الجمهورية الجزائرية الديموقراطية الشعبية

DEMOCRATIC AND POPULAR ALGERIAN REPUBLIC
MINISTRY OF HIGHER EDUCATION AND SCIENTIFIC RESEARCH

MUSTAPHA STAMBOULI UNIVERSITY OF MASCARA
FACULTY OF NATURAL AND LIFE SCIENCES
AGRICULTURAL SCIENCES DEPARTMENT

Hydroclimatology

Presented by:

Dr. Benali Benzater

Edition 2025

Faculty of Natural and Life Sciences (SNV)
Sidi Said, 1645, Mascara BP305 – Algeria
benzaterbenali@univ-mascara.dz

Foreword

This course material, titled "Hydroclimatology" is intended for first-year Master's students in Hydrogeology, within the training area of Earth and Universe Sciences (STU) at the Faculty of Nature and Life Sciences (SNV) of Mustapha Stambouli University of Mascara. This document aims to provide students with an introduction and in-depth analysis of fundamental concepts and phenomena related to Climatology. It is organized into seven (7) chapters: the first chapter deals with the atmosphere and its gaseous composition. The second details solar radiation and radiative balance. The third describes atmospheric and oceanic circulations. In the fourth chapter, atmospheric and oceanic circulations are explained exhaustively. The fifth and sixth are devoted to the study of climate and evapotranspiration. Finally, and in the last chapter, the document reserves a place for climate change and its impacts.

Table Of Contents

Programme Master 1 Hydrogéologie (Canevas mise à jour 2023)

Intitulé du Master : Hydrogéologie
Semestre : 01
Intitulé de l'UE : UEF13
Intitulé de la matière 2 : Hydroclimatologie

45heures (1h30 cours, 1h30 TD)
Coefficients : 02
Crédit : 04

Objectifs de la matière : Connaissance sur les termes de la climatologie et de l'hydrologie fondamentale
Connaissances préalables recommandées : Notions de climatologie, hydrologie et hydrogéologie.

Contenu de la matière :

Eléments de vocabulaire : temps, climat, météorologie, climatologie, hydrologie, échelles météorologiques,
Système climatique global :

- Analyse des différentes composantes du système climatique global: l'atmosphère (circulation) l'hydrosphère (océans), la géosphère et leurs interactions, l'albédo.
- Caractéristiques et rôle des composantes du système climatique global :approche, évaluation, mesure, variabilité, tendance à court, moyen et long termes;
- Fonctionnement du système climatique, répartition régionale des climats (classification de Wladimir Koppen, représentations graphiques, indices descripteurs), impacts des activités humaines sur les composantes climatiques et leurs conséquences: la fonte de la banquise et des glaciers, l'effet de serre, la couche et le trou d'ozone, les tornades, les pluies acides, les ouragans, la désertification, les déluges.

Hydroclimatologie et ressources en eau :

- Différentes composantes du cycle hydrologique à l'échelle planétaire.
- Le bassin versant.
- Evaluation des paramètres hydroclimatiques appliqués au bassin versant : température, vent, humidité, tensions de vapeur, rayonnement solaire, pression, précipitation, évaporation (évapotranspiration), infiltration.
- Métrologie utilisée en hydroclimatologie.
- Analyse des impacts des changements climatiques et des activités humaines sur les phénomènes hydrologiques extrêmes.

Sorties sur terrain
Travaux Dirigés

- Des séances de TD en relation avec le cours

Mode d'évaluation : évaluation continue et examen

Références :

- M. Tabeaud, 2008. La climatologie générale, Armand Collin
- Gérard Guyot, 2013. Climatologie del'environnement, Dunod.
- MUSY A. et C. HIGY, 2004. Hydrologie: une science de la nature. Presses Polytechniques et Universitaires Romandes, Lausanne.
- Hingray B., C. Picouet, A. Musy, 2009. Hydrologie 2: Une science pour l'ingénieur, Presses polytechniques et universitaires romandes, Lausanne.
- Anctil F., J. Rousselle, Lauzon N., 2012. Hydrologie – Cheminements de l'eau, 2e éd., Presses internationales polytechniques, Montréal

Master 1 Hydrogeology program (updated framework: 2023)

Title of the Master: Hydrogeology
Semester: 01
EU title: UEF13
Subject title 2: Hydroclimatology

45 hours (1h30 class, 1h30 tutorial)
Coefficients: 02
Credit: 04

Subject objectives: Knowledge of the terms of climatology and fundamental hydrology. Recommended prior knowledge: Concepts of climatology, hydrology and hydrogeology.

Content of the subject

- Vocabulary elements: time, climate, meteorology, climatology, hydrology, meteorological scales,

Global climate system:

- Analysis of the different components of the global climate system: the atmosphere (circulation) the hydrosphere (oceans), the geosphere and their interactions, albedo.
- Characteristics and role of the components of the global climate system: approach, evaluation, measurement, variability, short, medium and long term trend;
- Operation of the climate system, regional distribution of climates (Wladimir classification Koppen, graphic representations, descriptor indices), impacts of human activities on climatic components and their consequences: the melting of the ice floes and glaciers, the effect of greenhouse, the ozone layer and hole, tornadoes, acid rain, hurricanes, desertification, floods.

Hydroclimatology and water resources:

- Different components of the hydrological cycle on a global scale.
- The watershed.
- Evaluation of hydroclimatic parameters applied to the watershed: temperature, wind, humidity, vapor pressure, solar radiation, pressure, precipitation, evaporation (evapotranspiration), infiltration.
- Metrology used in hydroclimatology.
- Analysis of the impacts of climate change and human activities on the phenomena extreme hydrological conditions.

Field trips
Directed work

- Tutorial sessions related to the course

Method of assessment: continuous assessment and review

References

- M. Tabeaud, 2008. La climatologie générale, Armand Collin
- Gérard Guyot, 2013. Climatologie del'environnement, Dunod.
- MUSY A. et C. HIGY, 2004. Hydrologie: une science de la nature. Presses Polytechniques et Universitaires Romandes, Lausanne.
- Hingray B., C. Picouet, A. Musy, 2009. Hydrologie 2: Une science pour l'ingénieur, Presses polytechniques et universitaires romandes, Lausanne.
- Anctil F., J. Rousselle, Lauzon N., 2012. Hydrologie – Cheminements de l'eau, 2e éd., Presses internationales polytechniques, Montréal.

Global climate system

Climate derives from the Greek κλίμα which means « incline ». This refers to the first explanatory factor in the geography of climates: solar radiation and in particular its inclination (incidence) on the earth's surface. Climate refers to the average weather conditions over a long time period. The usual definitions contain these two notions and are based on the statistical distribution of instantaneous atmospheric conditions for a given period. Theoretically, the climate of a region is based on the average of at least thirty years of atmospheric conditions defined by temperatures, precipitation, atmospheric pressure, wind, etc.

Climatology, a branch of physical geography, is the study of climate, that is to say the succession of meteorological conditions over long periods of observations. Climate defines and explains the conditions of the atmosphere above a place in the medium and long term, while meteorology focuses on the short term and in particular forecasts over a few days. Climatology studies the components and variations of climates on the earth's surface.

Usually, climate designates the average, calculated over a long period of time (30 years, by convention, for meteorologists), of observations of parameters such as temperature, pressure, rainfall or wind speed, in a geographic location and on a given date (Intergovernmental Panel on Climate Change, IPCC, 2013). According to the World Meteorological Organization (WMO), climate is defined as the synthesis of meteorological conditions in a given region, characterized by long-term statistics of the variables of the state of the atmosphere (Elmeddahi, 2016).

Climate is an environmental component resulting from Earth-Ocean-Atmosphere interactions. Its variations impact human activities. The objective of this course is to lay the foundations for understanding the Earth's climate system. The main themes covered are: solar energy supply, greenhouse effect, general atmospheric circulation, and atmospheric water cycle.

Planet Earth is the structure of several spheres (Figure 1). These include the christosphere, the pedosphere, the biosphere, and especially the atmosphere and the hydrosphere. These last two mentioned respectively are the different domains of definition of Climatology and Hydrology.

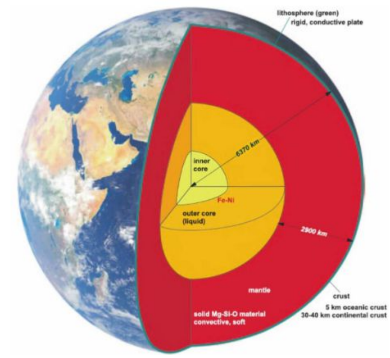


Figure 1. Structure of the terrestrial globe

Equatorial radius: 6,378.14 km, Polar radius: 6,356.78 km, Equatorial perimeter: 40,075.03 km,, Area: 510,067,420 km², Volume: 1.08321×1012 km³, Mass: 5974 billion billion of tonnes, Average density: 5.52 (water: 1), Average surface temperature: 14°C, Rotation period: 23 hours 56 minutes 4 seconds, Average orbit radius: 149.598 million km, Revolution period: 365 days 6 hours and 9 minutes, Rotation speed (at the equator): 1,674.38 km/h

Depending on the latitudes, the earth is made up of five major climatic zones (Figure 2):

- The surface convergence zone is also called the Inter Tropical Convergence Zone (ITCZ). It is located at the equator (0°).
- The northern mid-latitude zone. It is located between the equator (°) and latitude 30° north (Tropic of Cancer) of the northern hemisphere.
- The southern mid-latitude zone. It is located between the equator (°) and latitude 30° south (Tropic of Capricorn) of the southern hemisphere.
- The north polar zone corresponds to the Arctic Circle at the North Pole (90° N).
- The south polar zone corresponds to the Antarctic Polar Circle at the South Pole (90°S).

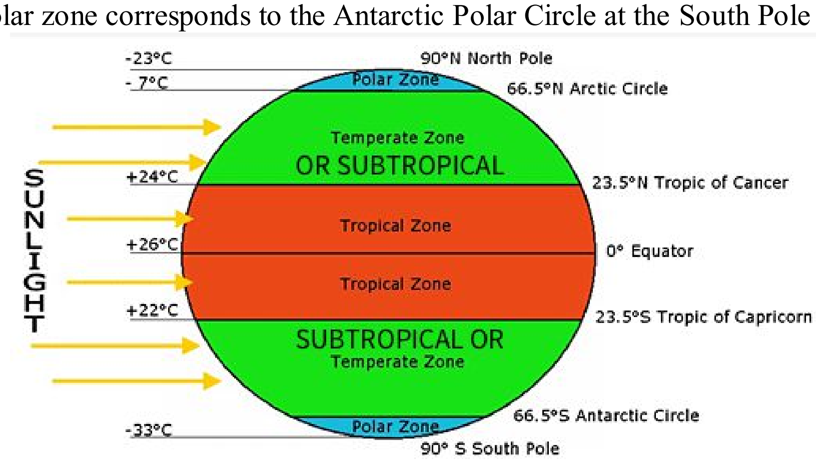


Figure 2. The major climatic zones

The four (04) seasons that we know result from the inclination of the earth's axis of rotation (Figure 3).

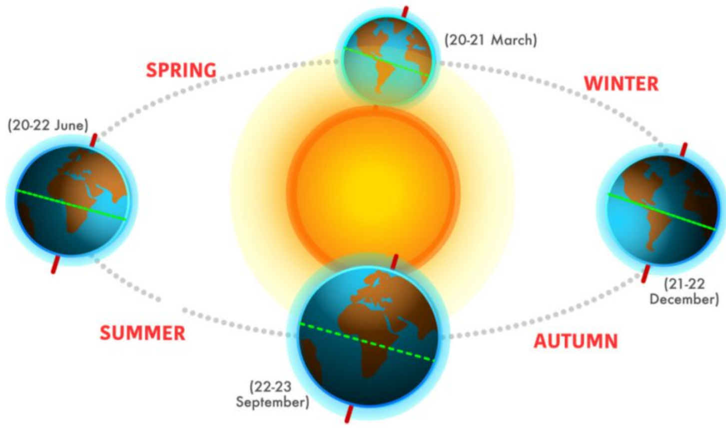


Figure 3. The astronomical seasons

Another impact of the inclination of the earth's axis of rotation is the unequal distribution of incident solar energy on different regions of the globe (Figure 4). The equatorial regions of the earth (ITCZ) receive more solar energy, which constitutes the only source of life on earth, than at the two (02) poles (Benzater, 2021).

Indeed, for the same quantity of energy (1 m^2 of flux) emitted by the sun to different regions of the globe, the receiving surface of this flux is minimum at the equator (1 m^2), maximum at the two (02) poles (56 m^2). This inequality between the equator and the 02 pole creates a radiative balance.

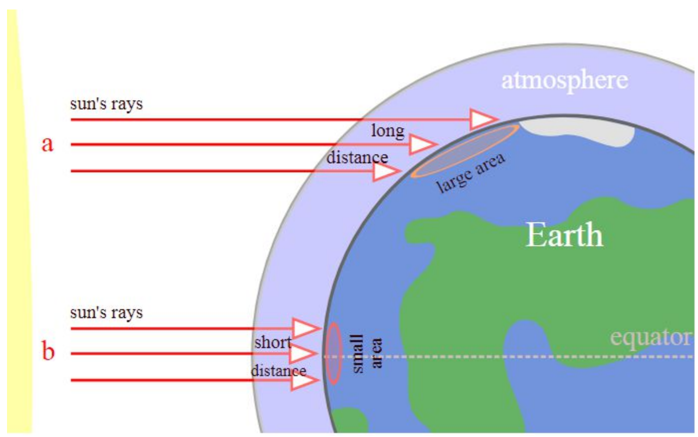


Figure 4. Solar impacts on different regions of the Earth

CHAPTER I

ATMOSPHERE AND ATMOSPHERIC PRESSURE

I- Chapter 1: Atmosphere and atmospheric pressure

1.1. Introduction

The atmosphere is the layer of air that surrounds the earth. It therefore rests on the entire surface of the terrestrial globe. From the point of view of physics, the atmosphere obeys the same laws as water because it is a fluid; the only difference is that its density is lower than that of water.

1.2. Composition and vertical structure

The earth's atmosphere has a film structure if we compare its thickness to the diameter of the earth. It is difficult to give it a mathematical depth because the system is open to the “sky”. The atmosphere (Table 1) is essentially made up of a gas mixture. It includes 78% nitrogen (N_2, important for the life and nutrition of plants), 21% oxygen (O_2, essential for respiration), nearly 1% argon (Ar), of the order of 0.03% carbon dioxide or carbon dioxide gas (CO_2), and proportionally trace amounts of helium, hydrogen, krypton, methane, carbon monoxide, neon, ozone and xenon. The mixture remains approximately constant, except around 30 to 40 km altitude where ozone (O_3) is concentrated in what is called “the ozone layer”.

Table 1. Composition of atmospheric air

Gases constituting dry air	Volume (%)	Molar mass
Nitrogen (N_2)	78,09	28,016
Oxygen (O_2)	20,95	32,000
Argon (A)	0,93	39,944
Carbon Dioxide (CO_2)	0,035	44,010
Neon (Ne)	$1{,}8\ 10^{-3}$	20,183
Helium (He)	$5{,}24\ 10^{-4}$	4,003
Krypton (Kr)	$1{,}0\ 10^{-4}$	83,07
Hydrogen (H_2)	$5{,}0\ 10^{-5}$	2,016
Xenon (Xe)	$8{,}0\ 10^{-6}$	131,300
Ozone (O_3)	$1{,}0\ 10^{-6}$	48,000
Radon (Rn)	$6{,}0\ 10^{-18}$	222,000

The succession of layers of the atmosphere from bottom to top is as follows (Figure 5).

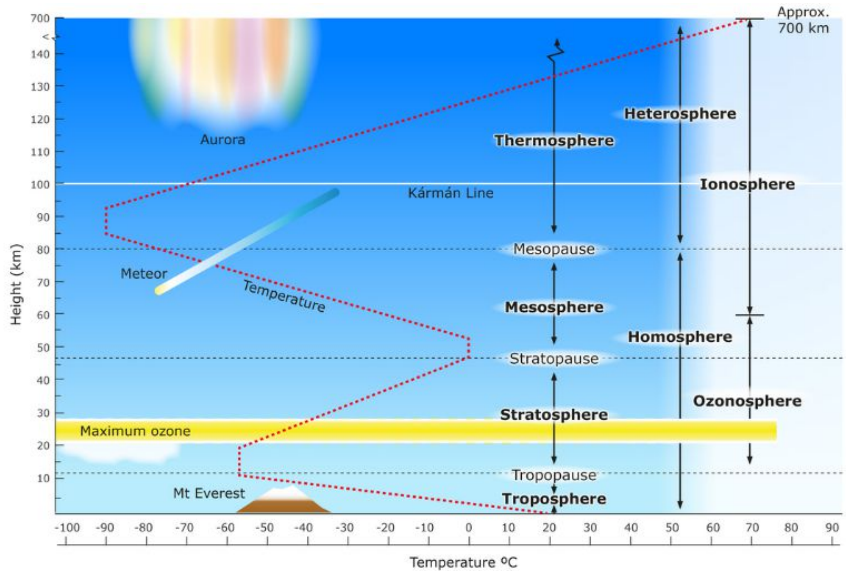


Figure 5. Atmosphere structure

1.2.1. Troposphere

Or lower atmosphere extends from 0 to 12-16 km: it is the most agitated part of the atmosphere with the presence of horizontal and vertical movements. In fact, most weather phenomena develop there. It contains ¾ of the atmospheric mass, all the water and solid bodies. Its temperature drops regularly by 0.65°C per 100 m (around 6°C every 1000 m) from 3 km. The troposphere is not homogeneous: the base or geographic layer (0 to 3 km) and the free layer (3 to 12-16 km). Note that the troposphere is thinner at the poles (6 to 9 km) than at mid-latitudes (around 11 km) or at the equator (12 to 16 km). Its upper limit is called tropopause.

1.2.2. Stratosphere and Mesosphere

The Stratosphere (12-16 to 50 km) and the Mesosphere (50 to 80 km): the stratosphere is the seat of fast winds and temperatures increase up to the top while in the Mesosphere, temperatures decrease up to the top (-90°; cooling). The stratopause marks the base of the Mesosphere. These two layers are characterized by an absence of water vapor, a relative humidity of 25%, by the rarefaction of gases and the presence of the majority of ozone (from 15-30 to 40 km). Ozone (O_3), the triatomic form of oxygen, has a maximum concentration from 15 km (above the polar regulations) and 30 km (at the equator): we speak of the ozone layer. Ozone is produced by the recomposition of free oxygen atoms dissociated by ultraviolet rays (0.1 to 0.4 µm) from the sun. It constitutes a filter against this harmful radiation, but its great instability can release oxygen. In these two parts of the atmosphere, we note a weakness of vertical movements and a predominance of horizontal movements of the air. They are the seat of east winds in summer and west winds in winter.

1.2.3. Thermosphere or Ionosphere

The upper atmosphere extends from 80 to 800 km above the Earth's surface. At 800 km, the air density is approximately 10^6 atoms/cm^3, compared to about $20x10^6$ $x10^9$ molecules/cm^3 at ground level. At such altitudes, these substances are not considered gases. Temperatures in this region are high, reaching approximately 1000°C. This layer is also known as the Ionosphere, as it is characterized by ionization of air particles and atomic dissociation (changes in the number of electrons in the atom producing free electrons). Ionization is due to electrified ultraviolet rays and X-rays coming from the sun. At around 80 km, the mesopause limits the ionosphere (Cosgrove, 2005).
The proportion of these different elements varies with altitude and this determines the vertical variation of climatic parameters such as temperature (Figure 6), humidity, pressure (Figure 7) (Koli et al., 2011).

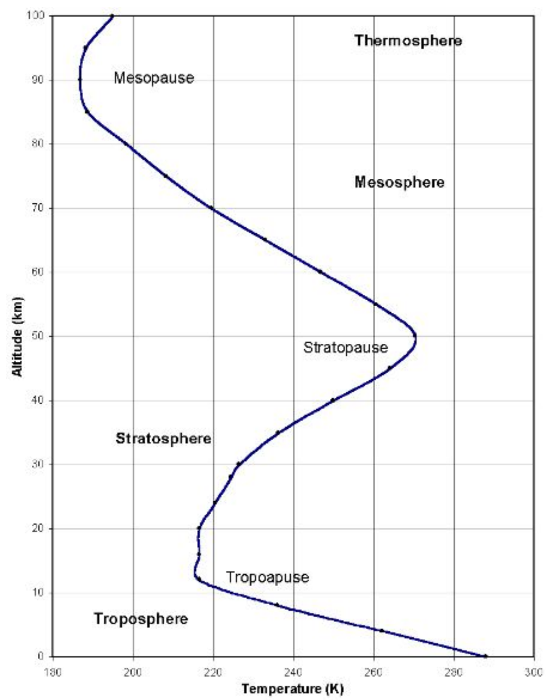


Figure 6. The atmospheric temperature gradient with altitude

Note that in the atmosphere, water is the only body encountered in the three states of matter: (a) gaseous (water vapor), (b) liquid (cloud droplets) and (c) solid (cloud crystals).). Other solid bodies, such as dust, ashes and sea salt crystals, are present in varying quantities depending on their source. They play an important role in condensation and absorption of solar radiation. However, taking into account the variation of temperatures, the proportion of certain gases and atmospheric pressure, four regions or superimposed layers are observed (Koli et al., 2011).

1.3. Atmospheric pressure or « air weight »

The mass of the atmosphere weighs approximately 1 kg/cm^2 (precisely 1.033 Kg/cm^2). The air exerts a pressure on the surface of the earth which is called atmospheric pressure which corresponds to the «weight of the air». If the atmosphere were distributed uniformly and equivalently all around the earth, it would exert a pressure of 1013.25 hPa; 1015 hPa representing the normal average pressure at sea level. In the lower layers, the pressure is high because the air density is high (Figure 7). The pressure drops with the attitude where the rarefaction of the air causes the Earth's attraction (Table 2).

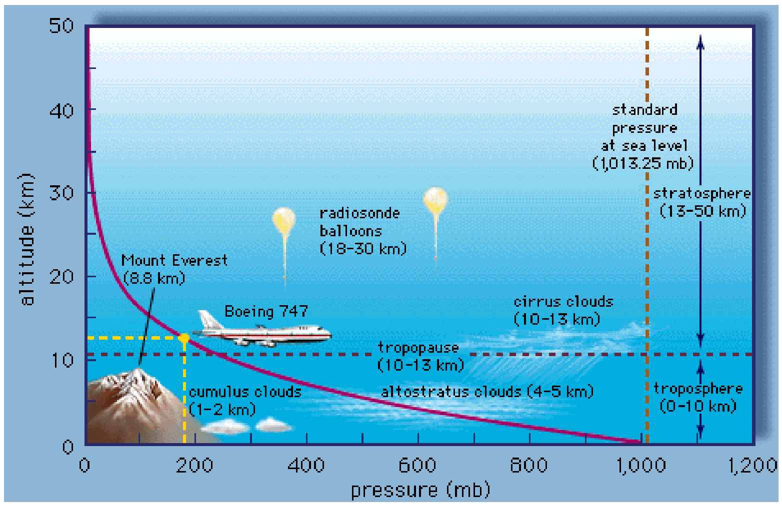


Figure 7. The atmospheric pressure gradient

Table 2. Variation of atmospheric pressure as a function of altitude

Altitude (m)	Pressure (hPa)
0	1013
5 000	700
10 000	300
16 000	100

The unit of measurement for atmospheric pressure is the hectopascal (hPa); formerly millibar (mb) where 1hPa=10 millibars. The measuring instrument is the barometer and the line which joins the points of the same pressure is called isobaric; but a series of equivalent pressures constitutes a pressure field. On a horizontal plane, the recorded pressure varies from normal: hot air is light and its pressure is low and cold air is heavy and its pressure is high. Note also that:

- A high pressure zone or anticyclonic area (pressure greater than 1,015 hPa) is coded A to define the high pressure regions where the air is subsiding (cold/dry or hot/dry);

- A low pressure zone or low pressure area (pressure less than 1,015 hPa) is coded D to define the low pressure regions where the air is ascending;

- In general, air flows from high pressure to low pressure.

1.4. Climate and climate system

According to the World Meteorological Organization (WMO), climate is defined as the synthesis of meteorological conditions in a given region, characterized by long-term statistics of the variables of the state of the atmosphere (Elmeddahi, 2016).

It is a complex dynamic system whose components interact with each other constantly. Its different components are mainly made up of the atmosphere, the lithosphere, the hydrosphere, the cryosphere (ice around the world) and the biosphere. These five (05) spheres are in permanent interaction via solar energy (Figure 8).

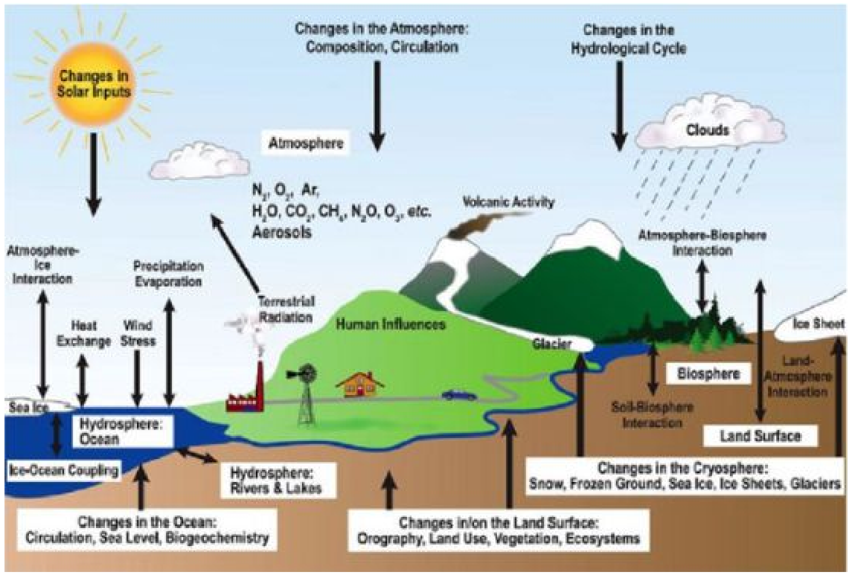


Figure 8. Components of the climate system and their interactions (IPCC, 2007)

The climate system includes the atmosphere, oceans and ice, and emerged land. Its operation is dominated by energy exchanges between the interior of the Earth system and the external solar source, and by exchanges between the three main compartments that make it up.

CHAPTER II

SOLAR RADIATION AND RADIATIVE BALANCE OF THE GLOBE

II- Chapter 2: Solar Radiation and Radiation Balance of the Globe

2.1. Solar Radiation

The fluid envelopes of the planet, that is to say the atmosphere and the hydrosphere, draw their energy from the sun. Solar emission, called solar constant, is 2 cal/cm^2/min. Solar radiation has a maximum in the short waves, the visible in particular. This flow of external origin is filtered by the atmosphere which absorbs, reflects and diffuses part of it. The loss is of the order of 50%, so on average only 1 cal/cm^2/min reaches the surface of the planet. Compared to the solar flux, the internal flux (geothermal energy) coming from the lithosphere is negligible due to the low conductivity of the rocks. A high water content in the atmosphere increases incident energy loss through albedo (reflection from ice clouds clearly visible in airplanes) and through absorption by water vapor, droplets and dust. The duration of illumination (sun is above the horizon) per 24 hours is constant at the equator (12 hours/24 hours); but everywhere else it decreases in winter and increases in summer. We recorded 24 hours of continuous illumination at 66° 33' (six months at the pole) (Ross et al. 2009).

2.2. Heat transfer modes

To raise the temperature of an object, it provides it with a certain amount of heat. It’s bringing him energy. Heat is expressed in joules and the quantities of heat exchanged per unit of time are expressed in «Watts», we're talking about power.
Example: The specific thermal heat of water being on average 4.19 kJ/kg.K, 419 kJ must be provided to heat a liter of water from 0°C to 100°C.
Heat in the atmosphere propagates in three (03) modes:

2.2.1. Conduction

This is a direct transfer. It is specific to solids. Conduction takes place over small thicknesses, in contact with the hot parts of the walls and the floor (Figure 9). Air is a poor conductor of heat. It can even be considered a good insulator.

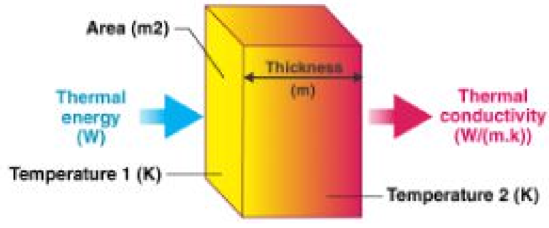


Figure 9. Conduction

2.2.2. Convection

Convection (Figure 10) is specific to fluids (liquid, gas, air). Heat is carried by the movement of a carrier fluid (liquid or gas). Convection can be natural (radiators, cumulus clouds, etc.) or forced.

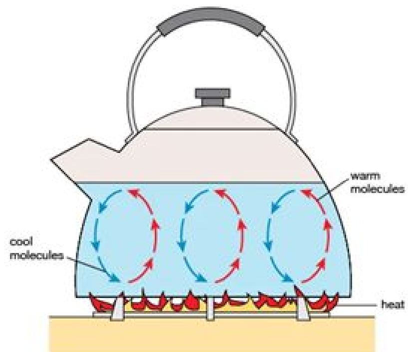


Figure 10. Convection

2.2.3. Thermal radiation

It is the transmission of heat without material support, but in the form of electromagnetic waves, such as light or radio waves. When an object is exposed to radiation, it heats up. Any object whose temperature is non-zero emits thermal radiation.

Depending on the temperature of the object emitting the radiation, it may appear to us: as very luminous... or, on the contrary, completely invisible. In both cases, we speak of « black body» radiation.

The energy emitted per unit of time (power), perpendicular to a unit surface element of the emitting body, is a function of the body temperature. It is in fact a flow of energy, and we speak of a «flow» of radiation (or « intensity»). It is expressed in Watts per m^2 ($W.m^{-2}$).

The energy distribution in the Earth-Ocean-Atmosphere (TOA) system schematically divides the planet into three groups: an energy surplus zone between 30°N and 30°S and two deficit zones beyond (Dhonneur, 1978). Energy transfers tend to restore the balance between these different zones (essentially meridian transfer) on the one hand, and on the other hand between the excess land surface with which the surface marine layers are associated and the deficit atmosphere (vertical transfers).

Transfers between the excess intertropical zone and the deficit zones of the middle and polar latitudes take place with a lag of one to two months compared to the apparent movements of the sun. Although most of these transfers are carried by the atmosphere, part is carried out by marine currents (Figure 11).

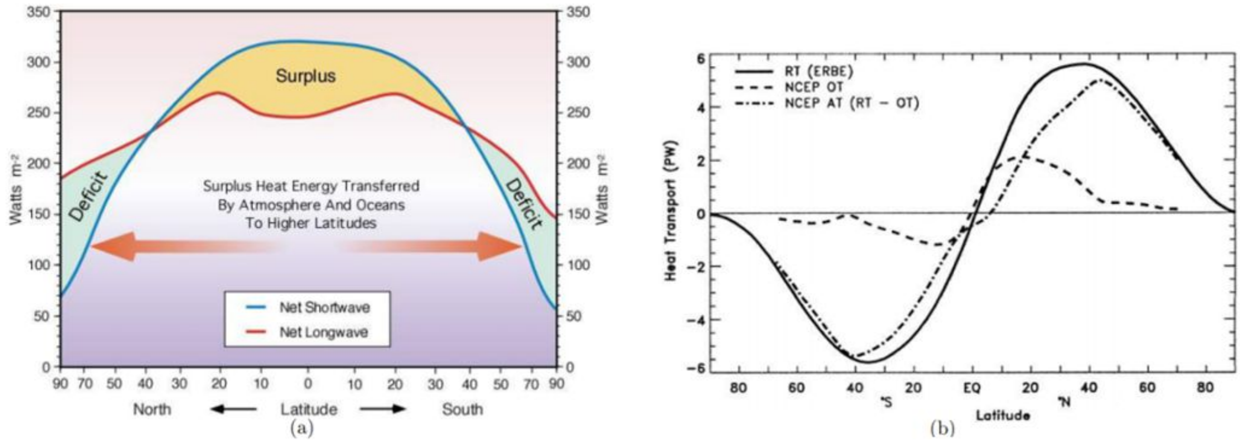


Figure 11. (a) Diagram representing the meridian section of the energy imbalance between the equator and the poles. Credits: NASA. (b) Meridian heat transport. RT: Total transport. OT: Oceanic component and AT: Atmospheric component. (Credits: Trenberth and Caron (2001)).

A warmer atmosphere changes precipitation patterns and increased climate variability. There will particularly be an increase in frequency and intensity of extreme phenomena, which will lead to a clearer succession of years of drought and floods.

One of the most important expressions of changes in the distribution of sea surface temperatures is known as the El Niño Southern Oscillation (ENSO) or El Niño. We now know that the El Niño phenomenon is triggered by a modification (decrease or reversal) of the trade winds which blow from east to west over the equatorial regions (Janicot, et al., 1993; Fontaine, et al ., 1998). The causes of this modification are still poorly understood, but its effects are felt at regular intervals (3 to 7 years can pass between 2 El Niño events) in the southern part of the Pacific Ocean.

CHAPTER III

ATMOSPHERIC AND OCEANIC CIRCULATION

III- Chapter 3: Atmospheric and oceanic circulation

3.1. Atmospheric circulation

Air is made up of a mixture of gases, the two main constituents of which are oxygen and nitrogen. It also contains a relatively large amount of water. Cloud masses are the direct manifestation of the presence of water in the atmosphere (atmospheric cloudiness).

In its gaseous state, water does not modify the transparency of the air, but when condensed into fine droplets or solidified into ice crystals, the atmosphere then becomes opaque.

If the degree of humidity of the air (partial pressure of water/saturation pressure) is less than 1, there is no condensation, the air is transparent. On the other hand, for a humidity level equal to 1, as water can condense, cloudiness can appear (cloud, mist or fog) (Figure 12).

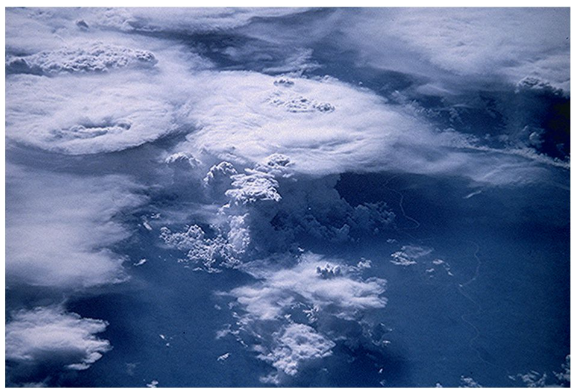


Figure 12. Water in the atmosphere

3.1.1. Dynamics of the Earth's atmosphere

The atmosphere is not a stationary sphere. It is dynamic. Atmospheric circulation follows a pattern. On the surface of the earth, there are pressure belts. These belts fall into two main categories: High Pressure belts and Low Pressure belts. The first are centers of convergence or departure and the second, the centers of divergence or arrival. We thus have several belts on the surface of the terrestrial globe: - the Polar and Tropical High Pressure belts - the subpolar and equatorial Low Pressure belts (Malardel, 2005).

3.1.1.1. The troposphere: a layer of unstable air

Located at the base of the atmosphere, the troposphere is driven by powerful movements which constantly stir the air: vertical movements, linked to temperature contrasts (convective processes) and horizontal movements, caused by differences in atmospheric pressure at ground level.

3.1.1.1.1. Vertical air movements

The density of air depends on its temperature: lighter warm air rises; on the contrary, the cold, heavier air settles towards the ground (Figure 13).

Thus, upon contact with the ground, in certain regions, the air heats up, therefore becomes lighter and rises: an upward movement occurs. As it rises, the air relaxes because the air pressure is lower and cools (decrease in temperature with altitude). The upward movement continues until the air has reached the temperature of the surrounding environment.

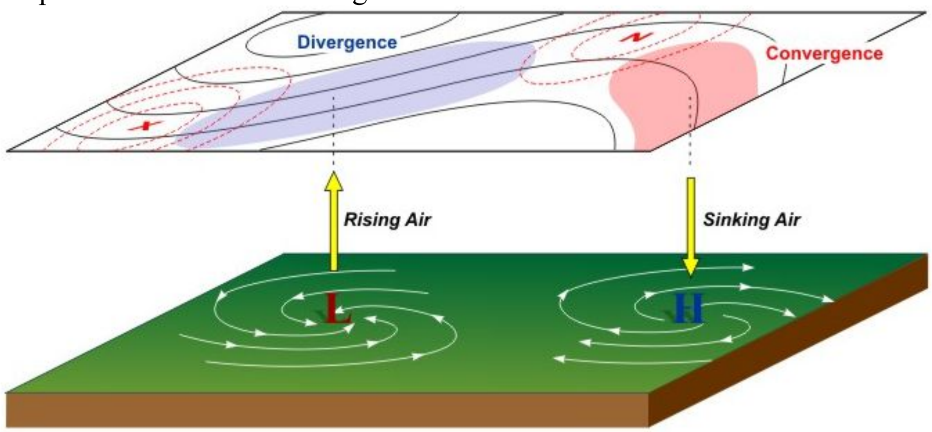


Figure 13. Ascent zone and subsidence zone

The extent of the movement will depend on the heating of the air at the start but also on its degree of humidity. Indeed, dry air sees its temperature decrease by 1°C every 100 m, whereas for air saturated with water, the temperature only decreases by 0.5°C every 100 m because the condensation of water at during the ascent releases heat.

Conversely, air colder than the ambient air, heavier, will descend towards the ground, compress and heat up: this is called subsidence (Figure 13).

3.1.1.1.2. Horizontal air movements

In the regions of ascent, the atmospheric pressure is lower than the average estimated at 1015 hectopascals, a depression forms. On the contrary, in subsidence zones, atmospheric pressure rises: an anticyclone forms.

The spatial distribution of high and low pressures varies during the year and constitutes the pressure field.

Meteorologists, for the purposes of forecasting, regularly draw up atmospheric pressure maps where the isobaric lines connect all the points which are at the same atmospheric pressure (Figure 14).

The wind is a horizontal movement of air caused by the force of pressure which tends to move the atmosphere from areas of high pressure to areas of low pressure to achieve a uniform pressure. This pressure force is perpendicular at each point to the isobaric lines, directed from high to low pressures and its intensity is greater as the pressure difference is high. It should therefore converge towards the center of a depression and diverge from the center of an anticyclone. However, we see that the wind on the ground circulates parallel to the isobaric lines. This is the consequence of the rotation of the Earth.

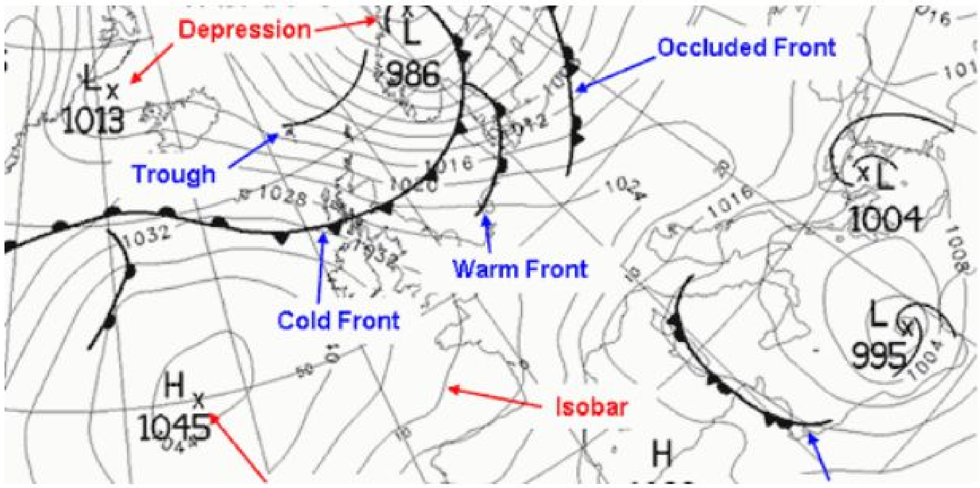


Figure 14. Weather maps (isobar lines)

In the northern hemisphere, the rotation of the Earth introduces an additional force, the Coriolis force, which causes any moving object to deviate to the right and in the southern hemisphere, to deviate to the left. This deviation is zero at the equator and maximum at the poles. This force, perfectly negligible in everyday life (trains remain on their tracks for example), is no longer so for major atmospheric and oceanic movements.

We can describe the movements of the atmosphere by assuming that at every point the pressure and Coriolis forces are balanced. We talk about the geostrophic hypothesis (Beau, 2012).

3.1.2. The Coriolis force

The Coriolis force is a law of kinematics, the statement of which is relatively simple: any particle moving in the northern hemisphere is deflected to its right (towards its left, in the southern hemisphere) (Figure 15).

The earth as a mobile emits kinetic energy through its movements of rotation and revolution. This creates winds or forces of various kinds on the surface of the globe. This is the case of the Coriolis force. Its action is preponderant in the study of winds (Beau, 2012).

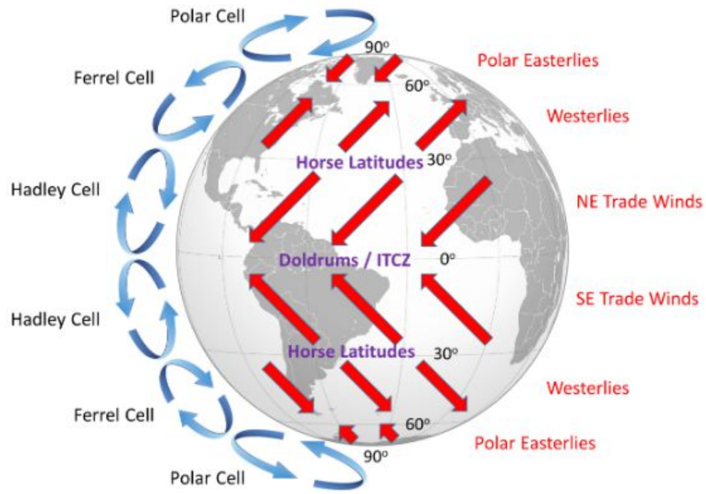


Figure 15. Action of the Coriolis force on wind direction

The Earth rotates around a north-south axis (Figure 16). Given the spherical shape of the terrestrial globe (second discovery), the linear speed of a point on its surface is not constant and depends on the latitude of this point: it increases starting from a pole, passes through a maximum at the equator, then decreases to the other pole:

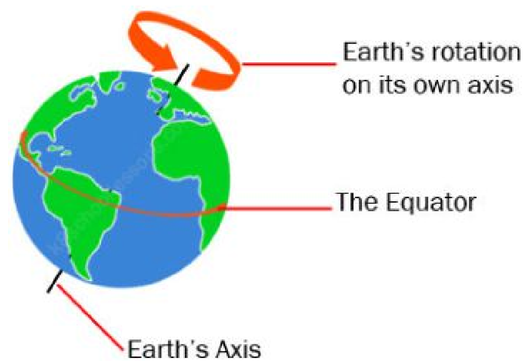


Figure 16. Rotation of the earth

So, if a mobile descends from the North Pole to the equator, it will be confronted with the increasing speed of the Earth's movement. As he continues his route towards the South Pole, he will see this speed decrease again.

Everything happens as if a walker leaves the center of a ride to reach the periphery, he will feel deviated in the opposite direction to the progress of the ride under his feet.

The direction of the Coriolis deviation does not depend on the direction of movement of the mobile on which it is exerted, but only on the direction of rotation of the support which gives rise to it. The direction of our diagram corresponds to the direction of rotation of the Earth in the northern hemisphere, with a rotation to the right (Malardel, 2005).

3.1.2.1. Action of the Coriolis force

In fact, this force is negligible in most cases, but becomes very important in certain phenomena, including the movement of air masses: the meteorological wind (due to the combination of factors influencing the force of Coriolis: low mass of particles, large scale of movement).

Furthermore, it is easily understood by referring to the example above, the faster the displacement, the greater the Coriolis deviation generated.

Finally, for there to be a Coriolis force there must be a change in speed of the support when moving on it. Therefore, the Coriolis force is maximum at the poles and negligible at the equator.

According to this hypothesis, the movement of the atmosphere does not occur perpendicular to the isobars but tangentially. Air does not flow from high to low pressures but revolves around low pressure centers and anticyclonic centers. In the Northern Hemisphere, the wind rotates counterclockwise around a low pressure center and clockwise around an anticyclone. In the southern hemisphere, wind movements are reversed.

The wind speed is even faster when the pressure differences are strong and when they occur over short distances (close isobaric lines). Latitude also plays an important role. This speed can be expressed by the Beaufort anemometer scale. The scale is graduated from 0 (zero wind) to 12 (hurricane).

3.1.3. General atmospheric circulation

The map (Figure 17) represents a modeling of the general atmospheric circulation at the surface. This shows a distribution of winds into six systems: the North-East and South-East trade winds, the west winds of the mid-latitudes and the east winds of the Polar Regions. These systems are separated by the intertropical convergence zone (convergence of equatorial low pressures or ITCZ and by the two extratropical convergence zones corresponding to low pressures near latitude 60° in each hemisphere.

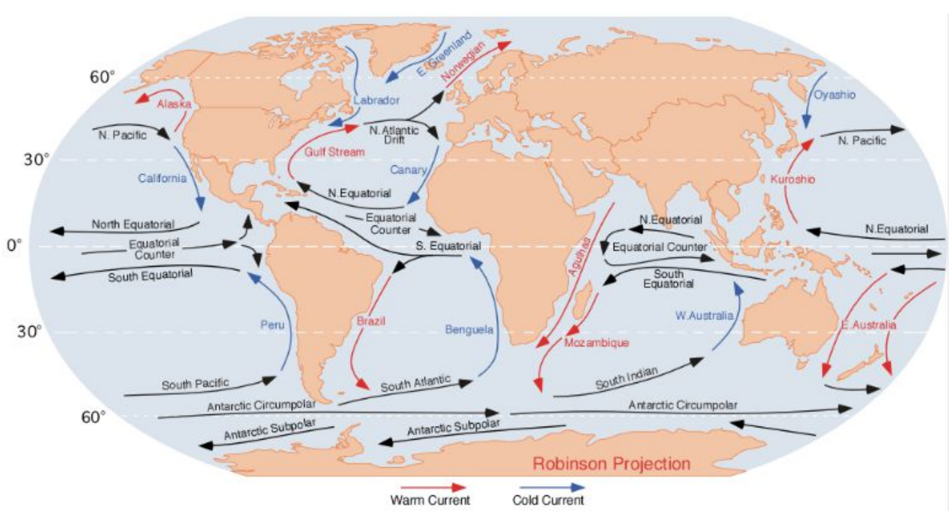


Figure 17. general atmospheric circulation

3.1.3.1. Convection cells

The main engine of atmospheric movements is the sun. This warms the Earth's surface, which in turn warms the surrounding air. Ascending movements are created, but as it rises, the air cools, approximately 1°C every 100 m in the troposphere, the layer of the atmosphere where almost all meteorological phenomena take place (Figure 18). The air then descends towards the ground. This circulation constitutes a convection current, classic in all fluids that are heated (a pot of water for example). Such circulation loops are called cells. The different cells are arranged in bands according to latitude: it is a zonal organization.

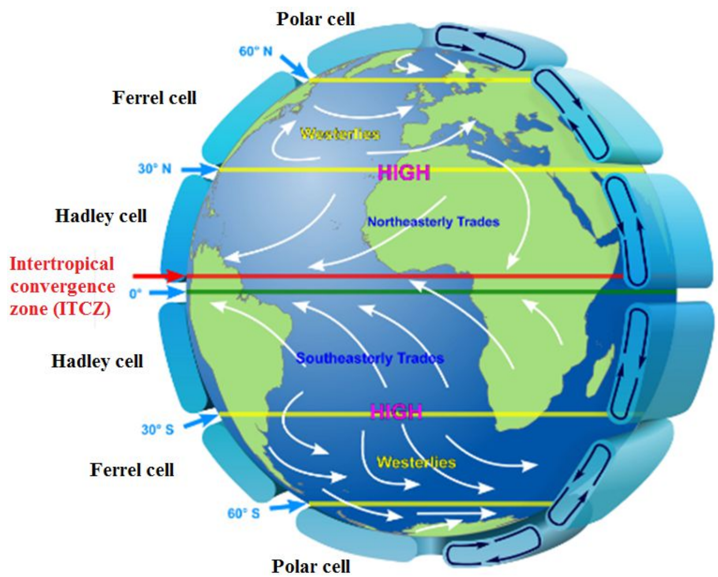


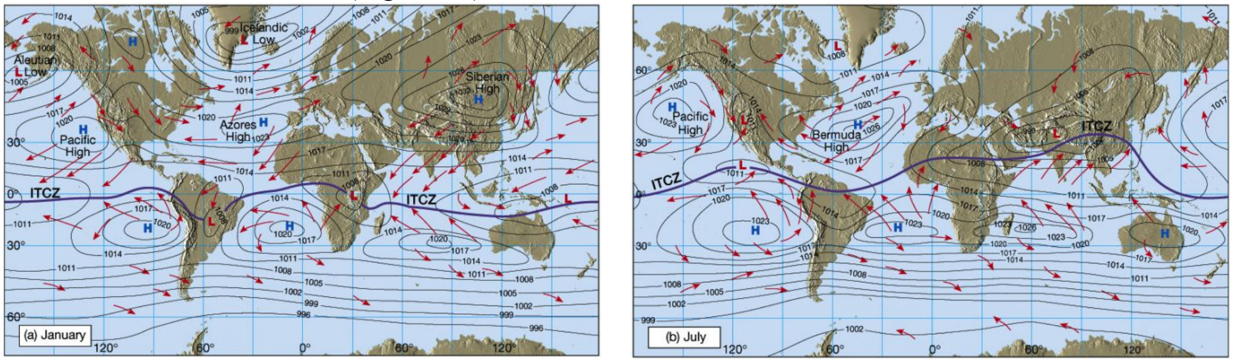
Figure 18. Convection cells

The proposed general circulation model includes six convection cells: two equatorial cells in the forward direction called Hadley cells, two cells with reverse circulation of the previous ones called Ferrel cells and two polar cells again with direct circulation.

The general atmospheric circulation thus defined ensures 70% to 80% of the transfer of energy between regions with a positive radiative balance and those with a negative radiative balance. It plays a considerable role in the water cycle, ensuring the transport of enormous quantities of water vapor. The movement of air masses conditions the climate of various regions of the planet (Sighomnou, 2004).

3.1.3.2. Seasonal variations in general atmospheric circulation

The general atmospheric circulation undergoes seasonal variations, a consequence of the position of the Earth relative to the Sun (Figure 19).



(a) January **Figure 19.** General atmospheric circulation (b) July

3.1.3.3. Hadley cells

In June-July-August, due to the inclination of the Earth in the northern hemisphere, the area which receives the most heat from the sun is located near 10° north latitude. It is therefore the southern hemisphere which has the greatest energy deficit. The South Hadley cell is then the most intense there.

As a result, the ITCZ moves northward. It brings rain to the Sahelian areas (south of the Sahara), while precipitation from temperate latitudes moves north. In December-January-February, it is in the northern hemisphere that the Hadley cell is most important. The cirrus veil over North Africa sometimes reaches Egypt. The movement of the ITCZ is towards the south. The dry season begins in the Sahel and rain falls in the northern Kalahari Desert (Bostwana), while precipitation linked to the polar front is responsible for the wet season north of the Sahara.

Over the course of the year, Hadley cells in each hemisphere move north and south in response to the position of the sun relative to the earth. The location of each cell is also influenced by temperature differences resulting from the uneven distribution of land and sea surfaces in each hemisphere. Oceans, which cover more than 70% of Earth's surface, have a greater heat capacity than continents and heat and cool more slowly. Temperature fluctuations are therefore much less significant over the oceans and the movements of the Hadley cell are relatively weak. Conversely, very large seasonal temperature changes on the continents lead to significant movement of the Hadley cell (Figure 20).

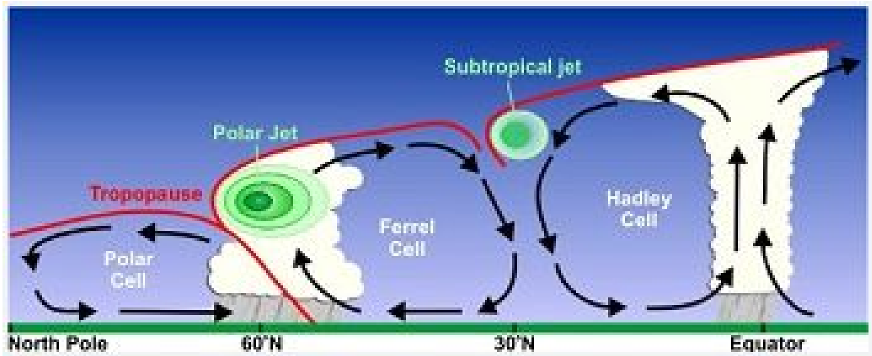


Figure 20. Circulation of air masses in the three convection cells

3.1.3.4. Tropical depression and cyclone

From the end of August and until mid-October, the cloud packs (cumulonimbus) that we see above the Gulf of Guinea can reach the American coasts and increase. We then speak of tropical depression and cyclone. For a cyclone to be created there must be a strong contrast between the temperature of the ocean and that of the atmosphere. Thus, the formation of a cyclone requires an ocean temperature above 26°C.

3.1.3.5. Polar front disturbances

Contact between cold air masses of polar origin and warm air masses of tropical origin occurs along the polar front. The polar front is not a continuous line. Its shape depends on pressure differences between continents and oceans. Its position varies in latitude depending on the seasons. It extends southward in winter (40° latitude) and is thrown northward in summer (Scandinavia) in the northern hemisphere. Polar air and tropical air constantly collide along the polar front. From this confrontation, the disturbances of the polar front are born, where "bubbles" of warm tropical air are raised and gradually integrated into the cold polar air.

The polar front therefore constitutes a place of thermal exchanges between hot zones and cold zones. A disturbance is always associated with a low pressure cell, a consequence of the lifting of warm, therefore light air by cold, dense air. A whirlwind (or vortex) appears, deepens and moves from West to East. Winds can reach up to 150 km/h depending on the severity of the depression. The low pressure zones near the 60th latitude North and South are more important in winter than in summer, which has the consequence of generating a faster and more virulent disturbed circulation from the West in winter.

3.1.3.6. The influence of continental masses

In the majority of cases observed, the main centers of action (anticyclone or depression) are split into several centers of action of greater or lesser importance. This fragmentation is the result of the thermal influence of large continental masses especially distributed in the Northern hemisphere. It therefore appears almost exclusively in the northern hemisphere.

In winter, the shortness of the day, the obliquity of the sun's rays and the persistent snow cover of the soil make the northern regions of Asia and America cold thermal centers, therefore zones of high continental pressure (anticyclone of Siberia for example) which disrupt the continuity of the low pressure axis of the 60th North latitude.

In summer, the subtropical zones of large continents store a lot of heat. As these regions are generally deserts, the air heats up very strongly. The air in the lower layers of the atmosphere (troposphere) then sees its density decrease as it heats up. Its pressure therefore tends to decrease. There is the formation of depressions of thermal origin (North America, India, Pakistan). The intertropical convergence zone then coincides more or less with these thermal depressions (in India for example). The subtropical anticyclonic belt is then interrupted.

3.2. Cloud formation

Water vapor is invisible. If you pass your hand through a thermal cloud (over a pan, which is often wrongly called smoke), you get a wet feeling.

At the top of the mountain, when we are in the clouds, we see a little drizzle all around us (lots of small droplets suspended in the air), those who have glasses will notice this very quickly.

A cloud is therefore drops of water suspended in the air (water is in the liquid state) or small ice crystals (water in the solid state) (Figure 21).

The appearance of the cloud depends on the light it receives, the nature, size and distribution of the particles which constitute it.

Figure 21. Cloud shapes

3.2.1. How do clouds form?

The heat provided by the sun causes the evaporation of water (from seas, lakes, rivers and oceans). The (invisible) water vapor rises. The maximum amount of water vapor depends on the air temperature. The warmer the air, the more water vapor it can contain (each temperature corresponds to a saturation value beyond which droplets appear.

The formation of a cloud is therefore due to the cooling of the air. How can the air cool down?

3.2.1.1. Air masses

The sun warms the ground, which in turn heats the air above it. Depending on the different situations, the air will therefore be heated more or less. There are two types of air masses: warm air masses and cold air masses (Figure 22).

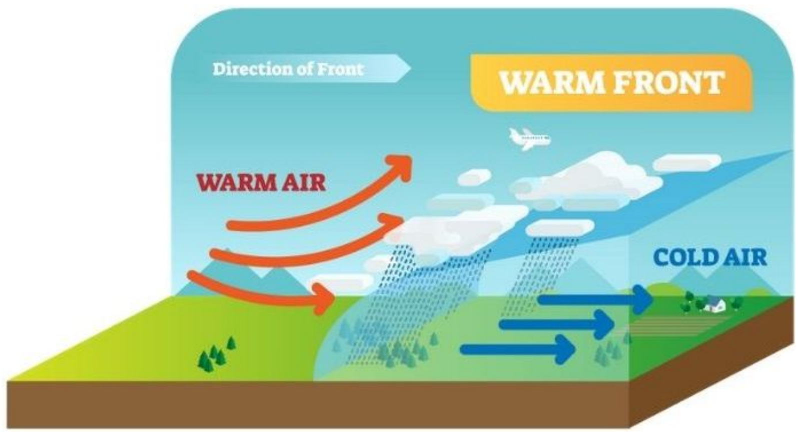


Figure 22. Air masses

3.2.1.2. The fronts

a) **Warm front:** The warm air is slowly lifted as it overtakes the cold air; the clouds tend to form in sheets, called stratus.
b) **Cold front:** The cold air mass quickly lifts the warm air mass, so high clouds, cumulus clouds, form.

3.2.1.3. Orographic uplift

On our island (Which Island ???), the northern part is mountainous. The relief forces the air mass to rise on its windward side. As the air mass rises, its temperature drops and can reach the saturation value. A cloud then forms on the windward slope (Figure 23).

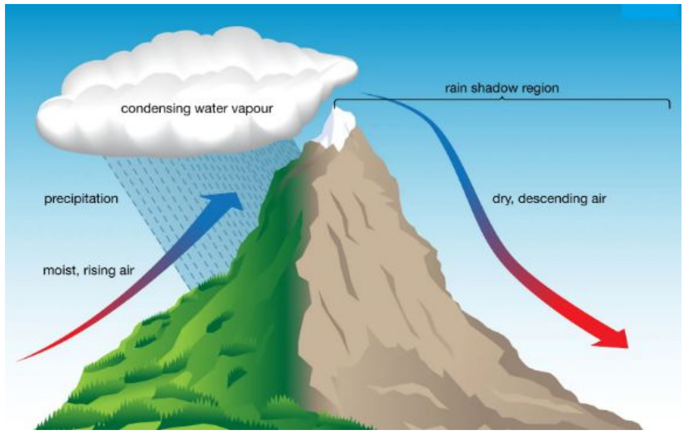


Figure 23. Orographic effect

3.2.1.4. Convection

The heating of the ground (or the oceans) is communicated to the air which, expanded and therefore lighter, begins to rise and cools (Figure 24). Convective clouds appear more easily when there is cold air at altitude (unstable air mass). The bases of such a cloud are horizontal, their tops change depending on the temperature. In the Antilles, they are present all year round.

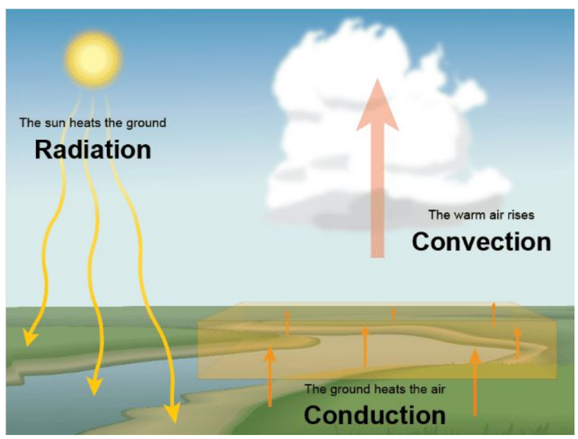


Figure 24. Atmospheric convection

3.3. Types of clouds

All of these phenomena lead to the formation of depression (low pressure zone located between two air masses). Here is the illustration: (Figure 25).

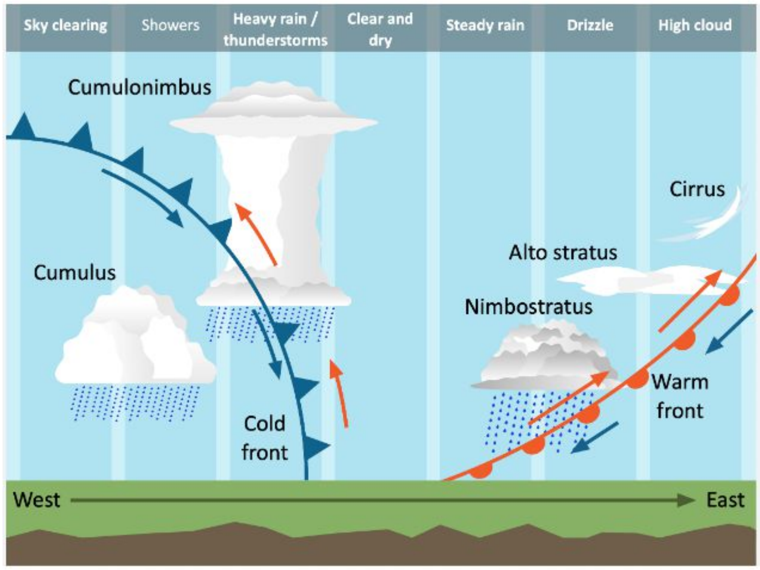


Figure 25. Atmospheric depression

Cloud names come from three aspects:

- their form, of which the three fundamental ones are:
  - **Cumulus** clouds (accumulation of droplets - in piles)
  - **Stratus** clouds (stratified-layered)
  - **Cirrus** clouds (shaped like a lock of hair)
- the altitude at which the base of the cloud forms,
  - High clouds (above 6000 m); designated by "**Cirrus**" or the prefix "**Cirro-**"
  - Clouds at mid-height (2000 - 6000 m); designated by the prefix "**Alto-**"
  - Low clouds (below 2000 m); without prefix
- their ability to produce precipitation.
  - Clouds producing precipitation have the prefix "**Nimbo-**"

Clouds are distinguished into 03 types according to the altitude of formation: (Figure 26)

- High clouds
- Clouds halfway up
- Low clouds

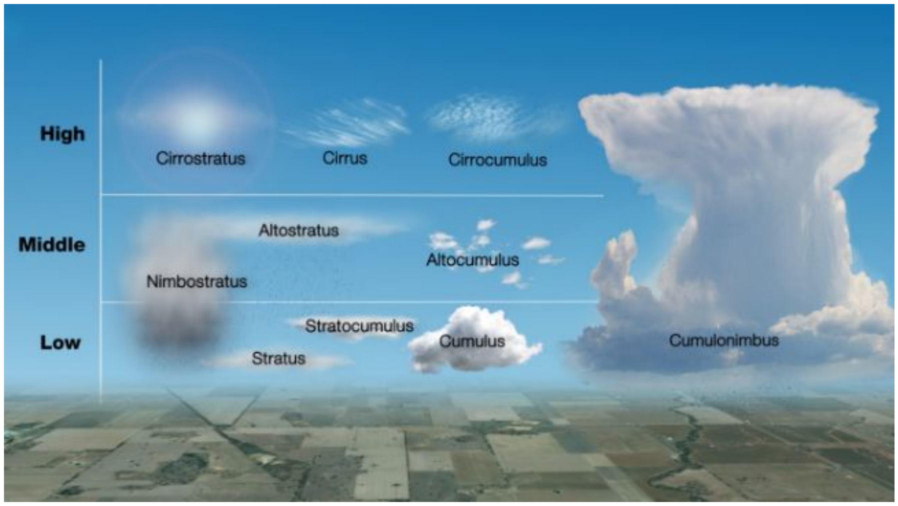


Figure 26. Different types of clouds

3.3.1. The high clouds

These clouds, from the upper level, appear between 6 and 13 km altitude in our latitudes. They are made of ice crystals.

3.3.1.1. Cirrus

These clouds have a fibrous appearance (Figure 27).

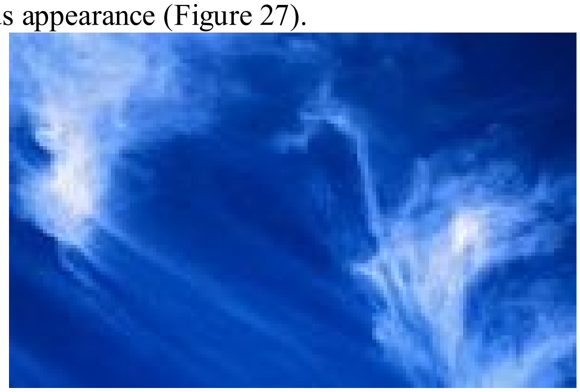


Figure 27. Cirrus clouds

3.3.1.2. Cirrocumulus

It is a sheet or thin layer cloud (Figure 28).

Figure 28. Cirrocumulus

3.3.1.3. Cirro-stratus

It is a transparent and whitish cloudy veil, with a fibrous (hairy) or smooth appearance, generally giving rise to halo phenomena (Figure 29).

Figure 29. Cirro-stratus

3.3.2. Clouds halfway up

Mid-level clouds appear between 2 and 7 km altitude in our latitudes.

3.3.2.1. Altocumulus

It is a white or gray cloud in a sheet or in a small packet (Figure 30).

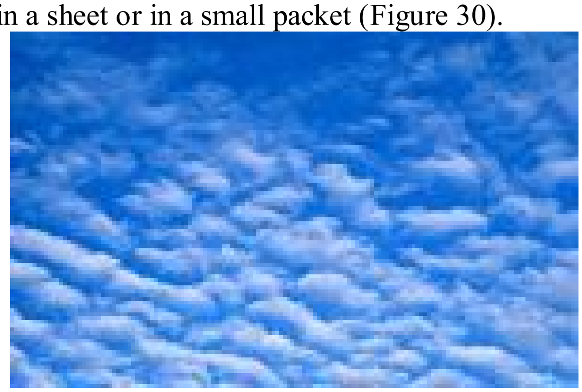


Figure 30. Altocumulus

3.3.2.2. Altostratus

It is a grayish or bluish cloud, with a streaked, fibrous or uniform appearance, can be accompanied by rain or snow (Figure 31).

Figure 31. Altostratus

3.3.3. The low clouds

Nimbostratus, cumulus and cumulonimbus: vertically developing clouds which can occupy several floors at the same time.

3.3.3.1. Stratus

It is a gray cloud layer (Figure32).

Figure 32. Stratus

3.3.3.2. Nimbostratus

It is a gray, often dark cloud layer (Figure 33).

Figure 33. Nimbostratus

3.3.3.3. Cumulus

These are separate clouds, generally dense and with well-defined white outline (Figure 34).

Figure 34. Cumulus

3.3.3.4. Cumulonimbus
It is a dense and powerful cloud, with considerable vertical extension, shaped like a mountain or enormous towers. It is often accompanied by precipitation of all kinds (Figure 35).

Figure 35. Cumulonimbus

3.3.2.5. Stratocumulus
It is a gray or whitish cloud, almost always having dark parts, composed of slabs, pebbles, rollers, etc., (Figure 36).

Figure 36. Stratocumulus

3.4. Effect of a cloudy sky

3.4.1. The day
During the day, the Earth's surface is heated by the Sun. If the sky is clear, almost all of the Sun's rays reach the ground. The ground heats up and in turn warms the air above it. This is how the air around you heats up (Figure 37).

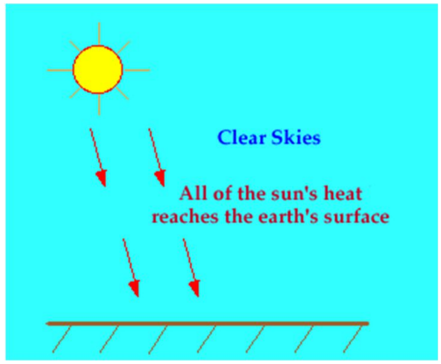


Figure 37. Cloudy sky effect (day)

On the other hand, if the sky is cloudy, part of the Sun's rays are reflected by the clouds (by water droplets and ice crystals) towards space. There will therefore be fewer solar rays reaching the ground to heat it. In other words, the ground will warm less if there are clouds than if there are no clouds. The surrounding air temperature will be lower (it will be cooler). If cloudy skies are forecast for the day, this means that the air temperature will be low.

3.4.2. At night

During the night, cloudy skies cause the opposite effect on air temperature. If the sky is clear, the rays emitted by the Earth's surface escape into space and the ground cools quickly.

If the sky is cloudy, part of the rays emitted by the Earth's surface are absorbed by the clouds. The clouds will in turn emit energy towards space and towards the Earth in the form of radiation. The ground absorbs the rays emitted by the clouds and warms up a little. Subsequently, the ground heats the air above it.

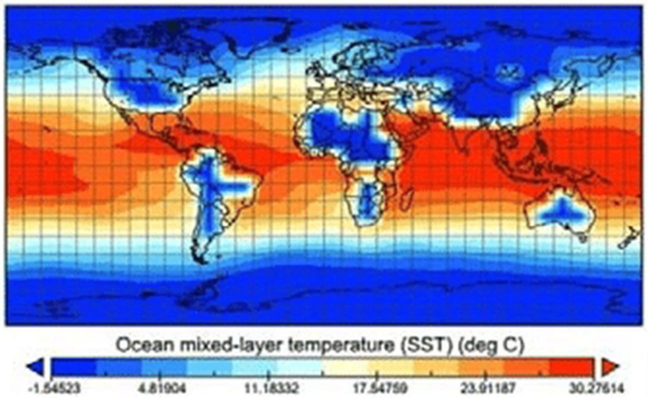
So if the night is cloudy, the air temperature cools less quickly than if the night was clear. This means it will be warmer that night. If cloudy skies are forecast for the night, we should expect to have a mild night.

3.5. Oceanic circulation

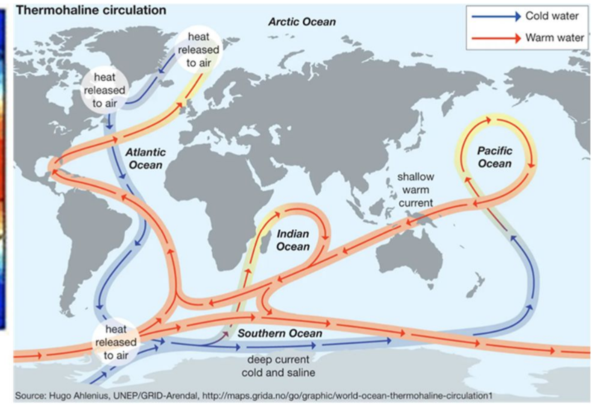
Also called thermohaline circulation (THC), it designates the movement of oceanic water masses under the action of the main deep currents, the Gulf Stream being one of the best known (Figure 38a). This circulation depends on temperature, salinity (hence its name) but also winds and tides (Figure 38b). The most studied phenomenon is the AMOC (Atlantic Meridional Overturning Circulation) which is observed in the Atlantic basin. This influences the regional climate, bad weather, temperatures, etc.

The oceanic system being more stable than the atmospheric system, it is less complex to study but without being easier to predict.

The ocean also plays an important role in the redistribution of energy from the tropics to high latitudes.



a)



b)

Figure 38. (a) Annual climatology of sea surface temperature (SST, Sea Surface Temperature in °C) calculated in the ERSST3 dataset. (b) Diagram of thermohaline circulation (taken from https://global.britannica.com/science/thermohaline-circulation)

Tropical oceans are much warmer than extratropical oceans. Thus, heat is transported much more slowly by the ocean than by the atmosphere: we speak of the thermal inertia of the oceans. 38a shows the annual distribution of global sea surface temperatures (SST). We observe more marked warming in tropical regions than in mid and high latitudes. In addition, in the tropics, the oceans are warmer at the western edges than at the eastern edges. This spatial structure is largely explained by surface circulation, mainly controlled by winds which explain the presence of cold water rising (called upwelling) on the eastern edges of the oceans. Note also the strong temperature gradients in the North Atlantic and the North Pacific at mid-latitudes which correspond to the meeting of cold subpolar waters and warm ocean currents (Gulf Stream for the Atlantic and Kuroshio for the Pacific). Large-scale ocean circulation is therefore responsible for significant heat transport (Figure 38b). Transport is different depending on the ocean basin (Trenberth et al., 2001). It is characterized, in the Atlantic, by a northward transport of warm and salty surface waters.

In the northern Atlantic regions, these waters are very dense and plunge towards the deeper layers of the ocean. This convection takes place in the Labrador Seas and in the subpolar gyres and the waters then move southwards at depth. AMOC (Atlantic Meridional Overturning Circulation) refers to the thermohaline circulation in the North Atlantic. It is often estimated as the maximum of the meridional current function at 30°N. It is closely linked to deep convection in subpolar regions: when convection is strong, the AMOC accelerates and vice versa. The AMOC is very important in the climate system, on the one hand because it redistributes a large part of the energy and on the other hand because it generates strong turbulent flows in the North Atlantic between the ocean and atmosphere.

It is very likely that the effects of climate change have led, since the 1950s, to a salinization of surface waters in mid-latitudes, due to high evaporation, and unlike, in tropical and polar regions, to softening of these surface waters due to precipitation. The average difference between high and low salinity regions thus increased from 1950 to 2008. Thus, the salinity of the Atlantic has increased while that of the Pacific Ocean and the Southern Ocean has decreased. This evolution of water salinity has an influence on the phenomenon of oceanic convection, when cooled and salty waters plunge towards the seabed in the North Atlantic.

The climate and its variations at the different temporal scales of a given region are determined by atmospheric circulation. Let us take the example of the North Atlantic Oscillation (NAO), which is the dominant mode of variability in the North Atlantic basin, and which exerts a considerable influence on the climate in Europe-Mediterranean. The impact of the NAO on the mean temperature and precipitation fields is reported by numerous studies (Trigo et al., 2002; Castro-Díez et al., 2002; Xoplaki et al., 2003). However, these studies reveal that the NAO, which is certainly the dominant mode of variability, only explains at most half of the total variance.

Indeed, the fluxes of solar energy received by ocean waters in tropical latitudes cause the warming of surface waters. To evacuate these excessive energy supplies, the heated water masses move under the effect of the winds towards the colder regions of the North Atlantic. This is how surface ocean currents are organized by carrying significant quantities of energy. Parallel to these surface currents, there is thermohaline circulation. The cold and salty waters of the North Atlantic, due to their density, plunge towards the ocean floor and will circulate there slowly, from the Atlantic to the Indian Ocean and the Pacific, during a cycle estimated to be around a thousand years old. Finally, they will warm up and rise to the surface to mix with the warmer waters circulating on the upper level. These different movements form a cycle which influences a number of climatic parameters and maintains the temperature of the globe. It is important to emphasize that the difference between the very long response times at the ocean level compared to those on the continents can lead to prolonged climatic anomalies (Sighomnou, 2004).

The analysis of the different weather elements is based first on observations made on the ground or at altitude. The need for air navigation, hence the need for short-term forecasting, has led to the creation of permanent weather stations, interconnected and constitute a network managed according to the directives of the World Meteorological Organization (WMO).

To define climates, we will therefore have to constantly rely on average data, as well as knowledge of extremes. This allows the amplitudes (amplitude of the absolute extremes of daily precipitation) to be measured. In addition, it is essential to research seasonal or instantaneous climate variability (Estienne et al., 1970).

All these definitions will be the entire better founded if they are based on longer and more homogeneous series of observations. Greenhouse gases outweigh natural climate variability. A new study shows that atmospheric circulation in the Southern Hemisphere is being altered by climate change. El Niño events no longer have the same impact on winds; the increase in global temperature is the most influential parameter (Benzater, 2021).

In the climate system, everything is linked. Thus, the modes of climate variability interact with each other. For example, El Niño events cause the SAM to enter a negative phase, that is, to attenuate atmospheric pressure at mid-latitudes. One of the big questions today is whether climate change modifies patterns of natural variability, such as El Nino, the NAO or the SAM. Although it is difficult to answer, Wenju et al. (2018) recently demonstrated that there was a link between changes in atmospheric circulation associated with the SAM and the increase in global temperature.

CHAPTER IV

HYDROCLIMATOLOGY AND WATER RESOURCES

IV- Chapitre 4: Hydroclimatology and water resources

4.1. Evaluation of hydroclimatic parameters

4.1.1. Precipitation

Until now we have assumed that when the air becomes saturated, the liquid water that appears remains in the volume element. This is only valid to the extent that the droplets can remain in suspension. If this is not the case, precipitation will appear (in liquid or solid form).

The measurement of precipitation is one of the most complex in meteorology because we observe a strong spatial variation depending on the movement of the disturbance, the location of the downpour, the topography and local geographical obstacles hindering its capture. Precipitation is generally expressed in height or blade of precipitated water per unit of horizontal surface area (mm). If we relate this water height to the unit of time, it is intensity (mm/h). Remember that: 1 mm = 1 l/m^2 = 10 m^3/ha. The measurement accuracy is at best around 0.1mm. In Switzerland, any precipitation greater than 0.5 mm is considered effective rain.

The cloud is an aerosol that can consist of air, water vapor, liquid water droplets and ice crystals. The droplet dimensions are very small. We assume that their diameter is of the order of 5 to 30 μ and their spacing of 1 mm. Their fall speed in calm air would be a few millimeters per second; however, the clouds are animated by turbulence whose instantaneous speeds are without any common measure (several meters per second). For water particles to fall, their speed must be significantly greater than the speed of the ascending currents (Figure 39).

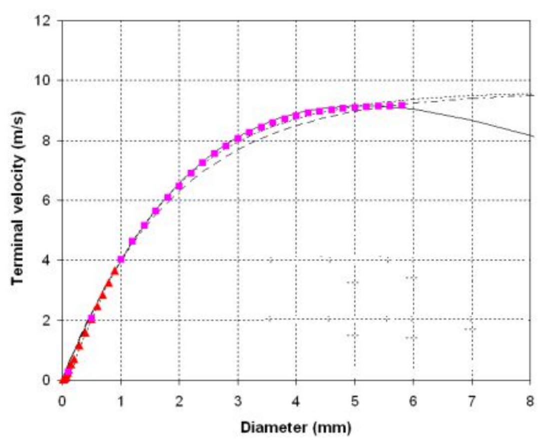


Figure 39. Rainfall velocity function and drop diameter

In Figure 39, we see that the drops must have at least a diameter of 0.5 mm to be able to cause rain. To form a raindrop, you therefore need approximately 106 elementary droplets.

4.1.1. Triggering of precipitation

The agglomeration of droplets (mechanisms rains) would occur, according to Figure 40, (the most probable and most generally accepted) (Laborde, 2013):

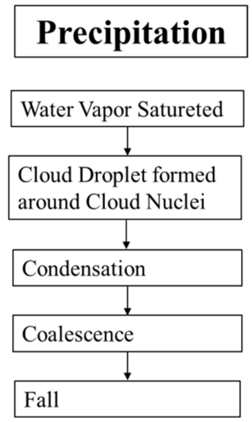


Figure 40. Mechanisms rain

Let's assume a "cold cloud" in which we encounter both ice crystals and supercooled water. The saturation vapor pressure being different in these two cases, the drop of supercooled water will vaporize in favor of the ice crystals. TOR-BERGERON were able to show that at -10º, all the liquid water in a cloud can gather in 20 minutes on ice crystals when their density is 1 crystal/cm^3. We thus obtain drops of approximately 0.1 mm in diameter.

In other cases where there is no presence of ice crystals, we think that the phenomenon is similar to that described by TOR-BERGERON. The role of crystals would be played by germs such as salt crystals, dust, etc.

Possibly, temperature differences between droplets could also explain preferential condensation from "hot" drops over "cold" ones. Thus, the drops can reach a diameter of 0.1 mm and acquire a sufficient falling speed to grow by "sweeping" other droplets. The drop thus initiated will increase in volume as it falls and increase its speed.

4.1.2. Precipitation study

The study of precipitation shows that, according to Figure 41, the clouds would originate above the oceans, then pushed by the winds would fall as rain on the continents, is false.

Indeed, we can assume that a cloud originating above the Atlantic pours approximately 100 mm onto Europe and Western Russia before dissipating beyond the Urals. However, a cloud only contains a maximum of around 20 mm of water. It is therefore only at most 20 mm which comes from the oceans and 80 mm which comes from the atmosphere above the continents.

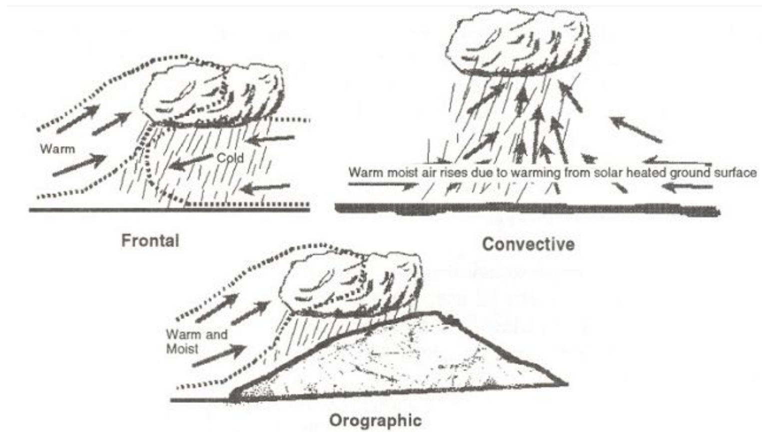


Figure 41. Different types of rainfall

In addition, there are frequent showers exceeding 20 mm. The cloud must therefore be continually replenished with water. This occurs when there are upward winds that carry masses of moist but unsaturated air toward the cloud formation zone. Such conditions arise when one or more of these three main cases are encountered:

4.1.2.1. Convective rainfall

If an air mass heats up near the ground, the temperature profile will change by increasing its gradient. There will then be instability and the appearance of convection cells. The moist, warm air will rise, relax and cool. When the dew point is reached, a cloud (cumulus) forms and if the lift is sufficient, a sufficient altitude can be reached to trigger precipitation (Figure 42). This type of rain corresponds to most precipitation in equatorial regions; it is also encountered in temperate climates in the form of summer thunderstorms.

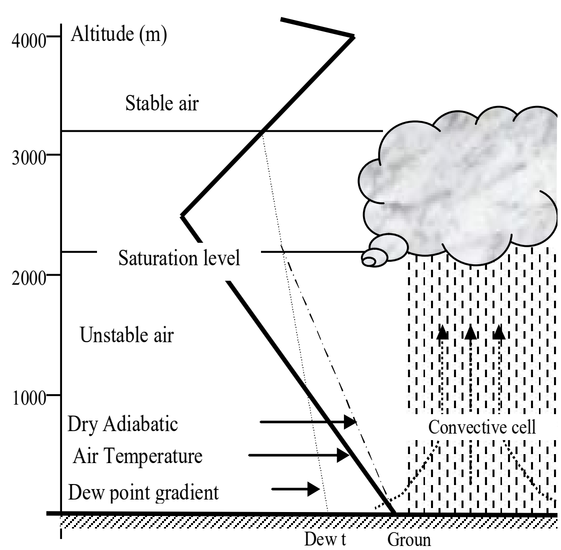


Figure 42. Convective rainfall

4.1.2.2. Orographic rainfall

If a horizontally moving air mass encounters a topographic obstacle (mountain range for example), the air masses rise and consequently cool. As before, we obtain precipitation in the form of rain but also, if the altitude is sufficient, snow. After passing the chain, the air will go down, compress and heat up. We then have hot and dry winds ("foehn" effect) (Laborde, 2013) (Figure 43).

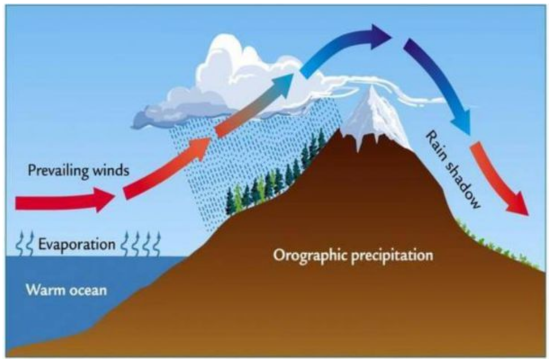


Figure 43. Orographic rainfall

4.1.2.3. Frontal rainfall

When several air masses of different properties meet, the warmest and most humid ones are pushed towards high altitudes where they cool and condense (Figure 44). These are the heaviest, longest and most frequent precipitations in temperate climates.

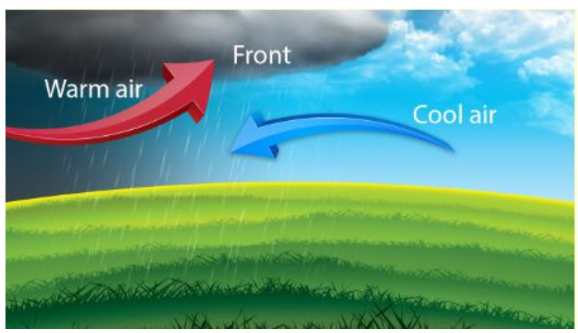


Figure 44. Frontal rainfall

4.1.3. Different types of precipitation

Precipitation depends on the diameter of the water drops and the speed of precipitation (table 3).

Table 3 Types of rain depending on drop diameter

Rainfall type	Drop diameters (mm)	Fall velocity (cm/s)
Drizzle	0,006 – 0,06	0,10 – 20
Light rain	0,06 – 0,6	20 – 100
Continuous rain	1 – 3	150 – 400
Downpour	4 – 6	500 – 800

Depending on the diameter of the water drops, we can distinguish the following different types of precipitation (Figure 45).

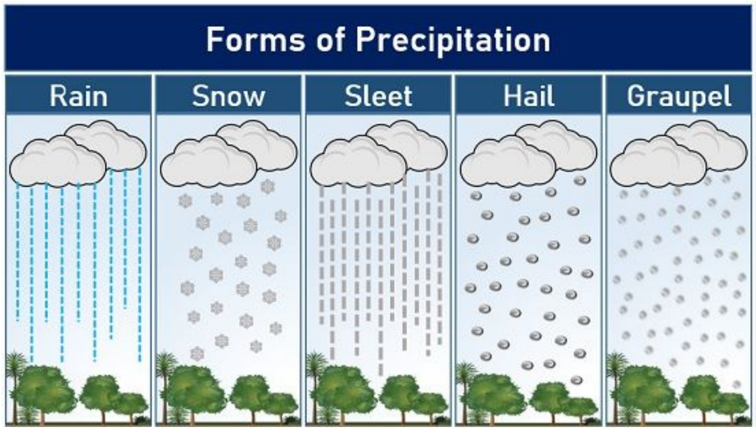


Figure 45. Different types of precipitation

4.1.4. Solid precipitation

They mainly occur in two forms:

4.1.4.1. Hail

It forms in cumulonimbus clouds around 5,000 m altitude with strong turbulence. The quantities of supercooled water that these clouds can contain suddenly solidify on contact with ice crystals. This phenomenon is relatively poorly understood. Its geographical extent is generally small (for example: 1 to 2 km by 10 to 15 km).

4.1.4.2. Snow

It is the main form of solid precipitation. It results from a slow and progressive condensation of water vapor at a temperature close to 0ºC. This condensation initially takes place in crystals shaped like a six-pointed star. If the crystals undergo partial melting, they clump together as they fall to form flakes.

4.2. Solar energy

The sun emits approximately constant energy. The average flux across a surface normal to the solar rays, located at the upper limit of the atmosphere, is called the "solar constant". At the average earth-sun distance (149.106 km), we assume that this constant is 1.396 k W/m^2.

Over time, solar energy arriving at the edges of the atmosphere undergoes:

Seasonal variations which are due on the one hand to the modification of the earth-sun distance between winter and summer and on the other hand, to the variation in the average incidence of solar rays. (These variations partly contradict each other);

Daily variations are due to the variable incidence of solar rays during the day and their absence during the night.

This energy flow is composed of radiation of different wavelengths (comparable to the radiation of a black body at 6,000º K). The spectrum extends very widely from ultraviolet to infrared. We can assume the following distribution of the powers emitted: 8% in the ultraviolet ($L < 0.4\ \mu$), 41% in the visible ($0.4 < h < 0.7\ \mu$) and 51% in the infrared ($h > 0.7\ \mu$).

4.3. Metrology used in hydroclimatology

Precipitation recording devices can be classified into several types:

4.3.1. Rain gauges

The rain gauge is the basic instrument for measuring liquid or solid precipitation. It indicates the overall precipitation precipitated in the time interval between two readings. The rain gauge is generally taken once a day (in Switzerland, every morning at 7:30 a.m.). The rainfall amount read on day d is assigned to day d-1 and constitutes its "daily rain" or "rain in 24 hours". If the rain gauge station is remote or difficult to access, it is recommended to use the total rain gauge. This device receives precipitation over a long period and the reading is done by measuring the height of water collected or by weighing. In the event of snow or hail, melting is carried out before measurement.

A rain gauge consists of a ring with a chamfered edge, the opening which surmounts a funnel leading to the receiver (bucket). To standardize methods and minimize errors, each country had to set the dimensions of the devices and the installation conditions. However, each country has its own type of rain gauge, although the characteristics are not very different. In France, the SPIEA type is used (receiving surface of 400 cm^2); in Switzerland, we use the Hellmann type rain gauge, with a surface area of 200 cm^2 (Figure 46).

Figure 46. Hellmann rain gauge

The quantity of water collected is measured using a graduated cylinder. The height above the ground of the rain gauge ring is also decisive for correct rain measurement. In fact, the effects of the wind create a water deficit, in the event that the rain gauge is in a high position. Also, despite the collection errors, the WMO standards (1996) recommend that the receiving surface of rain gauges (and rain gauges) be horizontal and located 1.50 m above the ground; this height allows the device to be easily placed and avoids splashing.

4.3.2. Pluviograph

The pluviograph differs from the rain gauge in that the precipitation, instead of flowing directly into a collecting container, first passes into a particular device (float tank, buckets, etc.) which allows automatic recording of the rain's instantaneous height precipitation. The recording is permanent and continuous, and makes it possible to determine not only the height of precipitation, but also its distribution over time and therefore its intensity. Pluviographs provide plots of cumulative precipitation amounts versus time. There are two main types used in Meteorology.

4.3.2.1. The siphon pluviograph

The accumulation of rain in a cylindrical reservoir is recorded by the elevation of a float. When the cylinder is full, a siphon is activated and empties it quickly. The movements of the float are recorded by a rotating drum at constant speed, surrounded by paper, and determine the plot of the rain gauge.

4.3.2.2. Tipping bucket pluviograph

This device (Figure 47) includes, below its water collection funnel, a pivoting part whose two compartments can receive water in turn (tipping buckets). When a specific weight of water (generally corresponding to 0.1 or 0.2 mm of rain) has accumulated in one of the compartments, the seesaw changes position: the first bucket empties and the second begins to fill (Figure 47). The tilts are counted either mechanically with recording on paper wrapped around a rotating drum, or electrically by pulse counting (for example MADD system): device allowing the acquisition of events in real time, developed by HYDRAM in 1983. Tipping bucket pluviographs are currently the most precise and most widely used (Figure 47).

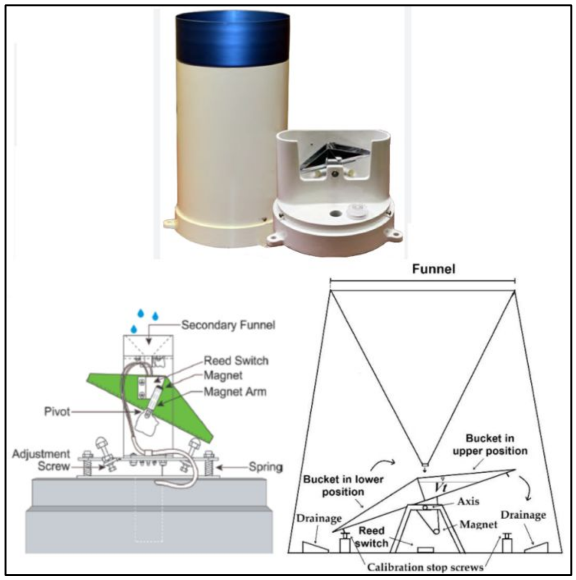


Figure 47. Pluviograph, tipping buckets and MADD recording system

4.3.2.3. Radar

Radar (Radio Detection And Ranging) has become an essential investigation and measurement instrument in atmospheric physics. The measurement of precipitation is made possible by the strong influence that hydrometeors exert on the propagation of short-wavelength electromagnetic waves (Figure 48). Radar thus makes it possible to locate and follow the movement of clouds. Some radars can estimate the intensity of precipitation, although with some difficulties due to calibration.

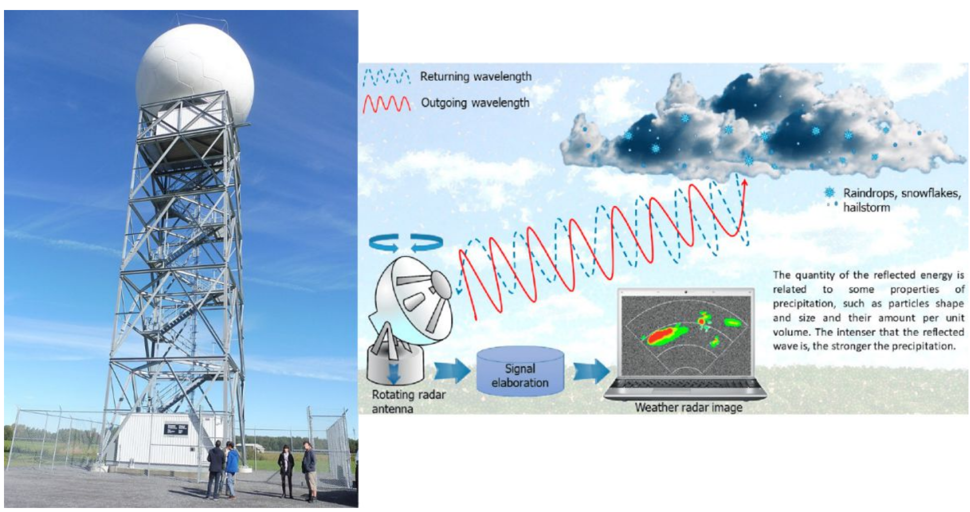


Figure 48. Weather radar principle of function

The essential advantage of radar, compared to a conventional network of pluviographs, lies in its capacity to acquire, from a single point, information on the state of precipitating systems affecting a vast region (--> 105 km^2). The range of a radar oscillates between 200 and 300 km.

However, many sources of error affect the quality of radar precipitation estimates. One of the sensitive points is the need to find an average relationship for the transformation of target reflectivities into precipitation intensity. Despite the uncertainty of the results, radar is one of the only instruments allowing real-time measurement over an entire watershed and is, therefore, very useful for real-time forecasting. It allows a good representation of phenomena within a radius of approximately 100 km (Laborde, 2013).

4.3.3. Snow gauge

Snow gauge (Figure 49) provide a vague idea of the water equivalent of snow precipitation. This is not sufficient, especially in mountain areas where it is interesting to know both the quantitative importance of the snowpack but also its condition (avalanche forecast, flood forecast, etc.).

Figure 49. Snow gauge

4.3.4. Solar radiation and duration of insolation

Solar radiation reaching the ground is commonly measured. The measurements relate on the one hand to the intensity of the direct radiation, and on the other hand to the overall radiation both in the form of diffuse radiation and in the form of direct radiation. The instruments used are referred to under the general name of actinometers. To measure net radiation, thermopile, blade or, more rarely, distillation pyranometers are used.

There are several devices, called heliographs, which evaluate the total duration of insolation for a station every day (Figure 50). They determine the sum of time intervals during which the intensity of direct solar radiation has exceeded a certain threshold.

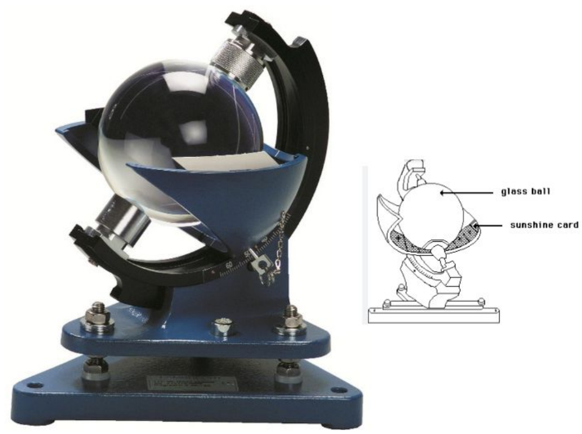


Figure 50. Campbell’s Heliograph

4.3.5. The temperature

4.3.5.1. Temperature measurement

Any thermometer only measures its own temperature. Particular care must therefore be taken so that it is in thermal equilibrium with the environment whose temperature we want to measure. Whether for measuring air or water temperature, it is therefore appropriate to protect the device from direct or indirect solar radiation. Air being a very poor conductor of heat, it is necessary to renew the air in contact with the thermometer; the shelter must therefore be ventilated. In our climates, natural ventilation through louvered walls is considered sufficient. Furthermore, the shelter will be painted white, glossy if possible, so as to limit its heating. Finally, the temperature measurement will be taken approximately 1.5 m from the ground.

For hydrological purposes, a mercury thermometer at 0.1°C is sufficiently precise, but to obtain average daily or monthly temperatures, it would require too frequent readings. We can then use maximum and minimum thermometers.

The maximum thermometer (Figure 51) is a mercury thermometer with a constriction at the outlet of the tank. When the temperature increases, mercury passes through it easily.

On the other hand, when the temperature decreases, the mercury splits and therefore maintains the indication of the maximum temperature reached. For the device to work properly, it should be installed in a position close to horizontal (approximately 2°). To reduce splitting after reading, simply "centrifuge" the device by hand.

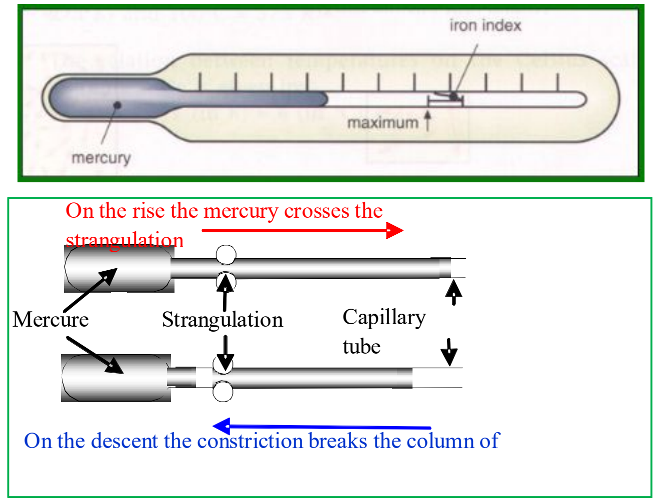


Figure 51. Maximum Thermometer

The minimum thermometer (Figure 52) is generally an alcohol thermometer whose capillary contains a small freely moving index finger. If the temperature increases, the alcohol rises in the capillary, flowing around the index finger which does not move. On the other hand, if the temperature drops, the alcohol will flow around the index finger until the meniscus reaches it. The capillarity forces are then sufficient for the meniscus to drag the index finger in its descent.

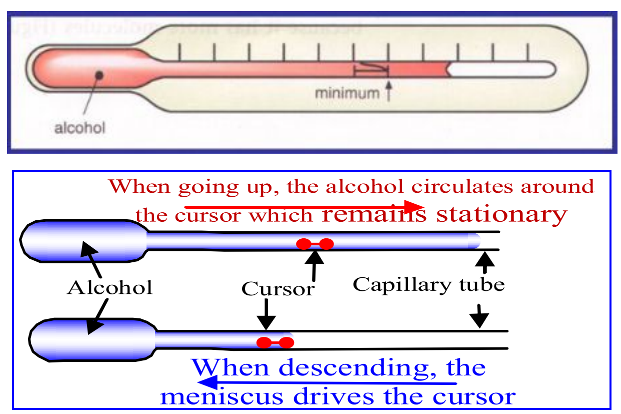


Figure 52. Minimum Thermometer

We generally see that the minimum temperature is reached some time before sunrise while the maximum temperature is observed around noon in the sun. Recording the maximum and minimum temperatures at the end of the day makes it possible to evaluate, to within a few tenths of a degree, the average daily temperature: $\bar{\theta}_j = (\theta_{min} + \theta_{max})/2 \pm 0.1°$

When a more detailed knowledge of the evolution of temperatures is necessary or if we do not have permanent observers, we use a thermograph. The sensitive organ is either a metal blade whose expansion is amplified, or two blades made of two metals with expansion coefficients as different as possible (bimetallic thermograph). The movements are amplified mechanically and recorded by a pen on a recording paper driven by a clockwork movement. Most often, recordings are weekly.

In meteorological stations, platinum resistance thermometer probes are very frequently used. Within the measurement range, the resistance of the platinum wire varies linearly with temperature. To measure water temperatures with a view to estimating evaporation, a simple mercury thermometer is used, submerged a few millimeters, but protected from solar radiation. In recent years, thermoelectric field probes have been developed. A digital display allows easy reading.

Measuring air temperature requires some precautions due to the disturbing effects, mainly those of radiation. It is therefore necessary to protect the thermometer by placing it under a weather shelter (Figure 53). These meteorological shelters generally house other instruments such as a barograph or a psychrometer for example. The shape and position of the shelter are standardized (2 m). The shelter must be painted white, with the door facing north and blinds (WMO standards).

Figure 53. Weather shelter

4.3.6. Air humidity

The psychrometer (Figure 54) consists of two mercury thermometers, one normal called "dry" and the other called "wet" whose reservoir is surrounded by foam moistened with water. The dry thermometer then indicates the temperature of the ambient air (t) while the wet thermometer records a lower temperature (t') due to the evaporation of water from the foam. The evaporation is all the more intense and this lower temperature than the air is drier. Obviously, for the measurement to be representative, it is necessary to prevent the steam emitted by the wet foam from stagnating around the thermometer, which would disrupt the measurement.

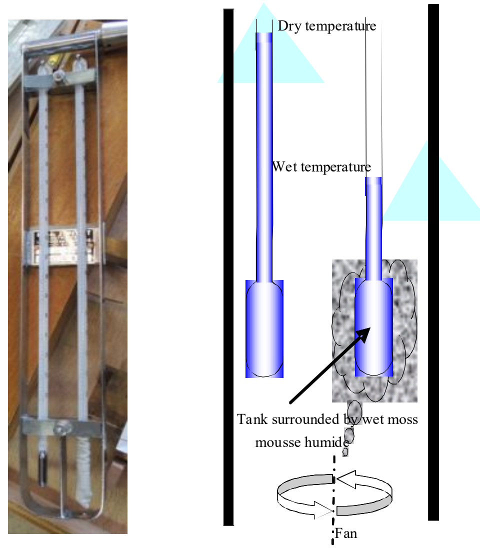


Figure 54. Psychrometer

We therefore preferably use forced ventilation psychrometers where air is introduced into the device by a small turbine driven by a spring motor. The advantage of this type of device is that it allows measurements that are completely reproducible and independent of the ventilation of the shelter or the movement of the observer.

4.3.7. Atmospheric pressure

There are various instruments that measure atmospheric pressure. We first distinguish the liquid barometer; mercury is most often used because of its density 13.6 times greater than that of water. We sometimes use a mechanical or aeroid barometer, installed under a meteorological shelter. It can be attached to a recording system (pen); we thus obtain a barograph measuring pressure as a function of time (Figure 55).

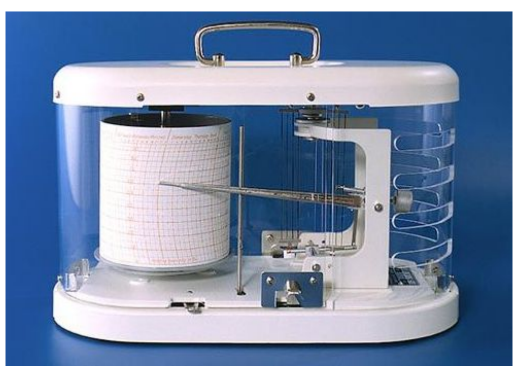


Figure 55. Baro-thermo-hydrograph

4.3.8. The wind

Wind measuring instruments are of two types; some evaluate speed, others direction. On the surface, anemometers measure wind speed. They are installed 10 meters above the ground, in a location clear of any obstacle (building, tree, etc.). The most frequently used are totalizing anemometers, made up of three or four branches ending respectively in a hemispherical cup. The system is also attached to a recording device to form a unit called an anemograph. For measurement at tropospheric altitude, we use a balloon filled with hydrogen which rises into the atmosphere. Knowing its speed of ascent and its horizontal displacement as a function of time, we easily calculate the speed of the wind which carries it. The direction of the wind is determined using a weather vane or a windsock. The wind direction is given according to the cardinal points.

Wind measurement is carried out at meteorological stations by anemometers recording instantaneous speeds, coupled with a weather vane giving the direction of the wind. For the hydrologist and except in special cases, we can make do with totalizing anemometers (Figure 56). In general, they have four hemispherical cups of 44 mm in diameter. The rotational movement caused by the wind, whatever its direction, is transmitted by an axis and a gear system to a counter directly indicating the number of kilometers traveled by the wind. Simply read this meter at the desired time interval (around once or twice a day) to be able to calculate the average wind speed.

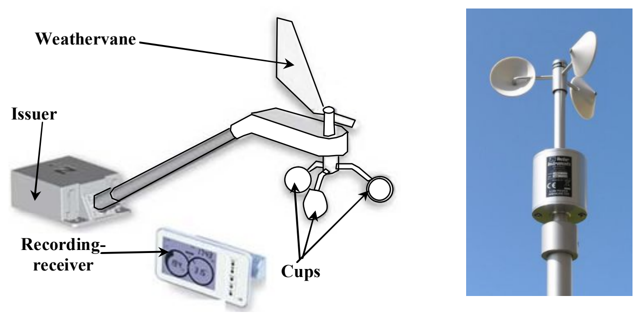


Figure 56. Anemometer

Wind speed varies significantly near the ground. Anemometers are generally placed 10 m above flat ground and at a distance from any obstacle equal to at least ten times the height of this obstacle.

4.3.9. Evaporation

Measurements of "evaporation" can be done in different ways depending on the goals pursued: estimation of evaporation from a reservoir, estimation of potential evaporation. Sometimes we even want to evaluate all of the evaporation and transpiration by the soil-plant system that is to say directly the real evapotranspiration.

4.3.9.1. Measurements of evaporation from a free surface

Different types of devices have been designed but with their faults and their qualities. The most used are:

4.3.9.1.1. Class A evaporation pan (Weather Bureau, U.S.A.)

This pan consists of a metal cylinder 121.9 cm in diameter and 25.4 cm in height (Figure 57). In this cylinder, a water thickness of 17.5 to 20 cm is maintained. The cylinder is supported by a grating 15 cm from the ground. The grating must allow good ventilation under the pan. This universally used tank only very partially satisfies the hydrologist because, due to its location in relation to the ground, it is very sensitive to temperature variations, its thermal inertia being low.

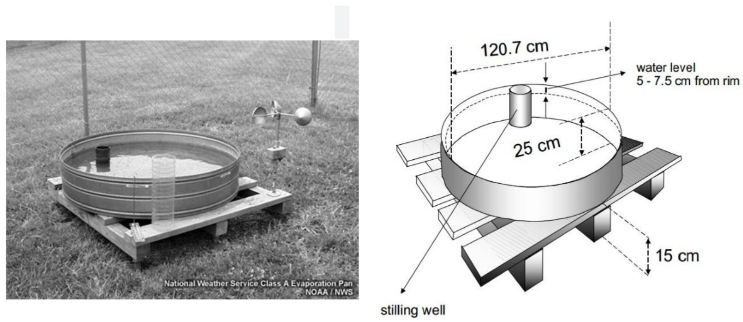


Figure 57. Class A evaporation pan

4.3.9.1.2. Colorado pan and ORSTOM pan

The Colorado pan and the ORSTOM pan which derives from it are square section tanks with a side of 92.5 cm (1 m for the ORSTOM tank), a height of 60 cm and buried 50 cm (Figure 58). The water is kept approximately 10 cm from the edge, or approximately at ground level. This device being buried and with a greater thickness of water, it has greater thermal inertia and is closer to natural conditions.

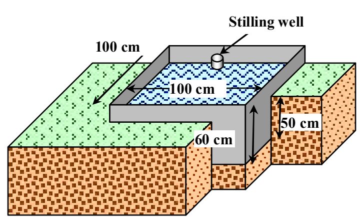


Figure 58. Colorado pan

4.3.9.1.3. CGI 30 pan

Similar in design to the Colorado pan, this tank, of Soviet origin, is the one recommended by the WMO. It is a cylinder 61.8 cm in diameter (3,000 cm^3) with a conical bottom. 60 cm deep, it is buried so that its collar protrudes 7.5 cm from the ground, the water level being kept bare at ground level.

4.3.9.2. Measuring evaporation from porous surfaces: atmometers

Altometers are intended to measure a broad characteristic of the evaporative power of ambient air. These devices should therefore have the following qualities: low thermal inertia, evaporating surface, flat, horizontal and with black body behavior, low disturbance of the wind speed field, not modifying the relative humidity of the ambient air in the vicinity of the device.

The "Black Bellani" is one of the devices that best corresponds to the qualities required of an atmometer. Evaporation takes place from a porous black porcelain surface 7.5 cm in diameter. This cup is supplied with water from a reservoir which is also used to measure the quantity of water evaporated. The small dimensions of the device make it possible to obtain low thermal inertia, and the reduced evaporation does not disturb the ambient humidity (Figure 59). Finally, the black color of the evaporating surface makes it possible to capture radiation over almost the entire spectrum. This device is installed without protection 2 m from the ground, in a representative area.

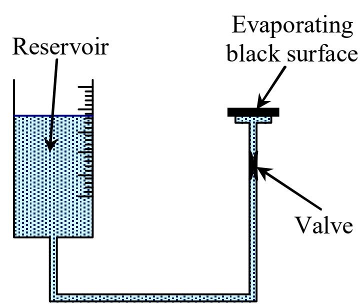


Figure 59. Altometer « Black Bellani »

The "Piche"

Among a large number of other atmometers, we would like to point out the Piche atmometer (Figure 60), although it only imperfectly meets the qualities required of a measuring device. It is used very frequently by agronomists. Its use is justified by the simplicity and low cost of the device. The evaporating surface consists of a film of white blotting paper, attached to the end of the U-shaped glass tube. This tube is used both for supply and for measuring evaporation. The blotting sheet is changed every day after reading the device. The Piche is placed inside the weather shelter; also the measurement depends a lot on the ventilation conditions.

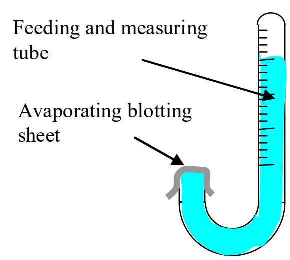


Figure 60. "The Piche" altimeter

4.4. Automatic weather station

4.4.1. The site

The site for installing an automatic weather station must meet certain criteria (Figure 61):

- be representative of the sector by being “normally” exposed to winds;
- be far from any singularity that is too close. We generally accept a minimum distance of four times the height of the obstacle.

Figure 61. Automatic weather station

These rules are not always easy to respect, particularly in the mountains and in the forest. Furthermore, if autonomous devices can be installed a priori at any point, rain gauges require them to be installed near the observer's residence. The choice of a site is therefore a compromise between technical, economic and human imperatives.

4.5. Calculation of average precipitation falling on a watershed

4.5.1. Arithmetic average

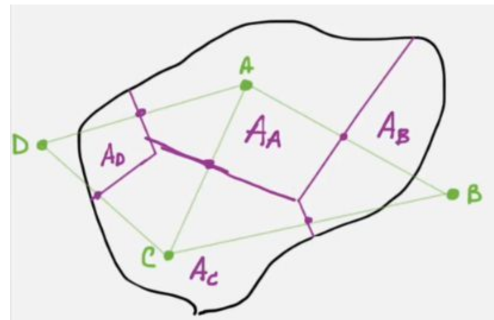
$$\overline{P} = \frac{\sum_{i=1}^{n} P_i}{N} \tag{1}$$

$\overline{P}$: Average annual precipitation, in mm;

P_i: Precipitation falling on rainfall station i, in mm

N: Number of rainfall stations.

4.5.2. Thiessen Method



The so-called Thiessen polygon is the most commonly used, because its application is easy and it generally gives good results. It is particularly suitable when the rain gauge network is not spatially homogeneous (rain gauges distributed irregularly).

This method makes it possible to estimate weighted values by taking into consideration each rainfall station. It assigns to each rain gauge a zone of influence whose area, expressed in%, represents the weighting factor of the local value. The different zones of influence are determined by geometric division of the basin on a topographical map (see). The weighted average precipitation Pmoy for the basin is then calculated by taking the sum of the precipitation Pi from each station, multiplied by their weighting factor (area Ai), all divided by the total surface area A of the basin. The average precipitation over the basin is written as (2) :

$$\overline{P} = \frac{\sum_{i=1}^{n} Pi \times A_i}{A} \tag{2}$$

$\overline{P}$: Average annual precipitation, in mm;

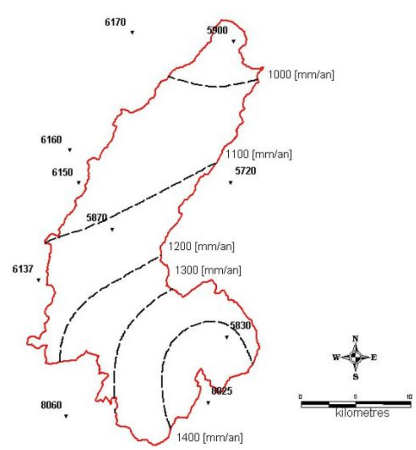
P_i: Precipitation falling on rainfall station i, in mm

Ai: Area of partial surfaces in km^2.

A: Total surface area of the watershed in km^2.

4.5.3. Isohyet Method

The isohyets are lines of the same rainfall (isovalues of annual, monthly, daily rainfall, etc.). Thanks to the rainfall values acquired at the basin stations and other neighboring stations, we can trace the network of isohyets. The outline of the isohyets is not unique like that of the contour lines. It must be drawn with maximum plausibility taking into account the region, the network, the quality of the measurement, etc. Today, there are automatic methods which trace isovalues using sophisticated statistical means (kriging technique).



Tutorials

TD N°1: Basic statistics for rainfall series

Consider the following rainfall series:
1- Calculate the annual rainfall.
2- Calculate the avergae interannual rainfall.
3- Calculate the average monthly rainfall.
4- Complete the table.

Year	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec	Annual total
1990	99.0	0.1	23.8	46.7	7.1	0.1	4.0	6.3	4.0	15.5	0.1	0.1	
1991	28.6	47.6	125.2	1,0	8.0	0.1	0.1	1.3	3.0	19.2	47.6	47.6	
1992	29.5	17.4	59.0	12.5	32.6	27.4	0.8	0.1	0.1	9.0	17.4	17.4	
1993	0.1	35.7	31.0	40.5	13.5	2.3	0.1	3,0	7.5	33.5	35.7	35.7	
1994	15.5	64.0	0.1	28.3	2.6	0.1	0.1	0.1	8.5	34.7	64.0	64.0	
1995	40.6	24.0	47.0	16,0	1.7	0.1	0.1	3.2	29.7	0.1	24.0	24.0	
1996	51.5	91.9	25.6	62.7	11.9	0.1	11,0	0.2	12.0	6.6	91.9	91.9	
1997	67.1	5.0	0.1	83.0	6.3	0.1	0.1	3.8	30.8	32.5	5,0	5.0	
1998	13.3	21.6	14.6	21.6	30.2	0.1	0.1	0.1	12,0	4.7	21.6	21.6	
1999	34.8	45.0	119.6	0.1	0.1	0.1	0.1	0.1	36.7	18.2	45.0	45.0	
2000	1.2	0.1	1.8	3.0	24.0	0.1	0.1	2.0	9.5	22.0	0.1	0.1	
2001	46.5	48.5	0.1	40.6	20.6	0.1	0.1	0.1	14.7	24.6	48.5	48.5	
2002	2.5	2.6	32.5	48.0	42.6	0.1	0.1	7.7	0.1	14,0	2.6	2.6	
2003	42,0	77.8	12.1	60.5	11.2	2.3	0.1	4.3	12.0	26.3	77.8	77.8	
2004	21.1	22.9	16,0	18.5	63.3	5.9	0.1	2.4	8.5	35.9	22.9	22.9	
2005	19.8	59.7	18.7	11.7	0.4	0.1	0.1	0.1	11.9	18.6	59.7	59.7	
2006	101.7	40.6	10.7	18.6	38.2	6.9	2.1	0.1	5.8	0.1	40.6	40.6	
2007	34.4	46.9	41.6	57.1	0.2	0.4	0.1	2.5	6.5	75.9	46.9	46.9	
2008	21.2	19.4	34.7	9.9	30.5	4.7	3.0	0.2	19.5	60.1	19.4	19.4	
2009	52.6	24.7	17,0	65.1	4,0	0.1	0.1	0.1	34.6	1.2	24.7	24.7	
2010	42.6	40.7	75.7	32.5	7.2	0.1	0.1	33.0	12.3	41.9	40.7	40.7	
Average ($\overline{X}$)													
Standard deviation (σ)													
Coefficient of variation (CV)													

Solution TDN°01

The average: $\overline{X} = \frac{1}{n}\sum X_i$ (3)

Standard deviation : $\sigma_x = \sqrt{\frac{1}{n-1}(X_i - \overline{X})^2}$ (4)

Coefficient of variation : $CV = \frac{\sigma_x}{\overline{X}}$ (5)

Year	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec	Annual total
1990	99	0,1	23,8	46,7	7,1	0,1	4	6,3	4	15,5	0,1	0,1	206,8
1991	28,6	47,6	125,2	1	8	0,1	0,1	1,3	3	19,2	47,6	47,6	329,3
1992	29,5	17,4	59	12,5	32,6	27,4	0,8	0,1	0,1	9	17,4	17,4	223,2
1993	0,1	35,7	31	40,5	13,5	2,3	0,1	3	7,5	33,5	35,7	35,7	238,6
1994	15,5	64	0,1	28,3	2,6	0,1	0,1	0,1	8,5	34,7	64	64	282
1995	40,6	24	47	16	1,7	0,1	0,1	3,2	29,7	0,1	24	24	210,5
1996	51,5	91,9	25,6	62,7	11,9	0,1	11	0,2	12	6,6	91,9	91,9	457,3
1997	67,1	5	0,1	83	6,3	0,1	0,1	3,8	30,8	32,5	5	5	238,8
1998	13,3	21,6	14,6	21,6	30,2	0,1	0,1	0,1	12	4,7	21,6	21,6	161,5
1999	34,8	45	119,6	0,1	0,1	0,1	0,1	0,1	36,7	18,2	45	45	344,8
2000	1,2	0,1	1,8	3	24	0,1	0,1	2	9,5	22	0,1	0,1	64
2001	46,5	48,5	0,1	40,6	20,6	0,1	0,1	0,1	14,7	24,6	48,5	48,5	292,9
2002	2,5	2,6	32,5	48	42,6	0,1	0,1	7,7	0,1	14	2,6	2,6	155,4
2003	42	77,8	12,1	60,5	11,2	2,3	0,1	4,3	12	26,3	77,8	77,8	404,2
2004	21,1	22,9	16	18,5	63,3	5,9	0,1	2,4	8,5	35,9	22,9	22,9	240,4
2005	19,8	59,7	18,7	11,7	0,4	0,1	0,1	0,1	11,9	18,6	59,7	59,7	260,5
2006	102	40,6	10,7	18,6	38,2	6,9	2,1	0,1	5,8	0,1	40,6	40,6	306
2007	34,4	46,9	41,6	57,1	0,2	0,4	0,1	2,5	6,5	75,9	46,9	46,9	359,4
2008	21,2	19,4	34,7	9,9	30,5	4,7	3	0,2	19,5	60,1	19,4	19,4	242
2009	52,6	24,7	17	65,1	4	0,1	0,1	0,1	34,6	1,2	24,7	24,7	248,9
2010	42,6	40,7	75,7	32,5	7,2	0,1	0,1	33	12,3	41,9	40,7	40,7	367,5
Average	36,5	35,1	33,7	32,3	17,0	2,4	1,1	3,4	13,3	23,6	35,1	35,1	268,3
Standard deviation (σ)	27,7	25,2	35,5	24,0	17,1	6,1	2,5	7,1	10,9	19,4	25,2	25,2	90,2
Coefficient of variation (CV)	76%	72%	105%	74%	101%	249%	235%	212%	82%	83%	72%	72%	34%

TD N°2: Short rains

In a watershed with a concentration time (Tc) equal to 28.41 min, we require:

1- Calculate the intensity of the rain (I)

2- Calculate short-term precipitation curves (frequency intensities) (IDF curves)

The Montana coefficients "a" and "b" are estimated by the ANRH, in the Oran region, according to the following table. The ANRH evaluated the coefficient a at 0.65

Retour period (T)	2	5	10	20	50	100
$a_{(T)}$	18.9	25.7	30.3	34.6	40.2	44.5

Solution TD N°02

The calculation of the Intensity-Duration-Frequency (IDF) curve (I) for different return periods (T) is estimated by the ANRH method (6):

$$I_{(t,T)} = \frac{a_{(T)}}{t^b} \tag{6}$$

I (t,T): Frequency intensity of rain in mm/h;

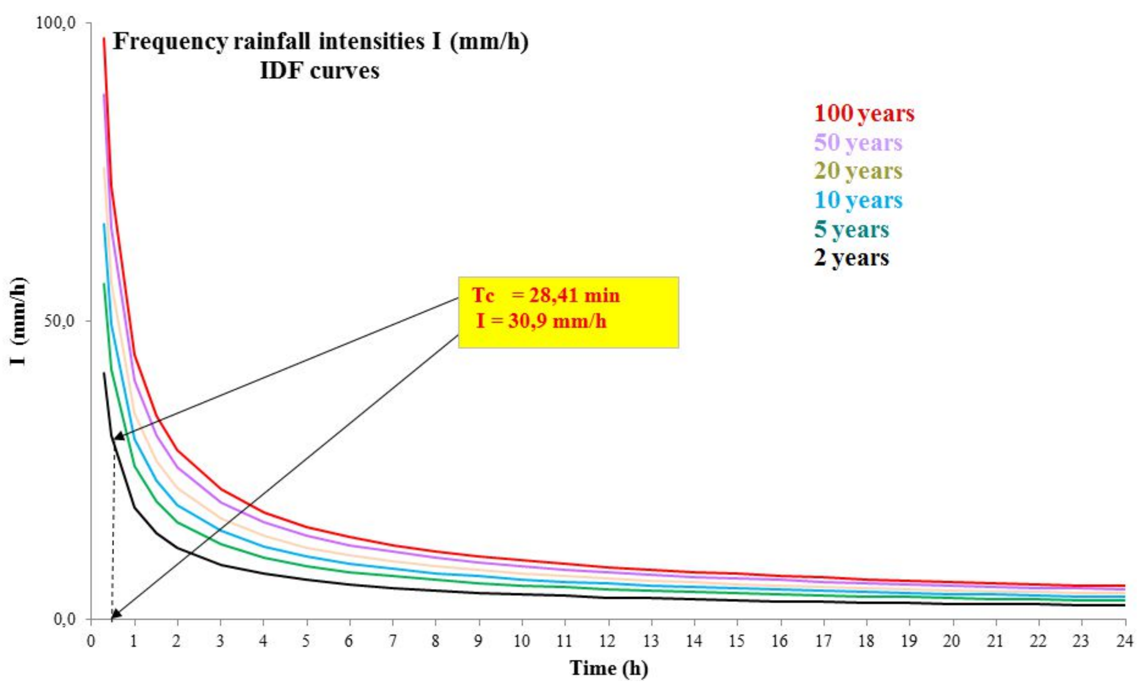
b: Climatic exposure, in our region (Oran), the ANRH assessed it at 0.65;

t: Concentration time in hours.

a and b: Montana coefficients

The calculation of Intensity-Duration-Frequency (IDF) curves (I) for different return periods (T) are presented in the following table and figure respectively.

t (H)	2 years	5 years	10 years	20 years	50 years	100 years
0,3	41,3	56,2	66,3	75,7	87,9	97,3
0,47 (Tc=28.41 min)	**30,9**	**42,0**	**49,5**	**56,5**	**65,7**	**72,7**
1	18,9	25,7	30,3	34,6	40,2	44,5
1,5	14,5	19,7	23,3	26,6	30,9	34,2
2	12,0	16,4	19,3	22,0	25,6	28,4
3	9,3	12,6	14,8	16,9	19,7	21,8
4	7,7	10,4	12,3	14,1	16,3	18,1
5	6,6	9,0	10,6	12,2	14,1	15,6
6	5,9	8,0	9,5	10,8	12,5	13,9
7	5,3	7,3	8,6	9,8	11,3	12,6
8	4,9	6,7	7,8	9,0	10,4	11,5
9	4,5	6,2	7,3	8,3	9,6	10,7
10	4,2	5,8	6,8	7,7	9,0	10,0
11	4,0	5,4	6,4	7,3	8,5	9,4
12	3,8	5,1	6,0	6,9	8,0	8,8
13	3,6	4,9	5,7	6,5	7,6	8,4
14	3,4	4,6	5,5	6,2	7,2	8,0
15	3,3	4,4	5,2	6,0	6,9	7,7
16	3,1	4,2	5,0	5,7	6,6	7,3
17	3,0	4,1	4,8	5,5	6,4	7,1
18	2,9	3,9	4,6	5,3	6,1	6,8
19	2,8	3,8	4,5	5,1	5,9	6,6
20	2,7	3,7	4,3	4,9	5,7	6,3
21	2,6	3,6	4,2	4,8	5,6	6,2
22	2,5	3,4	4,1	4,6	5,4	6,0
23	2,5	3,3	3,9	4,5	5,2	5,8
24	2,4	3,3	3,8	4,4	5,1	5,6



CHAPTER V

CLIMATE STUDY

V- Chapter 5: Climate study

Climate concerns changes in integrated weather conditions across the globe. The word "climate" derives from the Greek κλίμα which means "tilt", referring to the inclination of the earth's axis of rotation. Usually, climate designates the average, calculated over a long period of time (30 years, by convention, for meteorologists), of observations of parameters such as temperature, pressure, rainfall or wind speed, in a geographic location and on a given date (IPCC, 2013). According to the World Meteorological Organization (WMO), climate is defined as the synthesis of meteorological conditions in a given region, characterized by long-term statistics of the variables of the state of the atmosphere (Elmeddahi, 2016).

5.1. Climate classification

Climate classifications are systems that categorize the world's climates. A climate classification can be closely correlated with a biome classification, because climate has a major influence on life in a region. The most widely used is the Köppen climate classification system, first developed in 1884 (USDA, 2008).

There are several ways to classify climates into similar regimes. Climates were originally defined in ancient Greece to describe weather based on the latitude of a location. Modern methods of climate classification can be broadly divided into genetic methods, which focus on the causes of climate, and empirical methods, which focus on the effects of climate. Examples of genetic classification include methods based on the relative frequency of different types of air masses or locations within synoptic weather disturbances. Examples of empirical classifications include climate zones defined by plant hardiness, evapotranspiration, or associations with certain biomes, as in the case of the Köppen climate classification. A common disadvantage of these classification systems is that they produce distinct boundaries between the areas they define, rather than a gradual transition from climatic properties more common in nature (American Meteorological Society, 2008).

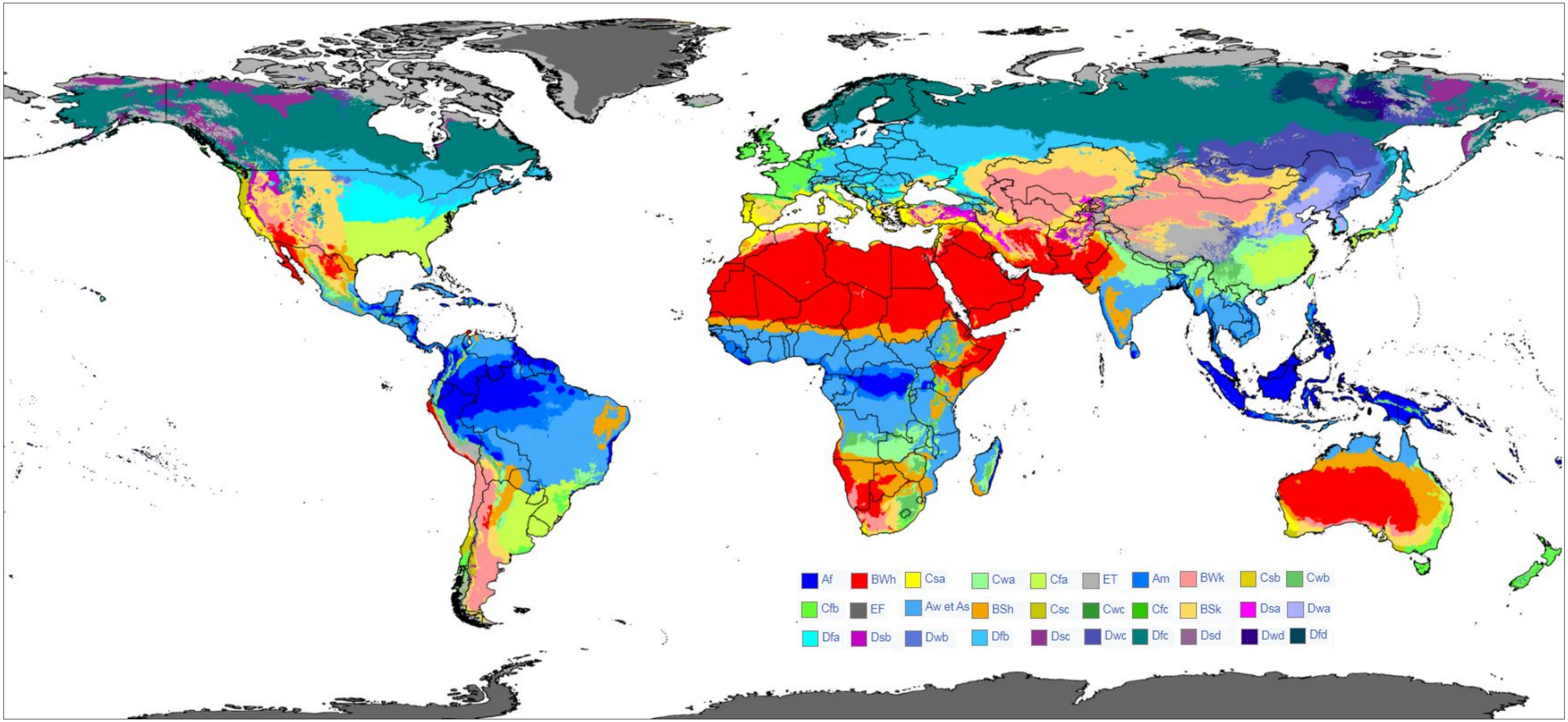
There are many methods of classifying climates. One of the best known is the Köppen classification (Table 3).

Table 3. Characteristics of the main climates in the world (Amyotte, 1995)

Climates	Characteristics
Climate equatorial	Concerns regions neighboring the equator. It is characterized by a only season, heavy precipitation (formation of cumulonimbus clouds), as well as an almost constant high temperature throughout the year, the average of which is 28°C.
Climate tropical wet	Characterizes the regions located up to 15 to 25 degrees north latitude and south. The monthly temperature is above 18°C all year round. We distinguish between a dry season and a wet season. The closer we get from the equator and the longer the wet season becomes.
Climate desert	Characterized by evaporation greater than precipitation. There vegetation is sometimes absent. It extends between 10 and 35 degrees latitude north and south.
Climate subtropical	Located at latitudes between 25 and 45°. These climates undergo the influence of tropical air masses during the summer months, their bringing strong heat. On the other hand, they experience a real cold season, even if it is moderate, under the influence of masses of polar air. Generally two types of climates can be qualified as subtropical: the Mediterranean climate and the humid subtropical climate.
Climate oceanic	Influenced by proximity to oceans, with hot and stormy summers and very cold and rather dry winters. It is marked by an amplitude low thermal (less than 18°C). Precipitation is generally of the order of a meter and above all well distributed. Some authors speak of a hyperoceanic climate where the amplitude average annual temperature is very low (less than 10°C).
Climate continental	Represented by a stronger thermal amplitude (exceeding 23°C). Regions presenting an intermediate thermal amplitude between oceanic climate and continental climate (around 20°C) are called degraded oceanic climate or semi-continental climate. Some authors speak of a hypercontinental climate (amplitude greater than 40 °C) where only the land influences the climate.
Climate Mediterranean	Is characterized by hot and very dry summers, and mild and mild winters. humid with violent precipitation. This climate owes its name to the proximity to the Mediterranean but can be found in other parts of the world (South Africa, Chile, etc.).
Climate subtropical humid	Also called "Chinese shade" is characterized by hot summers, very cold winters, high temperature range, heavy precipitation (greater than 1000-1500 mm/year) particularly in summer. This climate is typical of East Asia.
Climate subarctic	Corresponds to the designation “cold temperate climate without dry season with any hot months (+ 22°C).” This type of climate is only found in the northern hemisphere: central part of all of Canada, most of Russia and northeast China.
Polar climate	Is dry characterized by cold temperatures all year round. Strong wind and persist.

5.1.1. Köppen classification

The Köppen classification (Figure 62) is a classification of climates based on precipitation and temperatures. It was the botanist Wladimir Peter Köppen who invented it in 1900 by combining the world map of vegetation published in 1866 by Hermann Griesbach and the division of the climate into five zones by de Candolle.



Af: equatorial climate **Aw:** savannah climate with dry winter; **As:** savannah climate with dry summer (category sometimes used in analogy with Aw in the rare cases where the dry season occurs in the months when the sun is highest); **Am:** monsoon climate; **BS:** steppe climate (semi-arid); **BW:** desert climate; **Cf:** warm temperate climate without dry season; **Cw:** warm temperate climate with dry winter (Chinese); **Cs:** warm temperate climate with dry summer (Mediterranean); **Df:** cold continental climate without dry season; **Dw:** cold continental climate with dry winter; **Ds:** cold continental climate with dry summer (Mediterranean continental); **ET:** tundra climate**; EF:** ice sheet climate; **EM:** subpolar oceanic climate.

Figure 62. The Köppen-Geiger climate map

Other methods can be combined to characterize a climate using different climatic parameters (rain, temperature, humidity, wind, etc.).

5.1.2. Pluvio-thermal diagram

This diagram was developed by F. Bagnouls and Gaussen (Figure 63). The period is said to be dry when the average total precipitation for the month is less than or equal to twice the average temperature (P<2T) and vice versa. On the basis of this principle, the diagram is established by plotting precipitation and temperatures as a function of the months of the year, while taking into consideration that on the ordinate scale P=2*T.

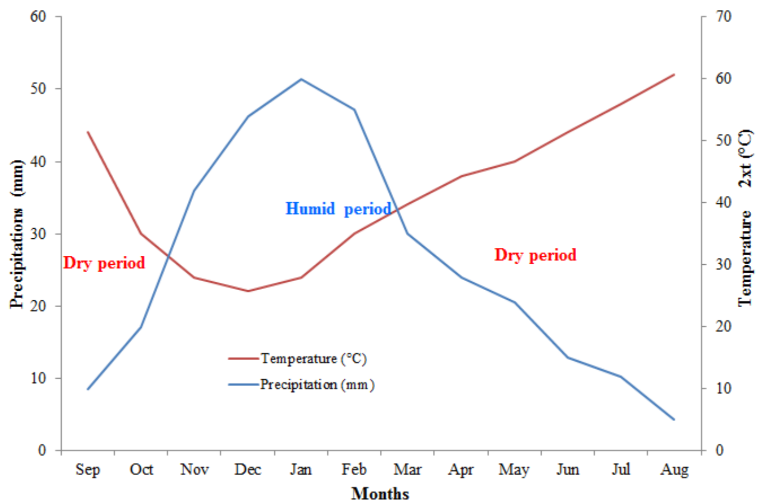


Figure 63. Bagnouls and Gaussen pluviothermal diagram

The establishment of a pluviothermal diagram for our study area, an average pluviothermal diagram is possible (Figure 63), which highlights six wet months (November to May) and six dry months (June to December) with the remark that the difference between temperatures and precipitation and is even greater during dry periods, compared to that of humid periods.

5.1.3. Ombrothermic diagram

Developed by EUVERTE, this method considers the action of heat and humidity with respect to biological activity. In fact, the water needs of plants follow an exponential progression and for a temperature increase of 6°C the water needs double. The ombrothermal diagram (Figure 64) has the months of the year on the abscissa and the temperatures on the ordinate in linear progression, and the precipitation on a logarithmic scale.

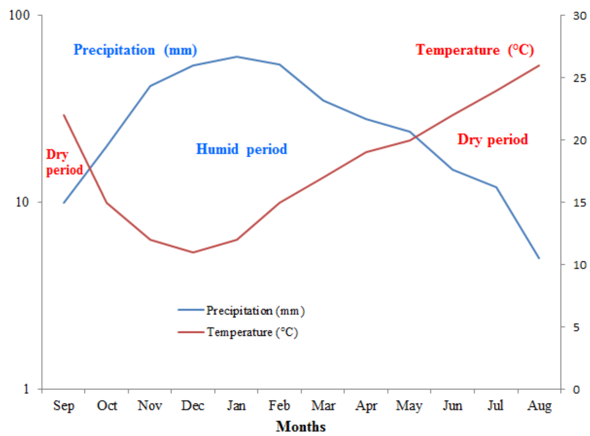


Figure 64. Ombrothermic diagram

5.1.4. De Martonne aridity index

The aridity index is a quantitative indicator of the degree of water scarcity present in a given location (John E. Oliver. 2006). We will calculate this index i using the DE MARTONNE formula (7) (Figure 65).

$$i = \frac{P}{T+10} \tag{7}$$

i: De Martonne aridity index;

P: average annual precipitation (mm);

T: average annual temperature (C°).

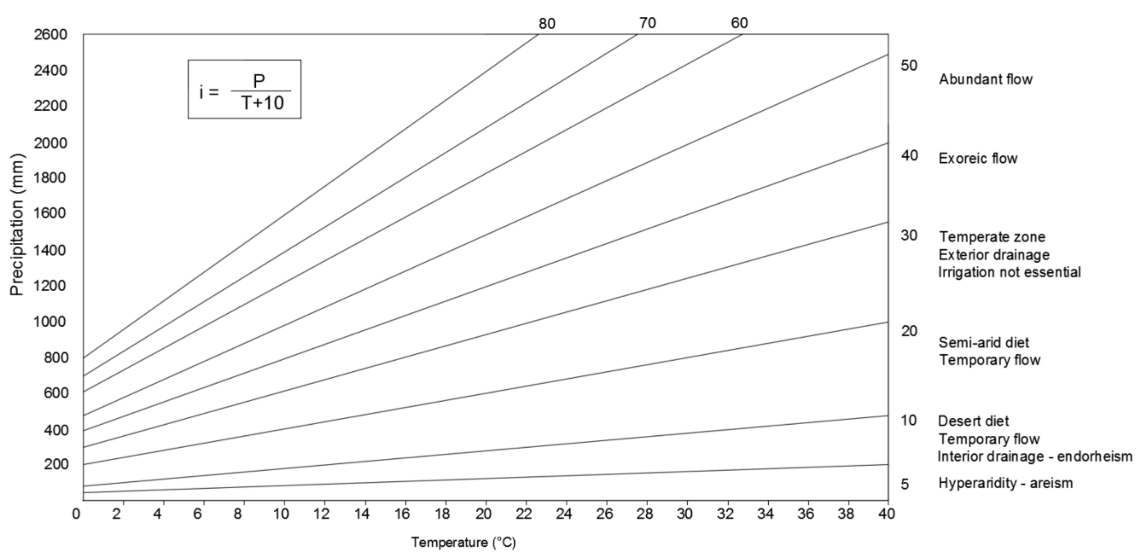


Figure 65. De Martonne aridity index

According to De Martonne's aridity index i, we have 5 types of climate:

- $i<5$: hyperarid climate.
- $5<i<7.5$: desert climate.
- $7.5<i<10$: steppe climate.
- $10<i<20$: semi-arid climate.
- $20<i<30$: temperate climate.

5.1.5. Emberger climagram

The Emberger Climagram (Figure 66) is an abacus comprising on the ordinate the values of the Emberger rainfall quotient (Q_2), and on the abscissa the values of the minimum average temperature of the cold season (T °C), where (8):

$$Q_2 = \frac{P}{\frac{M+m}{2} \times (M-m)} \times 1000 \qquad (8)$$

With :

Q_2: Emberger rainfall quotient.

P: average annual precipitation (mm).

M: average of the maxima of the hottest month, (°K).

m: average of the minimums of the coldest month, (°K).

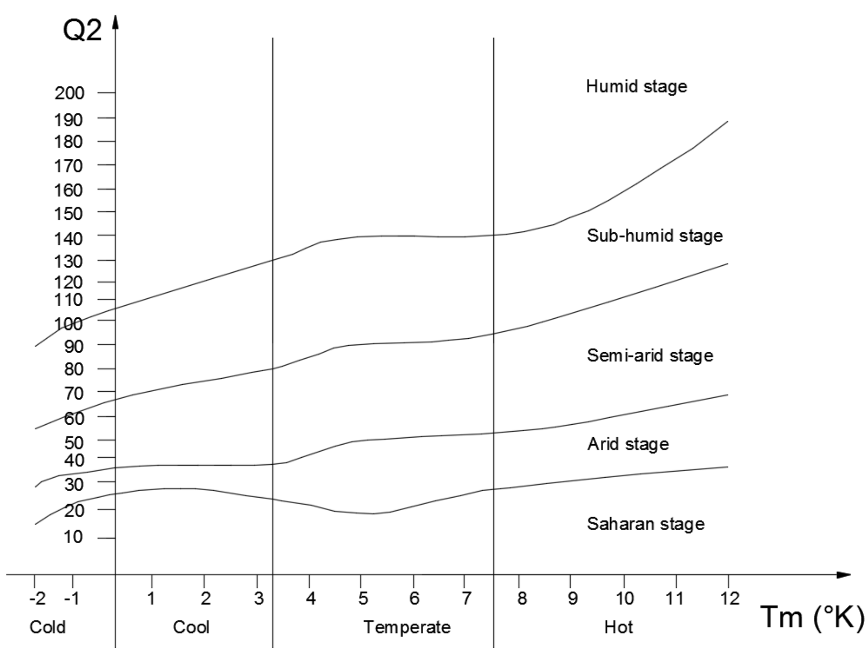


Figure 66. Emberger Climagram

5.1.6. Wundt abacus modified by Coutagne

This chart gives us the flow coefficient (%) and the water deficit (mm) as a function of precipitation (mm) and temperature (°C) (Figure 67).

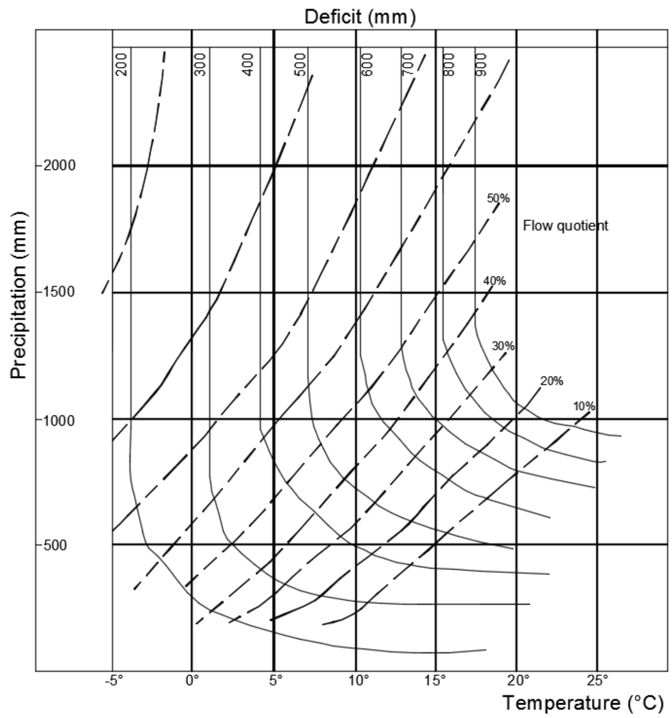


Figure 67. Wundt abacus modified by Coutagne

TD N°3: Climate study

We have a weather station "A" which has precipitation and temperature series (average series over 30 years). With an average minimum annual temperature of 10°C and an average maximum annual temperature of 26°C.

Parameter	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec	Annual
Temperature (°C)	8,4	10,0	14,1	11,4	16,4	23,9	25,5	24,6	23,9	19,1	10,4	9,0	
Rain P (mm)	36,5	35,1	33,7	32,3	17,0	2,4	1,1	3,4	13,3	23,6	35,1	35,1	

1- Calculate the average annual temperature.

2- Calculate the annual rainfall.

3- Convert the average annual temperature from °C to Fahrenheit and Kelvin.

4- Characterize the climate of this station according to the methods of Bagniouls and Gaussen, ombrothermic diagram, De Martonne aridity index, Emberger climagram and Wundt abacus.

Solution TD N°3

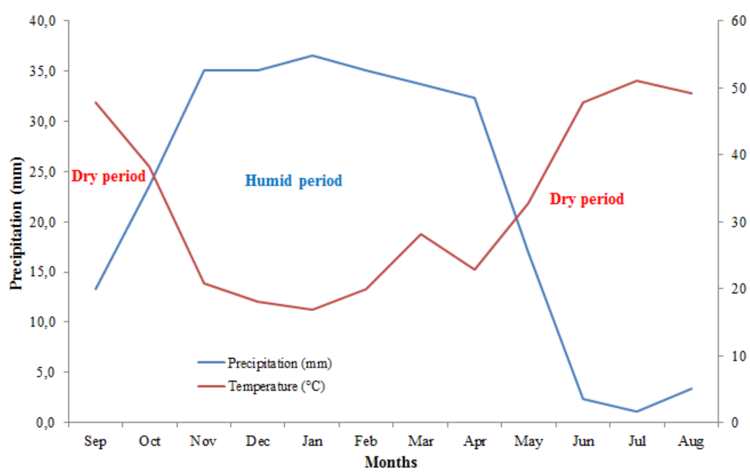
1 and 2 - Calculation of the average annual temperature and annual rainfall.

Parameter	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec	Annual
Temperature (°C)	8,4	10,0	14,1	11,4	16,4	23,9	25,5	24,6	23,9	19,1	10,4	9,0	**16,4**
Rain P (mm)	36,5	35,1	33,7	32,3	17,0	2,4	1,1	3,4	13,3	23,6	35,1	35,1	**268,3**

3- Convert the average annual temperature from °C to Fahrenheit and Kelvin.

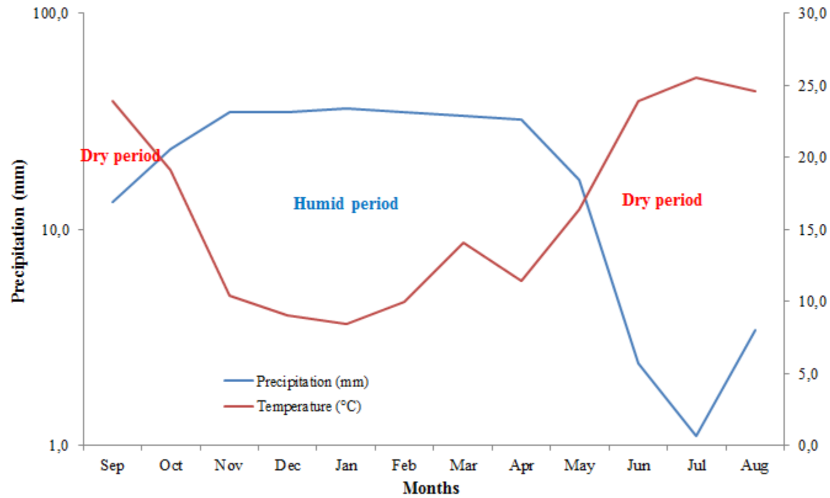
16.4°C = 60.8 F = 289.15 K

4-1. According to the Bagnouls and Gaussen pluviothermal diagram, the wet months of this station go from October to May. While the dry months are between May and September.



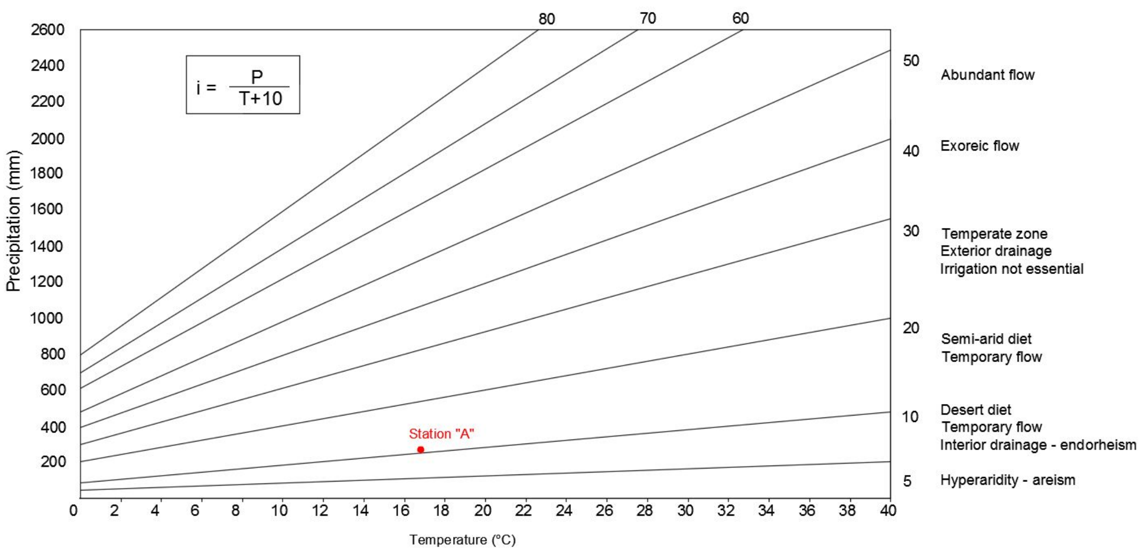
Bagnouls and Gaussen pluviothermal diagram Station "A"

4-2. The ombrothermic diagram gives us the same climate classification as that of Bagnouls and Gaussen.



Ombrothermic diagram Station “A”

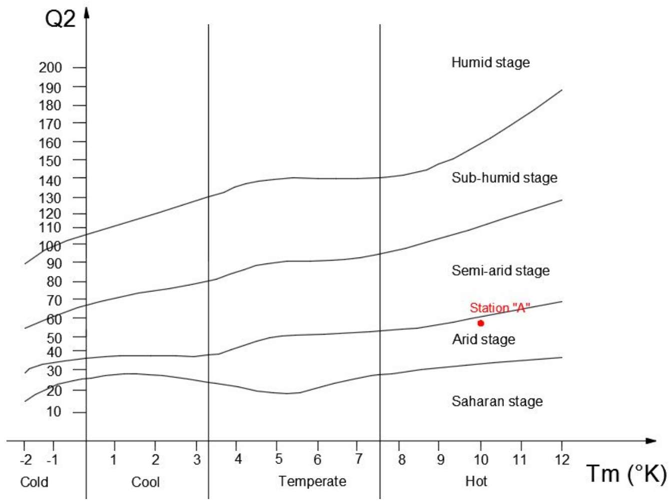
4-3. According to the De Martonne aridity index, the station “A” is located in the semi-arid climatic stage and is characterized by temporary flow.



De Martonne aridity index

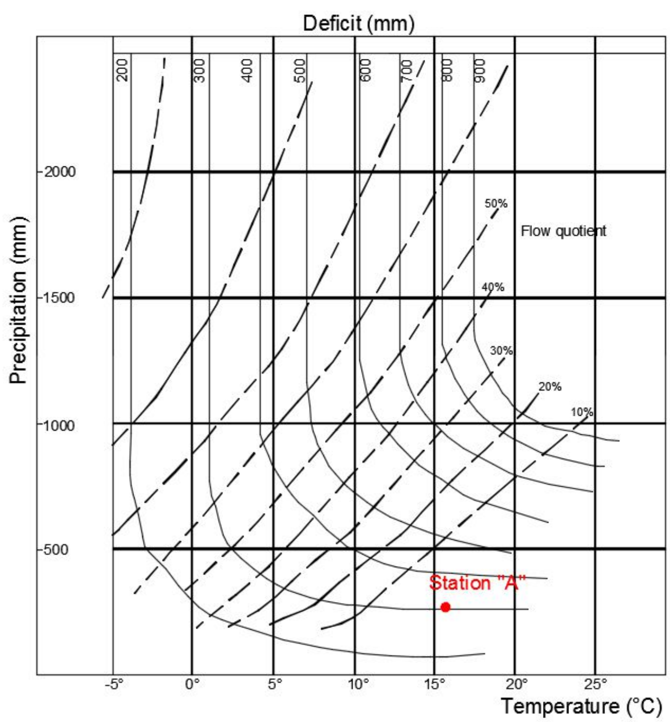
4-4- According to the Emberger climagram, station "A" is located in the arid climatic stage.

$$Q_2 = \frac{P}{\frac{M+m}{2} \times (M-m)} \times 1000 = 57.59$$



Emberger climagram

4-5- According to the Wundt chart, station "A" has a rainfall deficit of more than 300 mm/year with a flow rate of less than 10%.



Wundt abacus modified by Coutagne

CHAPTER VI

MEASUREMENTS AND ESTIMATION OF EVAPOTRANSPIRATION

HYDROLOGICAL BALANCE

VI- Chapter 6: Measurements and estimation of evapotranspiration

6.1. Concept of real and potential evapotranspiration (RET and PET)

We call real evapotranspiration (denoted RET), the quantity of water, generally expressed in millimeters, evaporated or transpired by the soil, plants and free surfaces of a watershed.
Potential evapotranspiration (hereinafter PET) is the quantity of water that would be evaporated or transpired from a watershed if the water available for evapotranspiration was not a limiting factor.

6.2. Direct measurements

Direct measurements of PET or RET are mainly done in agronomy where each particular type of crop is studied. The results of these measurements are difficult to use in hydrology because there is a very significant difference in scale between the surface of the test plot (a few square meters) and that of a watershed (tens of square kilometers). . Furthermore, the plantations used are generally not representative of the vegetation of a watershed.
The Etr measurement can be made on a lysimetric cell (Figure 68). We isolate a block of soil measuring a few square meters in surface area and approximately 2 m thick. This sample of land is drained at its base and the flow rates D leaving through the drains are recorded. On the surface, a collector goes around the plot and collects the runoff water, the flow rate Q of which is also recorded. Water inputs from rain P are measured with a rain gauge. Finally, we evaluate the stock of water R contained in the box, either by measuring the water contents in the soil with a neutron probe, or by mounting the box on a rocker system.

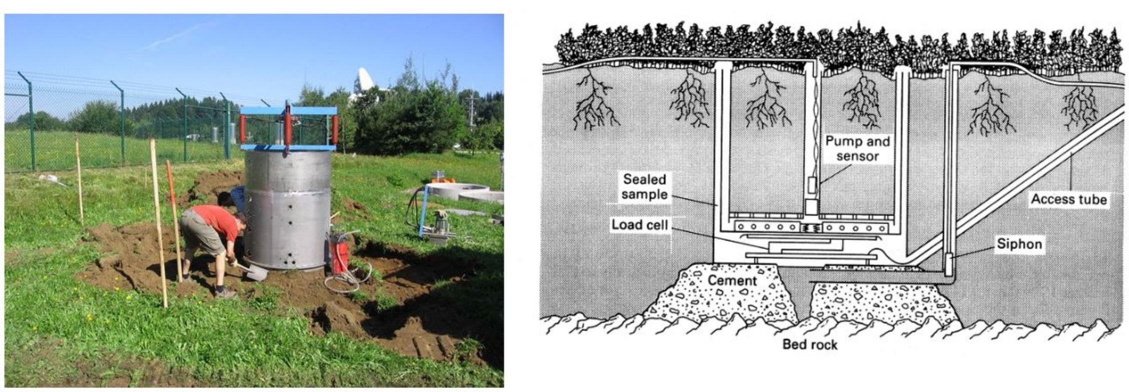


Figure 68. Lysimeter

A very simple assessment makes it possible to evaluate the RET over a time interval Δt since we must have the following relationship (9):

$$P = [Q + D + Etr] + \Delta R \tag{9}$$

$$\text{Input} = [\text{outputs}] + \text{reserve variation}$$

The same system as the lysimetric box, but we then speak of an "evapotranspirometer", allows the Etp to be measured. It is then sufficient to maintain a water level in the drains so that the available water is no longer a limiting factor for RET. We measure PET by writing the same balance as previously but the term D can be positive or negative.

6.3. Estimation of evapotranspiration (indirect measurements)

Several formulas make it possible to evaluate the PET from different climatological measurements. The most complete and complex is certainly the Penman formula based on the concept of energy balance. However, the number of parameters used by this formula (different temperatures, hygrometry, global radiation, albedo, etc.) means that its use is rarely possible given the available measurements.

6.3.1. Thornthwaite Formula

Thorntwaite also proposed a formula based primarily on air temperatures:

$$PET = 16.(10\ \textbf{Error!})^{a}\ .K \tag{10}$$

With:

$$i = (\ \textbf{Error!})^{1.5} \qquad \text{and} \qquad I = \textbf{Error!} \tag{11}$$

$$a = \textbf{Error!}I + 0.5 \tag{12}$$

t: Average monthly temperature of the month considered (°C);

PET: Potential evapotranspiration for the month considered (in mm of water);

K: Monthly adjustment coefficient.

Table 4 presents the monthly adjustment coefficients (K).

Table 4. Monthly adjustment coefficients (K)

Month	**J**	**F**	**M**	**A**	**M**	**J**	**J**	**A**	**S**	**O**	**N**	**D**
K	0,73	0,78	1,02	1,15	1,32	1,33	1,33	1,24	1,05	0,91	0,75	0,70

6.3.2. Other formulas

Other formulas for evaluating PET can be used. Let us cite, for example, the formulas of Bouchet, Blanet and Criddle, and Papdakis, the expressions of which will be found in works on climatology and agronomy.

We have already cited that of Penmann but several other more or less simplified or complicated versions have been proposed.

Based on another approach, we can cite the Bouchet formula which expresses PET as a function of the evaporation recorded at Piche and the Prescott formula based on the evaporation of a free water surface.

Overall, the use of these formulas is uncommon among French hydrologists.

6.4. Evaluation of real evapotranspiration (RET)

6.4.1. Turc Formula

Turc proposed a formula making it possible to directly evaluate the average annual RET of a basin from the annual rainfall amount and the average annual temperature (13):

RET = **Error!** (13)

With $L = 200 + 25\,t + 0{,}05\,t^3$

RET: Real evapotranspiration (in mm/year);

P: Annual rainfall amount (in mm);

t: Annual temperature (in ºC).

This formula is easy to use but unfortunately it only gives the order of magnitude of RET. Indeed, this formula allows the estimation of the "runoff deficit" which only comes close to the real evapotranspiration for relatively large catchment basins, without exchanges at the border and for observation periods long enough for the 'we can neglect variations in underground reserves. If possible, the following method will be preferred.

6.4.2. Simplified balance sheet according to Thornthwaite

This method is based on the notion of Easily Usable Water Reserve (hereinafter RFU). We accept that the soil is capable of storing a certain quantity of water (the RFU); this water can be taken up for evaporation via plants.

The quantity of water stored in the RFU is limited by 0 (the empty RFU) and RFU max (maximum capacity of the RFU which is of the order of 0 to 200 mm depending on the soils and basements considered, with an average of the order of 100 mm).

TD N°4: Water balance

Evaluate the simplified water balance according to Thornthwaite (hydrological or water).

Month	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec
Rain P (mm)	59	50	37	31	23	15	11	5	29	43	55	59
PET (mm)	3	8	33	61	90	94	101	109	67	35	14	5
RFU (mm)	100											
RET (mm)												
A.D. (mm)												
Runoff (mm)												

Solution TD N°4

We admit that the satisfaction of the PET has priority over the flow rate, that is to say that before there is a surface flow, the evaporating power must be satisfied (PET = RET). Then RFU completion also takes priority over flow (Runoff).

We thus establish an assessment on a monthly scale, based on the rain of month P, the PET and the RFU.

a) If P > PET, then:

- RET = PET (14)

There remains a surplus (P-PET) which is allocated first to the RFU, and, if the RFU is complete, to the flow Q (Runoff).

b) If P < PET:

• We evaporate all the rain and we take from the RFU (until emptying it) the water necessary to satisfy the RET, i.e.:

- RET = P + min (RFU, PET-P) (15)
- RFU= 0 or RFU+P-PET (16)

• If RFU = 0, the quantity (AD = PET - RET) represents the Agricultural Deficit, that is to say approximately the quantity of water that should be provided to the plants so that they do not suffer from drought.

Simplified assessment according to Thornthwaite

Month	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec
Rain P (mm)	59	50	37	31	23	15	11	5	29	43	55	59
PET (mm)	3	8	33	61	90	94	101	109	67	35	14	5
RFU (mm)	100	100	100	70	3	0	0	0	0	8	49	100
RET (mm)	3	8	33	61	90	18	11	5	29	35	14	5
A.D. (mm)	0	0	0	0	0	76	90	104	38	0	0	0
Flow (mm)	56	42	4	0	0	0	0	0	0	0	0	3

<- RFU -> empty

CHAPTER VII

CLIMATE CHANGE AND THEIR IMPACTS

7. Chapter 7: Climate change

7.1. Climate change

The climate has been in continuous variation since the appearance of the earth. These climate changes, by directly modifying the composition of the biosphere, the water cycle, plant cover, oceans and ice, have had multiple impacts on all living beings (Parrenin, 2002).

The recent paleoclimatology of our planet (the Quaternary) is marked by a succession of cold, glacial periods and warm, interglacial periods (Figure 69). Variations in solar emission (minimum solar spots correspond to episodes of cold in the Earth's climate) over the last thousand years explain historical climate variabilities: a warm period in the Middle Ages (476 to 1453), the Medieval optimum during the 11th to 14th centuries, then a cold period from the end of the 15th century to the middle of the 19th century called the Little Ice Age (1400 to 1850). This period is characterized, in Europe and North America, by long, harsh winters and short, humid summers, an extension of Arctic ice and global cooling of -1°C to -2°C. This period ended around 1860 around the same time as the industrial revolution. Since then, we have entered a warm period (Henson, 2008).

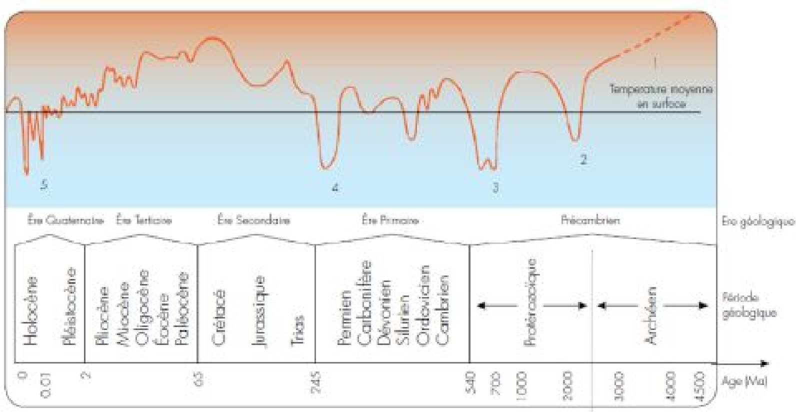


Figure 69. Variation in surface temperature throughout Earth’s history (Ibrahim, 2012)

Figure 69 shows that temperatures during the medieval climatic optimum were higher than today, while it was much cooler during the Little Ice Age (Ibrahim, 2012).

7.2. Distinction between climate variability and climate change

Climate variability is the variation in the mean state and other statistical variables (deviation from the mean) of the climate at all temporal and spatial scales. According to the IPCC (2013), variability results from natural internal processes within the climate system (internal variability) or from variations in external or anthropogenic forcing (external variability). Climate change is the statistically significant variation in the average state of the climate, persisting for an extended period of time. Climate change may be due to natural internal processes or external forcing, or to the persistence of anthropogenic variations in atmospheric composition or land use.

The United Nations Framework Convention on Climate Change (UNFCCC) distinguishes between "climate change" which can be attributed to human activities altering the composition of the atmosphere, and "climate variability" due to natural causes. Natural climate variability has purely geophysical and astronomical causes, as evidenced by the past history of climate: variations in solar activity, natural evolution of atmospheric composition, volcanic eruptions, and meteorite impacts (Lulu Zhang, 2015).

Climate change can manifest itself through:

- The drought which has affected the two tropical bands of our planet, especially the Sahelian regions of West Africa, since the 1970s (Sircoulon, 1987).
- Recent El Niño phenomena (Vandiepenbeeck, 1998).
- The findings of the World Meteorological Organization (WMO) on global warming estimated at 0.5°C since the middle of the last century (Cantat, 2004).

The ten-year average temperature (2001 – 2010) represents the highest average since meteorological instrument records began (WMO, 2013).

7.3. Climate change on a global scale

Since 1850, when instrumental records of global surface temperature began, the climate has warmed considerably, particularly over the last 30 years (Figure 70). Between 1905 and 2006, average warming is estimated at 0.74°C (0.56 ± 0.92°C) (IPCC, 2013), but this warming presents disparities at the local scale where continental regions experience faster warming than that of the oceans, and this warming is more significant at high latitudes in the Northern Hemisphere. The period between 1995 and 2006 appears to be the warmest since 1850 (Henson, 2008).

Figure 70. Earth temperature trend (1976-2000) (Henson, 2008)

7.4. Change in extreme events observed in the 20th century

The report of the IPCC (2007) highlighted that a warmer climate very likely leads to an increase in the frequency and/or intensity of certain types of extremes events. Since then, a considerable number of authors have studied the problem of the evolution of extremes.

For the entire northern hemisphere, a particularly pronounced increase in minimum temperature characterizes the period 1979-2003. Precipitation indices indicate a trend towards wetter conditions (Goubanova, 2007). For Northern Europe, Groisman et al. (2005) show an increase in the frequency of intense rains of 60% for summer (the season with the heaviest rains) and of 40% for the whole year. Alpert et al. (2002) examine data from 256 stations covering the last fifty years across the entire Mediterranean region, and detect on the one hand a decreasing trend in total precipitation and paradoxically an increasing trend in the intensity of extreme precipitation.

7.5. Impacts of climate change

7.5.1. The rise in surface temperature on Earth

The increase in surface temperature on Earth has been 0.8 ± 0.2°C since 1870. It remains notably different for the two hemispheres: stronger in the North and stronger at high latitudes (IPCC, 2014).

The conclusions of the IPCC AR5 are clear: human activities, in particular those emitting carbon dioxide (CO2), are mainly responsible for the warming of the climate system currently observed (95% chance). According to (Stocker et al., 2013), the global average of surface temperatures (land and oceans) increased by 0.85°C between 1880 and 2012.

Each of the last three decades has been warmer than the last. Measurements recorded the ten hottest years since 1850 between 1998 and 2010. Climate change will significantly increase the risks of extreme weather events (heat waves, extreme precipitation, flooding of coastal areas, etc.). In the RCP8.5 scenario, heat waves currently occurring around the world could double or triple in frequency (IPCC, 2013). This warming experienced a first phase between 1900 and 1940, followed by a stabilization (or even decrease) between 1940 and 1975, before resuming since 1975 (Figure 71).

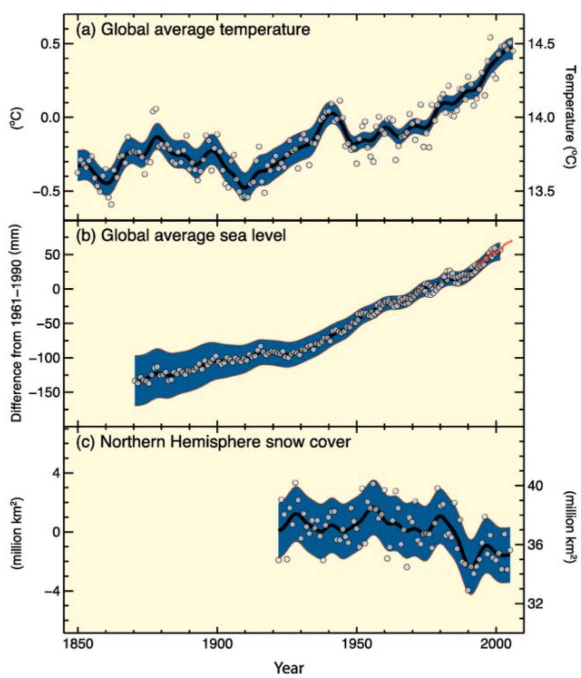


Figure 71. ***a)*** Global average surface temperature; (b) Global mean sea level; c) Snow cover in the Northern Hemisphere in March and April. All changes are relative to the corresponding averages for the period 1961-1990. The smoothed curves represent values averaged over a decade. Circles indicate annual values. Shaded areas are the uncertainty ranges estimated from an exhaustive analysis of known uncertainties (a and b) and time series (c) (IPCC, 2007)

7.5.2. Ocean temperature

Measured since the 1950s by commercial boats or oceanographic vessels (up to around 700 m depth) and more recently by the Argo profit buoy system, shows an overall average increase over the past few decades (Our planet.info, 2014).

The seas and oceans, like the land surface, are experiencing an increase in temperature, both at the surface and at depth. This warming is more marked in the upper layers (between 0 and 700 m depth) and near the surface (Figure 72). Thus, the first 75 meters of depth have warmed on average by 0.11°C per decade over the period 1971-2010 (IPCC, 2013).

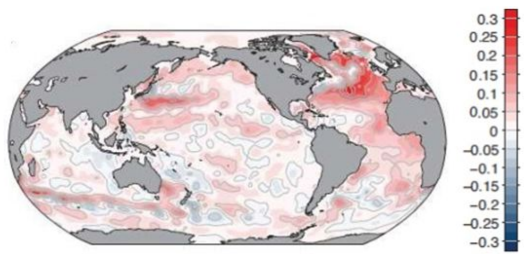


Figure 72. Trend of change in ocean temperature between 0 and 700 m (1971-2010) (in °C per decade) (Rhein et al., 2013)

It is almost certain that the surface ocean (up to 700 m deep) warmed between 1971 and 2010 (probably warmed between the years 1870 and 1971) (IPCC, 2013). According to projections, the greatest warming will occur in the surface ocean of the tropical and subtropical regions of the Northern Hemisphere. Deeper, warming will be most pronounced in the Southern Ocean. The most likely estimates of ocean warming over the first hundred meters are about 0.6°C (RCP2.6) to 2.0°C (RCP8.5) and about 0.3°C (RCP2.6) to 0.6 °C (RCP8.5) at a depth of approximately 1,000 m towards the end of the 21st century (IPCC, 2013).

7.5.3. Precipitations

From 1900 to 2005, precipitation increased significantly in eastern parts of North and South America, northern Europe, and northern and central Asia, but decreased in the Sahel, in the Mediterranean basin, in southern Africa and part of southern Asia. Clearly, humid regions tend to become more and more humid and dry regions increasingly drier (Cantet, 2009). On average, the IPCC group predicts, by the end of the century, more abundant precipitation and more frequent episodes of extreme precipitation in the continental masses of high and mid latitudes, and in humid tropical regions. Conversely, arid and semi-arid regions of mid-latitudes and subtropical regions will experience a decrease in precipitation and a worsening and increase in droughts (Cantet, 2009).

7.5.4. Seas and oceans

The seas and oceans, which cover nearly 70% of the earth's surface, are key players in the climate system due to the permanent exchanges of energy between them and the atmosphere. They are, among other things via ocean plankton, the main carbon sinks, ahead of forests, peat bogs and meadows. Since 1961, sea level has risen at an average rate of 1.8 mm ± 0.5 mm per year, and since 1993 at 3.1 mm ± 0.7 mm per year, due to thermal expansion and melting glaciers (Cantet, 2009).

7.5.5. Snow and ice

Satellite data show that, since 1978, the extent of Arctic sea ice has declined by an average of 2.7% ± 0.6% per decade, with a more marked decline in summer (7.4% ± 2.4% per decade). Mountain glaciers and snow cover have decreased on average in both hemispheres. These decreases are consistent with warming (Cantet, 2009).

7.5.6. Biological indicators

Biological indicators, such as movements of terrestrial or marine animal populations and changes in the dates of seasonal agricultural activities, also show the occurrence of global warming (IPCC, 2014).

7.5.7. The reduction in the surface of Arctic ocean ice

The sea ice, whose melting does not contribute to the rise in ocean levels, is another strong indicator of the acceleration of climate change: from 8.5 million km2 stable in the period 1950-1975, the surface area of sea ice experienced a very rapid decrease to 5.5 million km^2 in 2010 (Our planet.info, 2014).

7.5.8. Average ocean level

The average ocean level is another indicator that integrates the effects of several components of the climate system (ocean, continental ice, continental waters). Before 1992, sea level was measured by tide gauges along the continental coasts and some islands: the level of the oceans, as an annual average across the planet, rose at a rate of 0.7 mm/year between 1870 and 1930 and around 1.7 mm/year after 1930. Since 1992, measurements have been carried out by satellites: the rise in global average sea level is of the order of 3.4 mm/year (http://www.pensee-unique.fr/oceans.html le 22/04/2014).

7.6. Climate change on a global scale

Floods are among the natural disasters that cause loss and damage to property worldwide. They constitute the most distributed risk on the planet (White, 1999).

Climate change will increase both the frequency and intensity of extreme weather events, namely tropical cyclones, heat waves, flash floods, and droughts. While climate change has undoubtedly played a role in explaining this increase, we must not neglect the importance of two other factors: on the one hand, these disasters are better documented today than they were previously in the past, which increases their visibility; and on the other hand, the number of people exposed to these disasters has steadily increased due to population growth. Figure 73 shows vulnerability to climate disasters around the world.

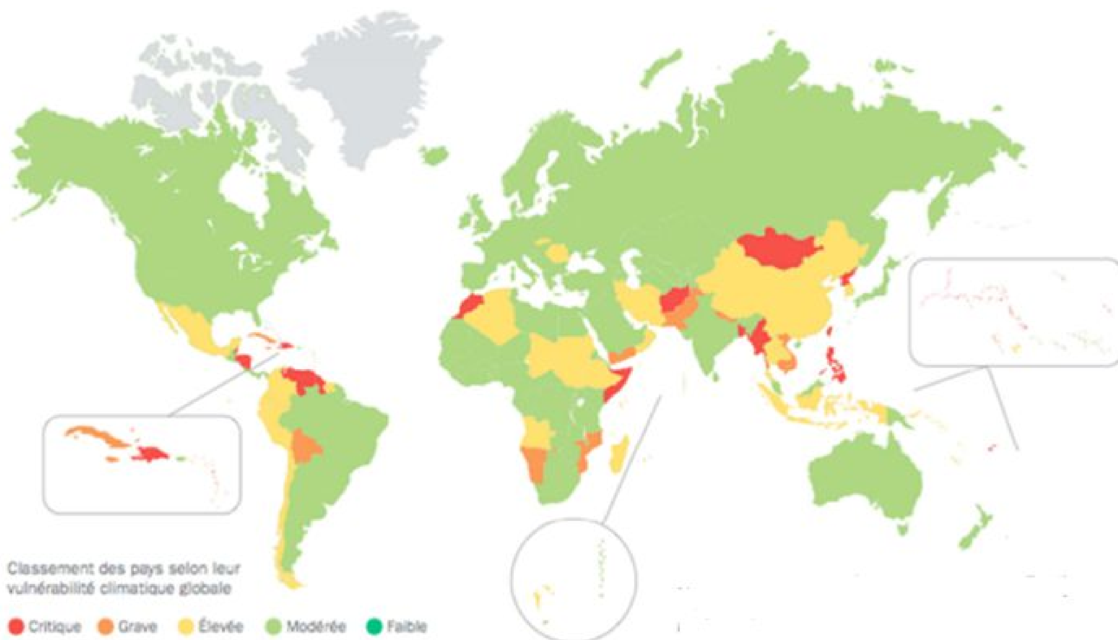
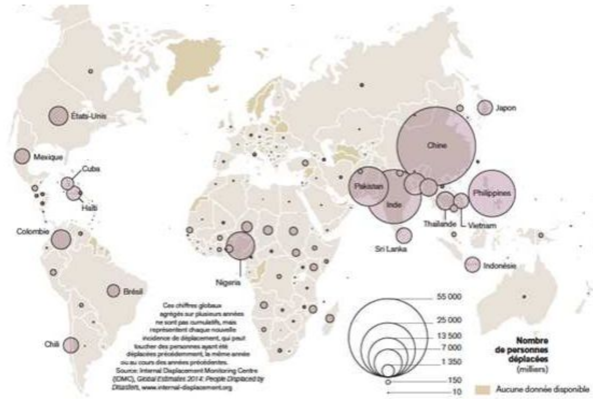


Figure 73. Vulnerability to climate disasters around the world

7.7. Impacts already detected

The first and main consequence of these disasters, in terms of security, obviously concerns the massive displacements of populations that they cause. According to data from the Internal Displacement Monitoring Center (IDMC), between 2008 and 2014, 26.4 million people were displaced each year by natural disasters, the equivalent of approximately one person per second (Figure 74).



Source : Ionesco et al., 2015

Figure 74. Global displacement caused by disasters (2008-2013)(IDMC)

Several climate simulations indicate that drought has appeared in the majority of countries in the Mediterranean basin since the beginning of the 1980s (IPCC, 2007; 2013) and in particular Algeria with the frequent appearance of extreme weather phenomena and natural disasters (Khoualdia, 2015).

According to (Zhang et al., 2007), an amplification of rainfall contrasts has been observed over the last 50 years, between the mid-latitudes where precipitation increases, and the subtropical regions of the northern hemisphere (case of Algeria) or on the contrary, precipitation decreases. In addition, an increase in extreme rainfall was noted in the northern hemisphere during this same period and demonstrates its anthropogenic origin (Benzater et al., 2021; 2019; Min et al., 2011).

7.8. The impacts of climate change in Algeria

Like many African countries, the consequences of extreme weather have not spared Algeria. Known for its arid and semi-arid climate, the region is extremely prone to climate change. Over the past 50 years, an increase in events due to extreme weather has been observed.

Among the phenomena recorded in climatological studies of the National Meteorology which testify to this change, we include an increase in the frequency of torrential rains, especially on the high plateaus (e.g. Ghardaïa and Béchar in 2009–2010), which led to floods for the first time. In 2020, daily rainfall may exceed the normal annual average in the south of the country (IPCC, 2013). Snowfall has decreased by 40% in several regions of Algeria, such as Tlemcen, Ouarsenis and Djurdjura.

Terrestrial and marine flora and fauna have been greatly affected by this increase: the change in environmental conditions is favorable and/or unfavorable to certain ecological factors compared to others, which leads to a change in environments and the species of the flora and fauna which constitute them. Change is expressed by the rarefaction and/or disappearance of species to the detriment of others, desertification and pollution, which leads to environmental degradation.

In Algeria and over the last three decades, the climate has had a negative influence on water resources. Its impact was felt on the waterways of the Macta watershed. Studies of annual precipitation during the period 1930-2002 show a clear decrease. (Meddi et al. (2009) demonstrated a reduction in the amount of water flowed by 28 to 36% compared to the two periods (1976-2002) and (1949-1976). Laborde (1995) showed four successive rainfall phases by analyzing data from 120 rainfall stations in northern Algeria:

- A long wet phase during which rainfall was above average, by 6% (1922-1938).
- A short dry phase from 1939 to 1946 in the west and center of the country, with a deficit of around 11%.
- Then a rainy phase which lasted until 1972.
- Since the end of 1973, a long dry phase begins and persists...

The rainfall map of the Algerian North, established by (Laborde, 1995) in collaboration with the National Agency for Hydraulic Resources (ANRH), detects a climate change in the Algerian North-West and particularly in the Macta, Tafna and Chellif basins, whose rainfall has significantly decreased in this region.

The evolution towards aridification of the climate in northern Algeria in general and in the Macta watershed specifically, shown by results (Laborde, 1995), has led to a flow deficit leading to a reduction surface flows and consequent drawdown of groundwater.

Paradoxically, and despite this decreasing trend in annual rainfall, northern Algeria is experiencing catastrophic floods due to extreme rains:

- December 21, 1930: start of the winter season, major 12-day flood followed by a terrible flood, was reported in Chlef, at the Pontéba Dam. In 24 hours, the volume of Chellif reached 200 hm^3.
- From March 28 to 31, 1974: exceptional rainfall in the wilayas of Algiers and Tizi-Ouzou (688 mm in 4 days) and 381 mm in one day at the Sakamody pass (Médéa).
- October 20, 1993: 22 deaths and 14 injured in Oued Rhiou.

- October 1994: 60 deaths and dozens of missing people, in several regions of the country, during a ten-day flood.
- October 24, 2000: The catastrophic floods of Sidi Bel Abbes (Wadi Mekerra), Tissemsilt Theniet El Had (Wadi Mesloub), Chlef and Ain Defla caused the disappearance of two people swept away by the Wadi Mekerra and important equipment in Theniet El Had and more than 24 deaths in the west of the country.
- On October 15, 2008, regional floods, affecting the Wilayas of Naama, Sidi Bel Abbes, Saida and El Bayadh and Ghardia, left 35 dead.

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