Morphological expressions of river sediment transport and their role in channel processes

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Abstract Relationships between erosion, sediment transport and deposition processes in river channels under different conditions of sediment yield formation and variable contribution of suspended (basin-derived) and bed (channel-originated) sediment are discussed. The role of the latter in total sediment yield and as a factor of morphological channel type differentiation is shown.

Key words channel deformations; channel processes; morphological type; runoff; sediment transport

INTRODUCTION

Sediment yield is the major driver of channel processes. The latter can also be considered as an expression of sediment yield (Makkaveev, 1955; Kondratiev *et al.* 1959). Fluvial erosion (mobilization of sediment particles from deposits or lithogenic substrate through their interaction with water flow to form the sediment load), sediment transport (movement of particles through the fluvial system) and sediment deposition (reverse transformation from one category to another) make up the complex of channel processes (Makkaveev & Chalov, 1986; Kondratiev *et al.*, 1982; Chalov, 1997). Sediment can originate from the channel itself if bed erosion occurs under the condition of virtual velocity V_{vir} excess over the non-eroding velocity V_n, i.e. $V_{vir} > V_n$, (Fig. 1). Sediment can also be basin-derived, if it is delivered to the channel from drainage basin sources. The latter mainly comprises suspended load, whereas the sediment of channel origin comprises bed load, although it may become suspended load if fine flood plain deposits, deposits in backwaters etc. are eroded. Furthermore, gullies, talus on mountain slopes, debris flows etc. provide a source of bed load in rivers.

Unlike water runoff, values of total sediment yield and contribution of its suspended and bed load components remain difficult to assess. Main reasons for that are insufficient number of monitoring points, short periods of observations, laboriousness of measurements, lack of reliable information on bed sediment transport and some others. The main aim of this paper is to consider relationships between sediment yield and forms of transport in river channels on one side and their expression in channel morphology and dynamics on the other, and to give geographical evaluation of these differences.

GEOGRAPHICAL VARIATION OF SEDIMENT YIELD AND RIVER CHANNEL MORPHOLOGY

Bed load and part of the suspended sediment load comprise the channel-forming sediment. Cessation of its transport results in the "prevailing content in the channel bed deposit" (Karasseff, 1975). However, the proportion of suspended load comprising the channel-

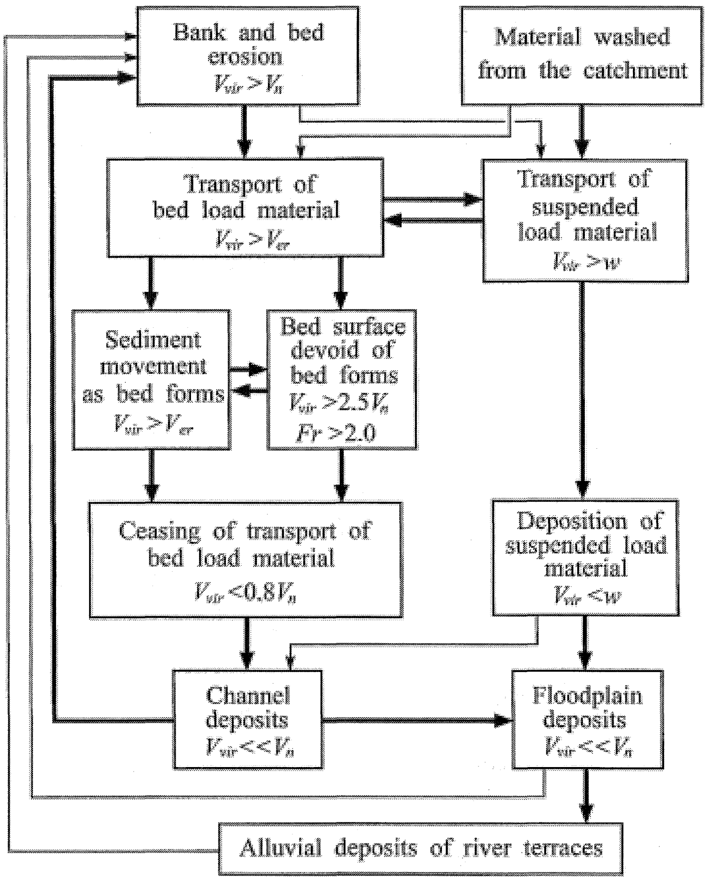

Fig. 1 Scheme of channel erosion interaction, transport and sediment accumulation in channel processes (arrow thickness defines primary or secondary role in material transport). V_{vir}, average virtual flow velocity; V_n, non-eroding velocity; V_e, eroding velocity; w, hydraulic size; Fr, Froude number.

forming sediment, as well as the ratio of suspended load to bed load yield W_S/W_B in the total sediment yield vary over a wide range of values. This range is controlled by the geographical laws of sediment yield formation, transport capacity R_{tr} conditions, and spatial–temporal changes of channel flow hydraulic characteristics (from axis to the peripheral channel parts, from high-water to low-water period etc.). The role of suspended sediment increases under the domination of finer material in the overall basin sediment yield. The coinciding tendency of increase is observed for total sediment yield W_{S+B} (Table 1), which increases from north to south and reaches its maximum in the rivers of southeastern Asia.

Suspended sediment-dominated rivers

The main sediment sources of the southern rivers are plateaus composed of loess or weathering crusts. As a result, basin-derived material dominates the sediment load, which is

Table 1 Annual sediment yield of several large rivers of Eurasia (based on Chalov *et al.*, 2000; Shtankova, 2002).

River-gauging station	Annual discharge Q, ($m^3 s^{-1}$)	Sediment load Suspended load W_R, ($t \times 10^6$)	Bed load W_G, ($t \times 10^6$)	Total W_{R+G}, ($t \times 10^6$)	$\frac{W_G}{W_{R+G}}$
Vyatka–Kirov		0.50	0.54	1.04	52.0
Lena–Kusur	7150	22.66	17.45	40.11	43.5
Sev. Dvina–Abramkovo	1930	1.3	0.7	2.03	35.0
Oka–Murom		1.54	0.40	1.94	20.5
Belaya–Birsk		2.62	0.54	3.18	12.1
Ob–Kolpashevo	4260	16.65	2.89	19.54	14.8
Ob–Barnaul	1470	8.10	0.57	8.67	6.6
Yangtze–Yichang	14300	512	13.08	525.1	2.48
Yellow River–Mijin	1523	1320	19.2	1321.9	0.14
Vichegda–Fedyakovo	1010	1	0.54	1.54	0.35

in turn dominated by the suspended component. It provides the maximum loading of flow by sediments ($R_{tr} \leq W_{S+B}$). Changes of R_{tr} and the hydraulic characteristics of river flow are caused by the irregularity of the temporal pattern of runoff. When river discharge decreases, a portion of the suspended load can be transformed into bed load. With $V_{vir} >> V_n$, bed load transport takes place without dune formation. Under such conditions bed sediment is deposited during the flood recession, forming continuous shoals. During low water periods its transport ceases completely, thus transforming bed sediment into channel deposits. Dried out shoals make high-level channels look like sandy fields extending along the rivers, among which the low-level channel forms meander (sediment bodies—point bars) or braids (sediment bodies—braid bars).

Subsequent re-entrainment of that sediment during the following high-water period causes wandering of channels, classified by the intensity of lateral deformation as absolutely unstable channels (the Yellow River, the Amu-Darya River): Lohtin index $L = d/I < 1.0$, where d is average sediment size and I is channel bed gradient (Chalov *et al.*, 2000). In turn, it causes channels to maintain relatively straight courses, since the numerous dried sediment bodies have no time to be fixed by vegetation and transformed into fragments of flood plain (meanders spurs, islands). This is illustrated by the plot of the changes in the position of the flow dynamic axis in the Yellow River (near Zhengzhou) during the high water period (Fig. 2). It shows that wandering occurs within a strip of more than 2 km wide, with an average rate of lateral channel shift from 25 to 144 m day^{-1}. These inter-annual dynamics are superimposed onto a background of longer-term changes (i.e. 3–5 years) within a strip up to 10.6 km wide (Chalov *et al.*, 2000). On the Amu-Darya River (Berkovich *et al.*, 1973) seasonal flow dynamic axis migration takes place within a strip up to 1.3 km wide with long-term position changes near the left or the right bank occurring every 7–10 years.

Condition $R_{tr} < W_{S+B}$ causes sediment deposition in the channel, i.e. preservation of sediment in the form of channel bed deposits. It results in aggradation of the river bed. For example, the Yellow River in its lower reaches aggrades at a rate of 4.5–7.7 cm year^{-1} (Chalov *et al.*, 2000), the Amu Darya River—3.4 cm year^{-1} (Zvetkova, 1963), and the Brahmaputra River—3.4 cm year^{-1} (Gusarov, 2002). Such river channels are buffered by natural levees, and even low-water level exceeds the elevation of the adjacent land (on the Yellow River up to 7 m).

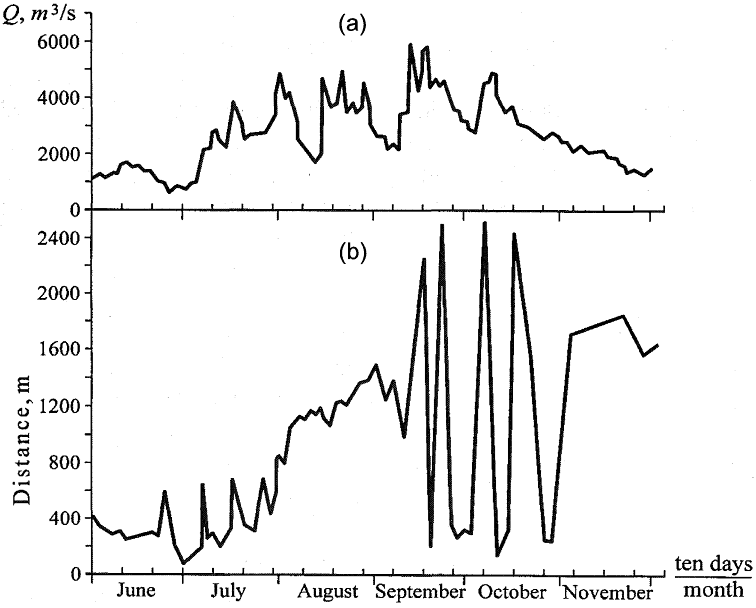

Fig. 2 Changes in the Yellow River flow dynamic axis through the high-water period (based on Chien, N. & Wan, Z., reproduced from Chalov *et al.* (2000)): (a) hydrograph; (b) flow dynamic axis position.

Bed load-dominated rivers

When coarser material (sand, gravel) is delivered into the channel, the basin-derived component of the sediment load decreases and bed load increases. As bed sediment availability increases, a hierarchy of bed forms is produced. Its structure depends on the flow discharge at different phases of the hydrological regime and position in the channel (axis or peripheral parts). During high floods, bed forms of all generations are actively transformed over the entire channel width. At low water macro- and medium-size bed forms (point bars and braid bars) stabilize, their higher parts emerge, and microforms (dunes) move only between the latter. The width of the belt of active sediment transport at low water may decrease up to two–three times.

Each dune is a depositional landform, in which sediment is temporarily stored, forming a stable nucleus. However, erosion on the dune stoss side leads to the entrainment of sediment, which accumulates again on the lee side. Active dunes as depositional channel bed forms are not necessarily evidence of the formation alluvial deposits. They may represent a temporary cessation of transport of sediment particles, which resumes with the process of dune movement as a whole. Nevertheless, each dune possesses a potential ability of transformation, partially or completely, from such seasonal deposits into longer-term alluvial deposits. It occurs through the influence of dunes on the kinematic structure of channel flow, causing erosion of channel banks opposite to drying dunes, vegetation development on dune surfaces and their transformation into flood plain segments without further movement. The stabilization of channel bed forms by vegetation is accompanied by deposition of suspended load material (with hydraulic size $w > V_{vir}$) on their surface and the initiation of thin flood plain deposits. Accumulation of suspended material is the dominant process on the flood plain surface. An increase in high-water suspended load (and, consequently, flow turbidity) causes the intensity of overbank deposition and the depth of the thin flood plain deposits to increase. As a result, the initial bends in the flow dynamic axis between point bars are transformed into channel meanders, and by flow divergence around braid bars into braided channels (point bars form meander spurs, braid bars—islands). Thus, a channel of a certain morphodynamic type is formed, where the further movement of bed forms determines lateral channel deformation (meander development, redistribution of discharge, deepening and shallowing of branches).

The major condition allowing such bed form transformation into channel form is the higher stability of sandy and gravelly alluvial channels (as a rule, Lokhtin index $L > 2$–3) in comparison with silty-sandy channels. Relatively low bed load transport is usually associated with straight channels, since there are no slowly moving macroforms of the channel bed to form initial planform irregularities. First order streams are usually characterized by relatively straight channels, since channel-flow interaction does not supply the stream with sufficient amounts of sediment to generate dunes. Relatively straight channels are also maintained by large rivers with unstable channels characterized by highly mobile bed forms. The latter have no time to be stabilized by vegetation to form new flood plain segments. This is also the case for rivers where suspended load material provides the channel-forming sediment (Chalov *et al.*, 1998). The largest bed forms (riffles) are characterized by long-term and seasonal changes: accumulation at high water (floods), erosion of the riffle crest at low water. However, the opposite situation occurs when a series of riffles/pools develops. Continuing movement of smaller-size bed forms during low water periods leads to an increase in the channel depth range and a decrease of riffle depth, causing shallowing at low water (Turykin, 1997).

Sandy and silt-sandy rivers are usually characterized by relatively balanced time-averaged values of load and transport capacity, i.e. $W_{S+B} \approx R_{tr}$. Usually R_{tr} slightly exceeds W_{S+B}. That is the main reason for the dominant tendency of slow (<1 mm year^{-1}) incision for many rivers. However, this can only be observed over historical or geological intervals of time. Possible reflections of this slow incision are steps in the flood plain topography as well as transformation of upper flood plain levels into unflooded terraces. The inverse situation (W_{S+B} slightly exceeds R_{tr}) is found in lower reaches and river mouths. As the excess of W_{S+B} over R_{tr} is usually low, rates of direct sedimentation are normally low and compensated by overbank accumulation on the flood plain. As a result there are no natural levees on the flood plain, but overlying flood plain segments (where younger deposits overlie older ones) or single-level flood plains can form (Makkaveev & Chalov, 1986).

The formation and movement of dunes in rivers with gravel-cobble bed load is very specific. Suspended load material does not take part in this process. During low-water periods, when $V_{vir} < V_n$, water turbidity decreases to around zero and sediment transport stops. On a flood rise, appreciable sediment transport only takes place after destruction of the armouring, when $V_{vir} > V_n$ (for relatively uniform armour-forming sediment particles). At this moment, the underlying mixed-size alluvium is characterized by $V_n << V_{vir}$. Avalanche-like growth of the bed load discharge takes place due to practically instant entrainment of finer material, as well as because of the decrease of V_n at increased sediment concentration in the flow. During the flood recession, transport of the largest particles ceases first. These produce armouring that overlies and protects the mixed-size deposit of the channel bed. On the other hand, because of the sediment coarseness and the presence of armouring, the hierarchy of bed forms is reduced. Usually it is limited to macro- and partly medium-sized forms. Those are largely more or less stable, and not eroded during flood recession and low-water periods. Therefore most gravel-cobble channels can be classified as stable or completely stable (Lokhtin index $L > 10–50$).

Usually in rivers with gravel-cobble alluvium $W_{S+B} << R_{tr}$. Because of a lack of sediment, bedrock channels (non-alluvial river reaches with channels cut in bedrock) can form (Chalov, 2003). Variations in geological structure then begin to play a leading role in determining channel morphology. For example, a change from volcanic to sedimentary rocks leads to an increased sediment supply, channel widening, and a decrease of R_{tr}. A similar situation is observed when mountain rivers enter piedmont areas or intermontane depressions. Rapid transition from constrained flow in a channel occupying the entire bottom of an incised gorge-like valley to relatively unconstrained flow in a widened braided channel takes place. The R_{tr} reduction here is also associated with a decrease in gradient. As a result, high sediment transport is replaced by the formation of dunes. These form gravel-cobble fields, within which the river is broken up into numerous braids (analogous to the sandy fields of the wandering rivers with complete domination of the suspended load described above).

In mountain rivers extremely high stream velocities determine specific modes of sediment transport and bed form movement. At relatively low channel gradients for Froude number Fr from 0.8 to 2.0, bed sediment is transported forming antidunes (so-called alluvial mountain channels with developed bed forms), whereas for $Fr > 2.0$ massive unstructured (so-called alluvial mountain channels without bed form generation) sediment transport is observed. At higher channel gradients, non-alluvial channels form (so-called channels with rapids and falls), which are able to transport extremely large rock fragments under the combined action of full flow energy (both kinetic and potential) and gravitational force (Makkaveev & Chalov, 1986).

Bank erosion returns the alluvial deposits comprising flood plains and terraces to sediment transport. Hence, flood plains also represent one of the expressions of sediment transport (Makkaveev, 1955). An increase of flood plain width and channel stability is connected with an increase in the amount of sediment preserved in flood plain deposits. At the same time, eroded fragments of unflooded terraces may become an additional source of sediment supply to the river channels. Intensive erosion of sandy terraces leads to the formation of a series of shallow riffles downstream, represented by either chaotic combinations of numerous various-sized dunes or the most complex (in terms of their morphology and deformation regime) braided reaches. Examples are the Severnaya Dvina River below Tolokonnaya Mountain (Chalov *et al.*, 2000) and the Zeya River below Belaya Mountain (Vinogradov, *et al.*, 2003).

CONCLUSIONS

River channel morphology, rates and tendencies of their deformations are determined largely by sediment yield and relative contribution of its suspended (basin-derived) and bedload (channel-originated) components. The most important characteristic appears to be size composition of channel-forming sediment and flow transportation capacity. Domination of different forms of sediment transport for material of various sizes results in morphological and dynamical differentiation of river channels with similar total sediment yield values under different natural conditions. On the other hand, similar channel forms can be produced under different conditions of sediment yield and transport.

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