

Sensitivity of runoff due to changes in the characteristics of the water balance in the Danube River region

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Abstract: Climate change is presently a widely discussed subject in relation to alterations in water storage capacity and the components of the hydrological balance within catchment areas. This research study was directed at two main objectives: 1. The indirect estimation of long-term mean annual runoff using an empirical model; 2. The determination of changes in the annual runoff regime of fifty Danube sub-basins. Monthly areal precipitation, discharges, and air temperature data from 1961 to 1990 were collected for selected headwater sub-basins of the Danube River. In the first part, Turc-type empirical equations for the estimation of the long-term average annual runoff R in the Danube basin were employed. The parameters of the empirical equations were determined through nonlinear regression. Given the underestimation of the actual (territorial, balance) evapotranspiration ET values determined from the balance equation, the precipitation totals were corrected by +10%. With a 10% increase in precipitation, the values of balance ET reached the values ET determined by the Budyko–Zubenok–Konstantinov method. In the second part, fifty equations for the estimation of changes in the average annual runoff, depending on increases in the air temperature and changes in the annual precipitation separately for each of the 50 sub-basins, were established. In conclusion, the results suggest that, on average, a 100 mm increase in the average annual rainfall in the Danube River headwater sub-basins, will cause a 50 mm increase in outflow, and a 1 °C increase in the average annual air temperature will lead to a 12 mm decrease in runoff.

Keywords: Danube River basin; Hydrological balance; Runoff; Climate change; Precipitation correction.

1 INTRODUCTION

The increased impact of anthropogenic activity on the climate of the Earth and the hydrological cycle have been observed since the middle of the last century. However, the strong variability of natural climates makes it difficult to estimate the human contribution to climate change and the hydrological cycle (Abbott et al., 2019). Direct anthropogenic effects include, e.g., the abstraction of groundwater or surface water or their redistribution. Indirect impacts caused by the effects of warming and regional human activities, such as changes in land use and land cover, can sometimes have more hidden consequences (Bonfils et al., 2020; Padrón et al., 2020). Changes in the behaviour of hydrological cycle components have also been detected in recent decades. Climate change in Europe is often associated with rising air temperatures, which shorten the snow cover period in mountainous regions and northern latitudes. Locally, there could be more rain and less snow in a warmer future in mountainous areas, which could severely affect the seasonal availability of water for users and the environment in general. Anthropogenic warming also increases the water capacity of the atmosphere, with consequences for increased vapour and evapotranspiration over the oceans and mainlands, which could cause more extreme rainfall and a more significant share of the total annual rainfall (Berghuijs et al., 2014;

Duethmann and Blöschl, 2018; Pulliainen et al., 2020; Thomas et al., 2016). The above-mentioned issues as well as a number of other factors affect many processes of the hydrological cycle in river basins, including changes in the surface and subsurface runoff, changes in regimes, reductions in runoff in various seasons of the year, etc. In recent years, several analyses of the assessment of changes in components of the hydrological cycle in Slovakia have indicated that their regime has recently changed (e.g., Blaškovičová et al., 2019; Fendeková et al., 2018b; Gera et al., 2017; Slezniak et al., 2021), and statistically significant but varying trends regarding their development have been observed (Đurigová et al., 2019; Halmová et al., 2019; Lapin et al., 2016).

There is also evidence of a link between the warming of the climate and potentially increased territorial vapour and droughts (Bonacci et al., 2023; Williams et al., 2020). These changing climate signals are expected to become stronger in the future, but their intensity will vary according to the climate regions or seasons (Cook et al., 2020).

One way to detect and document changes in hydrological patterns is to use the hydrological balance. A whole series of procedures have arisen from studying and generalising about the hydrological balance or taking direct measurements in many basins in different parts of the world. Many procedures have also been extended as tools for estimating the mean annual runoff in catchments despite a lack of direct observations, as reviewed by

McMahon et al. (2012). Historically, models primarily known for regionally valid estimations of evapotranspiration by various authors such as Schreiber (1904); Ol'dekop (1911); Turc (1954, 1955), and Budyko (1974) have been influential. The Budyko (1974) model, which was developed by a Russian climatologist, is based on an equation in which the relationship of the physical balance between the absorption and refraction of solar radiation (a thermal energy model) is used. The model is quite widespread and has been modified by many authors (e.g., Gan et al., 2021; Sposito, 2017). Other models that have been applied to a wide range of surface and atmospheric conditions are mentioned in, e.g., Brutsaert and Parlange (1998); Kahler and Brutsaert (2006); Lhomme and Guillioni (2006); Ozdogan et al. (2006); and Szilágyi and Józsa (2009).

However, the success of the detection of uneven states depends on the scales used. Furthermore, decreases over time and the shrinking of a territory increase the demand for a more extensive range of hydrological cycle data. Dynamic processes must also be considered. Various studies have also investigated trends in actual evapotranspiration based on the hydrological balance and generally found an increase in its values in recent decades (Domokos and Sass, 1990; Keszeliová et al., 2022; Trenberth et al., 2007). Global studies have found positive (but not always significant) trends in North and South America and Europe, but negative trends in Africa and Siberia (Gao et al., 2010; Milly and Dunne, 2002). Long-term time series for monitoring the hydrological balance (including territorial evapotranspiration) can enable the detection of changes in relevant processes and compare them.

Since the territorial values of actual evapotranspiration, especially on large spatial scales, cannot be directly measured, changes in the values over time and their conditioning factors are still featured in the literature; moreover, many hydrologists have reported actual evapotranspiration has increased regionally in many parts of the world (Duethmann and Blöschl, 2018; Zhang et al., 2010). A robust and straightforward option to estimate territorial evaporation is an evaluation of the hydrological balance equation at the basin level, especially when its long-term average is estimated. For example, changes in the water balance, runoff, and snow cover characteristics in the small mountain basin of Jalovecký Creek, Slovakia, in the hydrological years 1989–2018 were analysed by Holko and Danko (2018) and Holko et al. (2020). Their results indicated that daily air temperature data from two meteorological stations in the mountain range located at higher elevations show that the period 1989–2018 was warmer by 2.4 °C than the standard climatic period of 1951–1980. Majerčáková et al. (2004) also pointed out changes in the components of the hydrological balance on the territory of Slovakia. While the long-term average annual precipitation on the territory of Slovakia in the period 1931–1980 was 753 mm, it was 742 mm during the period 1961–2000. The runoff in 1931–1980 was 261 mm, but in 1961–2000 it had decreased by 25 mm to 236 mm. A comparison of the hydrological balance of the main sub-basins of Slovakia for the period 1931–1980 and 1961–2000 showed that in terms of the sensitivity and vulnerability of water resources, approximately 30% of the territory of Slovakia could be potentially very vulnerable (the Ipel', Slaná, Bodva River basins); 35% of the territory could be potentially moderately vulnerable; and the remaining part of the territory could be potentially not very vulnerable. Evaporation has increased throughout Slovakia, with a significant increase already in the northern regions of Slovakia. Garaj et al. (2019) and Keszeliová et al. (2021) dealt with changes in the water balance in some regions of eastern and western Slovakia. Halmová et al. (2022) analysed changes in the long-term water balance in the

small Slovak basins of the Krupinica and Litava Rivers in the period 1931–2020.

Estimates of evapotranspiration are usually based on on-site experimental assessments (e.g., evaporimeters and lysimeters) (Hobbins et al., 2004; Huang et al., 2020), spatial estimates using, e.g., remote sensing data (Gao et al., 2010), or outputs from hydrological rainfall-runoff models (e.g., Parajka et al., 2004; Zhang et al., 2010). In addition, spatial interpolation methods varying complexity from point estimates of actual evapotranspiration can also be used (Hobbins et al., 2001; Lapin et al., 2016). As a result of the above-mentioned problems, it is clear that if we introduce the assumption of water balance closure into the balance calculations and estimate one of the unknown components (e.g., *ET*), this estimate will contain uncertainties based on the methods used. However, for practical reasons, we cannot consider this as a reason for cancelling the use of the balance equation in hydrology. As a basis for the spatial estimation of territorial evapotranspiration to control the outputs from the hydrological balance, there is also a whole range of semi-empirical and empirical relations, which mostly have regional validity (see, e.g., Lavenne and Andréassian, 2018). The mostly used and applied methods around the world include, e.g., Thornthwaite's method, the Budyko equations, or the Bouchet method. Thornthwaite proposed a method for estimating potential evapotranspiration, assuming that average monthly temperatures are positive (Novák, 1995). The advantage of the method is that it enables the calculation of the evaporation in an area where only minimal meteorological data is available. A large number of scientific studies, which have been investigated in different conditions, have shown the accuracy of Thornthwaite's method (Shahidian et al., 2012). Bouchet (1963) developed the theory of a "complementary relationship", according to which actual evapotranspiration (*ET*) and potential evapotranspiration (*EP*) are in a complementary relationship, which was subsequently reworked into a practical application by several authors. Examples of corresponding models that were based on Bouchet's complementary relationship are Brutsaert and Stricker (1979), Morton (1983), and Hobbins et al. (2001).

For the calculation of the daily totals of potential evapotranspiration for days with positive average daily air temperatures, Turc (1954) suggested a relationship between potential evapotranspiration, the average daily air temperature, the average daily value of the intensity of short-wave radiation at the upper limit of the atmosphere, and the duration of sunshine during the day. According to Morton (1983), the model for calculating potential evapotranspiration introduced a new parameter, namely, the equilibrium temperature on the surface, so that data on wind speed is not necessary (Kovács et al., 2011).

Nowadays maps are constructed from the above-mentioned methods' outputs and point measurements for the spatial estimation of evapotranspiration values. These can then also be used to refine the hydrological balance of basins or to control the closing of the balance in the given basin. Within the framework of the Danube countries' regional cooperation, contour maps of the runoff have also been created, see, e.g., Stančík and Jovanovič (1988); Domokos and Sass (1990); Petrovič et al. (2010). In practice, these maps can be used to estimate, e.g., long-term averages of precipitation totals, and flow or climatic evapotranspiration; they can also be used in hydrological models. When creating such maps, the subjective knowledge of a map's authors, which cannot be formulated objectively, can be used, thereby making it possible to consider the local peculiarities of a hydrological regime.

Conceptually, visualizing a water balance on a watershed scale is a simple task. Realistically, it is quite difficult to produce

an accurate numerical estimate from measurement of all the variables in a water balance equation (Ye et al., 2012). Precipitation is the key variable in water budget calculations. The use of precipitation corrections is essential, especially if we want to calculate the balance evapotranspiration from the values of precipitation and runoff measured. In regions with high precipitation totals, even a small percentual error in precipitation can lead to significant deviations in actual evapotranspiration percentages (Mendel and Pekárová, 1983, 1995; Sevruc, 1982; Sevruc et al., 2008). According to several authors, yearly precipitation sums of precipitations are mostly increase by 8–16% after corrections (Adam and Lettenmaier, 2003; Lapin et al., 1991; Ye et al., 2012).

The main objective of this article is to identify potential changes in future runoff for the Danube River sub-basins, which would be uniformly based on changes in air temperature and precipitation. Available data from the database of monthly precipitation, flow rates, and air temperatures from 1961–1990 for 50 headwater sub-basins of the Danube River from the UNESCO Regional Cooperation of the Danube Countries project (Petrovič et al., 2006) was used. Based on the data measured from the selected Danube River headwater sub-basins, two types of empirical regression models for estimating the runoff were derived, i.e.,

- The first part of the paper focuses on a detailed analysis of the relations between the basic individual components of the hydrological balance (long-term average precipitation, evapotranspiration and runoff). Such an analysis helps us to estimate the parameters $a_1, \dots, a_5$ of the equations for the long-term annual runoff depth R .

- In the second part, we have derived simple multilinear equations for estimating the average annual runoff separately for each selected sub-basin of the Danube River basin. Based on the estimated coefficients a and b , this simple relationship makes it possible to immediately determine the effect of an increase or decrease in air temperature on the outflow (or the effect of an increase or decrease in precipitation on the outflow).

As in before mentioned, almost all components of the hydrological cycle are directly or indirectly affected by changing climatic conditions and human activity, which can also be reflected in the hydrological balance of the river basin. Therefore, some of the challenges of hydrologists, as well as climatologists, should be to detect changes in the components of the hydrological balance, analyse and update the relations of the components participating in the hydrological balance, and describe the possible consequences of their changes on the future hydrological conditions in the basin.

2 METHODOLOGY

When processing the hydrological and climatological series observed, a number of statistical methods were used to evaluate the time series. Multiple linear and non-linear regression were used to evaluate the mutual linear dependence of the individual components of the hydrological balance.

2.1 Relations between the long-term average precipitation, evapotranspiration, and runoff

The long-term outflow from a watershed is estimated using the hydrological balance equation, and the water circulation within an enclosed river basin system with concentrated runoff at the outlet is quantified by this equation. The only entry into a river basin is the atmospheric precipitation in the river basin. Therefore, the difference in the soil-water reserves at the

beginning and end of the balance period can be ignored for a sufficiently long period (e.g., a 30-year period). In such circumstances, we can employ a balance equation in the well-known form:

$$P = R + ET + \Delta S \quad (1)$$

where:

P – average annual precipitation total [mm],

R – average annual runoff depth [mm],

ET – actual (balance, territorial) evapotranspiration [mm],

ΔS – average total changes for time Δt .

Actual evapotranspiration can be estimated by several methods:

- i) From the hydrological balance Equation (1) as $ET = P - R$ (average total losses for time $\Delta t = 0$);
- ii) from the energy budget (e.g., by the Budyko method or the Penman method);
- iii) or from empirical relationships (e.g., the Turc (1954) equation).

Several practical empirical equations have been derived to estimate long-term actual (ET) and potential evapotranspiration (EP) based on the measured point data of ET , EP , precipitation (P), and temperature (T). As early as 1954, on the basis of the records of 254 catchments in different climatic regimes, Turc (1954, 1955) evaluated the long-term actual average evapotranspiration as a function of the precipitation (P in mm) and the temperature (T in °C). He proposed the following relations:

$$ET = \frac{P}{\left(a_4 + \left(\frac{P}{EP}\right)^{a_5}\right)^{1/a_5}}, \quad (2)$$

where:

$$EP = a_1 + a_2 T + a_3 T^3, \quad (3)$$

Turc derived the following parameters: $a_1 = 300$; $a_2 = 25$; $a_3 = 0.05$, $a_4 = 0.9$; $a_5 = 2$. The runoff R is deducted as $P - ET$.

Parajka and Szolgay (1998) focused on the estimation and spatial interpretation of the long-term mean annual actual and potential evapotranspiration on the territory of Slovakia using grid-based maps. From contour maps of the long-term mean annual potential and actual evapotranspiration, precipitation (uncorrected), elementary runoff, and the temperature for the period 1931–1960, a system of grid maps was constructed using interpolation methods. This system was tested using map algebra for consistency. It has been concluded that the contour maps do not form a cross-consistent system. The following empirical models for the calculation of the long-term mean annual potential and actual evapotranspiration have been derived from the grid maps:

$$ET = \frac{P}{\left(0.95 + \left(\frac{P}{EP}\right)^2\right)^{0.5}}, \quad (4)$$

$$EP = 373.19 + 34.22T - 2.94T^2 + 0.37T^3. \quad (5)$$

The accuracy of the Equations (4, 5) were tested against the Budyko–Tomlain model at 50 meteorological stations in Slovakia. The new formulae come closer to the Budyko–Tomlain values; however, they are overestimating EP and ET in the mountains and slightly underestimating them in the lowlands, which results in a smaller range and mean. In the studies by Hlavčová et al. (1999), Parajka (2000), and Parajka et al. (2004), the relationships (4) and (5) were updated for the more recent period of 1951–1980.

In Parajka (2000), equations for runoff R were derived for the territory of Slovakia based on nonlinear regression analysis using the following equations:

$$R = P \left(1 - \frac{EP}{(1.168(EP)^{3.273} + P^{3.273})^{1/3.273}} \right), \quad (6)$$

$$EP = 255.894 + 36.2587T + 0.1211T^3. \quad (7)$$

In the first part of this work, we used the long-term values of the hydrological balance components observed in 50 large sub-basins (2000–12000 km²) within the Danube River basin from the period 1961–1990, which were provided by Petrovič et al. (2006). From the dataset (R , P , and T) for the respective sub-basins, we employed a nonlinear regression to determine the parameters a_1 , ..., a_5 of the relationship for the long-term runoff depth R in the Danube basin, according to Parajka (2000) in the form:

$$R = P \left(1 - \frac{a_1 + a_2T + a_3T^3}{(a_4(a_1 + a_2T + a_3T^3)^{a_5} + P^{a_5})^{1/a_5}} \right), \quad (8)$$

where: P (mm) are the long-term average annual precipitation totals in the selected sub-basins;

T (°C) are the long-term average annual temperatures in the selected sub-basins; and potential evapotranspiration $EP = a_1 + a_2T + a_3T^3$.

2.2 Estimation of the annual runoff in the basins

In the second part of the study, we derived the parameters of the empirical models for each of the 50 sub-basins separately using multiple linear regression. A simple linear relationship was employed:

$$R_i = aP_i + bT_i, \quad (9)$$

where: R_i – average annual runoff depths in year i [mm];
 P_i – average annual precipitation totals in year i [mm];
 T_i – average annual temperature in basin in year i [mm];
 a , b – empirical parameters.

For estimating the parameters, 30-year datasets spanning from 1961 to 1990 from the Petrovič database (2006) were utilized. These simple functions enabled us to estimate the expected discharge for a given year, considering the increases/decreases in the air temperature and precipitation within a specific sub-basin.

3 STUDY AREA AND DATA USED

We focused on the Danube River basin in this study. According to the length (2860 km) of the main Danube River and its total drained area, which covers 817,000 km² (8% of its total European extent), it is understandable that this area is of great interest from the point of view of various aspects of historical and recent components of daily life. Within the Danube River basin there are also areas like a delta or similar to a delta, for which the surface runoff is a result of the flat topsoil, which practically cannot be measured. In such cases, the region is very often characterized by many irrigation and amelioration canals, and the water balance there should be estimated differently.

Approximately 600 monthly precipitation series and 125 monthly flow series from the Danube basin were collected by Petrovič during 1961–1990. The Danube area was divided into 109 balance regions. In our study, we used a database of the monthly precipitation, temperature, and discharge time series measured for fifty headwater sub-basins of the Danube basin during the period 1961–1990 (Table 1, Fig. 1; the numbering of the sub-basins follows the original database).

The headwater sub-basins of the Danube River (Table 1) are located in 13 European countries (Germany, Austria, Switzerland, the Czech Republic, Slovakia, Hungary, Ukraine, Serbia, Slovenia, Croatia, Bosnia–Herzegovina, Bulgaria, Romania), cover water basin areas from 2156.8 km² to 11881.6 km²; and have a minimum altitude of 7–967 m a.s.l., average altitude of 178–2302 m a.s.l., and a maximum altitude of 483–3652 m a.s.l. The location of the selected watersheds is shown in Figure 1, and their numbering is preserved according to the database by Petrovič et al. (2006, 2010).

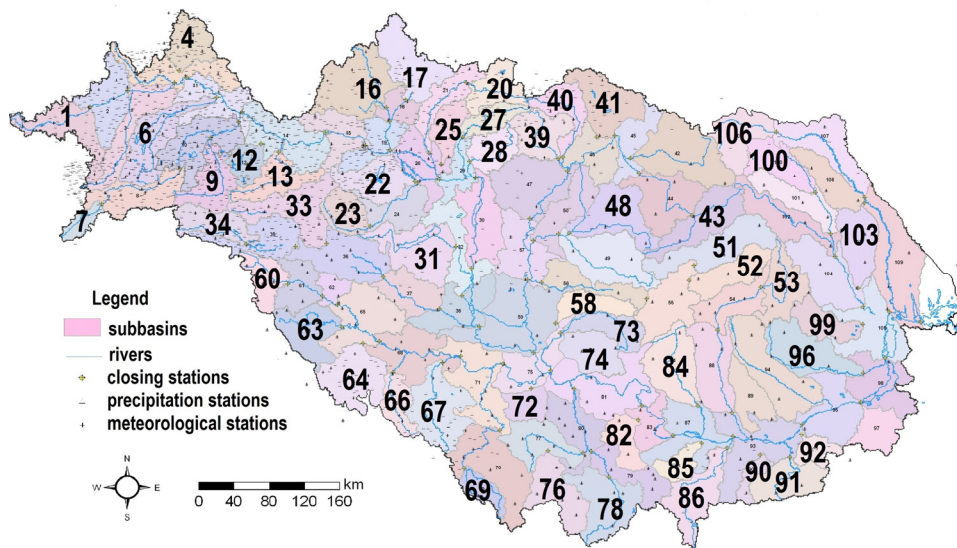


Fig. 1. Location of the fifty sub-basins 1–106 selected in the Danube River basin according to Table 1 (the basin numbering is preserved according to Petrovič et al. (2006, 2010)).

Table 1. List of the fifty selected headwater sub-basins in the Danube River basin with their morphological characteristics and long-term average runoff – R , uncorrected precipitation – P , and air temperature – T , potential evapotranspiration (Budyko-Zubenok) – EP , Petrović et al. (2006).

St. No. Petrović	RIVER (Period 1961–1990)	Station	Country	Area	Min elev	Mean elev	Max elev	EP [mm]	R [mm]	P [mm]	T [°C]
1	Donau	Neu-Ulm Bad Held	GE	7586.4	465	749	2401	623	560	1155	6.88
4	Naab	Heitzenhofen	GE	5414.8	334	497	933	622	296	813	7.18
6	Isar	Platting	GE/AT	8922.0	316	742	2557	618	617	1079	7.49
7	Inn	Kajetansbrucke	SW/AT	2156.8	967	2302	3652	490	854	1329	–0.49
9	Salzach	Burghausen	GE/AT	6627.7	352	1279	3369	546	1188	1479	4.29
12	Traun	Wels	AT	3505.0	309	842	2658	606	1154	1520	6.31
13	Enns	Steyr (Ortskai)	AT	6066.2	284	2302	3652	581	1188	1479	4.29
16	Dyje	Ladna	CZ/AT	11881.6	159	427	792	661	103	569	7.55
17	Morava	Straznice	CZ	9119.4	164	414	1469	639	211	724	7.56
20	Vah	Zilina-Budatin	SK	5684.3	332	855	2197	600	514	923	5.20
22	Mosoni-Duna	Abda	AT/HU	6951.6	106	293	1948	754	103	695	9.07
23	Raab	Kormend	AT/HU	4671.5	184	469	1698	692	225	817	8.22
25	Nitra	Nove Zamky	SK	4048.2	109	328	1234	747	128	670	8.69
27	Hron	Brehy	SK	3814.3	194	673	1768	653	392	798	6.02
28	Ipel	Ipelsky Sokolec	SK/HU	4831.6	115	313	1005	725	125	650	8.38
31	Sió	Simontornya	HU	10207.8	91	178	540	774	70	663	9.51
33	Mur	Spielfeld	AT	9523.2	244	1061	2742	639	493	1045	5.20
34	Drau	Villach	AT	5581.8	486	1663	3365	536	845	1336	2.25
39	Bodva	Felsozsoka	SK/HU	6439.3	107	365	1394	703	147	717	8.10
40	Hornad	Gesztely	SK	5312.2	108	507	1623	680	189	665	6.93
41	Bodrog	Streda n.B.	SK/UKR	11474.0	91	352	1439	559	304	723	6.26
43	Somes	Dej	RO	8816.9	232	648	2303	635	269	716	6.90
48	Barcau-Crisul	Korosladany	HU/RO	9218.0	80	270	1747	729	133	620	9.21
51	Mures	Ocna Mures	RO	9913.5	253	703	2100	643	228	647	6.57
52	Tarnava Mare	Mihalt	RO	6221.6	227	532	1800	635	148	568	6.67
53	Olt	Hoghiz	RO	7143.9	443	797	1792	599	236	632	5.78
58	Begej Plovn	Srpski Itebej	SM/RO	4226.1	76	180	1254	884	168	884	10.13
60	Sava	Litija	SI	4819.3	230	761	2440	615	1113	1694	7.62
63	Kolpa	Sisinec	SI/HR/BiH	7535.1	95	440	1348	686	756	1440	9.12
64	Una	Kostajnica	HR/BiH	9127.3	104	610	1677	700	769	1258	8.61
66	Vrbas	Delibasino selo	BiH	4218.5	141	843	1970	689	677	1026	8.92
67	Bosna	Modrica	BiH	10421.6	99	684	1937	752	549	975	9.99
69	Drina	Bastasi	BiH/SM	3090.2	425	1406	2296	556	1280	1542	4.95
72	Kolubara	Drazevac	SM	3575.5	71	286	1188	786	185	750	10.59
73	Tammis	Jasa Tomic	SM/RO	5294.5	74	418	2089	696	207	687	8.88
74	Caras-Nera	Virtual Nera	SM/RO	6100.8	7	279	1400	1000	180	764	10.60
76	Ibar	Lopatnica Lakat	SM	7552.1	225	891	2178	554	223	640	8.61
78	Juzna Morava	Korvin Grad	SM	9515.5	188	680	1840	714	196	728	9.43
82	Timok	Tamnic	SM/BG	4207.5	59	498	1788	698	210	709	9.14
84	Jiu	Podari	RO	9254.9	67	456	2519	707	308	797	9.07
85	Ogosta	Mizia	BG	3141.5	27	400	1792	810	239	819	10.73
86	Iskar	Orjahovo	BG	8286.0	31	686	2665	704	192	660	8.02
90	Osam	Izgrev	BG	2177.3	41	430	1967	685	198	810	9.43
91	Jantra	Karantzi	BG	6877.2	32	420	2282	783	196	779	10.59
92	Rusenski Lom	Bojichen	BG	2876.2	34	273	483	778	69	570	10.50
96	Ialomita	Slobozia	RO	9206.6	19	365	2505	722	161	651	9.17
99	Buzau	Racovita	RO	5111.1	40	533	1954	690	181	612	7.95
100	Siret	Lespezi	UKR/RO	6053.8	213	455	1470	635	204	650	6.96
103	Barlad	Tecuci	RO	6801.5	31	219	525	737	57	546	9.35
106	Pрут	Chernivtsi	UKR	6798.7	100	513	1848	482	329	798	6.52

4 RESULTS

As mentioned previously, we relied on the long-term values of the hydrological balance components measured from selected 50 headwater sub-basins (2157–11881 km²) within the Danube River basin for the period 1961–1990, as provided by Petrović et al. (2006). The hydrological balance in these sub-basins was processed by Petrović in monthly steps using the WATBAL model (Petrović, 2002; Yates, 1994). The monthly values of the potential evapotranspiration (EP_B) in the WATBAL model were

computed according to the Budyko–Zubenok–Konstantinov method (Konstantinov et al., 1971).

4.1 Estimation of the long-term runoff

First, we studied the dependency between the individual variables using a cross-correlation analysis. Figure 2 illustrates the relationships between the input data (long-term air temperature T , precipitation totals P , Budyko potential evapotranspiration EP_B , and runoff R), of the Danube sub-basins.

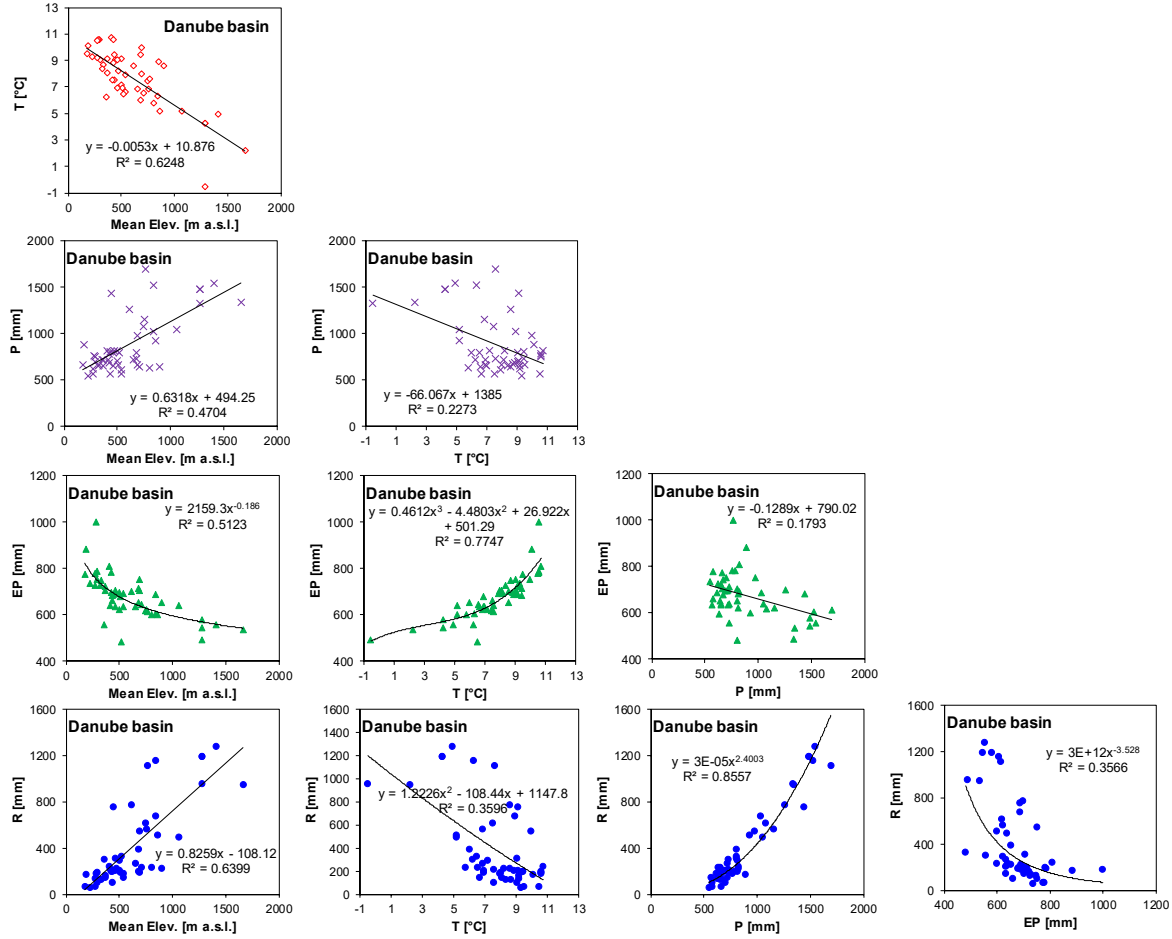


Fig. 2. Cross-correlation matrix, relationship between the input data (air temperature T , precipitation totals P , Budyko potential evapotranspiration EP_B , and annual runoff depth R), long-term values from fifty Danube sub-basins (period 1961–1990).

From the dataset of the long-term averages of the uncorrected precipitation, runoff, and temperatures (P , R , and T , Table 1) measured for the respective sub-basins (period 1961–1990), we employed nonlinear regression to determine the parameters of the relationship (8) for estimating the runoff depth in the Danube sub-basins. The parameters a_1 , a_2 , and a_3 that we estimated according to Equation (8.1) were compared with the parameters estimated by Turc (1954) and Parajka (2000) (Table 2).

$$R = P_{uncorr} \left(1 - \frac{319 + 16.8T + 0.137T^3}{(0.337(319 + 16.8T + 0.137T^3)^2 + P^2)^{1/2}} \right) \quad (8.1)$$

The dependence of the long-term potential evapotranspiration on the long-term air temperature, as demonstrated by different authors, is depicted in Figure 3a. It follows from Figure 3a that the potential evapotranspiration values determined by the Budyko–Zubenok–Konstantinov method are significantly higher compared to the values derived from the empirical equations based on the hydrological balance Equations (8.1). The likely cause is the underestimation of the long-term precipitation totals in the individual sub-basins within the Danube River basin. Petrovič et al. (2006) based their sub-basin precipitation estimates on uncorrected monthly precipitation datasets. The use of precipitation corrections is essential, especially in mountainous sub-basins. In regions with high precipitation totals, even a small percentual error in the precipitation can lead to significant deviations in the actual evapotranspiration

percentages (Mendel and Pekárová, 1983, 1995; Lapin et al., 1991; Sevruck, 1982; Sevruck et al., 2008). According to several authors, yearly sums of precipitation are mostly increased after corrections by 8–16%. Therefore, in the next step, we increased the long-term average precipitation by 5% and 10% and re-estimated the values of the parameters $a_1, \dots, a_5$ in Equation (8.1) by nonlinear regression. The estimated parameters are given in the last two columns of Table 2 and visually presented in Figure 3b. When increasing the precipitation by 10%, the EP estimates in (8.3) are comparable to the values of the Budyko method.

$$R = P_{corr5\%} \left(1 - \frac{399 + 13.1T + 0.188T^3}{(0.513(399 + 13.1T + 0.188T^3)^2 + P^2)^{1/2}} \right), \quad (8.2)$$

$$R = P_{corr10\%} \left(1 - \frac{480 + 9.8T + 0.242T^3}{(0.655(480 + 9.8T + 0.242T^3)^2 + P^2)^{1/2}} \right). \quad (8.3)$$

A comparison between the observed R_{obs} runoff values and the modelled R_{mod} runoff values based on the resulting Equation (8.3) for the precipitation totals (P_{corr}) increased by 10% is presented in Figure 4a–d. The highest positive deviations (lower predicted values) are observed in the basins of BiH, while, conversely, higher predicted runoff values are seen in the of Donau, Begej Plovn, Sava, Kolpa, and Mur sub-basins.

During the stability testing of parameters $a_1, \dots, a_5$ in Equation (8.3), individual stations were sequentially excluded from the

Table 2. Comparison of parameters a_1 , a_2 , and a_3 for calculating the long-term potential evapotranspiration using the regression relationship from the Budyko method, and according to the Turc (1954) model, the Parajka – Szolgay model (P-S) (Parajka, 2000), and the nonlinear regression Eq. (8) for the sub-basins of the Danube River.

	Budyko method	Turc (1954)	P-S (2000)	This study Eq. (8.1)	This study Eq. (8.2) P+5%	This study Eq. (8.3) P+10%
a_1	516.87	300	255.89	319.	399.	480.
a_2	5.698	25	36.258	16.8	13.1	9.8
a_3	0.206	0.05	0.1211	0.137	0.188	0.242
a_4		0.9	1.168	0.337	0.513	0.655
a_5		2.00	3.273	2.00	2.00	2.00

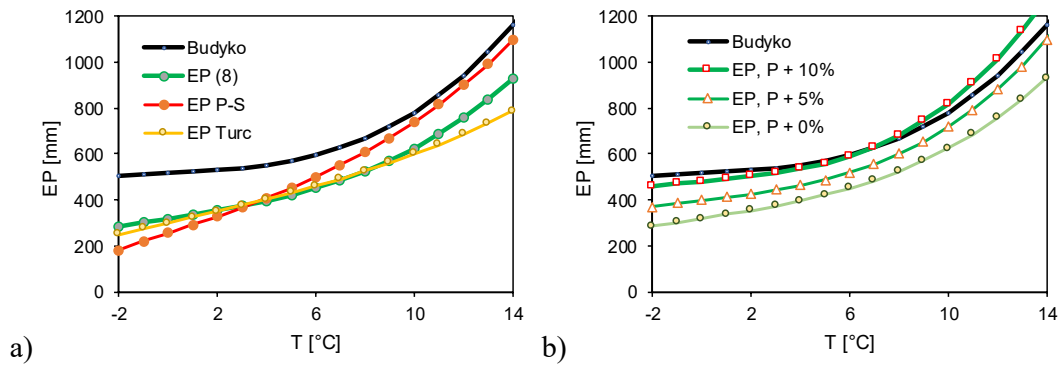


Fig. 3. a) Estimation of the long-term potential evapotranspiration in the sub-basins of the Danube basin according to different relationships (Budyko, Turc, P-S, this study and Table 2). b) Estimation of the long-term potential evapotranspiration in the sub-basins of the Danube basin according to Eq. (8.1–8.3) with different precipitation corrections (P+0% – without any correction, P+5% – increase in precipitation by 5%, P+10% – increase in precipitation by 10%).

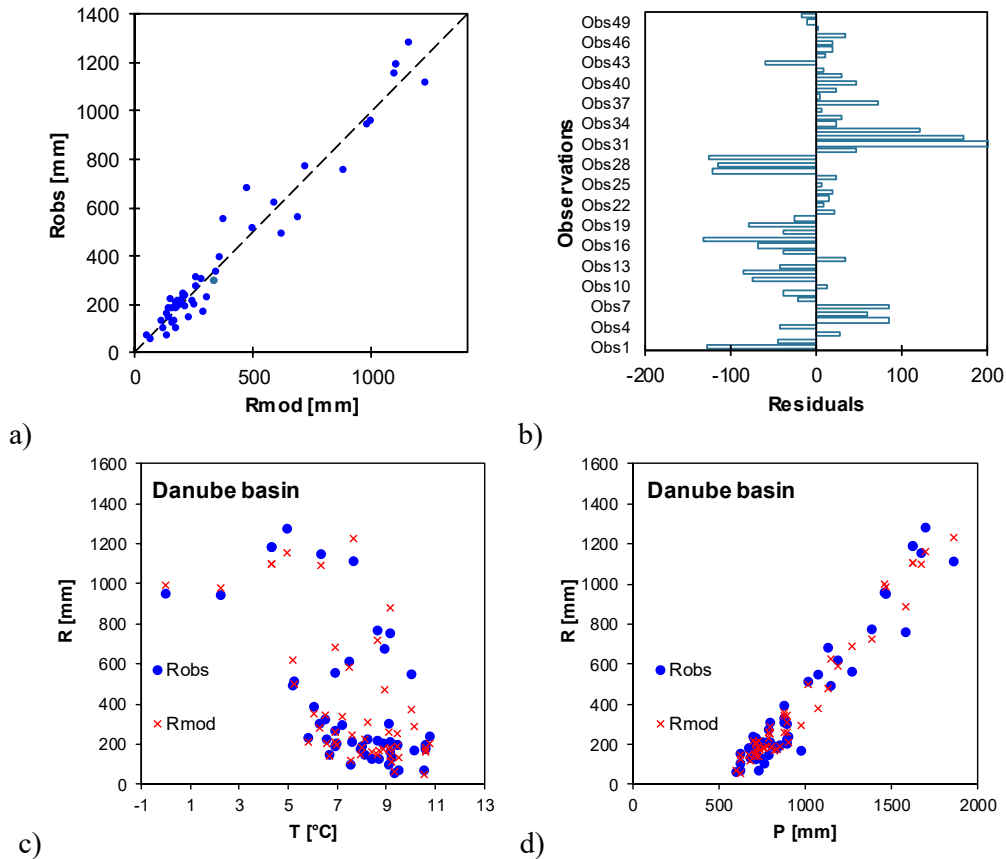


Fig. 4. A comparison between the runoff values measured and the runoff values modelled: a) Comparison of the long-term observed runoff (R_{obs}) and modelled runoff (R_{mod}). b) Residuals between the observed and modelled runoff. c) Relationship between the measured and modelled runoff depths and air temperature. d) Relationship between the measured and modelled runoff depths and corrected precipitation totals. Sub-basins of the Danube basin based on the relationships in Equation (8.3).

input data file. The parameters of the resulting equations are naturally most influenced by the distant outlier values, i.e., high-altitude basins No. 7 – Inn (average temperature $-0.49\text{ }^{\circ}\text{C}$) and No. 34 – Drau (average temperature $2.25\text{ }^{\circ}\text{C}$). Only two of such stations were available. In further research, it will be necessary to focus on the most accurate measurements of the components of the hydrological balance in high-altitude basins (Holko and Danko, 2018).

It is necessary to specify the purpose for which we want to derive indirect equations and what we want to model. If we want to determine a regional formula for estimating runoff depths based on the precipitation and flows measured, we do not need to correct the precipitation. However, if we want to estimate balance of actual basin evapotranspiration, it is necessary to correct the annual precipitation totals. The correction, i.e., increasing the precipitation, will result in an increase in the water balance evapotranspiration as well as a change in the runoff coefficient in the given basin.

4.2 Estimation of the change in the average annual runoff depending on the rising air temperature and falling precipitation in the Danube River basin in the period 1961–1990

In the second part of the study, we derived simple equations in the form of (9) separately for each of the 50 sub-basins within the Danube River basin based on 30-year datasets of the average annual runoff depths, precipitation depths, and air temperature values measured.

$$R_i = aP_i + bT_i. \quad (9)$$

We chose this simple relationship for the purpose of being able to immediately determine the effect of an increase or decrease in the air temperature on the runoff (or the impact of an increase or decrease in precipitation on the runoff) based on the coefficients a and b . Since we estimated the runoff depths, there was no need to correct the annual precipitation totals measured.

As an example, Fig. 5a presents the time series of the observed average annual runoff depths (R_i), the annual precipitation depths (P_i), and the average annual air temperatures (T_i) estimated in the Ipeľ River, i.e., the Ipeľský Sokolec basin, and the Danube, i.e., the Neu Ulm River basins.

Figure 5b on the left shows three equations for the dependence of the annual runoff heights on the average annual precipitation depth for the basin. The linear dependence of the runoff on the average annual air temperature in the basin is plotted in Figure 5b on the right.

The results of the statistical tests of the model parameters are presented in Table 3. The equation of the selected model for the Ipeľ River basin is:

$$R_{est} = 0.41 * P_i - 16.87 * T_i, \quad (9.Ipeľ)$$

and for the Danube at Neu Ulm

$$R_{est} = 0.601 * P_i - 19.28 * T_i, \quad (9.Danube\ NU)$$

It directly follows from (9.Ipeľ) equation that when the air temperature increases by $1\text{ }^{\circ}\text{C}$, the annual runoff decreases by about 17 mm. When the precipitation decreases by 100 mm, the runoff decreases by 41 mm. The time series of the observed annual outflows (R_{obs}) and the simulated values (R_{sim}) for Ipeľ and Danube are presented in Figure 5c.

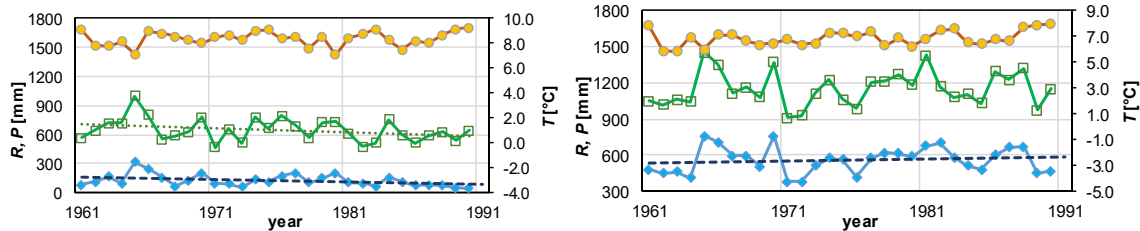


Fig. 5a. Time series of the observed average annual runoff depth (R , blue), the annual precipitation depth (P , green), and the average annual air temperature (T , yellow), period: 1961–1990, the Ipeľ River basin (left) and for the Danube, the Neu Ulm River basin (right). Long-term trends: dashed lines.

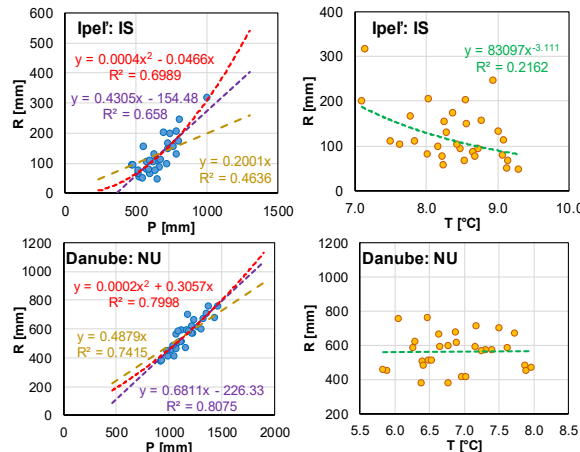


Fig. 5b. Dependence of the annual runoff on the total annual precipitation (left); dependence of the annual runoff on the average annual air temperature (right); the Ipeľ River basin (upper) and for the Danube, the Neu Ulm River basin (down).

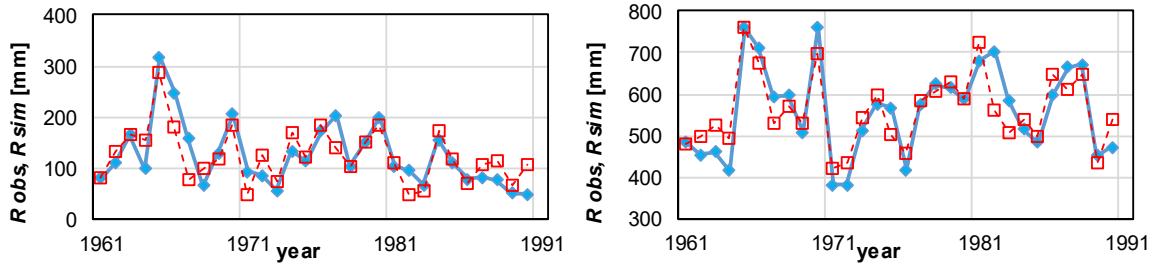


Fig. 5c. Measured (R_{obs}) and simulated (R_{sim}) annual runoff depths according to Equation (9), the Ipeľ River basin (left) and for the Danube, the Neu Ulm River basin (right), 1961–1990.

Table 3. Statistical parameters of the models selected.

Basin	Coefficients	Std. Error	t Stat	P-value *	Lower 95%	Upper 95%	Lower 80.0%	Upper 80.0%
Ipeľ	IS							
P_{obs}	0.410	0.047	8.704	0.000	0.314	0.507	0.348	0.472
T_{obs}	-16.867	3.702	-4.556	0.000	-24.450	-9.283	-21.726	-12.008
Danube	NU							
P_{obs}	0.601	0.055	10.974	0.000	0.489	0.713	0.489	0.713
T_{obs}	-19.283	9.229	-2.089	0.046	-38.187	-0.379	-38.187	-0.379

* the P -value should be less than 0.05.

The model's parameters in Equation (9) were then estimated by linear regression for all 50 sub-basins in the Danube basin based on the average annual outflows, precipitation, and air temperature measured. Table 4 summarizes the resulting coefficients of the model (9) a , b for the precipitation and air temperature in the Danube sub-basins. In the last two columns, the percentage changes in runoff are calculated for a 10% decrease in precipitation and a 1 °C increase in the air temperature. The coefficient b for the air temperature is shown on the map of the Danube basin in Fig. 6a.

From the results of the changes in runoff obtained in our study for the Danube region, we can see that when we assume an increase in the temperature of +1 °C, there will be a decrease in runoff in almost all the Danube River sub-basins. The highest decrease in runoff of about 10% was estimated in the sub-basins of the Tarnava Mare, Ipeľ, Tammis, Raab, Dyje, Siret, Barlad, Hornád and Somes Rivers. Only four river basins show an increase in the runoff but only by about 2–4%. These are the Inn, Drina, Ibar and Bagej Plovn Rivers of mountainous origins, with a high runoff depth.

5 DISCUSSION

The paper presents an estimation of changes in the runoff in 50 headwater sub-basins selected in the Danube River basin as a result of changes in the air temperature. A database of the monthly precipitation, flow rates, and air temperatures for the period 1961–1990 was used.

Runoff conditions are strongly controlled by climate, and any uncertainties in the projections about the future directly translate to uncertainties about future runoff. For example, in Slovakia, despite an increase in cumulative rainfall from 734 mm in 1961–1990 to 797 mm in 1991–2020, runoff decreased by 11 mm. Due to an increase in air temperature (by 1.1 °C) in the last 30-year period (1991–2020), the evapotranspiration balance increased by up to 74 mm from 494 mm in 1961–1990 to 568 mm in 1991–2020. The runoff coefficient decreased from 0.33 to 0.29 (Fendeková et al., 2018a; Halmová et al., 2022). This is in agreement with the results from other European countries. According to Blauhut et al. (2022), Europe in recent years has experienced significantly warmer years associated with a decrease in runoff

in several regions. Therefore, the first part of the paper focused on a analysis of the relations between the individual components of the hydrological balance, i.e., the average long-term precipitation, evapotranspiration and runoff. Such an analysis helps us to determine the parameters of the relationships (8.1) for indirectly estimating the long-term annual runoff depth R in the Danube River basin. These basic equations may have a wide range of applications in indirectly estimating runoff in areas without direct observations, as well as in tasks related to modelling the impact of climate change on changes in the long-term average annual runoff. The dependence of the long-term potential evapotranspiration on the long-term air temperature, as demonstrated by different authors, shows that the potential evapotranspiration values determined by the Budyko–Zubenok–Konstantinov method are significantly higher compared to the values derived from empirical equations (Parajka, 2000; Turc, 1954) based on the hydrological balance equation. The estimation of the long-term potential evapotranspiration in selected sub-basins of the Danube River according to empirical Equation (8.1) was also underestimated compared to the Budyko method.

Therefore, in the next step, we increased the long-term average precipitation by 5% and 10% and re-estimated the values of the parameters in the relationship for estimating the long-term runoff depths in the Danube basin. Two practical empirical Equations (8.2–8.3) with new parameters have been derived to estimate the long-term runoff depths (R), based on the temperature (T) and precipitation without any correction (P), with a 5% correction ($P \times 1.05$), and with a 10% correction ($P \times 1.1$).

The second part of this paper dealt with the construction of the simple multilinear empirical equations $R_{est} = a * P_{obs} + b * T_{obs}$, based on the measured data (1961–1990) from 50 headwater Danube River sub-basins. The coefficients of the multilinear empirical equations a and b were constructed separately for each sub-basin. The regression coefficients a and b immediately determine the effect of an increase or decrease in the air temperature on the outflow (or the effect of an increase or decrease in the precipitation on the outflow). Since we were estimating runoff depths, it was not necessary to correct the annual precipitation totals measured.

Table 4. Coefficients a and b of the linear regression Equation (9) and P -value of coefficients b , the estimated percentage change in runoff height (1) at an average annual increase in air temperature by 1 °C; (2) at a decrease in precipitation by 10% in the Danube basins in the period 1961–1990.

REGION	RIVER (Period 1961–1990)	R [mm]	P [mm]	T [°C]	k	coeff a	coeff b	P value * of coeff b	% change R at –10% P	% change R at +1°C
52	Tarnava Mare	148	568	6.67	0.26	0.58	–26.71	0.000	–22.0	–18.0
28	Ipel	125	650	8.38	0.19	0.41	–16.87	0.000	–21.3	–13.5
73	Tammis	207	687	8.88	0.30	0.65	–26.87	0.002	–21.5	–13.0
23	Raab	225	817	8.22	0.28	0.55	–26.84	0.000	–19.8	–11.9
16	Dyje	103	569	7.55	0.18	0.34	–12.04	0.002	–18.8	–11.7
100	Siret	204	650	6.96	0.31	0.56	–23.07	0.001	–17.9	–11.3
103	Barlad	57	546	9.35	0.10	0.21	–6.23	0.014	–20.1	–10.9
40	Hornad	189	665	6.93	0.28	0.50	–20.54	0.001	–17.5	–10.9
43	Somes	269	716	6.90	0.38	0.64	–27.11	0.001	–16.9	–10.1
72	Kolubara	185	750	10.59	0.25	0.49	–17.52	0.006	–20.0	–9.4
86	Iskar	192	660	8.02	0.29	0.50	–17.54	0.010	–17.3	–9.2
53	Olt	236	632	5.78	0.37	0.57	–21.50	0.033	–15.2	–9.1
48	Barcau–Crisul Repede	133	620	9.21	0.21	0.38	–11.07	0.003	–17.6	–8.3
39	Bodva	147	717	8.10	0.21	0.34	–11.81	0.050	–16.4	–8.0
41	Bodrog	304	723	6.26	0.42	0.62	–23.22	0.016	–14.8	–7.6
27	Hron	392	798	6.02	0.49	0.72	–29.62	0.100	–14.5	–7.6
33	Mur	493	1045	5.20	0.47	0.65	–36.24	0.009	–13.8	–7.4
31	Sió	70	663	9.51	0.11	0.18	–5.11	0.051	–17.0	–7.3
17	Morava	211	724	7.56	0.29	0.45	–15.14	0.002	–15.4	–7.2
63	Kolpa	756	1440	9.12	0.52	0.85	–50.77	0.000	–16.1	–6.7
51	Mures	228	647	6.57	0.35	0.50	–14.92	0.023	–14.3	–6.5
91	Jantra	196	779	10.59	0.25	0.41	–11.85	0.062	–16.4	–6.0
106	Prut	329	798	6.52	0.41	0.57	–19.02	0.027	–13.8	–5.8
82	Timok	210	709	9.14	0.30	0.45	–11.99	0.187	–15.2	–5.7
99	Buzau	181	612	7.95	0.30	0.42	–9.79	0.048	–14.2	–5.4
96	Ialomita	161	651	9.17	0.25	0.36	–8.24	0.031	–14.7	–5.1
4	Naab	296	813	7.18	0.36	0.49	–14.14	0.029	–13.5	–4.8
25	Nitra	128	670	8.69	0.19	0.27	–5.99	0.100	–14.1	–4.7
92	Rusenski Lom	69	570	10.50	0.12	0.18	–3.13	0.346	–14.7	–4.5
74	Caras–Nera	180	764	10.60	0.24	0.34	–7.77	0.013	–14.5	–4.3
85	Ogosta	239	819	10.73	0.29	0.42	–9.93	0.166	–14.4	–4.1
67	Bosna	549	975	9.99	0.56	0.76	–19.27	0.002	–13.5	–3.5
1	Donau	560	1155	6.88	0.49	0.60	–19.28	0.045	–12.4	–3.4
60	Sava	1113	1694	7.62	0.66	0.82	–35.87	0.005	–12.5	–3.2
84	Jiu	308	797	9.07	0.39	0.49	–9.41	0.111	–12.7	–3.1
64	Una	769	1258	8.61	0.61	0.75	–20.53	0.030	–12.3	–2.7
20	Vah	514	923	5.20	0.56	0.62	–10.86	0.280	–11.1	–2.1
9	Salzach	1188	1479	4.29	0.80	0.87	–22.73	0.400	–10.8	–1.9
6	Isar mit Muehlbaeche	617	1079	7.49	0.57	0.65	–11.10	0.210	–11.4	–1.8
78	Juzna Morava	196	728	9.43	0.27	0.31	–3.48	0.571	–11.6	–1.8
22	Mosoni–Duna	103	695	9.07	0.15	0.17	–1.71	0.440	–11.5	–1.7
34	Drau	845	1336	2.25	0.63	0.64	–6.61	0.817	–10.2	–0.8
90	Osam	198	810	9.43	0.24	0.26	–1.49	0.846	–10.6	–0.8
66	Vrbas	677	1026	8.92	0.66	0.70	–4.92	0.580	–10.6	–0.7
13	Enns	1057	1362	4.88	0.78	0.79	–4.12	0.870	–10.2	–0.4
12	Traun	1154	1520	6.31	0.76	0.77	–1.73	0.900	–10.1	–0.1
69	Drina	1280	1542	4.95	0.83	0.73	29.99	0.323	–8.8	2.3
76	Ibar	223	640	8.61	0.35	0.25	7.48	0.108	–7.1	3.3
58	Begej Plovn	168	884	10.13	0.19	0.12	6.37	0.009	–6.2	3.8
7	Inn	854	1329	–0.49	0.64	0.65	41.25	0.270	–10.2	4.8
	average	385	874	8	0.440	0.51	–12.6		–11.6	–3.3

* the P -value should be less than 0.05.

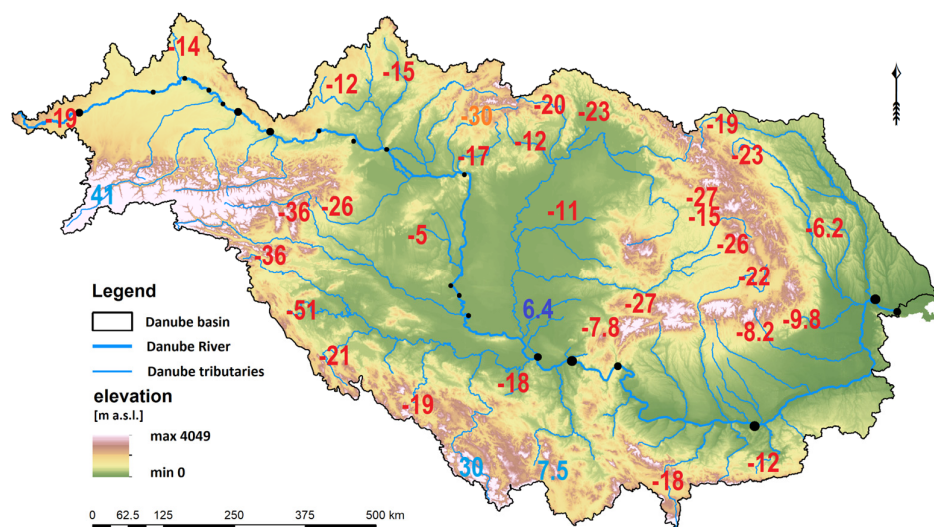


Fig. 6a. Estimated significant parameters b from the model (2) $R_{est} = a * P_{obs} + b * T_{obs}$ in selected sub-basins of the Danube basin. The red colour means a decrease, while the blue colour means an increase in the annual runoff in [mm] with an increase in temperature of +1 °C. Period: 1961–1990.

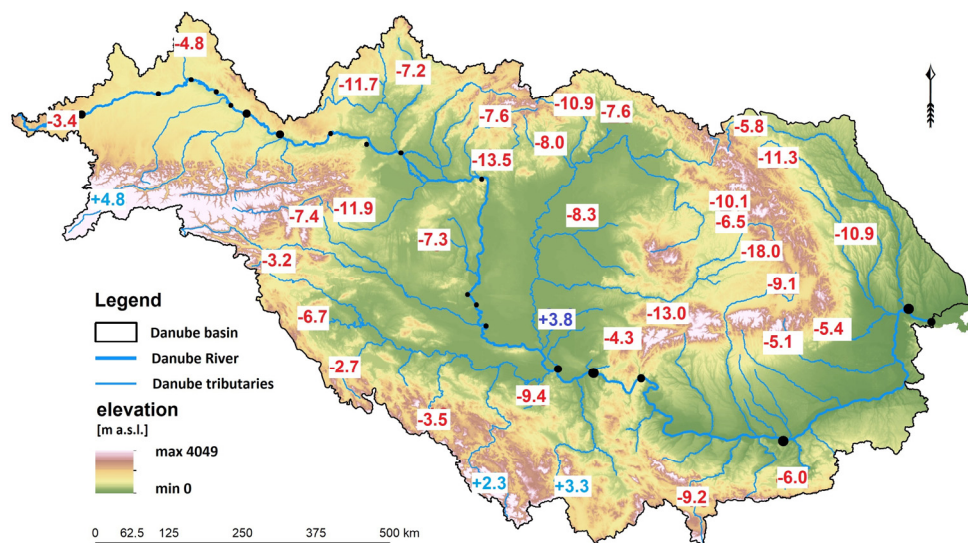


Fig. 6b. Estimated change in percentage of runoff at +1 °C in selected sub-basins of the Danube basin. The red colour means a decrease, while the blue colour means an increase in the annual runoff in [%] with an increase in temperature of +1 °C. Period: 1961–1990.

6 CONCLUSION

The runoff in the Danube region is highly sensitive to changes in temperature, due to shifts in snow processes as well as an increase in evapotranspiration.

The first part results indicated that in estimating runoff, any of the parameters of the Equations (8.1–8.3) can be employed. Correcting precipitation does not significantly alter the runoff estimate. However, by correcting (increasing) the precipitation, the runoff coefficient undergoes a significant change, e.g. decreasing on average from 0.39 to 0.35 in the Danube River sub-basins.

When estimating the long-term actual evapotranspiration, it is essential to correct the precipitation. Otherwise, we obtain underestimated evapotranspiration values referred to as the evapotranspiration index (ETI). It is always necessary to specify in a hydrologic balance whether the precipitation has been corrected or not.

From the results of the changes in runoff obtained from Eq. (9) for the Danube region, it follows that when we assume a rise in temperature of +1 °C, there will be a decrease in runoff in almost all the Danube River sub-basins. The highest decrease in runoff of about 10% was estimated in nine of the Danube sub-basins analysed, i.e., the Tarnava Mare, Ipel', Tammis, Raab, Dyje, Siret, Barlad, Hornád and Somes Rivers. These are basins with low annual runoff depths above 300 mm. In the future, we can expect earlier snowmelt and a lack of water storage in the spring and arid summer and autumn periods. Four river basins show an increase in the runoff but only by about 2–4%. These are the Inn, Drina, Ibar and Bagej Plovn rivers of mountainous origins. Glacier melt could have increased the runoff in the Inn River and other basins; also, the increased frequency of convective storms in the summer season could contribute to higher annual runoff. Generally, the increase in average annual rainfall by 100 mm in the Danube River sub-basins will cause an increase of 50 mm in the outflow, and the increase in the average annual

air temperature by 1 °C will cause a drop in the runoff by 12 mm.

Future increases in demands on water resources in basins must necessitate an adaptation strategy to be undertaken now. This task is essential and requires the close cooperation of climatologists, hydrogeologists, and water managers. However, it should be remembered that this is a simulation of a mathematical model under unchanged future vegetation conditions in the watershed. Such a simulation cannot accurately represent all the changes in a basin; they can occur in a basin with increased precipitation totals or air temperatures.

This research could be extended in the future by analysing and comparing runoff conditions in other periods and developing new climate scenarios from which we could subsequently create a regional scenario of changes in the runoff conditions in the basins.

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