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Courses

Hydrology **II**

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FOREWORD

Hydrology (from the Greek:, "hýdōr" meaning "water"; and "lógos" meaning "study") is the science that deals with all aspects of the water cycle, and in particular with exchanges between the sea, the atmosphere (oceanography, climatology...), the land surface (limnology) and the subsoil (hydrogeology), on earth (or potentially on other planets). The hydrologist contributes to the knowledge and management of water resources and their sustainability in relation to environmental watersheds.

The hydrologist uses the tools and concepts of the earth and environmental sciences, in particular physical geography, geology or civil and environmental engineering1.

Using a variety of analytical methods and resources, it collects and analyzes data, and where necessary models, to help solve water-related problems (overexploitation, pollution, salinization, environmental degradation affecting the water cycle, prevention and management of the consequences of technological and natural disasters, etc.).

This course is designed for *L3* students in the *hydraulics field*. The program includes four chapters:

- ➢ Probability and statistics
- ➢ Statistical and probabilistic study of precipitation
- ➢ Stream flow study
- ➢ Flood flow studies

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Chapter 1: Probability and statistical concepts

Chapter 1 Probability and statistical concepts

Introduction

Statistics is the discipline that studies phenomena by collecting, processing, analyzing, interpreting and presenting data in a way that can be understood by everyone. It is at once a science, a method and a set of techniques.

In practice, statistical methods and tools are used in areas such as :

- Geophysics, for weather forecasting, climatology, pollution, river and ocean studies;
- Demographics: the census provides a snapshot of the population at a g i v e n point in time, enabling us to to by the subsequently for surveys in at samples
  representative.
- economic and social sciences, and in econometrics: the study of the behavior of an Statistics are the key to understanding a population group or economic sector. Insee is working in this direction. Environmental issues also rely on statistical data;
- Ecology, for the study of plant communities and ecosystems.

Statistics enable us to use the information gathered to establish any causal relationships through interpretation and analysis.

A random phenomenon is a phenomenon with random variables, i.e. variables linked to chance, whose values cannot be known in advance.

Statistics are applied in almost every field of scientific activity. When analyzing data relating to a group of individuals or objectives, e.g. water levels, flow rates in a river, rainfall intensities, etc., it is often impossible or impractical to examine all the elements of the group, called a population; so a small part of the group, called a sample, is examined.

A sample can be representative of one, two, three or more variables, but never totally identical to the total population. The representativeness of a sample is always only partially verifiable. It's a relative notion.

Chapter 1 Probability and statistical concepts

Statistical analysis

The characteristics of a statistical series make it possible to characterize this series by highlighting information whose value gives an important indication of the series studied.

A statistical series is a sequence of individual data. For example, let's take the following series of annual precipitation from a weather station, Pen (mm) :

An	P(mm)	An	P(mm)	An	P(mm)	An	P(mm)	An	P(mm)
1996	794	2001	655	2006	804	2011	645	2016	549
1997	720	2002	720	2007	708	2012	645	2017	671
1998	652	2003	420	2008	755	2013	591	2018	800
1999	674	2004	645	2009	655	2014	698	2019	795
2000	804	2005	695	2010	549	2015	708	2020	704

Table 1: Average annual precipitation for a watershed

In general, these raw data are not organized. To be able to analyze such a series and highlight its essential characteristics, we proceed as follows:

Serial order

The values studied can be arranged in either ascending or descending order. The difference between the largest and smallest values is called the amplitude of the series.

A value is entered only once, and the number of times it has been observed is indicated opposite. This number is the value's actual or absolute frequency (n_i); for example, the absolute frequency of precipitation 794mm is 1.

We can also indicate the relative frequency (fi) for this value, which is the ratio between the frequency of the value and the total absolute frequencies $N=\sum n_i = 25$ of the series, so the relative frequency of precipitation 794is $\frac{1}{25}$ =(0) (.0) (4), table N°2 summarizes all calculation results.

Pi	Workforce or frequency absolute (ni)	Relative frequency (fi)	Pi	Workforce or frequency absolute (ni)	Relative frequency (fi)
420	1	0.04	698	1	0.04
547	2	0.08	704	1	0.04
591	1	0.04	708	2	0.08
645	3	0.12	720	2	0.08
652	1	0.04	755	1	0.04
655	2	0.08	794	1	0.04
671	1	0.04	795	1	0.04
674	1	0.04	800	1	0.04
695	1	0.04	804	2	0.08

Table 2: Classification of precipitation values

To highlight the characteristics of the series studied, we group them into value classes.

When you want to summarize a large quantity of raw data, it's convenient to distribute them into classes or intervals and determine the number of individuals belonging to each class, which is called the frequency or class size.

Number of intervals

In the case of a continuous quantitative characteristic, drawing up a frequency table involves first dividing the data into classes. This involves defining the expected number of classes, and therefore the amplitude associated with each class or class interval.

As a general rule, classes of the same amplitude are chosen.

For the frequency distribution to be meaningful, each class must contain a sufficient number of values (n_i) .

Various empirical formulas can be used to establish the number of classes in a sample of size *N (in our case N=25 values)*

STURGE's rule: Number of classes K= 1+ (3.3 log *N*)

The YULE rule: Number of classes K = $2,5\sqrt[4]{n}$ The interval between each class is then obtained as follows:

Class interval (CI)= (X max - X min) / Number of classes

with *X* max and *X* min, respectively the largest and smallest value of *X* in the statistical series.

In our case :

Sturge's rule: 1+ (3.3 log 25) = 5.61 Yule's rule:

$2,5\sqrt[4]{50}$ = 5.59

The two values are not very different, so we take 6 intervals.

Definition of the class interval :

$$IC = \frac{804 - 420}{6} = (6)\,(4)$$ mm rounded to 5 mm for convenience

From Xmin we obtain the class limits or class boundaries by successive addition of the class interval
For the first class in our case :
Lower limit = 420-0.5 = 419.5mm Upper limit =
419.5+64 = 483.5mm
Table N°03 summarizes all these operations

Class numbers	Terminals classes	Center classes	Number or absolute frequency (ni)	Relative frequency (fi=ni/N)
1	419.5≤ Pi< 483.5	451.5	1	0.04
2	483.5≤ Pi< 547.5	515.5	2	0.08
3	547.5≤ Pi< 611.5	579.5	1	0.04
4	611.5≤ Pi< 675.5	643.5	8	0.32
5	675.5≤ Pi< 739.5	707.5	7	0.28
6	739.5≤ Pi< 803.5	771.5	6	0.24

Table 3: Grouping of precipitation values

Graphic representations

Graphical representations have the advantage of providing immediate information on the general shape of the distribution. They make it easier to interpret the data collected.
For continuous quantitative characteristics, the graphical representation is *the histogram*, where the height of the rectangle is proportional to the number n_i. This is only true if the class interval is constant. In this case, the area under the histogram is proportional to the total number of individuals.

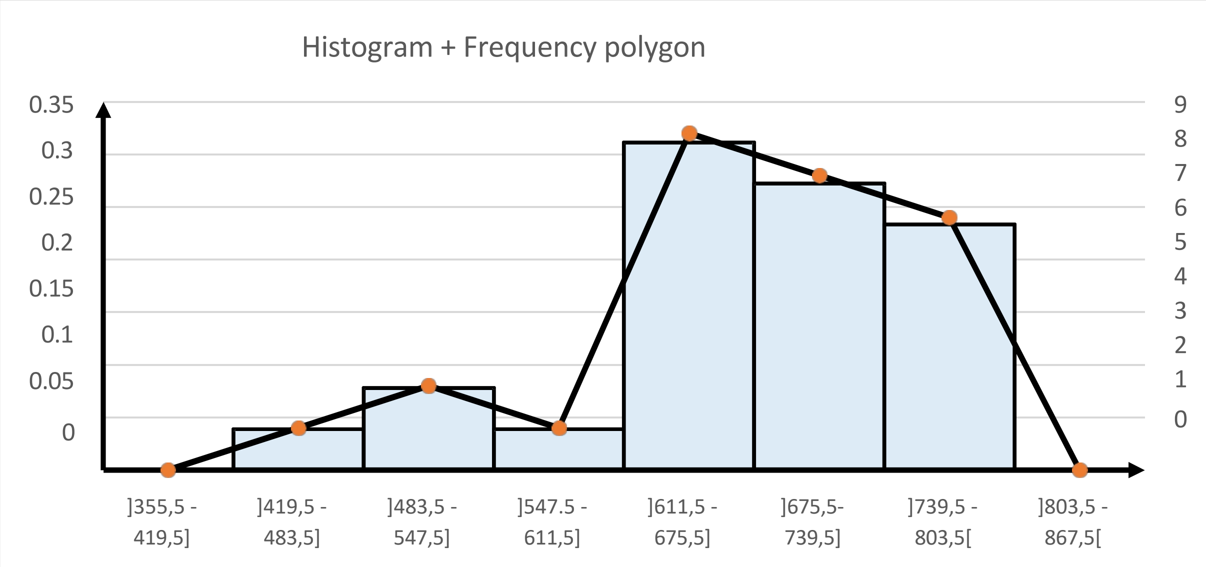


Figure 1: Cumulative increasing and decreasing frequency curve

Precipitation	Workforce cumulative	cumulative frequency croissant	Precipitation	Workforce cumulative	frequency cumulative decreasing
<419.5	0	0	>419.5	25	1
<483.5	1	0.04	>483.5	24	0.96
<547.5	3	0.12	>547.5	22	0.88
<611.5	4	0.16	>611.5	21	0.84
<675.5	12	0.48	>675.5	13	0.52
<739.5	19	0.76	>739.5	6	0.24
<803.5	25	1	>803.5	0	0

Table 4: Calculation of cumulative frequencies

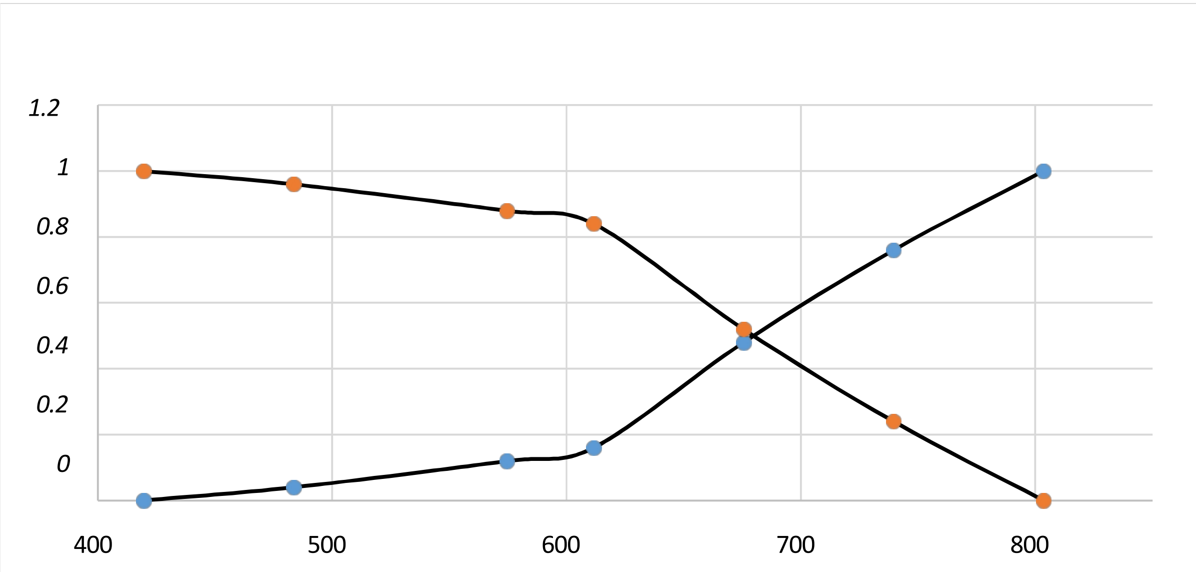


Figure 2: Curve of increasing and decreasing frequencies

In table N°4, we have calculated the cumulative frequencies up to the limits of the intervals

The sum of the frequencies of all values smaller than the upper limit of an interval is called the cumulative non-exceedance frequency (CNF): for example, 48% of the annual precipitation considered is less than 675.5 mm.

In addition, the sum of the frequencies of all values greater than the lower limit of an interval is called the cumulative exceedance frequency (FD): for example, 52% of the annual precipitation in our sample is greater than 675.5 mm.

FND +FD= 48%+ 52%= 100

A) Central parameters

There are three core values:

- mode
- the median
- the average

The mode or dominant value: is the most frequent value in a distribution. In our case, this value is 645mm.

The median

It can only be calculated for quantitative characteristics. With values ranked in ascending order, the median is the value of the characteristic that divides it into two sets of equal numbers: 50% of values are above it and 50% are below it.

$$Mediane = L_1 + \left(\frac{\frac{N}{2} (-) \left(\sum f_{im} \right)}{f_{ediane}} \right) * C$$

Where:

Middle class is that corresponding to N/2 =12.5 L_1: lower limit of

middle class= 675.5 mm

$\sum f_i$ Sum of absolute frequencies of all classes below class

median: 12

f_{median}: Frequency of the median class= 7

C: median class size= 64

In our case, the median is 680.07 mm.

The arithmetic mean

$$\bar{X} = \frac{\sum P_i}{N}$$

In our case the arithmetic mean equals $\bar{X}$ =682.08 mm

B) Dispersion parameters

The summary of a distribution given by a central value does not tell us anything about the dispersion of values around this central value, i.e. whether they tend to concentrate or disperse around it.

1. Standard deviation

$$\sigma Pi = \sqrt{Variance} = \sqrt{\frac{1}{N}\sum_{i=1}^{N}\left(P_i - \bar{P}\right)^2} = 90.92 \text{ mm}$$

2. Variance

$$Variance = \frac{1}{N}\sum_{i=1}^{N}\left(P_i - \bar{P}\right)^2 = 8312.82 \text{ mm}$$

3. Coefficient of variation

$$C_v = \frac{\sigma Pi}{\bar{P}} = 0.133 \text{ mm}$$

Frequency analysis

Frequency analysis is a statistical method of prediction which involves studying past events, characteristic of a given process (hydrological or other), in order to define the probability of future occurrences.

This prediction is based on the definition and implementation of a frequency model, which is an equation describing the statistical behavior of a process. These models describe the probability of occurrence of an event of a given value.

Frequency analysis calls on a variety of statistical techniques and is a complex process that needs to be handled with the utmost rigor. Its various stages can be very simply diagrammed as follows:

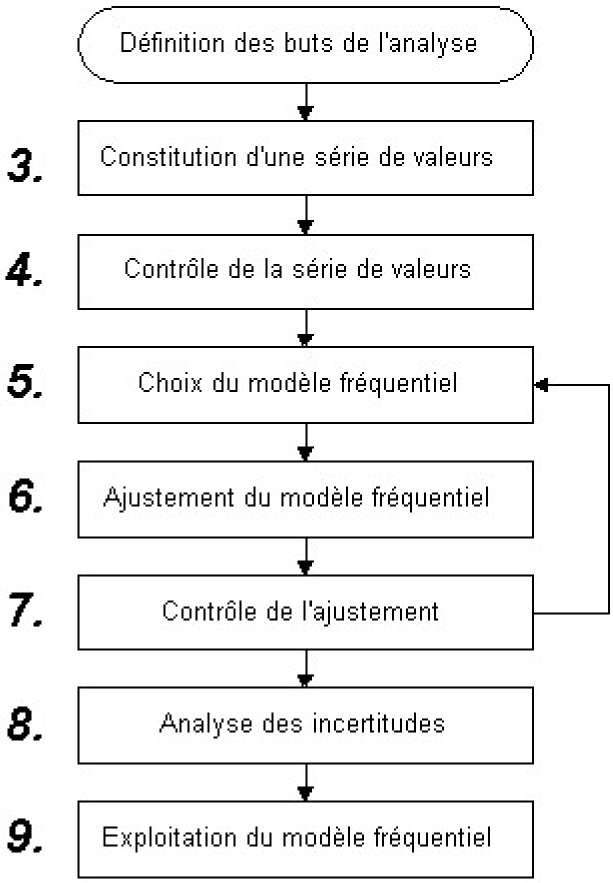


Figure 3: Main steps in frequency analysis.

Frequency model selection

The validity of the results of a frequency analysis depends on the choice of frequency model, and more specifically on its type. There are a number of ways to facilitate this choice, but unfortunately there is no universal, infallible method.

Normal law

The central limit theorem justifies the normal distribution as the distribution of a random variable formed by the sum of a large number of random variables. In extreme-value frequency hydrology, however, distributions are not symmetrical, which is an obstacle to its use. However, this law is generally well suited to the study of annual moduli of hydrometeorological variables in temperate climates.

Lognormal distribution

The lognormal law is advocated by some hydrologists, including V.-T. Chow, who justifies it by arguing that the occurrence of a hydrological event results from the combined action of a large number of factors which multiply . From then the random variable $X = X_1 \cdot X_2 \cdot \dots \cdot X_r$ follows a lognormal distribution. In fact, the product of r variables is reduces to the sum of r logarithms of these, and the central limit theorem asserts the log-normality of the random variable.

Gumbel's law

E.-J. Gumbel postulated that the double exponential law, or Gumbel's law, is the limiting form of the distribution of the maximum value of a sample of n values. Since the annual maximum of a variable is considered to be the maximum of 365 daily values, this law must be able to describe series of annual maxima.

It should be noted that the greater the number of parameters in a law, the greater the uncertainty in the estimate. In practice, it is therefore preferable to avoid using laws with three or more parameters.

Frequency model adjustment

In this part of the chapter, we'll be looking at techniques for fitting or calibrating a frequency model to a series of data: this involves defining the parameters of the chosen law. We'll use Gumbel's law, frequently used in hydrology, to model extreme events such as rainfall.

Introducing Gumbel's law

The distribution function of the Gumbel distribution is expressed as follows:

$$F(x) = \exp\left(-\exp\left(-\frac{x-a}{b}\right)\right). \qquad (1)$$

Where a is the position parameter, b the scale parameter

Let's assume the following reduced variable $u = \frac{x-a}{b}$ (2)

The distribution is then written as follows: $F(x) = \exp\left(-\exp(-u)\right)$ (3)

and $u = -\ln\left(-\ln\left(F(x)\right)\right)$ (4)

The advantage of using the reduced variable is that the expression of a quantile is then linear. In

Indeed, to find the value x_q of a quantile, corresponding to the distribution $F(x_q) = q$, as a function of the two parameters a and b it is sufficient to use the following relationship $x_q = a + bu_q$

Consequently, as soon as the points of the series to be fitted can be plotted on a system of x - u axes, it is possible to fit a straight line that best passes through these points and deduce the two parameters a and b of the law. There are various fitting methods:

graphical method (adjustment by eye or using statistical regression), method of moments, etc.

In practice, it's essentially a question of estimating the probability of non-exceedance F(xi) that should be attributed to each value xi. There are numerous formulas for estimating the distribution function using the empirical frequency. They are all based on sorting the series by ascending values, associating each value with its rank r. Simulations have shown that for Gumbel's law, Hazen's empirical frequency should be used:

$$\frac{r-0.5}{n} \quad (5)$$

where r is the rank in the ascending data series, n is the sample size, x[r] the value of rank r.

Remember that the return time T of an event is defined as the inverse of the frequency of occurrence of the event. Let :

$$T = \frac{1}{1-F(x_i)} \quad (6)$$

Approach and results :

Step 1: Preparing the peak flow data set.

* Sort values in ascending order.
* Assign a rank to each value.

Step 2: Calculation of the empirical frequency for each rank (Hazen, equation (5)).

Step 3: Calculate the reduced Gumbel variable "u" (equation (4)).

Step 4: Graphical representation of the pairs (ui, xi) of the series to be adjusted

Step 5: Fit a linear type relationship to the (ui, xi) pairs (figure 1) and deduce the two parameters a and b). With a graphical fit (by eye), we then have an estimate of the parameters a and b

Exercise

For catchment area X you are asked :

1) Fit the annual maximum precipitation series to a Gumbel distribution.

Adjust data graphically.

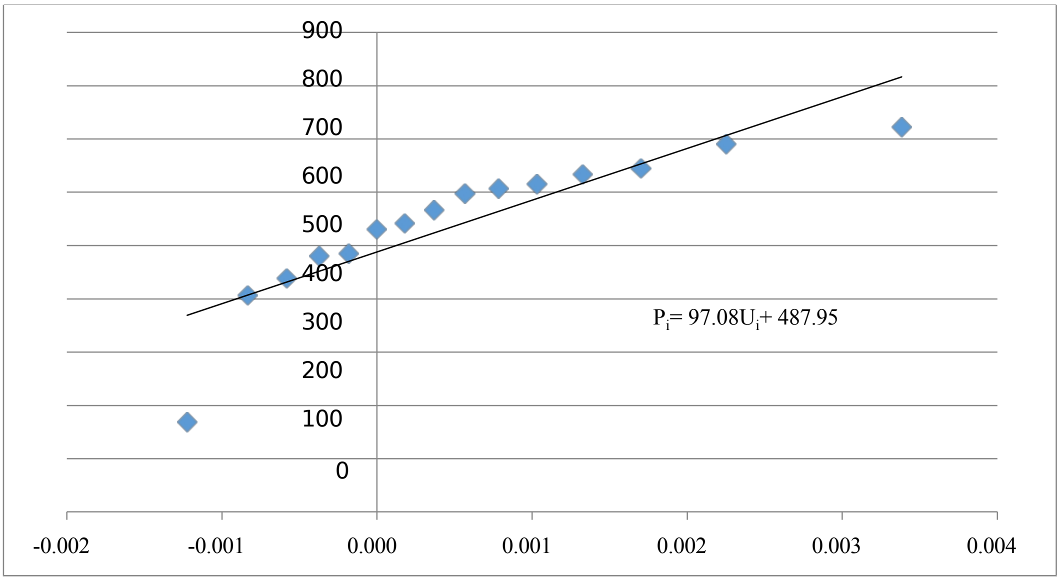
1) Estimate annual rainfall with return periods of 5, 20, 50 and 100 years.

Year	Precipitation (mm)
2005	607
2006	169
2007	407
2008	439
2009	480
2010	485
2011	531
2012	542
2013	567
2014	598
2015	615
2016	634
2017	645
2018	691
2019	723

Table 5: Average annual precipitation values for a watershed X

Answer

Precipitation	P in ascending order	rank	Frequency (fxi)	reduced variable Ui
607	169	1	0,0333	-1,224
169	407	2	0,1000	-0,834
407	439	3	0,1667	-0,583
439	480	4	0,2333	-0,375
480	485	5	0,3000	-0,186
485	531	6	0,3667	-0,003
531	542	7	0,4333	0,179
542	567	8	0,5000	0,367
567	598	9	0,5667	0,566
598	607	10	0,6333	0,784
615	615	11	0,7000	1,031
634	634	12	0,7667	1,325
645	645	13	0,8333	1,702
691	691	14	0,9000	2,250
723	723	15	0,9667	3,384



Return period T	100	50	20	5	2
F(xi)	0,99	0,98	0,95	0,8	0,5
variable Ui	4,6001	3,9019	2,9702	1,4999	0,3665
P (mm)for T	934,48	866,70	776,25	633,51	523,48

2 Presentation of the Normale law (Laplace Gauss law)

Normal distributions are very important in statistics. The curve representing their density function is called a Gauss curve or bell curve, because of its shape. It has an axis of symmetry at the mean or median.

The variable used is continuous, i.e. it can take on an indefinite number of values. This curve has two

parameters: μ and σ.

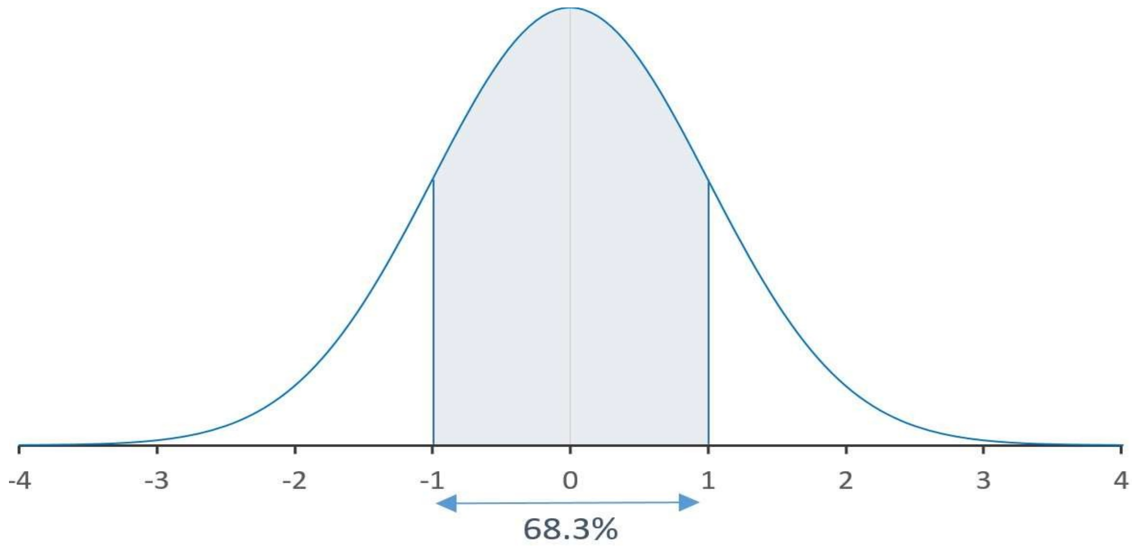
$$p(x) = \frac{1}{\sigma\sqrt{2\pi}}\, e^{\left(-\frac{(x-\mu)^2}{2\sigma^2}\right)}$$

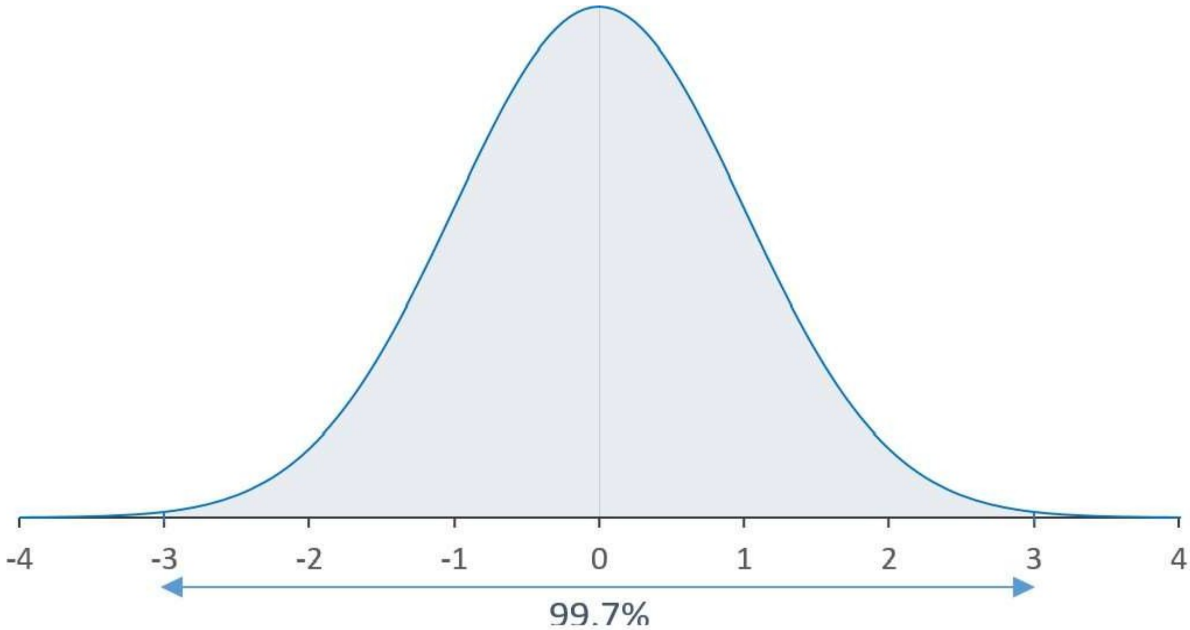
Its equation is ultimately very simple, since these 2 parameters alone suffice: μ and σ. The other elements of the equation are constants: Euler number *e* (2.71828) and Pi *π* (3.14159)

Mathematicians have simplified matters by calculating areas under a special normal distribution with parameters μ=0 and σ=1. This distribution is known as the reduced-centered normal distribution.

$$P(x) = \frac{1}{\sqrt{2\pi}}\, e^{-\frac{1}{2}U^2}$$

U is then said to be distributed according to a normal distribution with mean zero and variance 1.



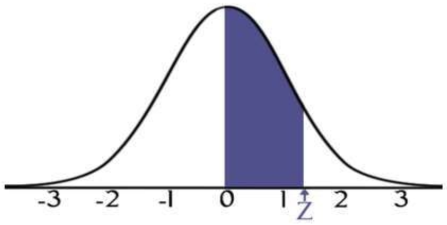


The two figures represent reduced-centered normal curves, with the areas between U = -1 and U = +1 and U = -2 and U = +2 and U = -3 and U = +3 plotted as "equal to 68.27%, 95.45%, 99.73% of the total area" respectively.

The following table shows the areas included or the normal curve between u = 0, and all positive values of U. From this table, we can obtain the area between any two values by using the symmetry of the curve with respect to U = 0.

Use of the center-reduced normal distribution table

Tables generally consist of a diagram and a table.



STANDARD NORMAL TABLE (Z)

Entries in the table give the area under the curve between the mean and *z* standard deviations above the mean. For example, for *z* = 1.25 the area under the curve between the mean (0) and *z* is 0.3944.

z	0.00	0.01	0.02	0.03	0.04	0.05	0.06	0.07	0.08	0.09
0.0	0.0000	0.0040	0.0080	0.0120	0.0160	0.0190	0.0239	0.0279	0.0319	0.0359
0.1	0.0398	0.0438	0.0478	0.0517	0.0557	0.0596	0.0636	0.0675	0.0714	0.0753
0.2	0.0793	0.0832	0.0871	0.0910	0.0948	0.0987	0.1026	0.1064	0.1103	0.1141
0.3	0.1179	0.1217	0.1255	0.1293	0.1331	0.1368	0.1406	0.1443	0.1480	0.1517
0.4	0.1554	0.1591	0.1628	0.1664	0.1700	0.1736	0.1772	0.1808	0.1844	0.1879
0.5	0.1915	0.1950	0.1985	0.2019	0.2054	0.2088	0.2123	0.2157	0.2190	0.2224
0.6	0.2257	0.2291	0.2324	0.2357	0.2389	0.2422	0.2454	0.2486	0.2517	0.2549
0.7	0.2580	0.2611	0.2642	0.2673	0.2704	0.2734	0.2764	0.2794	0.2823	0.2852
0.8	0.2881	0.2910	0.2939	0.2969	0.2995	0.3023	0.3051	0.3078	0.3106	0.3133
0.9	0.3159	0.3186	0.3212	0.3238	0.3264	0.3289	0.3315	0.3340	0.3365	0.3389
1.0	0.3413	0.3438	0.3461	0.3485	0.3508	0.3513	0.3554	0.3577	0.3529	0.3621
1.1	0.3643	0.3665	0.3686	0.3708	0.3729	0.3749	0.3770	0.3790	0.3810	0.3830
1.2	0.3849	0.3869	0.3888	0.3907	0.3925	0.3944	0.3962	0.3980	0.3997	0.4015
1.3	0.4032	0.4049	0.4066	0.4082	0.4099	0.4115	0.4131	0.4147	0.4162	0.4177
1.4	0.4192	0.4207	0.4222	0.4236	0.4251	0.4265	0.4279	0.4292	0.4306	0.4319
1.5	0.4332	0.4345	0.4357	0.4370	0.4382	0.4394	0.4406	0.4418	0.4429	0.4441
1.6	0.4452	0.4463	0.4474	0.4484	0.4495	0.4505	0.4515	0.4525	0.4535	0.4545
1.7	0.4554	0.4564	0.4573	0.4582	0.4591	0.4599	0.4608	0.4616	0.4625	0.4633
1.8	0.4641	0.4649	0.4656	0.4664	0.4671	0.4678	0.4686	0.4693	0.4699	0.4706
1.9	0.4713	0.4719	0.4726	0.4732	0.4738	0.4744	0.4750	0.4756	0.4761	0.4767
2.0	0.4772	0.4778	0.4783	0.4788	0.4793	0.4798	0.4803	0.4808	0.4812	0.4817
2.1	0.4821	0.4826	0.4830	0.4834	0.4838	0.4842	0.4846	0.4850	0.4854	0.4857
2.2	0.4861	0.4864	0.4868	0.4871	0.4875	0.4878	0.4881	0.4884	0.4887	0.4890
2.3	0.4893	0.4896	0.4898	0.4901	0.4904	0.4906	0.4909	0.4911	0.4913	0.4916
2.4	0.4918	0.4920	0.4922	0.4925	0.4927	0.4929	0.4931	0.4932	0.4934	0.4936
2.5	0.4938	0.4940	0.4941	0.4943	0.4945	0.4946	0.4948	0.4949	0.4951	0.4952
2.6	0.4953	0.4955	0.4956	0.4957	0.4959	0.4960	0.4961	0.4962	0.4963	0.4964
2.7	0.4965	0.4966	0.4967	0.4968	0.4969	0.4970	0.4971	0.4972	0.4973	0.4974
2.8	0.4974	0.4975	0.4976	0.4977	0.4977	0.4978	0.4979	0.4979	0.4980	0.4981
2.9	0.4981	0.4982	0.4982	0.4983	0.4984	0.4984	0.4985	0.4985	0.4986	0.4986
3.0	0.4987	0.4987	0.4987	0.4988	0.4988	0.4989	0.4989	0.4989	0.4990	0.4990
3.1	0.4990	0.4991	0.4991	0.4991	0.4992	0.4992	0.4992	0.4992	0.4993	0.4993
3.2	0.4993	0.4993	0.4994	0.4994	0.4994	0.4994	0.4994	0.4995	0.4995	0.4995
3.3	0.4995	0.4995	0.4995	0.4996	0.4996	0.4996	0.4996	0.4996	0.4996	0.4997
3.4	0.4997	0.4997	0.4997	0.4997	0.4997	0.4997	0.4997	0.4997	0.4997	0.4998

Example of fitting a normal distribution to a sample

For the watershed you are asked :

1) Fit the series of maximum annual flows according to a Gumbel distribution.

Adjust data graphically.

1) Estimate annual rainfall with return times of 5, 20, 50 and 100 years.

Year	Precipitation (mm)
2005	607
2006	169
2007	407
2008	439
2009	723
2010	531
2011	485
2012	634
2013	567
2014	615
2015	598
2016	542
2017	645
2018	691
2019	480

Answer

We propose to fit a normal distribution (Gauss distribution) to a given sample of annual rainfall. The steps to be followed are as follows

1- Calculation of empirical characteristics :

- Arithmetic mean $\bar{P}$
- Standard deviation σ

2- Value ranking :

The values in the sample are sorted in ascending or descending order, with each value assigned a sequence number starting from 1.

3- Calculation of experimental frequency

$$(f)\ (x)=\frac{n_i-0.5}{N}$$

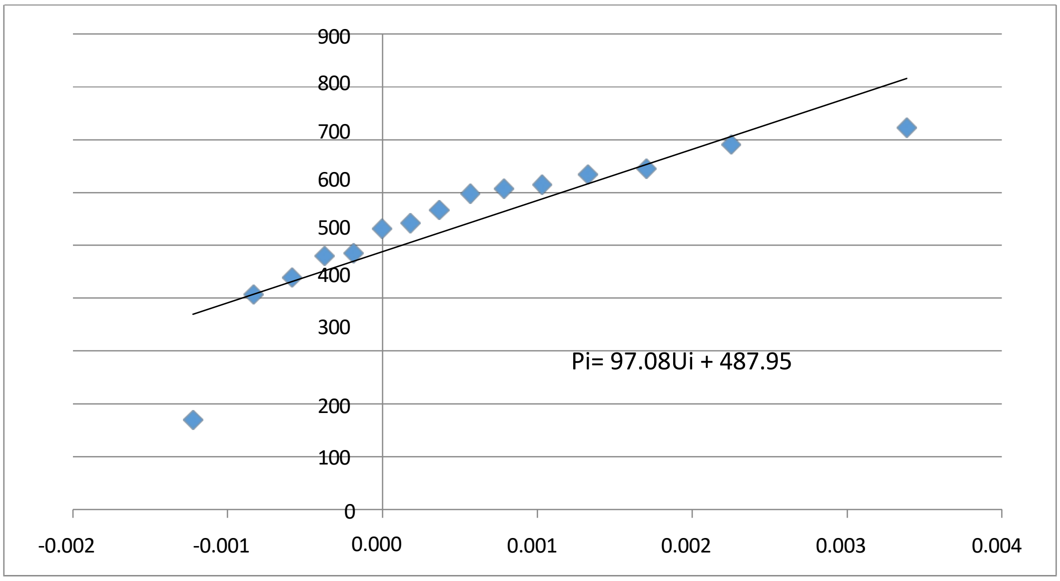
4- We deduce Ui values from the tables using the empirical frequencies F (x)

5- Henry curve: Pi= a Ui+b

6- Using this model Pi= a Ui+b we deduce all probable precipitation for different return times

The following table summarizes all calculation operations

Precipitation (mm)	P in ascending order	rank	Frequency f(x)	reduced variable U
607	169	1	0,0333	-1,224
169	407	2	0,1000	-0,834
407	439	3	0,1667	-0,583
439	480	4	0,2333	-0,375
480	485	5	0,3000	-0,186
485	531	6	0,3667	-0,003
531	542	7	0,4333	0,179
542	567	8	0,5000	0,367
567	598	9	0,5667	0,566
598	607	10	0,6333	0,784
615	615	11	0,7000	1,031
634	634	12	0,7667	1,325
645	645	13	0,8333	1,702
691	691	14	0,9000	2,250
723	723	15	0,9667	3,384



Pmoy	542,2 mm
standard deviation	136,93 mm

Return period T	100	50	20	5	2
F(xi)	0,99	0,98	0,95	0,8	0,5
Gauss variable Ui	4,6001	3,9019	2,9702	1,4999	0,3665
Precipitation (mm) for different T	934,48	866,70	776,25	633,51	523,48

Chapter 2: Statistical and probabilistic study of precipitation

Chapter 2Statistical and probabilistic study of precipitation

Analysis and representation of rainfall data at a station Climatological study :

The aim of the climatic study is to investigate the climatic parameters that determine surface runoff and flood generation.

In this chapter we will determine the climatic characteristic, i.e. :

1. Precipitation study :
   - Average annual rainfall
   - Monthly rainfall
   - Average seasonal rainfall
2. Temperature study :
   - Average annual temperature
   - Average monthly temperature
   - Monthly temperature variation by month ($T_{(MAX)}$, T_{MIN}, T Monthly)
3. evaporation :
4. Climate classification :
   - Rainfall-temperature method (relationship between precipitation and temperature)
   - General climate indices
     1. DEMARTONNE index (aridity)
     2. EMBERGER Index
5. Actual evapotranspiration (AET)
6. Evaporation potential (ETP)

Example: precipitation study :

Climatic conditions play a decisive role in r i v e r regimes, with precipitation playing a key role in feeding the watershed's runoff.

Average annual rainfall over 29 years (1984 to 2012):

The table shows the 29-year rainfall series from 1984 to 2012.

For this period, the maximum was observed in 1964 with 1066.0872 (mm), and the minimum in 2007 with 63.7105(mm).

We translate table (02) into a graph (figure 01) with years on the x-axis and annual precipitation on the y-axis.

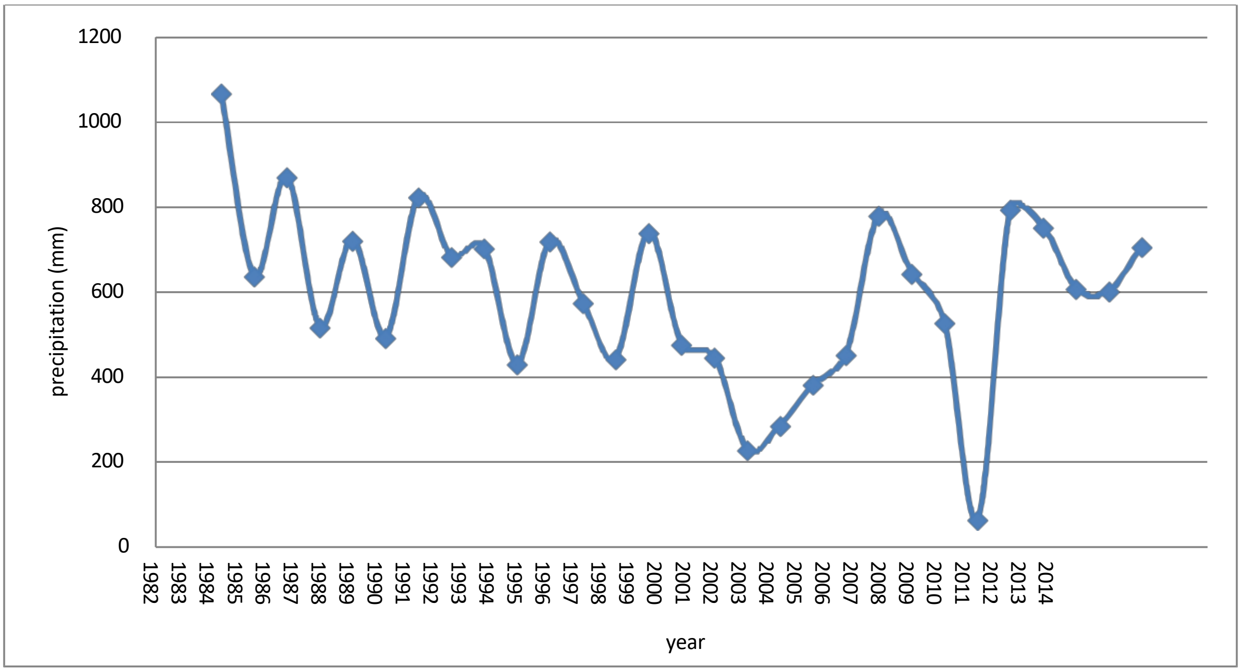


Figure 1: Distribution curve of annual precipitation (1984-2012) In table (02), we have shown the different mean annual precipitation.

Year	Pmoy ann (mm)	Year	Pmoy ann (mm)	Year	Pmoy ann (mm)
1984	1066,0872	1994	718,7669	2004	778,9
1985	635,8627	1995	573,5762	2005	641,9
1986	869,536	1996	441,7611	2006	526,6
1987	516,29934	1997	738,0683	2007	63,7105
1988	720,5634	1998	475,7240	2008	792,8
1989	491,438	1999	445,3618	2009	751,3
1990	822,1005	2000	228,09632	2010	607,1
1991	682,0016	2001	285,1771	2011	600,7
1992	702,18214	2002	381,805	2012	705,2
1993	429,5631	2003	451,3122		

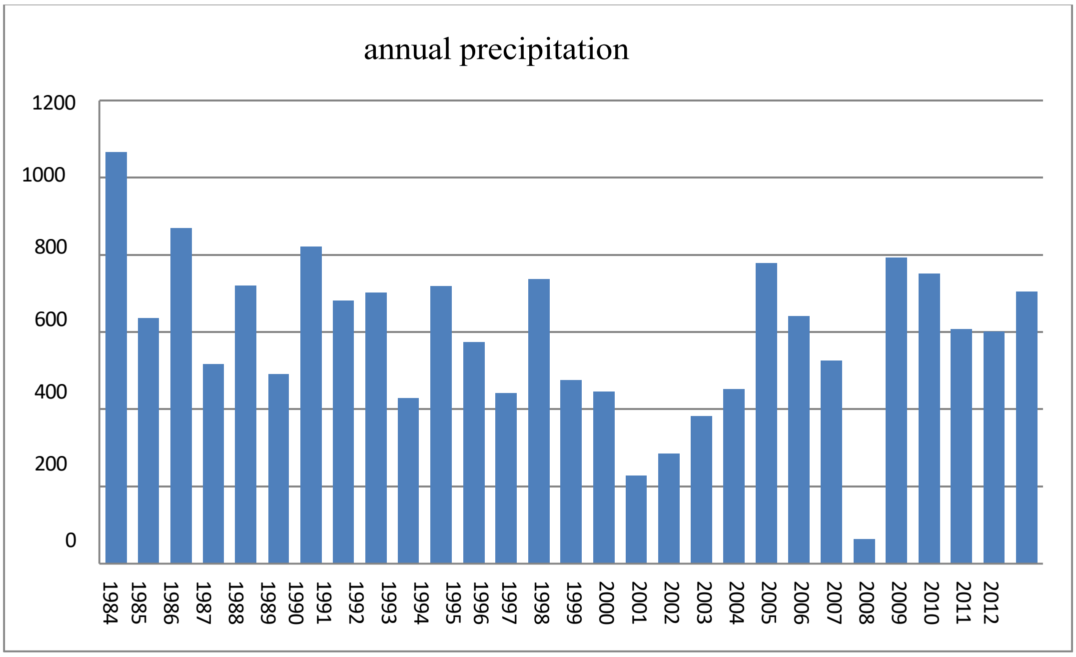


Figure 2: Histogram of annual precipitation (1984 to 2012) In table (03), we

have shown the different average monthly precipitation levels.

Month	SEP	OCT	NOV	DEC	JAN	FEV	MAR	AVR	MAY	JUN	JUIL	AOU	average
Monthly average	31,077 74	40,411	67,589 2	119,82 16	104,33 66	72,527 6	53,037 4	58,540 9	28,563 84	11,016 99	1,6809 52	2,5510 3	591,155

Table (02): Average monthly precipitation

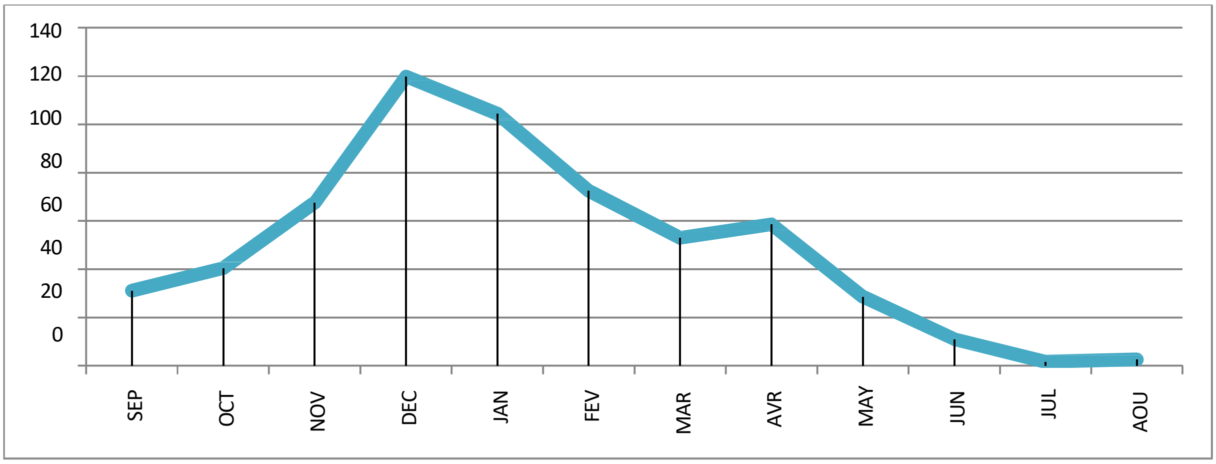


Figure 3: Monthly precipitation distribution curve

Average seasonal rainfall :
The table (n°04) shows the different seasonal precipitations.

Table N°04: Seasonal precipitation :

month	Autumn	Winter	Spring	Summer	Total
Seasonal average	139,07	296,68	140,14	15,24	591,154
%	23,52	50,18	23,70	2,579	100

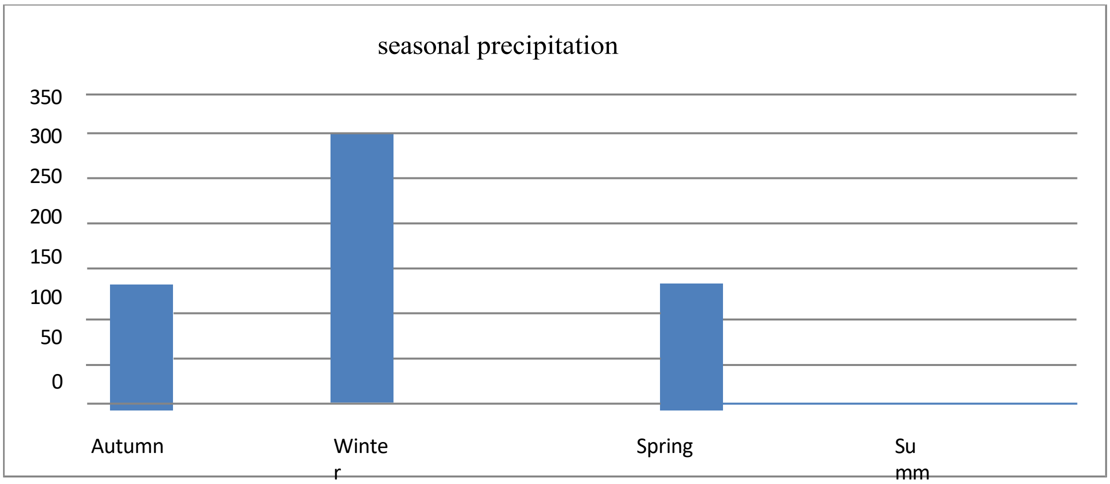


Figure 4:Histogram of seasonal precipitation

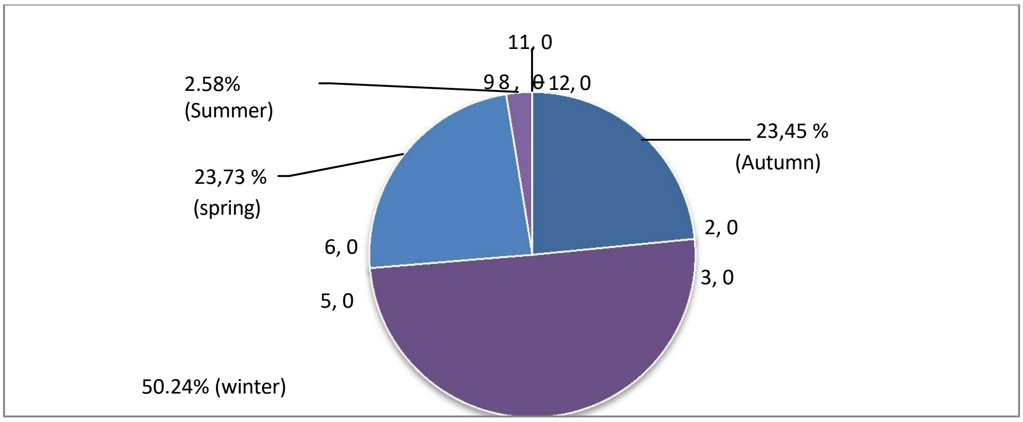


Figure 5: Percentage of seasonal precipitation

2. Temperature study :

Average monthly and annual temperatures have a direct impact on climate, interacting with other meteorological factors. Table (05) shows the temperature series over 29 years (1984 to 2012).

Year	SEP	OCT	NOV	DEC	JAN	FEV	MAR	AVR	MAY	JUN	JUL	AOU	Average
1984	25,31532	22,6863	16,671	13,176	12,922	13,879	14,5326	13,07085	19,46942	24,3808	28,2008	25,8076	19,176
1985	25,31752	22,7511	16,77	13,152	12,304	13,7692	14,5134	12,96865	20,73482	24,5796	28,9044	27,055	19,4017
1986	25,33182	22,8029	16,523	13,158	12,392	13,6228	14,5307	13,1475	19,75062	24,6648	29,048	27,8925	19,4054
1987	25,29002	22,4532	16,539	13,111	12,989	12,8725	14,5444	13,1621	20,74888	24,4944	28,4736	28,1598	19,4031
1988	25,31532	22,6475	16,688	13,129	12,498	13,696	14,4434	13,2205	20,73482	24,6506	28,5023	26,7342	19,3549
1989	25,32522	22,6345	16,523	13,075	12,392	14,0346	14,4406	13,1986	20,31302	24,5086	27,8418	26,9124	19,2666
1990	25,29772	22,6475	16,325	13,17	12,922	13,513	14,477	13,1621	19,25852	24,7358	27,1238	24,9701	18,9668
1991	25,29662	22,6993	16,671	13,18	12,41	12,8451	14,4643	13,21685	20,52392	24,2814	27,2674	26,164	19,085
1992	25,29882	22,6863	16,193	13,152	12,304	13,6228	14,4588	13,3081	19,32882	24,324	27,8418	25,8076	19,0272
1993	25,29992	22,7122	16,506	13,055	12,534	13,8058	14,3478	12,41385	19,39912	24,5796	27,77	26,9124	19,1113
1994	25,32192	22,7381	16,44	13,176	12,551	13,8516	14,4825	13,1402	20,80512	24,5654	33,0114	29,0152	19,9249
1995	25,3206	22,6397	16,506	13,091	12,47	13,879	14,4734	12,74965	20,20054	24,7528	28,0428	26,4313	19,213
1996	25,30322	22,7226	16,605	13,173	12,58	13,7601	14,3496	12,96135	19,9334	24,6392	27,612	26,6095	19,1873
1997	25,325	22,7485	16,387	13,091	12,887	13,513	14,4033	13,257	19,38506	24,5568	28,33	27,946	19,3191
1998	25,3294	23,7586	15,483	13,567	11,941	11,317	14,7555	13,549	16,9808	24,6108	27,0376	23,491	18,485
1999	25,3118	22,5413	15,582	13,448	11,941	8,8465	14,7009	13,549	23,7296	24,7812	25,458	29,5498	19,1199
2000	25,3382	23,4219	16,209	13,257	11,023	11,317	14,6008	13,111	22,3236	24,2132	27,4684	25,0948	18,9482
2001	25,303	22,5672	15,965	13,085	12,099	13,1653	14,477	13,2497	20,2849	24,3126	28,9762	26,6095	19,1745
2002	25,3272	23,0515	16,209	13,029	12,064	13,9888	14,5662	13,28985	19,70844	24,2842	28,7895	27,3223	19,3025
2003	25,325	23,2587	16,295	13,192	11,397	11,7379	14,498	13,01975	18,3165	24,2018	27,3966	23,7761	18,5345
2004	25,35	24,65	15,93	13,83	13,34	13,93	14,65	15,95	18,56	21,98	24,43	26,66	19,105
2005	24,92	22,57	16,61	11,83	9,77	8,91	17,97	19,3	22,19	24,76	28,43	27,56	19,5683
2006	25,55	24,31	18,05	13,73	9,91	11,32	14,17	19,51	23,37	26,29	30,11	27,65	20,3308
2007	26,43	21,66	15,89	11,87	13,58	14,83	14,55	17,81	21,77	25,9	29,31	30,17	20,3142
2008	27,22	23,37	16,67	13,09	13,04	12,2	13,8	17,41	21,99	23,38	29,97	31,55	20,3075
2009	25,73	21,86	16,9	14,73	12,64	12,33	15,4	15,41	25,52	26,84	30,07	30,11	20,6283
2010	26,3	20,96	15,86	13,33	10,64	13,84	14,93	17,02	19,7	23,83	28,5454	24,62	19,1313
2011	25,3558	22,9661	15,958	13,227	12,25	11,8	14,6873	13,8541	19,35694	23,9377	27,4397	27,7322	19,0471
2012	24,27	23,04	14,48	13,358	12,56	10,85	11,2	14,84	18,56	21,98	24,43	23,25	17,7349
Monthly average	25,428257	22,8122	16,325	13,188	12,219	12,7947	14,5316	14,27068	20,44644	24,4488	28,1321	26,9505	19,2957
MAX	27,22	24,65	18,05	14,73	13,58	14,83	17,97	19,51	25,52	26,84	33,0114	31,55	22,2884
MIN	24,27	20,96	14,48	11,83	9,77	8,8465	11,2	12,41385	16,9808	21,98	24,43	23,25	16,7009

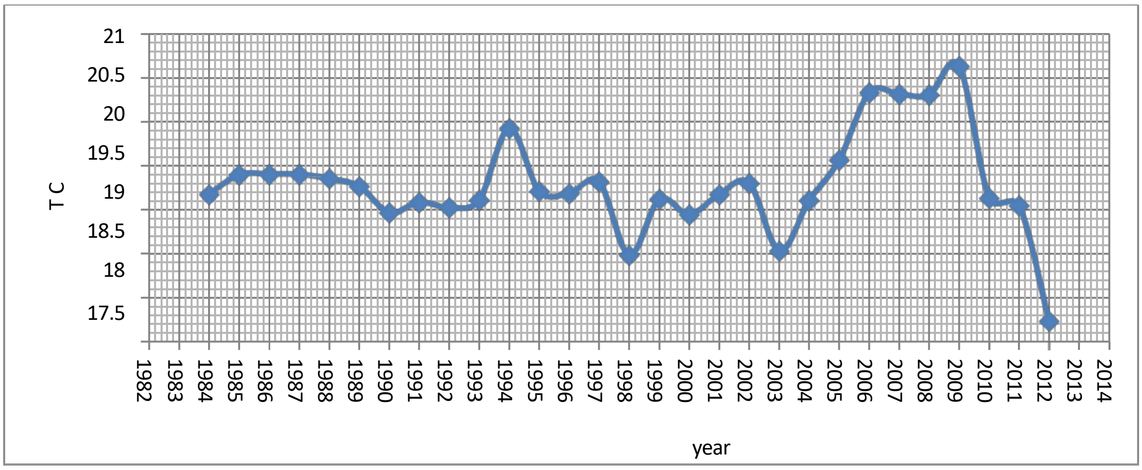


Figure 6: Average maximum and minimum temperatures and average monthly temperatures.
Table (06) shows average maximum and minimum temperatures, as well as average monthly temperatures.

Year	SEP	OCT	NOV	DEC	JAN	FEB	MAR	APR	MAY	JUN	JUL	AUG	Average
Monthly average	25,42	22,81	16,32	13,18	12,21	12,79	14,53	14,27	20,44	24,44	28,13	26,95	19,29
MAX MIN	27,22	24,65	18,05	14,73	13,58	14,83	17,97	19,51	25,52	26,84	33,011	31,55	22,28
	24,27	20,96	14,48	11,83	9,77	8,845	11,2	12,41	16,98	21,98	24,43	23,25	16,70

Exploration of table (06) shows that the coldest month is January (with 12.219°C) and the hottest is July (with 28.1321°C). The average annual temperature is (16.7009°C).

Figure 4 shows the months on the x-axis, and the temperatures (°C) on the y-axis.

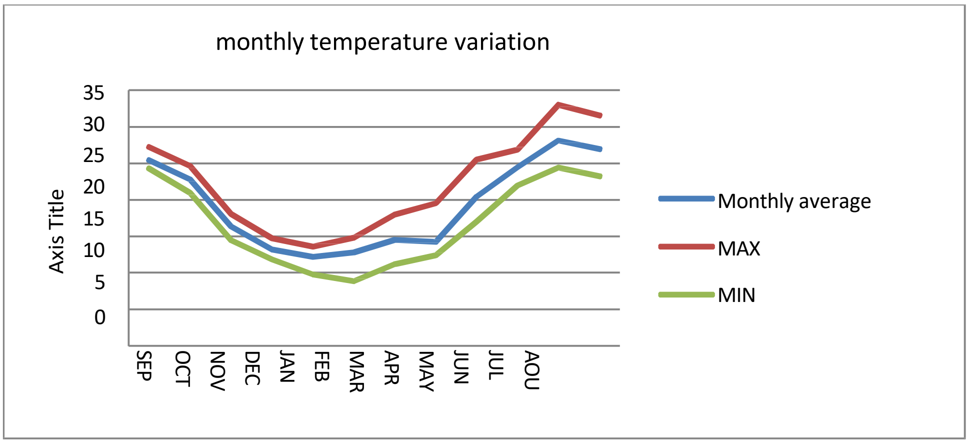


Figure 7: Temperature distribution curve (T_{Min}- T_{Max}- T monthly)

Climate classification :

Rainfall-temperature method :

According to GAUSSEN and BAGNOULS, a month is said to be dry if the average total precipitation is less than or equal to twice the average temperature *(P_2T)*.

This relationship makes it possible to draw up rainfall-temperature diagrams on which the temperature is plotted on a double scale with that of rainfall.

- If the temperature curve rises above the precipitation curve, we have a dry month.
- If the temperature curve falls below the precipitation curve, we have a wet month.

Table (7) shows average monthly temperatures and average monthly rainfall.

month	SEP	OCT	NOV	DEC	JAN	FEV	MAR	AVR	MAY	JUN	JUL	AOU	average
P (mm)	31,07	40,41	67,58	119,82	104,33	72,52	53,03	58,54	28,56	11,01	1,68	2,551	591,15
T C	25,42	22,812	16,3é	13,18	12,21	12,77	14,53	14,27	20,44	24,44	28,13	26,95	19,29

The rainfall diagram is shown in the following figure.

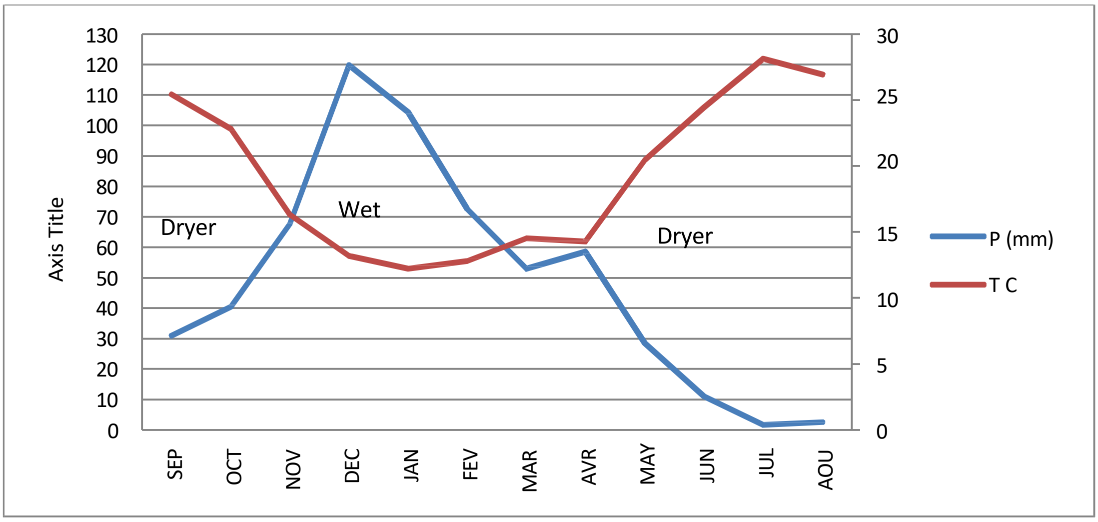


Figure 8: Rainfall curve

In table (08), you can see the difference between the months (Nov to Apr) with a humid climate and the months (May to Oct) with a dry climate.

month	SEP	OCT	NOV	DEC	JAN	FEV	MAR	AVR	MAY	JUN	JUL	AOU
P(mm)	31,07	40,41	67,58	119,82	104,33	72,52	53,03	58,54	28,56	11,01	1,68	2,55
T C	25,428	22,81	16,32	13,18	12,21	12,79	14,53	14,27	20,44	24,44	28,13	26,9
Rolation (T-P)	arid	arid	wet	wet	wet	wet	wet	wet	arid	arid	arid	arid

Arid climate

General climatic indices :

DEMARTONNE index (aridity):

DEMARTONNE in 1933 introduced an aridity index

$$I = \frac{P}{T+10}$$

I: DE MARTONNE aridity index.

P: mean annual precipitation (mm).

T: mean annual temperature (°C). From table (7):
The different values of I correspond to different types of climate I < 5 : the climate is hyper-arid;
5< I< 7.5 : desert climate;
7.5< I< 10: steppe climate; 10 < I <20:

semi-arid climate; 20 < I <30: temperate

climate.

P = 591.15496 (mm) T =

19.295663 (C°)

A DEMARTONNE aridity index is obtained.

I= 20.17892361

According to the distribution given by DEMARTONNE, the aridity index (I) is between 20 and 30, 20<I<30. The use of DEMARTONNE's aridity index chart (figure I-07) shows that our region has a temperate climate.
Indice D'EMBERGER :
D'EMBERGER index is calculated using the following formula:

$$Q = \frac{P}{\left(\frac{M+m}{2}\right)*(M-m)}*100$$

Q: EMBERGER index (rainfall-earthquake quotient)
P: average annual precipitation in (mm).
M: average maximum temperature of warmest month in (°K)
m: average minimum temperature of the coldest month in (°K)
P = 591,154955
T M = 31.55 C°+273 = 304,55 K° T m = 9.77 C°+273 = 282,77 K° Q = 9,24

Real evapotranspiration (ETR) :

Actual evapotranspiration refers to the loss of water in the form of water vapour, and is a complex phenomenon that combines the physical evaporation of water contained in the soil with transpiration processes, corresponding to the use of atmospheric water and water in the atmosphere by plants.

Runoff deficit is defined as the difference between precipitation and runoff at watershed scale. To achieve an acceptable estimate of runoff deficit values, we use empirical methods.
TURC method:

$$ETR = \frac{P}{\sqrt{0,9+\frac{P^2}{L^2}}}$$

AET: actual evapotranspiration (mm).

P: mean annual precipitation (mm).
L : 300+ 25T+ $0.05T^3$.
T: mean annual temperature (°C).

T = 19.29566348 C° P=

591.154955(mm)

Therefore: L= 1141.602195

Then : ETR= 546.956255 (mm)

Potential evaporation:

The emission of water vapour, or evapotranspiration, considered as a
loss by hydrogeologists, takes place in all environments. It is the result of two phenomena, one physical - evaporation - and the other biological - transpiration. Evaporation takes place in the atmosphere during rainfall on the surface of lakes and rivers, and also in bare soil.
.evapotranspiration is due to the presence of vegetation cover.

THORNTWAITE method :

The THORNTWAITE formula can be used to determine a potential evapotranspiration (PET) for each plant.

$$FTE = 1.6 \left(10 \frac{T}{I}\right)^{a}$$

ETP: Evapotranspiration potential (cm).
T: Mean annual temperature (°C).
I: sum of monthly thermal indices for the year.

month	SEP	OCT	NOV	DEC	JAN	FEV	MAR	AVR	MAY	JUN	JUL	AOU
T(C°)	25,43	22,81	16,32	13,18	12,21	12,79	14,53	14,273	20,44	24,44	28,13	26,95
I	11,73	9,955	5,99	4,344	3,86	4,14	5,02	4,89	8,43	11,05	13,67	12,81

$$I = \left(\frac{T}{5}\right)^{1,514}$$

$I = \sum_{I}^{12} in$

T= 19.29566348 C° I =

95.941611342

a = $6.75*10^{-7}*I^3 - 7.71*10^{-5} + 0.4989 + 1.792*I$ a =

2.0104591987

TheTHORNTWAITE formula

ETP= 212.2814631 (Cm)

A. Study of rainfall series homogeneity

The term "inhomogeneities" refers to unnatural variations caused by changes in observation networks. Homogenization consists in detecting and then correcting these variations. For example, precipitation measurements are affected by wind and the Venturi effect produced above the rain gauge cone. So any modification affecting the wind will induce jumps in the data (a change in location, sensor shape or height above ground). A change in the immediate environment of a station, induced for example by urbanization, reforestation, etc., will also have an impact on the data.

In practice, it is quite difficult to determine whether a break in a series represents a change in regional climate or inhomogeneity. There are two types of information that can be used to support either explanation:

Records from neighboring stations (if any): data from one station are often compared with those from neighboring stations, to avoid climate change being interpreted as inhomogeneity.

The series to be tested for homogeneity is called the base series. Neighboring series are assumed to be climatically similar to the base series (often observations from geographically neighboring stations). They must also be homogeneous, otherwise inhomogeneities in one of them could be attributed to the base series.

Metadata (if any): Is historical information about the conditions under which data at a station was recorded. It consists of station recordings, meteorological yearbooks, inspection sheets, photographs of the station and its environment, an interview with the person in charge of a station, etc.

Correcting for inhomogeneities in climate series is crucial for all studies involving climate variables. In particular, high-quality precipitation series will enable decisions to be made based on reliable data in various climate change research studies.

Correcting errors

a) The double totals method

The principle of the method is to check the proportionality of values measured at two stations. One of the stations (station X) is the base or reference station, assumed to be correct. The other station (Y) is the station to be checked. The method consists of the following steps:

- Calculate cumulative precipitation for suspect station X

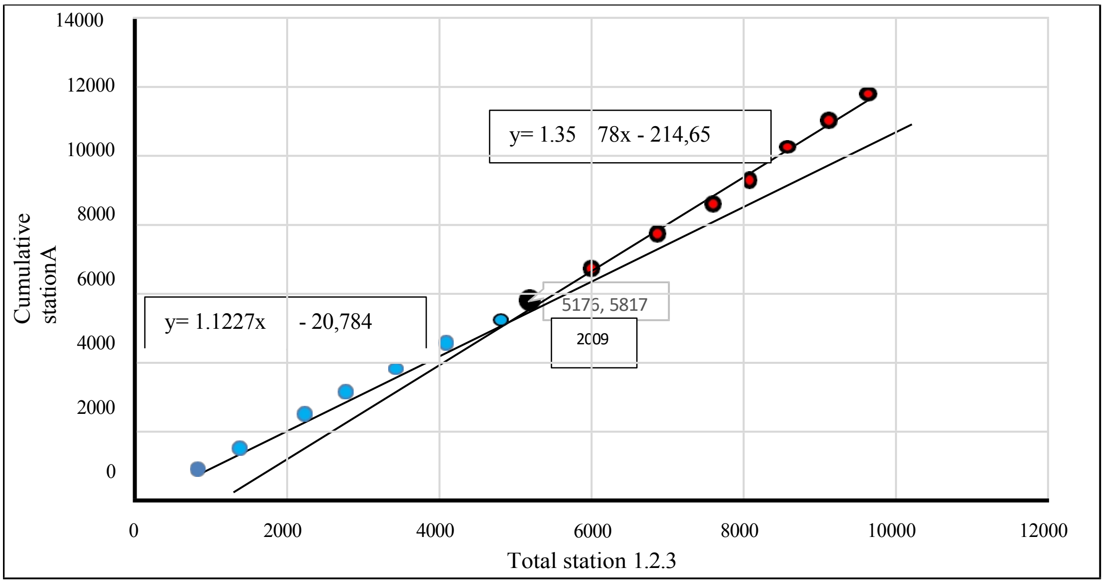
- Calculate the mean annual rainfall and add up the averages for stations 1, 2, 3 and 4, n

- Plot the cumulative precipitation for the suspect station against the cumulative precipitation for the reference station. If several reference stations exist, a base station can be formed from the average of the data recorded at these stations.

The principle of the test is that a change due to meteorological causes will not change the slope of the curve, since neighboring stations will be affected. Only changes caused by systematic errors at the station to be controlled would lead to a change in the slope of the graph.

Example:

Year	Station X	Station 1	Station 2	Station3	Average station 1,2 and 3	Cumulative stationX	Cumulative Avg. Station 1.2 and 3
2002	869	800	850	750	800	869	800
2003	596	543	553	551	549	1465	1349
2004	994	856	860	858	858	2459	2207
2005	643	546	536	538	540	3102	2747
2006	736	655	656	660	657	3838	3404
2007	734	674	679	678	677	4572	4081
2008	699	702	704	700	702	5271	4783
2009	546	389	394	396	393	5817	5176
2010	953	806	830	824	820	6770	5996
2011	882	837	842	844	841	7652	6837
2012	945	728	731	737	732	8597	7569
2013	694	459	458	460	459	9291	8028
2014	875	516	526	524	522	10166	8550
2015	849	546	536	538	540	11015	9090
2016	791	511	507	515	511	11806	9601



On the graph, we can see that the points are aligned on two different line segments, i.e. there was a break in the line during 2009. It is assumed that the displacement (or other cause of error) occurred in 2009. Data measured after 2009 are considered good, and only the previous data (2002;2003,2004,2005,2006,2007,2008 and 2009) need be corrected.

Calculate the slopes m1 for the segment on the right containing data from 2016 to 2009, and m2 for the segment containing data from 2008 to 2002. The ratio of the slopes m1/m2 is calculated, and used to multiply the data from 2009 to 2002 in order to correct them. The results are as follows:

$$slope_e^{Corrected} = \frac{m_1}{m_2} slope^{Measured}$$

m1: slope of the portion of the graph at correct, *m2*: slope of the reliable of the graph

$$P_{corrected} = 0.8268 P_{measured}$$

Station A corrected
869
596
994
643
736
734
699
451,4328
787,9404
729,2376
781,326
573,7992
723,45
701,9532
653,9988

Several statistical tests are used to check the homogeneity of a statistical series

Statistical tests

Almost all tests (Mann-Whitney test, Kruskal-Wallis-Wilcoxon test, etc.) assume that the distribution of the random variable under study follows the normal distribution. Since this condition is not always satisfied, all these tests are called non-parametric because they do not require the estimation of the mean and variance of even the values xi , but only their rank in the ordered series of all values. In this part of the course we will study the Wilcoxon test.

Wilcoxon test :

The principle of this test is as follows: If sample X comes from the same population Y, sample XUY (union of X and Y) also comes from it.

The procedure is as follows:

Given a series of observations of size N from which we draw two samples X and Y: N1 and N2 are respectively the sizes of these samples, with N=N1+N2 and N1≤N2.

We then rank the values in our series in ascending order, and will only be interested in the rank of each of the elements in the two samples in this series. If a value is repeated more than once, we associate the corresponding average rank with it.

We then calculate the sum $W_x$of the ranks of the elements of the first sample in the common series:

$$W_x = (\sum) Rank\ x$$

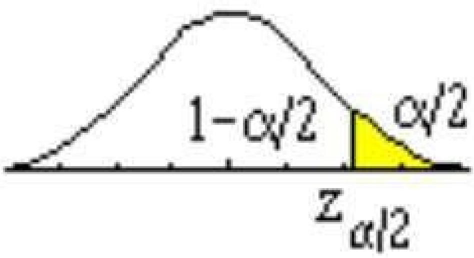
Wilcoxon showed that , in the case where the two samples X and Y constitute a homogeneous series , the quantity $W_x$lies between two limits $W_{max}$and W_{min}, given by the following formulae :

$$W_{min} = \frac{(N1 + N2 + 1)N1 - 1}{2} - Z_{1-\alpha/2}\sqrt{\frac{N1N2(N1 + N2 + 1)}{12}}$$

$$W_{max} = (N1 + N2 + 1)\,N1 - W_{min}$$

$Z_{(1)(-)(\alpha)(/2)}$represents the value of the centred reduced variable of the normal distribution corresponding to $(1)(-)(\alpha)(/2)$

(at the 95% confidence level where α =5%), we have $Z_{(1)(-)(\alpha)(/2)}$))



We will use the Wilcoxon test to check the homogeneity of the annual precipitation data from the Bordj Boou Naàma station (Tissemssilt wilaya) at the 5% singnification level.

The data are shown in the table below:

Year	P(mm)	Year	P(mm)	Year	P(mm)	Year	P(mm)	Year	P(mm)	Year	P(mm)
1	641.2	8	582.1	15	780.2	22	953.8	29	340.7	36	570.3
2	659.1	9	827.3	16	685.1	23	801.7	30	819.5	37	758.5
3	1176.9	10	530.4	17	500.9	24	709.8	31	391.6	38	550.5
4	557.1	11	1125.3	18	1030.3	25	519.8	32	618.8	39	522.2
5	367.5	12	659	19	898.7	26	1006	33	720.7	40	416.1
6	410.5	13	787.7	20	1085.4	27	838.5	34	712.2		
7	1014.8	14	641.8	21	588.7	28	826	35	458.5		

To make the calculations easier, we start by dividing our rainfall series into two samples of respective lengths, N1 = 18 values and N2 = 22 values.

In the first column, we enter the first sample X; in the second column, we enter the second sample Y; in the third and fourth columns, we enter respectively the ranks and the classified values of the original series and, in the fifth column, the origin of the value in the series, i.e. we note whether it comes from the X or Y sample.

1	2	3	4	5	
X	Y	Rank	XUY	Origin	Rank X
641.2	898.7	1	340.7	Y	
659.1	1085.4	2	367.5	X	2
1176.9	588.7	3	391.6	Y	
557.1	953.8	4	410.5	X	4
367.5	801.7	5	416.1	Y	
410.5	709.8	6	458.5	Y	
1014.8	519.8	7	500.9	X	7
582.1	1006	8	519.8	Y	
827.3	838.5	9	522.2	Y	
530.4	826	10	530.4	X	10
1125.3	340.7	11	550.5	Y	
659	819.5	12	557.1	X	12
787.7	391.6	13	570.3	Y	
641.8	618.8	14	582.1	X	14
780.2	720.7	15	588.7	Y	
685.1	712.2	16	618.8	Y	
500.9	458.5	17	641.2	X	17
1030.3	570.3	18	641.8	X	18
	758.5	19	659	X	19
	550.5	20	659.1	X	20
	522.2	21	685.1	X	21
	416.1	22	709.8	Y	
		23	712.2	Y	
		24	720.7	Y	
		25	758.5	Y	
		26	780.2	X	26
		27	787.7	X	27
		28	801.7	Y	
		29	819.5	Y	
		30	826	Y	
		31	827.3	X	31
		32	838.5	Y	
		33	898.7	Y	
		34	953.8	Y	
		35	1006	Y	
		36	1014.8	X	36
		37	1030.3	X	37

		38	1085.4	Y	
		39	1125.3	X	39
		40	1176.9	X	40
				Sum Row X= 380	

$\sum Rows\ (x) = 380$

$W_{min} = 296.4$

$W_{max} = 441.6$

Given that $Z_{1-\frac{1}{\alpha}}$ =1.96 for a significance level α= 5%

We check the ingality: $W_{(min)} (\Sigma rang\ (x)(W_{max}$

In other words: 296.4<380<441.6

The ingality is therefore verified and our series is therefore homogeneous.

Chapter 3: Streamflow studies

Introduction:

How are river flows measured? What is a hydrometric
station?

A hydrometric station is a device for observing and measuring the water level or flow in a watercourse.

Hydrometry is the term used to describe all the techniques used to measure the various parameters that characterise the flow of natural or artificial watercourses and pipes. The two main variables that characterise flow are :

- The dimension de la surface d'eau libre, notée H et exprimée en mètre. Sa mesure concerne la

water level measurement.

- The flow rate of a watercourse, noted Q and expressed in m3/s or l/s, representing the total volume of water flowing through a straight section of the watercourse during the unit of time considered. It is measured by flowmetering.

The water level in a channel is easy to observe, but is only representative of the section under observation and may be subject to changes over time. Only the flow variable physically reflects the behaviour of the catchment area, and can be interpreted in time and space.

The levels of rivers, lakes or reservoirs are used directly to forecast run-off, to delimit areas at risk of flooding and to design hydraulic structures. Through their relationship with river flows or the volumes of water contained in reservoirs and lakes, water levels constitute the basic information for determining flows or stocks. The criteria for choosing the location of the station must reflect the final objective of the observations and the accessibility of the site, taking into account the geometric and hydraulic properties of the reach. Hydraulic conditions are also an important factor in the choice of site along watercourses, particularly when water levels are used to calculate flows. Stations on lakes and reservoirs are normally located close to the outlets, but far enough upstream to avoid the influence of the phenomenon of lowering of the water level due to the increase in velocity.

Concept of instantaneous flow

Flow - the volume of water passing through a section of watercourse in one unit of time - is expressed in cubic metres per second (m^3/s).

$$Q=\frac{v}{t}$$

There are two types of method for measuring flow through a given section of water:

- the direct method: "Volumetric" methods (or capacitive gauging) allow the flow rate to be determined directly from the time required to fill a container of a given capacity with water. In view of the practical aspects inherent in this method of measurement (size of container required, uncertainty over the measurement of time, any special equipment required), this method is generally only used for very low flows, a few l/s at most.

- Indirect methods allow the flow rate to be determined by measuring another variable such as the water level or flow velocity.

Velocity field" methods involve determining the flow velocity at different points in the cross-section, while measuring the surface area of the wetted cross-section.

These techniques require specific equipment (reels, poles, salmon, current meters, etc.) and staff trained in their use.

Physico-chemical" methods take into account variations in certain physical properties of the liquid (concentration of certain dissolved elements) during flow. These methods generally involve injecting a substance in solution into the watercourse and monitoring changes in its concentration over time. These methods are known as "dilution" or "chemical" methods.

All these flow measurement methods generally require a river flow regime, except for chemical gauging, which is appropriate in the case of torrential flow.

2. Measuring water level

The most common way of determining flow is to measure the water level (head). The water level in a channel is easy to observe, but is only representative of the section under observation and may be subject to changes over time. Only the flow variable physically reflects the behaviour of the catchment, and can be interpreted in time and space. Generally speaking, we do not have a direct and continuous measurement of flow, but a recording of variations in water level in a given section (hydrometric station). The curve of water level versus time H=f(t) (called a limnigram) is then converted into the curve of flow Q=f(t) (called a hydrograph) by establishing a rating curve Q=f(H).

Once the rating curve has been established, the flow rate can be obtained at any time from the head of water observed in the measuring section.

The water level is measured discontinuously using water level gauges or continuously using water level gauges.

Limnimetric scales:

These are graduated rulers placed on the banks of rivers and wadis to mark the water level. A water level gauge requires an observer to take readings. Readings are taken 3 or 4 times a day during periods of low water (low flows) and every 10 minutes during periods of high water.

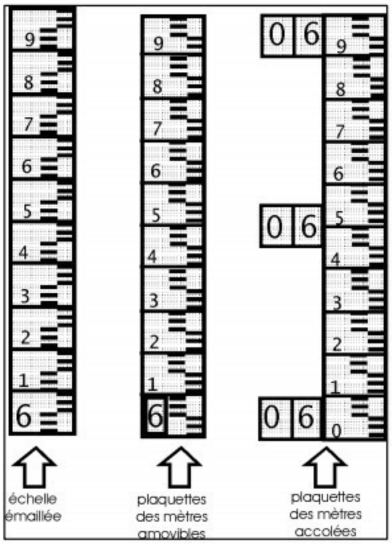


Figure 1: Echelle limnimétrique

Level gauges:

The float level gauge: this is a device that holds a float on the surface of the water by means of a counterweight, a cable and a pulley. The float follows water level fluctuations, which are plotted on a graph attached to a rotating drum (at a rate of one revolution per 24 hours, per week or per month).

Float gauges are easy to use, but are difficult to use in rivers with a high solid load because of silting up of the float well or the water intake on the river.

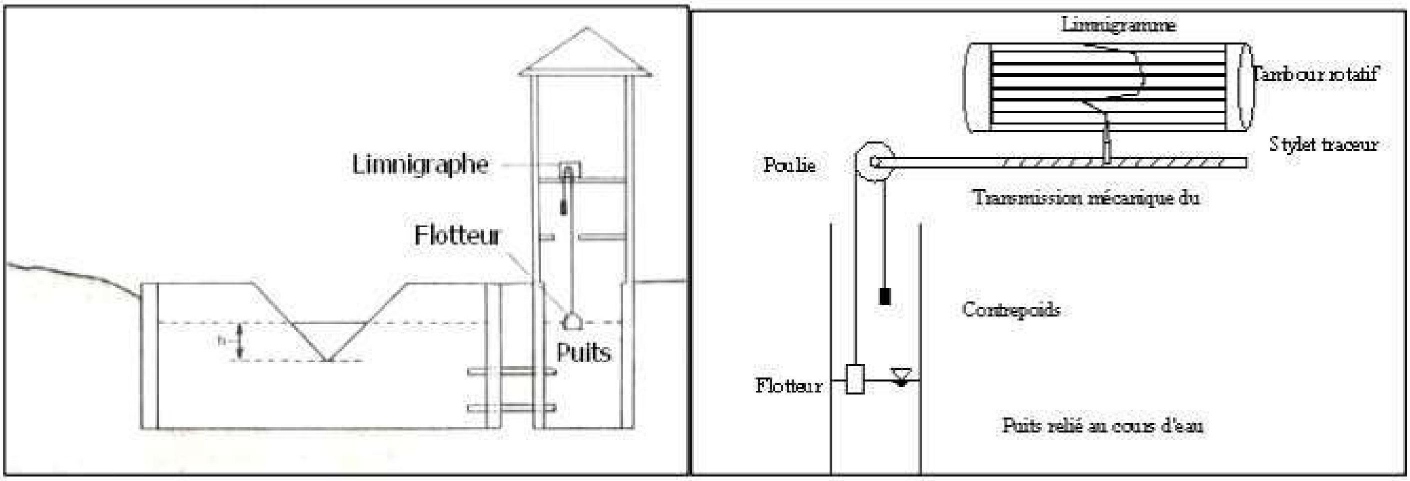


Figure 2 : Limnigraphe à flotteur

Flow measurement (gauging)

1. Gauging by exploring the velocity field

Flow can be measured by measuring the velocity at different points in the flow section.

Flow velocity is never uniform across the cross-section of a river. The principle of

The principle of this method is therefore to calculate the flow rate from the velocity field determined in a cross-section of the river (at a certain number of points, located along vertical lines).
points, located along verticals judiciously distributed across the width of the watercourse). In parallel with this exploration of the velocity field, the cross-sectional profile of the watercourse is recorded by measuring its width and taking depth measurements.
The flow rate Q in a flow section S of a river can be defined from the mean velocity V perpendicular to this section by the relationship :
Several methods can be used to determine the average velocity of the water.

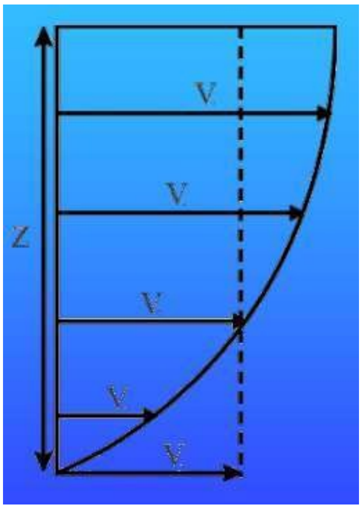
a. Gauging with a reel

The hydrometric reel is used to measure the point velocity of the flow. A reel has a rotating element (propeller) which, when immersed in the water, rotates at a speed that increases with the speed of the water. The flow speed is measured at each point from the speed of rotation of the propeller at the front of the reel (number of revolutions n per unit of time). The function V = f (n) is established by a calibration operation (reel calibration curve).

V= speed in m/s n= revolutions/second V = a +
b n

a= value of V for n= 0, and b= pitch of the rotating propeller

The number of measurements on a vertical line is chosen so as to obtain a good description of the speed distribution on this vertical line. As a general rule, 1, 3 or 5 measurements will be made depending on the depth of the bed.



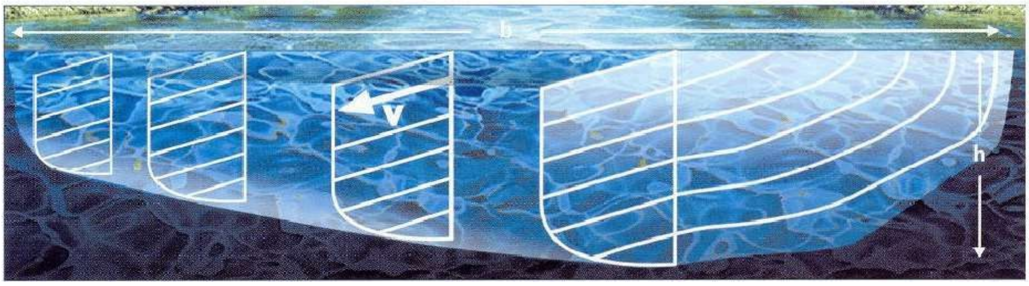


Figure 3: representation of the flow of a river

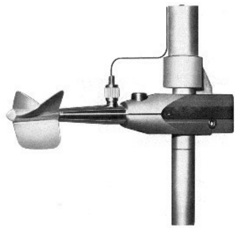


Figure 4: reel

b. Float gauging :

When reel gauging cannot be carried out because of excessive or too shallow velocities and depths, or the presence of material in suspension, it is possible to measure the flow velocity using floats. This method involves measuring only the velocities in the surface layer of the flow (the first 20 centimetres or so). Floats can be either artificial (plastic bottles) or natural (trees, large branches, etc.). The horizontal displacement of a surface float over a period of time t is used to determine the velocity of the surface flow. Several float velocity measurements must be taken. The average of these measurements is then multiplied by an appropriate coefficient to obtain the average velocity of the section element. In general, the average velocity in the section is of the order of 0.4 to 0.9 times the surface velocity.

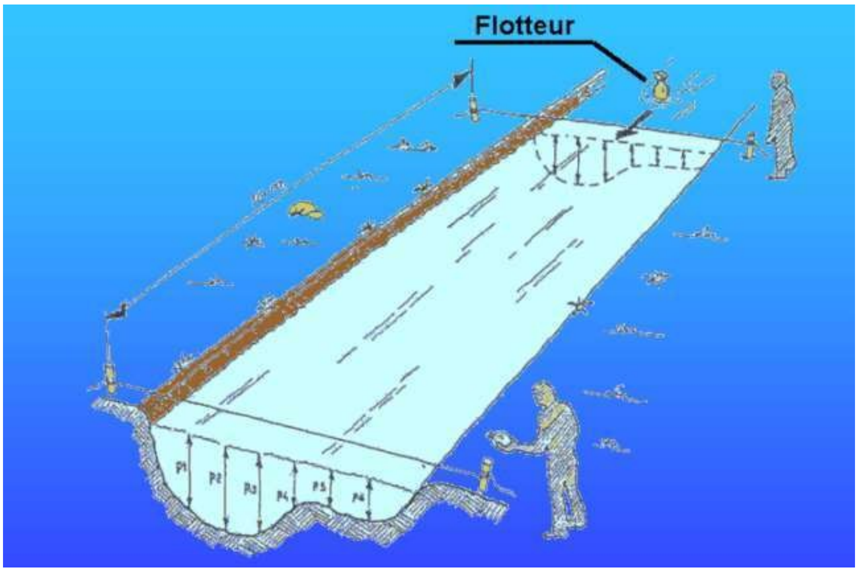


Figure 5: Float gauging

Calculating flow by exploring the velocity field :

The velocity field can be explored continuously or discontinuously. This is generally done in two different ways:

Point method:

At each fixed point on the chosen vertical, the velocity is measured for at least 30 seconds. Five points per vertical are desirable. Speeds are measured at 20%, 40%, 60% and 80% of depth, near the surface and near the bottom. The average speed over the entire section is calculated by graphical or numerical integration.

Integration method :

This consists of directly measuring an average speed per vertical section. To do this, the operator must let the speed-measuring device descend and then rise along a vertical line at a constant, slow speed. At the end of the operation, the operator will record the time and number of revolutions made by the device.

Profondeur de l'eau d (m)	Points de mesure de la vitesse	Vitesse moyenne (m/s)
0.3-0.6	0.2 d à partir de la surface	$V = V_{0.2}$
0.6-3	0.2 d et 0.8 d	$V = \frac{1}{2}(V_{0.2} + V_{0.8})$
3-6	0.2 d, 0.6 d et 0.8 d	$V = \frac{1}{4}(V_{0.2} + 2 \cdot V_{0.6} + V_{0.8})$
>6	surface, 0.2 d, 0.6 d, 0.8 d et fond	$V = \frac{1}{10}(V_s + V_{0.6} + 3 \cdot V_{0.8} + V_f)$

The average speed of the flow and therefore the flow rate can be calculated by graphical or numerical integration. The following methods can be used to estimate flow rate from velocity measurements:

Parabola method

Isotach method

Calculation example

Reel gauging carried out at the Sidi Aich station on the Soummam wadi on 5 October between 1.40 pm and 2.20 pm gave the following results:

Time	S_i	X_i	P(m)	N	t (sec)	n	V (m/s)	q_i	Obs
13h40	S_0	0.0							RD
	S1	1.0	0.03 0.10 0.28 0.36	168 165 125 0	25.9 25 25				Bottom
	S2	4.0	0.03 0.2 0.38 0.46	200 160 100 0	25 25 25				Background
	S3	7.0	0.03 0.20 0.35 0.43	219 207 153 0	25 25 25				Bottom
	S4	10	0.03	139	25				

			0.1 0.18 0.26	129 109 0	25 25				Bottom
	S5	12.0	0.03 0.075 0.155	37 28 0	25 25				Background
14h20		12.5							RG

S_i: section; X_i: abscissa of section S_i; P: depth of section S_i; N: number of turns of the reel; n=N/t: number of turns per second; V: water velocity; q_i: specific flow; RD: right bank; RG: left bank; Given that the reel calibration formula is :

$$V = 0.1319\, n + 0.032 \quad \text{for} \quad n \leq 2.93$$

$$V = 0.1360\, n + 0.020 \quad \text{for} \quad n > 2.93$$

Calculate the flow rate measured by this gauge

Help: the calculation steps are as follows:

1) Calculate the velocities using one of the above formulae
2) For each section, plot the depths on graph paper and the respective velocities on the horizontal axis.
3) The curve obtained is equal to the specific flow in (m2/s) for the section in question.
4) The specific flow rates found in b are plotted on graph paper as ordinates and the distances to the right bank of each section are plotted as abscissae.
5) The area under the curve is equal to the wadi flow Q
6) Determine the wetted section
7) Calculate the mean velocity of the flow

Answer:

The results of the calculations are shown in the following table:

IX – 1 – a - On commence par calculer la vitesse correspondant à chaque mesure en calculant d'abord le nombre de tours par seconde que fait le moulinet et ensuite la vitesse de l'eau en appliquant la formule d'étalonnage correspondante. On porte ces valeurs sur le tableau.

IX – 1 – b - Sur du papier millimétré on porte en abscisses les vecteurs vitesse et en ordonnées la profondeur correspondante pour chaque section afin d'obtenir le débit spécifique de qi de chaque section Si (voir Fig. 1).

IX – 1 – c - Sur une nouvelle feuille de papier millimétré on porte en abscisses les abscisses des verticales auxquelles ont été faites les mesures et en ordonnées les débits spécifiques qi trouvés en b.

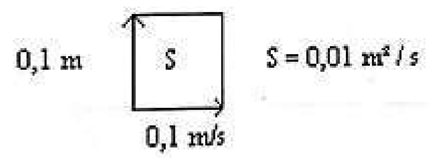
La surface sous la courbe représente le débit de l'oued pendant le jaugeage. On trouve Q = 38,84 m3/s.

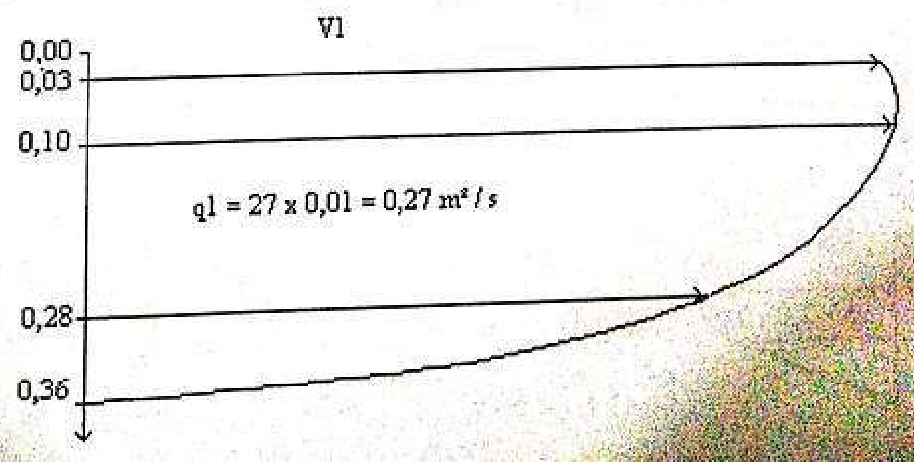
IX – 2 - La section mouillée est déterminée graphiquement (Fig. 2). On trouve S = 42,5 m².

IX – 3 - La vitesse moyenne de l'écoulement est:

V = Q / S = 38,84 / 42,5 = 0,92 m/s.

Echelle :





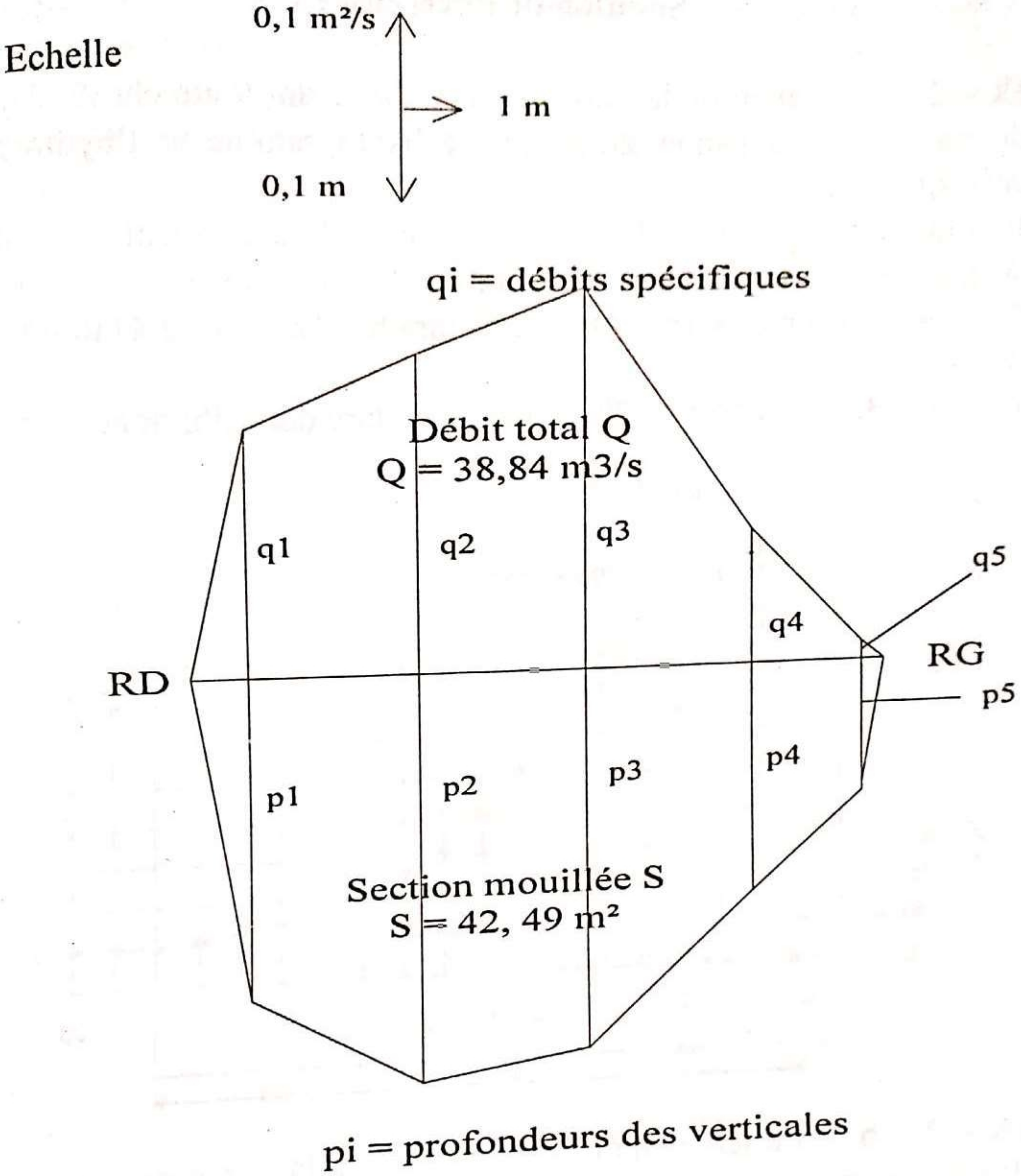


Fig 2 : Calcul du débit et de la section mouillée

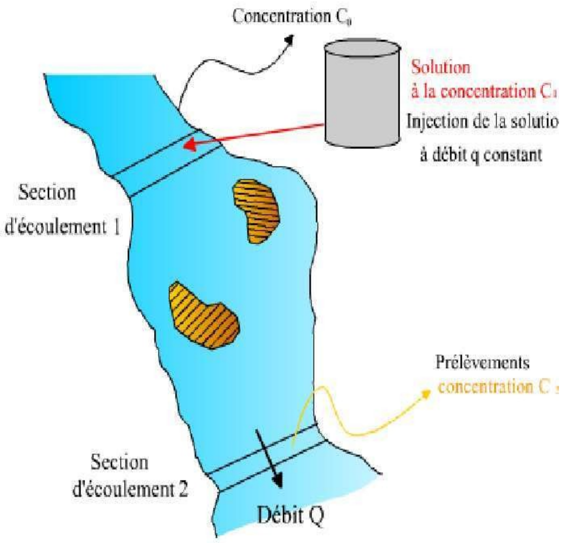
2. Determining the flow using calibrated structures

The purpose of building a weir or a calibrated channel to determine the flow of a watercourse is to obtain a relationship between the water level H and the flow Q that is as stable as possible, and in principle without gauging in the field. The flow rate is then obtained using hydraulic formulae and calibration on models. Gauging channels and calibrated weirs are used in particular in the case of small rivers with narrow, unstable, boulder-filled beds and shallow draughts, for which the installation of gauging stations and the use of reel gauges are not recommended. They operate according to the laws of classical hydraulics.

Fig: Triangular weir in minc34 wall and Venturi channel.

3. Dilution gauging

This method of gauging by dilution applies to torrents or steep rivers where the flow is turbulent or where there is no section suitable for reel gauging. The general principle is to inject a concentrated solution of a tracer (salt, dye, etc.) into the river and to find out how much of this solution has been diluted by the river, by taking water samples downstream of the injection point. *Fig. Principle of dilution gauging; procedure.*



The following conditions must be met before the integration or dilution methods can be applied:

- the flow of the river must remain more or less constant during the measurement
- the tracer must pass in its entirety through the sampling site
- At the sampling height, the mixture must be such that the same quantity of tracer passes through each point in the section of the river.

Different mineral or organic tracers are used, such as fluorescein or rhodamine. Depending on the flow to be assessed, the same tracers will not be used.

1. Constant flow injection method

The constant flow injection method consists of injecting a known constant flow *q* of a tracer solution, at concentration $C_{(1)}$ (stock solution), into the watercourse for a specified time. The duration of the injection must be such that the concentration C_2 of the tracer at the sampling section remains constant for a certain period of time, known as the "plateau". Based on the following assumptions :

Fig. Constant flow gauging.

- the flow rate Q in the watercourse is constant during the measurement (steady state),
- the flow rate q of the tracer at the sampling section is equal to that of the injection (no losses), and negligible compared with *Q*,
- the mixture is homogeneous at the sampling section,

then, assuming conservation of the tracer mass, we have :

$$Q = q \times \left(\frac{C_1}{C_2}\right)$$

2. Integration method (instantaneous injection)

This method involves injecting a volume *V* of tracer in concentrated solution C_1 into a point in the watercourse. At the end of a sufficiently long trajectory to ensure good mixing with the water in the river, samples are taken for the entire duration *T of* the tracer cloud's passage.

Samples are taken at several points in the sampling section so as to provide an average value for the concentration C_2, which changes as a function of time and sampling point.

Integrating the various concentration values $C_{(2)}(t)$ over time gives an average value $C(_{2)}(t)$.

$\overline{C_2}$

Assuming conservation of the tracer mass, the flow rate can be expressed as follows:

$$Q = \frac{M}{\int_0^T C_2(t)\,dt} = \frac{V \times C_1}{T \times \overline{C_2}}$$

With :

Q: stream flow [l/s or m3/s] ;

M: mass of tracer injected [g]; $M = V \,.\, C_1$;

V: volume of solution released into the watercourse [l or m3] ;

C_1: concentration of the solution released into the watercourse [g/l] ;

$\overline{C_2}$ average concentration of tracer in the samples, obtained by integration [g/l] ;

$C_{(2)}(t)$: concentration of the sample taken at time t [g/l];

T: sampling time [s].

Introduction to hydrological regimes

Records of a river's flow over a long series of years show systematic seasonal variations (position of high and low water) as a function of the main factors influencing the flow: rainfall regime, nature of the catchment area, geographical location, infiltration, etc. The hydrological regime of a river is defined by its flow regime. The hydrological regime of a watercourse summarises all its hydrological characteristics and how they vary. It is defined by the variations in its flow, usually represented by a graph of average monthly flow (calculated over a number of years and also known as "inter-monthly" flow or monthly modulus).

The figure below represents of the values of modules monthly mean of certain rivers around the world.

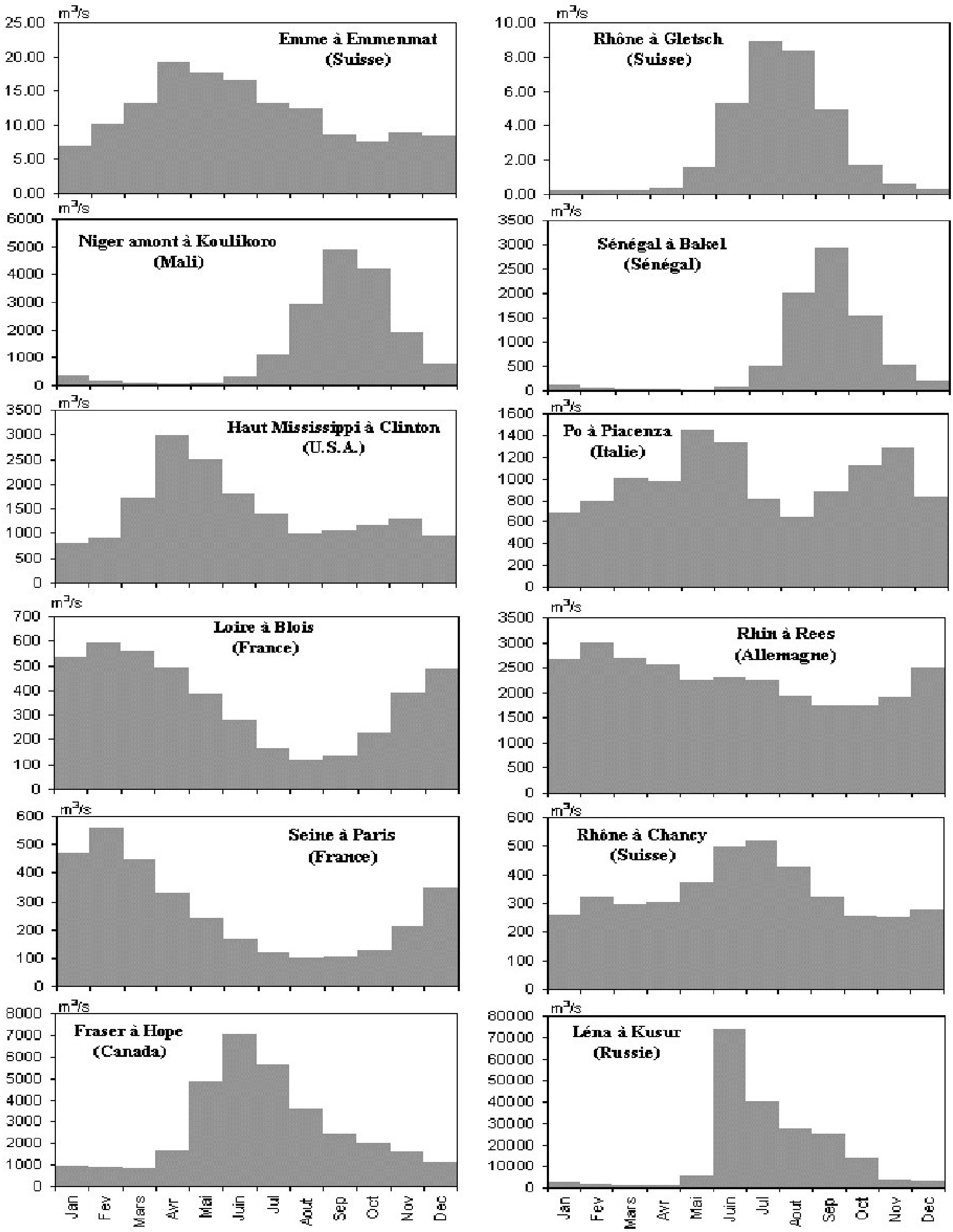


Fig. . Average flow rates (in m3/s) of some of the world's rivers

We also use the monthly flow coefficient, which is defined as the ratio of the average monthly flow to the inter-annual modulus (inter-annual average calculated over a number of years). This represents the percentage distribution of monthly flows over the year.

$$C_m(\%) = \frac{\textit{Débit mensuel moyen}}{\textit{Module Interannuel}} \cdot 100$$

The annual flow coefficient is also defined by the following ratio:

$$C_a(\%) = \frac{\textit{Lame moyenne écoulée}}{\textit{Pluie moyenne annuelle}} \cdot 100$$

The curve of monthly flow coefficients for the average year makes it possible to highlight the systematic nature of seasonal variations, and to compare rivers with one another. Knowledge of this coefficient is also of great interest for estimating the volumes flowing during a season in order to size a reservoir. Similarly, the curves of relative flow frequencies over a long series of years define the seasonal variation in flow quantiles (Fig. 3.10). The curves marked 10, 25,..., 90 % indicate the monthly flow values that have respectively 10, 25,..., 90 chances out of 100 of not being reached or exceeded.

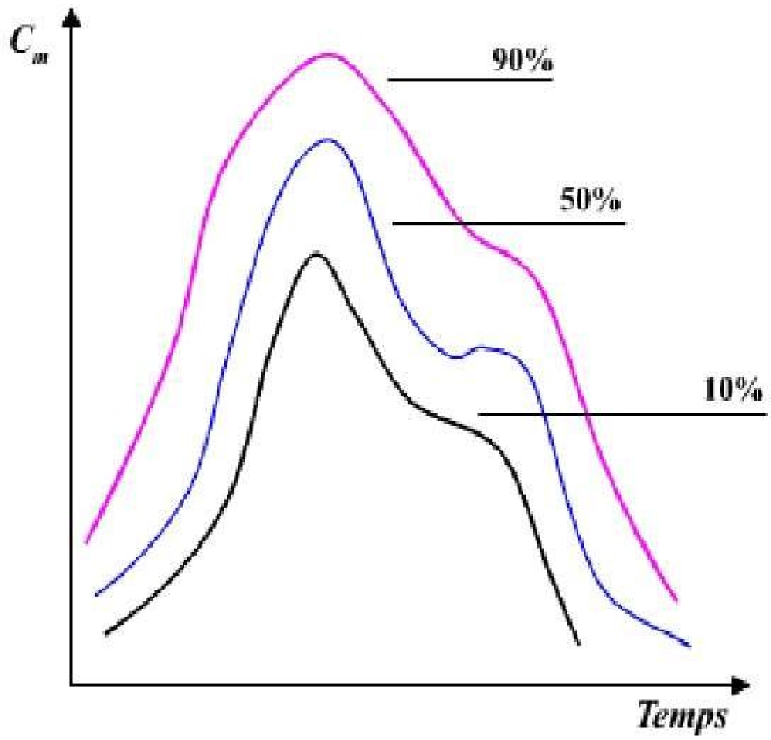


Fig. Example of frequency curves (frequency of non-exceedance) for monthly flows.

Classification of regimes Consequently, it is possible to characterise a catchment area and its flow by adopting a classification of the watercourse regime based on the pattern of the systematic seasonal fluctuation of the flows it presents, on the one hand, and on its mode of supply, on the other, i.e. the nature and origin of the high water (pluvial, snow or glacial). The monthly distribution of flows is then used to classify the flow regime of a watercourse called the hydrological regime. One of the simplest classifications of river hydrological regimes is that of Pardé (1933), who distinguishes three types of regime:

- Simple regime: characterised by a single annual alternation of high and low water (one monthly maximum and one monthly minimum in the course of the hydrological year) and, in general, by a single mode of supply
- Mixed system: 2 maxima and 2 minima per year, corresponding to several modes of supply.
- Complex regime: several extremes and supply modes.

Chapter 4: Study of flood flows

Introduction

The desire to investigate floods stems from their formidable impact when they reach a significant scale. An in-depth analysis of floods is useful for understanding their genesis, the various incidents that trigger them and the mechanisms involved in their formation.

Floods are caused by a wide range of events, the best known of which are intense or prolonged rainfall, the failure of dam structures, etc.

1. Basic information

1.1. Definition of a flood: a flood refers to an abundant quantity of water that is transported, with or without overflow, by a hydraulic system such as lakes and streams, which represents an exceptional event. Generally speaking, it is defined by its peak flow, its volume and its duration.

1.2. Why study floods?

- ✓ Management of the hydrological risk due to flooding (flood risk)
- ✓ Management of dependent surrounding systems (possibility of water supply, groundwater recharge);
- ✓ Management of the catchment areas from which the flooding originates;
- ✓ Management of the hydraulic system concerned (watercourse, tributaries and outlets, natural or artificial lakes, pools, depressions);

1.3. Events or Phenomena causing floods: Floods can be triggered by the following phenomena or events:

- ✓ Dam failure ;
- ✓ Extraordinary melting of snow or ice (due to equally extraordinary temperatures) or "normal" melting. extraordinary temperatures) or "normal" melting combined with other events
- ✓ extraordinary precipitation (liquid or solid) in intensity or duration;

1.4. Characteristics associated with floods :

A river floods when *its* "*critical flood level*" is reached at any point in the forebay. This threshold can be indicated by a fixed water level gauge

The maximum water level reached is the *flood peak*, corresponding *to* a *maximum flood flow* or *peak flow*. It is temporarily marked by a *high-water mark* (mud on walls, rubbish hanging from branches) and is sometimes recorded later by a *high-water mark*, which can be listed in a national database of high-water marks.

This maximum water level moves downstream with *a flood wave*.

The *recession* is the lowering of the water level until the flow returns to the minor river bed. If this occurs very slowly (several hours at the maximum water level or very close to it), it may be referred to as *a flood plateau* rather than a flood peak.

In hydrology, the whole of the flood-flood phenomenon can be characterised at a characteristic point, (a section of control ex: a bridge) by a hydrograph whose slopes characterise the speed or slowness of the variations. The flood peak is then often linked to a flood return period.

The hydrograph

The hydrograph can be represented by a curve or a table, which shows the fluctuation in the flow of a watercourse in relation to time at a specific point. Flows have been measured at the given point

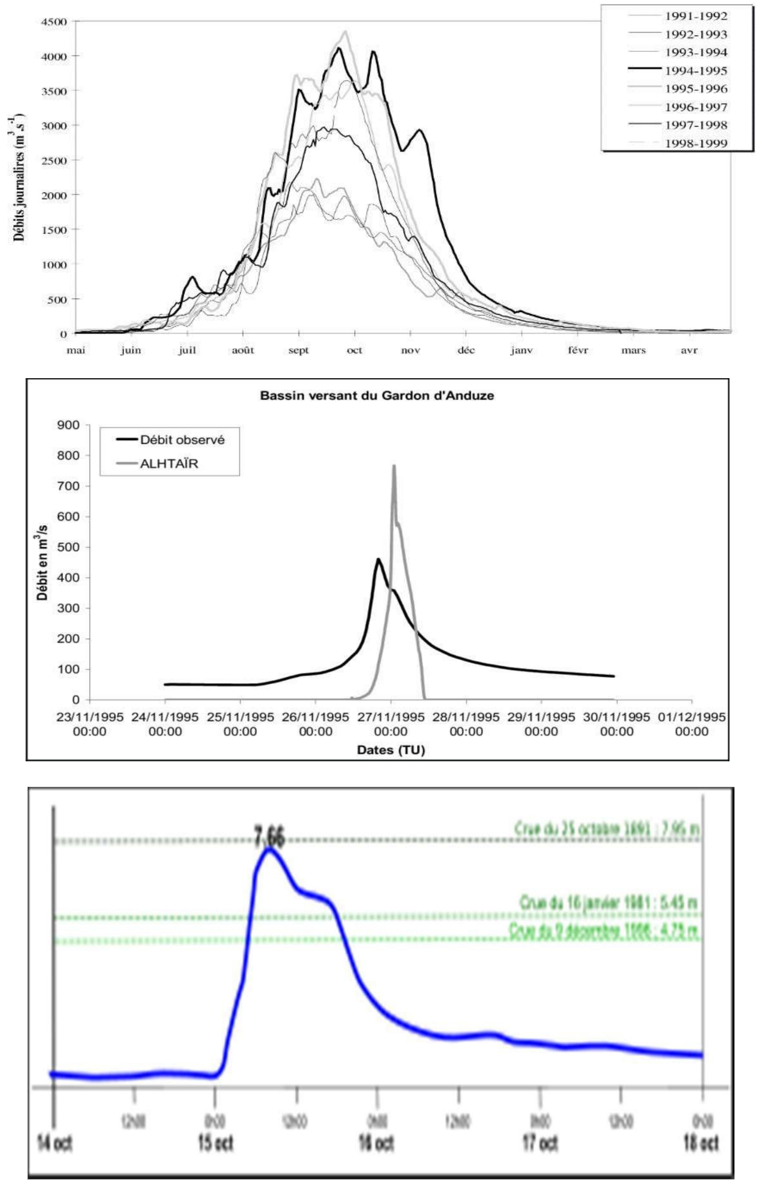


Figure 1: Example of the shape of a hydrograph

The hydrograph is an expression of the complex interaction between physiographic and climatic characteristics that dictate the dynamic association between rainfall and discharge in a specific catchment.

There are two distinct classifications of hydrographs that need to be remembered: the annual hydrograph and the hydrograph generated by a rainfall event.

Rainfall-generated hydrograph

A hydrograph is a graphical representation of the instantaneous flow at a point in the watercourse as a function of time (figure).

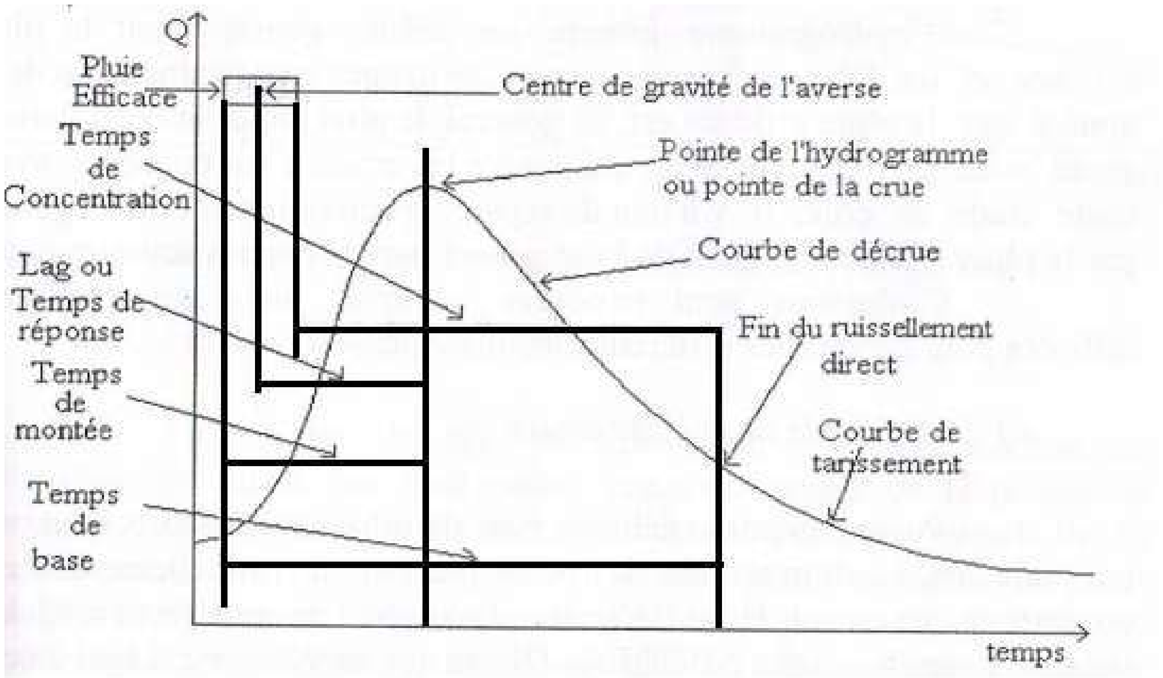


Figure 2: Presentation of a hydrograph

The hydrograph above shows :

- The rise time, between the start of direct runoff and the peak of the flood;
- The response time or "Lag" between the centre of gravity of the "effective" rainfall (i.e. the portion of the rainfall that is completely transformed into runoff) and the peak of the hydrograph;
- The base time or runoff duration, between the start of the effective rainfall and the end of the direct runoff;
- The time of concentration, between the end of the effective rainfall and the end of the direct runoff;
- The net rainfall or effective rainfall is the part of the shower that has runoff, the balance equation gives :

$$P= I + E + F+ S+ P_{net}$$

Where :

I: interception by plant cover

E: 2evaporation

F: infiltration

S: storage in depressions

P_{net}: net rainfall=effective rainfall=direct runoff

Separation of hydrograph components

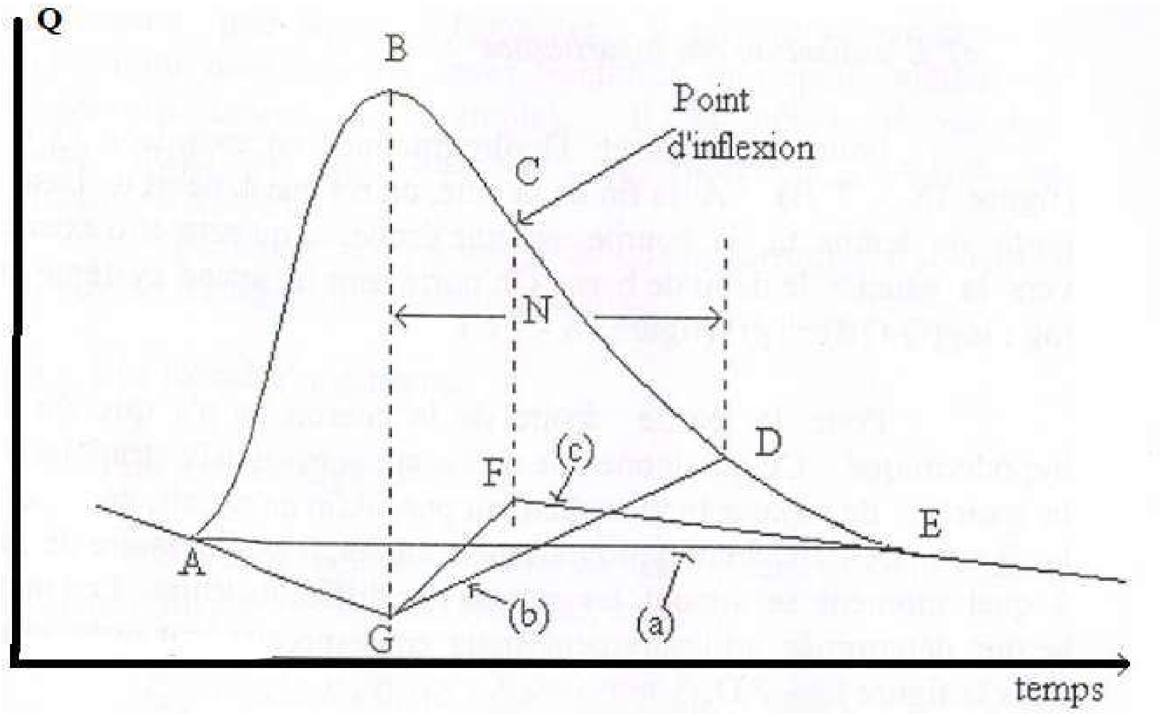


Figure 3: Separation of hydrograph components

1) The straight line method

To distinguish the base flow from the direct runoff, a horizontal line connects point A, where the direct runoff originates, to point E, where it ends. The volume under the curve ABCDEA, which is equivalent to the actual volume of rainfall, represents the amount of direct runoff (see figure above).

2) The fixed or constant base method

Direct runoff should cease after a designated period, noted N, following the peak of the hydrograph. The pre-existing base flow should continue until it crosses the vertical line passing through the top of the hydrograph (point G). At this point, a line segment GD is constructed, with D positioned at a distance equivalent to N from point G. The volume included under curve ABCDGA (as shown in the figure above) represents the volume of direct runoff.

3) The variable slope method

The base flow curve before the start of direct runoff is extrapolated to the time of the peak of the hydrograph (point G), and the base flow curve after the end of direct runoff is extrapolated back to the time of inflection point C (line EF). A line segment joins these two points G and F. The direct runoff is equal to the volume under curve ABCDEFGA (see figure above).

4) The ϕ index method (recharge rate)

If, for a given rainfall event, we have the hyetogram, the ϕ index and the total hydrograph generated by this rainfall event, we can determine the effective rainfall using a method

Once the effective rainfall (direct runoff) has been found, an area equal to that of the direct runoff is graphically subtracted from the total area of the hydrograph. The result is equal to the base flow.

The ϕ index is defined as the average rainfall intensity above which the volume of rainfall is equal to the volume of surface runoff (flow).

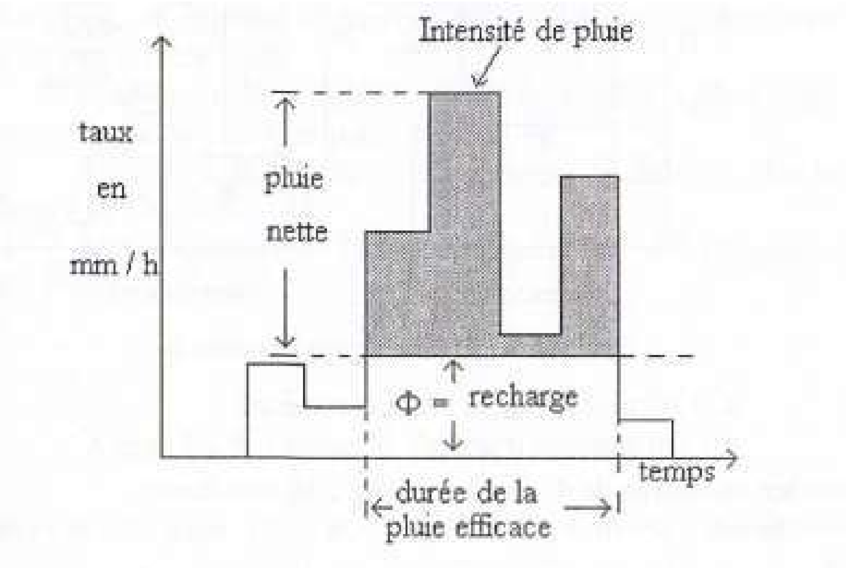


Figure 4: Diagram defining the ϕ index

In other words, it is the average intensity above which any excess rainfall is found in the form of run-off at the outlet.

In the figure above, the unshaded area below the line represents all losses, including water in depressions, evaporation and infiltration.

Time of concentration

Time taken for a particle of water coming from the part of the basin furthest "hydrologically" from the outlet to reach it. $t_{(c)}$ can be estimated by measuring the time between the end of net rainfall and the end of direct run-off (i.e. the end of surface run-off).

The choice of the appropriate method depends on topographical and rainfall factors, as well as scale factors linked to the size of the catchment.

a) :Empirical formulas 1-
Giandotti formula

$$T_c = \frac{6L\sqrt{A}}{0,8\sqrt{H_{moy} - H_{min}}}$$

Where :
Tc: Time of concentration (h)

A: Catchment area (Km2).
L: Length of main talweg (Km).
H_{avg}: Average height of catchment (m). H_{min}:
Minimum height of catchment (m).

2- Kirpich formula

$$T_c = 0,38\left(\frac{L}{\sqrt{I}}\right)^{0.77}$$

Where :

I: Average slope of the main trench

L: Length of main talweg (Km).

3- The Algerian formula

This was determined by Melles Saadi Cherif and Tamani, in their final year project at the USTHB-IGC, in 1992

$$T_c = 0.0055\ A + 0.1657\ L + 0.0078\ D_H + 0.821$$

Where :

Tc: Time of concentration (h)

A: Catchment area (Km2). L: Length of

main talweg (Km). $D_H = H_{moy} - H_{min}$

H_{avg}: Average height of catchment (m). H_{min}:
Minimum height of the catchment in (m).
This formula was determined on the basis of an analysis of "downpour" events recorded in 15 catchment areas in northern Algeria.

b) Transformation of rainfall into a flood

hydrograph

The transformation of rainfall into a flood hydrograph involves the successive application of two functions, known respectively as the production function and the transfer function:

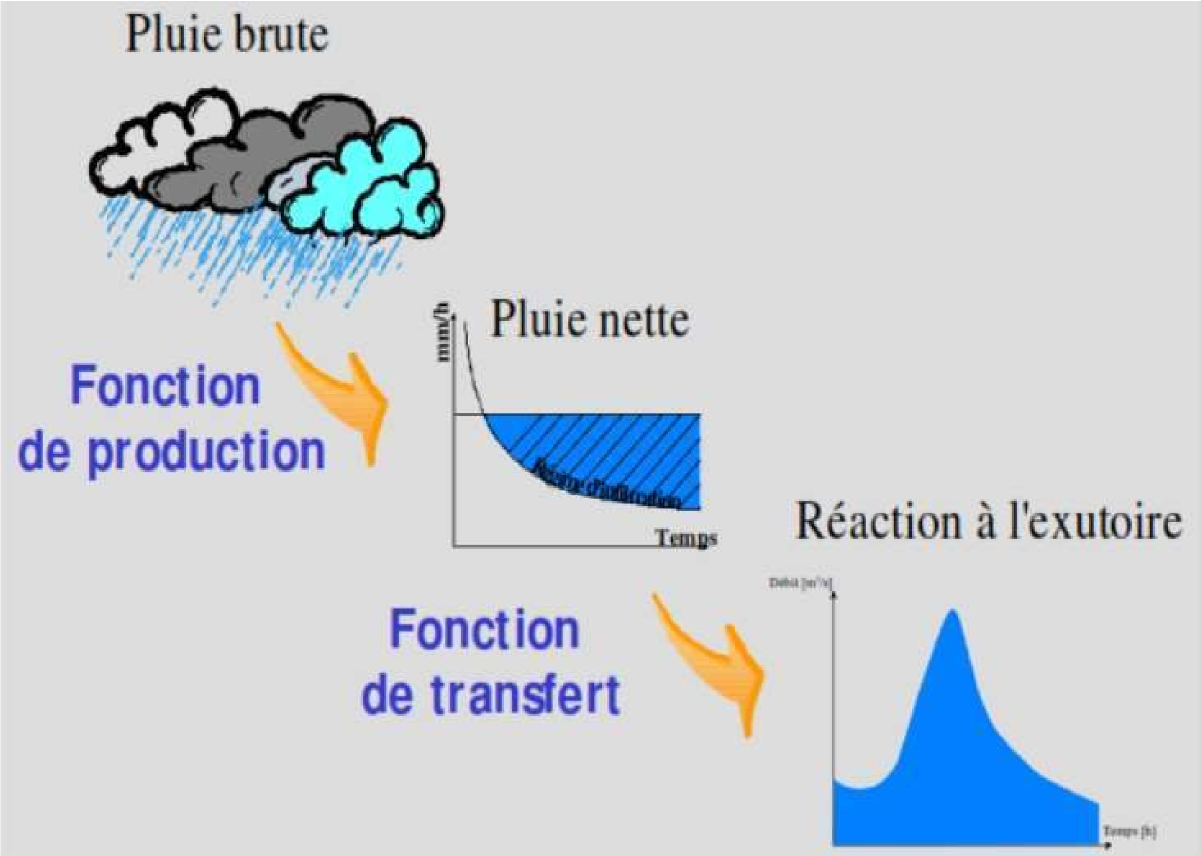


Figure 5: Transformation of rainfall into a flood hydrograph

Production function: this is used to determine the net rainfall from the gross rainfall. The net rainfall is the fraction of raw rainfall that is totally involved in the runoff.

Transfer function: used to determine the flood hydrograph resulting from a rainfall event (often considered as the net rainfall).

Analysis of rainfall-runoff events requires knowledge of a certain number of characteristic elements of the flood (characteristic shape and duration). A downpour, defined in time and space, falling on a catchment area of known characteristics, and under given initial conditions, causes a defined hydrograph at the outlet of the catchment area in question.

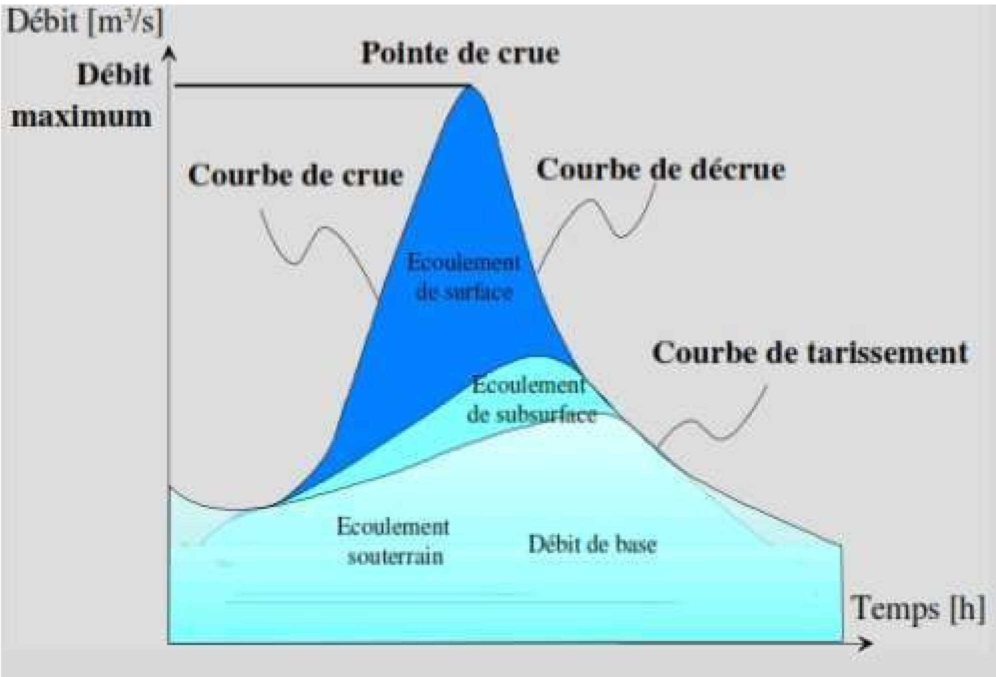


Figure 6: Shape of the flood hydrograph The

shape of the flood hydrograph is characterised by :

- The flood rise or concentration curve
- The flood peak or crest of the hydrograph
- The deflooding curve
- The drying-up curve

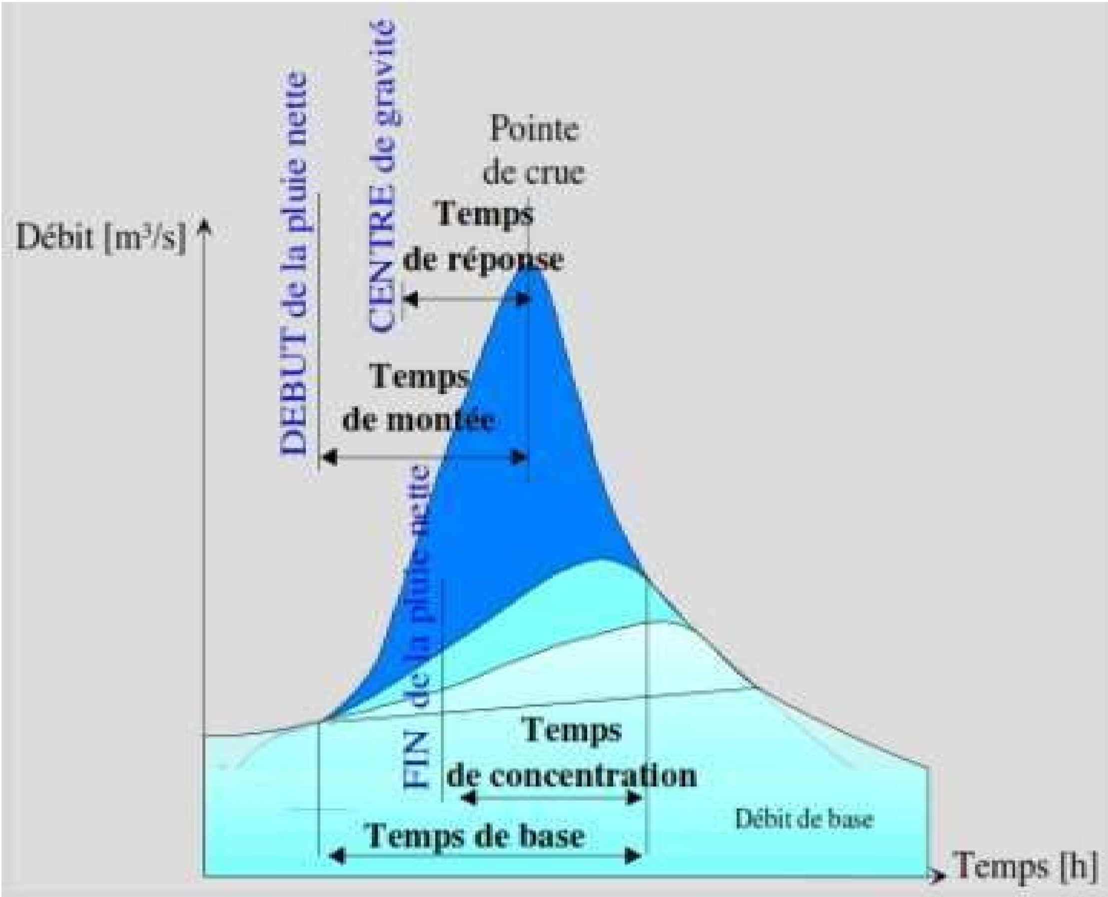


Figure 7: Characteristics of the hydrograph The

characteristic times of the flood hydrograph are :

- Basin response time
- Time of concentration
- Rise time
- Base time
- Basin response time *t p* (or "lag") - The time interval between the centre of gravity of the net rainfall and the flood peak or sometimes the centre of gravity of the hydrograph due to surface runoff.
- Time of concentration tc - Time taken for a particle of water coming from the part of the basin "hydrologically" furthest from the outlet to reach the latter. tc can be estimated *by* measuring the time between the end of net rainfall and the end of direct runoff (i.e. the end of surface runoff).
- Rise time tm- Time elapsing between the arrival at the outlet of the fast flow (detectable by the limnigraph) and the maximum of the hydrograph due to surface runoff.
- Base time tb- Duration of direct runoff, i.e. the length on the time abscissa of the base of the hydrograph due to surface runoff.

The area between the delayed runoff curve and the flood/decrease hydrograph represents the volume runoff. This volume, expressed in blade of water, is equal by definition to the volume of net rainfall. However, as the distinction between delayed subsurface runoff and direct surface runoff is relatively blurred, it is not uncommon to consider a volume of direct runoff equivalent to that of the net rainfall, defined as the area between the curve of the flood/decrush hydrograph and that of the subsurface runoff.

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