HYDROLOGICAL PROCESSES
Hydrol. Process. (2010)
Published online in Wiley Online Library
(wileyonlinelibrary.com) DOI: 10.1002/hyp.7901

An approach to the scaling problem in hydrological modelling: the deterministic modelling hydrological system

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Abstract:

A brief description of the Deterministic Modelling Hydrological System (DMHS) 'Runoff–Erosion–Pollution' proposed by the first author is presented. This system is being developed with the aim of giving it a universal character so that it can be applied in mountainous and flat terrain, and in basins of different natural climatic zones regardless of their size. The main feature of the model is its independence of the scaling problem. The basis of our approach consists of a simple theory of runoff elements. This is different from the generally accepted use of partial differential equations such as the Saint Venant equation for surface and channel flow and the Boussinesq equation for underground waters describing the water movement from runoff formation origins to the basin outlet. The results of runoff simulation for six mountainous watersheds of different sizes across Eastern Siberia within the Lena River basin and their statistical evaluation are presented. The selected river basins ranged in size from about 40 km^2 (small scale) to the entire Lena River basin (2·4 million km^2), classified as a large-scale basin.

KEY WORDS deterministic modelling hydrological system; scale problem; the principle of universality; concept of runoff elements; Lena River basin

Received 27 September 2009; Accepted 15 September 2010

INTRODUCTION

The need for large-scale hydrologic modelling was first expressed by Eagleson (1986). Current research points out the main spheres of natural sciences where the improved understanding of hydrological process on a large scale has become increasingly important, such as proper representation of hydrological feedback to the climate system in global circulation models (Wood *et al.*, 1992; Evans, 2003), water resources applications (Lettenmaier *et al.*, 1999), and the evaluation of anthropogenic effects due to land-use change (Vörösmarty *et al.*, 2000; Rost *et al.*, 2008).

Within the climate change agenda, large-scale modelling is especially required for those remote areas of the world which are still characterized by poor availability of any data but strongly affecting the globe climate system, such as the cold environment basins in the Arctic (Serreze *et al.*, 2000) or tropical watersheds in South America (Makarieva and Gorshkov, 2007). Current alteration of the hydrological cycle due to different impacts is evident and widely documented (Yang *et al.*, 2002; Haddeland *et al.*, 2007).

The examples of large-scale hydrological models developed in the last years can be found in Singh and Frevert (2002). They include the VIC type models (Liang *et al.*, 1994); the global model by Hanasaki *et al.* (2007a,b); ISBA-MODCOU (Habets *et al.*, 1999); WATFLOOD (Kouwen and Mousavi, 2002); WATCLASS (Soulis *et al.*, 2000); the MESH modelling system (Pietroniro *et al.*, 2007); SWIM (Krysanova *et al.*, 1998), MGB-IPH model (Collischonn *et al.*, 2007), and the SHE model (Abbott *et al.*, 1986a,b) applied on a large scale by Andersen *et al.* (2001). There are also the global water-balance models, such as WBM (Vörösmarty *et al.*, 1996), Macro-PDM (Arnell, 1999), WGHM (Doll *et al.*, 2003), and WASMOD-M (Widen-Nilsson *et al.*, 2007). They are usually applied on a global resolution, for instance $0{\cdot}5^\circ \times 0{\cdot}5^\circ$, but may be used also in regional studies (Arnell, 2005; Lehner *et al.*, 2006).

Much of the work in large-scale hydrologic modelling has been driven by the need to incorporate subgrid heterogeneities in the land-surface components of climate and atmospheric models. The land-surface models use increasingly complex descriptions of the physical mechanisms, requiring the specification of a large number of parameters controlling water and heat fluxes (Franks *et al.*, 1997), thereby running the risk of falling into the equifinality problem defined by Beven (2006). Despite the complexity of these models, their performance from the hydrologic perspective does not match that of simpler, hydrologically dedicated models (NOAA, 2010). Arguably, reliable representation of the hydrological processes by these models requires the application of upscaling techniques to transfer information from small

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scales (where the parameters are actually observed) to larger scales where the models apply.

A view of large-scale hydrologic modelling would be incomplete without a mention of the treatment of the subsurface and groundwater processes which are of considerable importance in water resources planning and management. Clearly, an exhaustive review of the literature on groundwater models is beyond the scope of this paper. However, we would like to make the following points. Virtually all large-scale hydrologic models have a detailed representation of the surface processes, while the subsurface processes tend to be considerably more simplified. Groundwater contribution to streamflow has been approached in most contemporary models by splitting flow between a fast (surface and subsurface) and a slow component (groundwater). Of late there has been a trend to couple high-resolution surface models with complex groundwater models, as is the case of Facchi *et al.* (2004) and Markstrom *et al.* (2008).

The split-flow technique does not take into consideration the different levels of interaction between groundwater and surface flow that is a function of the basin scale. The more complex methods have a different problem, namely that of the estimation of the parameters of the model for large basins. This paper will describe our approach in the Section on 'Concept of Runoff Elements'.

The above paragraphs illustrate why scaling is one of the main problems of contemporary mathematical modelling of hydrological processes.

Klemes (1983) was one of the first to state that one of the main reasons for slow progress in runoff modelling of natural river basins is inadequacy of observed dominant effects at different scale watersheds of limited continuity.

Dooge (1986, 1999) stated that the scaling problem exists from a scientific point of view: fully 'complete' hydrologic science must cover scale ranging from molecular (10^{-10} m) to global (10^8 m). The equation of continuity is linear and therefore can be rescaled without the need to estimate new parameters and their variance in new scales. This might lead to a new fundamental theorem at all scales. All other equations, however, are not linear and therefore complicate the scaling problem as well as lead us to the problem of complex non-linear feedback processes. Hydrologic theory remains theoretical until it is confirmed with reliable data related to scaling issues.

Kundzewicz (1993) argues that the representation of processes on a micro scale might be much more complicated than at the macro scale and therefore scaling will remain the centrepiece of hydrology. Refsgaard *et al.* (1996) stated that a universal scaling system has not been developed and is not expected to be developed in the near future.

Beven (1996) links the scaling issue with parameters of distributed models, suggesting that the parameterization improvement should begin with data collected at the scale required for modelling. Wood (1995) attributed the scaling problem to 'heterogeneity in land-surface characteristics' on a small scale.

The scaling problem is considered to be fundamental as this is the main source of uncertainties introduced by modelling. In general, it could be summarized in the following main statements:

- Parameters of macro scale models are generalized parameters at the micro scale.
- Relationships and equations are different for different scales.
- Equation parameters are different at different scales.
- A universal scaling methodology, allowing transition from one set of scale parameters to any other, is highly desired and still undeveloped.
- Data are to be collected at the scale required by modelling.

This summary suggests that the scaling problem could be at odds with the idea that the process of runoff formation in any point of space is controlled by the same fundamental laws of physics. The latter implies that mathematical theory and parameters must be the same for all scales if the model is adequate for the natural processes.

In the developed Deterministic Modelling Hydrological System (DMHS) 'Runoff–Erosion–Pollution' (Vinogradov, 1988; Vinogradov and Vinogradova, 2008, 2010), often called 'Hydrograph' (Semenova and Vinogradova, 2009), the scale problem, which is related mainly to methods of mathematical description of water movement from the places of runoff formation to the basin outlet, has been minimized. Our research focuses on the application of the DHMS in natural catchments with sizes ranging from the order of 10^{-1} km^2 to the order of 10^6 km^2.

A different idealization of runoff generation and its transformation phenomena is the basis of our approach. Two main ideas are constantly being taken into account while designing and constructing this model. They are as follows:

- The necessity to reach a relative balance in searching for the simplest solutions while aiming to describe the natural processes and laws adequately
- The maintenance of a general approach, i.e. the description of the whole set of any possible situations in runoff formation processes, catchments of any scale (from water-balance plot to the Earth land-surface), mountains and plains, any geographical zone, and use of minimum standard meteorological information

The principle of universality is based on the fact that physical laws are universally applicable. Therefore, it should be possible to develop models that can be used in any geographical setting. Further discussion on this topic can be found in Vinogradov and Vinogradova (2008, 2010) and Semenova and Vinogradova (2009). The principle of universality should not lead to neglect of runoff formation peculiarities even in the most specific conditions.

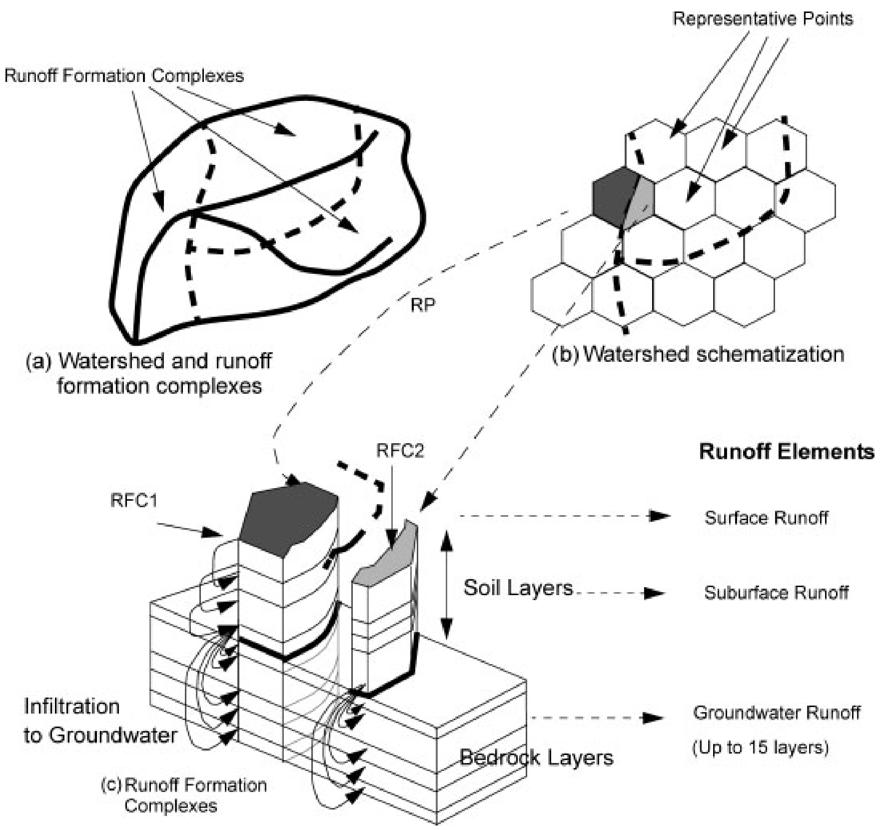

Figure 1. Basin schematization in the DHMS

Nowadays, it is striking to see the predominance of similarities between different proposed models rather than their differences. That is why we draw attention to an old but important warning expressed by Richard Courant "... as exact physical laws are just the idealizations and as any given physical situation can be idealized by many different methods it is important to be able to choose acceptable idealizations" (Courant and Hilbert, 1937).

Within the frame of search for an acceptable idealization, we should discuss the question about the ways and forms of water movement after precipitation or snow melt reaches the river basin surface, and above all, the approaches for the mathematical description of this movement.

This paper intends to demonstrate that the scaling problem can be avoided by using appropriate algorithms of runoff modelling. It illustrates the solution to the problem using basins of different sizes as examples. Six nested mountainous watersheds of various sizes differing from 40·2 to 2 430 000 sq km within the Lena River basin were chosen as study objects because of their variety of landscapes, geomorphology, and climate characteristics.

We defined our experiment by estimating the model parameters to small headwater watersheds (40·2 km^2) within the larger Lena River basin. Then we applied the estimated parameters to the larger watersheds that contained the smaller ones, until we arrived at the overall Lena River watershed ($2{\cdot}4 \times 10^6$ km^2). These simulations were performed without change in the model's structure and without resorting to calibration procedures for most of the model's parameters. So far, the results from the application of the DMHS in the runoff simulation of different mountainous watersheds show that the approach is valid, although, certainly, more tests are required to fully prove it.

The paper is organized as follows: the basic model structure, input data requirements, and model output are covered under the Section on Brief Model Description; the Section on Methodology of Water Movement in the Basin describes the model approach to the formation of surface flow and infiltration, the concept of runoff elements, and flow routing; the Section on Basin Schematization explains the system of representative points (RPs), the system of runoff formation complexes (RFCs) and how the models treat the problem of a heterogeneous snow cover; model parameters are described in the corresponding section; the study area and the results are in the Section on Model Implementation; and, finally the conclusions of the study are in the last section.

BRIEF MODEL DESCRIPTION

Basic model structure

It is important to give the readers a concept of the basic DMHS elements. These will be described in more detail below.

In the horizontal dimensions, the model divides the watershed into a number of RPs. Although the area represented by each point could be arbitrary, we have chosen each area to be a regular hexagon (Figure 1). Also, in

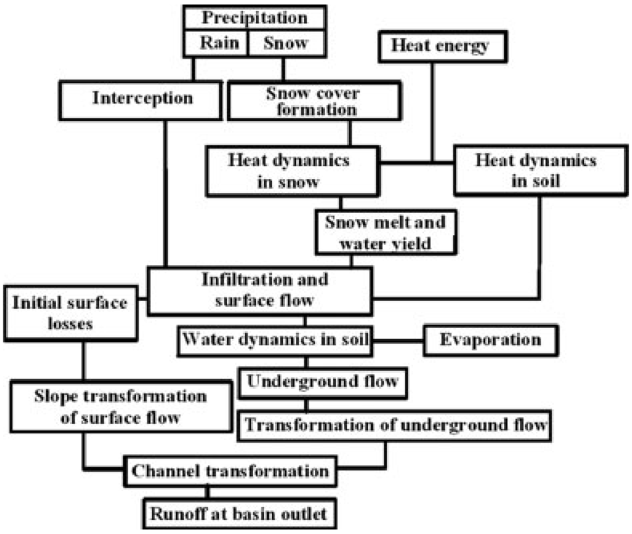

Figure 2. Block diagram of the DHMS modules

the horizontal dimensions, the watershed is divided into RFCs, which are assumed to be homogeneous for soils, vegetation, topography, hydrology, etc. and which may cover from a fraction of an RP to several RPs.

In the vertical direction, the model represents the soil column with at least 3 strata (usually, up to 10–15), for which energy and water balance are computed, and whose physical properties (model parameters) are arranged by RFCs. In addition, the model considers 15 layers for the deeper groundwater flows, for which only the water balance is computed (see the Section on Runoff Elements below).

The model describes all the components of the land hydrological cycle, including precipitation and its interception; snow accumulation and melting; evaporation from snow, soil, and vegetation cover; surface flow and infiltration; soil water dynamics and flow; heat dynamics and phase change in soil layers; underground flow formation, and slope and channel flow transformation; and flow discharge (Figure 2). For a detailed description of approaches used in the model, see Vinogradov (1988, 2003a,b,c,d) and Vinogradov and Vinogradova (2008, 2010).

Forcing data requirements and model output

The model input consists of standard meteorological information, such as values of air temperature and relative humidity, and precipitation.

The model algorithm includes the following computation routines: precipitation and its interception; snow accumulation and melting; evaporation from snow, soil, and vegetation cover; surface flow and infiltration; soil water dynamics and flow; heat dynamics and phase change in soil strata; ground flow formation, and slope and channel flow transformation; and flow discharge.

As the model describes the complete land hydrological cycle, it has various output results. First of all, there are the continuous runoff hydrographs at the outlet, from any part of the basin, from a specified landscape, or any set of RPs. In addition, the model also presents the distributed state variables, reflecting water and heat dynamics in soil strata and snow cover. The model can be run at time intervals of 1 day or less, although it only has been tested with daily data. Results from the model including spatial and temporal distribution of water balance elements such as precipitation; evaporation from snow, soil, and vegetation cover; and surface, soil, and underground runoff can be obtained at any averaging time interval.

METHODOLOGY OF WATER MOVEMENT IN THE BASIN

In this section, we discuss three main processes which are at the heart of our model. They are infiltration, water movement in the soil layer, formation of classical surface and subsurface flow; slope (surface, subsurface and underground) inflow to channel network; and flow routing. We focus on these three problems because they are particularly sensitive to scale dependencies in conventional physically based models.

Formation of surface flow and infiltration

Let us consider the following simple approach. We imagine such a picture: during rainfall, drops fall down on an elementary plot occasionally and independently of each other. The quantity and volume of drops per unit of area and time are determined by the rainfall intensity. It means that the rainfall intensity serves as an argument which, together with the maximum possible infiltration rate (i.e. infiltration coefficient f_0), determines the relative size of the infiltration area. We assume that the increment of the infiltration area corresponding to an increment of rainfall intensity i decreases proportionally to this area. Then it follows that the intensity of surface flow q is determined by Equation (1):

$$q = i - f = i - f_0[1 - \exp(-i/f_0)]. \qquad (1)$$

Contrary to approaches in which rainfall produces soil surface ponding if, and only if, the rainfall intensity exceeds the hydraulic conductivity of the saturated soil, our use of Equation (1) indicates that, in fact, surface runoff can form even in a case when the rainfall intensity is lower than the infiltration coefficient (hydraulic conductivity of saturated soil).

Equation (1) allows us to calculate the surface flow formation by a given pluviogram. It is useful to get the equation for computation of surface flow for the whole period of rainfall apart from the initial losses. With this purpose, let us turn to the possibility of transforming the stochastic process of rainfall intensity $i\{t\}$ to that of the rate of surface flow generation $q\{t\}$ (for details, see Vinogradov, 1988, pp. 266–270).

A stochastic function (or process) is such a function in which the variable is random regardless of any argument (in this instance time, t) value. For convenience, we consider some random process:

$$i^*\{t\} = i\{t\}/I \tag{2}$$

where

$$I = (1/T)\int_0^T i\{t\}\mathrm{d}t \tag{3}$$

is the mean rainfall intensity for the period of rainfall T. The fact is that the random process $i^*\{t\}$ has acquired the so-named ergodic property, when each of its realizations is a valid element of a single stationary process whose statistical average is equal to 1. It is known that the distribution law of $i\{t\}$ can be suitably represented by an exponential distribution:

$$\varphi(i) = (1/I)\exp(-i/I) \tag{4}$$

$$\varphi(i^*) = \exp(-i^*) \tag{5}$$

Herein, the asterisk notation (*) along with any argument indicates that the last is divided by I.

Accounting for the physical Equation (1), the probabilistic average of random process such as the difference between the rainfall intensity and infiltration is calculated as

$$M(q^*) = \int_0^\infty \{i^* - f_0^*[1 - \exp(-i^*/f_0^*)]\}\varphi(i^*)\mathrm{d}i^* = 1/(f_0^* + 1) \tag{6}$$

Finally,

$$M(q) = I^2/(I + f_0) \tag{7}$$

Then the value of surface flow formation H_q during the period T is the following:

$$H_\mathrm{q} = H^2/(H + f_0 T) \tag{8}$$

where H is the precipitation depth, and equals IT.

The applicability of Equation (1) was confirmed by various data of artificial irrigation experiments (actually, initially it was derived from those data) and verified by the observation of rainfall and surface flow processes at small watersheds in Central Asia with the surface flow generation (Vinogradov, 1988, pp 89–92).

The use of a constant infiltration rate is a key assumption of our model, and it is the main difference between our approach and the variable infiltration assumption, whether it is the traditional Richard's equation-based approaches such as, for instance, tRib, (Ivanov *et al.*, 2004), or non-linear infiltration approaches such as the Sacramento model (Burnash and Ferral, 1974) and the Precipitation-Runoff Modeling System (PRMS) (Leavesley *et al.*, 1983), to cite just a few. The variable infiltration rate can be observed under pressure head, which only occurs under ponding (if the underneath soil is dry), or on a dry floodplain, or under exceptionally heavy precipitation intensities. Then the traditional models assume a uniform wetting front across each grid cell.

In our approach, the soil column can be presented by the number of soil strata (the number can be from three to how many the modeller decides). Let us call the top soil stratum of the soil column the first stratum, the next one is the second stratum and so on. Each soil stratum has its own properties (porosity, maximum water-holding capacity, hydraulic conductivity. . ., Table IV). In Equation (1) the hydraulic conductivity of the first stratum is used. Therefore, the depth of this first stratum can be set to some minimum depth in order to represent the surface of the soil column, although typically we use greater depths (e.g. 10 cm).

Equation (1) is used to calculate the amount of water available for surface flow. After the surface depression storage is filled up, the excess water forms the surface flow. The rest of the water infiltrates into the first stratum, filling it to its maximum water-holding capacity. Once this stratum becomes saturated, water flows to the next stratum. Excess water in each stratum contributes to the subsurface flow. It can happen not only if the infiltration rate is low but also in the case when the lower stratum is frozen. That is why the water balance of the soil strata is accompanied by energy balance in the model. To allow for cracks, preferential infiltration paths, worm holes, etc. and those features that do occur in nature, it may be necessary to calibrate the infiltration coefficient.

Finally, water leaving the last discrete soil stratum is redistributed between the different aquifer layers. This distribution is controlled by means of coefficients that require manual calibration.

Concept of runoff elements

An important requirement while applying the mathematical modelling is the following: "the chosen method for the solution of a task must be such that only such a data which can be obtained with a required reliability be meant for input. If initial data cannot be achieved accurately enough, then in many cases it is expedient to change the approach usually simplifying it. . .", or ask yourself the question: "is it not easier to measure directly the value which theoretically can be also calculated?" (Myshkis, 1994).

We used these ideas for the solution of a problem such as slope (surface, subsurface and underground) runoff generation. The approaches to calculate the runoff movement at the slopes from the place of its generation to the channel net using unavailable data about inclinations, morphometry, roughness etc. seem to us utopic. As the depths of the surface slope flows are measured in millimetres and centimetres, the spatial step should be only a cut above. High temporal preciseness of physically based models is doubtful as the processes of small temporal scales are smoothed out by the rough spatial resolution. The other problem is that the results of calculations cannot be compared with the real processes as spatial-temporal observations for the water flowing down over the slope surfaces are rarely carried out. The model parameters, estimated in reverse against the

Table I. The system of runoff elements

Number	Type of runoff	a^* (m^{-1})	Time (τ)	Outflow intensity ($dm^3\ s^{-1}\ km^2$)	Water storage (mm)
—	Surface	1000	17 m	10^5	4·6
—	Subsurface	100	2·8 h	10^4	24
1	Rapid ground	10	1·2 days	10^3	69·3
2		3·162	3·7 days	464	121
3		1	11·6 days	215	195
4	Ground	0·3162	1·2 months	100	301
5		0·1	3·8 months	46·4	454
6		0·03162	1 year	21·5	674
7	Upper underground	10^{-2}	3·2 years	10	995
8		$3{\cdot}162 \times 10^{-3}$	10 years	4·64	1464
9		10^{-3}	32 years	2·15	2152
10	Deep underground	$3{\cdot}162 \times 10^{-4}$	100 years	1	3161
11		10^{-4}	320 years	0·464	4640
12		$3{\cdot}162 \times 10^{-5}$	1000 years	0·215	6812
13	Historical underground	10^{-5}	3200 years	0·1	10 000
14		$3{\cdot}162 \times 10^{-6}$	10 000 years	0·0464	14 678
15		10^{-6}	32 000 years	0·0215	21 450
$\sum$(1–15)					67 166

observed runoff in the basin outlet, are the subject not of systematization, generalization, or normalization; often they are not realistic.

The concept of runoff elements, briefly stated here, offers the possibility of a unified methodological approach to modelling the surface, subsurface, and the underground runoff of different layers. The solution to the multiple-scale problem by directly including the basin area into the algorithms of conversion of runoff element parameters into coefficients of main calculating equations is presented below.

The water flowing before it reaches the river network is dispersed over the runoff elements. These are the natural formations originating as a result of the interaction between water on one hand, and soil cover and the upper layer of the lithosphere on the other hand. The runoff elements can be surface, subsurface (soil), and underground. Their linear dimensions change over an extremely wide range: from several centimetres (at the surface of eroded slopes) to many (tens, hundreds, thousands) kilometres (in the underground lithosphere structures).

The theory of runoff elements is very simple. The basis is the usual water balance relation

$$dW/dt = S - R \tag{9}$$

where W is the water volume that is accumulated by runoff element (m^3), S and R—inflow and outflow to/from it ($m^3\ s^{-1}$). There is the non-linear relation between W and outflow discharge R described in the model by an empirical equation:

$$R = b[\exp(aW) - 1] \tag{10}$$

From Equation (10) it is possible to derive the corresponding equation of the outflow hydrograph from runoff elements of a given layer, although the derivation is too elaborate to be included here (Vinogradov, 1988):

$$R = (S + b)/\{1 + [(S - R_0)/(R_0 + b)] \exp[-a\Delta t(S + b)]\} - b \tag{11}$$

Here, R_0 is the initial value of runoff R and S is the input rate ($m^3\ s^{-1}$); Δt is the computational time interval (s) during which S is constant; where a, b—hydraulic coefficients (which determine the conditions of outflow) with dimension m^{-3} and $m^3\ s^{-1}$ In the general case, we can assume that the number of runoff elements is proportional to the basin area F (m^2) or a fraction of it ΔF, as we will see below and then $a = a^* \times F^{-1}$ and $b = b^* \times F$. The coefficients a^* and b^* are the subject of our further attention.

It is useful to attach to the coefficient a^* the status of a conditional constant systematized by types of flow and the coefficient b^* to be included in the list of main parameters of the runoff model. The units are the following: a^*—m^{-1}, b^*—$m\ s^{-1}$, F—m^2. The product $ab = a^*b^* = \tau^*$ can be named as the specific time of discharging of a runoff element. The specific values of outflow q and water storage J are also determined by the values of the conditional constant a^*, parameter b^* and are connected by the relation

$$J = \ln(q/b^* + 1)/a^*. \tag{12}$$

Next, we offer a probable idealization—hierarchical sequence of layers of runoff elements arrangement which take part in river inflow. It does not contradict known processes, phenomena, and laws.

All specific values of runoff elements are determined by the conditional constant a^* and the median value of parameter $b^* = 10^{-6}\ m\ s^{-1}$. The 'constant' itself is sequentially determined by the expression $a^* = 10^i$, where $i = 3$ for surface flow, $i = 2$ for subsurface flow,

Table II. Characteristics of average flow velocity (m s^{-1}) observed at the hydrometric stations in the main rivers of Eastern Siberia

The range of basin scales (km^2)	The Lena River		The Yana River		The Indigirka River		The Kolyma River	
	V	C_V	V	C_V	V	C_V	V	C_V
<100	0·94	0·42	1·28	0·22	1·4	0·16	1·60	0·14
100–1000	1·31	0·34	1·31	0·32	1·52	0·18	1·62	0·15
1–10 000	1·10	0·42	1·60	0·32	1·77	0·21	1·90	0·11
10–100 000	1·55	0·20	—	—	2·15	0·19	2·14	0·23
>100 000	1·72	0·19	—	—	—	—	—	—
Average	1·40	0·30	1·39	0·28	1·52	0·25	1·65	0·21
The number of observational stations	228		50		46		117	

C_V, absolute variation coefficient of averaged velocities.

and further for different layers of underground flow from $i = 1$ to $i = -6$ with step $\Delta = 0, 5$.

Let us accept a logical and expected assumption: the infiltration capacity of water-holding rocks naturally decreases with the increase in depth. At the same time, two empirical facts should be taken into account—the decrease in outflow rate and the simultaneous increase in water storage with depth in groundwater aquifers. So we postulate the following hierarchical system of layers located each under another layer of runoff elements, feeding the river and corresponding types of underground flow (Table I). Therefore, the model assumes that the groundwater runoff is modelled by having different constituents of the groundwater flow, and the contributions to each from the bottom layer of the soil is controlled by parameters that need calibration. Notice that, likely, historical underground runoff (i.e. layers 13–15) can be referred to as 'hydrological illusions' but their inclusion gives some completeness to the proposed schematization.

It is interesting to compare the total sum of specific stores of water in the system of underground runoff elements of different layers which amounted to 67 166 mm (without 15th layer 45 716 mm) with the data of different sources: from 45 000 to 70 800 mm. Those are typical values for each layer, computed from the a^* and b^* parameters.

Channel flow and lag time

Now let us discuss the task of routing water from the place of its appearance in the system of the river network to the basin outlet.

The concept itself is very easy. It consists of two assumptions:

- The lag time of water channel run to the basin outlet is assumed to be a constant for each point chosen in the basin.
- We use directly measured mean flow velocities at cross-sections of the river channel. Such data was being published in Russia in hydrological year-books.

To illustrate the proposed approach, Table II presents the velocity values for the four large rivers of Eastern Siberia. These velocities are computed by averaging the velocities corresponding to the top 10% of flow values for each of the rivers.

We have drawn a conclusion that a great many factors affecting the acceleration or conversely slowing down of flow velocity surprisingly lead to general order, steadiness, and relative constancy as a result of some 'self-regularity'. The values in Table II indicate that the departures of flow velocities from their average value are insignificant and not too dependent on basin size and spatial variability. In this way, the averaged minimum value of lag time (corresponding to maximum velocities) is taken as the calculating value. This is due to the fact that short time lags correspond to peak limb of flow hydrograph. At the same time, we assume that the underestimated shift of hydrograph shape during the period of low flow does not affect the resulting runoff significantly.

Perhaps a more appropriate velocity to define travel time would be wave celerity (Beven, 1979; Romanowicz *et al.*, 2006). However, with the data available to us in these rivers, it was impossible to estimate wave celerity and we have found that the velocity computed as explained above works quite well in most cases. This topic needs additional research.

BASIN SCHEMATIZATION

This section explains in more detail the concepts of RPs and RFCs that were introduced earlier in the Section on Brief Model Description. In addition, we also describe how the model treats the problem of heterogeneity of snow cover.

System of RPs

Some words about the spatial-calculating schematization of the river basin should be mentioned. If we are able to mathematically and algorithmically describe the runoff formation processes at the local point (or better, at some elementary unit) within the river basin area, then there is the need to formulate the principle that some multitude of these points can completely represent this river basin. In other words, a regular system of points within the watershed is required. There are not as many variants of that system. For our purposes, a hexagonal grid seems to be appropriate because it possesses the property that the centre point of each grid cell is equally distanced from the six neighbouring points at a distance ΔL. We call the points which evenly cover the basin area and are located from each other, the RP. Each RP has under its 'control' the hexagonal area. We call it RP-area:

$$\Delta F = 0{\cdot}866\Delta L^2 \tag{13}$$

The basin area F, the number of points n, and the distance between the neighbouring points (the size of the hexagonal grid) are related between one another by the following ratio:

$$n = 1{\cdot}1547F/\Delta L^2 \tag{14}$$

The RP is characterized by geographical coordinates, altitude above sea level, aspect, and surface inclination. The forcing data of meteorological stations is interpolated into the RP. The interpolation methods generally are those required for the modelling procedure while preparing the information. In such a way, the RP-areas are the equal equivalent elements of a river basin (apart from those which are crossed by a water divide).

It is difficult to recommend a number of RPs for each given basin. But it is obvious that it should be non-linearly related to the basin area. As a rule of thumb, the number of RPs can be estimated according to the following equation:

$$n = kF^{0{\cdot}3}(1 + \Delta H) \tag{15}$$

where F is the basin area (km^2) and ΔH is the difference of altitude in the basin (km). The value k can vary from 0·5 to 1·5 depending on the task, object complexity, landscape heterogeneity, and availability of information (especially meteorological).

System of RFCs

The basin map with the ordered set of RPs is combined with the scheme of RFCs to which the information about most of model parameters is related. The RFC is the part of the river basin which is relatively homogenous regarding topography, geomorphology, geology, pedology, geobotany, ecology. We assume the process of runoff formation to be uniform within the range of one RFC and its quantative characteristics can be averaged. In essence, the RFC is similar to concepts such as the hydrologic response units of PRMS (Leavesley *et al.*, 1983) or the approach taken by the SLURP model (Kite and Kouwen, 1992). The system of RFCs in the basin is the subject for generalization depending on scales of mapping and modelling. It is supposed that all parameters of DMHS defining the RFC in entirety are fixed within its range and change step-wise at RFC borders.

In general, properties within the RFC are assumed to be uniform. However, other characteristics have heterogeneity that cannot be ignored, such as the snow cover.

Heterogeneity of snow cover

The heterogeneity of snow cover depth and consequently of some other of its properties is the first in importance. Snow water equivalent (SWE) for river basins of middle and high latitudes is one of the most important elements in the system of characteristics of the hydrological cycle. It determines not only the possibility of water inflow to the watershed but also governs many quantitative relations of hydrometeorological processes in soil and snow. Snow drifts and blizzards redistribute snow across the territory, filling gullies, narrows, gorges, and crevices. The resulting heterogeneity of SWE should be taken into account.

It is appropriate to assign several additional 'calculating' points (cPs) characterized by its own value of SWE. The calculating points are attached to representative ones (RPs) and do not have exact locations. They refer to any point on the surface and exist only in a statistical way.

The general scheme is the following. For this paper, we used the precipitation sum Y^* for a given RP-area. For accounting of spatial heterogeneity, the spatial distribution of snow is approximated by using typically five quantiles corresponding to the centres of equal intervals at the probability scale: 0·1; 0·3; 0·5; 0·7; 0·9 (the normal distribution law is assumed). If necessary, a sixth calculating point is added to the five 'quantile' ones and it corresponds to the snow accumulation in a system of gullies. As a result, we have

$$Y_1^* = Y^*/[m_1(m_2 - 1) + 1] \tag{16}$$

at the territory that surrounds the gullies after part of the snow has drifted into the system of gullies

$$Y_2^* = m_2 Y_1^* \tag{17}$$

Here, m_1 is the fraction of the RFC area covered with gullies and m_2 is the ratio of the snow depth at gullies and the surrounding territory.

Thus, the snow redistribution, which takes place mainly not actually during the snowfall but afterwards, is imitated simultaneously with snowfall. The layer of solid precipitation at the five quantile points is calculated by multiplying the module coefficients k_p with accepted variation coefficient $C_v(Y^*)$ by Y_1^*

$$Y_p^* = k_p Y_1^* \tag{18}$$

The value of k_p depending on $C_v(H^*)$ is determined by the equation

$$k_p = 1 + U_p C_v(Y^*) \tag{19}$$

where U_p is the quantile of normalized normal distribution. The magnitudes of k_p for some values of $C_v(Y^*)$ are presented in Table III.

The variation coefficient $C_v(Y^*)$ is usually estimated by the materials of snow surveys.

Table III. The values of the coefficient k_p

Interval	The middle of the interval	U_p	Values of k_p for each $C_v(Y^*)$				
			0·1	0·2	0·3	0·4	0·5
0·0–0·2	0·1	−1·282	0·872	0·744	0·615	0·487	0·359
0·2–0·4	0·3	−0·524	0·948	0·895	0·843	0·790	0·738
0·4–0·6	0·5	0	1·000	1·000	1·000	1·000	1·000
0·6–0·8	0·7	0·524	1·052	1·105	1·157	1·210	1·262
0·8–1·0	0·9	1·282	1·128	1·256	1·385	1·513	1·641

THE MODEL PARAMETERS

Any model can be characterized by the set of its parameters. Their list testifies to the extent of the factors governing the runoff formation process which are taken into account. And of course, it almost completely determines the necessary information that should be prepared for any other realization of the model at a given river basin.

The list of DMHS main parameters and the way of their estimation is presented in Table IV. They are divided into groups according to the four elements of a river basin which are defined in conformity with their functional role in the system of the surface hydrological cycle.

All parameters from the first and second groups and related to the three upper underground layers from the third group are individual for each RFC. Other parameters from the third group are determined by more large-scale geological and hydrogeological structures.

Here we do not mention the parameters related to other specific tasks, such as modelling of erosion and pollution process.

MODEL LIMITATIONS AND UNTESTED FEATURES

One potential limitation of the model is the acquisition of the soil profile properties in a format that is suitable for the model. While the information in Russia is directly applicable to the model, in other countries it may prove difficult to translate their soil descriptions to those used by us. In some cases, it was difficult to get the soil profiles in the studied basins.

Another model limitation is its routing scheme, which is not applicable to rivers subject to backwater effects. However, the model can produce the runoff to be input as lateral inflow to fully dynamic models where this is a necessity.

The model is capable of modelling watersheds with additional features, but, to date, they have not been tested. These include watersheds in arid or semi-arid climates; areas where groundwater is very close to the ground; glacially dominated watersheds and watersheds in which lakes, reservoirs, and swamps are an important component; and urban-dominated watersheds.

MODEL IMPLEMENTATION

Study area

In this paper the efficiency of the DMHS is demonstrated by its application to the six watersheds within the Lena River basin. The selected watersheds are shown in Figure 3 and their characteristics are summarized in Table V. The watersheds are not only of different sizes but also contain different landscape characteristics within themselves, as shown in Table V.

The Lena River basin has complex geological and varied relief structures. Mainly, it is a mountainous land. The climate of the study area is strong continental. The absolute temperature range exceeds 100 °C, while mean annual temperature is below zero. The temperature inversions are typical for mountainous conditions during the winter period. Precipitation distribution is extremely irregular as within the basin territory, especially in its mountainous parts, and also during the year. The main amount of precipitation is observed during the warm season. The winter period is prolonged and lasts from 6 to 8 months; it is characterized by hard frosts. Intensive snow melting ensues when mean daily temperature changes to above-freezing. The study basin is situated in the zone of continuous permafrost. The depth of soil thawing fluctuates over a considerable range. The main landscape type is taiga (mainly mountainous), which is dominated by larch.

Forcing data

The DMHS was originally developed in Russia and is prepared for the use of the information available within this country. The simulations for the Lena River basin were conducted for the period 1966–1984, which is characterized by the historical maximum of number of meteorological stations. We used the observational data on daily temperature and precipitation from 210 meteorological stations, 180 of which are located within the basin (Table V). We used available average monthly values of relative humidity for the cold period and daily values for the rest of the year.

The accuracy of precipitation measurements in the northern latitude significantly affects the results of runoff simulations (Tian *et al.*, 2007). The interpolation of the forcing data has become a rather complicated task as, in the study basin, one meteorological station on average covers an area of 12 000 km^2 and more than 50% of

Table IV. List of the model parameters (general recommendations about parameters estimation can be found in Vinogradov (2003b, c, d)

#	Parameters	Way of estimation
	I. Vegetation and surface:	
1–4	Four phenological dates	Characterize the phases of vegetation growth; may be obtained in the literature. Trapezoidal "phenological" approximation is applied in the model.
5–6	Maximum and minimum values of seasonal shadow fraction by vegetation cover	Characterize the changes of parameters within the phases of vegetation growth; may be obtained in literature
7–8	Maximum and minimum interception water capacities	
9–10	Maximum and minimum landscape albedo	
11–12	Maximum and minimum coefficients of potential evaporation	$k = E_0 \cos\alpha / d\Delta t$, where E_0—potential evaporation, d—effective air humidity deficiency, Δt—time period, α is slope inclination. It fluctuates usually in the range of $(0.3 \div 0.6)10^{-8}$ m (mbar s) depending on the type of evaporating surface. For snow cover, water, bare soil, rock-talus complex value can be considered to be constant. For vegetation it can have an annual course in concordance with the parameters 1–4.
13	Coefficient of evaporation from the interception storage during the maximum development of vegetation cover	The same as previous but for conditions of maximum development of vegetation
14–15	Maximum and minimum values of the snow redistribution coefficient	Values may be obtained by analyzing the data of snow surveys
16	Spatial variation coefficient of SWE in snow cover	
17	Spatial variation of infiltration capacity of upper soil layer	May be obtained from the literature for small experimental watersheds
18	Maximum ponding fraction	May be obtained from the literature, visual and aerial photo observations of the basin
19	Maximum surface depression storage	Obtained from the literature; or calibrated (preferably, for small watersheds)
20	Hydraulic parameter of surface runoff elements	$= 10^{-6}$; may need calibration
21	Orographic shadow fraction	Obtained from digital elevation models
	II. Discrete soil strata (unsaturated zone)	
22	Density	Typical values for soil types may be obtained from literature, soil surveys; usually do not require any further calibration
23	Porosity	
24	Maximum water holding capacity	
25	Infiltration coefficient	Values can be obtained from literature; may be calibrated against runoff at small watersheds
26	Specific heat capacity	Typical values may be obtained from literature, soil surveys or estimated by soil texture; do not require any calibration
27	Specific heat conductivity	
28	Index of ice content influence at infiltration	4—sand, 5—loam sand, 6—loam, 7—clay
29	Contribution ratio to evaporation	The contribution ratio of the first soil stratum K_1 changes from 0.1 (deep penetration of vegetation roots) to 0.5 (for sand soils with lack of vegetation). For other soil stratum K_i is calculated as the following $K_i = K_1(1 - K_1)^{i-1}$
30	Hydraulic parameter of soil runoff elements	$= 10^{-6}$; may need calibration
31	Infiltration coefficient from soil stratum to groundwater	Values can be obtained from the literature; geological information about mother rock is important; may need calibration
32–36	Five parameters (average, two phases and two amplitudes) describing temperature at maximum available depth (usually 3.2 m in Russia)	Available in climate reference books or soil surveys; or could be estimated from soil temperature observations

Table IV. (*Continued*)

#	Parameters	Way of estimation
	III. Saturated zone (in case of shallow groundwater)	
37–39	Thickness, porosity and specific water yield coefficient in the groundwater flow area	Can be obtained in literature; or calibrated at small watersheds (this module of the model is being refined)
40–41	Height of capillary raise and index of nonlinearity in the equation of capillary moisture capacity	
	IV. The system of underground runoff elements (specific for each 15 layers of underground water—up to 15)	
42	Hydraulic parameter	$= 10^{-6}$; may need calibration
43	Values of redistribution of water volume among modelling groundwater layers	Need calibration against observed runoff; usually may be easily transferred to the basins in the same conditions without changes; can be systematized for different hydrogeological conditions
	V. Other parameters	
44	Lag time from each RP to the basin outlet	See Section Channel flow and lag time; if there are no observations, initial velocity can be estimated as 1.5 ms^{-1} and then adjusted.

RP, representative point.

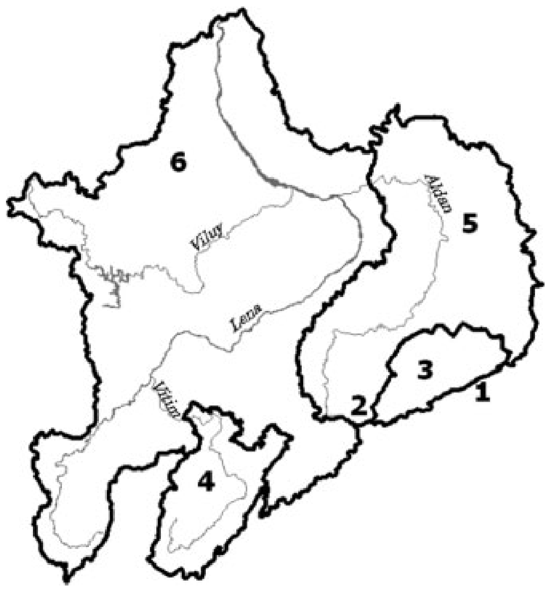

Figure 3. Study area with location of different basins (numbers 1, 2 represent small watersheds)

basin is of complex mountainous relief; here, a simple triangular interpolation of precipitation from meteorological stations into RPs, commonly used in flat areas, was not acceptable. According to Adam *et al*. (2006) the increase of precipitation in the mountain ranges of the Lena River basin can reach 30–40% compared with the observed values in the valley depressions depending on elevation and slope aspect.

The problem of interpolation of precipitation in a mountainous basin was addressed in this research using the procedure of normalizing of daily precipitation by the annual mean values. Main patterns of precipitation dependence on elevation and relief features were described in terms of pluviometric gradients. Curves depicting the dependences of precipitation as a function of elevation were built for each of the defined mountain areas; accordingly, the annual precipitation values for each RP were estimated.

It is known that the proportion of unrecorded precipitation is high for the polar regions. According to Yang and Ohata (2001), the underestimation of precipitation may reach 10–65% of the annual value in the Lena River basin, which is approximately 30–330 mm for the various watersheds. This is due to wind blowing, evaporation, instrument wetting, and other losses. Underestimation of solid precipitation associated with blowing can reach 50–100% in the northern latitudes (Tian *et al*., 2007). As such, the systematic underestimation of solid precipitation in the winter period may result in total losses of the spring flood runoff in May–July up to 5–25% of the annual runoff.

Therefore, we introduced separate areal correction coefficients for solid and liquid precipitation, for each specific sub-basin. On an average, correction coefficients were 1·2 for solid and 1·1 for liquid precipitation; they are in good agreement with the results obtained by Yang *et al*. (2005).

Initial conditions

Precise knowledge of initial conditions may be of great importance to correctly model the basin response at the storm event scale (Noto *et al*., 2008). In the long-term response within continuous simulations, especially for the large-scale basins, the impact of the initial conditions seems to be not so considerable after a 1-year simulation warm-up period.

In the DMHS, the following initial conditions are necessary to start the modelling procedure: antecedent soil wetness conditions (amount of liquid water and ice in every discrete soil stratum); temperature of each soil stratum; state of snow cover (SWE, density, temperature, and saturation index); and amount of water in each of layers of groundwater runoff elements.

Table V. Description of watersheds used in the study: the basin number is coordinated with Figure 3

No	River—outlet	Watershed area (km^2)	Mean daily discharge (m^3 s^{-1})	Watershed features	Number of RP	Number of M (M1)
1	Katyryk—Toko	40·2	0·50	Alpine taiga	4	1 (1)
2	Timpton—Nagorny	613	10·0	Alpine taiga	16	1 (1)
3	Uchur—Chyulbu	108 000	1252	Alpine taiga and mires	49	9 (3)
4	Vitim—Bodaybo	186 000	1660	Alpine tundra, shrub tundra, taiga	71	43 (28)
5	Aldan—Verkhoyansky Perevoz	696 000	5590	Alpine stone talus, alpine tundra, cedar shrubs, taiga	69	40 (26)
6	Lena—Kusur	2 430 000	16 660	Giant watershed, different climatic and landscape conditions	128	212 (180)

RPs, representative points; M, used meteorological stations; M1, within the basin area.

As a rule, we start the simulations during the snow-free period of the year, preferably in autumn, during the flow recession stage. That way, the snow cover state variables may be set to zero. The soil wetness is set to the value of maximum water-holding capacity (or a half of it). Depending on the studied area and the depth of individual soil strata, the moisture can be set as liquid or solid (ice). Temperature of soil strata is estimated accordingly to known annual dynamics. Our modelling practice shows that those initial conditions do not play much of a role for the whole period of simulations and there is not much of a problem in coming back and refining them (e.g., set up the soil completely frozen if the observed spring melt flow identified those conditions).

When the parameter of redistribution of water volume among modelling groundwater layers (Table IV) is estimated for the basin, the initial volume of water storage in each layer can be evaluated according to the system of runoff elements. This initial condition has considerably more impact on the results of flow simulations on the large-scale basins as the recession time of water volumes stored in deeper layers may be large.

Calibration, validation, and adjustment of the parameters

A split sample technique is the most common approach used for the calibration and verification of hydrological models. During the calibration stage, the model parameters are optimized based on the evaluation of the discrepancy between the simulated and observed hydrological characteristics. At the following validation stage, the modelling procedure is conducted with the use of calibrated parameters but for a period different from the calibration one.

While using the DMHS model we apply a different approach. Instead of the 'calibration' term, we use a manual 'adjustment' concept which can be applied to a small group of specially chosen parameters. Manual adjustment (or correction) may be carried out at *a priori* defined narrow ranges of parameter variation based on comparison of observed and simulated values of flow, soil temperature, and moisture; SWE; active layer thickness; and other available information. Usually the period of 1–3 years turns out to be enough to adjust the model parameters for a specific basin. The simulations for the whole period of available observations are conducted continuously and with the single set of estimated parameters.

In this study the following parameters were manually adjusted, using between 1 and 3 years of data, in a narrow predetermined range away from their *a priori* values while simulating the specified period:

1. Evapotranspiration coefficients, coefficient of solar radiation influence on effective air temperature, maximum water storage capacity, and infiltration coefficient of different soil layers were corrected at small basins and then used for large-scale ones without modification.
2. Indexes of incoming water content distribution between modelled groundwater layers were defined for each basin separately but in accordance with the theory of runoff elements, which describes the hierarchical sequence of underground runoff elements layers participating in runoff contribution, assuming that rates of inflow decrease and water store increase with depth (Vinogradov, 1988). In this way, the DMHS is able to divide groundwater flow into components related to different underground water layers. Of course, they are still conceptual model-specific components and may not be directly related to specific flow aquifers and paths, but, rather, they reflect a hypothetical view of groundwater hydrology that does not contradict available observational information. On the other hand, the validity of the widely accepted physically based equations for describing water movement in the saturated zone in conditions of complete lack of information for the large-scale basins seems to us to be unrealistic.

RESULTS

The DMHS was applied at six nested watersheds of the Lena River (Table V). Continuous simulations of runoff

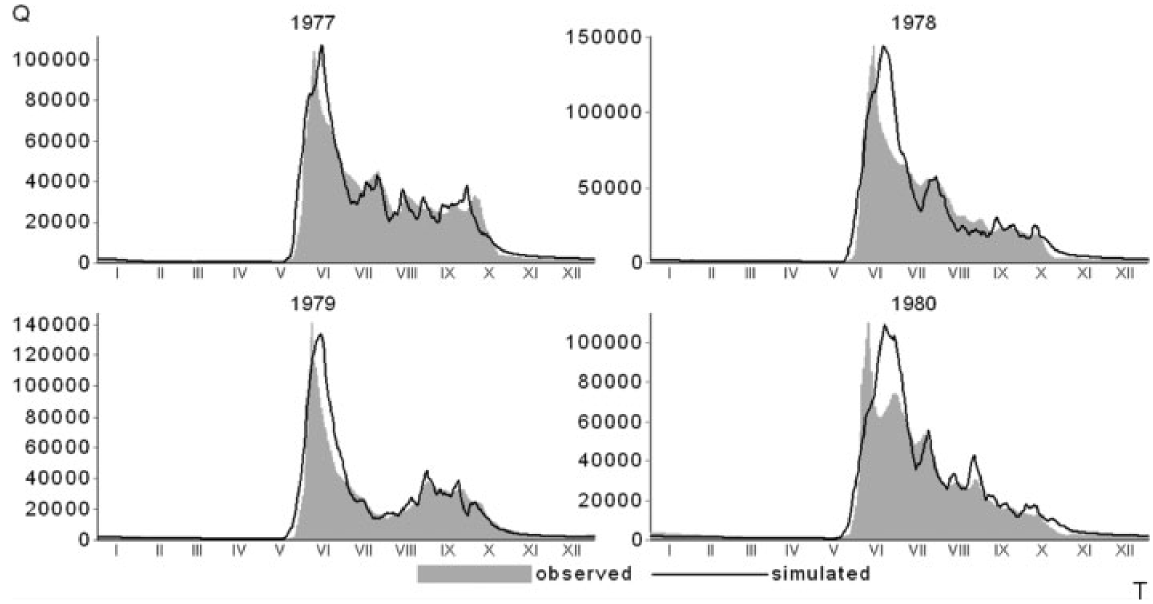

Figure 4. Observed and simulated hydrographs (m^3 s^{-1}), Lena at Kusur, 1977–1980

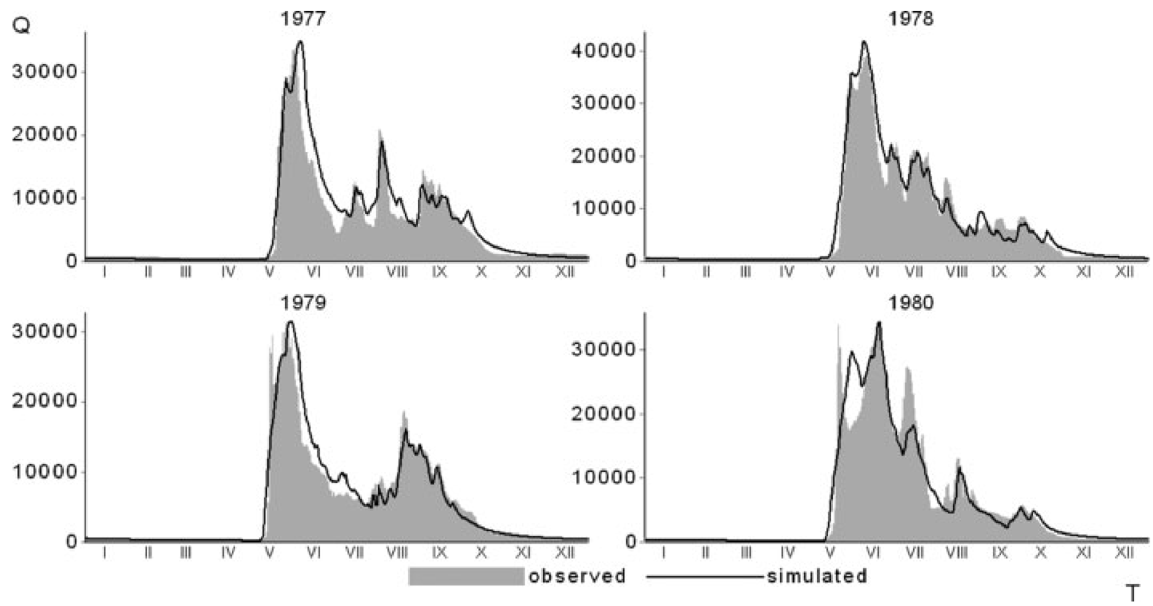

Figure 5. Observed and simulated hydrographs (m^3 s^{-1}), Aldan at Verkhoansky Perevoz, 1977–1980

formation processes with 24-h time steps were performed for the different watersheds for periods ranging between 8 and 19 years, starting in 1966 for the Timpton River, and all simulations ending in 1984. The comparison of the simulated and measured hydrographs for studied watersheds is shown in Figures 4–9. Table VI illustrates the values of statistical characteristics for observed *versus* simulated daily and annual flow layers for six basins (simulation period of 10 years). In general, daily Nash–Sutcliffe efficiencies (Ef) exceed 0·60, and for 4 it exceeds 0·8. The average relative error in its absolute value (Er) for the annual flow simulations is within the 5–10% range.

During the implementation period, the values of annual water balance were calculated as the following: total precipitation amounted to 465 mm for the entire Lena River basin, 538 mm—Aldan River watershed, 550 mm—Vitim River watershed, 600 mm—Uchur River watershed, 700 mm—Timpton River watershed, and 675 mm for the Katyryk basin, taking into account the correction factors introduced to solid and liquid precipitation. Total evaporation from the surface of the studied basin ranged from 150 to 180 mm in the northern part, exceeding the value of 250 mm in the south-eastern mountainous regions, which is about 30–50% of the total precipitation. The values of water balance are consistent with Ma *et al*. (1998) and Yang *et al*. (2002).

The Lena River basin is the largest watershed in our study. Observed and calculated runoff hydrographs have a shape of the Eastern Siberian type which

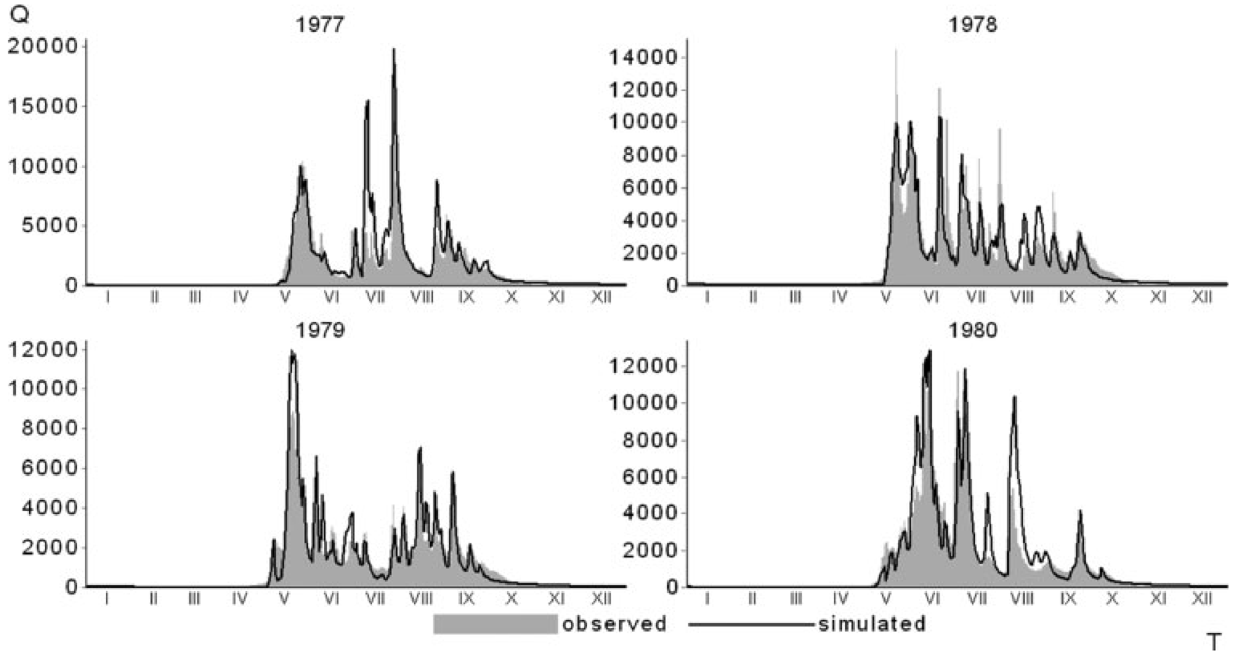

Figure 6. Observed and simulated hydrographs ($m^3\ s^{-1}$), Vitim at Bodaybo, 1977–1980

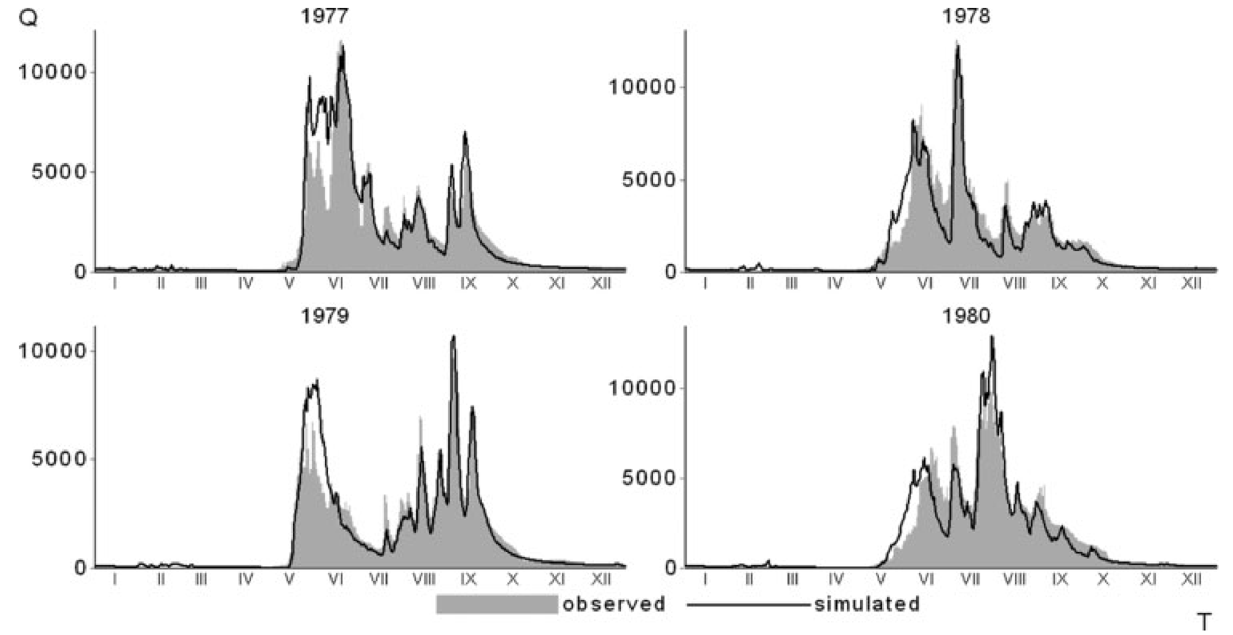

Figure 7. Observed and simulated hydrographs ($m^3\ s^{-1}$), Uchur at Chyulbu, 1977–1980

is characterized by pronounced period of snowmelt with a maximum runoff volume, rain floods in the summer-autumn period, and extremely low flow in winter (Figure 4). Average Ef for the studied period amounted to 0·84 for daily and 0·96 for annual values. Average daily Er is 34% within the range of 23–48%; for annual values average Er is about 7% with the amplitude of 6–8%. Thus, the results of the calculations demonstrate sustainable high consistence with the observed dynamics of the annual runoff.

The calculated and observed hydrographs of the Lena River have a timing mismatch and some discrepancy in terms of runoff volume. Differences in the magnitude of the annual runoff reach 2–20 mm (which is equivalent to 1–4% of annual precipitation), and during the calculation period, except in 1977, there has been general overestimation of flow, although it is not considerable. The largest mismatches with the measured values are observed within the period of snowmelt. Calculated hydrographs have a more 'smoothed' shape of the peaks compared with the observed ones. Although, in every year in the simulation results, the flow rise due to snowmelt appears earlier than the corresponding observations, their maximum values are shifted to a later date, as compared with those measured. The delay amounts up to 15 days in 1981 with an average value of 9 days. In general, the

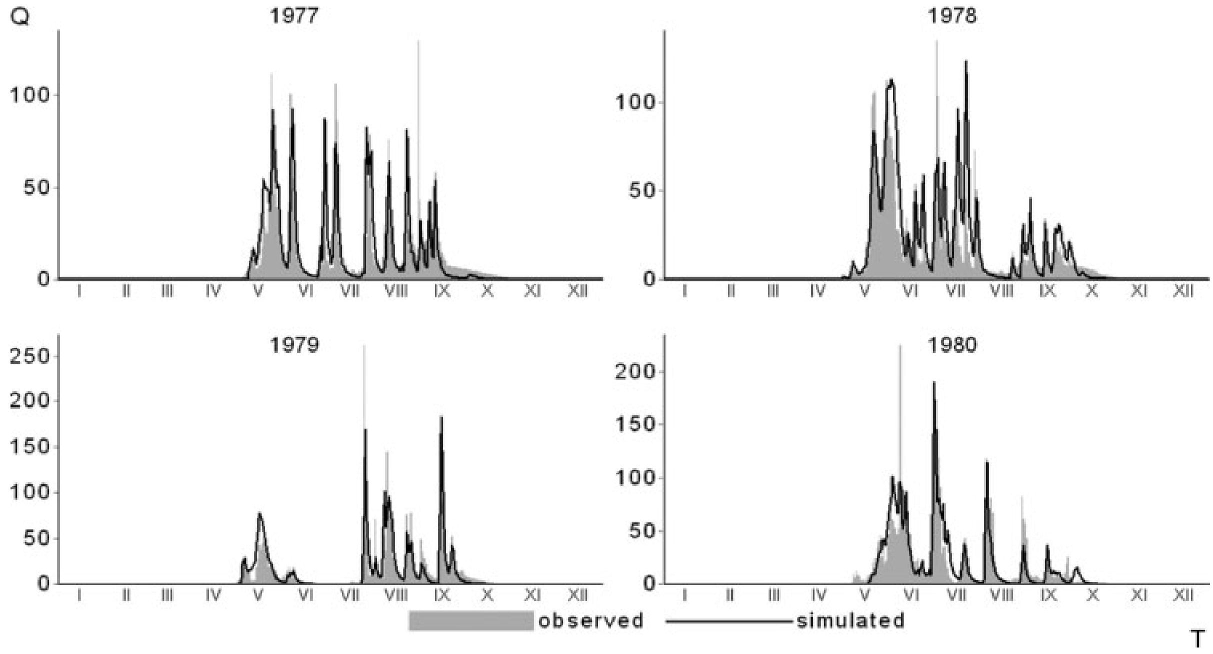

Figure 8. Observed and simulated hydrographs ($m^3 s^{-1}$), Timpton at Nagorny, 1977–1980

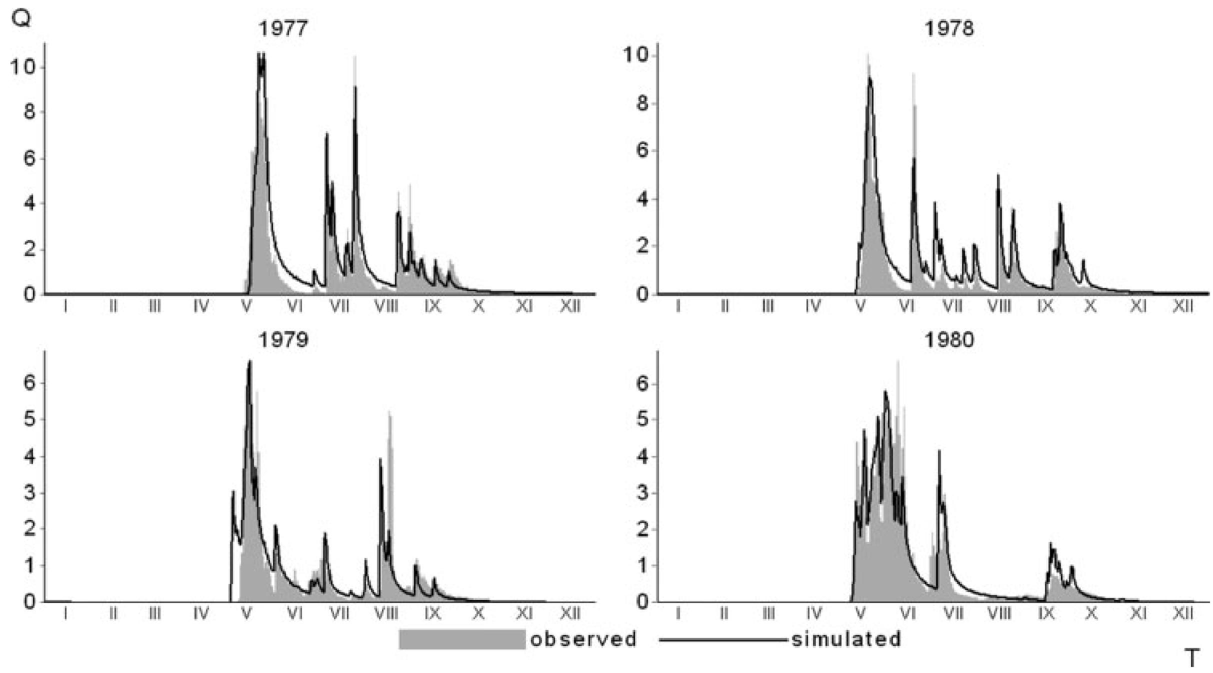

Figure 9. Observed and simulated hydrographs ($m^3 s^{-1}$), Katyryk at Toko, 1977–1980

total volume of snowmelt runoff is overestimated with the underestimation of maximum flows up to 0·7–17·4% (with absolute value of 19 000 in 1983 $m^3 s^{-1}$).

Similar behaviour of calculated and observed runoff hydrographs during snowmelt is observed in the Aldan River basin (Figure 5), e.g. in 1980 and 1984, but to a less degree than for the Lena River. For smaller basins, the discrepancies occur mainly not in the timing of the rise and fall of the branches of flooding but in the volume of the peaks.

Simulated values of low flow occurring from November to May are lower than the observed ones. These play a minor role in the formation of annual runoff volume, but considerably affects the daily values of Er. The difference between observed and calculated flows during winter amounts up to 700 $m^3 s^{-1}$, which is about 50% in relative units. This is due to the almost complete lack of information about groundwater such as the number of layers, their depths, and timing and discharge rates.

The Aldan River flow regime also belongs to the Eastern Siberian type (Figure 5). The simulation results are more precise than for the Lena River. Thus, daily Er is about 26%, annual—5%. Daily Ef amounts to 0·90 and 0·93 for the annual values, which is the highest among

Table VI. Statistical characteristics of runoff simulations in the Lena River basin watersheds (daily/annual)

Basin	Period	R_{obs} (mm)	R_{calc} (mm)	Δ (mm)	Δ_{abs} (mm)	RMSE	Ef	r	Er, %		
									Mean	Minimum	Maximum
Lena at Kusur	1977–1984	0·67	0·71	0·04	0·16	0·32	0·84	0·94	34	23	48
		246	260	14	15	20	0·96	0·99	7	6	8
Aldan at Verhoyansky Perevoz	1970–1984	0·71	0·66	−0·05	0·26	0·33	0·90	0·95	26	16	43
		256	260	4	12	29	0·93	0·97	5	1	15
Vitim at Bodaybo	1968–1984	0·77	0·77	0·00	0·20	0·35	0·84	0·93	31	23	41
		282	280	−2	16	69	0·93	0·98	9	6	15
Uchur at Chyulbu	1977–1984	1·04	1·05	0·01	0·32	0·64	0·81	0·93	36	27	53
		379	381	2	30	49	0·95	0·98	8	2	14
Timpton at Nagorny	1966–1984	1·38	1·42	0·04	0·74	1·76	0·66	0·85	49	31	80
		503	516	13	36	101	0·85	0·94	9	1	26
Katyryk at Toko	1974–1984	1·15	1·05	−0·10	0·54	1·24	0·64	0·85	38	26	50
		422	388	−36	38	79	0·88	0·97	10	1	24

R_{obs}, observed runoff value; R_{calc}, calculated runoff value; Δ, average deviation ($R_{calc} - R_{obs}$); Δ_{abs}, average deviation in the absolute value calculated as $\Delta_{abs} = (\sum_{i=1}^{n} |R_{calc}^{i} - R_{obs}^{i}|)/n$, where R_{calc}^{i} and R_{obs}^{i} are the calculated and observed runoff at day (or year) i for daily (or annual) values; n is the number of the days in the year (or the number of years); RMSE, mean square deviation; Ef, Nash-Sutcliffe Efficiency; r, correlation coefficient; Er, relative error in its absolute value.

all basins. In average, annual flow is underestimated by 4 mm, with a maximum of 12 mm. We associate the better accuracy of simulations for the Aldan River with the positive impact of the estimation of the parameters for the small and medium size watersheds (Katyryk, Timpton, and Uchur rivers) located in the basin, and used in the larger basin.

The Vitim River is a large basin where mountains occupy almost the entire territory. Its average altitude is about 1150 m, reaching almost 3000 m at the Kodarsky range. The flow regime of this river belongs to the Far Eastern type. It is characterized by high rainfall-generated sharp floods during the summer period, which can be merged with the spring snowmelt rise, and often exceed them in magnitude. Flow during the winter period is very low.

The basin has a relatively high availability of meteorological data. We were able to obtain the data of 43 stations, 28 of which are located within the basin. At the same time, a large number of stations (as compared to other basins) are located at altitudes of 1000 m and higher, allowing for a detailed analysis of precipitation distribution within the mountainous ranges.

The results obtained for this basin are as follows: daily Ef is 0·84, annual—0·93; Er—31% and 9%, respectively. The average difference between observed and calculated flow was 2 mm, with its maximum value of 16 mm. Analysis of the results (Figure 6) shows that the timing of peaks is better simulated than their absolute values. We relate it with the difficulties of precipitation measurements and interpolation in mountainous areas.

The Uchur River is a tributary of the Aldan River. The shape of the hydrographs reveals the high intensity of the processes in this mountainous basin. It also has the Far Eastern type of hydrograph with a distinct snowmelt which is almost immediately merged with the runoff from a number of large storm rainfall floods (Figure 7). Accuracy of simulations obtained for this basin is acceptable. Daily Ef value is 0·81 and the annual value is 0·95, while Er reaches in average 36% for daily and 8% for annual flows. The highest discrepancy of observed and calculated flow runoff relates to the period of recession of floods (September–October) and to overestimation of winter flow. Absolute volume difference for the studied period did not exceed 30 mm per year.

The Timpton and Katyryk Rivers, with a catchment area of 613 and 40·2 km^2, respectively, are located within the Aldan River basin. These small watersheds are entirely mountainous, which respond quickly to any intense precipitation (Figures 8 and 9). The data of meteorological stations located at the outlets of the watersheds were used for the simulations.

Calculated flow hydrographs are characterized by discrepancies mainly in the maximum values of flow during floods. Average absolute volume errors (36 mm for Timpton and 38 mm for Katyryk River) exceed those of larger basins.

The Ef for these small watersheds were 0·66 and 0·64 for daily and 0·85 and 0·88 for annual values. Average Er has grown to 49 and 38%, with maximum values of 80% for the Timpton River in 1981 and 50% for the Katyryk River in 1979. Maximum values of Er can be found during the recession of flood peaks in which the measured flows have very steep shape of curves and ups and downs while computed hydrographs are characterized by more gradual flow increase/decrease. This leads to the rise of Er values (up to 200%) with relative low absolute flow values (less than 5–10 $m^3 s^{-1}$ for the Timpton River and less than 0·5 $m^3 s^{-1}$ for the Katyryk River).

DISCUSSION

Although one can see that the results are quite stable for the whole studied period and for basins of different sizes, we find several reasons for the additional important discrepancies between simulations and real flow observations due to model imperfection. Among those reasons are the poor and uneven meteorological data and the difficulty in accounting for the heterogeneity of land cover in small watersheds

Runoff simulation results for small watersheds strongly depend on forcing data and the impossibility to take into account the local precipitation distribution in complex orographic structures.

Systematic and random changes of the characteristics of soil and vegetation are observed within each basin. These systematic changes are taken into consideration by selecting suitable RFCs in the presented model, similar to the HRUs used by Leavesley *et al*. (1983). In such a way, the parameter values are generalized and become representative of a given landscape. The approach is similar to those implemented while developing topographic maps of different scale. It is necessary to use it when area extension should be compensated by reducing the information contents. In hydrological modelling, the level of parameter generalization depends not only on the information available for assessing these parameters but also on the actual need for that level. For large scale, the uncertainties provided by the land heterogeneity are smoothed within the basin territory.

In Semenova (2010) it was shown that the generalization of physically observable parameters may result in a worse match of the observed and simulated flow values for individual, especially small, basins. At the same time, those parameters proved to be reliable for the whole scale range of basins with satisfactory and stable results, and, likely, for all weather and resulting runoff conditions, even those outside of the range of values used for verification. The results of this paper are in consistence with those conclusions.

Examining the Nash–Sutcliffe statistics Ef in Table VI, one can observe that the model performance improves with the watershed size. This result is consistent with the findings by Merz *et al*. (2009), in which they discuss model performance across different scales, although based on a different model. The same pattern was shown for other large-scale Siberian basin of the Kolyma River by Semenova (2010). We associate it with the fact that there is a mutual compensation of uncertainties of influencing factors in the large basins.

Small intercomparison

Much work has been reported on hydrologic modelling of small watersheds in the arctic, but very little work can be found on modelling the Lena or other northern basins of similar size, especially for Russian territory with daily temporal resolution. Therefore, it is difficult to compare the results of this work with that of other researchers. The only example of large-scale hydrologic modelling for some of the studied basins we could find was the work by Su *et al*. (2005), whereby the VIC model was applied over a 100 km EASE-Grid across the pan-Arctic domain. For comparison, Ef calculated for *monthly* values by the VIC model are 0·92 for Lena at Kusur and 0·88 for Aldan at Verkhoyansky Perevoz, the *daily* values for the same basins within the results of DMHS are 0·84 and 0·90.

CONCLUSIONS

How can we prove the proposed approaches and methods? How can we shake the *status quo* with the scaling problem in hydrology? It can be done only by implementing the model across many scales, climates, and landscapes. This topic is well covered in Andréassian *et al*. (2009) and references therein.

The DMHS has proven to be capable of simulating runoff for watersheds of different scales over the Lena River basin, and under different landscapes and meteorological conditions. The simulations time span ranged from 8 to 19 years. The ability to transfer most of the parameters from small to middle and large basins without calibration, low requirements of input meteorological data, and model performance validate the model as a viable alternative to traditional 'physically based' models.

For some additional examples of the model application in Eastern Siberia, see Semenova and Vinogradova (2009); for other applications in different parts of Russia, see Semenova (2010) and Vinogradov and Vinogradova (2010).

On the basis of the obtained results, one can conclude that this model could be an appropriate foundation for climate change impact assessment on hydrological characteristics, particularly in the permafrost areas, and, given the ability to transfer parameters from neighbouring basins, also for applications to some more practical tasks, like simulation of ungauged basins.

As future work, it is important to continue to apply the model to different geographical areas throughout the world, in order to validate our assertions of model universality, and, if necessary, to further refine the model structure and to generalize the parameters. To this end, we are testing the model in the Distributed Model Intercomparison Project-2 (Office of Hydrologic Development, 2010) Oklahoma basins, and will be testing it in the Sierra Nevada basins. We are also applying the model to three separate basins in which the VIC model will be applied. This year, we will test the model against the tRib model (Ivanov *et al*., 2004) in an experimental watershed in Idaho. We will be also testing it against two Canadian models, CRHM (Pomeroy *et al*., 2007) and MESH (Pietroniro *et al*., 2007), at a Yukon river head watershed. We are also applying the model to three Rhine River tributary watersheds.

ACKNOWLEDGEMENTS

We thank the editor of this special issue, Dr Hannah Cloke, Prof. Keith Beven, and two anonymous reviewers for their constructive comments and valuable suggestions. Their input made this paper considerably better.

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