Some spatial patterns in the water balance structure for the river basins within European Russia

Yuri A. Salo

Northern Water Problems Institute, 50 A. Nevskogo St., 185030 Petrozavodsk, Republic of Karelia, RUSSIA
e-mail: yuris@nwpi.krc.karelia.ru

ABSTRACT
This paper presents the results of estimation of basic annual and mean annual water balance characteristics (precipitation, evapotranspiration, and runoff) in 182 river basins within European Russia and neighbour countries in the period 1950-1985. The aim of the paper is to assess spatial (first of all, latitudinal) variability in the change of water storage, and both of the runoff and the evapotranspiration coefficients, which characterize the water balance structure as a whole. It was found over an annual time scale that a change of water storage in high latitudes is strongly dependent on annual air temperature and, to a lesser extent, on precipitation. Conversely, the annual amount of precipitation is the main factor controlling the change of water storage in southern, relatively dry regions. It is shown that the change of water storage is more temperature-sensitive in northern river basins than in southern ones. Also, a semi-empirical curve and a simple formula are suggested to estimate typical zonal ("standard") values of the runoff coefficient as a function of the dryness index within the study area over a long-term scale.

KEYWORDS
River basin; European Russia; water balance structure; change in water storage; runoff coefficient; evapotraspiration coefficient.

1. INTRODUCTION

The water balance structure of river basins depends on both zonal and azonal factors. The main zonal factor is the geographic location (or – more specifically – the latitude) of a catchment. In general, the incoming solar radiation, surface albedo, mean annual temperature, potential evaporation, and evapotranspiration decrease with increasing latitude. Conversely, there is a noticeable downward latitudinal trend in mean annual river runoff characteristics from north to south. The effect of azonal features (topography, sea-level elevation, lake coverage, forest coverage of a catchment, etc.) results in essential deviation of major climate and water balance elements from their common tendencies in a given natural zone.

The problem of both zonality and azonality in the context of variability of the water balance structure has been discussed and studied in detail by Bulavko (1971), Kuzin (1973), Babkin and Vuglinsky (1982). Principal spatial patterns in heat and moistening regimes of the territory of Russia as a whole and the factors behind these patterns were described, in particular, by Budyko (1974), and Zhakov (1982).

It has been recently found for several river basins located in the northern taiga zone (Republic of Karelia and the Kola Peninsula, Russia) that annual changes in water storage within a basin depend both on the area-averaged annual air temperature and on the annual amount of precipitation (Salo, 2005). Therefore, in the relatively cold and humid northern regions, the air temperature is the main control of water accumulation within a catchment. It has be noted also that in these conditions, the correlation between the air temperature and the change in water storage has a negative sign, and absolute values of the correlation coefficient reach 0.7 or more.

Thence, one of the goals of the present study is to assess the latitudinal (i.e. zonal) differences in the degree to which air temperature and precipitation affect changes in water storage over an annual time scale within European Russia (ER). Traditionally, the water balance structure over a long-term scale is defined through both the runoff coefficient and the evapotranspiration coefficient, which characterize the relative shares of river runoff and evaporation from the drainage area in the water budget. It was therefore interesting to estimate the runoff coefficient as a function of the Budyko's dryness index within the studied territory, where the dryness index varies from 0.34 in the far north to 2.93 in the south of ER (Budyko, 1974).

2. DATA AND METHODS

Data from instrumental measurements and calculations of basic annual climate and water balance characteristics (air temperature, precipitation, evapotranspiration, and runoff) in 162 river basins evenly distributed over European Russia were used in this study. In addition, twenty basins located in the Baltic states, Belarus and Ukraine were also included in the analysis. All these 182 catchments are situated in various physiographic conditions (Figure 1).

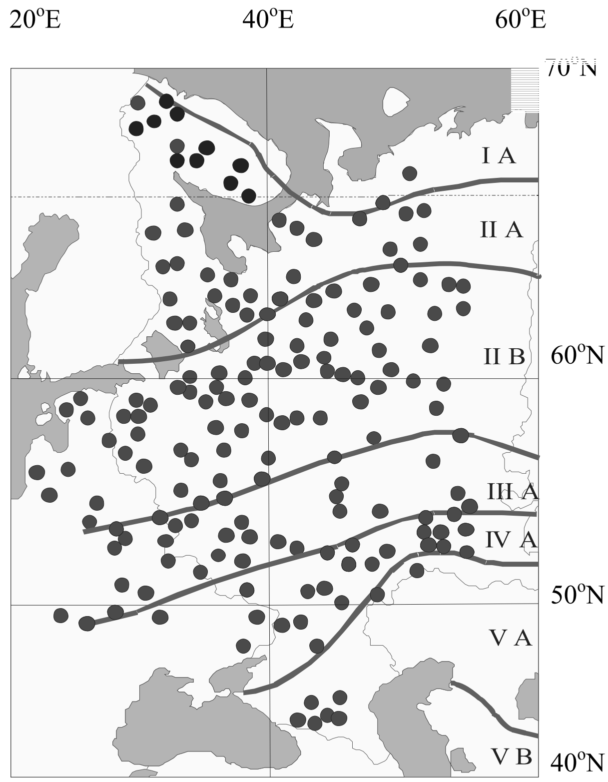

Figure 1. Map of the study area and locations of the investigated catchments.

According to Kaminski's climatic classification (Drozdov *et al.*, 1989), they are as follows: tundra (marked as I A in Figure 1), paludified taiga (II A), central forest zone (II B), forest-steppe zone (III A), chernozem belt (IV A), semi-desert (V A), and desert zone (V B). The studied rivers vary from 1,000 to 20,000 km^2 in drainage area, and from 17 to 533 mm in annual runoff. Mean annual air temperature in the catchments ranges from -2.9 to 11.1°C, and the precipitation norm – from 350 to 854 mm per year. For each of basin, data for the same 36-year period (1950-1986) were taken into consideration.

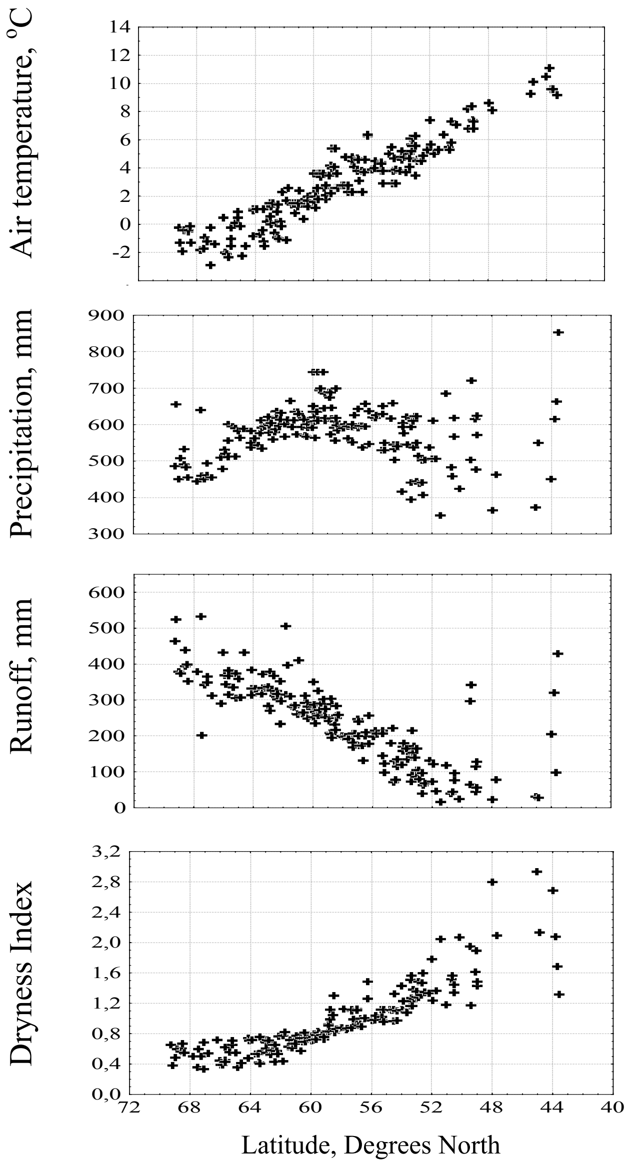

Figure 2. Average annual air temperature, precipitation, total runoff, and dryness index as a function of the latitude of the studied river basins.

As seen from Figure 2, there is a clear tendency in the change of basic climate and water balance elements from zone to zone of the study area, except for mountain and semi-mountain catchments (foothills of the Caucasus far in the south, below 45°N, and the Ural Mountains in the east of ER) due to the strong effect of elevation a.s.l. and to peculiarities in the formation of the hydrological regime in mountainous areas.
In this study, the water balance equation over an annual time scale is drawn as follows:

$$P - E - R \pm W \pm \varepsilon = 0, \quad (1)$$

where P is precipitation; E is evapotranspiration; R is total runoff (sum of surface and subsurface runoff); W is change in both surface and subsurface storage; and ε is the integral component including the balance

discrepancy, error of each water balance element measured or estimated, and error due to unknown elements not included in Equation (1). All values are in mm (10^{-3} m).
The formula after Ol'decop for estimation of evapotranspiration E is used as follows:

$$E=E_o tanh(P/E_o), \quad (2)$$

where $tanh(P/E_o)$ is the hyperbolic tangent function of the ratio of precipitation to potential evaporation E_o. The latter value has been expressed as an empirical formula $E_o=E_o(T)$, as follows (Salo, 2007):

$$E_o=329 +62T+2.14T^2, \quad (3)$$

where T is mean area-averaged annual air temperature in a given basin in a given year, ^oC.
The ratio of potential evaporation, E_o to precipitation, P commonly known as the aridity or dryness index (φ) after Budyko (1974). Regions with $\varphi>1$ are broadly classified as dry since the evaporative demand cannot be met by precipitation. Conversely, similarly regions with $\varphi<1$ are classified as wet. The index may also be related to climatic regimes in a broad sense, e.g. arid, semi-arid, sub-humid, and humid regions are defined by the aridity index ranges of $12>\varphi>5$; $12>\varphi>5$; $5>\varphi>2$; $1> \varphi>0.75$; and $\varphi <0.75$, respectively (Arora, 2002).
In our study, the Budyko's formula, $\varphi = E_o/P$, was used as the basic formula to estimate zone-to-zone changes in the water balance structure.

3. WATER BALANCE STRUCTURE FOR CATCHMENTS WITHIN ER

First, the correlation of the change in water storage, W with annual air temperature, T and precipitation, P was estimated for each of the 182 basins for the same period of 1950-1986. Herewith, it was assumed that value ε in Equation 1 is included in W and its contribution to the sum ($W+\varepsilon$) is negligible.
Figure 3 illustrates latitudinal regularity in the coefficient of correlation between W and T (Figure 3a) and between W and P (Figure 3b).
The coefficient of correlation between W and T is negative for all basins tested and r(W, T) reaches up to -0.8 for the catchments located in the tundra and paludified taiga zones. It can be concluded that in high latitudes, annual air temperature exerts a stronger influence on the accumulation/discharge of water in catchments, and accounts for up to 65 per cent of the variability in W, while in southern latitudes – for not more than 10 per cent.
There is no clear regularity in the common latitudinal tendency in the correlation between the change in water storage and annual precipitation (Figure 3b). We can only estimate that the correlation between W and P generally increases from north to south, and that precipitation influences changes in water storage more noticeably in the steppe and semi-desert zones in comparison with the tundra and forest regions.
The relationship between the coefficients of determination $r^2(W, T)$ and $r^2(W, P)$, and the dryness index $\varphi = E_o/P$ has also been studied. As one can see from Figure 3c, the leading climate characteristic influencing W in wet regions ($\varphi<1$) is air temperature. As the dryness index decreases, the influence of T upon W lessens nonlinearly, so that for semi-arid regions ($\varphi>2$) the effect becomes insignificant. Taking a common tendency for an increase in $r^2(W, P)$ with increasing index φ (Figure 3d) into consideration, we can conclude that the effect of precipitation on the change of water storage in wet regions is at least not so strong as in semi-arid and arid regions.
Secondly, water balance structure was studied in relation to the dryness index. Over a mean annual time scale, the water balance equation can be written in a non-dimensional form as follows:

$$K_R + K_E = 1, \quad (4)$$

where K_R is the runoff coefficient, K_E is the coefficient of evapotranspiration. All values in Equation 4 are calculated as annual means.

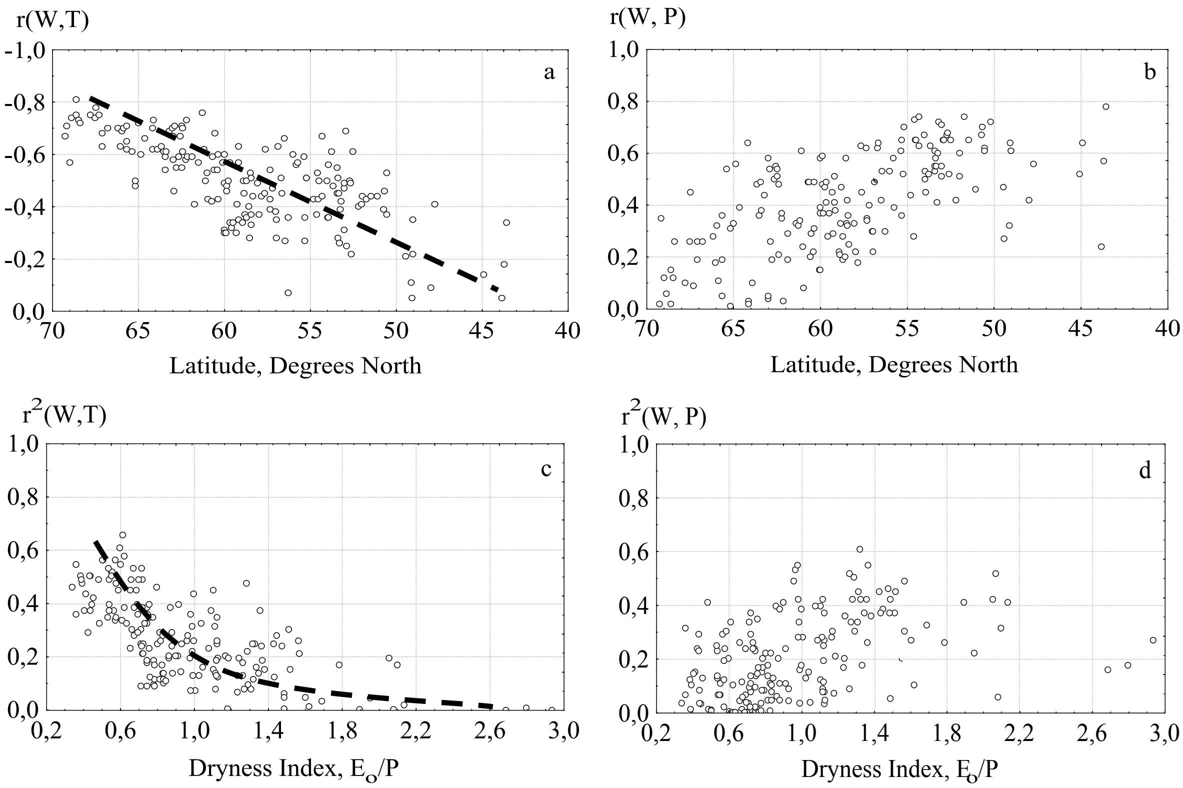

Figure 3. Coefficients of correlation (r) and determination (r^2) of water storage W with annual air temperature and precipitation as functions of the latitude and the dryness index.
and total runoff as a function of latitude of studied river basins.

Two variants of K_R and K_E calculations were carried out. In the first case, initial precipitation, P (corrected only by wetting and evaporation losses) was used to calculate these coefficients as K_R=R/P and K_E = E/P, respectively. In the second case, precipitation was corrected for each of the 182 river basins using the water balance approach and calculation scheme proposed by Salo (2005, 2006). The advantage of this method is that it enables finding mutual conformity between corrected precipitation P_c, runoff R, and evapotranspiration E_c calculated using P_c, in accordance with the trinomial water balance equation:

$$P_c - R - E_c = 1, \qquad (5)$$

$$K_R^* = R/P_c, \quad K_E^* = E_c/P_c,$$

$$K_R^* + K_E^* = 1. \qquad (6)$$

Thus, in the second case, the coefficients K_R^* and K_E^*, too, were considered in this study.
Figure 4 demonstrates a non-linear decrease of runoff coefficients in relation to reduction of the dryness index in these two cases. One can see from the left-hand graph that for a dryness index of $\varphi<0.8$, there is a noticable rise in K_R values. On the contrary, where corrected precipitation values have been taken into account, there is full functional correspondence between the runoff coefficient and the dryness index. It is important that the curve in Figure 4b agrees well with a theoretical limit $K_R \rightarrow 1$ under $E_o/P \rightarrow 0$.

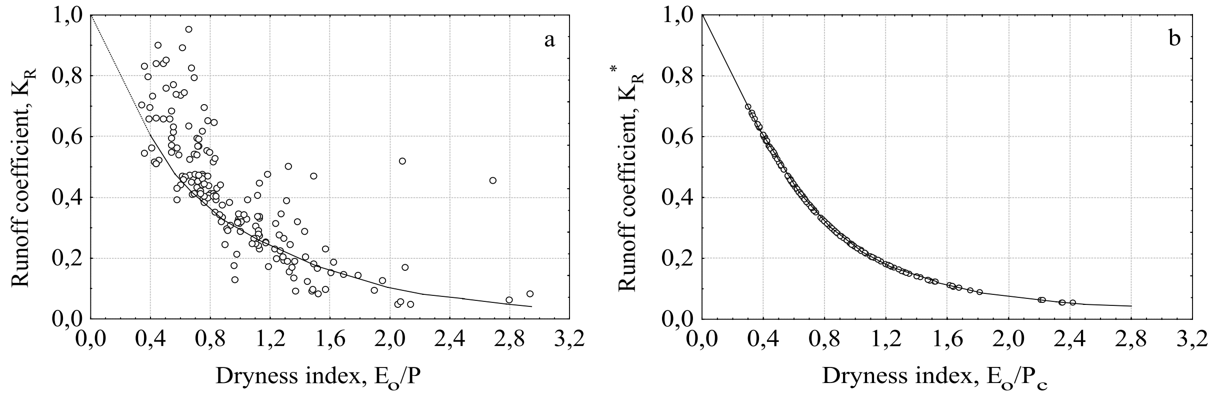

Figure 4. Graphs of the dependence of runoff coefficients $K_R = K_R(E_o/P)$ and $K_R^* = K_R^*(E_o/P_c)$.

Thus, the curve plotted in Figure 4b can be considered to be semi-empirical, suitable for estimating typical (or "standard") values of the runoff coefficient as a function of the dryness index in ER. In the analitical form, the dependence can be presented as follows:

$$K_R = \exp(-1.4\cdot\varphi), \tag{7}$$

where $\varphi = E_o/P_c$ is the Budyko's dryness index calculated using corrected precipitation P_c.

By analogy with Figure 4b and taking Equation 6 into account, the semi-empirical curve $K_E^* = K_E^*(E_o/P_c)$ can also be plotted easily.

We can assume that any (whether positive or negative) deviation of actual runoff coefficients from"standard" ones would be due to the effect of local (i.e. azonal) factors, in particular, to the percent cover of lakes and mires in a catchment. This assumption, in turn, requires an additional study.

4. CONCLUSIONS

A key finding of this study is that the difference in the degree of effect of annual air temperature and precipitation on water storage has been detected and estimated. In high latitudes, air temperature is the leading climate parameter controlling accumulation/discharge of water in catchments, whereas annual precipitation plays the second part. Conversely, in the relatively dry arid and semi-arid regions, precipitation affects annual changes in water storage more than air temperature.

Consequently, water storage is more temperature-sensitive in northern river basins than in southern ones. In general, this conclusion agrees well with estimations of catchment's sensitivity to present-day climatic conditions as well as to possible climate change (Vinnikov, 1986; Kovalenko, 1993; Bobylev *et al.*, 2003).

Also, a semi-empirical curve and a simple formula of the relationship between the runoff coefficient and the Budyko's dryness index have been obtained. They provide a possibility to estimate typical zonal values of the runoff coefficient as a function of the dryness index within the study area over a long-term scale.

The results reported above relate to European Russia, which is characterized, in general, as a relatively flat territory. Applicability of the results to mountain and semi-mountain regions is uncertain and requires that the characteristics typical of these regions (sea-level elevation, stream gradient, mean watershed slope, etc.) are included into consideration. Probably, other formulae would be more suitable for calculating potential evaporation and evapotranspiration in mountainous catchments.

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