Fluvial response to climate change: a case study of northern Russian rivers

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Abstract The cold regions of North Eurasia include very sensitive fluvial systems. Rapid changes in climate are reported for these areas. The aim of this study is to propose a framework for fast climate-driven predictions of fluvial systems, and to apply it for rivers of the northern part of the East European Plain and West Siberian Plain. The general approach consists of integrating outputs from climate models into a hydrological model, and then driving a catchment and morphodynamic model using output from the hydrological model. Modelled by AOGCMs, future climate shifts are the drivers of significant changes in surface flow. Predictions of an up to 25% decrease in annual runoff by the middle of the 21st century enables us to forecast changes in sediment migration rates, stream energy and water-channel boundary interactions, changes in channel morphology and channel patterns shifts using corresponding physically-based equations. Whereas high dimensional models are still computationally too expensive for long-term morphological predictions, simple 1-D equations enable us to make assessments of channel system response. We tested a suit of 1-D models to estimate fluvial response to climate change for the middle of the 21st century of medium and large rivers draining the north of Russia. Comparison with regional predictions for other territories is the special task of the study.

Key words climate change; runoff calculations; fluvial systems; sediment load; channel patterns

INTRODUCTION

A fluvial system represents a combination of processes and forms expanding from hillslopes through rills and gullies to rivers. The indirect influence of climatic factors on fluvial systems appears in change of those factors that control fluvial processes and forms. A review of the literature provides insight into previous understanding of fluvial system evolution. The essence of this stability and its failure is given by Velikanov (1958): "*Interaction and mutual control between fluid flow and solid boundaries leads to certain combinations of stream hydraulic conditions and channel morphology*". The controlling factors described by Leopold *et al*. (1964) were width, depth, velocity, slope, discharge, sediment size, suspended sediment concentration, and channel roughness.

Climate change effects on fluvial processes should be considered through conceptual process-based modelling which aims to join together the spatial structuring, variability and scaling effects. Though a variety of studies have been done in fluvial geomorphology, the modelling capability in this area is still highly imperfect, even in terms of the general vision and understanding of the driving forces and links. Consideration of the joint impacts of climate change through controlling factors for the entire system emphasizes the nonlinear nature of fluvial response, and the possibly severe and synergistic effects that come from the combined direct effects of climate change. River runoff changes and other factors affect fluvial system evolution, such as rates of evapotranspiration, precipitation characteristics, plant distributions, sea level, glacier and permafrost melting, and human activities.

The modelling approach considered in this paper comprises use of climate model output which drives a hydrological model, and output of the hydrological model drives a catchment and morphodynamic model. The problem is tackled by applying a suite of 1-D models to the fluvial response to climate change. High-resolution models are computationally too expensive for long-term morphological predictions over extended river reaches. The main aim of this study is to develop a framework for fluvial processes, including climate-driven modelling and application of a 1-D approach in the rivers of the East European and West Siberian Plains. Based on a literature review and recent studies, we describe fluvial system response using governing 1-D equations. These equations were previously validated for the rivers of the past environments and proved their high adequacy. Validations of the models for the present or past time scales (testing against "real" data) is necessary, but beyond the scope of this short paper.

STUDY AREA

The study focuses on the rivers of two northern areas of the Russian Federation: the East European and West Siberian plains. The East European plain together with the Northern European Plain constitutes the European Plain. It is the largest mountain-free part of the European landscape. The East European plain rivers (Volga, Dnepr, Don, Neva, Severnaya Dvina and Pechora and their tributaries) are very important fluvial systems from both economic and ecological perspectives. The study area includes the watersheds of the Caspian Sea – Volga River and its tributaries (Kama, Oka, Vyatka, Sura), Azov and Black Seas – Don and Dnepr, White and Barents Seas – Onega, Mezen, Severnaya Dvina and Pechora. The West Siberian Plain is another extensive flat territory of Russia, which is represented mostly by the Ob watershed and its confluents (Irtysh, Tom, Chulym, Ket, Tavda, Tura, Sosva, Biya) and also some small watersheds of the northern rivers: Nadym, Pur and Taz. The flow of some rivers (Biya, Katun, Tom) is formed in mountain conditions, but most of the territory is a flatland. All the studied rivers are located in the northern part of Eurasia and flow into the Arctic seas, and are thus considered as cold region.

RUNOFF CALCULATIONS

Climate-driven changes in river flow will likely cause changes in the fluvial system. Its estimation is crucial for predicting sedimentation rates and morphology changes. Results from global atmospheric circulation models and their interaction with the ocean (atmospheric–ocean general circulation models, AOGCM) were used to assess surface runoff changes. The climate-driven hydrological model is based on the water balance equation:

$$\overline{Y} = \overline{P} - \overline{E} \tag{1}$$

where $\overline{Y}$, $\overline{P}$ and $\overline{E}$ are average values (mm) of river flow, precipitation and evapotranspiration, respectively. Evapotranspiration $\overline{E}$ is often evaluated using the equation proposed by V. S. Mezentsev (Kislov *et al*., 2008):

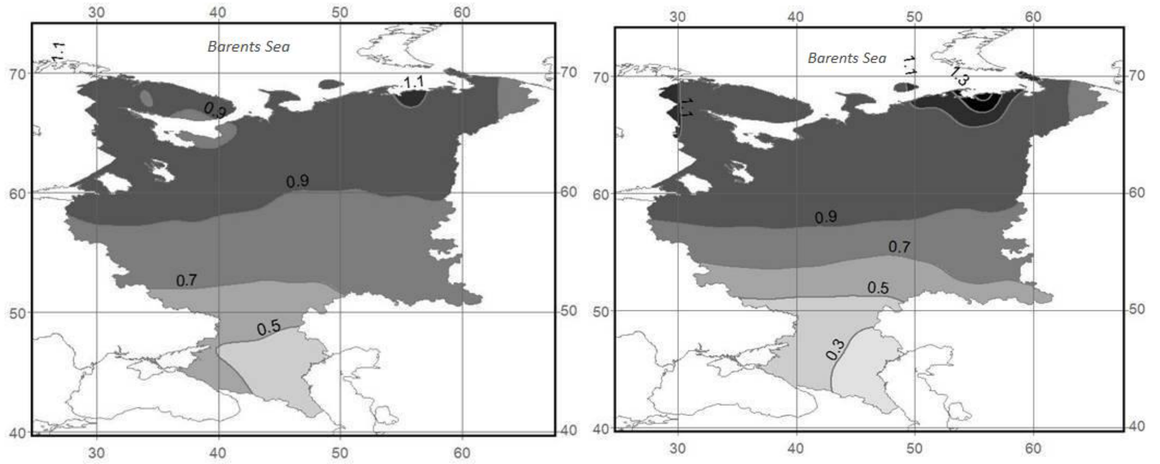
$$\frac{E}{E_o} = \left\{1 + \left(\frac{E_o}{P}\right)^n\right\}^{-1/n} \tag{2}$$

where E_o is evapotranspiration, and n is the correction factor, which is usually applied as $n = 3.8$ for East European Plain, and $n = 4$ is appropriate for Siberia (West Siberian Plain) (Kislov *et al*., 2008). Evapotranspiration E_o depends on the sum of positive monthly average temperatures T_o:

$$E_o = a\,T_o + b \tag{3}$$

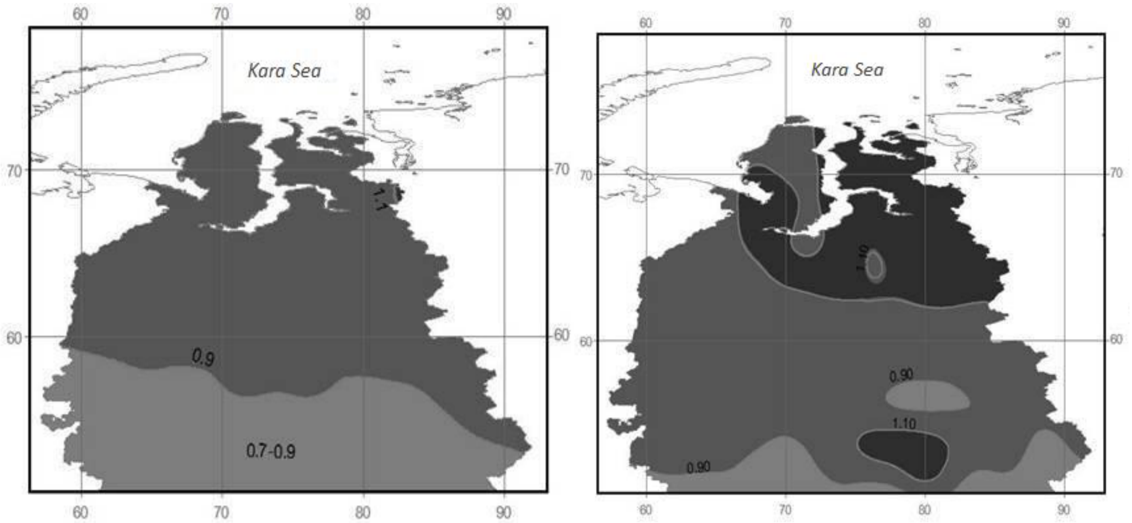
where a and b are regional empirical coefficients (Kislov *et al.*, 2008). The approach was previously validated for the period of 1961–1989. Using the data for 21st century from the AOGCMs, runoff changes were evaluated for the middle of the 21st century (2050) and the end of the 21st century (2100). Calculated runoff values for the baseline (1961–1989 years) and predicted periods were compared using runoff changes rate: $K_y = Y_{predicted}/Y_{base}$.

The results (Figs 1 and 2) indicate a general reduction in average annual runoff in the southern part of the territory (south of around 55°N), and an increase in the northern part. Thus water flow will increase in high latitudes in the watersheds of the Severnaya Dvina, Mezen, Pechora, Nadym, Pur and Taz rivers. In the central part of the East European Plain, the model suggests a 10–15% decrease in annual runoff, while for southern watersheds (Dnepr, Don) the decrease attains 25% by the middle of the 21st century. These changes also occur across a large part of the West Siberian Plain. Water flow was found to reduce across central and southern parts of the Plain. This study



(a) (b)

Fig. 1 Relative runoff changes on the territory of East-European Plain by the middle (a) and the end (b) of the 21st century



(a) (b)

Fig. 2 Relative runoff changes on the territory of West-Siberian Plain by the middle (a) and the end (b) of the 21st century.

attempts to quantify several sources of uncertainty, and to show that the effects of model uncertainty on the estimated change in runoff were generally small relative to the differences between scenarios and the assumed change in global temperature by 2050.

CATCHMENT PROCESSES AND SEDIMENT DELIVERY

Evidently, future catchment processes are strongly affected by changes in precipitation. Further changes in land-use management practices could represent the reaction of farmers to climate shifts in implementing different crops and land use. There are a few methods which can be used for the estimation of the rate of main driving forces – surface erosion. The well-known empirical Universal Soil Loss equation (Wischmeier *et al.*, 1971) is:

$$A = R_f K_f L_f S_f C_f P_f \tag{4}$$

where A is soil loss in tons acre^{-1} year^{-1}, R_f is rainfall factor, K_f is soil erodibility factor, L_f is slope-length factor, S_f is slope-steepness factor, C_f is cropping management factor, and P_f is the erosion control practice factor. A similar approach resulted in the Modified Universal Soil Loss Equation, MUSLE (Wischmeier *et al*., 1971):

$$E = R_e K_e L_s S_g C_u P_c \tag{5}$$

where E is the mean of a couple of years of annual eroded soil mass from a unit of surface area, R_e is annual mean erosion from rainfall and runoff, K_e is coefficient of erodibility, L_s is nondimensional coefficient of slope length, S_g – nondimensional coefficient of slope gradient, C_u is nondimensional coefficient of type of cropping and land use, and P_c is a non-dimensional coefficient of erosion control practice. In the context of fluvial processes response to climatic change, all these parameters are regarded as changing variables, but R_e contributes most to changes in erosion (Kanatieva *et al*., 2010). The main task for this work is to integrate the results of erosion modelling to the in-channel sediment data, which in the case of linear relationships, could be done using rate of sediment delivery and unit area discharge relationships (Dedkov & Mozjerin, 1984). Estimation of the amount of mobilized sediment that actually reaches the stream network depends on the delivery ratio *DR*, which determines the fraction of eroded sediment that is transferred from one location to another (i.e. field to field, or field to stream). It depends on the distance between the sediment source and channel, slope and roughness of the flow path, and availability of surface runoff. The latter provides background for climate-induced change assessments. The *DR* could be empirically estimated (Dickinson *et al*., 1986) as:

$$DR = \lambda\left(\frac{H_c I^{1/2}}{n_r L}\right)^{\beta} \tag{6}$$

in which H_c is a hydrological coefficient expressing the ability of a certain area to generate surface runoff, I is the slope, L is distance between sediment source and channel, n_r is a roughness coefficient that depends on the type of land use, and λ and β are empirical parameters, being 9.53 and 0.79, respectively.

Where the data of sediment delivery is absent, the most useful method is connected with relationships with water discharge Q. The average relationship between river discharge and suspended sediment concentration is described by a sediment rating curve, which can be found through empirical statistical relationships (Syvitski *et al*., 2000). The degree of $Q_s = f(Q)$ relationship is about 2–3. This leads to a 5% decrease in suspended sediment transport for the northern rivers (Severnaya Dvina, Onega, Mezen), and a 15–30% decrease for the central and southern rivers (Kama, Don, Irtysh).

IN-CHANNEL PROCESSES

In-channel changes appear in sediment load, bed-load particle stability, dune length and height, channel patterns (straight, meandering or braided), riverbank erosion, riverbed erosion and channel slope. The scale and intensity of these processes are influenced by a number of factors controlled by climate through water runoff changes.

Total sediment transport

The amount of total sediment load can be calculated through the Makkaveev (1955) equation:

$$R = A\, I^{n1} Q^{m} \tag{7}$$

where A is the erosive coefficient estimated for the gauged river basins (Makkaveev, 1955; Chalov & Shtankova, 2000), I is slope, Q is discharge, and n_1 and m are regional coefficients which depend on stream hydraulics (m usually equals 2 for plain rivers and 3 for mountain rivers; n_1 is about 1). Through the regional combination of m and A, the predictions were made. The results are shown in Table 1. For rivers where runoff decrease is expected, the total sediment load shows the same decreasing trend: for the Don by 40%, Kama 17% and Vyatka 22%. The Biya River demonstrates stable sediment transport rates ($R_{pr}/Ro = 1$) in the absence of significant flow changes.

Table 1 Total sediment transport changes for periods of the middle and the end of the 21st century.

River	Q_{pr}/Qo 2050	Q_{pr}/Qo 2100	A	m	R_{pr}/Ro 2050	R_{pr}/Ro 2100
Don	0.69	0.53	2.5	1.44	0.59	0.40
Kama	0.89	0.87	0.31	1.63	0.83	0.80
Vyatka	0.86	0.92	1.13	1.62	0.78	0.87
Biya	1.00	1.00		3.45	1.00	1.00

Ro – total load for the period (1961–1989), R_{pr} – total load for prediction period (2050 and 2100).

Bed load

Climate instability transforms the characteristics of bed load. A distinction can be drawn between transport in gravel-bed and sand-bed rivers. For alluvial sand-bed rivers, the best results to estimate sediment load are indicated through application of a structural approach (Alexeevskiy, 1998). The latter usually represents the moving of grains in groups such as ripples and thus forming the hierarchy of bed forms (riffles). Usually five types of ripples from small to large (Alekseevskiy, 1998) can be identified on rivers of different size. Their full height and shifting velocity depend on stream order N, in accordance with the equation:

$$(h_g)_i = \alpha N_{sh}^{\gamma} \tag{8}$$

where α and γ are the empirical coefficients, and i is bed form type. Stream order is calculated using (Scheidegger, 1966) N_{sh} = $\log_2 S + 1$, where S is the number of streams, the length of which is less than 10 km, located in the upstream of the drainage basin. Climate-induced changes of Q_o causes stream order transformation because of the logarithmic relationship between stream order and discharge: $N_{sh} = f(Q_o)$. Bed forms move along the stream with a certain velocity, C_g. The analysis of the bed form order of different type demonstrates the seasonal character of the relation between C_{gi} and stream order N_{sh}. During floods, bed forms move with the velocity:

$$C_g = \alpha_1 V_f N_{sh}^{\gamma_1} \tag{9}$$

where V_f is annual stream velocity during flood, and α_1 and γ_1 are empirical coefficients determined for each type of bed form (Alexeevsky, 1998).

The approach described was applied for the Chulim River (Ob basin), Severnaya Dvina and Oka rivers to estimate future changes in sediment transport. The relative changes in bed load were calculated using the Velikanov (1958) formula:

$$G_y = k\,\rho_n\,b_a \sum(h_i - C_i) \tag{10}$$

By the middle of the 21st century, due to flow changes, bed load will decrease by 1% for the Chulim and Severnaya Dvina, and 3% for the Oka. By the end of the century it seems to decrease by 2% for the Chulim and Severnaya Dvina, and by 5% for the Oka, in comparison with the present time.

Channel aggradation/degradation

Sediment delivery, sediment transport and transport capacity alterations enhance either sediment deposition or bed incision. Further changes in longitudinal profile occur indicating channel aggradation or degradation. Combination of the stream order–water discharge and the mass storage equations enables estimation of bed elevation change Δz (Alekseevskiy *et al.*, 2006):

$$\Delta z(x) = -\frac{1}{B\sigma_o}\frac{\Delta t}{\Delta x}\Delta[\alpha_2 e^{\beta_2 N} + \alpha_3 e^{\beta 3N} \tag{11}$$

where B is channel width, σ_o is density of bed deposits (kg/m^3), α_2, α_3 and β_2, β_3 are regional coefficients, and N is stream order. Concerning $N = f(Q_o)$, the correlation between Δz and Q_o changes is determined and can serve as the basis of bed elevation shift analysis on the extensive river sections under different scenario of Q_o changes. For the rivers of the Severnaya Dvina watershed, a runoff decrease of 5% will lead to 0.1–0.3 m of channel aggradation. In the Don catchment a significant runoff decrease will induce 0.5–1.0 m of channel aggradation.

Riverbank migration

The instability of water regime transform intensity and probability of planform changes are important. The latter phenomenon is estimated through riverbank migration rates. Berkovitch (1992) claimed that riverbank erosion C_b (m year^{-1}) could be assessed as a function of annual water discharge Q:

$$C_b = K_1(Q^2 I/dH_b) \tag{12}$$

where I is channel slope; d is average sediment size, mm; H_b is the height of the washed bank relative to low water level, m; and K_1 is an empirical coefficient (m^3/s)$^{-1}$. Riverbank migration rates were estimated for the studied rivers. The results are shown in Table 2. Increase (5–15%) in the intensity of bank migration is expected for northern rivers (such as Sev. Dvina, Peshora, Neva, Taz, Pur, Nadym). Conversely bank migration will become slower on rivers of central southern areas, and reduce by from 10 to 50%.

Table 2 Rates of riverbank migration changes according to equation (12) and (13) by the middle and the end of 21st century.

River	C_{b2050}/C_{b1990}	C_{b2100}/C_{b1990}	BC_{x2050}/BC_{x1990}	BC_{x2100}/BC_{x1990}	River	C_{b2050}/C_{b1990}	C_{b2100}/C_{b1990}	BC_{x2050}/BC_{x1990}	BC_{x2100}/BC_{x1990}
Volga	0.86	0.81	0.94	0.92	Nadym	0.98	1.24	0.99	1.06
Kama	0.85	0.83	0.94	0.93	Pur	0.97	1.24	0.99	1.06
Dnepr	0.67	0.45	0.86	0.73	Taz	1.07	1.33	1.02	1.08
Don	0.59	0.40	0.82	0.71	Biya Katun	1.00	1.00	1.00	1.00
Sev.Dvina	1.04	1.10	0.96	1.01	Tobol	0.67	0.65	0.90	0.89
Pechora	1.04	1.12	0.99	1.04	Ishim	0.64	0.61	0.89	0.88
Chulym	0.95	0.84	0.99	0.95	Irtysh	0.82	0.83	0.95	0.95
Ket	0.99	0.94	1.00	0.98	Ob	0.90	0.91	0.97	0.98

C_b – bank migration rate according to equation (12) using the approach of Berkovich;
BC_x – bank migration according to equation (13), Bartley's approach.

Riverbank erosion (BC_x) can be estimated from bankfull discharge, which is determined by a process of basin-wide regionalization of the 1.58-year recurrence interval flow on the annual maximum time series (Bartley *et al.*, 2004). Because woody riparian vegetation is known to reduce bank erosion rates, a simple ramp function is introduced to exclude the proportion of bank with riparian vegetation (PR):

$$BC_x = 18(1 - PR)\,(Q_{1.58})^{0.6}\,L_x \tag{13}$$

Calculations for rivers of the East European and West Siberian Plain indicate (Table 2) that future changes mainly represent decreases of riverbank erosion. The most considerable changes are predicted for the Don and Dnepr rivers due to significant flow decrease (more than 25% by the end of the century). On most other rivers, these rates will be between 1% and 8%. There are also a few rivers with stable rates of bank erosion: the Ket, Taz, Biya and Katun. Although the results from both equations gave similar directions of change, the absolute values in a few cases differed by up to 20%. The Berkovitch equation (12) gave large values in bank migration, whereas according to Bartley's approach (13), only slight fluctuations in bank migration are predicted.

Channel patterns

The direction and intensity of bed deformations and bed form dynamics changes can lead to channel pattern shifts. Channel patterns recognition is typically based on the assumption that the transitions between the different channel types are threshold-governed processes. Conventional QI-type models predict the existence of either multi-thread or single-thread meandering and straight channel types using data on discharge Q and channel slope I. The more advanced model by Van den Berg (1995) takes into account the size of the sediment material. QI-type models can be used to predict channel pattern development. A probabilistic approach indicates that QI values (unit stream power) will still be in the transition zone around the critical threshold where two channels types may be expected.

Previous study on channel pattern prediction under future climatic conditions for 16 rivers of Severnaya Dvina and Pechora basin, using Van den Berg's model (Anisimov *et al*., 2008) found potential transformations of the channel types, from single- to multi-thread, at 4 of the 16 selected locations in the next few decades, and at 5 locations by the middle of the 21st century. Hypothetical scenarios of 10%, 15%, 20% and 35% runoff increase were used. The results are confirmed by the recent investigations of several rivers (Alexeevsky *et al*., 2006) where channel pattern changes were studied during 1960–2000 and were found to correspond with climate flow rate shifts. Thus for the small Protva River (Volga basin) the cumulative difference curve depicts a growth of flow rate during 1960–2000: $M = \frac{1}{n}\sum\frac{(K_i - 1)}{c_v} = 0.17$. For the period 1975–2000, M equals 0.75; this is the period of steady flow rate growth. A single braided reach transformed into a multi-thread channel with a few islands. Comparison of predicted and virtual channel pattern changes shows an absence of total transformation from a meandering channel to a braided or straight one.

COMPREHENSIVE ASSESSMENT OF FLUVIAL SYSTEM RESPONSE

The results obtained show the synergistic effects of the fluvial system's response to climatically-driven water discharge changes. Simulations for a given basin demonstrate that increase of surface runoff enhances surface erosion and sediment delivery, which is followed by sediment transport alterations. The latter contribute to distinctions in mass storage over channel reaches, caused in longitudinal profile evolution. Due to channel degradation decrease, a change from meandering and braided channel patterns to straight channels is usually observed. Otherwise, in the aggraded channel reaches, braided channels occur (Alekseevskiy, 1998). The meandering-braided threshold shows that climate decrease of Q_{max} (I = const) causes the transformation from the braided to meandering channels

The results of this investigation show the complex reaction of the fluvial system to climate changes. A number of regionally-based predictions of climate-driven changes of fluvial processes have been made recently (Asselman *et al*., 2003; Gomez *et al*., 2009; Verhaar *et al*., 2010) to describe exact phenomena of the fluvial system (Table 3). It can be seen that in different regions of the world, flow and channel process fluctuations may be of different amount, and even of different trend. Asselman *et al*. (2003) suggested that sediment load transport of the River Rhine by 2100 could decrease by 13%. Morphological simulations for the 21st century of three tributaries of the Saint-Lawrence River (Verhaar *et al*., 2010) predicted an overall increase in volumes of bed material that will reach the Saint-Lawrence River, as well as an effect on the longitudinal profile extending up to 10 km from the confluence with the Saint-Lawrence River. Decrease of the mean flow in the Waipaoa River (Gomez *et al*., 2009) by an average of 13% in the 2030s, and 18% in the 2080s, will lead in the 2030s to the decline of bed load to 6.3±16.1 kt year^{-1}, further rising to 9.4±20.1 kt year^{-1} in the 2080s (after aggradation reduces the amount of accommodation space and modifies the long profile of the simulated river). The results depend mostly on the type of climate prediction used the given region, as this underlies the fluvial process model. Choosing the certain fluvial response approach also plays an important role in the resulting values, but the contribution is less significant.

Table 3 Overview of regional-scaled predictions of fluvial processes in 21st century.

Region	Time scale	Climate-driven water discharge scenario	Suspended load	Bed load	Bed elevation	Riverbank erosion	Authors
River Rhine	2100	Annual average discharge increase	Decrease by 13%				Asselman, *et al.*, 2003
Tributaries of the Saint-Lawrence	2010–2099	Reference base level 1 cm decrease		increase in average bed material delivery	Degradation		Verhaar *et al.*, 2010
Waipaoa River, New Zealand,	2010–2030s	Q_{mean} decrease by 13%	Either decline by 1 Mt/year or increase by 1.9±1.1 Mt/year.	Decrease 6.3±16.1 Kt/year	Aggrade by 0.31 m		Gomez *et al.*, 2009
	2010–2080s	Q_{mean} decrease by 18%		Increase 9.4±20.1 Kt/year	Aggrade by 0.85 m		
Northern Russian rivers (Onega, Pechora, Severnaya Dvina, Pur, Taz)	2050	Q_{mean} decrease by 3–5 %	Decrease	Decrease by 5–7%	10 cm aggradation	Increase by 2–3 %	Present study
	2100	Q_{mean} increase by 3–5 %	Increase	Increase by 5–15%	20 cm aggradation	Increase by 10 %	

CONCLUSION

There is evidence that the global climate is changing. Nowadays, hypotheses of global warming are largely applicable. Climate-induced changes of flow were assessed by a hydrological model which is able to recalculate meteorological data to runoff rates. AOGCMs modelling data and relationships between meteorological and hydrological parameters were used. Through direct and indirect climate influences, river runoff will decrease for the most part of northern Eurasia. The outputs of these assessments underlie the fluvial model, which was proposed for us through the review of existing approaches for predicting fluvial system behaviour. Further consideration of a variety of fluvial processes was made and the prediction models were tested. In a few cases, it was shown (e.g. bank erosion) that various models show similar directions of changes, but the absolute values of the changes could be different. It was found that 1-D assessments give physically-based results which could be easily joined to characterize the singular nature of the fluvial system. The suite of models could successfully cover the whole range of processes which govern the behaviour of the catchment–river system. The special task remaining for future work is to link channel pattern evolution and certain components of the fluvial system.

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