Article

Long-term water balance investigations in the Danube-Tisza Interfluve of Hungary, Europe

József Szilágyi [1,*]

[1] Department of Hydraulic and Water Resources Engineering, Faculty of Civil Engineering, Budapest University of Technology and Economics, Műegyetem Rakpart 1, Budapest, H-1111 Hungary;
* Correspondence: szilagyi.jozsef@emk.bme.hu

Abstract

Increasing aridity of the Danube-Tisza Interfluve of Hungary has led to shallow groundwater decline threatening traditional farming practices, making wetlands disappear and saline lakes dry up. The long-term (1950-2024) water balance of the region is estimated with the help of 0.1° resolution monthly evapotranspiration estimates of the complementary relationship method, as well as spatially interpolated well-measurements of unconfined groundwater levels from 1950-2017. Recharge to the groundwater, obtained as the difference in annual precipitation and evapotranspiration, gradually decreased from around 130 mm in the early 1950s to 20 mm by 1978 then increased slowly to about 47 mm by 2008 and finally dropped again afterwards to its lowest, 14 mm level in 2024. The area is estimated to lose around 55 ± 18 mm of water annually: 80-90% of it as discharge to the Danube and Tisza River as well as deep seepage to the underlying regional aquifer, and about 10-20% as baseflow contribution to the scattered streams of the interfluve. The loss is 10 mm above the mean annual recharge rate of 1950-2017 (13 mm for 1950-2024), producing the observed overall 2-2.5 m drop in unconfined groundwater levels. As long as recharge rates stay predominantly under 55 mm/yr, there remains little hope that the unconfined groundwater of the interfluve could return to the 1950s' level.

Keywords: Danube-Tisza Interfluve; potential recharge to groundwater; groundwater depletion; water balance

1. Introduction

In recent years there have been renewed interest and intense public discourse about the "desertification" of the Danube-Tisza Interfluve [1–3], a unique, sandy region of Hungary (Figure 1), with an area of about 12,000 km^2. An accurate number for the areal extent is hard to define, as there is no natural or artificial surface boundary for the northern edge of the region. Reports of declining unconfined groundwater levels have started to emerge in the 1970s and by the 1990s the observed decline has reached more than 3 m in about 6% of the area [4–6]. The conditions have not improved over the past 30-some years as sustained groundwater declines are being discussed ever since [7–9].

Different authors have reached various conclusions about the possible drivers of the observed regional groundwater decline. Even the same authors have shifted their opinions (see examples in [7]) over time about the relative significance of the possible causes which are typically listed as: climate variability/change, afforestation, unconfined groundwater use, hydrocarbon mining and confined groundwater extraction with an assumed induced deep seepage from the unconfined aquifer, declining gauge levels in the flanking rivers, and canalization/drainage works.

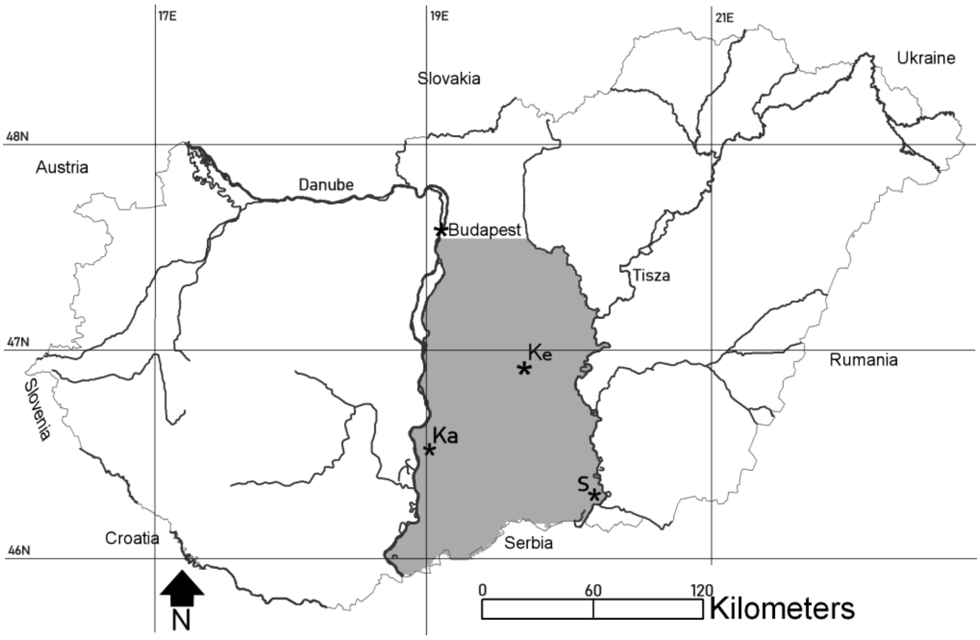

Figure 1. Location of the Danube-Tisza Interfluve within Hungary. Asterisks denote gauging stations (including Budapest) with monthly precipitation (1950-1970) data: Ke –Kecskemét, Ka –Kalocsa, S –Szeged. Gridded HungaroMet precipitation data at 0.1° spatial resolution are available only after 1970.

This study's aim is not another attempt for ranking the different contributing factors by their estimated weight in causing the observed decline as has been done by several authors in the past [6,10,11]. Such ranking can never be fully objective, not the least because these weights may and almost certainly have been changing with time. Rather, the objective of the present work is to offer a common framework in the form of a simplified water balance of the region (by making use of the latest developments in evaporation research) that future more complex investigations may be based on. The foundation of this framework is the groundwater elevation data that were obtained from well-observations by geostatistical analysis [8,9] for the 1950-2017 period and contain monthly groundwater storage values above an arbitrary reference level for the plateau region (areal extent of 8,360 km^2). When dividing these volumes by the corresponding area it is assumed here that the change in the resulting depth values is representative of the entire interfluve. In this sense the focus is on the temporal change of these values and not on any individual depth value which is relative only (over the arbitrary datum).

2. Methods

The general long-term water balance of the study region can be formulated as

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

where P and ET are the area-averaged mean annual precipitation and evapotranspiration rates, respectively, ΔS the change in water storage and Q discharge from the area. ΔS can be further divided into storage change in the unsaturated (ΔS_{us}) and saturated (ΔS_s) zones. In long-term water balances the change in the unsaturated zone storage (especially when the shallow groundwater is within several meters below the surface) may become negligible compared to the sum of in- and outgoing water fluxes, the latter being able to increase without constraints by the length of the period considered. This is so because the former has a physical limit, i.e., its thickness multiplied by the porosity, and restricted further when the water rarely falls below field capacity due to periodic and relatively frequent infiltration events. The same cannot be assumed for saturated-zone storage when a significant long-term trend is being observed in groundwater elevations, as the case for the Danube-Tisza Interfluve.

Discharge (Q) from the area can be categorized into i) baseflow contribution to the Danube and Tisza River (Q_{DT}) as natural boundaries of the interfluve; ii) baseflow contribution (Q_s) to the small streams scattered across the interfluve; iii) deep seepage (Q_v) toward the underlying confined aquifer through a semipermeable layer, and; iv) groundwater pumped directly from the unconfined aquifer (Q_p) and released into streams and canals leaving the area as stream/canal flow. None of these discharges are immediately considered negligible in this study. The only discharge however that is considered negligible by all investigators [4–10], takes place across the northern and southern boundary (Figure 1) of the interfluve. Figure 2 illustrates the basic processes that affect the water balance of the study area. The hydrogeology of the region is most certainly more complex [12] than what is sketched in the illustration, but an overall deep seepage certainly exists [4].

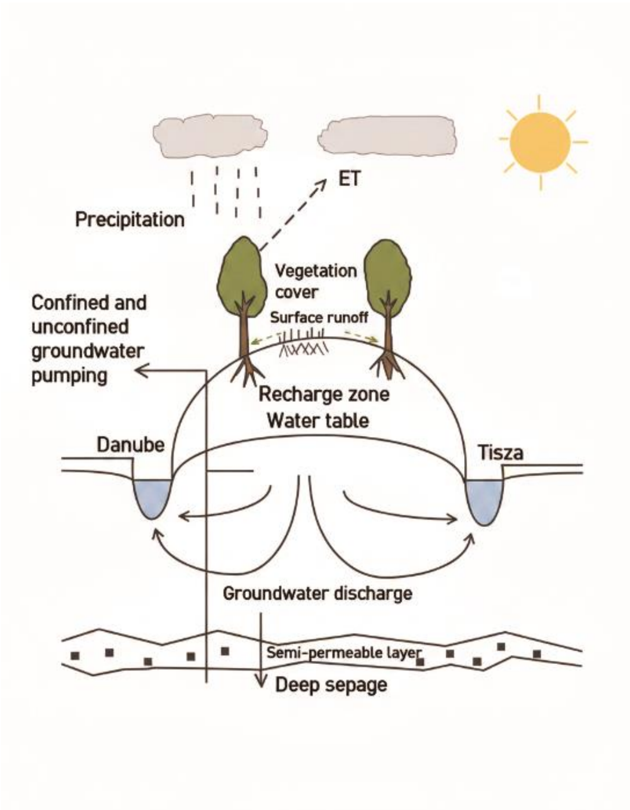

Figure 2. Sketch of the basic processes affecting the water balance of the study area. Surface runoff here includes groundwater contribution to the scattered streams (not shown) of the region.

With the above components, Equation (1) can be reformulated as

$$P - ET = \Delta S_s + Q_{DT} + Q_s + Q_v + Q_p \quad (2)$$

where the $P - ET$ difference acts as potential recharge (R) to the unconfined groundwater. Actual recharge is potential recharge less surface runoff that actually leaves the area. Considering the high porosity and high infiltration rates of the sandy soils of the region combined with a gentle topography, such surface runoff (without baseflow) can be considered negligible [13]. Therefore, potential recharge from here on is treated as actual recharge to the groundwater over the Danube-Tisza Interfluve.

Monthly precipitation at 0.1° resolution after 1970 came from the HungaroMet website (https://odp.met.hu/climate/homogenized_data/gridded_data_sries/daily_data_series/). Before 1971, the monthly precipitation sums of four stations were utilized: Budapest, Kecskemét, Kalocsa and Szeged (Figure 1). The Thiessen polygon method was applied for spatial interpolation of the station values.

Monthly evapotranspiration rates at 0.1° came from the complementary relationship (CR) of evaporation [14,15], driven by ERA5-Land data [16] of mean monthly air and dew-point temperature, wind velocity, net radiation and surface pressure values. The complementary relationship is based on the intricate feedback of land evapotranspiration and humidity of the air. It compares the evaporation rate (called potential evaporation, E_p) of a small water body to that of the wet land (called wet-environment evaporation, E_w) of regional extent. The larger the difference between the two, the smaller actual land ET is (hence the complementarity). The CR as applied, is the culmination of nearly half a century of CR research, started by Morton in the 1970s [17,18]. Recent studies [14,19] proved that the currently employed version of the CR is the most accurate of the several formulations available today worldwide, including Morton's original WREVAP-program estimates [18].

Precipitation and ET rates could also have come from ERA5-Land, but water balances of medium-sized watersheds across Hungary indicated [20] that the HungaroMet precipitation data is more accurate. Also, a global-scale study [21] revealed that even an earlier, calibration-free version of the CR outperformed ERA5 ET estimates. In this study the latest version of the CR is employed the way it was calibrated against water-balances of several Hungarian watersheds and eddy-covariance ET measurements [15]. Details of the method are included in Appendix A.

As the P, ET and ΔS terms of Equation (1) are known for 1950-2017, Q can be estimated on an annual basis. However, a quarter of those annual values become negative owing to uncertainties in the P, ET and ΔS values (Figure 3). Certainly, a negative discharge value in any year is questionable due to the general topography of the interfluve, groundwater levels being 20-40 m above those near the draining rivers of the Danube and Tisza. Because of this large difference in groundwater levels it can be assumed that discharge toward the flanking rivers (Q_{DT}) is more-or-less constant on the long term, even with the observed few meters of decline over the decades. As Q_{DT} is considered the dominant component of Q [6], the latter can also be treated as a constant on a long-term basis. While unconfined groundwater pumping has substantially increased after 1960, Q_p is not considered as a significant loss term [8]. This is especially so because its possible growing contribution (in the range of 0 - 5 mm/yr) to Q may have been balanced by the dropping Q_s rates as the groundwater continuously declined. About the deep seepage component (Q_v), not much concrete information is available, not to mention its long-term trend. Confined groundwater extraction and hydrocarbon mining existed before 1950, therefore Q_v has most likely been present (at an unknown level) over the entire study period.

The constant annual discharge, Q, was estimated by minimizing the root-mean-square-error (RMSE) between the "observed" annual groundwater storage (S) of [9] and its estimate, $\hat{S}$, such as

$$\left[\frac{1}{N}\sum_{i=1}^{N}(S_i - \hat{S}_i)^2\right]^{1/2} \xrightarrow[f(Q^t)]{} min \tag{3}$$

where $\hat{S}_i$ is obtained as

$$\hat{S}_i = S_0 + \sum_{j=1}^{i}\left(R_j - Q^t\right) \qquad i = 1, \ldots, N \tag{4}$$

where N is number of years (= 68) with observed storage, and Q^t is a trial value of the constant discharge. f indicates that the minimum value of RMSE depends on Q^t while

$$S_0 = S_1 - (R_1 - Q^t)\ . \tag{5}$$

Here S_0 ensures that the first value of the storage estimate, $\hat{S}_1$, equals S_1. The trial values in mm/yr were taken from the interval of (1–100), incremented by unity.

3. Results

Optimization of the constant discharge value resulted in $Q = 55$ mm/yr, which is practically the same as what the inversion of Equation (1) yields (Figure 3) for the mean of the annual discharge values (i.e., 55.32 mm/yr). Figure 4 displays the time series of the different drivers of *ET* and the corresponding water balance components. Annual mean air temperature was dropping until about 1980, similar to net radiation, precipitation, potential and wet-environment evaporation rates. While precipitation rates were falling, *ET* rates were growing (probably due to more favorable within-year distribution of precipitation and/or land cover change) until about 1980, which led to the observed medium rate of decline (Figure 5) in groundwater storage. The rate of decline became more severe after 1980 until it levelled off around 1995 (Figure 5), just to continue declining again afterwards. Interestingly, the difference between precipitation and *ET* grew mildly between 1980 and 1995, yet, the decline got more intense during the same period. Between 1980 and 2010, recharge to the groundwater increased slightly, but started to drop again after that, leading to the plight of recent years. Equation (4) with $Q = 55$ mm/yr predicts continued groundwater decline after 2017, the year when available water level observations end, thus casting doubt on any possible claim that water levels could have somehow levelled off after 2005.

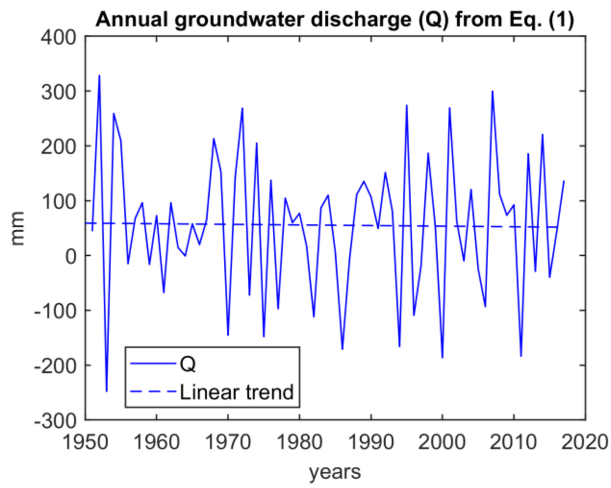

Figure 3. Equation-(1)-inverted annual groundwater discharge (Q) estimates for 1951-2017 and their linear trend. The trend is not significant ($p = 0.97$). The average is 55.32 mm/yr.

Table 1 lists the mean values of the variables in Figure 4 for different time periods. The 1950-1970 and 1971-1992 periods are chosen because studies in the 1990s compared these two periods historically. It is seen that the 1971-1992 period was the driest having the lowest mean annual precipitation, coinciding with the largest observed groundwater decline and the smallest recharge rate. Interestingly, mean *ET* rates constantly increased through time, producing a similar increase in dew-point temperature which must increase with the humidity of the air, as a result of enhanced *ET* rates. These are the only variables (together with the wet-surface temperature) that increased monotonically between 1950 and 2024 (Figure 4).

Table 1. Spatially averaged mean annual values and their standard deviations of the hydro-meteorological variables employed in this study. T_a –air temperature, T_{ws} –wet-surface temperature, T_d –

dew-point temperature, u_2 –2-m wind speed, R_n –net radiation at the surface (in water-depth equivalent), E_w –wet-environment evaporation, E_p –potential evaporation, ET –evapotranspiration, P –precipitation, R –recharge to the groundwater. Q is the water-balance estimated (constant) discharge value calibrated by the well-measurement derived annual groundwater storage values of [9].

	1950 - 1970	**1971 - 1992**	**1950 - 2017**	**1950 - 2024**
T_a (°C)	11.0 ± 0.83	10.88 ± 0.65	11.2 ± 0.83	11.4 ± 0.95
T_{ws} (°C)	10.1 ± 0.56	10.27 ± 0.50	10.4 ± 0.67	10.6 ± 0.78
T_d (°C)	5.12 ± 0.72	5.60 ± 0.55	5.64 ± 0.78	5.76 ± 0.86
u_2 (m/s)	2.33 ± 0.09	2.30 ± 0.08	2.31 ± 0.08	2.31 ± 0.08
R_n (mm/yr)	788 ± 21.3	774 ± 16.0	796 ± 28.4	800 ± 31.0
E_w (mm/yr)	653 ± 17.7	641 ± 15.3	660 ± 25.7	665 ± 28.2
E_p (mm/yr)	930 ± 77.8	874 ± 45.4	915 ± 69.7	922 ± 70.6
ET (mm/yr)	489 ± 40.9	509 ± 25.9	510 ± 38.5	511 ± 39.1
P (mm/yr)	561 ± 93.1	522 ± 74.9	555 ± 110	553 ± 107
R (P-ET) (mm/yr)	72.0 ± 84.4	13.2 ± 68.1	45.2 ± 94.4	42.0 ± 91.6
Q (mm/yr)	67	43	55	(55)

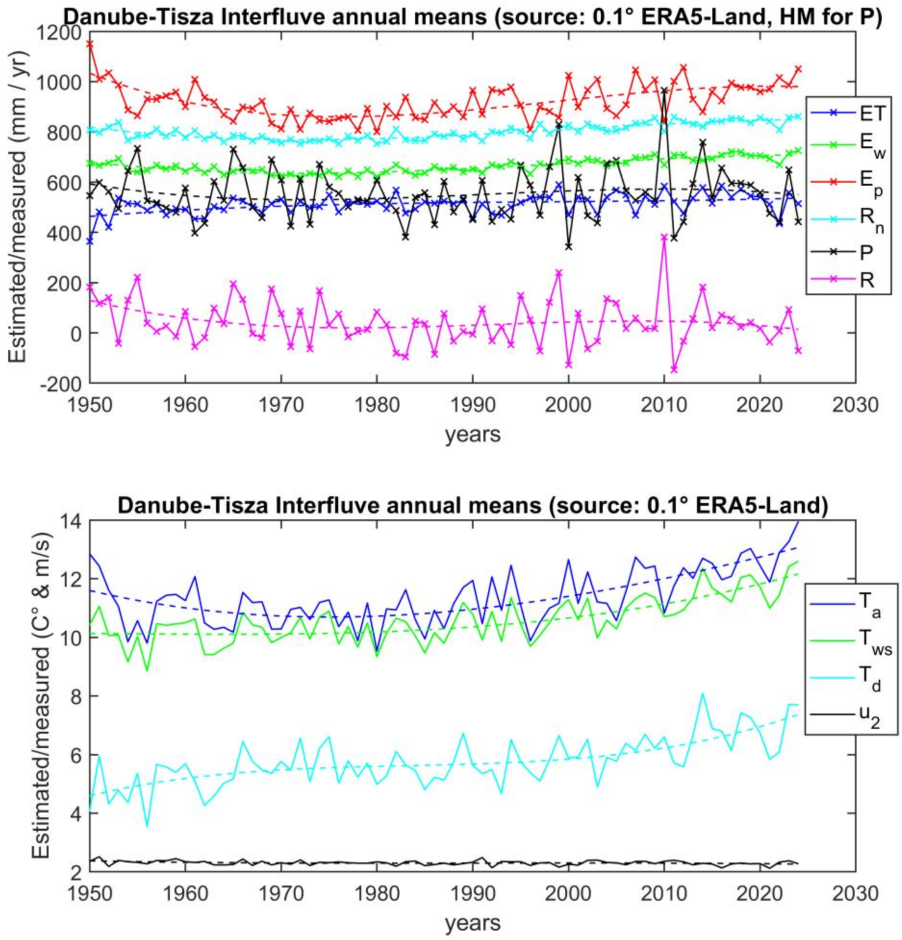

Figure 4. Measured/modeled hydrometeorological variables (1950-2024) averaged across the study area. ET –evapotranspiration, E_w –wet surface evaporation, E_p –potential evaporation, R_n –net radiation at the surface (in water-depth equivalent), P –precipitation, R –recharge to the groundwater

(i.e., $P - ET$), T_a –air temperature, T_{ws} –wet-environment surface temperature, T_d –dew-point temperature, u_2 –2-m wind speed. HM –HungaroMet. The dashed lines denote the 3rd-order best-fit polynomials. All time series display a positive linear trend, significant at the 5% level, except P, R, and u_2 where the negative linear trends are not significant.

The Equation-(3)-estimated Q values for the two intervals differ (67 and 43 mm/yr) but their mean is 55 mm/yr, and when the two periods are combined, the resulting discharge value becomes 59 mm/yr, within 10% of the 55 mm/yr value obtained for the longest possible period (i.e., 1950-2017) with groundwater level measurements. Naturally, the shorter the period for which Q is estimated, the larger the expected uncertainty, as with any estimate. At the same time, natural long-term variability in Q cannot be dismissed but even if existed (however unlikely as seen in Figure 3) it could not be captured with the present methodology.

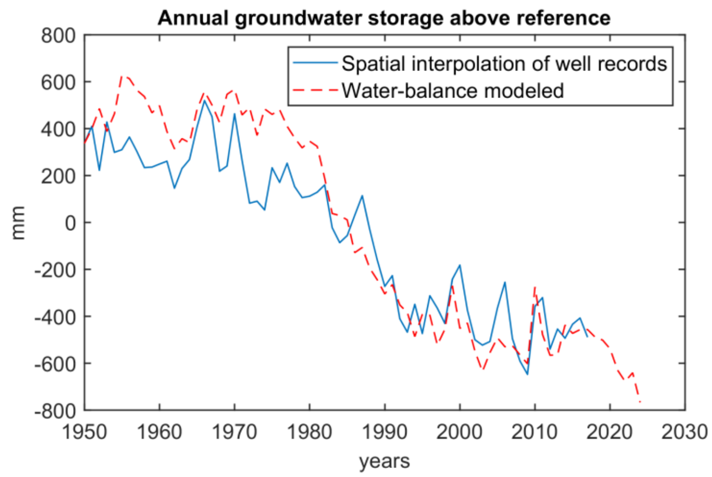

Figure 5. Annual groundwater storage (1950-2024) per unit area above an arbitrary reference level (data from [9]). RMSE = 172.31 mm, R^2 = 0.9.

Figure 6 displays the spatial distribution of the mean annual recharge (R) estimates as P - ET. In general, recharge rates decrease from west to east across the region driven by a similar spatial pattern in precipitation values. The Thiessen polygons due to missing grid values of precipitation before 1971 most certainly contributed to the sharp change in precipitation, and thus, in recharge values along the middle of the region. Overall, only 42 (45 for 1950-2017) mm of water recharges the interfluve annually, which is 13 (10) mm below the 55 mm/yr mean annual discharge rate, leading to the observed decline in shallow groundwater levels.

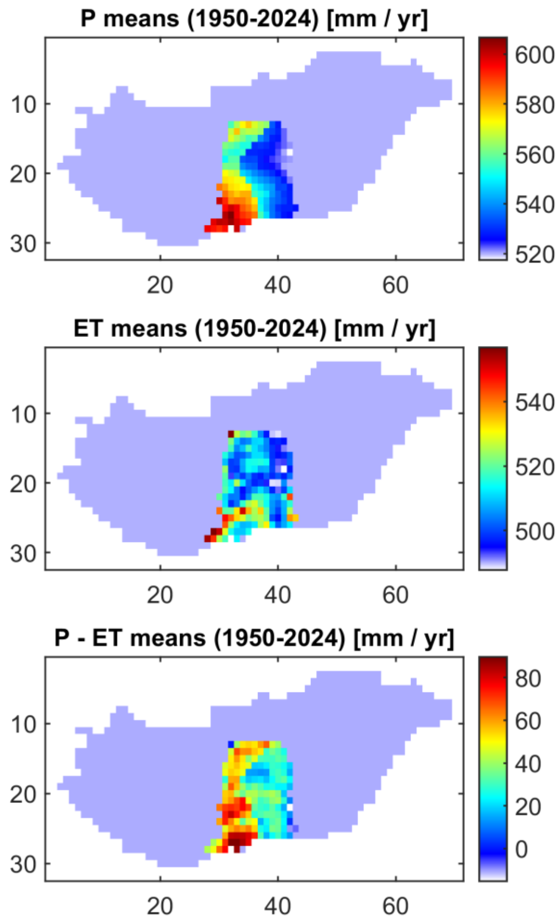

Figure 6. Spatial distribution of the long-term mean annual precipitation, *ET* and recharge (*P* – *ET*) values. The regional averages (from top to bottom panel) are 553, 511, and 42 mm/yr, respectively.

The spatial distribution of the estimated long-term linear trends for *P*, *ET*, and *R* are displayed in Figure 7. Since *ET* rates increased faster than precipitation in most cells, recharge to the groundwater has dropped over a large majority of the region.

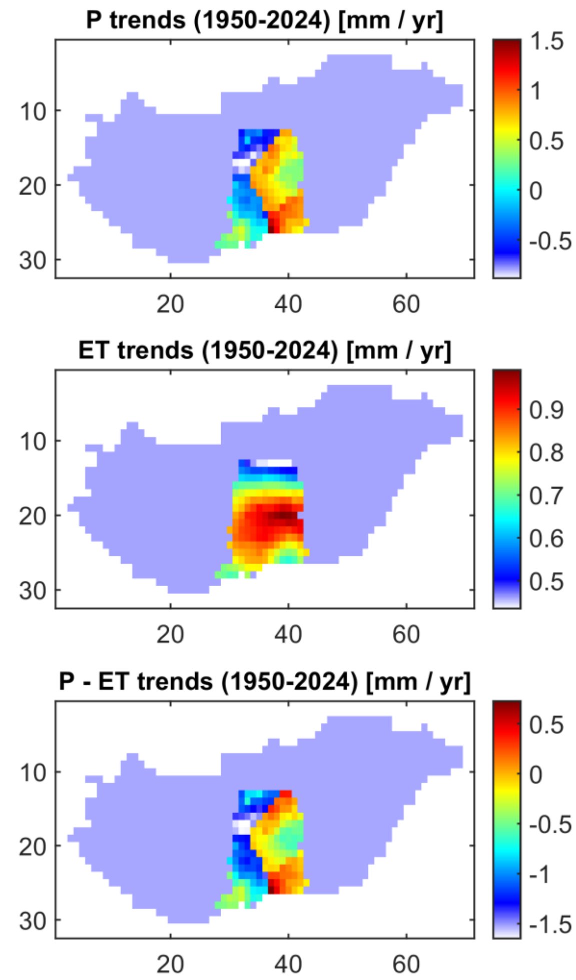

Figure 7. Spatial distribution of the linear trend values. The regional averages (from top to bottom panel) are 0.78, 0.26, and -0.52 mm/yr, respectively.

4. Discussion

On the question of how the present long-term mean annual discharge estimate compares to previous studies, [5] can be mentioned yielding an average value of 43 mm/yr for Q over the 1976-1985 period. [6] via a coupled surface- and groundwater-balance 2D hydrologic/hydraulic model estimated Q_{DT} as 45 mm/yr for the 1951-1970 period. [22] obtained a Q_{DT} estimate of 24 mm/yr for the 1970s based on seepage and tritium-content measurements of the groundwater. At the same time the deep seepage rate, Q_v, was estimated [23] to be in the range of 7-14 mm/yr after taking into account the ratio of the central plateau and total area of the interfluve.

Lacking continuous discharge measurements, [24] estimated average surface runoff, including Q_s, to be ~13 mm/yr for the southern part of the region. [9] obtained modeled Q_s as ~7 mm/yr (average of a dry and wet year) for the Dong Stream, in the South-Eastern part of the region.

An error estimate of 18 mm/yr as standard deviation (σ) can be derived for the 55 mm/yr Q value. This is so because Q is expected to be a positive number (due to the arch-like shape of the shallow groundwater table drained by the Danube and the Tisza River at the sides) and for a normally distributed random variable the interval defined as the mean plus/minus 3σ almost certainly contains all possible values of the variable in question. A σ = 18 mm/yr value thus eliminates the possibility that the long-term mean value of Q could be negative. The lower and upper boundaries, i.e., 37 and 73 mm/yr this way encompass the above 43 mm/yr estimated value of Q by [5] as well as the sum of the Q_{DT}, Q_s, and Q_v estimates.

5. Conclusions

With the help of the latest CR-estimates of monthly ET rates and spatially interpolated well measurements of groundwater levels, the long-term mean annual discharge rate (Q = 55 ± 18 mm/yr) could be estimated for the Danube-Tisza Interfluve. A significant depletion of the shallow groundwater took place over 1950-2017 and modeled to extend until today as the long-term mean annual recharge rate is estimated to be 10-13 mm/yr below Q. As long as recharges rates stay in general below this long-term discharge rate, no improvement can be expected in groundwater levels over the area, especially so if the widening gap (starting in the late 1970s, Figure 4) between potential evaporation and ET [25] persists or becomes even more severe in the future.

Before any large-scale mitigation approach could be successfully planned and implemented in the Danube-Tisza Interfluve, the general water balance of the area must be known. This study, by its specification of the long-term mean discharge and annually varying recharge rates, is one step into that direction. It is hoped that future, more complex regional investigations, studies and modelling efforts will take advantage of the present findings.

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Conflicts of Interest: The author declares no conflicts of interest.

Appendix A

ET rates (mm/d) were estimated by the Complementary Relationship (CR) of evaporation in the form [14] of

$$\frac{ET}{E_p} = 2\left(\frac{w_i E_w}{E_p}\right)^b - \left(\frac{w_i E_w}{E_p}\right)^{2b-1} \quad . \tag{A1}$$

The potential evaporation rate, E_p (mm/d), is defined by the Penman equation [26]

$$E_p = \frac{\Delta Q_n}{\Delta+\gamma} + \frac{\gamma f_u[e^*(T_a)-e_a]}{\Delta+\gamma} \tag{A2}$$

where Q_n (expressed in water-depth equivalent of mm/d) is the available energy (= $R_n - G$) at the surface, R_n (mm/d) the net radiation, while G (mm/d) the ground heat flux, the latter negligible at a temporal averaging of a day or longer. Δ [= $4098e^*(T_a + 237.3)^{-2}$] denotes the slope of the saturation vapor pressure (e^*) curve (hPa / ºC) at the measured air temperature, T_a. The empirical wind function, f_u [mm/(d hPa)], is traditionally formulated [27] as $f_u = 0.26(1 + 0.54u_2)$. Here u_2 (m/s) is the horizontal wind speed at 2-m above the ground/canopy surface and can be estimated by a power function [27] from measurements (u_h) at h meters above the surface as $u_2 = u_h (2 / h)^{1/7}$, and γ [= $c_p p/ (0.622L)$] is the psychrometric constant (hPa / ºC) where c_p [J/(kgºC)] is the specific heat of air under constant pressure, L (J/kg) the latent heat of vaporization and p atmospheric pressure (hPa). e_a (hPa) is the actual vapor pressure, i.e., e^* evaluated at the dew-point temperature (T_d).

The wet-environment evaporation rate, E_w, is often estimated by the Priestley-Taylor equation [28] as

$$E_w = \alpha \frac{\Delta(T_w)Q_n}{\Delta(T_w)+\gamma} \tag{A3}$$

where α is the dimensionless Priestley-Taylor parameter. The wet-environment air temperature, T_w, can be estimated by the wet-surface temperature, T_{ws}, provided the latter is capped by T_a. The implicit equation [29] for T_{ws} can be written as

$$\frac{Q_n - E_p}{E_p} = \gamma \frac{T_{ws} - T_a}{e^*(T_{ws}) - e_a} \tag{A4}$$

requiring some iterations.

The Priestley-Taylor parameter, α, in Equation (A3) is related to T_w [30] as

$$\alpha = \frac{\Delta(T_w)+\gamma}{\Delta(T_w)+c\gamma} \tag{A5}$$

where c (-) is a constant with a value of 0.4 recommended by [30] and kept in this study.

w_i (-) in Equation (A1) is defined as $w_i = (E_p^{max} - E_p) / (E_p^{max} - E_w)$ where E_p^{max} is the maximum achievable rate of E_p (under the given Q_n term) by Equation (A2) when the drying land is completely devoid of moisture (e_a = 0) and the air attained a temperature of T_a^{dry}. T_a^{dry} can be obtained as $T_a^{dry} = T_a + e_a/\gamma$ [15].

Equation (A1) contains only one parameter, b (-), which [15] calibrated to be 1.8 with the help of eddy-covariance ET measurements and water-balances of small to medium-sized watersheds across Hungary.

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