partial \overline{C}}{\partial x} + v\frac{\partial \overline{C}}{\partial y} = \frac{1}{h}\frac{\partial}{\partial x}\left(hD_x\frac{\partial \overline{C}}{\partial x}\right) + \frac{1}{h}\frac{\partial}{\partial y}\cdot\left(hD_y\frac{\partial \overline{C}}{\partial y}\right) - S, \tag{1}$$

where $\overline{C}$–depth-average suspended sediment concentration; u, v–depth-average horizontal and vertical flow velocities; D_x, D_y–coefficients of turbulent diffusion in x and y directions; h–water depth; and S–deposition/erosion term (kg/m^3/s).

For the study of water flow, when sediment deposition is in process, the term S is of great importance. It is introduced through the bed boundary conditions. Formulae for the calculation of suspended sediment deposition in riverbed flows are based on the concepts of transportable sediment concentration, critical bed-shear stress for deposition, or critical flow velocity.

In the case of silt and clay sediments, the formula containing the shear stress critical for sediment deposition is usually applied:

$$S = \left(1 - \frac{\tau}{\tau_{cr}}\right)\frac{w\overline{C}}{h_*}, \quad \tau < \tau_{cr} \tag{2}$$

where S–decrease in sediment flow rate due to sediment deposition (kg/m^3/s); τ–bed-shear stress; τ_{cr}–critical value for sediment deposition of bed-shear stress; w–fall velocity of sediment particles; and h_*–distance from the bottom to the centre of sediment mass, calculated by the numerical integration of vertical distribution profile of sediment concentrations.

The bed-shear stress can be determined by

$$\tau = \rho g \frac{v^2 n^2}{h^{1/3}} \tag{3}$$

where v–average flow velocity; h–water depth; ρ–fluid density; g–gravity acceleration; and n–hydraulic roughness coefficient (Maning's coefficient).

When combining Equations (2) and (3), the deposition term S can be expressed by

$$S = \left[1 - \left(\frac{v}{v_{cr}}\right)^2\right]\frac{\overline{C}w}{h_*} \quad v < v_{cr} \tag{4}$$

where v_{cr}–average stream velocity critical for sediment deposition.

In this way the expression based on critical for sediment deposition flow velocity is determined.

In the case of sand sediments the formula for calculation of sediment deposition rate, $D = S{\cdot}h$, based on the concept of suspension transport capacity or transportable sediment concentration is usually applied:

$$D = \frac{Q_s - Q_{tr}}{Q}w = \left(1 - \frac{C_{tr}}{\overline{C}}\right)\overline{C}w \quad C_{tr} < \overline{C} \tag{5}$$

where D–deposition rate per unit of the bottom area (kg/m^2/s); Q_s–sediment flow rate (kg/s); Q_{tr}–sediment transport capacity (kg/s); Q–water discharge (m^3/s); $\overline{C}$–actual average sediment concentration; and C_{tr}–transportable sediment concentration.

Transport capacity or transportable sediment concentration of suspended sediment may be calculated according to well-known methods (Zamarin 1948; Velikanov 1958; Bagnold 1966; Grishanin 1969; Karaushev 1969; Coleman 1986; Van Rijn 1993; etc.). Formulae proposed by the authors are created on the basis of experiments performed in different flow conditions, therefore they give some different results.

The formulae discussed are created for calculation of suspended flow on the sandy bed. For such flows it is characteristic for the bottom sediment load with high sediment concentration. The sediment deposition process begins when the flow in this bottom sediment layer is saturated, i.e. when the transportable sediment concentration is achieved. The critical state of flow is estimated in Equations (2), (4) and (5) by the factor in brackets. The concept of critical velocity or transportable concentration cannot be applied for sediment deposition calculations for flow with grass-covered bottoms where neither the bed load of sediments nor high concentrations necessary for flow saturation are observed. The unique sediment deposition is constantly going on at the bottom in the flow over grass. Having measured the water turbidity, it was determined that suspended sediment concentration exceeded the average concentration at the grass level by only about 1.2 times, while the concentration near the bottom of the river Minija (tributary of the Nemunas River in its delta) was eight times higher (Vaikasas and Rimkus 1996). Consequently, the boundary conditions for sediment deposition in riverbed on grassed flood plains are quite different. Therefore the calculation equations, in both cases must be different as well.

Suspended sediment deposition in grass-covered floodplains

Stream velocities are low between the grasses. Each sediment particle settles down when it gets among the vegetation (similarly as in still water). Thus the ability of grass to entrap sediments increases. As a result, sediment deposition is constantly going on even when flow velocities are rather high or sediment concentration is low (Figure 1).

Under such conditions, sediment deposition in the grass layer is proportional to the fall velocity of sediment particles w and sediment concentration at the grass layer C_b, which forms the sediment concentration between the grasses. When this concentration is estimated, the sediment deposition rate per unit of the bottom area can be calculated as follows:

$$D = k_{cor} w C_b \tag{6}$$

where C_b–sediment concentration at the flow bottom; and k_{cor}–correction coefficient estimating possible difference between sediment concentrations in the grass layer and at the flow bottom (considering that the sediment deposition might be affected by the state of the grass).

For the calculations with Equation (6), sediment concentration at the flow bottom is to be estimated and vertical sediment distribution along the depth is to be considered. The laboratory investigations (Kouwen and Unny 1973; Christensen 1985; Temple 1986; Kouwen 1987; Yurchuk 1999) have shown that, under the conditions of flow over grass as well as over sandy bottom, the logarithmic velocity distribution along the depth is observed. Consequently, the turbulent structure of flow created by turbulent vortices and the vertical sediment distribution in both cases would be similar. Therefore, for the calculation of concentration C_a on grass-covered flow bottom the Rouse equation is applied:

$$C = C_a \left(\frac{h-y}{y} \frac{a}{h-a} \right)^z \quad z = \frac{w}{\beta k u_*} \tag{7}$$

where C–sediment concentration at the distance y from the bottom; C_a–concentration at the distance $y = a$; z–Rouse number; $k = 0.4$–Van Karman number; u_*–shear velocity; and β–ratio of the sediment and momentum diffusion coefficients ε_s and ε_m.

For the hydraulic calculations the water depth h of the flow over a grass-covered floodplain should be measured from the surface of grass layer with the hydraulic equivalent thickness (Figure 1), which may be calculated according to the methodology discussed in the articles by Rimkus (1999, 2000a, b). An approximate thickness h_{eq} of this layer is about $0.7h_{gr}$ (h_{gr}–thickness of grass layer).

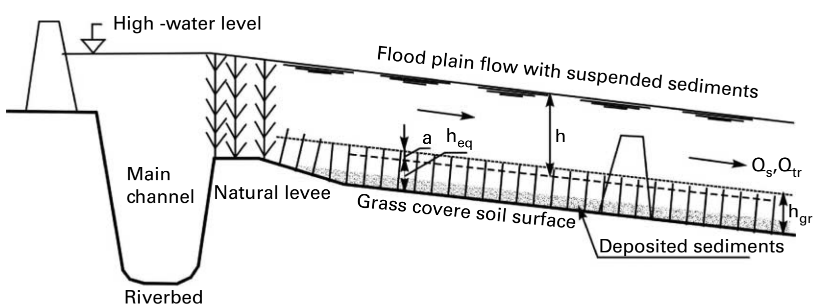

Figure 1 Longitudinal profile of grass-covered floodplain

The value a in the Rouse equation is usually accepted to be equal to the thickness of a bed load layer. For the flow above grass, this value may be considered as the difference between the thickness of the natural and hydraulic equivalent grass layer (Figure 1). Thus

$$a = 0.3h_{gr}. \tag{8}$$

The concentration C_a in these conditions becomes equal to the searched sediment concentration C_b at the surface of the grass layer. Thus $C_b = C_a$. Usually the amount of sediments in the water flow is expressed by the average in the stream concentration $\overline{C}$. Consequently, for the calculations in Equation (6) the ratio $F = \overline{C}/C_a$ is to be estimated. The formula for the calculation of this ratio is obtained through integrating the following equation:

$$dq = C_a v_y dy \tag{9}$$

where dq–sediment discharge through the depth element dy; v_y–flow velocity at the distance y from the bottom; and C_a–sediment concentration at the distance a from the bottom, calculated according to the Rouse equation (7).

Then the following expression is obtained:

$$F = \frac{\overline{C}}{C_a} = \left(\frac{a}{h-a}\right)^z \left[\int_a^h \left(\frac{h-y}{y}\right)^z v_y dy\right] \cdot \frac{1}{\int_a^h v_y dy}. \tag{10}$$

In this formula the flow velocities v_y can be estimated, for example, according to their logarithmic distribution.

Applying the ratio $F = \overline{C}/C_a$, Equation (6) becomes as follows:

$$D = k_{cor}\overline{C}w/F. \tag{11}$$

The shortcoming of Equations (7) and (10) is that they estimate the concentration $C = 0$ at the water surface. This property of the formula originates from the expression of diffusion coefficient, which is used when creating the Rouse equation. The Rouse equation was derived having integrated a one-dimensional diffusion–convection equation of steady flow, which expresses the equilibrium between the sediment deposition and the ability of flow to lift the sediments. It is as follows:

$$wC + \varepsilon_s \frac{\partial C}{\partial y} = 0 \tag{12}$$

where ε_s–diffusion coefficient of sediments.

The Rouse equation is obtained having integrated this equation when the following parabolic dependence for ε_s is applied:

$$\varepsilon_s = \beta k u_* y\left(1 - \frac{y}{h}\right). \tag{13}$$

The parabolic distribution Equation (13) is based on a linear shear stress distribution and a logarithmic velocity profile. The aforementioned shortcoming of the Rouse formula arose because this expression was applied. The analysis of measured concentration profiles (Coleman 1986) shows that expression Equation (7) is always exact when $y < 0.5\,h$. For $0.5 < y/h < 1$, the constant ε_s is more convenient:

$$\varepsilon_s = 0.25\beta k u_* h. \tag{14}$$

Having integrated Equation (12) with the ε_s expression from Equation (14), the following formula for vertical sediment distribution is obtained (Van Rijn 1993):

$$C = C_a\left(\frac{a}{h-a}\right)^z \exp\left(-4z(y/h-0.5)\right) \quad y \geq 0.5h \tag{15}$$

When applying both Equations (7) and (15) for more exact calculations, integration in formula (10) is to be performed only from $y = a$ to $y = h/2$. Then Equation (15) is to be integrated between the limits $h/2 < y < h$. Consequently, the ratio $F = \overline{C}/Ca$ is defined as follows:

$$F = \frac{\overline{C}}{C_a} = \left(\frac{a}{h-a}\right)^z \left[\int_a^{h/2} \left(\frac{h-y}{y}\right)^z v_y dy + \int_{h/2}^{h} \exp\left(-4z\left(y/h-0.5\right)\right) v_y dy\right] \frac{1}{\int_a^h v_y dy}. \tag{16}$$

During the investigations in the Nemunas delta, the thickness h_{gr} of a grass layer bent by snow was about 20 cm. When the hydraulic equivalent thickness of the grass layer equals $0.7\,h_{gr}$, the value a is equal to $0.3\,h_{gr} = 6$ cm. Average water depth h was about 2.0 m in the studied floodplain, and thus the ratio a/h was equal to 0.03. As the test calculations show, the change of the ratio a/h has no significant effect on the values of the ratio $\overline{C}/C_a$. This is because the maximum values of sediment concentration C are observed at the bottom where the flow velocities are low; therefore they cannot affect the average values of $\overline{C}$ significantly. Thus the average value of the ratio a/h may be used for calculations in the whole studied floodplain area.

To do the calculations with the Rouse equation, the estimation of the value of coefficient β is needed. According to the investigations carried out by Cellino and Graf (1999), this value for sediment particles having a small diameter $d < 0.1$ mm is less than unity, as such particles cannot be thrown to the outside of turbulent eddies by centrifugal forces. Therefore dependence of β on the sediment diameter is also insignificant. It depends on the distance from the bed when the sediment concentration is close to capacity. For small concentrations it is constant along the depth. The depth-averaged value β was received depending on the sediment concentration (Figure 2). Higher sediment concentrations determine lower β values, as the turbulence is being suppressed in this case. Our investigations were performed with low concentrations (12–22 mg/l), and therefore $\beta = 0.6$ (Figure 2).

The values of factor F calculated with Equations (10) and (16) when $u_* = 0.1$ m/s are given in Table 1. They do not differ significantly; therefore formula (10) can be applied if such precision is sufficient. For fine particles with $d < 0.02$ mm the value of F is close to unity, therefore it is little influenced by the selected value of β or the kind of formula for F calculation.

Sediments transported by the river flow usually contain aggregates formed by cohesive clay particles. Their deposition was calculated applying an equivalent sand diameter in fall

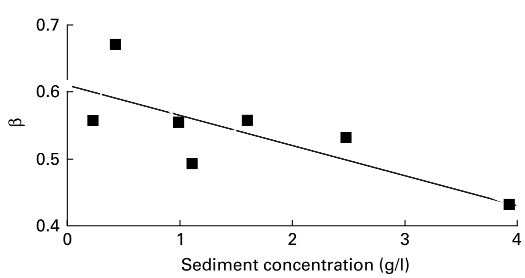

Figure 2 Dependence of parameter β on the mean sediment concentration $\overline{C}$ (according to Cellino and Graf (1999))

Table 1 The values of factor F obtained with Equations (10) and (16)

Diameter of particles, mm	0.002	0.005	0.01	0.02	0.05	0.1
Factor F according to Equation (16)	0.9996	0.997	0.989	0.956	0.768	0.365
Factor F according to Equation (10)	0.9994	0.995	0.986	0.945	0.732	0.347

velocity calculations, i.e. the diameter of sand particles with the same fall velocity. This diameter, as well as the diameter of loose, not round, clay particles, can be estimated by sedimentation type of grain-size investigation. When the sediment deposition is calculated, it is very important to use the same formulae for the calculation of fall velocity of sediment particles that were applied when the grain-size investigations were performed. This is necessary for exact calculations.

Sediment concentration in Lithuanian rivers is rather low ($\leq 0.5\,kg/m^3$). Fall velocity of clay particles with high concentration (more than $10\,kg/m^3$) is to be calculated according to special formulae.

As mentioned above, sediment deposition may depend on the state of the grass layer. The depositing sediments first settle on grass stems and leaves on the surface of the grass layer. The bottom vortices are able to raise up these particles into the stream again. This process is more intensive when grass stems are bent by the flow in the direction of the stream; then the coefficient k_{cor} will also be lower. However, stream velocities are usually low in the flood plains and the grass is bent only by snow here. Naturally, the grass stems are bent in different directions, i.e. not only along the stream. The sediment particles washed from grass leaves by the bottom flow velocities appear among the grass stems. Most of them continue settling down. Thus the coefficient k_{cor} will be slightly below 1.0 in this case. Such were the conditions in the case of our investigations. The influence of the state of the grass has not been investigated yet; therefore sediment distribution and deposition processes in grass-covered floodplains need to be studied further. Unfortunately, no more floods suitable for investigations have occurred up to now. The problem for further studies to be performed is that the investigations need to be carried out in natural flow conditions, because the exact laboratory modelling of meadow flows with fine silt and clay sediments would be very complicated.

Calibration of formulas

According to the discussion above, equations (2)–(5) created for flows on sandy bottom cannot be applied for the modelling of the flow over meadows because of different boundary conditions. However, they are used in recent mathematical models (Young *et al.* 1987; MIKE 21 1995; etc.) created for the calculation of suspended flow.

We could not use the existing mathematical models for our calculations because they are closed for alternations; therefore it was impossible to include our formulae in these models. Thus, we had to create our model adapted for our investigations (Rimkus and Vaikasas 1999a, 1999b).

A 23-km long interval of the Nemunas floodplain valley was used for the field investigations, mathematical modelling and calibration. The areas under the flow are covered with grass vegetation. The valley relief was taken from maps with scales 1:2000 and 1:5000. Water sampling points and cross sections used for the calculations of sediment and flow strips are shown in Figure 3 (scales of length and width are different). Field turbidity measurements were taken during spring floods in the period 1994–1996. The sediment concentrations measured in several cross-section points are presented in Table 2. Water

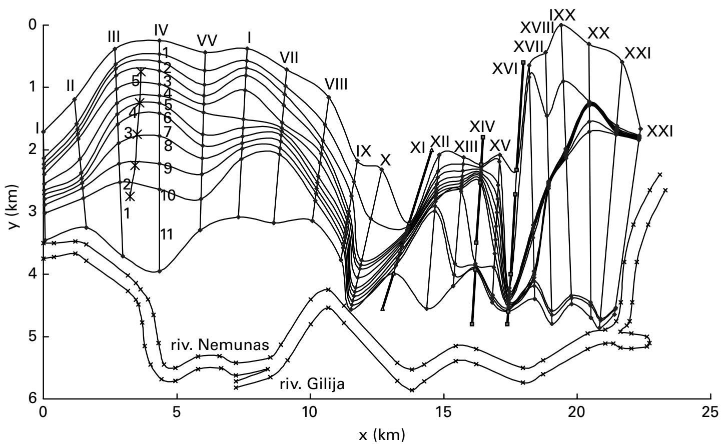

Figure 3 Distribution of computerised stream strips at the investigated section of the Nemunas delta (the flood of 1996): I–XXII–cross-section numbers; 1–5–numbers of measurement posts; and 1–11–numbers of flow strips

discharge flowing *via* the floodplain was 507 m^3/s during the flood investigations. The flood was already decreasing slightly. The flow velocities were 0.1–1.0 m/s in the flood plain.

The necessary for flow calculations for the shape of valley cross sections was presented in the tables as dependence of bottom levels in characteristic points on the distance from the axis line. Water flow for calculation was divided into flow strips with equal discharges. The water discharges in cross sections and in flow strips were calculated adding up in the cross sections the elementary discharges $\Delta Q = vh\Delta x$ (v–velocity; h–water depth; and Δx–length of cross-section elements) until the discharge of next flow strip in turn was achieved. Thus the strip width and average flow velocity in the strip were estimated. Flow velocity is calculated with Manning's formula $v = R^{2/3} i^{1/2}/n$. Increase of water level in adjacent cross sections was estimated according the one-dimensional SaintVenant's equation:

$$-\frac{dz}{dl} = \frac{v^2}{C^2 h} + \frac{d}{dl}\frac{v^2}{2g}. \tag{17}$$

Sediment deposition was calculated along each flow strip, and thus the distribution of settled sediment in the valley area was determined and our model became quasi-two-dimensional. A similar calculation method was applied for example in the model MIKE 11.

Further the calibration results are discussed.

Equation (5) was tested applying Zamarin's equation for transportable sediment concentration C_{tr} (Zamarin 1948):

$$C_{tr} = 0.022 \frac{v}{w_0} \sqrt{\frac{Riv}{w}} \tag{18}$$

Table 2 Measured concentrations, of turbidity, in mg/l

III						XXII
Point numbers						
1	2	3	4	5	Average 1–5 points	
8.0	10.8	12.6	14.2	15.0	12.5	22.0

where v – average stream velocity, m/s; R – hydraulic radius, m; i – channel slope. $w_0 = 0.002$ when $w < 0.002$ m/s (for silt and clay); $w_0 = w$, when $w \geq 0.002$ m/s (for sand).

According to this equation, no suspended sediments were able to deposit during the modelled flood in the Nemunas delta valley, though field measurements showed contrary results. Such calculation results were determined because the transport capacity C_{tr} calculated with Equation (18), appeared to be higher than the actual concentrations in the flooded valley.

To test Equations (2) and (4), which are applied for the calculation of the deposition of silt and clay particles (only these particles were found in our case), the critical velocity value for the deposition v_{cr} was obtained as follows (MIKE 21 1995):

$$v_{cr} = 1.25w\frac{h^{1/6}}{n\sqrt{g}}. \tag{19}$$

No sediment deposition was obtained by applying these formulae either. Thus, the study results showed that the formulae derived for riverbeds could not be used for flows over meadows, as the transport capacity calculated with them is too large.

Equation (5) was also tested by Bagnold's equation adjusted by Young *et al.* (1987) for sediment deposition calculations in small channels:

$$q_{tr} = \eta k \tau v^2 / w \tag{20}$$

where η – effective sediment transport factor; k – transport capacity factor; τ – shear stress; and q_{tr} – suspended transport capacity rate in the flow.

According to the above-mentioned modification, the following expression was obtained:

$$\eta = 0.74\left[\frac{(\gamma_s - \gamma_w)d}{\tau}\right]^{1.98} \quad k = 0.001 \tag{21}$$

where γ_s and γ_w – specific weight of sediment and water; and d – particle diameter.

Equations (20) and (21) have been applied for the AGNPS model to calculate sediment deposition in concentrated storm water streams from agricultural fields (Young *et al.* 1987). The value of sediment transport capacity calculated with these formulae is less than its value obtained from Equations (18) and (5). Therefore in applying Equations (20) and (21), the sedimentation process was found to appear in the Nemunas delta valley. With the help of those formulae the calibration was performed. For this the composition of sediment fractions brought by the river stream into the entry section XXII has been chosen by calculations so that sediment concentration in the test section III after the sediment deposition until this section would be equal to the value measured here. Suitable formulae should give then the real sediment composition in the entry cross section. Calibration results received from those formulae and our equations are compared in Figure 4.

Applying Equations (5), (20) and (21), it was necessary to include in this composition not only silt fractions, but also a large amount of fine sand particles ($d = 0.05$ mm). Otherwise the total transported amount of sediments obtained might have been much smaller than the measured value. According to the results obtained by these formulae mainly sand sediments would have been brought into the valley, although no sand fractions were found in water samples taken during the flood in 1996. These results show that formulae resulting in low transportable concentrations cannot ensure correct calculation results either.

The grain size composition found from our Equations (11) and (16) was natural (Figure 4). It was received with coefficient $k_{cor} = 0.8$.

Four sediment size fractions were chosen for the calibration. However, field measurements were only taken in two sections, though for the required calculations the number of investigated control sections should also be four. Therefore, the concentrations of

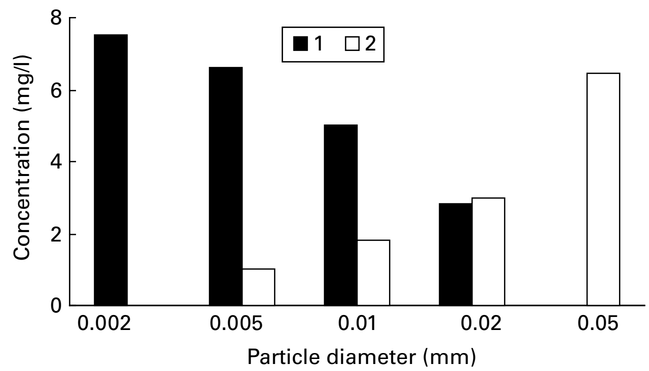

Figure 4 Comparison of calculated sediment composition in the entrance section: 1 – calculation with our formulae (11) and (16);and 2 – calculation with formulae (5), (23) and (24)

the rest of the fractions were chosen so that the selected concentrations and the ones estimated by calculations would appear to be in regular dependence (Figure 4). This regularity shows that the selection of concentration values is made successfully. Unfortunately, a more thorough field study has not been carried out yet, as recent floods in the Nemunas delta have been too low.

Calibration was also performed considering the measurement data from the flood in 1994. This was the first year of our investigations when we obtained the data that was used to plan our further work. Measurements were taken in one point of each cross section. Sediment concentrations were low during this flood; therefore they were less suitable for the investigation. The concentrations of turbidity were found to be 10.0 and 3.4 mg/l in cross-sections XXII and III, respectively. The results of calibration were similar, i.e. sediment compositions calculated with Equations (20) and (21) at the entrance section were not exact either; they contained mostly fine sand, whilst no sand was found in any of the water samples.

Calculation by calibration sediment composition in entry cross section depends on the applied meaning of k_{cor}. This meaning is to be selected so that the calculated composition would be similar to the one measured in this cross section. The sediment composition in water samples taken during the investigation in 1996 was not estimated, as the sediment concentrations were too small for the instruments available in our institute at this time. Therefore the supposed $k_{cor} = 0.8$ meaning was chosen for calibration. Sediment compositions received with other k_{cor} meanings seemed also natural. For the investigation of sedimentation process by measurement of decreasing sediment concentration along the flow is possible only in the long strip of flood plain and when the flow is sufficient high. Thus these investigations could not be repeated in the Nemunas delta till now. Therefore it was decided to perform the investigations for this aim by estimation of the amount of sediments deposited during the flood. It is necessary for it to repeat the measurements of sediment concentration and grain size composition at the grass surface during all the flood. It was made during the spring flood 2006 in the river Nevęžis close to our institute. The instrument used for measurements of sediment concentration was based on infrared radiation. The applied work method with this instrument is described in the paper of Rimkus (2000a, b; Rimkus *et al.* 2004).

This flood was small, and the places prepared to gather the deposited sediments were flooded only for the point with not mowed grass with stem length 0.4 m. Thus the investigations became few. The investigated place was flooded for 6 days, i.e. from 30 March to 4 April. The summary amount of deposited sediments washed from the grass samples in

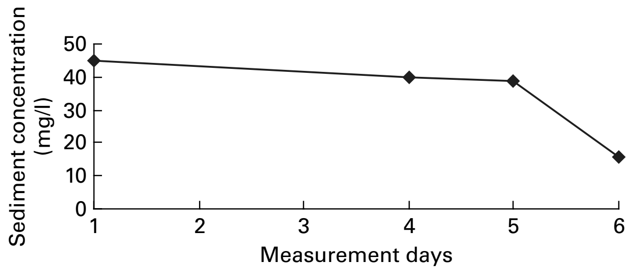

Figure 5 Sediment concentration exchange over the investigated point during the measurement days

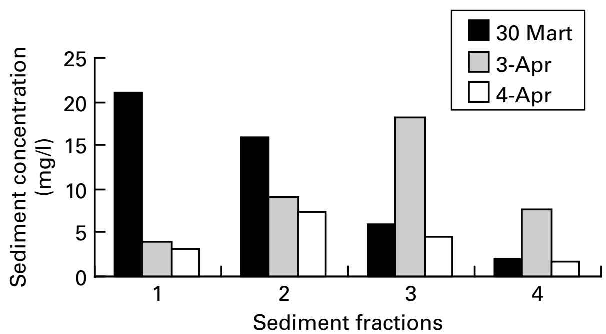

Figure 6 Grain size composition of sediments during the flood at measurement point. Sediment fractions: $1-d = 5-10\,\mu m$, $2-d = 2-5\,\mu m$, $3-d = 1-2\,\mu m$, $4-d < 1\,\mu m$

this place was 179 g/m^2. The change of sediment concentration and sediment grain size composition in the water over the studied point measured during the flood is shown in Figures 5 and 6. Sediment concentration was decreasing with the decrease of flooding. Summary deposition of sediments calculated with Equations (11) and (16) equal to their measured amount was determined when the coefficient $k_{cor} = 0.7$ was applied. It corresponds to the investigated grass state. The meanings of k_{cor} in other states of grass surely will be different; therefore these investigations must be continued.

Sedimentation values calculated within the studied area of the delta during the flood of 1958 and the period 1950–1991 are given in Table 3. These calculations showed that nearly 40% of total suspension transported by the river was deposited in the studied part of the Nemunas delta. Thus the floodplains of rivers overgrown with grass are very important as sediment traps. This phenomenon should be estimated by sediment transport calculations. Obviously, special formulae are needed for the calculations of sediment deposition in flows over meadows at least with approximate meanings of the coefficient k_{cor}.

Table 3 The sedimentation values (t/ha) in the studied area of the Nemunas delta during the flood of 1958 and the period 1950–1991

	Diameter of sediment particles (mm)				
Calculation cases	**0.005**	**0.01**	**0.02**	**0.05**	**Total**
Flood of 1958	56.4	175.7	332.1	187.6	751.8
Sedimentation during the period 1950–1991	240.4	555.7	435.0	622.4	2253.5

Conclusions

1. Sediment deposition in the flows over grass-covered valleys cannot be calculated by methods developed for riverbeds because of different bottom sediment motion and deposition conditions. The main factors determining the high deposition rate include low velocities in the grass and the ability of grass to entrap sediments. New formulae are suggested for sedimentation calculations in this case.
2. Floodplains of rivers overgrown with grass increase significantly the deposition of suspended sediment; therefore their existence is to be estimated by suspension transport calculations.

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