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Research Article

GENETIC DETERMINATION OF THE MAIN COMPONENTS OF THE MEAN INTERANNUAL FLOW WADIS IN THE SEMI-ARID CLIMATE OF THE MAGHREB

M. Ladjel

L. R. H. A. E., University of Bejaia, Algeria

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SUMMARY

North of the Maghreb, the flow of the wadis is generated by the rains. The hydro-climatic network is not dense enough to reasonably estimate the flow in any watershed where the wadis are not gauged. There is a need to develop methods for the acceptable estimation of river flow. The average flow of wadis is conditioned by a combination of climatic factors and local physicogeographic and hydrogeological factors. The influence of these local factors is dominant on the small and medium watersheds through the subterranean supply, where the climatic factors determine the superficial flow. The mean interannual flow (MIF) is composed of a regional climatic flow and a potential local flow. Regional climate flow differs from the actual climate flow in a watershed. The objective of this research is the genetic determination of the main components of the average interannual flow of wadis in the semi-arid climate of the Maghreb.

Keywords: flow, rain, climate, evapotranspiration, infiltration.

Author Correspondence, e-mail: ladjel_mahmoud@yahoo.fr

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1. TOPICALITY OF THE PROBLEM

In semi-arid areas, the flow of wadis is generated mainly by liquid precipitation. The latter undergo different quantitative and qualitative transformations on the slopes. The different losses of rainwater are observed on the slopes, such as: infiltration, retention in micro-depression, retention in the soil, vegetation cover and evaporation. Only a part of this rainwater can flow and reach the wadi bed, by both the superficial and underground channels. The estimation of the mean interannual flow (MIF) remains dependent on the quantity and the quality of hydrometric data. Meteorological data are relatively abundant and of better quality. It is a question of developing the genetically verified relationship between river flow and rainfall. In countries with humid climates, further north, hydrologists have focused only on large and medium rivers where climate contrast is slow and for which the application of the hydro-climatic equation $P_o = E_o + ETR_o$ is appropriately used for estimating fluvial flow [7,10]. At the level of small catchment areas, this equation requires a particular interpretation that must guarantee genetic plausibility to determine the mean interannual flow.

In the current climate situation, these researches must also take into account the expected climate change that influences the genesis of river flow in semi-arid regions. Recent researches confirm the reality of global warming [1]. At the same time, a probable increase in precipitation is expected [2]. In the regions of the Maghreb, the temperature trend is in line with the global situation [3,4]. After the long drought observed (1975-2002), the beginning of a much rainier period is observed in Algeria. The different climate models predict a probable increase in precipitation and temperatures in the Maghreb [5]. In principle, the hydrological flow model should maintain the same structure in view of the invariability of local factors.

2. RESEARCH OBJECTIVE

Sometimes, in hydrological practice, inadequate applications of the water-climate balance equation can be encountered. Therefore, we propose our interpretation, based on the verification and genetic identification of the main elements of this equation and which are based on the hydro-climatic database, in the semi-arid climate of North Africa.

Consider a territory receiving a quantity of rain P_o, where the hydrogeological conditions are uniform. In small watersheds, this rain P_o breaks down into surface runoff $E_{\sup}$, infiltration $\inf_1$ and evapotranspiration ETR_1. In average watersheds, rainfall P_o is decomposed into shallow flow $E_{\sup}$, subsurface flow $E_{st,2}$, infiltration, $\inf_2$ and evapotranspiration ETR_2. While in large catchment areas, rainfall P_o is broken down into river flow (climatic) $E_o = E_{\sup} + E_{st}$ and reference evapotranspiration ETR_o [12]. The surface flow component $E_{\sup}$ is identical in all watersheds and depends directly on the amount of precipitation and surface properties of the slopes. The groundwater component E_{st} arises from a given area $S_{cr,1}$ and continues to increase with increasing watershed area, ie with the attainment of horizons draining groundwater.

Evapotranspiration ETR_i is of course proportional to the amount of rainfall and increases with the increase in the surface area of the catchment areas. It depends on soil characteristics and vegetation cover. For the different areas of the watersheds, we have the following inequality $ETR_o > ETR_2 > ETR_1$. For given climatic conditions, the interannual mean reference evapotranspiration ETR_o represents a maximum value of the region that can be estimated by one of the formulas, known in climatology. On the other hand, by comparing the equations of the water balance of a small catchment area $P_o = E_{\sup} + ETR_1 + \inf_1$ and a large catchment area $P_o = E_{\sup} + E_{st} + ETR_o$, we obtain $ETR_o = ETR_1 + (\inf_1 - E_{st})$. This actually shows that $ETR_o > ETR_1$.

For any watershed, the expression of hydrological balance is given by the water balance equation [6, 7]. For a large watershed, the water balance equation is written as follows $P_o = E_{c\lim} + ETR_o$, where the three terms of the equation are genetically climatic, that is the amount of rain P_o is entirely spent on boundary potential climatic flow $E_{c\lim}$ and reference evapotranspiration ETR_o. Hydrometric measurements have shown that under semi-arid climatic conditions and in a watershed of any size, river flow E_o is always greater than climatic flow $E_{c\lim}$. The difference $\Delta E = E_o - E_{c\lim}$ is potentially generated by some of the rain that has infiltrated the basement, when it should theoretically be evaporated. This is called a potential local flow addition E_{loc}. Then, the evapotranspiration that corresponds to this flow E_{loc} is called local evapotranspiration ETR_{loc}. This evapotranspiration is caused, under semi-arid climatic conditions, by the difference $P_o - E_{loc}$. Hence the climatic flow at the level of a watershed $E_{c\lim,BV}$ which is expressed by:

$$E_{c\lim,BV} = P_o - ETR_{loc} \quad (2)$$

Obviously, evapotranspiration ETR_{loc} is lower than the value of the regional reference evapotranspiration ETR_o.

3. STATUS OF THE QUESTION

River flow reflects the ability of the watershed surface to transform rainfall, due to the process of water runoff. The surface runoff process is determined by the amount of water falling on the surface, by evapotranspiration from the earth's surface and by the underlying local factors that cause water runoff. In general, flow is a universal characteristic and depends on the climate, which varies with the latitude of the locality [6,7,12].

The use of the water balance method is fundamental in the study of hydrological regimes. In the case of a watershed, this method shows the relationship between water inflow and outflow, taking into account the variation in reserves over a given time interval or for each phase of the water regime. This method allows direct hydrological estimation of a water element, even in the absence of appropriate observations [6].

For a watershed, the water balance equation, for a given time interval Δt, is written in general terms and includes the main components of the balance, expressed as a water slide [7]:

$$E = P - E_v \pm \Delta U \quad (3)$$

where: E - river flow blade at the river outlet, in mm; P - precipitation over the catchment area, in mm; - E_v evaporation of the surface of the catchment area, in mm; $\pm\Delta U$ - variation in moisture reserves in the catchment area, in mm.

This equation is applicable to watersheds without natural or artificial flow regulation and for rivers that have a complete groundwater supply and are generally medium and large rivers [7]. For small rivers, there is a constant lack of groundwater supply in the river system.

Hence, the water balance equation can be represented by:

$$E_{\sup} = P - E_{st.d} - E_{st.nd} - E_{v} \pm \Delta U \qquad (4)$$

where: $E_{\sup}$ - surface runoff; $E_{st.d}$ - groundwater flow, drained by the river; $E_{st.nd}$ - undrained groundwater flow, exiting beyond the limits of a small watershed.

In very small watersheds, the first draining horizon is so deep that the component $E_{st,d}$ tends to zero. In equations (4), (5), the value ΔU is not constant because the moisture reserves in the basin decrease during the low water years and increase in the wet years. As a result, the average interannual value ΔU tends to zero. In this case, the equation of the water balance (4) of a medium and large river basin, with a natural regime, is of form:

$$E_{o} = P_{o} - E_{v.o} \qquad (5)$$

where: E_{o}, P_{o}, $E_{v.o}$ - interannual mean values of the components of flow, precipitation and evaporation of the surface of the catchment area.

For a small catchment area, the water balance equation (5), over a long period of time, takes the form:

$$E_{o} = P_{o} - E_{st.nd,o} - E_{v,o} \qquad (6)$$

Mean interannual river flow (MIF) is a stable hydro-climatic feature that determines the potential water resources of a given watershed or of a given region. Some hydrologists consider the IME of medium-sized rivers to be a climatic characteristic [10]. Indeed, the equation of the water balance of the river basin of an average river, for a long interannual period, is written in the form [7]:

$$E_{o} = P_{o} - ETR_{o} \qquad (7)$$

where: E_{o}, P_{o} and ETR_{o} are the interannual averages, respectively the values of flow, rain and evapotranspiration.

The supply of groundwater to rivers begins from an area S_1 and continues to grow to an area S_2. In addition, the values of S_1 and S_2, in mountain areas and in the plain, depend on the relief and hydrogeological conditions of the region. Above the limit S_2, the influence of the relief gradually diminishes, the watershed drains all the groundwater and the river flow is equal to the sum of the surface and groundwater components, i.e.: $E_o = E_{\sup} + E_{st}$.

The theoretical concept of climatic flow is established for the first time by Mezentsev V. B., hydro-climatologist from Omsk in 1957 [8]. It is a flow generated solely by climatic factors: average precipitation P_o and average evapotranspiration ETR_o, in a given geographical landscape. It is expressed by the following hydro-climatic balance equation:

$$E_{c\lim} = P_o - ETR_o \tag{8}$$

with $P_o -$mean interannual rainfall, in mm and $ETR_o -$mean interannual reference evapotranspiration, in mm.

The disciples of Mezentsev V. B. applied this design to determine the mean interannual flow over a large part of Western Siberia [8, 9, 10]. The latter confirm that the results are identical with the flows measured at the river outlet. While the climatic flow is conditioned by the total absence of rainwater losses by infiltration. It should be known that this is only for large watersheds, whose draining substratum coincides with the level of the outlet of the watercourse.

Later, in his work on the influence of underlying surface factors on the formation of mean flow in Ukraine, Loboda N. S. developed methodical approaches for its quantitative estimation, based on maps of isolines of climate flow, determined by the hydro-climatic equation. She considered climatic flow $E_{c\lim}$ as an integral characteristic of water resources that reflects the zonal specificities of the distribution of annual flow, conditioned by climatic factors. But she found that for the majority of small and medium-sized rivers in the semi-arid zone of Ukraine, the climatic flow $E_{c\lim}$ was different from the values of the mean river flow E_o. The relative deviations of the calculated $E_{c\lim}$ and actual values E_o reach, in some distinct cases, 30% and more [11].

As a rule, these differences are conditioned by the influence of underlying surface factors, which occur, particularly in small and medium-sized watersheds, with limited and unstable groundwater supply. For the first time, on the basis of the genetic concept of annual flow formation, developed by Befani A. N. and his disciples [13,14], the principles for determining the components of the water-heat balance on the territory of the Republic of Moldova are implemented [15].

For the physicogeographical conditions of Ukraine, Loboda N. S. noted the inequality of the two types of flows: real E_o and climatic $E_{c\lim}$ and proposed a conversion coefficient $k = \frac{E_o}{E_{c\lim}}$ of the climatic flow $E_{c\lim}$ into real flow E_o. This coefficient can be less than one or greater than one. Loboda N. S. proposes expressions for this coefficient, for the plain landscape, the coefficient depends on the surface area of the catchment area and for the mountainous landscape, the coefficient depends on the average altitude of the catchment area[11].

In Moldova, the assessment of the influence of anthropogenic factors on the water-heat balance has been reduced to the use of the anthropogenic influence coefficient k_a, which is the relationship between the characteristics of the water-heat balance E_a and the value of climate flow $E_{c\lim}$ [15]. Consequently, this made it possible to evaluate the modified characteristics of the components of the water-heat balance using the product of the coefficient of anthropogenic influence k_a by the climatic flow $E_{c\lim}$, i. e. $E_a = k_a E_{c\lim}$. The model adopted allowed to evaluate, separately, the impact of each factor (regulation of the flow, agricultural activities, urbanization, irrigation and water consumption) and their total impact, mainly on the water resources. However, the mean interannual flow (MIF) estimated by the hydro-thermal balance equation $E_o = P_o - ETR_o$, in terms of heat and moisture ratios, reflects only the influence of climatic factors, excluding the influence of factors. underlying surface.

4. STUDY AREA AND BASELINE DATA

The North of Algeria represents less than one fifth of the total surface area of the country and includes the different river basins, the subject of this study.

The relief of this part of the country, characterized by two mountain ranges: the Tellian Atlas in the northern part and the Saharan Atlas in the southern part. At the foot of the Tellian Atlas are the coastal plains. Between the Tellian Atlas and the Saharan Atlas are the high plains with a semi-arid character and a continental climate. Given this natural diversity, the country's climate is highly contrasted from one region to another. Thus, in the North the climate is Mediterranean, mild and not very windy. In the center of the country, the reliefs are characterized by a mild climate where it is not uncommon to see in winter snow in the mountains of Kabylia and Aurès. The study region is therefore jointly subjected to a Mediterranean climate in the North and a predominantly Saharan climate in the South. In the Tell, annual rainfall amounts vary from 100 mm to 1200 mm; average temperatures are around 30°C in summer and around 10°C in winter. In summer the Chehili, a dry and very hot wind blows from the Sahara.

The climatic and hydrological data, used in the present research, are collected in the database of the National Hydric Resources Agency of Algeria. These data are collected for 104 watersheds in northern Algeria, with an area ranging from 16 km^2 to 4060 km^2. The series of hydro-climatic observations are not always of the statistically recommended size and are flawed [21,22]. It should be noted that the average data for potential rains and evapotranspiration are taken directly from maps drawn up by the National Agency for Hydric Resources of Algeria. While the data on average river flows for watersheds are the average values of annual flows, collected in the directories of the same agencies. For series of observations ranging from 20 to 40 years and over.

The data which are at our disposal, are only those provided by the specialized organizations: meteorological and hydrological. The only effective, quantitatively verifiable tool is the water balance equation, at the level of a watershed. It is precisely the absence of certain specific data that prompted us to carry out this research.

5. RESEARCH METHOD

Generally speaking, for the annual time interval, the annual water balance equation for a watershed of any size is written as:

$$P_a = E_a + ETR_a + \inf \pm \delta_a \qquad (9)$$

with: P_a – annual rainfall, in mm; E_a – annual flow, in mm; ETR_a – evapotranspiration from the surface of the catchment area, in mm; $\inf$ – water slide infiltrated into the subsoil, in mm and $\pm\delta_a -$variation in the volume of water stored in the catchment area. For a sufficiently long time interval, including an integer number of hydrological cycles, this equation takes the following form:

$$P_o = E_o + ETR_o + \inf_o \tag{10}$$

Obviously, for large watersheds or for a large territory, the mean interannual flow E_o is equal to the regional climatic flow $E_{c\lim}$, i.e. infiltration is almost nil.

The hydro-climatic balance equation is then written as follows:

$$P_o = E_{c\lim} + ETR_o \tag{11}$$

The analysis of hydrological data from the northern Algerian catchment basins showed that the measured river flow E_o is always higher than the regional climatic flow $E_{c\lim}$. The differences between these two components of flow $\Delta E = E_o - E_{c\lim}$ constitute the local potential groundwater supply of the region, which is conditioned by the amount of rainfall and the surface and hydrogeological conditions of the watersheds [23]. Analysis of data from a number of watersheds, located in climatic regions at more latitudinal latitudes, has shown that deviations $\Delta E = E_o - E_{c\lim}$ can be negative [17-19].

From the graph below (fig. 1), it can be seen that the differences in the flow $\Delta E = E_o - E_{c\lim}$ actually decrease with the increase in the area of the watershed S, since the increase in the areas of the watersheds, the value of the actual river flow E_o tends towards the value of the regional climatic flow $E_{c\lim}$.

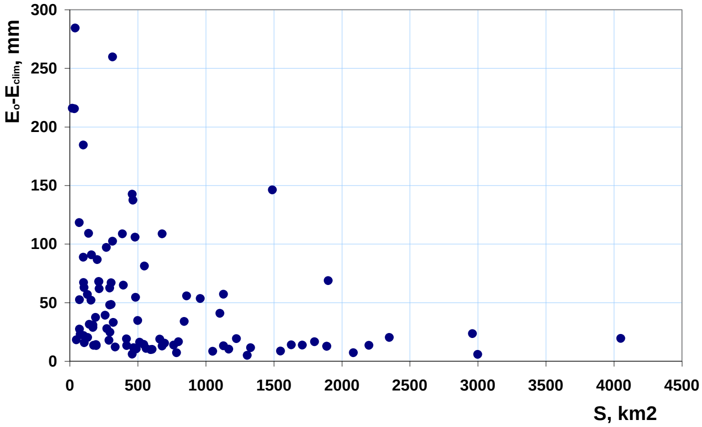

Fig.1. The dependence of $E_o - E_{c\lim} = f(S)$

Moreover, the equation of the water balance of a watershed of any size, is written:

$$P_o = E_o + P_r \tag{12}$$

with: P_r − flow losses, in mm.

And since we have $E_o = E_{c\lim} + \Delta E$. Hence, the balance equation (12) takes the form:

$$P_o = E_{c\lim} + \Delta E + P_r \tag{13}$$

with: ΔE − positive flow difference, which is now called local flow and is designated by the symbol E_{loc}.

Comparing the two equations (12) and (13) allows us to see the equality $ETR_o = \Delta E + P_r$, and to deduce the contribution added to the flow ΔE is a part of rain infiltrated deep into the subsoil and which has not had time to evaporate. On the other hand, the flow losses P_r are always lower than the reference evapotranspiration ETR_o for small and medium watersheds. Equality $P_r = ETR_o$ is verified only for large watersheds. Ultimately, the actual river flow E_o consists of a regional climatic component $E_{c\lim}$ and a potential local component E_{loc}, namely: $E_o = E_{c\lim} + E_{loc}$. For all the watersheds examined, local flow E_{loc} is always higher than regional climatic flow $E_{c\lim}$ [21]. The dependence between these components $E_{loc} = f(E_{c\lim})$ is narrow (Fig.4).

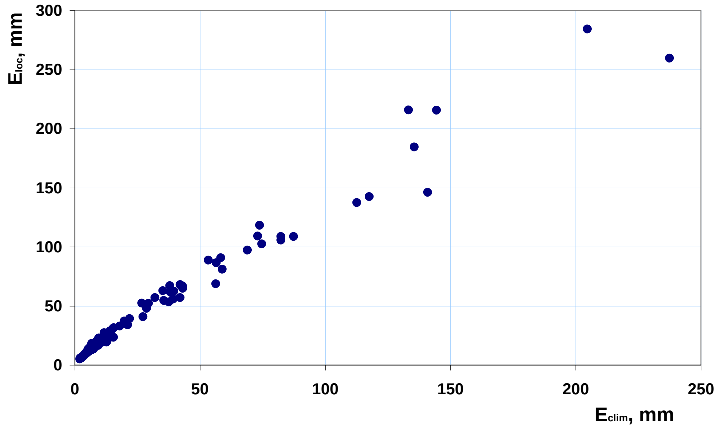

Fig.4. Dependency $E_{loc} = f\left(E_{c\lim}\right)$

This relationship can be expressed as follows $E_{loc} = k\, E_{c\lim}$, with $k-$ a proportionality coefficient. In a previous work [23], we carried out the analysis of the dependence between the coefficient k with the area S, with the average altitude of the watersheds H_o and the climatic component of the flow $E_{c\lim}$. We obtained a relationship $k\left(S, H_o, E_{c\lim}\right)$ characterized by a coefficient C_k, which depends on the latitudinal zonality and which has been mapped. But the analysis of the dependence between the coefficient k and the different factors showed that the best correlations are obtained with the surfaces S of the catchment basins and also with the climatic flows $E_{c\lim}$ (fig. 5 and fig. 6). By logarithmic anamorphosis, we obtained a dependence $\ln(k/S) = f\left[\ln(S)\right]$, with a high correlation coefficient $r = 0.977$.

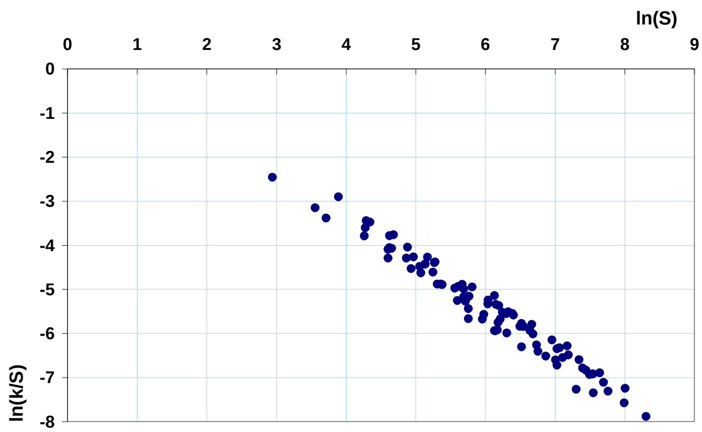

Fig.5. Dependency $\ln\left(k/S\right) = f\left[\ln\left(S\right)\right]$

Similarly, we obtained a dependency $\ln(k/E_{c\lim}) = f[\ln(E_{c\lim})]$, with a high correlation coefficient $r = 0.997$.

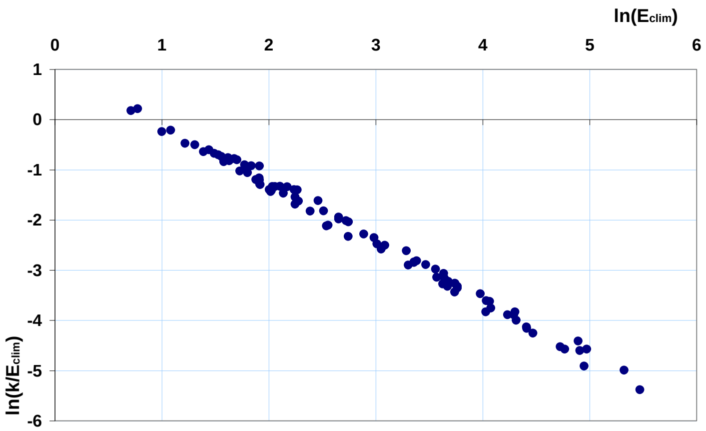

Fig.6. Dependency $\ln(k/E_{c\lim}) = f[\ln(E_{c\lim})]$

Finally, the grapho-analytical analysis and multiple regression allowed the same relationship to be obtained, expressing the proportionality coefficient k as a function of the watershed area S and climatic flow $E_{c\lim}$, in the following form:

$$k = \frac{k_{\max}}{S^{1/10}\, E_{c\lim}^{1/5}} \tag{14}$$

The value $k_{\max}$ is the maximum regional value of the coefficient k, i. e. $k_{\max} = 5.80$. Finally, river flow can be estimated directly by the following expression:

$$E_o = \left(1 + \frac{k_{\max}}{S^{1/10}\, E_{c\lim}^{1/5}}\right) E_{c\lim} \tag{15}$$

The evaluation of the flow, by this relationship, using the same hydro-climatic observation data, is characterized by an average relative error of 0.30%. The relative error of estimating the river flow of an ungauged wadi depends on the relative errors of rainfall P_o and potential evapotranspiration ETP_o.

6. SURFACE FLOW AND UNDERGROUND FLOW

Thus, one can express equations of the water balance, for the different sizes of watersheds:

a- small catchment area, we have $P_o = E_{\sup} + \inf + ETR_1$, with:

$$E_o = E_{\sup} \tag{16}$$

b- medium sized watershed, we have $P_o = E_{\sup} + k\,E_{st} + (1-k)\inf + ETR_2$, with:

$$E_o = E_{\sup} + k\,E_{st} \tag{18}$$

c- large catchment area, we have $P_o = E_{\sup} + E_{st} + ETR_o$, with:

$$E_o = E_{\sup} + E_{st} \tag{19}$$

Fluvial flow E_o, at the outlet of a river, is composed of superficial flow $E_{\sup}$ and subsurface flow E_{st}.

The superficial component of the flow is formed mainly by runoff, generated by the net part of the rains, during the floods. While the other homeland of the flow, which is drained to the outlet, is called a groundwater flow. At the level of a watershed, part of the rainfall, infiltrated into the subsoil, does not contribute to the formation of evapotranspiration in the latter, either an effective rainfall $P_o - E_{loc}$, or:

$$ETR_{loc} = f\left(P_o - E_{loc}\right) \tag{20}$$

Consequently, the climatic flow of this watershed is expressed by:

$$E_{c\lim,BV} = P_o - ETR_{loc} \tag{21}$$

The surface flow is equal to the climatic flow of this watershed $E_{\sup} = E_{c\lim,BV}$, that is to say:

$$E_{\sup} = P_o - ETR_{loc} \tag{22}$$

The groundwater component of the flow is directly calculated by the expression:

$$E_{st} = E_o - E_{\sup} \tag{23}$$

7. ESTIMATE OF FLOW LOSSES

River flow losses P_r consist of losses by evapotranspiration from the surface of the watershed ETR_{BV} and losses by infiltration $\inf$, i.e. $P_r = ETR_{BV} + \inf$. We know that infiltration losses are proportional to rainfall amounts and decrease with increasing surface area S and average watershed elevation H_o. These losses are proportional to the quantities of losses P_r, ie $\inf = f(P_r, S, H_o)$. The losses by infiltration are expressed by a monome of the form:

$$\inf = \left(1 - \frac{P_r}{ETR_o}\right) P_r \qquad (24)$$

Evaporation losses in a watershed are determined by the following expression:

$$ETR_{BV} = \frac{(P_o - E_o)^2}{ETR_o} \qquad (25)$$

The evapotranspiration values ETR_{BV} in the watersheds, estimated by formula (25), show the decrease with the increase in the area of the watersheds. The dependency graph $ETR_{BV} / ETR_o = f(S)$ shows that ETR_{BV} represents 45% the reference evapotranspiration ETR_o for small catchment areas and represents 90% the reference evapotranspiration ETR_o for large catchment areas (fig.6).

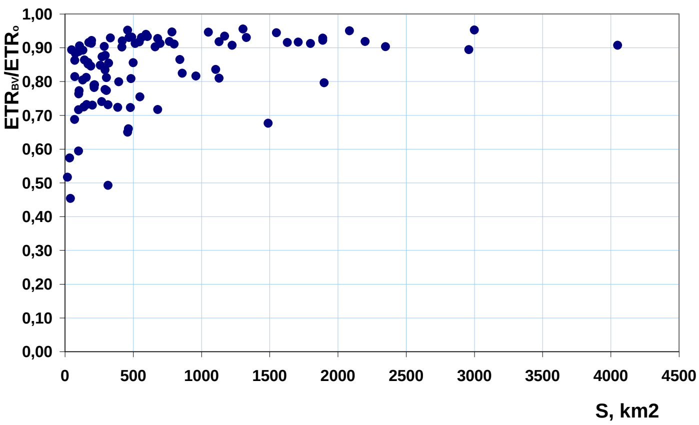

Fig.6. Dependency $ETR_{BV} / ETR_o = f(S)$

Losses of flow by infiltration and evapotranspiration are proportional to rainfall amounts and increase with increasing rainfall (Fig. 7).

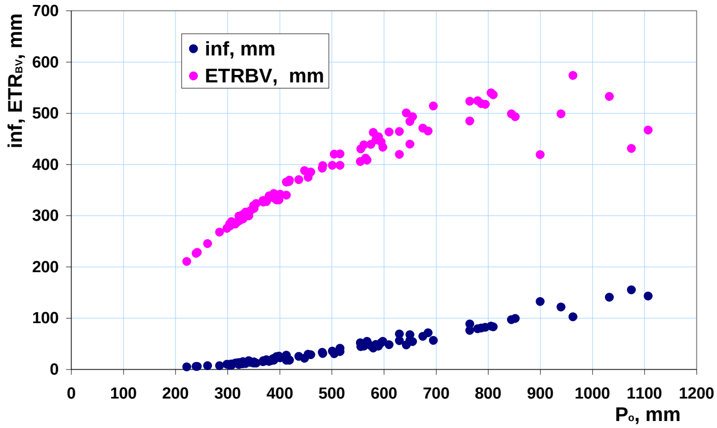

Fig.7. The dependencies $\inf = f(P_o)$ and $ETR_{BV} = f(P_o)$

When the amount of rain is minimal, evapotranspiration takes up all the rain. In the case of northern Algeria, rains in the order of 100 mm are entirely expended by evapotranspiration and losses by infiltration are almost nil. Losses by infiltration range from 2.2% to 14.5% of total losses.

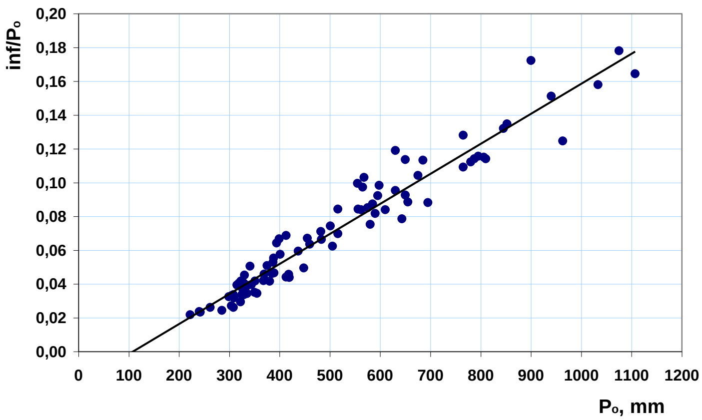

Fig.8. Dependence $\inf / P_o = f(P_o)$.

The graphical analysis of the dependence $\inf / P_o = f(P_o)$ shows that the average minimum value of the rain below which losses consist solely of evaporation is of the order of 100 mm (fig.8).

8. DISCUSSION OF THE RESULTS

The decomposition of river flow into potential climatic flow and potential local flow and the analysis of the dependence between them made it possible to identify the influence of local factors, which is limited to the surface area of the watersheds and the potential climatic flow. We have shown that evapotranspiration in small and medium-sized watersheds is always lower than the reference evapotranspiration. The surface component of the flow is genetically equal to the estimated climatic flow for the watershed and the groundwater component is equal to the difference between the river flow and the surface component of the flow. Losses by infiltration, estimated by the proposed relationship, verify their reduction with the increase in the area and altitude of the watershed.

They become minimal and stable beyond 2000 km^2 and above 1000 m altitude and they tend to reach zero when the rains are of the order of 100 mm and are entirely expended by evapotranspiration.

9. CONCLUSION

The flow of Maghreb wadis is generated by liquid precipitation, it is conditioned by climatic factors and physico-geographical and hydrogeological factors. We propose an interpretation of the hydro-climatic equation, by genetic identification of the main elements of this equation, based on hydro-climatic observations, in a semi-arid climate. In small and medium-sized watersheds, rainfall is broken down into surface runoff, infiltration and evapotranspiration. So, in large watersheds, it breaks down into potential climatic flow and reference evapotranspiration. For the same amount of rain, evapotranspiration is inversely proportional to the catchment area. In large catchment areas, the amount of rainfall is entirely spent on potential climatic flow and reference evapotranspiration, with zero infiltration losses. In a semi-arid climate, river flow is always greater than climatic flow. The difference between fluvial flow and climatic flow is generated by some of the rain that has infiltrated into the subsoil, whereas it should be evaporated. This difference is called local potential flow. Flow is the potential groundwater supply of the region, proportional to the amount of rainfall and conditioned by local surface and hydrogeological factors. It is inversely proportional to the area of the watershed, since as the area of the watersheds increases, the actual river flow and the regional climate flow become equal. The evapotranspiration corresponding to the local flow is called local evapotranspiration. Thus, the climatic flow, at the level of a catchment area, is expressed by $E_{c\lim,BV} = P_o - ETR_{loc}$. On the other hand, the flow losses are always lower than the reference evapotranspiration for small and medium watersheds. Their equality is verified only for large watersheds. Thus, fluvial flow is represented by regional climatic flow and potential local flow. There is a close relationship between these two flows that can be expressed by the relationship $E_{loc} = k\, E_{c\lim}$ The coefficient k has been strongly related to catchment areas and climatic flows $E_{c\lim}$. River flow can be estimated directly by the following expression $E_o = \left(1 + \frac{k_{\max}}{S^{1/10}\, E_{c\lim}{}^{1/5}}\right) E_{c\lim}$, with an average relative error of less than 1%. The superficial component of the flow is equal to the climatic flow in the watershed, ie $E_{\sup} = E_{c\lim,BV}$. The groundwater component of the flow is calculated by the expression: $E_{st} = E_o - E_{\sup}$. River flow losses consist of losses by evapotranspiration from the surface of the watershed and losses by infiltration. The average minimum value of rain below which losses are constituted solely by evapotranspiration is about 100 mm and losses by infiltration are almost nil.

In the context of global warming, recent climatological studies have predicted a simultaneous increase in temperatures and precipitation for the northern Maghreb region. The factors which are likely to be affected by this warming are the plant cover and the surface layer. from the ground, through the influence of actual evapotranspiration. Consequently, our methodical approach remains valid for determining the relationship between the climatic component and the local component of the flow, not only for the mean values but also for expressing their dynamic variability over time.

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