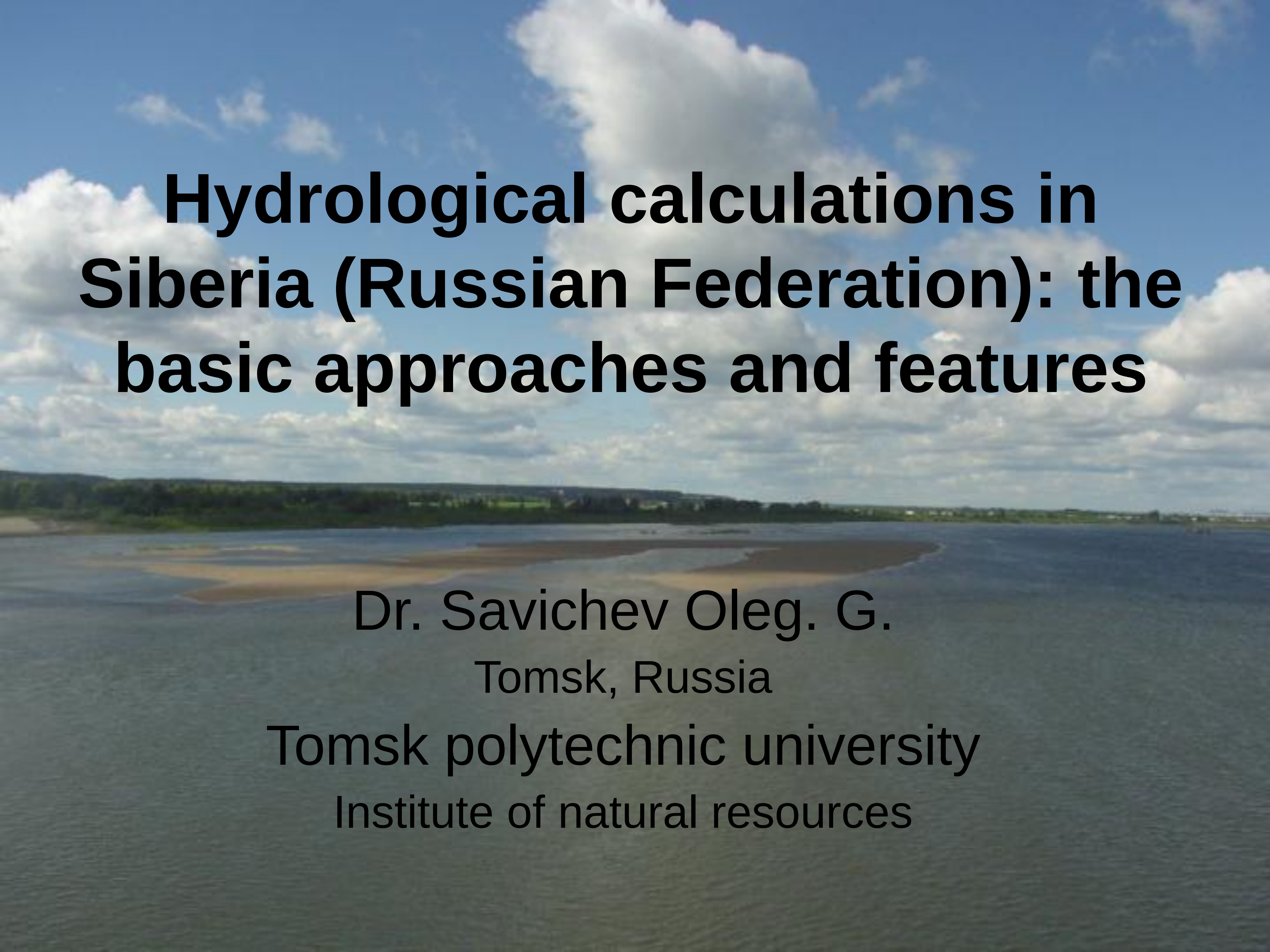

Hydrological calculations in Siberia (Russian Federation): the basic approaches and features

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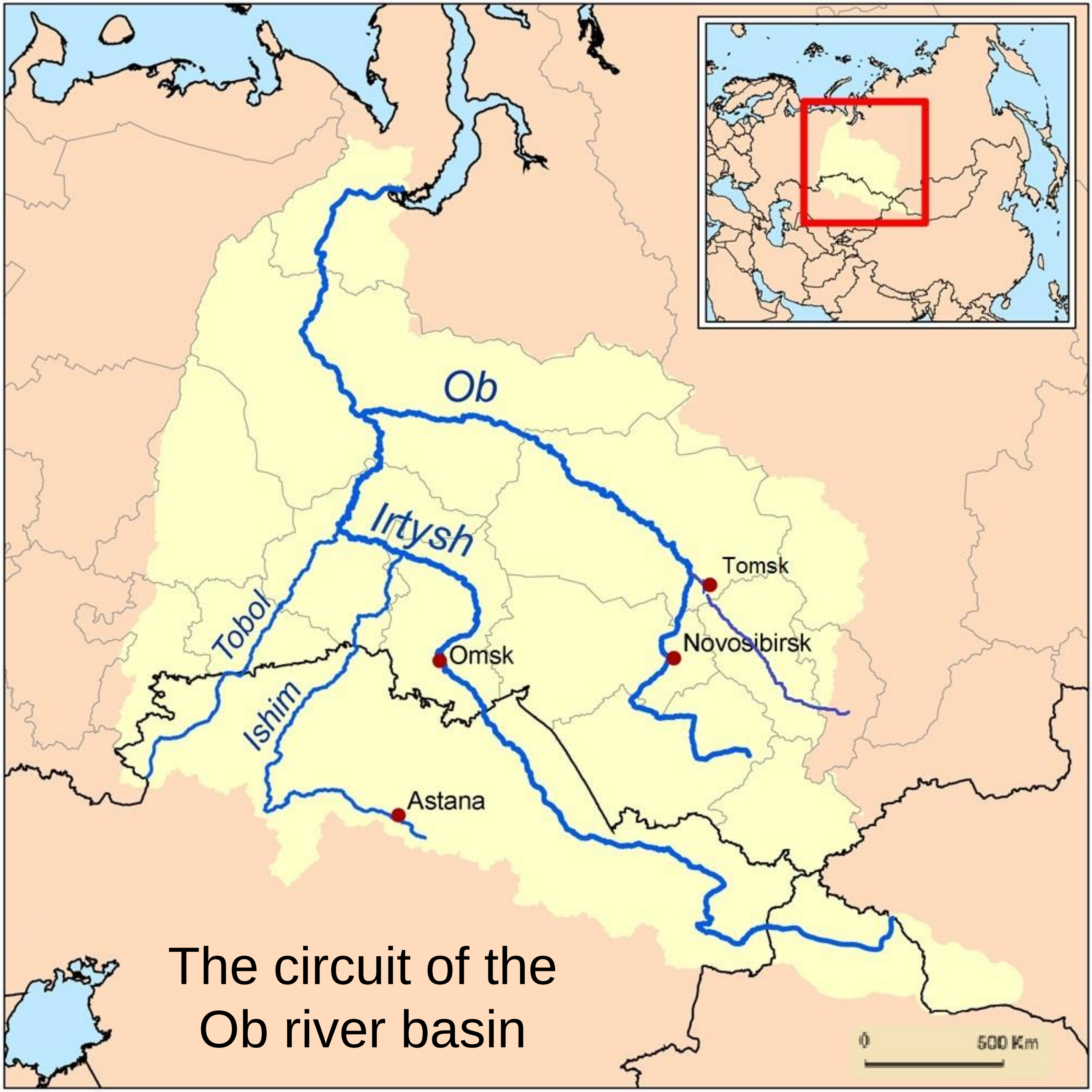
Short information about methods of hydrological calculations in the Russian Federation



The circuit of the Ob river basin

Hydrological Calculations (or an Engineering Hydrology) are the important part of a hydrology which connects together methods of hydrodynamics, mathematical statistics and probability theory, hydrochemistry, geography.

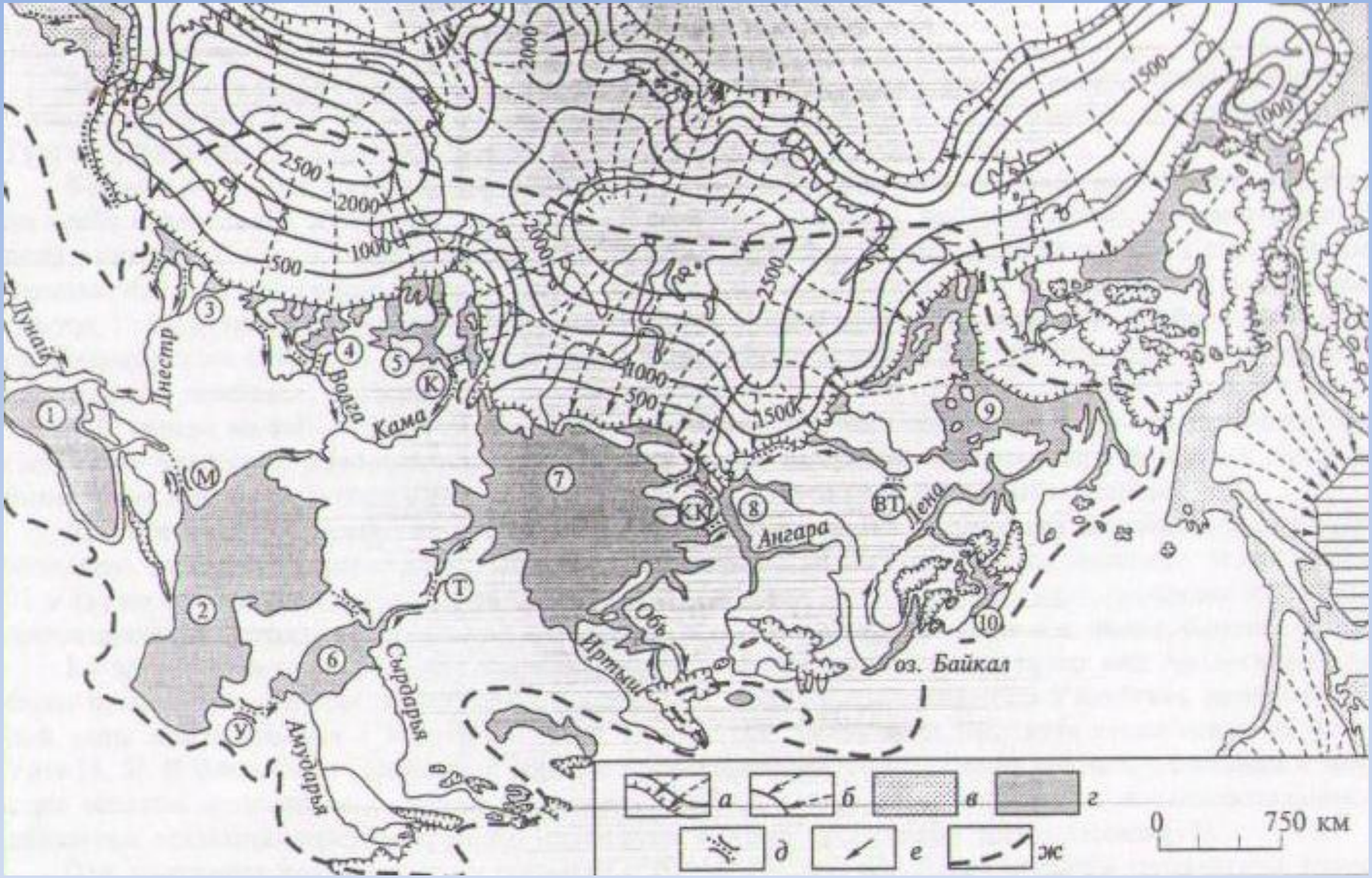
Result of this connection should become simple enough methods in use which allow to receive parameters of a hydrological regime with required accuracy.

Simply – does not mean badly. So – it is necessary to distinguish key parts of researched system and the most informative parameters of their condition.

Frequently it is enough to receive the information on limiting conditions. But it is necessary to understand, that all water systems are always changing.

For example, 20000-15000 years ago in the north and the south of Siberia there were glaciers, 15000-10000 years ago - the density of a river network has strongly increased and “bogs” (*marshes*) have appeared. Within the last 11000-8000 years the bogs are extended (about 1 mm per year upwards) and they gradually “kill” the rivers. But water in bogs is too much abundant and it should flow somewhere. Therefore it form new marsh rivers and lakes.

The sequence of process is shown in photos.



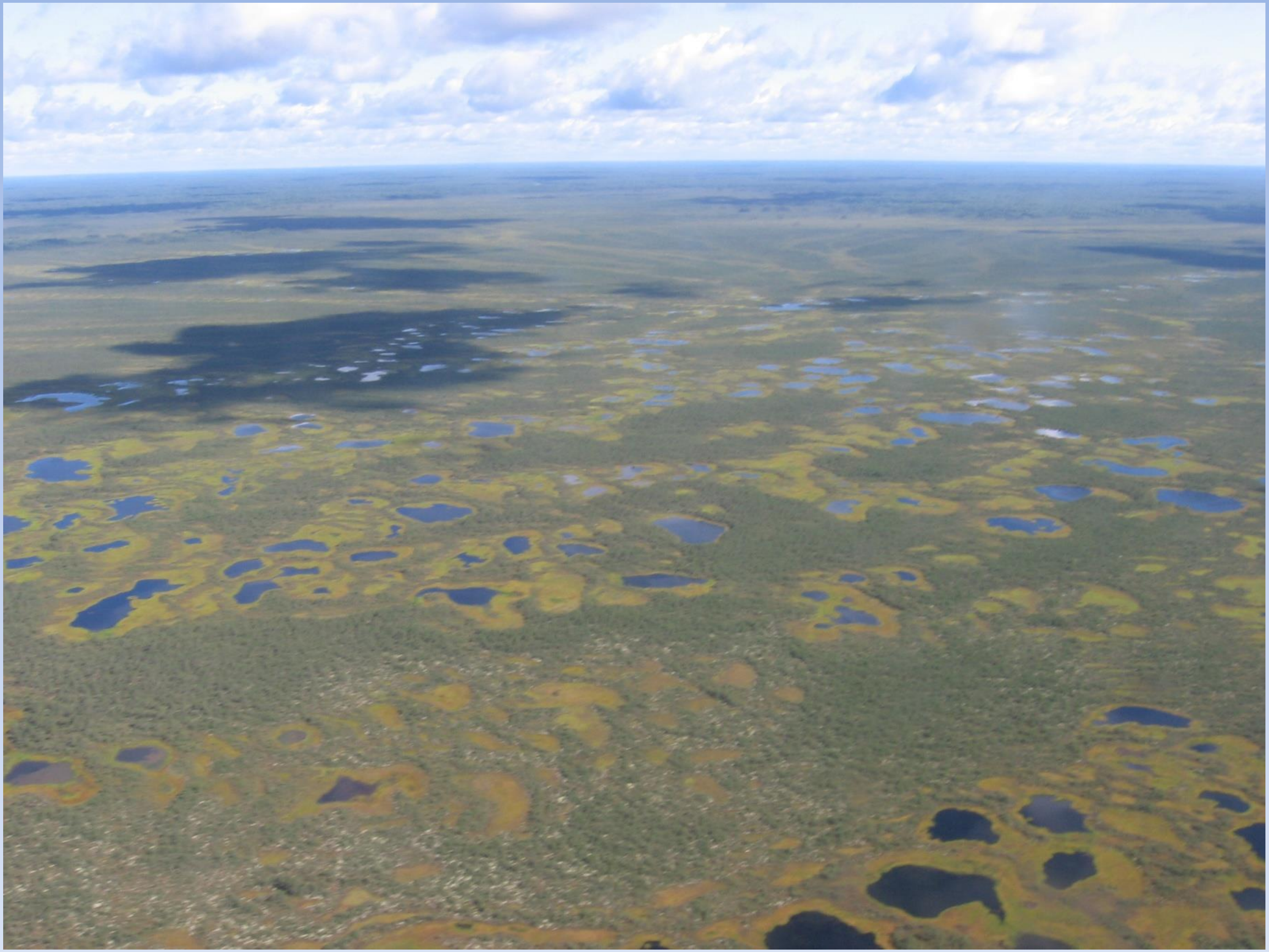
“Stage 1”. The main elements of the transcontinental system Snowmelt Runoff of the last glaciation (by [Volkov, Kazmin, 2007]); а - ice sheets and mountain-cover complexes; б - floating ice shelves; в - drained ice shelves; г - elements of runoff; д- meltwater in streams, ж - border basins drain systems;

6 – Aral lake; 7 – Mansyiskoe (Obskoe) lake

“Stage 2”. The Mouth of Vasugan river in low water

“Stage 3”. Meander of the Vasugan river

"Stage 4". Marshy valley of the Parabel river tributary, with the lack of stream flow in the most part of the year



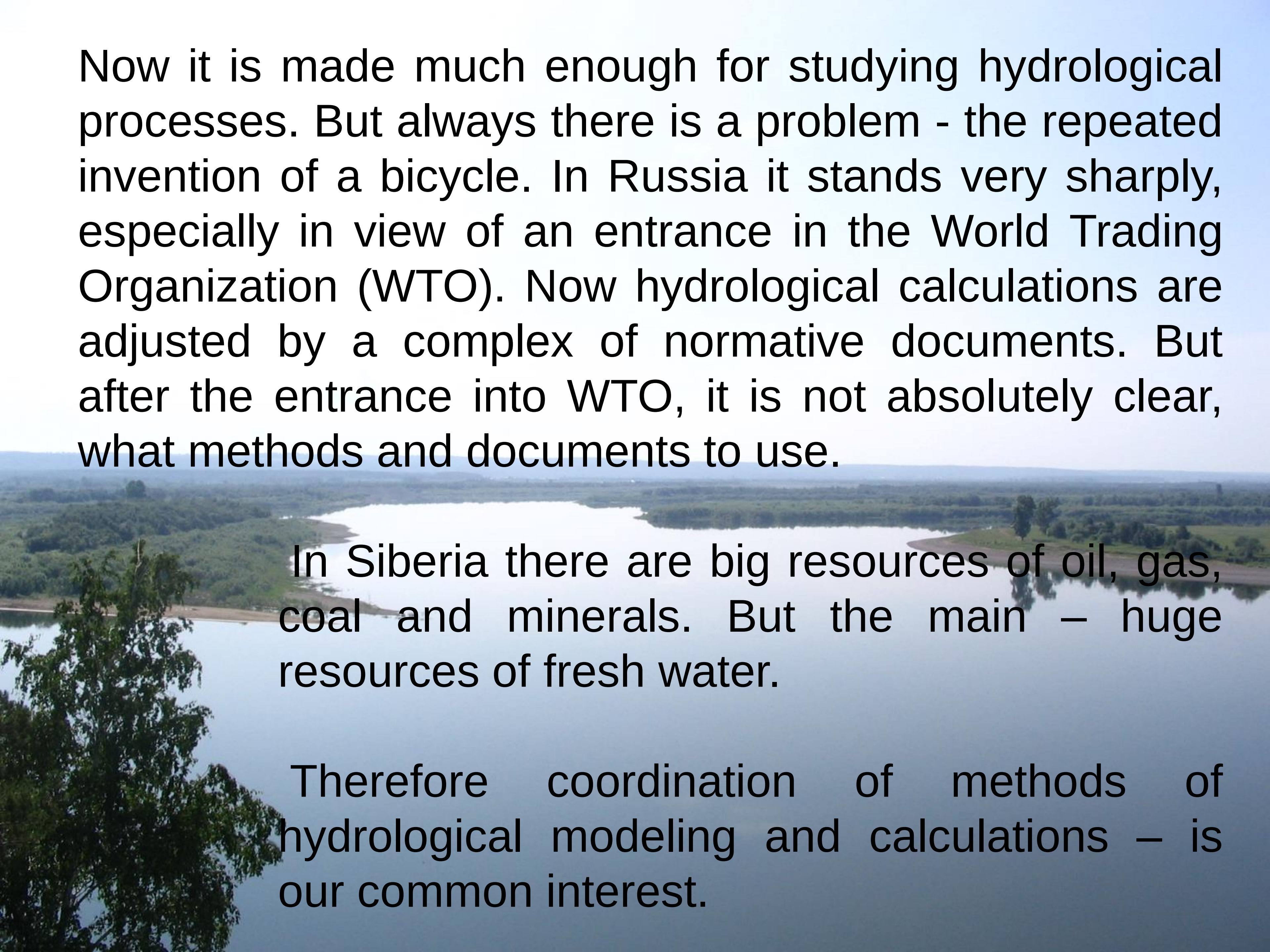
“Stage 5”. Oligotrophic bogs in the Parabel’ river basin (left tributary of the Ob river)

If we want to understand, what will be the water resources (in Siberia or in all Eurasia), then how we should study them? What time scale to choose? What methods of research to use?

It is possible to assume, that it is necessary to create complex models and to understand hydrological processes.

But also it is necessary to carry out the geographical analysis, to cut off superfluous and to apply them to concrete objects.

The second direction is the basic contents of my researches in the part of hydrological calculations and the geographical analysis in hydrochemistry. The basic objects of my research - the rivers and bogs in the Siberia.



Now it is made much enough for studying hydrological processes. But always there is a problem - the repeated invention of a bicycle. In Russia it stands very sharply, especially in view of an entrance in the World Trading Organization (WTO). Now hydrological calculations are adjusted by a complex of normative documents. But after the entrance into WTO, it is not absolutely clear, what methods and documents to use.

In Siberia there are big resources of oil, gas, coal and minerals. But the main – huge resources of fresh water.

Therefore coordination of methods of hydrological modeling and calculations – is our common interest.

1. Briefly about methods used in Russia

Hydrology key concept is Flow – the water movement on the Earth surface, as well as in the thickness of soil and rock in its cycle in the nature. It is distinguished the following kind of flow:

1) surface, overland, subsurface, ground flow;
2) rainfall, snow-melted, ice, ground flow;
3) river and channel flow.

The characteristics of Water Flow are:

1) Volume of Runoff W (m^3/period);
2) Specific Discharge M_W (l/(sec km^2));
3) Depth of Runoff Y_W (mm/period).

A particular case of the runoff volume is the Water Discharge Q (m^3/sec, l/sec).

Each water body is formed by the set of regularly recurring changes such as Hydrological Regime, reflected in long-term, seasonal and daily fluctuations of the following characteristics: 1) level, discharge and volume of runoff (Water Regime), 2) ice (Ice Regime), 3) water temperature (Thermal Regime), 4) sediment transport (Sediment Regime) 5) change of the channel and floodplain (River Bed Regime and Floodplain Evolution), 6) concentration of dissolved substances (Hydro-Chemical Regime). Water Regime is a key element of the Hydrological Regime.

Some special terms: ice dam – congestion of ice floes (ice jam – congestion of frazil ice) in a channel during an ice drift which causes compression of water section and rise of a water level.

1.1. Expert analysis of initial hydrological information

The first stage of research is the expert analysis of data.

In Russia the definition of the calculated hydrological characteristics should be based on the data of hydro-meteorological observations. They are realized by Roshydromet and other survey and design organizations which have special licenses. In the absence of observational data it is necessary to carry out hydro-meteorological researches.

Data of hydrometric observations are tested for authenticity. Unreliable data are excluded.

In the Russia there are often used Geographical-Hydrological Method.

This method involves:

determination of qualitative and quantitative correlation between landscape elements and flow;

detection of geographical regularities in the hydrological regime, and relations between hydrological characteristics;

extrapolation of these dependencies in unexplored objects.

Geographical-Hydrological Method application is based on the analysis of Geo-flow formation conditions and selection of river analogues.

The river-analogue is the river with a similar hydrological regime and conditions of flow formation.

In the selection of analogues we should consider the following conditions:

1) homogeneity of the Water Regime;
2) geographic proximity;
3) homogeneity of runoff formation conditions;
4) average height of the catchments should not differ significantly,
5) absence of factors that distort the natural river flow significantly.

1.2. Statistical analysis of initial hydrological information

Proper use of statistical methods in the Hydrological Calculations assumes obligatory analysis of baseline information on accident, homogeneous and accord between empirical and theoretical distribution curves. This analysis includes the following main stages:

1) formulation of null and alternative hypotheses;
2) determination of significance level and critical region;
3) decision.

The hypotheses formulate on the results basis of preliminary information analysis and the existing physical theory of the nature phenomenon.

1.3. Choice of a design scheme

Hydrological Calculations are realized in three cases:
1) the presence of observational data; 2) at their failure; and 3) in their absence.

1.3.1. Basis of the Calculation in the presence of observational data

The method will consist in a choice and use of function of distribution of probabilities. For each type of objects of construction exceeding probabilities are stipulated. Using these values and function of distribution, we find parameters of a flow.

Select of distribution function type is on the basis of the accord analysis of the theoretical and empirical curves.

Empirical exceeding probability P_e (%) of hydrological characteristics is determined by the formula:

$$P_e=100\cdot m/(N+1), \qquad (1.3.1.1)$$

where m – number of members of hydrological characteristic series, ranged in the descending order; N – total volume of the series.

In Russia, as a rule, three-parameter distribution is applied. These are Kritsky-Menkel distribution in any ratio C_s/C_v; Pearson distribution of type III for $C_s/C_v \geq 2$, log-normal distribution for $C_s \geq (3C_v + C_v^3)$. Distribution parameters of Kritsky-Menkel and Pearson Type III are mean Q_{av}; variation coefficient C_v; ratio of asymmetry coefficient to variation coefficient C_s/C_v. The parameters is determined by observation series of hydrological characteristic with using method of the maximum plausibility and method of moments.

At the present time it often is used the method of moments:

$$Q_{cp} = \frac{\sum_{i=1}^{N} Q_i}{N}, \qquad (1.3.1.2)$$

$$C_v^* = \sqrt{\frac{\sum_{i=1}^{N} (k_i - 1)^2}{N - 1}}, \qquad (1.3.1.3)$$

$$C_s^* = \frac{N \cdot \sum_{i=1}^{N} (k_i - 1)^3}{C_v^* \cdot (N - 1) \cdot (N - 2)}, \qquad (1.3.1.4)$$

where $k_i = Q_i / Q_{cp}$; C_v* и C_s* – biased estimates of variation and asymmetry coefficients respectively.

When C_v*<0.6 and C_s*<1.0 variation and asymmetry coefficients are allowed to determine according to the formulas (1.3.1.3; 1.3.1.4) without the introduction of bias amendments. Otherwise, the variation C_v and asymmetry C_s coefficients are determined by formulas:

$$C_v=(a_1+a_2/N)+(a_3+a_4/N)\,C_v^{*}+(a_5+a_6/N)\cdot C_v^{*2}, \quad (1.3.1.5)$$

$$C_s=(b_1+b_2/n)+(b_3+b_4/n)\cdot C_s^{*}+(b_5+b_6/n)\cdot C_s^{*2}, \quad (1.3.1.6)$$

where $a_1, ..., a_6; b_1, \ldots, b_6$ – coefficients determined for Pearson distribution of Type III on special tables in accordance with autocorrelation coefficient $r(1)$. Biased estimate of autocorrelation coefficient $r^*(1)$ is determined by the formula (1.3.1.7), unbiased estimate is done according to the formula (1.3.1.8):

$$r^*(1) = \frac{\sum_{i=2}^{N}\left(Q_i - Q_{cp,1}\right)\cdot\left(Q_{i-1} - Q_{cp,2}\right)}{\sqrt{\sum_{i=2}^{N}\left(Q_i - Q_{cp,1}\right)^2 \cdot \sum_{i=1}^{N-1}\left(Q_i - Q_{cp,2}\right)^2}}, \quad (1.3.1.7)$$

$$r(1) = -0{,}01 + 0{,}98\cdot r^*(1) - 0{,}06\cdot r^*(1)^2 + \\ + \frac{1{,}66 + 6{,}46\cdot r^*(1) + 5{,}69\cdot r^*(1)^2}{N}, \quad (1.3.1.8)$$

where

$$Q_{cp,1} = \frac{\sum_{i=2}^{N} Q_i}{N-1}, \quad Q_{cp,2} = \frac{\sum_{i=1}^{N-1} Q_i}{N-1}$$

The period of observation should be sufficient if the period is representative, and the relative standard error of calculated hydrological characteristic does not exceed 10% for the annual and seasonal runoff and 20% for the maximum and minimum flows.

The standard error of average is determined according to the formulas (1.3.1.9) or (1.3.1.10), the standard error of the variation coefficient for $C_s = 2C_v$ – by the formula (1.3.1.11).

$$\sigma_{Qcp} = \frac{\sigma}{\sqrt{N}} \cdot \sqrt{\frac{1+r}{1-r}}, \ r<0.5. \qquad (1.3.1.9)$$

$$\sigma_{Qcp} = \frac{\sigma}{\sqrt{N}} \cdot \sqrt{\frac{1+\frac{2\cdot r}{N\cdot(1-r)}\cdot\left(N-\frac{1-r^N}{1-r}\right)}{1-\frac{2\cdot r}{N\cdot(N-1)\cdot(1-r)}\cdot\left(N-\frac{1-r^N}{1-r}\right)}}, \ r \geq 0.5. \qquad (1.3.1.10)$$

$$\sigma_{C_v} = \frac{C_v}{N + 4 \cdot C_v^2} \cdot \sqrt{\frac{N \cdot \left(1 + C_v^2\right)}{2}} \cdot \left(1 + \frac{3 \cdot C_v \cdot r^2}{1 + r}\right), \quad (1.3.1.11)$$

where N – volume of observations; r – autocorrelation coefficient.

If the relative standard error is more then the specified limits and the period of observation is not representative, it is necessary to use method of hydrological characteristics determining in failure of observational data.

1.3.2. Basis of the Calculation of insufficient observational data

In case of observational data insufficiency the calculations are realized by the extension of existing short series involving observational data points of analogs (on the same river or others).

Methods of analogue choosing provides:

1) natural zoning analysis of the study area and the choice of possible analogues in the area with similar hydrological and physiographic conditions;

2) calculation of the matrix of paired correlation coefficients;

3) regression analysis for following conditions:

$N' \geq (6\text{-}10)$;

$R \geq R_{\text{кр}}\ (\approx 0.7)$;

$R/\sigma_R \geq A_{\text{кр}}$,

$k/\sigma_k \geq B_{\text{кр}}$,

where N' – number of joint years;

R – coefficient of pair or multiple correlation;

k – coefficient of regression;

σ_k – standard error of regression coefficient;

R_{cr} – critical value of the pair or multiple correlation;

$A_{\text{cr}}(\approx 2)$, $B_{\text{cr}}(\approx 2)$ – critical values of ratios R/σ_R and k/σ_k.

1.3.3. Basis of the Calculation in the absence of observational data

When observational data are not available, the distribution parameters and calculated values are determined by using the following basic methods:

1) water balance;
2) hydrological analogy;
3) averaging in the homogeneous region;
4) construction of contour maps;
5) construction of regional dependencies between flow characteristics and physical and geographical factors of catchment or flow.

1.4. Calculations of Water Flow

1.4.1. Annual runoff

1.4.1.1. Determination in the presence of observational data

The annual runoff is calculated by applying of analytical probability distribution functions. Parameters estimate of analytical distribution curves are: mean, coefficient of variation *Cv,* and ratio of asymmetry coefficient to variation coefficient *Cs/Cv.*

These parameters set on the base of observation series by approximately Maximum plausibility Method and Method of Moments.

1.4.1.2. Determination in insufficiency of observational data

In practice two most used method of flow calculation are applied: 1) by the distribution parameters of the analogue; and 2) by restoring series of annual values.

In the first case, bringing the average to a longer period is calculated as:

$$\overline{Q}_N = \overline{Q}_n + r \cdot \frac{\sigma_n}{\sigma_{n,a}} \left(\overline{Q}_{N,a} - \overline{Q}_{n,a} \right), \qquad (1.4.1.2.1)$$

where Q_n, Q_{na} – means of hydrological characteristics for the study and analogue objects for joint observational period; Q_N, Q_{Na} – normal annual of runoff for *N*-period; σ_n, σ_{na} – standard deviations.

The relative standard error is determined by the formula:

$$\varepsilon_{\overline{Q}_N} = \frac{100 \cdot \sigma_n}{\overline{Q}_N \cdot \sqrt{n} \cdot \sqrt{1 + r^2 \cdot \frac{\sigma_{N,a}^2}{N \cdot \sigma_{n,a}^2} - 1}} \,, \qquad (1.4.1.2.2)$$

The coefficient of variation $C_{v,N}$ is determined by the formula:

$$C_{v,N} = \frac{\sigma_n}{\overline{Q} \cdot \sqrt{1 - r^2 \cdot \left(1 - \frac{\sigma_{n,a}^2}{\sigma_{N,a}^2}\right)}} \,, \qquad (1.4.1.2.3)$$

where $\sigma_{N,a}$ – standard error of river-analogue for the *N*-year period.

In the second case, the annual values are restored by regression dependencies, and then the distribution parameters are calculated.

Calculated values of the asymmetry coefficient Cs and autocorrelation $r(1)$ take for the river analogues.

1.4.1.3. Determination in the absence of observational data

The parameters are determined by: 1) river analogues, and 2) empirical dependencies with use of climate parameters and other factors.

In the first method, the annual runoff for the gravity center of the flat part of the catchment is determined by linear interpolation between the flow contour. The average annual value of mountain river runoff is determined by the district flow dependencies on the average height of the catchment. The coefficient of variation *Cv* is determined by the map of isolines or the regional empirical formulas. In the second method the annual runoff is calculated as the sum of runoff depths for seasonal components during homogeneous periods.

1.4.2. Maximum flow

1.4.2.1. Determination in the presence of observational data

Calculated characteristics of the maximum river flow in snow melt flood and rainfall flood are determined as the annual runoff. Specificity of the calculations is concluded in the way of initial series forming and the use of amendments in some cases.

In impossibility of divide the maximum discharges of snow melt flood and rain flood it is allowed the construction of the common curves of distribution.

1.4.2.2. Determination in insufficiency of observational data

Characteristics of peak flow, as in the case of calculating of annual runoff, are determined on the basis of regression analysis:

1) the parameters of the analogue distribution; and

2) restored a series of values by years.

1.4.2.3. Determination in the absence of observ. data

Calculation formulas of maximum charges is generally divided into three main groups:

1) Reducing formulas reflect a decrease in runoff modulus with increasing catchment area;

2) Limit Intensity formulas connect the maximum flow Q_{max} with the greatest intensity of water run-out from the snow cover or rain for the travel time, and

3) Volumetric formulas characterize the relate of maximum charges with flow volume or flow depth, and with duration of floods.

In fact, in all three cases it is told about the dependence of the form:

$$Q_{\max} = \frac{W}{\tau} , \qquad (1.4.2.3.1)$$

where W – volume of flow for the characteristic time; τ – lag time.

1.4.2.3.1. Snow Melt Flood

The basic design equation (reducing) for determination of the maximum flow of the spring flood $Q_{p\%}$ (m^3/sec) with exceeding probability *P*% has the form:

$$Q_{p\%} = K_0 \cdot h_{p\%} \cdot \mu \cdot \delta \cdot \delta_1 \cdot \delta_2 \cdot A \,/\, (A + A_1)^n, \qquad (1.4.2.3.1.1)$$

where K_0 – parameter characterizing the concerted action in the spring flood; defined as average value on basis of several river analogues data by reverse path from (1.4.2.3.1.1); $h_{p\%}$ – runoff depth of spring, mm, it depends on variation coefficient *Cv*, ratio *Cs*/*Cv* and average depth of runoff h_0; μ – coefficient; δ, δ_1, δ_2 – coefficients considering the influence of lakes, forest and bogs on the maximal water discharge; *A* - catchment area of the river to the current alignment, km^2; A_1 – additional area, km^2; *n* - the exponent of reduction.

Coefficient δ determined by the formula:

$$\delta = 1/(1+C\,A_{oz}), \qquad (1.4.2.3.1.2)$$

where C – empirical coefficient. Average weighted rate of lake area A_{oz} is determined by the formula:

$$A_{oz} = \sum_{i=1}^{n} \left(100 \cdot S_i \cdot A_i / A^2 \right), \qquad (1.4.2.3.1.3)$$

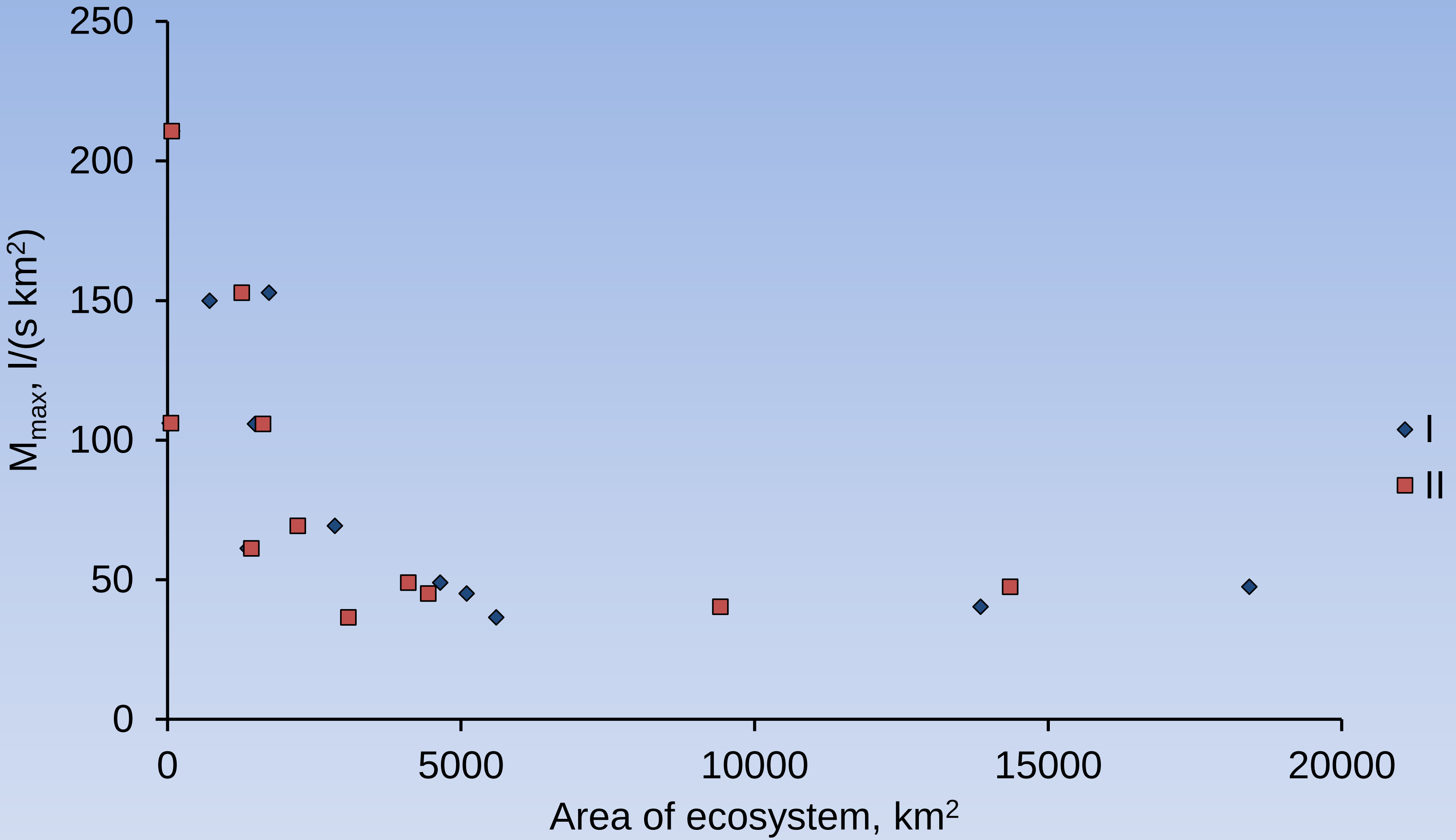
where S_i – mirror lake area, km^2; A_i – catchment area of a lake, km^2. Coefficient δ_1 is determined by the formula:

$$\delta_1 = \frac{\alpha}{(A_{л} + 1)^{n'}}, \qquad (1.4.2.3.1.4)$$

where n' – coefficient of reduction; α – empirical coefficient; $A_{л}$ – relative area of forest. Coefficient δ_2 is determined by the formula:

$$\delta_2 = 1 – \beta \cdot \lg(0.1 \cdot A_{б} + 1), \qquad (1.4.2.3.1.5)$$

where β – empirical coefficient; $A_{б}$ – relative area of swamp in the river basin (%).



The Dependence between a maximum specific discharge (M_{max}) and the area of a forest (I) and swamps (II) for conditions in a taiga zone

In the presence of the meteorological data water discharges of small rivers in the spring floods is allowed to determine by the simplistic formulas, for example:

$$Q_{\max,P} = \frac{\lambda_P \; B \left(a \; R_{T,1\%} + b\right)^{5/3} \; \sqrt{J}}{n_{cp}}, \quad (1.4.2.3.1.6)$$

where $Q_{\max,P}$ – maximum water discharge with exceeding probability P; λ_P – transition coefficient from water flow with exceeding probability 1%. The value of n_{cp} can be defined as average weighted value of surface roughness catchment ecosystem coefficients.

The flooded right flood-plain of the Tom river upstream Tomsk in April in 2010 year

1.4.2.3.2. Rainfall Flood

Calculation formula type for determination of the maximum flow in rain floods is chosen, taking into account the catchment area and the availability of analogue (table 1.4.2.3.2.1).

Table 1.4.2.3.2.1. The application of design formulas to determine the maximum flow of rain flood with given exceeding probability

Formula Type	Formula	River Catchment Area
I	Reducing equation (1.4.2.3.2.1) in the presence of the river analogue	> 200 km^2
II	Reducing equation (1.4.2.3.2.3) in the absence of the river analogue	The same
III	Limit flow intensity formula (1.4.2.3.2.4):in the presence of river analogue, and in the absence of river-analogue	< 200 km^2
IV	Volumetric, genetic and other formulas based on the calculation of runoff from rainfall	> 0 km^2

The design formula of the type I for determination $Q_{p\%}$ in the presence of one or more river analogues has the following form:

$$Q_{p\%} = q_{p\%,a} \cdot \varphi_{м} \cdot (\delta \cdot \delta_2 / \delta_a \cdot \delta_{2a}) \cdot A, \qquad (1.4.2.3.2.1)$$

where $q_{p\%,a}$ – specific max term charge of river analogue with estimated exceeding probability $P_{\%}$ (m^3/sec km^2); $\varphi_{м}$ – coefficient allowing for the reduction of the maximum runoff module of rainfall flood ($q_{1\%}$) with increase of catchment area (A, km^2) or channel lag time (τ_p, minute); it is calculated depending on the value $\eta_{ф}$ (correlation of catchment form coefficients for the study river and the river analogue).

Channel Lag Time τ_p (h) for the hydrologically studied rivers is determined by the formula:

$$\tau_p = \frac{100 \cdot L}{V} = \frac{100 \cdot L}{m_p \cdot I_p^m \cdot Q_{1\%}^{0.25}}, \qquad (1.4.2.3.2.2)$$

where V – the maximum of average speed of channel lag time (m/sec).

The design formula of the type II in the absence of river analogues has the form:

$$Q_{p\%} = q_{200} \cdot (200 / A)^n \cdot \delta \cdot \delta_2 \cdot \delta_3 \cdot \lambda_{p\%} \cdot A, \qquad (1.4.2.3.2.3)$$

where q_{200} – module of max term charge with annual exceedance probability $P = 1\%$, and for the catchment area equal to 200 km^2 with $\delta = \delta 2 = \delta 3 = 1.0$; $\delta 3$ - the transition rate from the maximum term for the mountainous areas.

The design formula of the type III for determination $Q_{p\%}$ in catchment area less 200 km^2 has the form:

$$Q_{p\%} = q'_{1\%} \cdot \varphi \cdot H_{1\%} \cdot \delta \cdot \lambda_{p\%} \cdot A, \qquad (1.4.2.3.2.4)$$

where $q'_{1\%}$ – relative modulus of the maximum term charge with annual exceeding probability P = 1%, representing the ratio:

$$q'_{1\%} = \frac{q_{1\%}}{\varphi \cdot H_{1\%}}, \qquad (1.4.2.3.2.5)$$

This value is determined depending on the hydro-morphological characteristic of the channel Φ_p (1.4.2.3.2.6) and the length of the slope lag time $\tau_{ск}$; φ – collecting runoff coefficient; $H_{1\%}$ – the maximum diurnal depth of precipitation with exceeding probability P=1% (mm); determined from the data of nearest meteorological stations.

$$\Phi_p = \frac{100 \cdot L}{m_p \cdot I_p^m \cdot A^{0.25} \cdot (\varphi \cdot H_{1\%})^{0.25}}, \qquad (1.4.2.3.2.6)$$

In the absence of rivers analogues collecting runoff coefficient φ for lowland rivers is carried out by the formula:

$$\varphi = \frac{C_2}{(A+1)^{n_3}} \cdot \varphi \cdot \left(\frac{I_{ck}}{50}\right)^{n_2}, \qquad (1.4.2.3.2.7)$$

where C_2 – empirical coefficient; φ_0 – collecting coefficient for a conditional catchment with the area 10 km^2.

The flood on the Tom river, Eushta village

1.4.3. Minimum runoff

1.4.3.1. Determination in the presence of observational data

Determination of the minimum water discharges is carried out as annual runoff. In the presence of zero values the empirical exceeding probabilities are determined by the formula:

$$P = \frac{n_1 \cdot P_1}{n_1 + n_2}, \qquad (1.4.3.1.1)$$

where P_1 – the probability of an element in a series of non-zero values; n_1 and n_2 – sample volumes with non-zero and zero values. For the calculations it is used the minimum daily mean, monthly mean or 30-daily discharges (non-calendar) observed in winter and (or) summer-autumn seasons. Monthly minimum means are used if they do not exceed the 30-daily more than 10%.

The Tributary of the Middle Ob river

1.4.3.2. Determination in insufficiency of observational data

The calculations are the same as in the case of runoff rate. Specificity is the preparation of input data.

1.4.3.3. Determination in the absence of observational data

Minimum water discharge in large and medium rivers is determined by interpolation between points of observation, taking into account lateral inflow data and meteorological research. Minimum flow of small rivers $Q_{P\%}$ (m^3/s) is calculated from:

$$Q_{P\%} = b\cdot(A \pm A_1)^{m}\cdot\delta_1\cdot\delta_2\cdot\lambda_{P\%}, \tag{1.4.3.3.1}$$

where A – catchment area, km^2; A_1 – additional catchment area; δ_1 – coefficient of lake influence; δ_2 – coefficient of boggy area influence;

$$\delta_1 = 1 / (1 - c\cdot A_{oz}), \tag{1.4.3.3.2}$$

where c – empirical coefficient; A_{oz} – relative lake catchment;

$$\delta_2 = 1 + \beta^{*}\lg(0{,}1A_{б} + 1), \tag{1.4.3.3.3}$$

where β^{*} – empirical coefficient depending on the type of swamp; A_b – relative area of swamps in the catchment.

In the mountainous areas the minimum flow is determined from the dependence of the modulus of the minimum 30-day runoff from the average height of the catchment.

The Part of the Vasugan bog

1.4.4. Annual distribution of stream flow

1.4.4.1. Determination in the presence of observational data

Determination of calculated calendar annual distribution of flow with duration of observation series n equal to 15 years and more, is carried out by the following methods:

1) compound;
2) real year;
3) average flow distribution for the years of the characteristic scales of water content.

The calculations of the distribution are made by water years, beginning usually from April.

When the observation period n is from 15 to 30 years, three year groups are divided:

- wet years ($P < 33.3\%$);
- average water years ($33.3\% \leq P \leq 66.7\%$);
- dry years ($P > 66.7\%$).

For the duration of observations over 30 years, five groups are divided:

- very wet years ($P \leq 16.7\%$);
- wet years ($16.7\% \leq P \leq 33.3\%$);
- average water years ($33.3\% \leq P \leq 66.7\%$);
- dry years ($66.7 \leq P \leq 83.3\%$);
- very dry years ($P > 83.3\%$).

The simplest method is the Method of Average Flow Distributions for water year. It is based on the calculation of average relative distribution of monthly runoff.

These distributions are typical for each group of characteristic of water flow years. For the calculated distribution it necessary to multiply the monthly relative flow and annual runoff with concrete exceeding probability.

The snow melt flood and winter low water on the Tugoyakovka river (the tributary of the Tom river)

1.4.4.2. Determination in insufficiency of observational data

The short series of the observed runoff values for water year, seasons and months are extended to long period. Further calculations are carried out in the presence of observational data.

1.4.4.3. Determination in the absence of observational data

In the absence of hydrometric observational the annual runoff distribution is determined from the river analogues according to the district schemes and regional dependencies.

1.5. Calculations of the characteristic water levels

1.5.1. Determination in the presence of observational data

Highest water levels are determined by analytical distribution curve of exceeding probability during the period of long-term observations. At the heterogeneity of the highest water levels, empirical distribution curves can be used.

If it is necessary, the calculated levels are transferred up or down the river by one of three ways: 1) use of water discharge curves $Q = f(H)$; 2) use of relation curves between corresponding water levels; and 3) use of longitudinal profile of water surface, taking into account its slope at high water.

Transfer of water levels on the longitudinal profile of water surface is produced within small areas along the lengths (1-3 km). If the highest levels occur in the ice period, then their transfer is realized by the dependence $Q = f(H)$ for the open channel and water flow calculated as:

$$Q'_{P\%} = \frac{Q_{P\%}}{k_Q} \quad , \qquad (1.5.1.1)$$

where $Q_{\mathrm{p}\%}$ – water discharge; k_Q - winter rate.

Determination of maximum levels of lake water is carried out on exceeding probability level distribution curves by the same method, as for the rivers

1.5.2. Determination in insufficiency of observational data

Extrapolation of empirical curve of maximal levels is performed by using analytical functions of the probability distribution of the water discharges.

Calculated water levels of lakes are found by extrapolating the analytical probability curve of water volumes V in the lake. Coordinates of the dependence $V = f(H)$ set during the survey.

1.5.3. Determination in the absence of observational data

Estimated highest water levels is performed by the curve $Q = f(H)$ using the water discharge $Q_{P\%}$. Discharge curves are constructed by the formula:

$$Q = \frac{\omega}{n} \cdot h^{\frac{2}{3}} \cdot I^{\frac{1}{2}}\ , \qquad (1.5.3.1)$$

where ω – area of channel or floodplain cross-section for level H, m^2; n – roughness coefficient, s/m$^{0.33}$; h - average depth of water in the channel or floodplain, m; I – slope of the water surface. Curves $\omega = f(H)$ and $h = f(H)$ are constructed by the results of measurements. The coefficient n and slope I are found from special tables or they are measured.

For determination of the highest water levels in the case of ice dams and jams, preliminary special research is required. The possibility of ice dam formation is defined by the following characteristics: 1) freezing for conditions of moving of the ice cover upstream (characteristic of the rivers that flow from south to north, or go from the mountains to the plain); 2) the fracture of the longitudinal profile with a sharp decrease in the mouth of the slope, the presence of a channel narrowing, a sharp turn, islands; 3) upstream slope of the water surface is more 0.05 ‰; 4) intensive and prolonged (6 days or more) frazil ice drift (flow from deep lakes and reservoirs; 5) large autumn water flow (specific discharge is more 3 l/s·km^2).

The highest level $H_{Z.p\%}$ of ice dam (jam) is calculated by the formula:

$$H_{Z,P\%} = \left(\mu \cdot I_{Q_{Z,P\%}}^{0.3} - 1\right) \cdot h_{Q_{Z,P\%}} + H_{Q_{Z,P\%}}, \quad (1.5.3.2)$$

where μ – coefficient of ice dam or ice jam; $I_{QZP\%}$, $h_{QZP\%}$ and $H_{QZP\%}$ – slope (‰), depth of the river (m) and water level (m) for discharge $Q_{Z,P\%}$ and ice-free channel.

The highest water levels in the movements of the ice are estimated by the curve $Q=f(H)$ through the calculated discharges using the formula:

$$Q`_{\mathrm{P\%}} = Q_{\mathrm{P\%}}\,\eta / K_Q, \quad (1.5.3.3)$$

where η – ratio coefficient of water discharges in the ice drift and at the maximum of spring flood $Q_{\mathrm{P\%}}$; K_Q – coefficient characterizing the change in the hydraulic characteristics of water flow due to ice. These coefficients are determined by the Method of Analogy.

The road flooding is the result of ice dam
in the Tom river in 2004 year

Definition of some terms:

ice dam – congestion of ice floes (ice jam – congestion of frazil ice) in a channel during an ice drift which causes compression of water section and rise of the water level.

To calculate the maximum water levels of flow-through lakes in the humid area the following dependence is used:

$$\overline{\Delta H} = \beta \cdot \sqrt{\frac{A}{\Omega}}, \qquad (1.5.3.4)$$

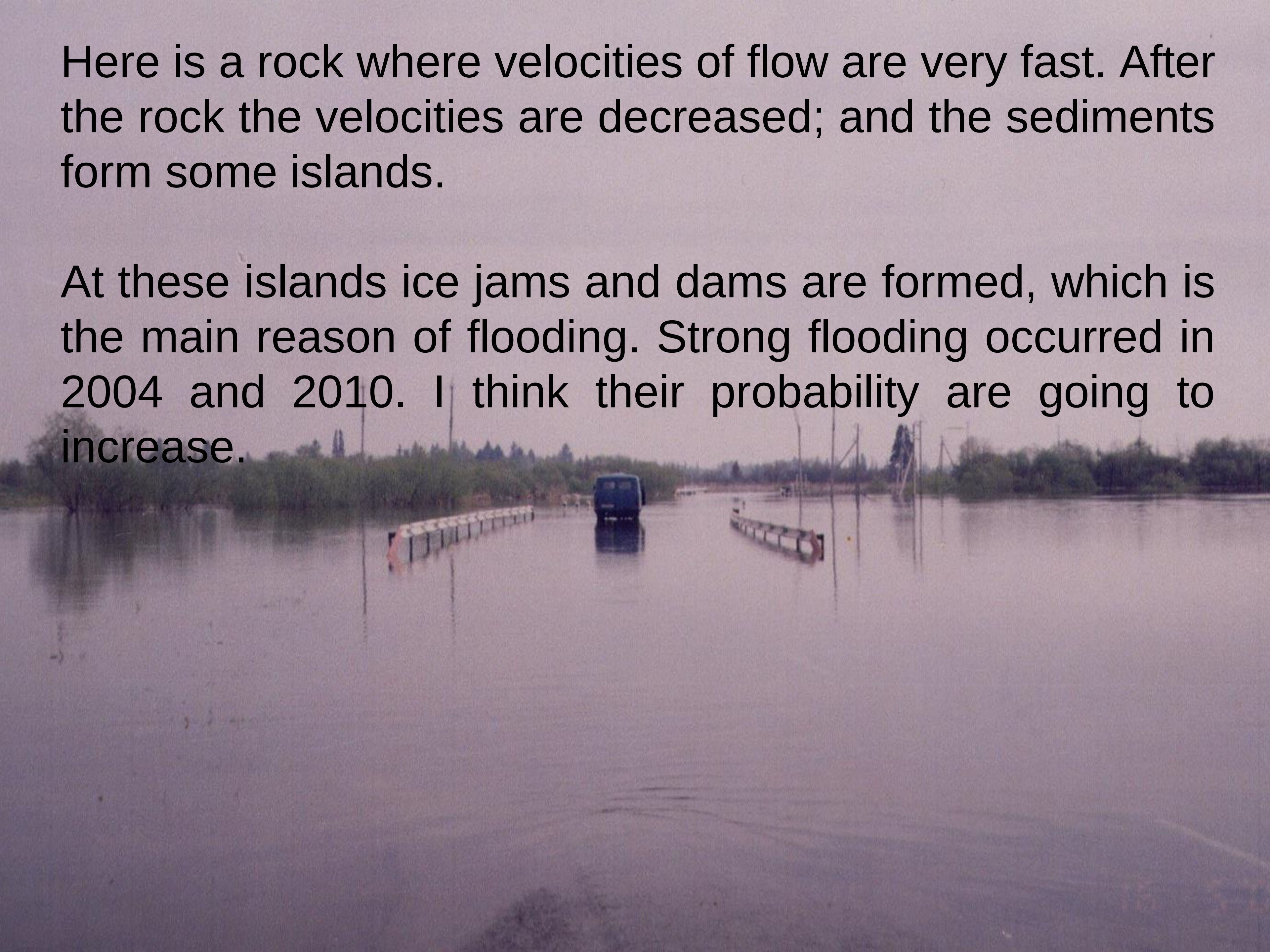
where ΔH – mean spring-summer level rise in the lake above the threshold of runoff (cm); A – catchment area of the lake (km^2); Ω – lake mirror area (km^2); β – coefficient determined from observations at nearby lakes. The dependence (1.5.3.4) is applicable for the ratio $A/\Omega < 250$. The transition from the middle long-term level raising with the calculated probability is carried out on distribution curves with the parameter *Cv* and the ratio *Cs*/*Cv*, determined by analogues. In the values of the calculated levels amendments to the surge $\Delta H_{\text{н}}$ and wind $\Delta H_{\text{в}}$ are introduced.

1.5.4. An example of a the calculations of water level in the ice dam

The Tom' river (tributary of the Ob river) has a direction of current from the south to the north. In the south (mountains of the Shoria and the Kuznetsky Alatau) the most part of water flow is formed. It is the reason of natural ice jams and flooding at Tomsk.

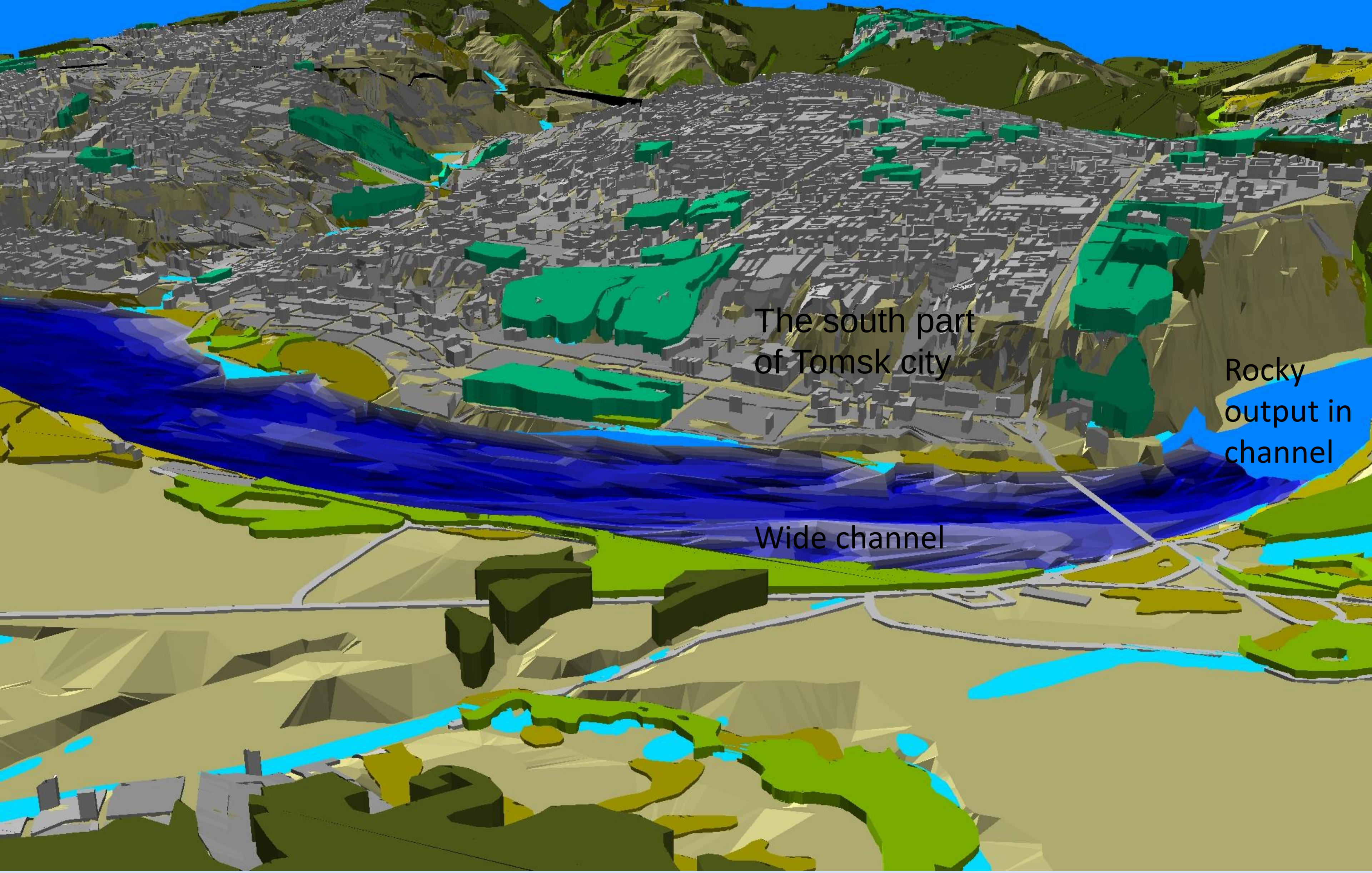
In 1950-1980’s years in the channel of the Tom' river sands and gravels were extracted. Therefore depth of bottom of the Tom river decreased on 2-2.5 m. The probability of flooding decreased also.

In the 1980th year extraction of gravel in the channel of the Tom' river stopped. But at Tomsk some dangerous situation is formed.

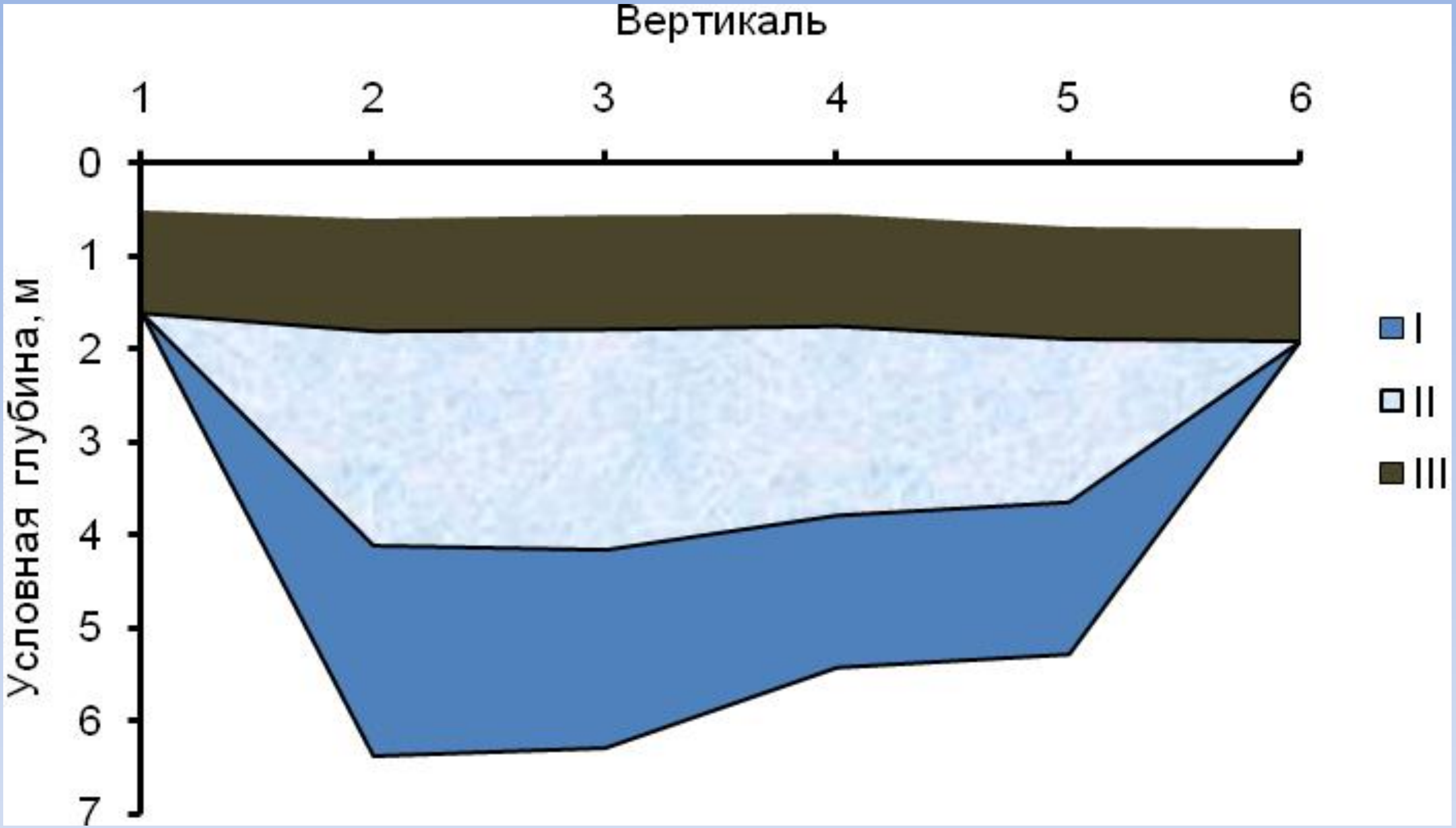


Here is a rock where velocities of flow are very fast. After the rock the velocities are decreased; and the sediments form some islands.

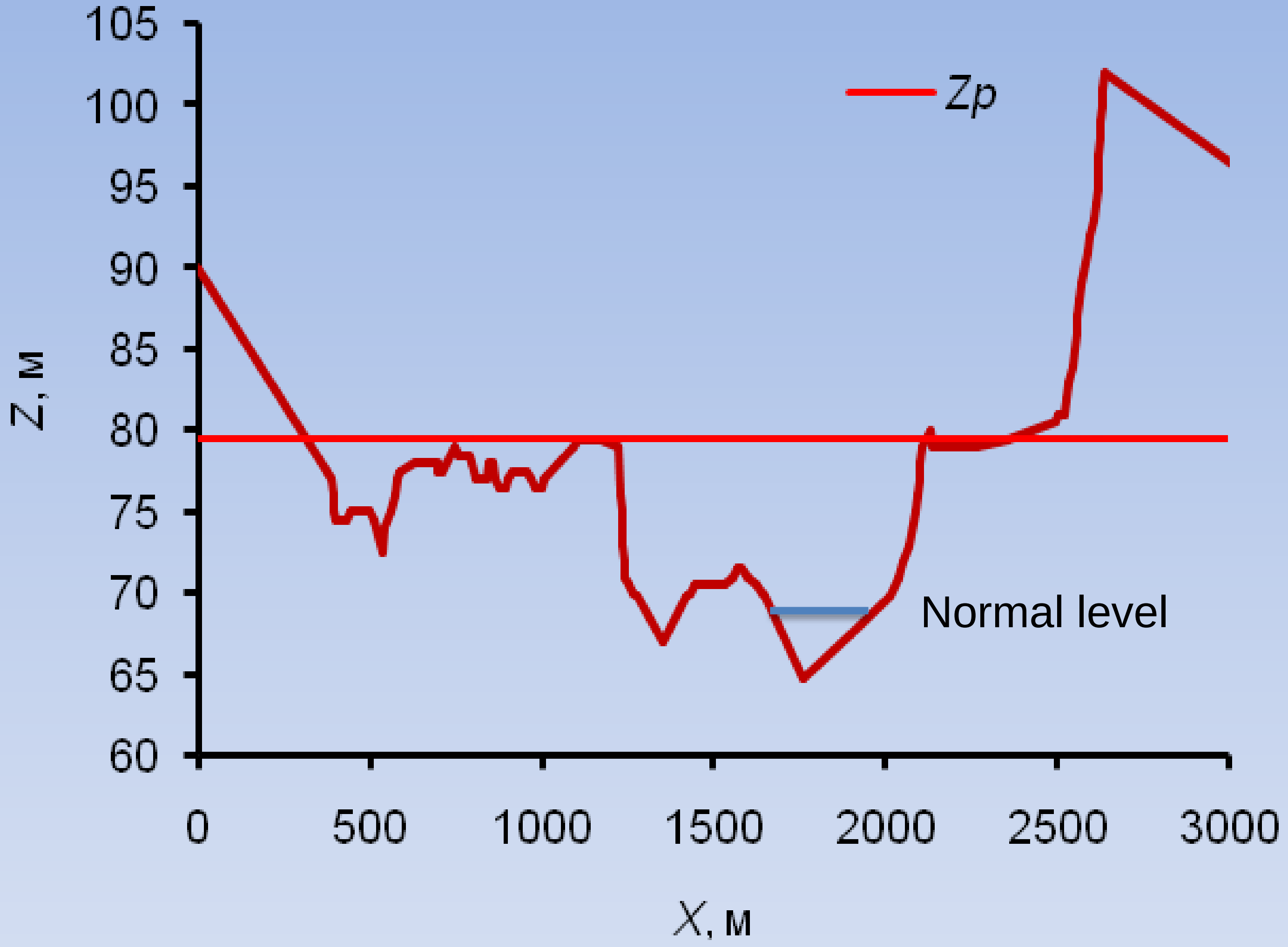
At these islands ice jams and dams are formed, which is the main reason of flooding. Strong flooding occurred in 2004 and 2010. I think their probability are going to increase.



Conditions of formation of ice dams and ice jams on the Tom' river at Tomsk city



Cross-section profile of the Tom river by Tomsk upper the utility bridge in 12.03.2010; I – water; II – frazil ice; III – ice (it is initial stage of ice dam and jam)



Schematic profile of the Tom river under conditions:
Q_p=7071 m^3/sec; η=0.411; $К_{wint}$=0.41

The ice dam and water entrance on the flood-plain upstream Tomsk in April, 2010 (beginning of flooding)

The ice dam and water enter on the left
Flood-plain at Tomsk in April, 2010 (peak of flooding)

1.6. Calculations of the sediment runoff and channel deformation

1.6.1. Classification of river bed evolution

In accordance with the hydro-morphological theory of channel process, developed by the State Hydrological Institute (SHI), all possible schemes of channel deformation of lowland rivers should be subdivided into seven types (fig. 1.6.1.1).

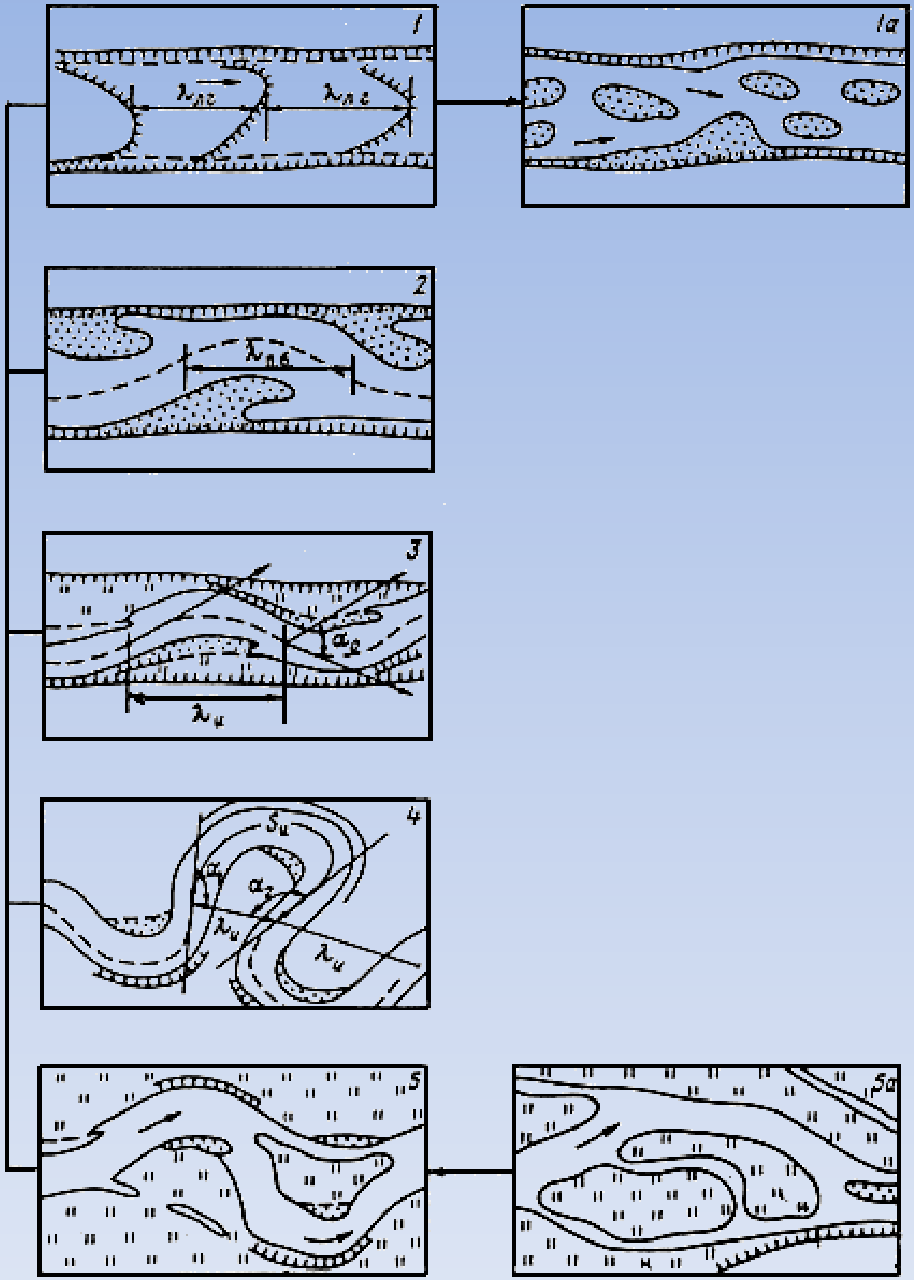


Fig. 1.6.1.1. Types of river bed evolution and their main meters:
1 – ribbon ridge type ($\lambda_г$ - step of ribbon ridges); 2 – “pobochnevy” type ($\lambda_{пб}$ - by-step); 3 - limited meandering ($\lambda_и$ – bend step, α_0 – angle of the bend rotation); 4 - free meandering ($S_и$ – length of the bend, $\lambda_и$ – bend step, α_1 – angle of entry, α_2 –angle of exit); 5 – incomplete meandering; 1a – river bed with many arms; 5a – floodplain with many arms

Ribbon Ridge type (variant of straight channels) of channel process is met in small and average lowland rivers (and the individual arms), built of sand. It occurs rarely. The velocity of ridge moving is determined from the dependence:

$$C_g = 0.019 \cdot V \cdot Fr^3, \qquad (1.6.1.1)$$

where C_g – velocity of ridges offset (m/s); V - average velocity of (m/s); Fr - Froude number.

$$Fr = \frac{V}{\sqrt{g \cdot H}}$$

“Pobochnevy” (shoal; variant of straight channels**) type** of channel process is met in the lowland and submountain rivers, built of sediment of any size. It occurs mainly on the straight and poorly meander sections of the rivers (and the individual arms).

Limited meandering (variant of meandering channels) is the type of channel process, mostly distributed in lowland rivers. It is characterized by angle of rotation to 120°. The free development of the horizontal channel deformation at this type of channel process is limited by the impossibility of washing out of valley slopes.

Free meandering (variant of meandering channels) is the most distributed type of channel process in lowland rivers. It usually develops in the broad river valleys, the slopes of which do not restrict the free development of horizontal deformations.

Incomplete meandering (variant of meandering channels) is a kind of free meandering. It is characterized by the channel which straightens a meander. It occurs on strong flooded lands in the flood-time.

An example of complex of free meandering and ribbon ridge type; the Vasugan river

Floodplain with many arms (variant of braided channels) is the further development and complication of incomplete meandering. It is characterized by a broad flood plain. Channel processes on each arm can develop under the laws of any type.

“Oseredkovy” type (variant of braided channels) of channel process (channel with many arms) is available at the sites of the lowland, mountain and submountain. It is characterized by prostrate channel, on which in flood period the channel forms are moving: “oseredki” (islands without vegetation), shoals and ribbon ridges.

1.6.2. Suspended Sediment Runoff

1.6.2.1. Determination in presence of observational data

Determination of the characteristics of suspended sediment runoff in the presence of observational data in general is similar to the corresponding calculation of the basic characteristics of water flow.

1.6.2.2. Determination in insufficiency or absence of observational data

In case of observational data insufficiency and significant correlation, the charges of suspended sediment can be restored by observation data of water runoff. Restoration of series is similar to the corresponding calculation of water flow.

In absence of data, the total river sediment flow can be estimated by analyzing the equation of channel deformation, which has the form (if width of river ≈ const):

$$\frac{\partial G_1}{\partial x} - G_2 + m_0 \cdot B \cdot \frac{\partial z}{\partial t} = 0 \ , \qquad (1.6.2.2.1)$$

where G_1 – runoff by the flow of sediment;
G_2 – bottom or rising up;
m_0 – relative density of soils and sediments;
B - width of the river, m;
x - coordinate of the longitudinal displacement;
z - elevation bottom; t - coordinate time.

In absence of observational data charge of suspended sediment G_2 can be calculated by the **method of A.V. Karaushev**:

$$G_2=(u+k_u)\cdot S_e-k_u\cdot S_v, \qquad (1.6.2.2.2)$$

where u - hydraulic size of sediment, m/s; S_v - soaring turbidity, g/m^3; S_e - average turbidity at the end of the element Δx, g/m^3; k_u – coefficient of equilibrium:

$$k_u=\frac{u\,\Gamma}{1-\Gamma}, \qquad (1.6.2.2.3)$$

where Γ - hydro-mechanical parameter of sediment, calculated for the *i*-fraction of sediment :

$$\Gamma_i=Б_i\cdot\Phi_i, \qquad (1.6.2.2.4)$$

Function $Б_i$ and Φ_i are defined by A.V. Karaushev, depending on the hydraulic size of sediment u, the vertical fluctuations v_z, the average flow velocity, and the values of Chezy coefficient C_{Ch}.

Turbidity S_v (in g/m^3) is calculated by the formula

$$S_{v.} = 150\, a\, N\, \eta^2 \frac{\overline{v}^2}{\overline{h}}, \qquad (1.6.2.2.5)$$

$$N = \frac{M\, C_{ш}}{g}, \qquad (1.6.2.2.6)$$

$$M = \begin{cases} 0.7\, C_{Ch} + 6 & \text{for } 10 \le C_{Ch} \le 60 \\ 48 & \text{for } C_{Ch} > 60 \end{cases}, \qquad (1.6.2.2.7)$$

where h – average depth of the watercourse, m; $a = S_A / S_t$; g – acceleration due to gravity, m^2/sec; η – coefficient calculated by the formula:

$$\eta^2 = \frac{0.53\, C_{Ch} - 4.1}{C_{Ch} - 2}, \qquad (1.6.2.2.8)$$

Carrying capacity S_t (g/m^3) is calculated by the formula:

$$S_t = \Gamma \cdot S_v, \qquad (1.6.2.2.9)$$

Average turbidity S_e at the end of the element Δx is determined by:

$$S_e = S_t + (S_b - S_t) exp\left(-\frac{B \cdot (u + k_u)}{Q} \Delta x \right), (1.6.2.2.10)$$

where S_b – average turbidity of water in the beginning of part Δx, g/m^3; Q - water flow, m^3/sec.

1.6.3. Bed Load Runoff

1.6.3.1. Determination in the presence of observational data

In the presence of observational data Bed Load Runoff (*G*1) is determined from the dependence $G_1=f(Q)$ for the average daily water discharges Q. Then the runoff is calculated for a decade, month, year. The method of further calculations is the same as for the runoff of suspended sediment.

1.6.3.2. Determination in insufficiency or absence of observational data

In case of insufficiency or absence of observational data, Bed Load Runoff G_1 can be calculated by different methods. Further the methods of I.I. Levi, G.I. Shamov, V.N. Goncharov are stated.

Method of I.I. Levy. The calculation of the value G_1 for rivers, sediments which are sand and gravel, should be carried out according to the formula:

$$G_{1(L)} = 0.002 \cdot B \cdot \bar{d} \cdot \left(\bar{v} - v_{0(L)}\right) \left(\frac{\bar{v}}{\sqrt{g\,\bar{d}}} \right)^{3} \left(\frac{\bar{d}}{\bar{h}} \right)^{\frac{1}{4}}, \quad (1.6.3.2.1)$$

where d – average diameter of sediment, m; $v_{0(L)}$ – velocity of not washing.

$$v_{0(L)} = 1.3 \cdot \sqrt{g\,\bar{d}}\; lg \frac{12\,\bar{h}}{d_{90}}, \qquad (1.6.3.2.2)$$

where d_{90} – diameter of particles with probability 90%.

Method of G.I. Shamov. The calculation of Bed Load discharge in cases of the sandy and sandy-gravelly sediment it is recommended using the G.I. Shamov formula:

$$G_{1(Sh)} = k\,B\left(\frac{\overline{v}}{v_{0(Sh)}}\right)^{3}\left(\overline{v} - v_{0(Sh)}\right)\left(\frac{\overline{d}}{\overline{h}}\right)^{\frac{1}{4}}, \quad (1.6.3.2.3)$$

where *k* - coefficient taking into account the heterogeneity of the load; parameter v_0 is determined by the formula:

$$v_{0(Sh)} = 3{,}7\,\overline{d}^{\frac{1}{3}}\,\overline{h}^{\frac{1}{6}}. \qquad (1.6.3.2.4)$$

Method of V.N. Goncharov. V.N. Goncharov has shown for sediments with the diameter from 0.2 to 10 mm is advisable to apply this design formula:

$$G_{1(G)} = 1.2 \cdot (1+\psi) \cdot \bar{d} \cdot v_{0(G)} \left(\frac{\bar{v}}{v_{0(G)}} \right)^{4.33}, \quad (1.6.3.2.5)$$

where ψ – parameter of turbulence, determined from the special tables of V.N. Goncharov; parameter $v_{0(G)}$ is determined by the formula:

$$v_{0(G)} = \lg\left(\frac{8.8 \cdot \bar{h}}{d_5} \right) \cdot \sqrt{0.57 \cdot g \cdot \bar{d} \cdot \frac{\rho_G - \rho_W}{\rho_W}}, \quad (1.6.3.2.6)$$

where d_5 – diameter of particles with probability 5 %; ρ_W and ρ_G – density of water and sediment.

The confluence of Tom and Ob rivers, the snow melt flood

1.6.4. Vertical Deformation of Riverbed

1.6.4.1. Determination in the presence of observational data

Vertical Riverbed Deformations are determined on the basis of cross-channel profile combination, taking into account the errors of the vertical strain δ_z:

$$\delta_z = \sqrt{\delta_g^2 + \delta_n^2 + \delta_{co}^2}\ , \qquad (1.6.4.1.1)$$

where δ_g – depth measurement error in the riverbed shooting; δ_n – error of depth determination on the plan; δ_{co} – error of cross-channel profile combination (δ_g =0.3 м; δ_n=0.5·Δ_0, where Δ_0 – step between isobaths, m; δ_{co}(м)= 0.5·M_B/1000, где M_B – scale of the profile. If δ_z>0.5 z_{np}, then δ_z=z_{np}+2·δ_z).

1.6.4.2. Determination in insufficiency or absence of observational data

In the absence of measurements an important characteristic of vertical deformations is the lowest level of channel limited erosion profile $H_{min.lim}$. It can be determined by the formula:

$$H_{min.\ lim} = H_{min} - H_g - \Delta_g - \delta, \qquad (1.6.4.2.1)$$

where H_{min} – min bottom elevation within the bend; H_g – height of slipping ridges; Δ_g – additional deformations of the bottom; δ – error of depth measurements. Value H_g is determined by the formula:

$$\text{if } h < 1 \text{ m then } H_g = 0.25 \cdot h, \qquad (1.6.4.2.2)$$

$$\text{if } h > 1 \text{ m then } H_g = 0.2 + 0.1 \cdot h,$$

$$\Delta_g = 0.1 \cdot 1.3 \cdot (h_{5\%} - h), \qquad (1.6.4.2.3)$$

where $h_{5\%}$ – river depth at exceeding probability 5%; h – depth of the river.

The Bank Erosion on the Chulim river (till 50-70 m/year)

1.6.5. Horizontal Channel Deformation

Planned deformations are determined on the basis of maps and topographical materials comparison in view of an average common error δ_y:

$$\delta_y = 0.001 \cdot M \cdot \sqrt{\delta_c^2 + \delta_{np}^2 + \delta_{co}^2 + \delta_{ob}^2}\ , \qquad (1.6.5.1)$$

where M – scale coefficient of combined plans; δ_c – error of channel planned measurement; δ_{np} – errors of reduction of maps to the one scale; δ_{co} – error of map overlapping; δ_{ob} – measurement error of the coast line displacement on the combined plans. Average errors of the planned survey is usually taken as 0.5 mm in the plan scale.

The average error of map reduction to the common scale (in millimeters of the reduced scale) should be equal:

$$\delta_{np} = 0.5 \cdot \sqrt{\left(n_M^2 + 1\right)}, \qquad (1.6.5.2)$$

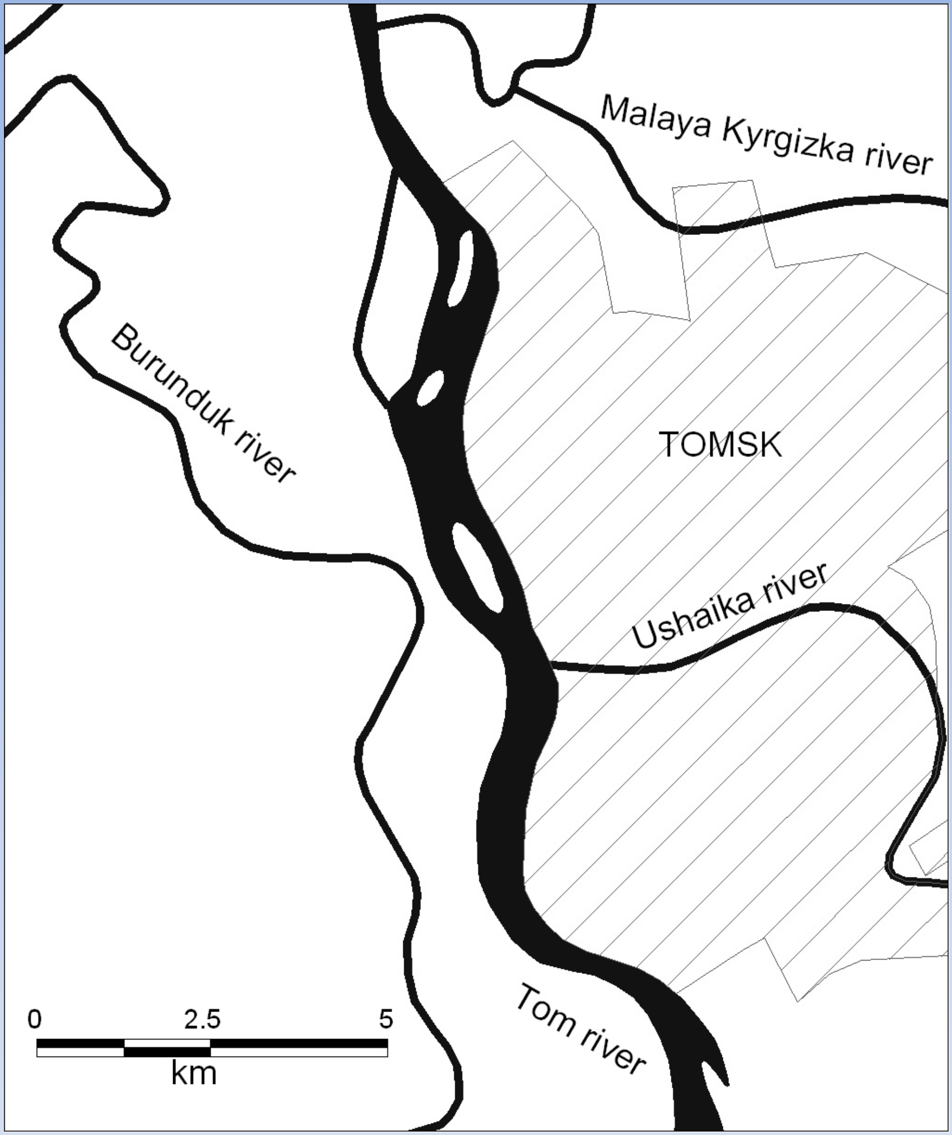
where n_M – ratio of the biggest scale coefficient to the smallest; 0.5 mm – average error of the characteristic length. Combined survey should lead to a larger scale. The error of map overlapping δ_{co} is taken equal to 1 mm in the plan scale in presence of common and time constant benchmarks in the maps.

1.6.6. An example of a sediment runoff

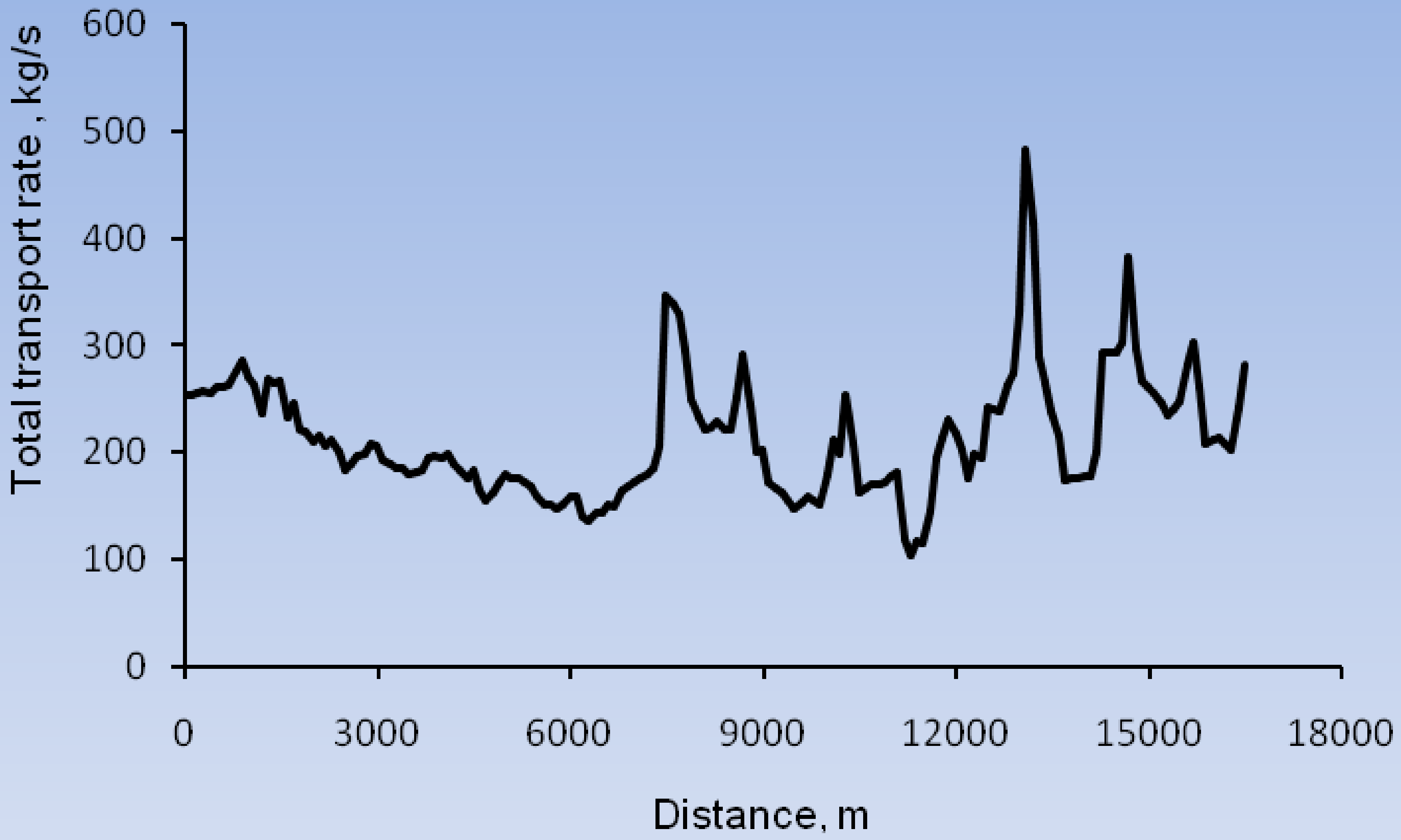
An example – model of vertical deformations of a channel of the Tom' river at Tomsk city for planning change of a channel.

This model was based on the numerical solution of equation (1.6.2.2.1).

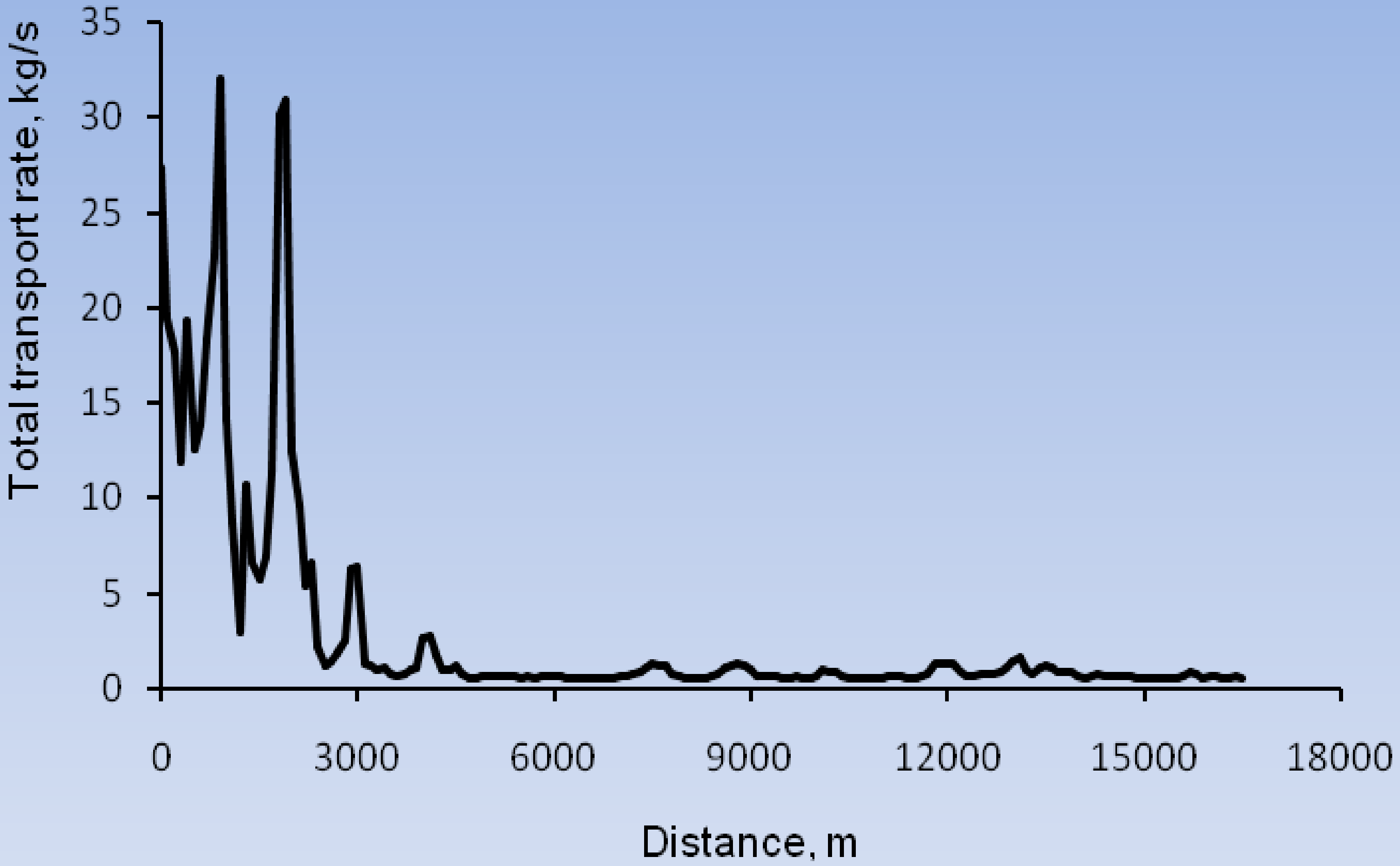
The value of the runoff of suspended sediment was calculated by the method A.V. Karaushev, and the calculation of sediment runoff are executed by the methods of G.I. Shamov, I.I. Levy, K.I. Rossinsky and V.N. Goncharov.



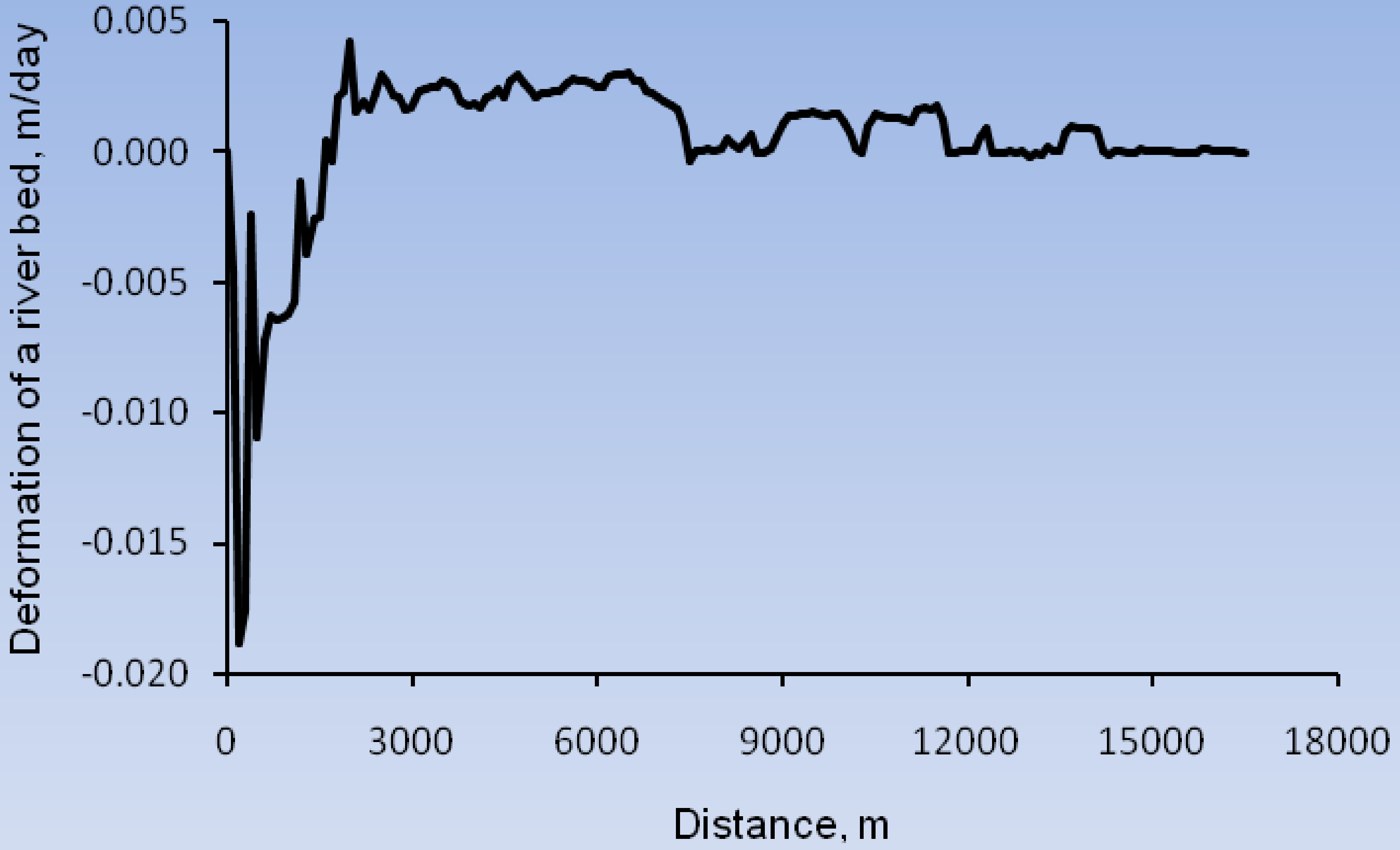
The circuit of a researched site of the Tom‘ river; points from a river mouth (from 74.8 km to 58.3 km)



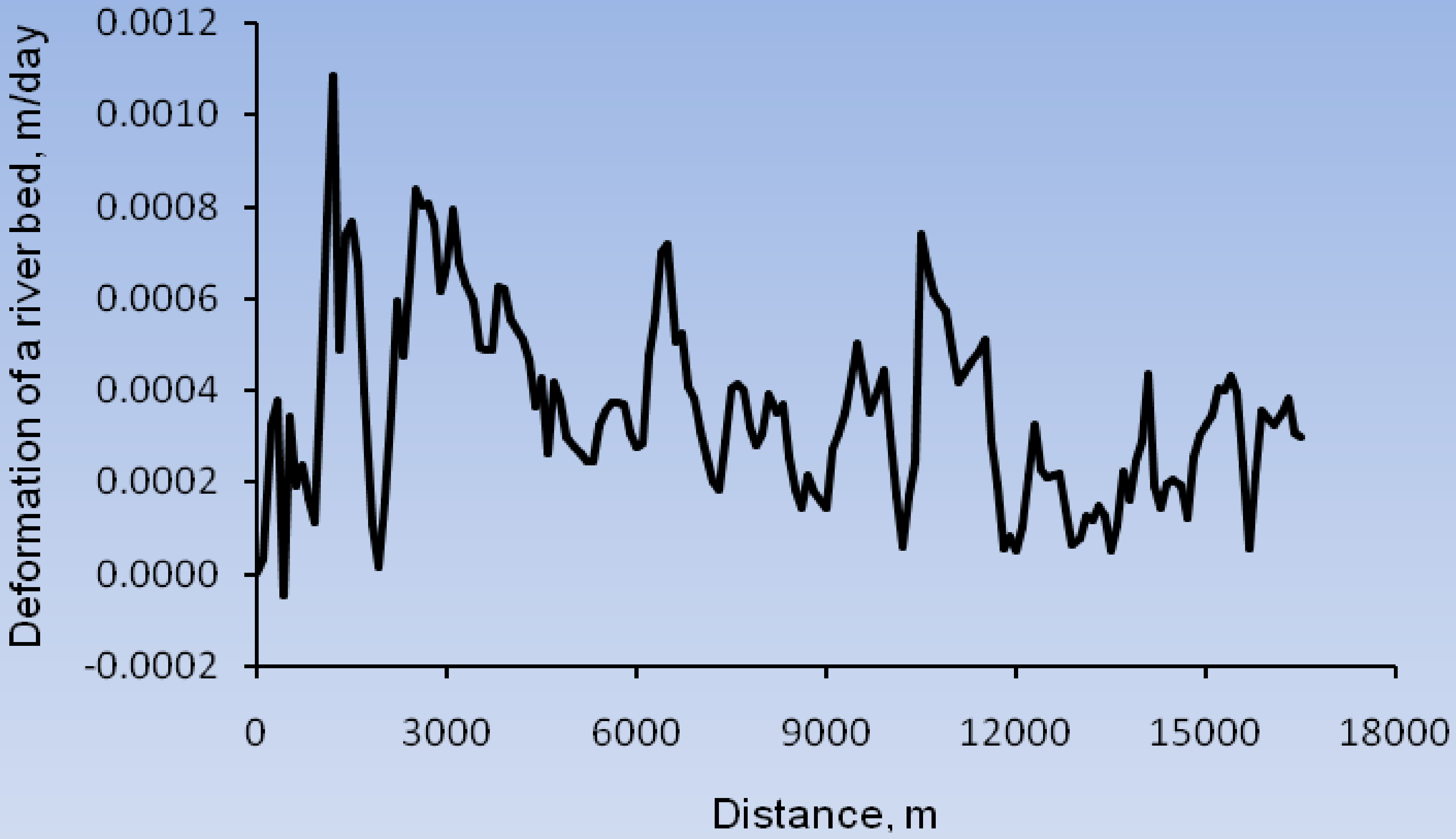
Changing the total runoff of sediment of the Tom river on part 74.8-58.3 km from the mouth; *Q*=6860 m^3/s (average maximum)



Changing the total cost of sediment of the Tom river on part 74.8-58.3 km from the mouth; *Q*=472 m^3/s (low water)



Change in mean cross-sectional deformation intensity bottom of the Tom river on the part 74.8-58.3 km from the mouth; *Q*=6860 m^3/s (average maximum)



Change in mean cross-sectional deformation intensity bottom of the Tom river on the part 74.8-58.3 km from the mouth; Q=1090 m^3/s (annual discharge)

2. Features of mathematical modeling of hydrological processes and the calculations in Siberia

Features of mathematical modeling and hydrological computations in Siberia are defined by following:

1. Siberia is a part of the Northern Eurasia. Conditions of the flow formation in this huge area are extremely varied.
2. Hydrological and meteorological studies of Siberia are generally weak. Network of hydrological observation is rare.
3. The role of swaps in the formation of an average, maximum and minimum flow is not enough studied .
4. Methods of swamp water level computation are remained at the level of the sixties (1960); and it does not satisfy the demands of practice and theory.

The Aк-Turu river (the Mountain Altai)

The Ob river near the mouth of the Tym river

5. In the permafrost the displacement and change of the water body boundaries on the territory are possible up to several tens meters per year.
6. One of results of roads and pipelines construction is superfluous humidifying of territories nearby engineering objects. Sometimes this effect incorrectly connects with the global warming.
7. Actually the observations of lake water levels in forest, forest tundra and tundra-forest areas are absent.
8. Actually the observations of sediments and deformations of river bed are absent.
9. Roshydromet observations of ice and thermal regime of rivers do not allow to assess the formation conditions of ice jams and ice dams.

The Bog without name with forest island in basin of Chaya river

10. There are very serious problems with access to hydro-meteorological information in Russia.

11. The rivers of Siberia are often characterized by large amplitude of levels and flow variations (10 m and more), ice jams and hanging ice dams, and in some cases by large bank deformation (more 10 m/year). This sets great difficulties with the construction and operation of some engineering objects in the river valleys.

12. Russia's current methods of the calculations do not focus on new types of information, associated with distant methods; and they do not allow to estimate changes in hydrological processes under climate change conditions for certain.

Thanks for your attention