Combined Parameterisation of Orography-Induced Precipitation and Runoff for Regional Hydroclimatic Studies

Andrey B. Shmakin, Andrey Yu. Mikhailov and Sergei A. Bulanov
Institute of Geography, Russian Academy of Sciences,
Staromonetny St., 29, Moscow 109017 Russia.

Abstract

A parameterisation scheme of heat/water transfer on land surfaces, taking into account orographical factors, was used for the plotting of regional maps of the water balance components in the Ganges-Brahmaputra basin. The commonly available sets of data relating to land cover and climatic parameters were drawn upon, and some specific sets were created for the modelled territory as well. The resolutions of initial data in terms of time (1 month) and space (1x1°) are rather rough, as there are no more detailed data for the region. In spite of poor resolution, the main regional features of the water balance components in winter and summer were reproduced satisfactorily. The calculated total runoff from the basin was in the same order as measured in winter and in the same volume as measured in summer. The relief height variability was found to be more important for the regional water cycle and its influence on the climate system than the absolute values.

Introduction

The processes of land-atmosphere interactions are strongly influenced by orography in many aspects. The role of orography is felt by the scientific community to be among the most important factors to be considered in hydroclimatic models and/or calculations.

In studies of energy water exchange on land surfaces, it is necessary to take into account the spatial scales of orographical features most important for the processes. It is more or less well known that the mutual disposition of main mountain ranges (or hill chains) and main valleys of approximate lengths from one to 100km result in the spatial distribution of precipitation fields. The formation of runoff and the secondary reflection and emission of radiation fluxes between relief slopes are influenced mainly by the structure of elementary valleys and water divides measuring from 100m to 10km. The lateral transport of water by rivers is associated mainly with macro-slopes of sizes from 10 to 1,000km. In order to include these processes in hydroclimatic studies, one should: 1) develop special parameterisation schemes describing results of the processes according to external atmospheric/hydrologic conditions and internal land cover properties; 2) have an initial natural data base containing the information about relief features on the given scale.

It must be noted that, when studying rather complicated and mutually dependent processes such as the heat water transformations at the land surface, one usually needs to find a compromise between model requirements and actually available data. In our case. it was necessary to obtain the temporal course of heat-water balance components for a significant territory featuring different natural conditions. There was no detailed land cover and climatic information covering a large mountainous region with spatial and temporal resolution good enough for the physically based modelling. The regular climatic information available via Internet consists of global data sets with 0.5x0.5° spatial resolution at one-month intervals, while the commonly available soil vegetation data sets (e.g., Wilson and Henderson-Sellers 1985) have an even poorer spatial resolution of 1x1°. For calculations, one needs to use resolutions corresponding to the worst ones among all used data sets. Thus, our regional estimates of heat water balance components were made for grid cells of 1x1° at one-month intervals, though the parameterization scheme was developed and tested at much smaller intervals (no more than one day) and smaller land-cover patches. Nevertheless, we obtained quite interesting results even with such poor resolutions due to the unique heterogeneity of both land cover and climatic conditions within the tested region (the whole basin of the Ganges and Brahmaputra). In future, it is planned to improve at least the disadvantages connected with rough temporal resolution by inclusion of the "weather generator" procedure into the scheme.

Calculation Methods and Initial Data

Orographic Precipitation

The orography-induced precipitation was calculated by the model developed by Mikhailov (1985, 1986). Unlike many other methods that take into account the 'altitudinal precipitation gradient', this one obtains the spatial precipitation field from the two-dimensional field of orographical vertical wind velocity as the main governing factor. The vertical wind speed is parameterised as a function of two horizontal wind speed components (such as may be taken from aerological data) and coefficients representing the disposition of spatial ranges and depressions around the given point. It is suggested that the relief features are important for precipitation events if situated at a distance no less than 50km from the point of calculation, and their relative weight in the coefficients depends on the distance. The orographic precipitation intensity may be taken to be equal to:

$$P_{OR}=A\sigma qF(u\iint h(r,\varphi)r\cos\varphi G(r,\sigma)d\varphi dr+v\iint h(r,\varphi)r\sin\varphi G(r,\sigma)d\varphi dr) \qquad (1)$$

where,

A is a coefficient, $\sigma=p/1000$ (p - atmospheric pressure), q is the air specific humidity, u and v are horizontal wind velocity components, and $G(r,\sigma)$ is an influence function with the maximum at some distance between 50 and 150km from the given point. Note that it is enough to calculate the integrals only one time for the given territory.

For the orographic precipitation modelling, we used the ETOPO5 relief data base obtained from UNEP/GRID. It consists of relief heights for the entire globe (including ocean depths) with the spatial resolution of 5x5 minutes made by polynomial interpolation from data measured for 10x10 minutes. In spite of some disadvantages (obvious mistakes in separate areas, height values divisible by some numbers only), this data base provides a good general overview of the disposition of the main ranges and depressions which is useful for calculations of orographic precipitation.

Thus, we obtained a spatial field of climatic orographic precipitation for the part of Eurasia delimited by 23° and 31° north latitude and 77° and 95° east longitude with a five-minute resolution. Then we were forced to average it out over 1x1° grids in order to be able to use the precipitation values in the regional water-heat balance estimations. The result, given such poor spatial resolution, was that our precipitation values were nearly the same as those compiled by Leemans and Cramer (1991). Hence, there is no need to calculate the orographic precipitation by complicated physically-based models with detailed resolution if results are planned to be roughly averaged.

Heat/Water Transformations at the Land Surface

For regional calculations, we used the parameterisation scheme developed earlier (Shmakin et al. 1993), with some modifications. The scheme takes part in the PILPS programme, which provides rich facilities for its testing and calibration (Pitman et al. 1996). In the scheme, all main processes typical of the hydrological cycle in the mountains were taken into account: four mechanisms governing the formation of runoff (Hortonian surface runoff, Dunne's runoff from the contributing areas, subsoil runoff, melt runoff); soil-water transfer between layers; evapotranspiration (considering all components of the energy budget); water transport by rivers; replenishment of soil water storage by rivers in the lowlands. The scheme allows one to reproduce the temporal course of main energy water budget components for a territory with many grid cells occupied by different land-cover types and connected by rivers. Thus, the temporal and spatial variabilities of both land surface features and heat water fluxes are taken into account, though each cell may be thought of as consisting of several spatially homogeneity land-cover types according to the classification of Wilson and Henderson-Sellers (1985).

The hydrological cycle at the land surface was reproduced in the same manner as in Shmakin et al. (1993). We suggest that the relative soil water content of the grid cell is equal to the fraction of the territory occupied by saturated soil (such saturated zones are situated in the valleys and adjacent parts of water divides - so-called 'contributing areas'). In the case of an increase in soil-water content, the saturated fraction 'ascends' the elementary valley slopes, and its square increases. If the soil water content decreases, one witnesses the opposite situation. In bringing the elementary valley morphology into the scheme, we propose the conventional relief image on an appropriate scale. One may postulate that the profile perpendicular to a valley consists of a V-form valley and a water divide which is flat in general (in a rather dissected area, the flat part of a water divide between two valleys may disappear). It is also hypothesised that any given terrain consists of a network of the valleys perpendicular to each other. Almost all types of land relief can be presented under such a conception. As for the numerical characteristics of the given territory, its enough to know three morphometrical parameters of relief dissection: the mean depth of elementary valleys; the mean length of elementary valleys per unit square; and the mean fraction of the unit profile occupied by the flat water divide. For the present work, a data set containing these parameters with the spatial resolution 1x1 degree was created for the main territory of Eurasia (except some peninsulas) by analysis of topographical maps. For the description of water transport by rivers and secondary replenishment of soil water storage by rivers in the lowlands, we also created two data sets of 1x1 degree resolution: the main runoff directions between grid cells and the fraction of incoming river runoff which can penetrate the soil over considerable areas of the cells.

The water balance components are calculated in the scheme by the following procedures. The interception of precipitation by vegetation is parameterised according to Noilhan and Planton (1989):

$$w_i = \min[P * \tau, 0.1 * (LAI+SAI) * \sigma_f];$$
$$P_s = P - wi/\tau \qquad (2)$$

where,

w_i is the layer of intercepted water [mm], P - precipitation intensity, τ - time step, LAI and SAI - leaf area index and the same for stems and branches, σ_f is the fractional vegetation cover at the given grid cell, and P_s - precipitation intensity coming to the soil surface.

The surface runoff from the contributing areas due to the Dunne mechanism is equal to:

$$r_w = P_s * w_1, \text{ if } w_1 > w_2;$$
$$r_w = P_s * w_1 + (w_2 - w_1) * (Z - \theta * d_1/\tau), \text{ if } w_1 < w_2, \qquad (3)$$
$$Z = \min[P_s, K_{si}],$$
$$K_{si} = K_s * (b * \psi_s/d_1 + \cos x),$$

where,

w_1 and w_2 are the relative soil-water contents in the first and second soil layers respectively, θ is soil porosity, d_1 is the depth of the upper soil layer, K_{si} is the soil infiltration capacity (Entekhabi and Eagleson 1989), K_s is the soil hydraulic conductivity at saturation, b is a parameter constant for the given soil texture according to Clapp and Hornberger (1978), ψ_s is the matrix potential at saturation, and x is the steepness of the slope of the elementary basin.

The Hortonian runoff is equal to:

$$r_h = (P_s - K_{si}) * (1 - w_1) \qquad (4)$$

The subsoil runoff is equal to:

$$r_g=(1-L_p) * K_s * w_2^{2b+3}, \quad (5)$$

where,
L_p is the fraction of flat water divide in the typical profile across the elementary valley (it is assumed that subsoil runoff can take place from valleys only).

The increase of soil-water storage due to precipitation can be imagined as the 'ascending' of the saturated fraction of the territory up the slopes of elementary valleys and then as spreading over the flat water divides. We calculate the width of such zones of soil water increase assuming that their volume is equal to the volume of a saturated soil layer filled by the rainfall or melting snow:

$$\begin{aligned} &d_{pr}w_1=h * (1-w_1)/ d_1 \\ &d_{pr}w_2=(h-d_1) * (1-w_1)/ d_2, \text{ if } h>d_1 \quad (6) \\ &h=Z * \tau/\theta, \end{aligned}$$

where,
d_2 is the depth of the second soil layer (usually taken as the root zone depth or 1 m). Of course, the relative soil moisture cannot exceed unity, and all water excess is directed to the runoff.

The decrease of the soil moisture due to evapotranspiration and subsoil runoff is calculated by the same approach assuming that the saturated zones diminish at the water divides or descend by the elementary valley slopes.

The evapotranspiration consists of four components: evaporation of precipitation intercepted by vegetation; evaporation of snow cover (these two processes have the intensity of potential evaporation from a wet surface); and evaporation from the soil and transpiration. The potential evaporation is calculated by an iterative solution of the heat balance equation, and the turbulent transfer coefficient is calculated according to Thom (1975):

$$E_0=(q_s(T_s)-q_a) * u * \rho * f * \kappa^2/\ln^2((z-d_0)/z_0), \quad (7)$$

where,
$q_s(T_s)$ is the saturation specific humidity at the given surface temperature, q_a is the air specific humidity calculated under the assumption that the relative air humidity is equal to the climatic cloudiness fraction (we had no exact information on the climatic air humidity in the studied area), u is wind velocity, ρ is air density, κ is von Karman constant, z is the height of calculations (it was 30 metres in order to calculate the fluxes above the forests; the air temperature was recalculated for this level), d_0 and z_0 are the roughness length and zero plane displacement respectively, and f is the atmospheric stability factor:

$$f=(1-16 * Ri)^{0.75}, \text{ if } Ri<0,$$

$$f=(1-5 * Ri)^2, \text{ if } Ri>0, \quad (8)$$

where,
Ri is the Richardson number.
The evaporation from the soil is equal to:

$$E_s=w_1 * (E_0-E_i-E_{sn}) * (1-\sigma_f), \quad (9)$$

where,
E_i and E_{sn} are the evaporation of intercepted precipitation and snow cover respectively. Their intensity is equal to E_0, but they cannot exceed the interception storage and amount of snow, respectively. Thus, if a snow cover exists ($E_0=E_{sn}$), evaporation from the soil surface is absent.

Transpiration can occur only if the leaf area index (LAI) is positive. Its intensity from both soil layers is linearly proportional to the relative soil water content (parameterisation of the stomatal resistance mechanism is not used):

$$E_{t1}=w_1 * (E_0-E_i-E_{sn}) * L_w * \sigma_f * r_1,$$
$$E_{t2}=w_2 * (E_0-E_i-E_{sn}) * L_w * \sigma_f * (1-r_1), \quad (10)$$
$$L_w=1-\max[0.,w_i-E_i * \tau-0.1 * SAI * \sigma_f]/(0.1 * LAI * \sigma_f),$$

where,
r_1 is the relative fraction of roots in the upper soil layer. The term L_w allows one to set the transpiration from the leaves covered by intercepted precipitation to zero.

The soil water storage can be replenished from incoming rivers:

$$d_r w_2=\Delta * R<(1-w_2), \quad (11)$$

where,
R is total river runoff formed at the given grid cell and coming from the upper cells and Δ is a special coefficient specially defined for each grid cell and equal to the fraction of the river runoff which can penetrate the soil over considerable areas of the cell. The information for the latter data set can be found in Korzoun (1974).

The soil water storage in each layer after each time step is equal to the algebraic sum of the previous soil water content and all its changes during all corresponding processes: infiltration of rainfall and melted snow, evapotranspiration, subsoil runoff, soil water replenishment from rivers, and vertical diffusion between layers (the latter process is assumed to work in such a manner that the relative soil moisture becomes constant vertically).

In the given realisation of the scheme, we also evaluated the solar and infrared radiation fluxes, as well as the radiation fluxes between slopes, by the methods developed earlier (Krenke et al. 1991 Shmakin and Ananicheva 1990), using the climatic cloudiness information as well as the atmospheric transmissivity data at the given altitude according to Pivovarova (1977). The snowmelt (if any) and surface infrared emission are calculated by an iterative solution of the heat balance equation. The soil heat flux is assumed to be proportional to the net radiation, with a coefficient constant for the given land-cover type. The turbulent sensible heat flux is calculated by using the same equation as for potential evaporation, but proportionally to the vertical gradient of the air temperature instead of the air specific humidity.

The Relief Data in the Modelled Territory

The general relief field in the modelled territory is shown in Figure 1. The territory includes the whole basin of the Ganges and Brahmaputra with the steepest Himalayan macroslope and adjacent lowlands. The numbers along the axes represent coordinates by latitude and longitude.

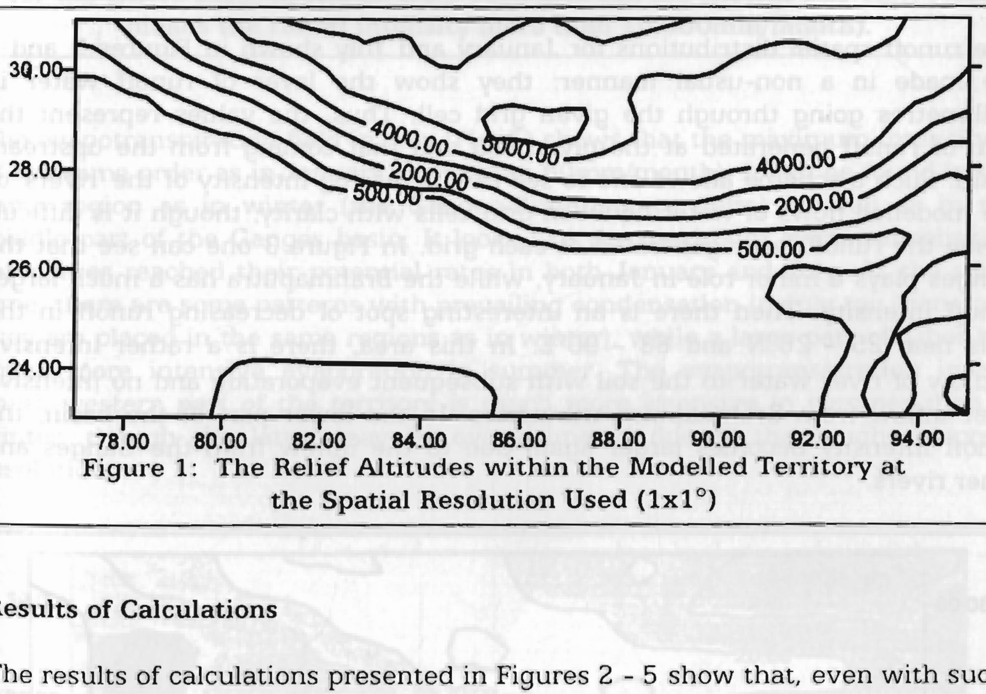

Figure 1: The Relief Altitudes within the Modelled Territory at the Spatial Resolution Used (1x1°)

Results of Calculations

The results of calculations presented in Figures 2 - 5 show that, even with such poor meteorological data (average climatic regime only, without weather changes during a month), rough land cover information (1x1°), and a rather simple calculation scheme, the main features of the distribution of regional water/energy balance components were reproduced quite satisfactorily, and some interesting conclusions can be made. For example, in Figure 2 one can see that, even in high mountains in the middle of winter, weak evapotranspiration prevails, while there are some patterns with condensation. The most intensive evapotranspiration occurs in the Ganges and Brahmaputra valleys, and especially in their delta (due to the water transport from other parts of the basin by the rivers).

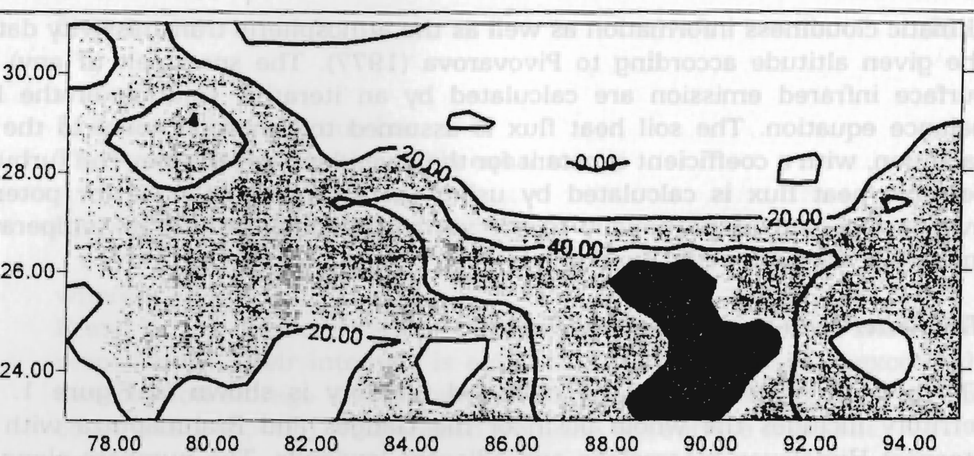

Figure 2: The Evapotranspiration Field in January, mm/Month (The dark colour means evapotranspiration of more than 60mm/month).

The runoff spatial distributions for January and July shown in Figures 3 and 4 are made in a non-usual manner: they show the layer of runoff water in millimetres going through the given grid cell. Thus, the values represent the sum of runoff generated at the given cell and that coming from the upstream basin. Such a scheme allows one to see the increasing intensity of the 'rivers' or the modelled flows of water between grid cells with clarity, though it is difficult to see the runoff layer generated at each grid. In Figure 3 one can see that the Ganges plays a minor role in January, while the Brahmaputra has a much larger runoff intensity. Then there is an interesting spot of decreasing runoff in the cells near 25 - 26°N and 88 - 90°E. In this area, there is a rather intensive outflow of river water to the soil with subsequent evaporation and no intensive river inflow from Brahmaputra tributaries. In the lower part of the basin, the runoff intensity becomes larger again due to the inflow from the Ganges and other rivers.

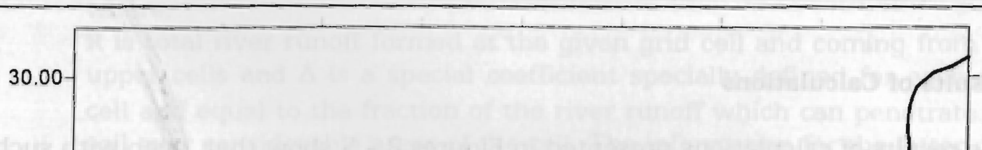

Figure 3: The Runoff Field in January, mm/Month (The accumulated' values for the whole basin upstream from each grid cell are shown).

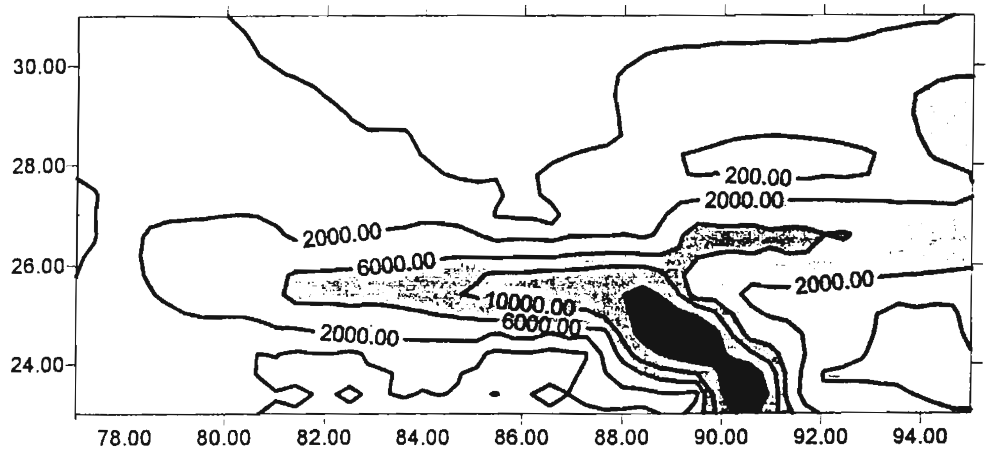

Figure 4: The Runoff field in July, mm/Month (The "accumulated" values for the whole basin upstream from each grid cell are shown. The dark color means the runoff intensity more than 16,000mm/month).

The evapotranspiration field in July (Fig. 5) shows that the maximum intensity is of the same order as in January (more than 60mm/month). It can be found in the same region as in winter (the Ganges-Brahmaputra delta) as well as in the middle part of the Ganges basin. It looks like the maximum evapotranspiration intensities reached their potential rates in both January and July. At the same time, there are some patterns with prevailing condensation in July too (generally they are placed in the same regions as in winter), while a large part of Tibet has much more intensive evaporation in summer. The evapotranspiration in the south-western part of the territory is much more intensive in summer than in winter, though the latter may be overestimated due to the rough temporal resolution.

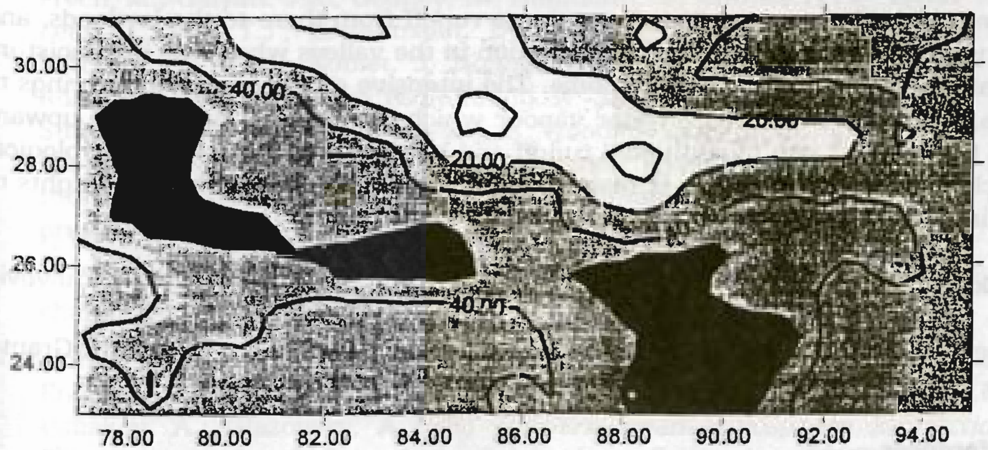

Figure 5: The Evapotranspiration Field in July, mm/Month (The dark colour indicates evapotranspiration of more than 60mm/month).

The runoff field in summer shows that the Ganges plays a more important role in the total runoff from the basin than the Brahmaputra. The runoff intensities are an order more than in winter. Tibet is rather passive in runoff generation, while it gives more water to the rivers in winter due to the melting snow.

The total runoff from the Ganges-Brahmaputra basin calculated by the parameterisation scheme can be verified against the measured runoff data. For such verifying, we should compare the runoff values for more or less the same periods of time. Our time step is one month, so the calculations for all grid cells are carried out with, for example, January meteorological forcing, and the 'runoff wave' moves through all cells which operate with January forcing too, though in reality the lower parts of the basins obtain the runoff brought by rivers nearly a month later than when it was formed in the upper part of the basin. Hence, we compared the runoff calculated for January with the February measurements and so on. The comparison shows that the order of total runoff is calculated accurately, and in summer even the absolute calculated value is very close to the measured one: the calculated monthly runoff volume in January was $38km^3$, while the measured runoff in February was $21km^3$. In July, we calculated the total runoff as $378km^3$, and the measured value in August is $369km^3$. Thus, the model estimations are more or less accurate in spite of such poor temporal and spatial resolutions.

The results also gave us an opportunity to analyse the role of secondary regional water cycles in the climate system in dependence on some relief features. As can be seen from Figures 1-5, the most intensive evapotranspiration is situated not in the low mountain belt where the heaviest precipitation occurs, but in the wide river valleys. The reason is that, in our model, the lateral water transfer by rivers and soil water replenishment from rivers are taken into account, and these are mechanisms which turn regions of high precipitation into ones of high evaporation. For such an effect, the mutual disposition of mountains and adjacent lowlands is necessary. Only with this combination of relief forms can one observe considerably increased precipitation in the first chain of mountains due to significant upward air flows, large runoff from there to the lowlands, and consequently increased evapotranspiration in the valleys where the soil moisture is affected by the river water income. The intensive evapotranspiration brings to the atmosphere additional water vapour which, condensing due to the upward air flows, gives rise to additional runoff and so on. Thus, the role of hydrological cycles in regional climates is mainly connected not with the absolute heights of relief, but with its spatial variability.

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