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Mechanisms driving precipitation changes in North Africa, under past, present and future climate

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Declaration

I hereby declare that this thesis is my original work and it has been written by me in its entirety. I have duly acknowledged all the sources of information which have been used in the thesis.

This thesis has also not been submitted for any degree in any university previously.

Arab Djebbar
September 2021

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CHAPTER 1

North Africa

1.1 Geography

Egypt, Libya, Tunisia, Algeria and Morocco are the five countries adjacent to the Mediterranean Sea in North Africa (Figure 1.1). The Sahara Desert, the most prominent characteristic of the North African landscape, runs along the region's southern part. Except in Egypt, the Sahara acts as a regional divide between North and Sub-Saharan Africa. The Sahara, the Atlas mountains in the west and the Nile river and Delta in the East are the three major geological features of North Africa. The Atlas mountain range runs through Algeria, Tunisia and Morocco. Mountains evolved over millions of years when the African continent collided with the Iberian Peninsula in a convergent plate boundary, giving the mountains their current shape [1]. The Sahara's main land types include ergs, large seas of sand that often develop into vast dunes; the Hammada, a flat rocky plateau without soil or sand; and the reg, a level plain of gravel or small stones. The Sahara Desert encompasses the southernmost parts of Morocco, Algeria and Tunisia, as well as the majority of Libya. Outside the desert, Libya has only two regions: Tripolitania in the northwest and Cyrenaica in the northeast [2].

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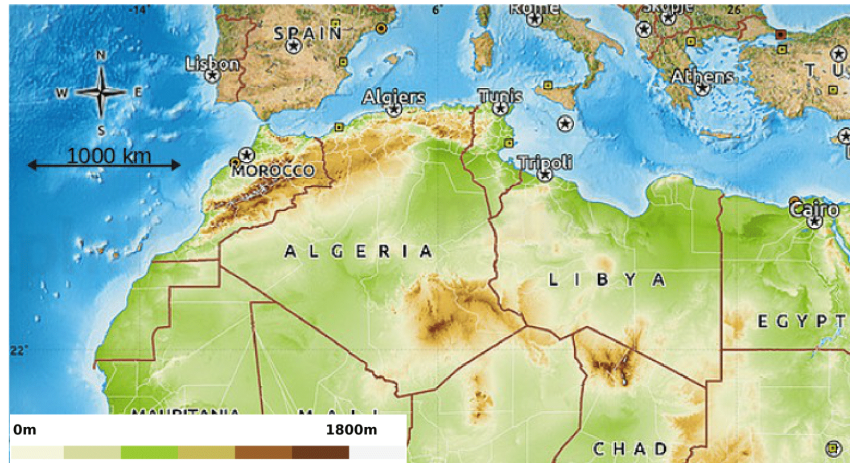


Figure 1.1: Physical map of North Africa. Differences in land elevations relative to the sea level are represented by color. Green color represents lower elevations, orange or brown indicate higher elevations. source: www.maphill.com/africa/maps/.

The North African plateau is largely made up of an ancient rock plateau that is now occupied by the Sahara desert. The internal region of the plateau is divided by a few isolated mountain massifs: Ahaggar, Tibesti, which include the High Atlas and Toubkal Mount (4,167 m) [3]. This vast region is bordered on the west by a plane that circles the Senegal River's estuary, on the north by Libyan and Tunisian planes and on the northwest by the Atlas Mountains. Several large rivers flow near the Sahara's limit, with a tropical climate. They have generally a single high water season that corresponds to the tropical rainy season. After entering Sudan, where it receives waters of the Blue Nile and the Atbara river, the Nile reaches Egyptian territory through a narrow valley and then flows through a vast plain to the delta and finally into the Mediterranean Sea [4].

1.2 North Africa Climate

1.2.1 Past and present climate

North Africa has been classified into two main climate types, with the Mediterranean temperate climate (Csa and Csb) dominating the coastal parts (from Morocco, Algeria, to part of Libya) and the semiarid and arid climate (Bwh and Bwk) dominating the South (Köppen-Geiger climate classification [5]) (Figure 1.2).

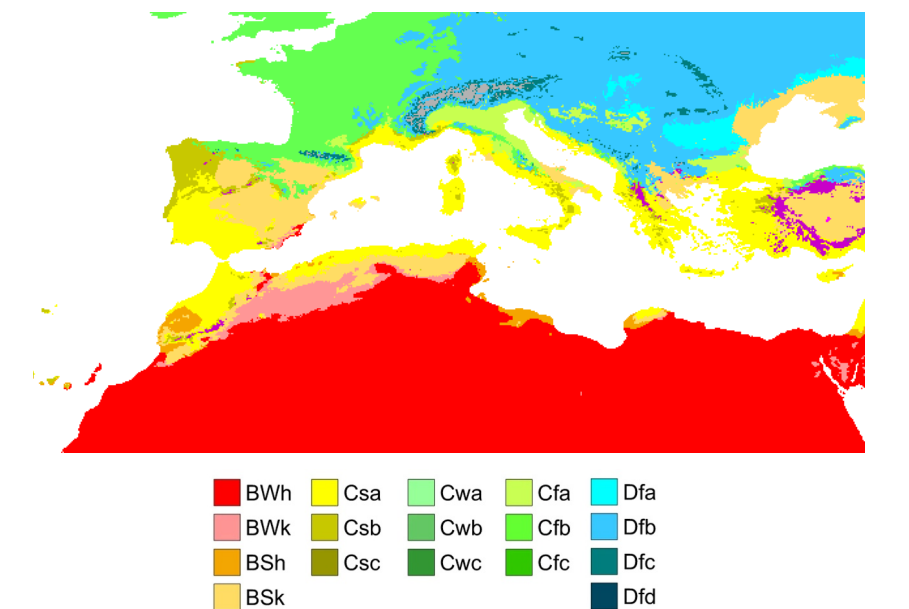


Figure 1.2: Climate types over North Africa based on the Köppen–Geiger climate classification system at 1–km resolution [6]. The significance of the symbols are given in the following table:

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Symbol	Significance		
BWh	Arid, desert,hot	Cwc	Temperate, dry winter, cold summer
BWk	Arid, desert,cold	Cfa	Temperate, no dry season, hot summer
BSh	Arid, steppe,hot	Cfb	Temperate, no dry season, warm summer
BSk	Arid, steppe, cold	Cfc	Temperate, no dry season, cold summer
Csa	Temperate, dry summer, hot summer	Dfa	Cold, no dry season, hot summer
Csb	Temperate, dry summer, warm summer	Dfb	Cold, no dry season, warm summer
Csc	Temperate, dry summer, cold summer	Dfc	Cold, no dry season, cold summer
Cwa	Temperate, dry winter, hot summer	Dfd	Cold, no dry season, very cold winter
Cwb	Temperate, dry winter, warm summer		

Table 1.1: Köppen–Geiger climate classification classes [6].

North Africa's precipitation is marked by rainy winters and dry summers. The region receives moderate to low annual rainfall (less than 100 mm in the south to around 600–800 mm in the north and high mountains) [5] (Figure 1.3). In Egypt, for example, precipitation is only significant in the northern Mediterranean coast (about 180 mm/year) and it is extremely low in the rest of the country's desert territory [7]. The mean temperature in the winter, ranges between 10°C and 20°C and in the summer, it is between 25°C and 35°C [5] (Figure 1.4). The temporal variability of temperatures is more uniform than that of precipitation, for example, [8] divided the Mediterranean Sea coast from Morocco to Tunisia into five homogeneous regions of similar precipitation variability in January and in three for temperature [8]. It's over the winter periods that both precipitation and temperature vary the most between North Africa countries.

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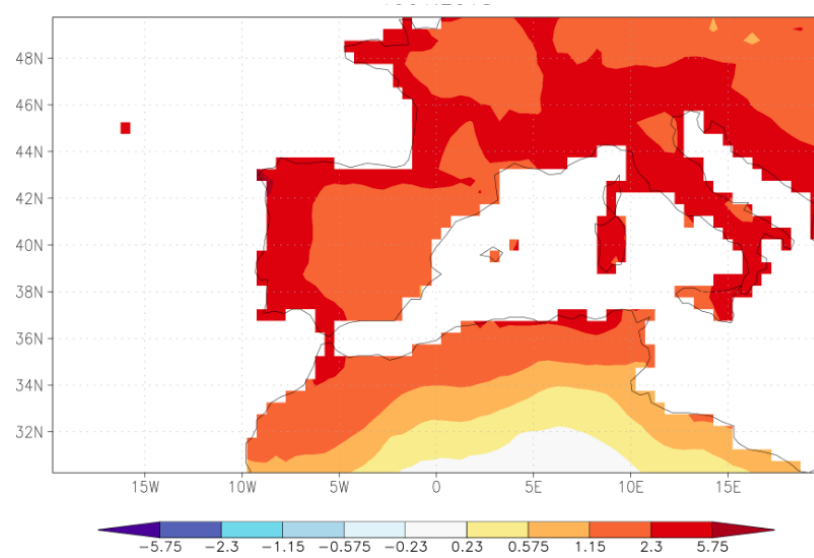


Figure 1.3: November-April observed (CRU-TS4.04) [9] precipitation means (mm/day) over 1979-2019.

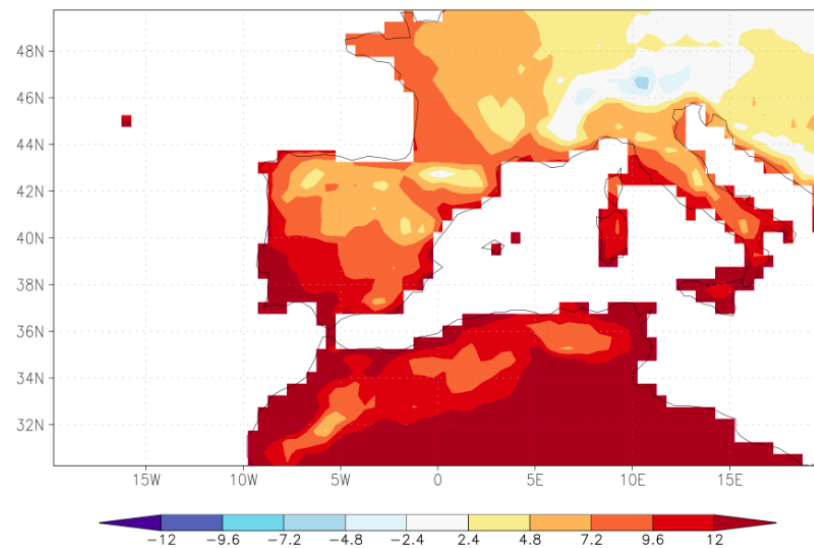


Figure 1.4: November-April CRU-TS4.04 [9] temperature mean (°C) over 1979-2019.

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Figure 1.5 represents Sea Level Pressure (SLP) means over winter season 1979-2019. They show a high pressure centred over the Azores islands (west of Portugal, known as the sub-tropical or Azores high) and low pressure centred over Iceland (the sub-polar or Icelandic low). The difference between pressures at these two point measures the North Atlantic Oscillation (NAO), a teleconnection pattern that is well-known and one of the first ones to be discovered [10]. The SLP mean over winter (November-April) shows an extension of the Azores high pressure towards the western part of North Africa (Morocco and West of Algeria) which is associated with a difference in SLP between the West and the East of North Africa (Figure 1.5).

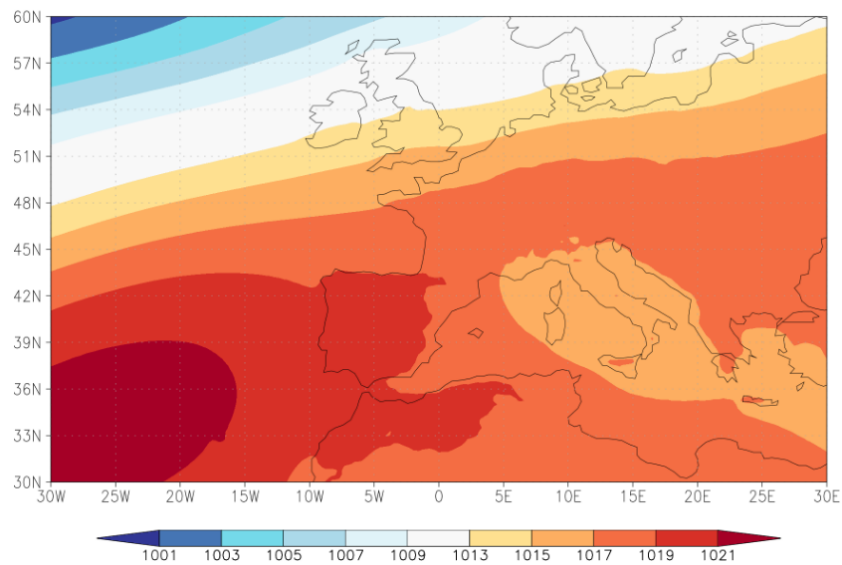


Figure 1.5: November-April, ERA5 [11] Sea Level Pressure means (mb) over 1979-2019.

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The winds generally flow from North to North-West over North Africa countries in winter and from East to South-East in summer season. The winds are relatively calm during the summer period over the region. The sirocco (Shargui), a dry, warm, southerly wind that blows all year (Figure 1.6), is a distinctive climatic characteristic of North Africa. Siroccos are at the maximum of their strength in the spring, when winds can reach gale force (defined as sustained wind speeds of 17 to 24 m/s). The winds originated over the Sahara Desert, sweep North over North Africa, across the Mediterranean Sea, towards Southern Europe [8]. Siroccos are caused by the west-to-east progression of extratropical cyclones, where a low pressure center drives southerly winds in a warm sector and thereby generates the sirocco winds across the Mediterranean [12].

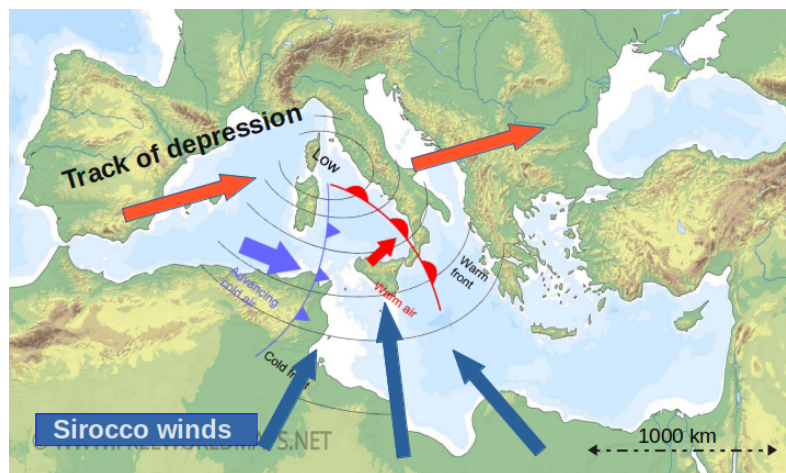


Figure 1.6: Sirocco winds: dry, hot air from the Desert, flowing northward into the Mediterranean basin.

source:<https://www.freeworldmaps.net/europe/mediterranean/mediterranean-physical-map.jpg>.

Scientists rely on instrumental records that are only a few hundred years long at best and longer geologic records that span thousands to

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millions years to understand the climate dynamics. These records also allow to study the potential range and impacts of future climate change, as well as how these changes may affect ecosystems and society. The reconstruction of various Quaternary climatic phases has been possible thanks to the study of climate history documented in lake sediments in various Saharan zones. The sediments of Lake Megafezzan and the Tunisian chotts are records evidence of major climate changes during the global glacial cycles, alternated with brief interglacial periods [13]. About 10,000 years ago, during the Holocene Climate Optimum, the Sahara was covered in a savanna with abundant vegetation. The African Monsoon moved north during this period, resulting in the creation of a series of wetlands that enabled men to settle in the region [14]. The Sahara's hyperarid climate stabilized around 4,000 B.P. after a period of desiccation that started around 6,000 B.P. The African monsoon started to move south about 5,000 B.P., causing the area to completely dry up.

1.2.2 Recent change in North Africa climate

Owing to its high vulnerability and low adaptive capability, North Africa overall, is one of the most vulnerable regions to global warming [15]. Extensive studies have shown that different parts of North Africa are getting drier and hotter [16]. [16] showed a decrease in precipitation between February to April over 1950 to 1999. While the decline in annual precipitation over North Africa extends until present time, we note a strong increase in T2m temperature between 1975 to 2019 (Figures 1.7 and 1.8). [17] have calculated a set of climate indices, based on daily temperature and precipitation data from the GCOS ¹. They found positive trend in the number of warm days and nights in the northern Mediterranean African countries (Algeria, Egypt, Libya, Morocco and

¹Global Climate Observing System (GCOS) Surface Network (GSN) data set for North Africa and coordinated by the WMO Commission for Climatology (CCI) Expert Team on Climate Change Detection and Indices (ETCCDI) (cccma.seos.uvic.ca/ETCCDI/)

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Tunisia) from 1951 to 2003, as well as negative trends in the number of cold days and nights. In certain parts of northern Algeria and Tunisia, they also show positive, but non-significant, trends in severe precipitation events (drought and high precipitation events). [18] observed positive and significant trends in annual mean temperature across Morocco and Tunisia (0.01° to $0.03^{\circ}\text{C}/\text{year}$) and significant negative trends in annual precipitation in northern Morocco using data from the CRU TS2.0 dataset. During the period 1950–2000, they noticed an increase in the occurrence of mean drought intensity and span in the north of the region [18]. In addition, decadal temperature analyses have shown a strong increase in warming rate across North African, over the last 50 to 100 years, according to the Fifth Assessment Report (AR5) of IPCC [19]. Some other studies analysing the Mediterranean hydroclimate [20, 21] showed that the internal climate variability also contributes to the present observed precipitation winter decline.

It is worth to note that North Africa region is among the regions with scarce climate data. Consequently, using the different available gridded data set (Delaware [22], GPCC [23]...) is of great benefit to estimate the uncertainty in observation products. However, it is difficult to provide an accurate diagnosis of the uncertainties in the observations since it is always likely that some unknown mistake has polluted the data and there is no method to account for such unknowns quantitatively. The data does, however, include a number of recognized flaws. Land data uncertainties can be classified into three categories [24]: (1) station error, which is the uncertainty of individual station anomalies; (2) sampling error, which is the uncertainty in a grid box mean due to estimating the mean from a few point values; and (3) bias error, which is the uncertainty in large-scale precipitation (temperature) due to systematic changes in measurement methods.

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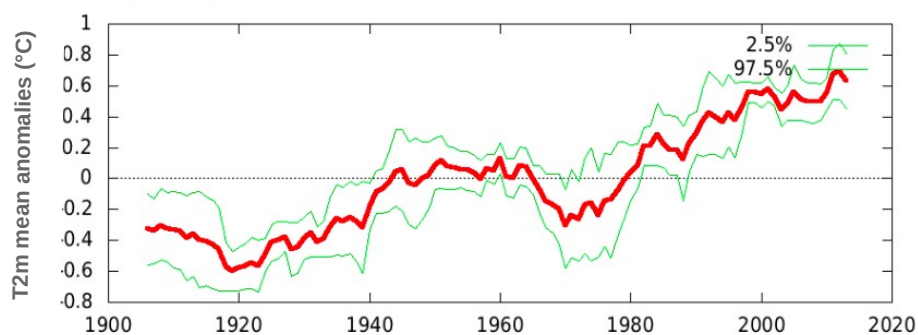


Figure 1.7: CRU_TS_4.04 (1900-2019), 12 years running T2m mean anomalies (red line), averaged over North Africa land (10W-11E, 30N-36N). In green color is represented both, the 2.5% and 97.5% confidence intervals.

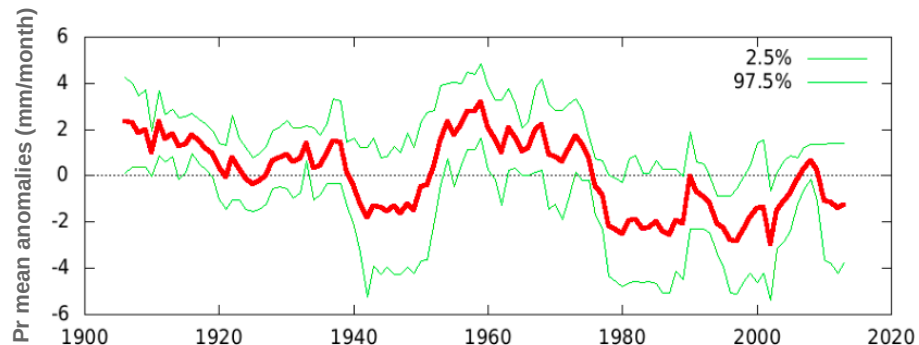


Figure 1.8: CRU_TS_4.04 (1900-2019), 12 years running precipitation mean anomalies, averaged over North Africa land (10W-11E, 30N-36N)(red line). In green color is represented both, the 2.5% and 97.5% confidence intervals.

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The high spatial and temporal variability of the seasonal mean temperature and total precipitation in the Mediterranean basin is controlled by several mechanisms [25]. The particular location of the Mediterranean, a semi-enclosed sea surrounded by lands, in a subtropical zone coupled with complex geography, high mountain ridges, peninsulas and islands, makes its climate highly sensitive to atmospheric circulation variability. In winter, North Africa's hydroclimate is strongly influenced by disturbances coming from the North Atlantic and controlled mostly by the action of the downward branch of the Hadley cell [26]. Additionally, the cold surface water flow associated with the Canary flow, which is a current flowing southwestward along the northwest African coast, from 30° N to 10° N and from offshore to 20° W [27], affects mainly that coast [28].

Previous works have substantially contributed to the description of the atmospheric patterns that are related to precipitation variability in the Mediterranean basin on different timescales [29, 30, 31, 32, 33, 34]. These studies have shown that a large part of the spatial and temporal variability of rainfall is closely linked to a few large-scale atmospheric modes of variability that will be defined explicitly in the next section, mainly the North Atlantic Oscillation (NAO), the East Atlantic (EA) pattern, the Mediterranean Oscillation (MO) and the Western Mediterranean Oscillation (WeMO) [29, 30, 32]. It is suggested in [35] that the heavy precipitation events in the east of the Iberian Peninsula are linked to the NAO, MO and WeMO negative phases. Furthermore, [36] showed that the increased precipitation along the Iberian Peninsula is mainly due to prevailing negative phases in the NAO and MO. More generally, the western part of the Mediterranean region is characterized by a strong anti-correlation between winter precipitation and the NAO, which is related to the transport of moisture by storm tracks [26, 29, 30, 37]. The development of Mediterranean storminess and precipitation is indeed strongly linked to the atmospheric circulation over the Atlantic

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and Europe [33, 35, 36, 37]. It was shown in [34] that NAO variability in winter explains up to 30% of the decadal winter precipitation changes in a region covering parts of Spain, Morocco, Southern France, Italy and the Balkans.

Additional studies have underlined the importance of other atmospheric modes of variability (that will be defined explicitly in the next section), in particular the Upper-Level Mediterranean Oscillation index (ULMOi) and the East Atlantic (EA), East Atlantic/West Russia (EATL/WRUS) and Scandinavia (SCAND) patterns [38, 39]. They have shown that these four teleconnection patterns have a substantial impact on the frequency of Euro-Mediterranean extreme precipitation. Additionally, variations in zonal wind speed at the 850 hPa level explain a large fraction of the interannual changes in area-averaged Mediterranean precipitation and storminess, with stronger westerly winds (positive wind anomalies) associated with positive precipitation anomalies [40]. During the cold season (November–April), the Mediterranean region is subjected to weak atmospheric depression derived from Europe, which, together with a high-pressure zone over North Africa and the Sahara, induces a strong westerly flow [40]. The subtropical high-pressure zone over North Africa might be viewed as a side of the descending branch of the Hadley cell, which, with climate warming, is expected to enlarge poleward [37]. This phenomenon seems remarkably robust over the Mediterranean region [40], but for reasons that are still not well understood.

Atmospheric Circulation Indices

Atmospheric circulation is characterized in terms of circulation indices, generally based on normalized SLP differences between two distant locations where observations are selected. We used four circulation indices obtained from station-based observations for the recent period (1979–2005). These indices are NAO, MO, WeMO and U850 (Figure 1.9) and are defined as follows:

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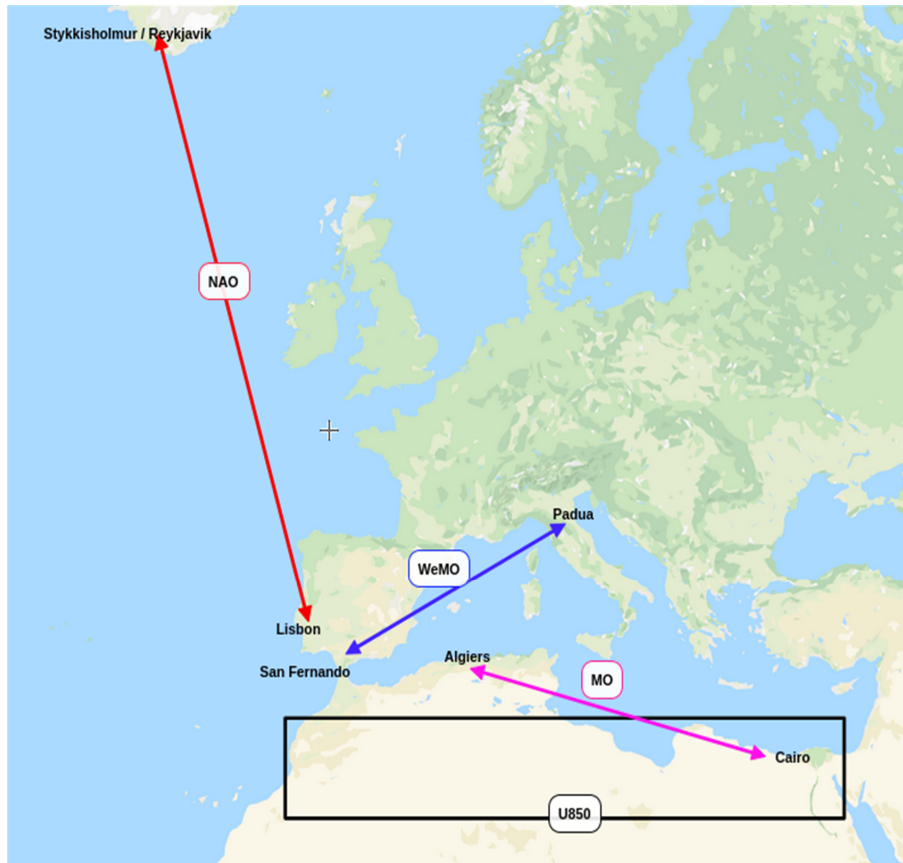


Figure 1.9: Circulation indices, North Atlantic Oscillation (NAO), Western Mediterranean Oscillation (WeMO) and Mediterranean Oscillation (MO) and zonal wind speed at 850 hPa (U850, averaged over black box as in [40]) (map source: <https://www.scribblemaps.com>).

1. The NAO index is calculated as the difference in normalized SLP between Lisbon, Portugal and Stykkisholmur/Reykjavik, Iceland (Figure 1.9). Normalization is obtained by removing the long-term mean and dividing by the long-term standard deviation of the interval 1864–1983. This is done to avoid the series being impacted by the greater variability of the northern station [29, 41].

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2. The MO index was calculated by [42, 43, 44] as the normalized pressure difference between Algiers (36.4° N, 3.1° E) and Cairo (30.1° N, 31.4° E) (Figure 1.9). The MO has been viewed as the dominant regional low-frequency atmospheric pattern impacting rainfall in the Mediterranean region [33, 45, 46].
3. The WeMO index is defined as the difference between the standardized surface pressure values recorded at Padua (45.40° N, 11.48° E) in northern Italy and San Fernando (Cádiz) (36.28° N, 6.12° W) in southwestern Spain (Figure 1.9) [47].
4. The zonal wind index (U850) is related to the eddy-driven jet and storm track action over the North Atlantic [40]. This index is obtained by averaging the monthly zonal wind over 25° N–33° N and 10° W–23° W (Figure 1.9).

Figure 1.10 shows the time series evolution of NAO, WeMO and MO indices over 1950-2013. According to Mann-Kendall test for trend detection (see Chapter 3), there is no significant trends in the time series, except for WeMOI, where a decreasing trend starting 1960 is detected (with significant $p - value = 0.02 < 0.05$).

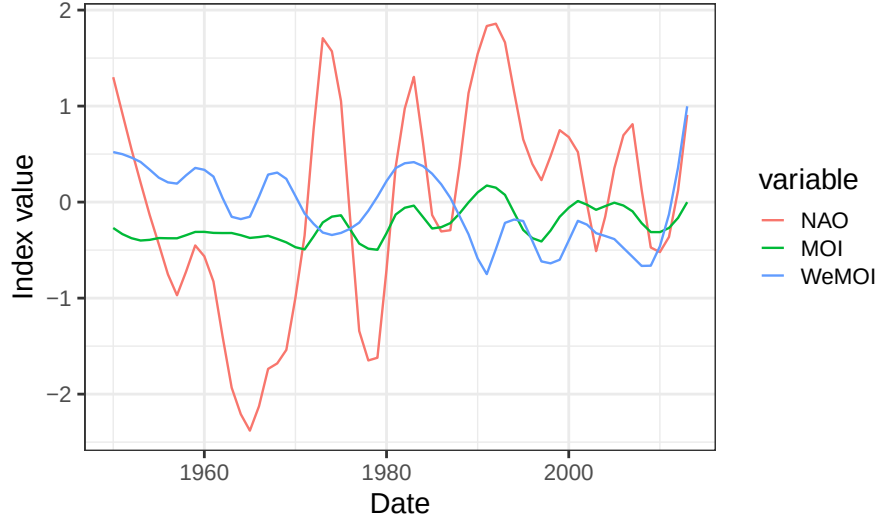


Figure 1.10: Smoothed 12 months time series of North Atlantic Oscillation (NAO), Western Mediterranean Oscillation (WeMO) and Mediterranean Oscillation (MO), over 1950-2013.

Computation of the atmospheric indices

1. NAO, MO and WeMO indices

To compute the different atmospheric indices (NAO, MOI, WeMO) shown in Figure 1.9, we calculate the difference in standardized sea level pressure (SLP) or geopotential height (Gz) anomalies between two reference points (Table 1.2).

For instance for the NAO: $A_{(m,y)}$ and $D_{(m,y)}$ are the monthly means of the pressure with m the month ($m = 1$ to 12) and y the year ($y = y_1$ to y_2). For each month are calculated the mean and the standard deviation, for A (the Azores high pressure) and D (the Icelandic low pressure) separately and over a reference period (over the three periods of the last millennium).

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$$\overline{A_m} = \frac{1}{y_2 - y_1 + 1} \sum_{y_1}^{y_2} A_{m,y} \text{ et } \sigma_m = \sqrt{\frac{1}{y_2 - y_1} \sum (A_{m,y} - \overline{A_m})^2} \quad (1.1)$$

Then each monthly data is normalized by subtracting the average of the corresponding month and by dividing by its standard deviation:

$$A'_{m,y} = \frac{A_{m,y} - \overline{A_m}}{\sigma_m} \quad (1.2)$$

We do the same task for $D_{m,y}$.

NAO index for a given month m and year y is defined by:

$$NAO_{m,y} = A'_{m,y} - D'_{m,y} \quad (1.3)$$

The same method is applied for MO and WeMO indices, using SLP (or Gz).

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NAO	<i>Portugal (Lisbon)</i>	<i>Iceland (Stykkisholmur/Reykjavik)</i>
Latitude	38.76	64.13
Longitude	-9.09	-21.95

MO	<i>Algeria (Algiers)</i>	<i>Egypt (Cairo)</i>
Latitude	36.40	30.1
Longitude	3.10	31.4

WeMO	<i>Italy (Padua)</i>	<i>Spain (San Fernando)</i>
Latitude	45.24	36.28
Longitude	11.52	-6.12

Table 1.2: Location of the reference points defining the atmospheric modes (NAO, MO and WeMO).

2. U850 index

The U850 index is calculated as the area averaged (over 10W-20E, 25N-32N) zonal wind velocity anomalies at 850 hPa, as in [48].

We retrieved all the needed data set from observations, reanalysis and (PMIP/CMIP5) climate models over the three studied periods of the last millennium (850-1850, 1850-1950 and 1979-2013).

The links between these atmospheric modes of variability (NAO, MO, WeMO and U850 indices) and precipitation, averaged over the three study areas (see chapter 2), are evaluated for the 6-month November–April mean using Pearson’s correlation coefficients. This period accounts for the majority of annual rainfall.

Analyses over the past decades have suggested that the Mediterranean

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drying in winter over the period 1960–1990 was due to the multidecadal variability of the NAO, with a much smaller part related to radiative forcing in the form of rising CO₂ and global warming [34]. In contrast, [49] showed that human-induced greenhouse gas and aerosol forcing are the principal causes of winter drying in the Mediterranean region. This is confirmed by a recent study [50], which also pointed out that rainfall is more reactive to black carbon (BC) than to well-mixed greenhouse gases (WMGHGs). In addition to regional warming, BC and WMGHGs appear to reduce precipitation by inducing a sea level pressure (SLP) anomaly similar to the positive phase of NAO, with lower SLP at high latitudes and higher SLP at mid latitudes [50]. This leads to a northward movement of the jet stream and storm tracks, increasing precipitation in northern Europe while decreasing it in the Mediterranean.

The Mediterranean Sea is a semi-enclosed basin with variable depth. The surrounding orography generates cold, dry northern regional winds (Tramontane, Mistral, Bora...) that blow over the Mediterranean Sea, causing heat and freshwater losses due to evaporation and latent heat transfer [51]. These Mediterranean features influence water budget and thermohaline circulation over the basin and may lead to major impacts on the climate system, including heavy rain and cyclones that contribute to precipitation change in North Africa [25].

1.2.3 Future climate changes in North Africa

Representative Concentration Pathways scenarios (RCP) are scenarios that comprise emissions time series and concentrations of greenhouse gases (GHGs), aerosols, gases from chemicals actives and land cover [52]. The term "representative" point out to the fact that every RCP characterizes just one of many possible scenarios for reaching the specified radiative forcing features. The "pathway" term stresses that it's not solely the long-run concentration levels that are important, but also the

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time taken to get there [52]. In the Fifth IPCC Assessment, four RCPs are used as the foundation for the climate projections and estimates provided in WGI AR5 Chapters 11 to 14 [52]:

- RCP2.6 One pathway in which radiative forcing rises up to 2.6 Wm^{-2} before 2100 and then decreases.
- RCP4.5 and RCP6.0 Two intermediate stabilisation pathways wherein radiative forcing stabilizes around 4.5 Wm^{-2} and 6.0 Wm^{-2} after 2100.
- RCP8.5 One high pathway for which radiative forcing exceeds 8.5 Wm^{-2} by 2100 and rises farther for long time.

According to RCPs (RCP6.0 and RCP8.5), large parts of North Africa will warm by more than 2°C by the end of the century, relative to the late 20th century mean annual temperature [19]. The IPCC (AR5) also found that land temperatures in North Africa are likely to rise faster than the global land average, particularly in arid regions. In terms of predicted temperature extremes, there is a strong likelihood that heatwaves and warm spell periods will rise in North Africa, implying that periods of hot days will last longer by the end of the century [19].

[53] showed a projected decline in precipitation over Northern Africa by the end of the 21st century in both RCP6 and RCP5.8 scenarios. The drying/warming at the annual scale over northern Africa (along with North of Morocco, Algeria, Libya, Egypt and Tunisia) is a reliable feature in the global and regional climate change projections for the 21st century according to the IPCC AR5 (RCP6 and RCP8.5 scenarios) [53]. However, [53] showed a positive trend in precipitation in the autumn (September to November) in some areas of northern Algeria and Morocco. Precipitation projections are less certain than temperature projections and have high seasonal and spatial dependence [54]. Figures 1.11 and 1.12 show precipitation decline over 2050-2100 (relative to 1985-2005) in both winter (November-April) and summer (May-October)

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seasons according to CMIP5 models mean (RCP8.5) scenarios. The projected RCP8.5 precipitation decrease scenario is about 10 to 30 %, going from the East (Tunisia) to the West (Morocco) of North Africa [55]. Figures 1.13 and 1.14, show a projected (2050-2100) temperature increase (relative to 1985-2005 period) of about 2 °C in winter to 4 °C in summer, over North Africa (scenario RCP8.5). Nevertheless, the outcomes of the different simulation models vary in terms of the underlying atmospheric dynamics, as well as the time horizons over which they occur, leaving these projections with considerable uncertainties. [55].

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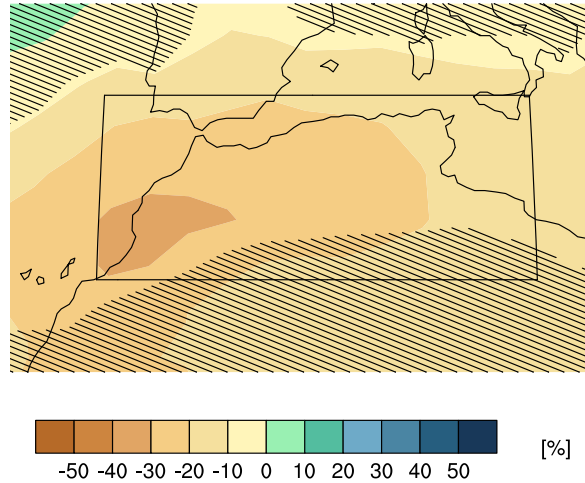


Figure 1.11: Mean CMIP5 (40 models, RCP8.5) winter (November-April) precipitation change over 2050-2100 relative to 1985-2005. The hatching represents areas where the signal is smaller than one standard deviation of natural variability (calculated from: <http://climexp.knmi.nl/>).

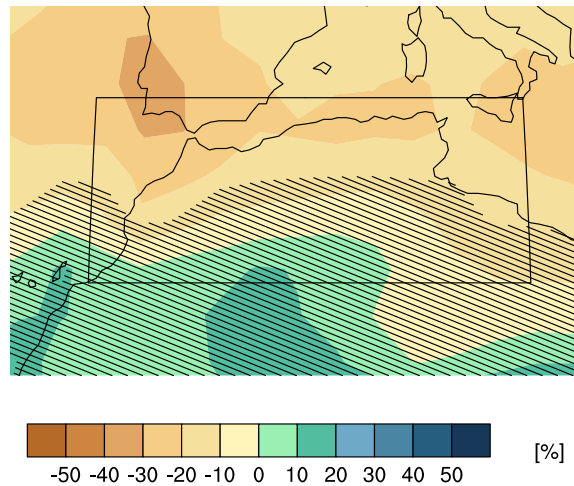


Figure 1.12: Mean CMIP5 (40 models, RCP8.5) summer (May-October) precipitation change over 2050-2100 relative to 1985-2005. The hatching represents areas where the signal is smaller than one standard deviation of natural variability (calculated from: <http://climexp.knmi.nl/>).

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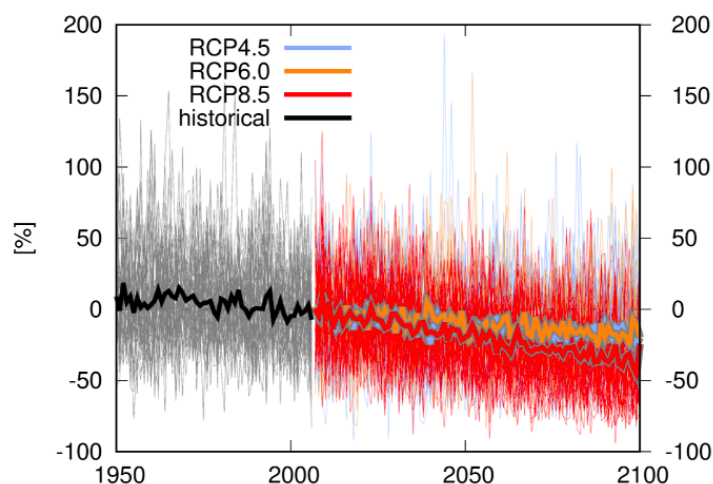


Figure 1.13: Precipitation change in North Africa (30N-36N, 10W-11E), November-April 1950-2100 relative to 1985-2005 AR5 CMIP5 (40 models). For each scenario one line per model is shown plus the multi-model mean (calculated from: <http://climexp.knmi.nl/>).

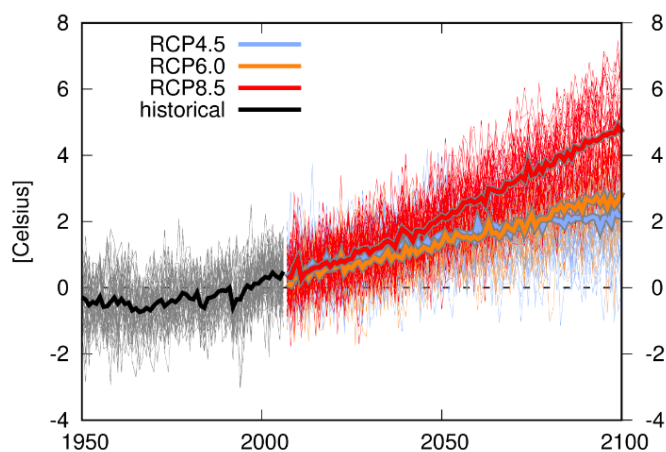


Figure 1.14: Temperature change in North Africa (30N-36N, 10W-11E), November-April 1950-2100 relative to 1985-2005 AR5 CMIP5 (40 models). For each scenario, one line per model is shown plus the multi-model mean (calculated from: <http://climexp.knmi.nl/>).

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Under future climate change (RCP8.5 scenario), a significant part of winter precipitation decline and increased drying will occur more in the northwestern part of Africa (Morocco) compared to the eastern parts (Figure 1.11), from 30% to 40% in (December–February) and from 20% to 25% over eastern Mediterranean [56]. [57, 58, 59, 56] showed that changes in low-level circulation at regional scale may have a main control in winter Mediterranean drying, caused by increased moisture divergence by the time-mean flow due to anomalous high pressure circulation (Figure 1.15). [40] argued that precipitation changes is well identified by a simple wind index reflecting regional circulation, reinforcing the hypothesis that future Mediterranean hydroclimate trends are mainly induced by modifications in regional atmospheric circulation. The magnitude of the strength of the anticyclonic anomaly and that of the associated wind field, is strongly correlated to the regional precipitation decline across models [40]. [60] showed that the anomalous high pressure in the Mediterranean, may cause the projected drying by increasing atmospheric stability in the region and suppressing Mediterranean cyclones, as discussed in [61, 40]. [60] showed that this anticyclone circulation over the eastern Mediterranean, expands from November to April.

Among hypotheses which may explain the anomalous surface anticyclonic circulation over the Mediterranean, [62] showed that, it may be due to a relative temperature decrease over the Mediterranean Sea surface. Over the winter, the Mediterranean Sea is on average warmer than the surrounding land due to water's higher thermal capacity. Future climate projections, however, show a robust and steady relative cooling of the Mediterranean Sea with respect to adjacent land as a result of increased warming over land due to climate change [63].

[57] showed that changes in the stationary waves under increasing greenhouse gases, is also specifically connected with the high pressure anomalies projected by CMIP5 models above the Mediterranean.

Changes in global atmospheric circulation, for instance the Hadley cell

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expansion [64] and the subsequent shift of the North Atlantic storm track [65, 66] among others, have previously been proposed as drivers of precipitation reduction over the North Africa region.

However, in abrupt CO₂ experiments, the role of the Hadley cell shift and storm track changes was contested because they occur at much faster timescale than the Mediterranean drying [67]. The zonal symmetry of Hadley cell shift is also very different from the Mediterranean Sea Level signal, so it is not obvious why the Mediterranean is particularly vulnerable in climate projections.

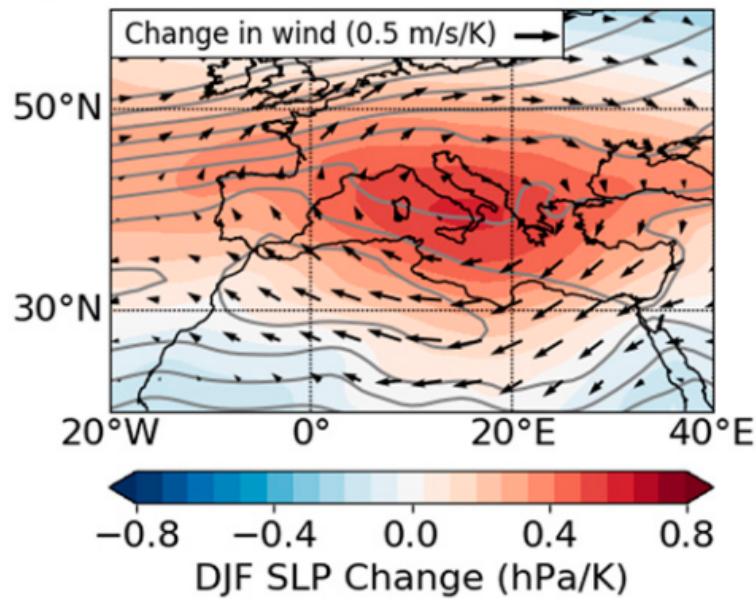


Figure 1.15: CMIP5 winter (DJF) multimodel mean projected change in Mediterranean SLP ($\Delta SLP/\Delta T$), under RCP8.5 (2071–2100 minus 1976–2005). Historical DJF SLP contours are also shown in (from 1010 to 1024 hPa every 2 hPa; the closed contour over North Africa is 1024 hPa), source: [68].

1.3 Projected impacts of climate change

Droughts events

Drought is a creeping natural phenomenon that slowly sneaks up and impacts many sectors of the economy and operates on many different time scales [69]. It occurs in all climatic zones, but its characteristics vary significantly from region to region. It differs from aridity that is restricted to low rainfall regions and is a permanent feature of climate. Defining drought is therefore difficult [70]. The Glossary of Meteorology [71] defines a drought as “a period of abnormally dry weather sufficiently prolonged for the lack of water to cause serious hydrological imbalance in the affected area”. Compounding factors, such as poverty and inappropriate land use, increase vulnerability to drought. When drought causes water and food shortages, there can be many impacts on the health of the population, which may increase morbidity and result in death [72]. In the period from 1970 to 2012, drought caused almost 680 000 deaths, due to the severe African droughts notably in 1975, 1983 and 1984 [72]. It is common for droughts to be classified into a set of categories [73].

The meaning of the drought differs from the meteorologist to the hydrologist, or an economist. Drought was divided into four types by the American Meteorology Society: meteorological, agricultural, hydrological and socioeconomic droughts [74] (Figure 1.16). The first three approaches are concerned with determining how to quantify drought as a physical phenomenon. The last deals with drought in terms of supply and demand, following the consequences of water scarcity as they spread through socioeconomic structures. [73].

(i) **Meteorological drought** is defined as a prolonged lack of precipitation over a given region. Evaluating meteorological droughts is based on the precipitation metric [75, 76]. [76] define meteorological drought as a water deficit caused by a difference between precipitation and evaporation. Other approaches analyze drought duration and intensity

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in relation to cumulative precipitation shortages [76, 77].

(ii) **Agricultural drought** typically represents the period of decreasing soil moisture and consequent crop loss, that has little to do with surface water resources. Soil moisture deficit is influenced by a number of causes, including meteorological and hydrological droughts, as well as variations in actual and potential evapotranspiration. Plant water demand is influenced by the temperature, the biological characteristics of the plant, its stage of development and the physical and biological properties of the soil. To study agricultural droughts, multiple drought indices based on the integration of precipitation, temperature and soil moisture have been created.

(iii) **Hydrological drought**: A period of limited surface and subsurface water supplies for existing water uses of a given water resources management system is referred to as a hydrological drought. For hydrologic drought studies, stream-flow data has been a commonly used metric [78, 76, 79, 80]. Geology is one of the key influences driving hydrological droughts, according to regression analyses comparing droughts in stream-flow to catchment resources [81, 82].

(iv) **Socioeconomic drought** takes into account the impact of the previous drought types such as meteorological, agricultural or hydrological drought, on the production and availability of certain economic goods (e.g. fruits, vegetables, cereals,...) [83].

In North Africa, drought has become a long-standing and ongoing issue [84]. According to [85], North African dry areas face lower rainfall amount and higher rainfall variability than coastal regions, which may result in several years of drought. Under the Representative Concentration Pathway RCP8.5 [86, 87, 88, 89], North African countries (particularly Morocco, Algeria and Tunisia) are consistently projected to become global hot-spots for drought by the end of the twenty-first century [86, 87, 88, 89]. The projections for an increase in severe drought conditions across the Mediterranean and Northern Africa are reliable across a number of studies [90]. Examples of direct effects of drought in

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North Africa includes crop yield decrease, increased fire risk, increased livestock mortality, decline in water levels [91]. Drought could also serve as a trigger, increasing population unrest, amplifying disparities in access to water and facilities and reinforcing marginalization myths. [91]. Droughts harm the poor rather than the rich [91].

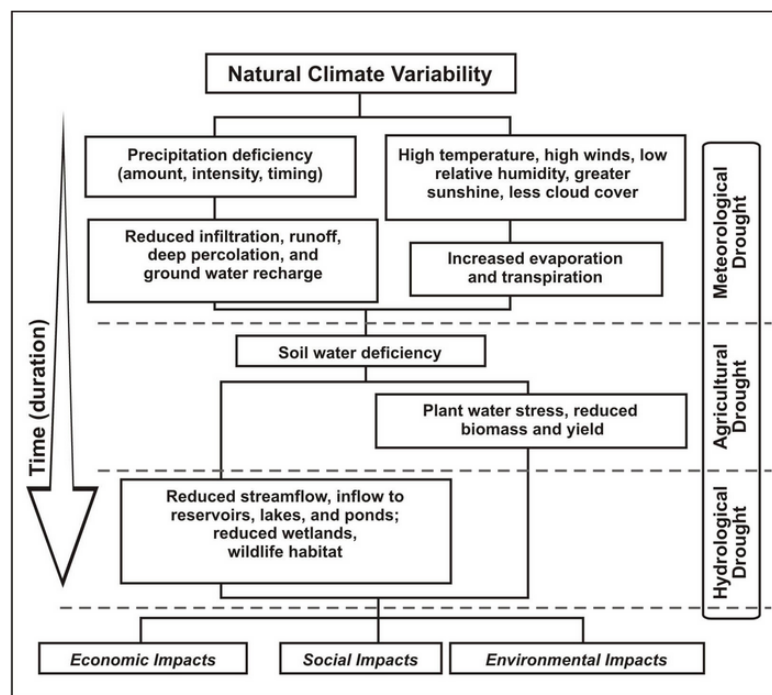


Figure 1.16: The relationship between meteorological, agricultural, hydrological and socio-economical droughts and its impacts. (Source: National Drought Mitigation Center) [92].

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Water supply

The south part of the Mediterranean basin has seen a substantial increase in mean temperature in recent decades, which increases the region vulnerability to climate change. The climate change will exacerbate existing water scarcity in North Africa [93]. Water resources are subjected to high spatial and temporal variability and are a major obstacle to the region's economic development. North Africa countries are within the 15 most water-stressed countries worldwide, according to [94]. Freshwater scarcity is the first among the serious problems affecting North Africa, owing to its effect on food security, economic growth, poverty reduction and sociopolitical stability [94]. Natural water stress, which is intensified by increased water demand as a consequence of the North Africa region's mainly (hyper-)arid climate, is exacerbated by population development, urbanization and growth. According to [95], climate change is estimated to account for 22% of increased future water storage in North Africa by 2050, while socioeconomic conditions will account for 78% of increased projected water scarcity. Reduced snowpack in the Atlas Mountains is due to a combination of warming and reduced precipitation [96]. Due to a rise in potential evapotranspiration from warmer climates, decreases in water resources will typically be greater than the associated reductions in precipitation (e.g. river runoff or groundwater recharge) [96].

1.4 Objectives

Many efforts have been addressed to understand the hydroclimate variability in the Mediterranean basin. However, a lot of works is still remaining to better explain the mechanisms responsible of precipitation changes over North Africa region. This region, which is expected to be highly impacted by climate changes (hot spot), suffers from the lack of scientific studies devoted to its specific climate.

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Consequently, this thesis work will be devoted to better understand the main mechanisms responsible of precipitation variability in North Africa region (going from Morocco to Tunisia). Our main question is: **Are the atmospheric mechanisms that dominate precipitation changes over the whole Mediterranean basin the same over North Africa?** To achieve our goal, we will decompose this main question into two parts, by answering the following questions:

- *Question.1: What are the atmospheric patterns that govern precipitation changes in North Africa?*

As shown previously in different studies, precipitation change in the Mediterranean basin is related to different atmospheric patterns (NAO, WeMOI, MO,...) depending on the study area. Additionally, most of these studies do not go back further in time. Here we will analyse the link between the precipitation variability in North Africa, with these standard atmospheric mode of variability, over the last century and last millennium. The goal is to understand the past changes in precipitation, compared to the present climate and future projections. We will dig in more details by dividing North Africa region into three sub-regions (West-Center-East), to see whether the link of precipitation change with those atmospheric patterns is the same from the West to East. Additionally, this will let us evaluate the robustness of these links over time.

- *Question.2: Is the North Africa precipitation response to increasing GHG the same over the Western to Eastern parts, or are there differences in the sub-regions (Morocco, Algeria to Tunisia)?*

As showed in different studies, the changes in low-level circulation at regional scale may have a main control in winter Mediterranean drying, caused by increased moisture divergence by the time-mean flow due to anomalous high pressure circulation. In this part, we will analyse the atmospheric circulation changes under the present and future climate. The goal is to understand whether these

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changes are the same going from the western to the eastern North Africa.

These questions will be addressed in this present work, to contribute to the benefit of North Africa in preparedness and management of water resources.

In addition to the importance of this part of scientific research, we have not neglected the practical side. In order to allow meteorological and climatological services (e.g.: National office of meteorology ONM) as well as water resource managers (National Agency for Hydraulic Resources) to have an easy-to-use drought monitoring and forecasting tool, we will develop a drought web application based on R-Shiny. This application will allow the daily monitoring of droughts and/or excess rainfall in Algeria, using Synoptic observation data sets. This application is described in the last chapter of the thesis.

CHAPTER 2

Robustness of the Link between Precipitation in North Africa and Standard Modes of Atmospheric Variability during the Last Millennium

Note: This Chapter is published as journal article (<https://www.mdpi.com/2225-1154/8/5/62>). A part of this introduction is moved to the general introduction, to avoid repetitions.

2.1 Introduction

Recent hydroclimate changes in North Africa can be put into perspective by analyzing drought variability over the Common Era (CE, year 0–present). This has been done using a variety of proxies, including tree rings [97, 98, 99, 100, 101], sediment cores [102, 103] and networks including various proxies [104, 105, 106]. Compared to other Mediterranean regions, little is known about North Africa.

In [107] a reconstruction of precipitation was developed covering the past millennium in the Atlas Mountains of Morocco (a part of the mountain range in the Maghreb separating the Mediterranean and Atlantic coastlines from the Sahara Desert, reaching 2500 km through Morocco,

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Algeria, and Tunisia). They showed that in the late 20th century, drought was exceptional in the context of the prior 500 years. A comparison between general circulation model (GCM) output and proxy-based precipitation and temperature reconstructions over the past 500 years showed that while no model accurately simulated the reconstructed changes, reasonably good agreement was found after 1650, with models taking into account changes in volcanic forcing, solar irradiance, and greenhouse gases [108].

The Old World Drought Atlas (OWDA) is a gridded ($0.5^\circ \times 0.5^\circ$) reconstruction based on tree-ring data providing an estimate of the June, July, August (JJA) Palmer Drought Severity Index (PDSI) for Europe and the Mediterranean region over the past 2000 years [109]. The PDSI provides integrated information about relative dryness and wetness and is more representative of the conditions impacting tree growth than, for instance, precipitation. This is why it was chosen as the target for reconstruction over past centuries. However, in North Africa, there are few tree-ring time series included in the OWDA database. Only two time series of *Cedrus atlantica* (cedar) ring width data are used to recreate long-term changes in the PDSI in Morocco over the past 953 years [107].

The main goal of this study is to provide a better understanding of the variability of precipitation in the North African region (from Morocco to Tunisia) over the past millennium and to compare those long-term changes with more recent ones observed over the past century. Only a few studies have addressed the issue of long-term rainfall variability in this region, as most were limited to the last 20th century period. Additionally, most of that research focused on the whole of the Mediterranean basin or its northern part (Euro-Mediterranean region), with very little information specific to North Africa. The potentially different behavior in North Africa compared to the rest of the Mediterranean region justifies our focus on this important area.

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In our manuscript, we divide the region of North Africa into three sub-regions (Figure 2.1) from west to east to take into account the potentially decreasing influence of the North Atlantic. The links between winter precipitation in sub-regions over the past millennium and the atmospheric modes of variability are separated into three periods (850–1850, 1850–1950, 1979–2005) to test the robustness of the correlations over time. We focus here on the NAO, WeMO, and MO indices in addition to the U850 zonal wind index introduced in [40], because a preliminary analysis showed that they are the best candidates among existing indices to explain the variability in our region of interest.

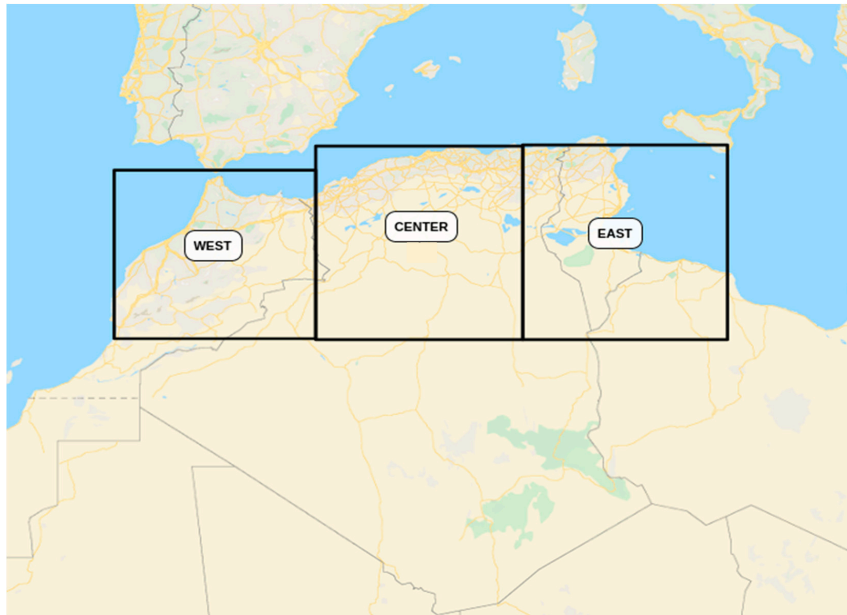


Figure 2.1: Study area. Precipitation is averaged over solid boxes (western, central, and eastern North Africa) (source: <https://www.scribblemaps.com>).

The outline of this chapter is as follows: Section 2.2 describes the atmospheric indices, data, and methods. Section 2.3 discusses the links

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between these indices and winter precipitation variability in the North Africa region and compares model results and reconstructions over the last millennium. Finally, conclusions are given in Section 2.4.

2.2 Data and Methods

The present work is based on an analysis of winter (November to April) precipitation variability over North Africa (Figure 2.1) derived from observations, reanalysis, and model simulations over the last millennium. Several gridded datasets of monthly precipitation, SLP, temperature, and zonal wind speed at a pressure level of 850 were used (Tables 2.1 and 2.2). These data were remapped to the same grid at 1° resolution and the values over the ocean were masked to only consider the land areas.

The main study area is the North Africa region. This area is divided into three boxes, because the literature [35, 36, 110] suggests that the link between atmospheric patterns (NAO, WeMO, MO) and precipitation is highly heterogeneous in the region. The western part (left box) is mainly influenced by the North Atlantic Oscillation (NAO), which is not the case for the central part, which is found to be a transitional zone between the North Atlantic perturbations and the eastern part (right box), which is mainly under the impact of perturbations coming from the east, and the Mediterranean Oscillation (MO).

The three boxes are defined to roughly match the regions with decreasingly direct influence of perturbations coming from the North Atlantic. For simplicity and to avoid any bias that could come from a more specific choice, they have the same extent in longitude but differ in latitude to only cover the North Africa region (only land area, without the ocean in the box).

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2.2.1 Statistical Tests

In addition to standard correlation analysis, we applied the modified Mann–Kendall test [111] and Pettitt’s test [112], two widely used non-parametric tests to detect trends and change points in time series of climatic and hydrological parameters, respectively. The modified Mann–Kendall is a variant of the Mann–Kendall test [113, 114], which can be calculated by the equation:

$$S = \sum_{i=1}^n \sum_{j=1}^{i-1} \text{sign}(x_i - x_j) \quad (1)$$

where n is the total length of data, x_i and x_j are sequential data values, and the function $\text{sign}(x_i - x_j)$ gives the following values:

$$\text{sign}(x_i - x_j) = \begin{cases} 1, & \text{if}(x_i - x_j) > 0 \\ 0, & \text{if}(x_i - x_j) = 0 \\ -1, & \text{if}(x_i - x_j) < 0 \end{cases} \quad (2)$$

The statistic S is approximately normally distributed with the mean $E(S)$ and variance $\text{Var}(S)$ computed as follows:

$$\begin{aligned} E(s) &= 0 \\ \text{Var}(S) &= 1/n \left[n(n-1)(2n+5) - \sum_t t(t-1)(2t+5) \right] \end{aligned} \quad (3)$$

where $\sum_t$ means summation over all associated numbers of values.

In this test, the null hypothesis H_0 is verified if there is no trend in a dataset of n independent randomly distributed variables of equally likely ordering. The use of this method comes with two advantages. First, it is a nonparametric test and does not require the data to be normally

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distributed. Second, the test has low sensitivity to abrupt breaks due to inhomogeneous time series [115].

Throughout the time series analysis, it is important to consider autocorrelation, defined as the correlation of a variable with itself over successive periods of time, before checking for trends. The modified Mann–Kendall takes account of autocorrelation, another reason to choose this statistical method.

The magnitude of the trend is estimated using Sen’s slope estimator, also called the Theil–Sen estimator. The Theil–Sen estimator of a set of two-dimensional points (x_i, y_i) as defined by [116] is the median m of the slopes $(y_j - y_i)/(x_j - x_i)$ determined by all sample point pairs.

Pettitt’s test for change detection, developed by Pettitt [112], is a non-parametric method that is useful for detecting abrupt changes in climate records [112]. According to Pettitt’s test, $x_1; x_2; x_3, \dots, x_n$ is a set of observed data that has a change point at t if $x_1; x_2; x_3, \dots, x_t$ has a distribution function $F1(x)$ that is different from the distribution function $F2(x)$ of the second part of the sequence $x_{t+1}; x_{t+2}; x_{t+3}, \dots, x_n$. For this test, the nonparametric test statistics U_t can be defined as:

$$U_t = \sum_{i=1}^t \sum_{j=t+1}^n \text{sign}(x_t - x_j) \quad (4)$$

This test is commonly used because of its sensitivity to breaks in the middle of any time series [117]. Change point detection is also an important method to determine the period over which a significant trend has occurred in a time series [112].

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Observations and Reanalysis Data	Version and Resolution	Description	Reference
GPCC V7	0.5° latitude x 0.5° longitude global grid (720 × 180)	Monthly gridded land-surface precipitation from rain gauges built on GTS-based and historic data from 1901–2013	[23]
Delaware	0.5° latitude x 0.5° longitude global grid(720 × 360)	Monthly global gridded high-resolution station (land) data for air temperature and precipitation from 1900–2013	[22]
20th century reanalysis (20cRE)	2.0° latitude x 2.0° longitude global grid(180 × 90)	Monthly gridded precipitation Geopotential height Zonal wind at 850 hPa (u-wind) from 1901–2013	[118]

Table 2.1: Observations and reanalysis used in this study.

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In the present study, the modified Mann–Kendall test and Pettitt’s test were used in R packages to analyze precipitation, wind speed (U850), and PDSI variability in North Africa over the last millennium (850–2005).

2.2.2 Observation Datasets and Reanalysis

Several gridded datasets of observations and reanalysis of monthly precipitation were analyzed (Table 2.1). The Global Precipitation Climatology Center (GPCC) analysis V7 [23] has a resolution of 0.5° latitude by 0.5° longitude. It is based on rain-gauge data only and spans the period 1901–2005. The University of Delaware database, using high-resolution station (land) data [22], has a resolution of 0.5° latitude by 0.5° longitude and covers the period 1901–2005. The NOAA-CIRES 20th Century Reanalysis database is a global reanalysis dataset that contains objectively analyzed four-dimensional weather maps and their uncertainty from the mid-19th century to the 21st century [118].

2.2.3 Model Simulations

The simulations used here were performed in the framework of the Palaeoclimate Modeling Intercomparison Project (PMIP3) [119] and Coupled Model Intercomparison Project (CMIP5) [120], whose goal is to coordinate climate modelling activities performed in various centers worldwide. As indicated by their names, PMIP focuses on past periods while CMIP is devoted to the recent past and the future as well as control and sensitivity experiments. PMIP and CMIP activities are coordinated, in particular to use the same model version for analyzing past and recent conditions.

The World Climate Research Program (WCRP), which coordinated these projects, has made available a multimodel dataset intended to improve our knowledge of climate variability. The global scientific community has examined PMIP3/CMIP5 model outputs to produce results that

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support the Intergovernmental Panel on Climate Change (IPCC) Fifth Assessment Report [\[121\]](#).

The models used in this study are those for which the variables of interest were available at the time of analysis for both the past millennium and the past century, covering 850–1850 and 1850–2005, respectively. These models are: CCSM4, MPI-ESM-P, IPSL-CM5A-LR, GISS-E2-R, CESM1, and BCC-CSM1 (defined in Table [2.2](#)).

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Model name	Institution	Period Covered	Data Availability During the Study	Atmospheric Resolution Horizontal Vertical	Reference
CCSM4	National Center for Atmospheric Research (NCAR), USA		850-2005	280 × 200 27	[122]
MPI-ESM-P	Max Planck Institute for Meteorology (MPI-M), Germany	Past millennium	940-2005	192 × 96 47	[123]
IPSL-CM5A-LR	Institut Pierre-Simon Laplace	Past century	850-2005	96 × 96 39	[124]
GISS-E2-R	NASA Goddard Institute for Space Studies		850-2005	250 × 200 40	[125]
CESM1	NCAR, USA		850-2005	144 × 96 27	[126]
BCC-CSM1	Beijing Climate Center, China Meteorological Administration		850-2005	128 × 64 26	[127]

Table 2.2: Modelling centers, characteristics, and references of Coupled Model Intercomparison Project (CMIP5) models used in this study.

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Climate simulations covering the last millennium are driven by both anthropogenic (well-mixed greenhouse gases, ozone, tropospheric aerosols, land use) and natural (solar, volcanic, orbital) climate forcing. Forcing related to volcanic aerosols is obtained from different sources for GISS-E2-R and MPI-ESM-P [128] and the other four Global Simulation Models (GSMs) [129]. The input of volcanic aerosol in the stratosphere after a major volcanic eruption is known to have a major climate effect on the annual to decadal timescale, from observed and historical as well as modeling [130] studies. Observed and reconstructed changes in greenhouse gases (CO_2 , CH_4 , and N_2O) driving past millennium [131, 132] and historical [133] simulations are the same in all models. Both tropospheric ozone and aerosol variation are considered in historical experiments in CCSM4, CESM1, GISS-E2-R, MPI-ESM-P, and BCC-CSM1 and are based on the dataset described in [134], except in IPSL-CM5A-LR, which considers a pre-industrial aerosol concentration. See [135, 136, 137] for more details on the climate-forcing reconstructions used for past millennium experiments and their implementation.

The ability of CMIP5 models to explain the Mediterranean hydrological cycle during the 20th century has been investigated in detail [26, 57]. Their results show that most CMIP5 models can reasonably represent the observed Mediterranean precipitation, evaporation, and moisture fluxes in winter.

2.2.4 Drought Index

The choice of using the PDSI [138] is mainly related to the fact that the hydroclimate related proxy-based reconstructions (such as OWDA) [109] do not directly reconstruct precipitation but rather PDSI. The PDSI incorporates precipitation and temperature and uses a provision and supply model to estimate changes in soil moisture availability over time. The index varies between negative values for dry conditions and positive values for wet conditions. Values below -4 or above $+4$ indicate extreme

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drought or wet spells, respectively [138].

Droughts in North Africa are induced by both lack of precipitation and high evaporation rates. Thus, it is important to use an index that considers soil water availability for vegetation, rather than strictly climatic variables [139, 138].

In order to compare the reconstructed hydrological changes in North Africa (Morocco and Algeria, where reconstructions are available) over the last millennium (850–2005) with CMIP5 model simulations, we computed the PDSI using the MATLAB tool developed in [140]. This application allows calculation of the PDSI with the use of four types of input data: temperature, precipitation, available water capacity (AWC) of the soil [140], also known as field capacity, and the latitude of the location of interest. Note that this part of the analysis could only be achieved using results from the CCSM4, MPI-ESM-P, and CESM1 models due to data availability (the other models do not have the AWC variable).

For the 20th century, instrumental PDSI (1901–2005) from OWDA is from [109] (online resource, accessed on 10 September 2018: <http://drought.memphis.edu/OWDA/ExtractSingleInst.aspx>).

In addition, to further investigate precipitation variability in the region over the last millennium, we calculated Pearson's correlation coefficients between OWDA and the calculated PDSI from CMIP5 models, then between each CMIP5 calculated PDSI for the whole period of the last millennium, as shown in Supplementary Table S1.

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2.3 Results and Discussions

Overall, the CMIP5 precipitation patterns fit the observed annual climatological rainfall cycle in North African regions (Figure 2.2), although some models (in particular GISS-E2-R) overestimate precipitation, while others (IPSL-CM5A-LR) underestimate it. This gives credence for using the CMIP5 models in our analysis.

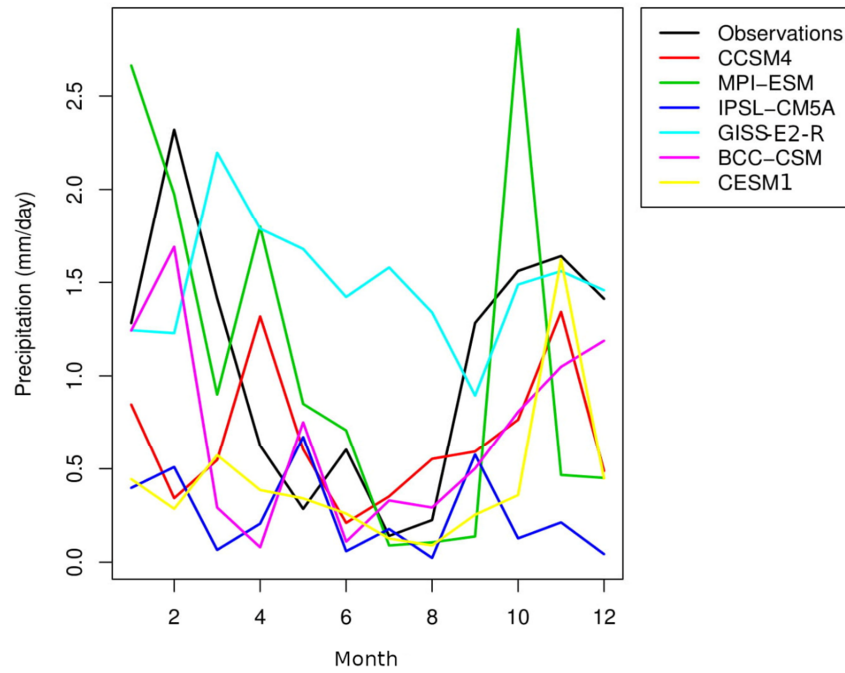


Figure 2.2: Seasonal climatological cycle of precipitation datasets: observations (Delaware) [22] and six Coupled Model Intercomparison Project (CMIP5) climate simulations over 1979–2005 in North Africa (between -5° W– 5° E and 34° N– 36.5° N).

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2.3.1 Link between Winter Precipitation and Atmospheric Modes of Variability in North Africa, 850–2005

The correlations between winter precipitation (observation, reanalysis, and models) and the atmospheric indices (NAO, MO, WeMO, and U850) over the past millennium are presented in Figures 2.3 and 2.4 (exact numbers are given in Supplementary Tables S2–S5). Figure 2.3 shows the spatial distribution for one dataset as an example, while analysis of all datasets for averages over sub-regions and periods is shown in Figures 2.4 and 2.5. For the analyses of different datasets, we determined if correlations were significant for the sub-regions and provide their ranges.

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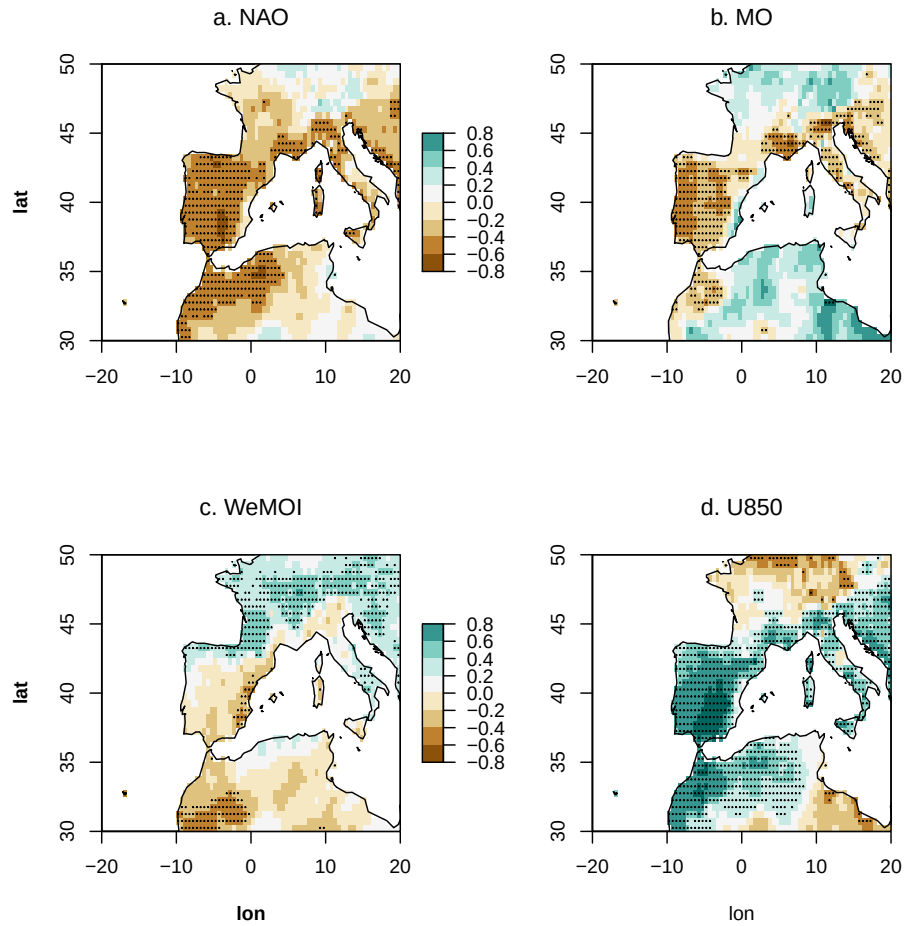


Figure 2.3: Pearson's correlation coefficients between observed precipitation (Delaware) [22] and (A) NAO, (B) WeMO, (C) MO, and (D) U850 for the period November-April 1979–2005. Dotted areas represent significant correlations at 95% confidence interval.

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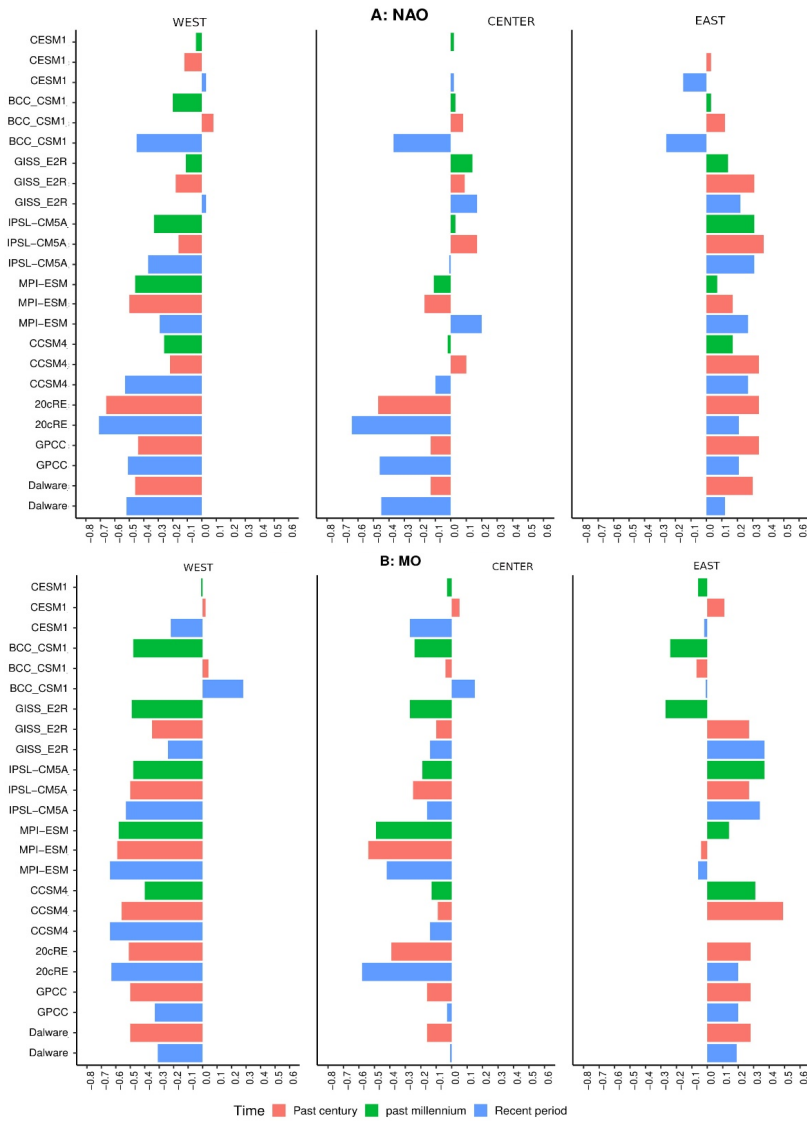


Figure 2.4: Pearson's correlation coefficients between winter (November–April) precipitation (observations, reanalysis, and CMIP5 simulations) and (A) NAO and (B) MO indices for western, central, and eastern North Africa during the recent period (1979–2005), the past century (1850–2005), and the past millennium (850–1850).

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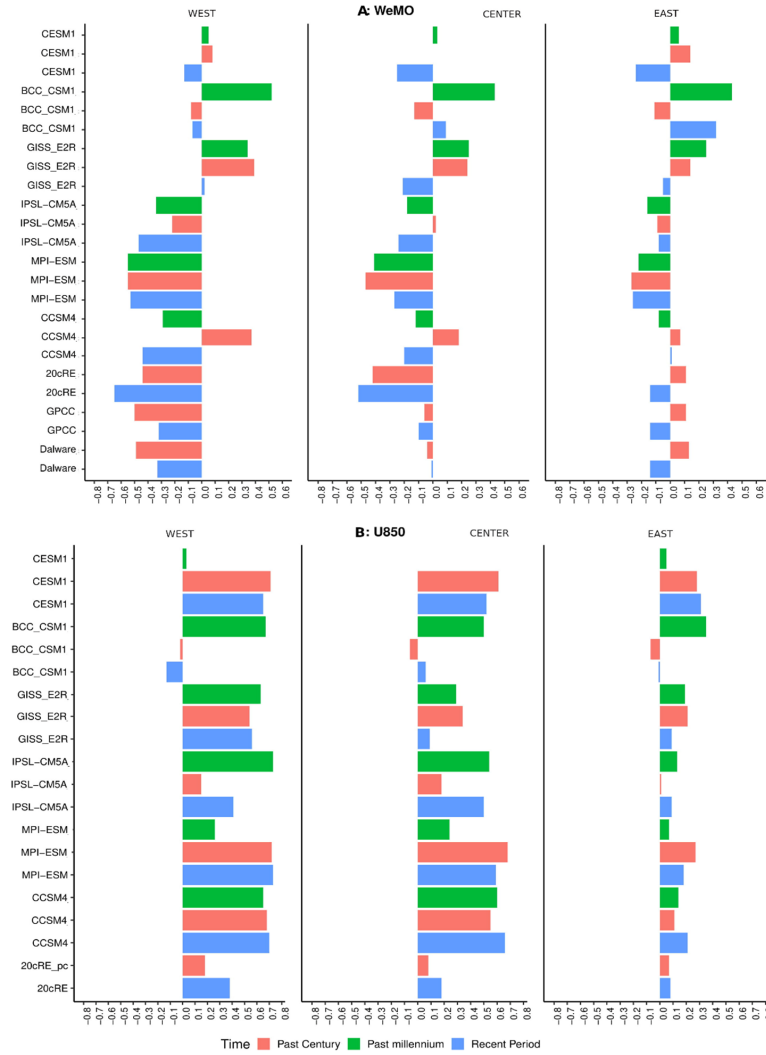


Figure 2.5: Pearson's correlation coefficients between winter precipitation (observations, reanalysis, and CMIP5 simulations) and (A) WeMO and (B) U850 indices for western, central, and eastern North Africa during the recent period (1979–2005), the past century (1850–2005), and the past millennium (850–1850).

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Recent Period (1979–2005):

The negative correlation coefficients between the NAO and the winter precipitation data in the western part of North Africa are significant except for MPI-ESM, GISS-E2-R, and CESM1 models (Figure 2.4A, Supplementary Table S2). The correlations become less significant going from west to east. The range of correlations varies from -0.37 to -0.71 on the west coast of Morocco, -0.45 to -0.64 in the north of Algeria, and -0.38 in Tunisia (only one dataset). These values are in agreement with previous results [26, 29, 30, 35, 49] that found negative correlations between winter NAO and precipitation in the 20th century in the western region of the Mediterranean. This is also consistent with [29, 34], which argued that the NAO is the main atmospheric circulation pattern explaining the hydroclimate variability in this region.

The MO and WeMO show similar spatial patterns to the NAO in the western part of North Africa (Morocco coast) with high significant correlations (except for GISS-E2-R, BCC-CSM1, and CESM1 models). For both indices, the range of correlations varies from -0.31 to -0.65 in the west of Morocco, -0.42 to -0.58 in northern Algeria, and -0.38 to 0.38 in northern Tunisia (Figures 2.4B and 2.5A, Tables S3 and S4 in Supplementary Materials).

In contrast, the correlations between winter precipitation and U850 are positive and highly significant in more datasets than those found for NAO, WeMO, and MOI. The range of correlations varies from 0.38 to 0.72 in the western part, 0.50 to 0.65 in the central part, and 0.18 to 0.30 in the eastern part (Figure 2.5B, Supplementary Table S5).

Past Century (1850–1950):

The correlation coefficients between NAO and winter precipitation are significant in five datasets, which range from -0.21 to -0.66 in the western part (Morocco coast), -0.46 in northern Algeria (only one

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dataset), and -0.27 to 0.36 in northern Tunisia with more datasets than in the recent period (6 datasets) (Figure 2.4A, Supplementary Table S2).

As in the recent period, the MO shows a similar atmospheric pattern to the NAO in the western part of the study area (Morocco) with significant negative correlations varying from -0.35 to -0.59 and -0.24 to -0.54 in northern Algeria, becoming positive in northern Tunisia in five datasets (more than for the recent period), with values ranging from 0.26 to 0.48 .

The correlation of the WeMO with precipitation datasets shows a similar atmospheric pattern in the western part (Morocco coast) as in the recent period, but differs in the central and eastern parts of North Africa, with more significant correlations alternating between negative and positive values (discrepancies between datasets) compared to the recent period (Figure 2.5A, Supplementary Table S3).

Correlation coefficients between U850 and winter precipitation are similar for the recent period in most of the precipitation datasets, with high positive values ranging from 0.18 to 0.71 on the Morocco coast, 0.17 to 0.68 in northern Algeria, and 0.11 to 0.27 in northern Tunisia.

Past Millennium (850–1850):

The correlation of the NAO with the precipitation datasets in the western part (coast of Morocco) are similar to the periods previously mentioned. All correlations are significant except CESM1 model data. They vary from -0.19 to -0.46 on the Morocco coast, -0.10 to 0.13 in northern Algeria, and 0.16 to 0.31 in Tunisia. The correlations are less significant going from west to east (Figure 2.4A, Supplementary Table S2).

For the MO and WeMO, the correlations are significant for most datasets in all parts of North Africa. MO correlations vary between negative -0.12 and positive 0.36 values going from west to east. In contrast, the correlation between the WeMO and precipitation presents high variability between positive and negative values in the three parts of

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North Africa (Figures 2.4B and 2.5A, Supplementary Tables S3 and S4).

As for the past century and the recent period, the U850 correlations are highly positive and significant in the majority of datasets and the three parts of North Africa. The correlations generally range from 0.13 to 0.73 (Figure 2.5B, Supplementary Table S5).

These results show that the link between winter precipitation variability and the three atmospheric patterns (NAO, WeMO, and MO) over the past millennium is stronger and more stationary in the western region of North Africa than in the central and eastern parts, where it is overall weak. This result appears robust over time with no large changes observed for the three studied periods, except for the WeMO. Compared to the recent period, the MO index has a quite similar pattern to that during the past century and past millennium, as seen for NAO. On the contrary, WeMO shows a different pattern during the past century and the past millennium over the whole region. U850 is characterized by a stronger positive link with winter precipitation than the other atmospheric patterns for all study periods considered.

This is consistent with [40], which analyzed CMIP5 climate model projections (2070–2099) to quantify the role of atmospheric circulation in the Mediterranean precipitation response to climate change in the cold season (November–April) using the RCP8.5 climate scenario. They found that the zonal wind U850 is strongly associated with area-averaged Mediterranean precipitation in North Africa and is a useful diagnostic to explain the precipitation change.

2.3.2 Precipitation, Wind, and PDSI Analysis over the Last Millennium in the North African Region

According to the modified Mann–Kendall test [111], a significant decreasing trend is found in the simulated (November–April) precipitation between 1800 and 2005 in the three North African regions (except for

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BCC-CSM1 and CESM1 models in the western part) (Table 2.3 and Figure 2.6). No trend was found in the precipitation model results for the period 850–1800 (Figure 2.6). In order to detect the change point corresponding to the year when these significant trends started, we applied Pettitt's test [112]. The results showed a change point around 1800–1850 for the PMIP3/CMIP5 model simulations in the three regions. Furthermore, a decreasing trend is found in the observations and reanalysis starting around 1950 depending on the region (Table 2.4 and Figure 2.6). The magnitude of trends in the annual mean was determined using Sen's slope estimator. Sen's slope estimator indicates a higher magnitude of the decreasing trend in the observations and reanalysis (from -0.001 to -0.008 mm/day/year) than in the CMIP5 data (from -0.0002 to -0.001 mm/day/year). This is consistent with the analysis of [141], which documented a decline in annual precipitation in the Mediterranean region during the 20th century (1901–2009) in the Coupled Model Intercomparison Project phase 3 (CMIP3) simulations at a rate of -0.007 mm/day/decade. The range of slope values is higher in the models than in the observations and reanalysis. Furthermore, for the same CMIP5 model precipitation data, there are differences in the magnitude of trends between the three parts of North Africa. This confirms the difference in variability of precipitation over the study period in the three parts (West, Center and East) of North Africa.

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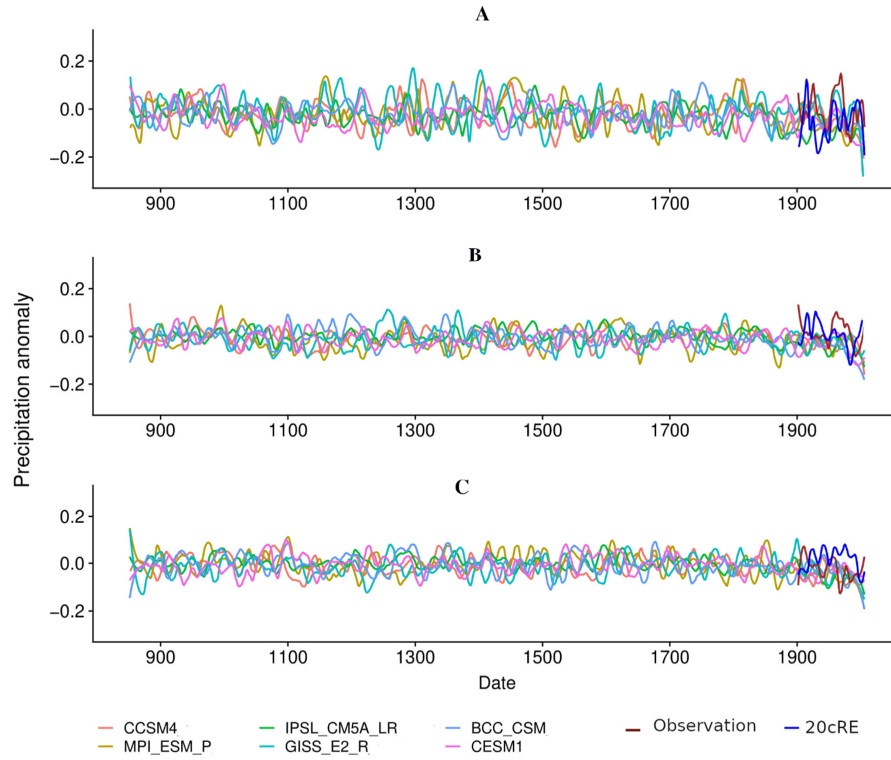


Figure 2.6: Loess-smoothed (30-year window) time series of monthly averaged precipitation anomalies (mm/day) for (A) western, (B) central, and (C) eastern parts of North Africa over the past millennium for PMIP3/CMIP5 model simulations and over the 20th century for observations (in brown) and reanalysis (in blue).

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Period 1800-2005			
Region	Time Series	P-Value	Slope (mm/day/year)
Western	CCSM4	0.01	-0.0006
	MPI-ESM-P	0.003	-0.001
	IPSL-CM5A-LR	0.0008	-0.0006
	GISS-E2-R	0.04	-0.0004
	BCC-CSM1	0.1	
	CESM1	0.1	
Central	CCSM4	0.02	-0.0002
	MPI-ESM-P	0.007	-0.0006
	IPSL-CM5A-LR	0.0002	-0.0004
	GISS-E2-R	0.2	
	BCC-CSM1	0.001	-0.0004
	CESM1	0.003	-0.0005
Eastern	CCSM4	0.04	-0.0002
	MPI-ESM-P	0.002	-0.0008
	IPSL-CM5A-LR	0.02	-0.0005
	GISS-E2-R	0.007	-0.0009
	BCC-CSM1	0.003	-0.0006
	CESM1	0.01	-0.0004

Table 2.3: Results of modified Mann–Kendall test for precipitation trend detection for six Palaeoclimate Modeling Intercomparison Project (PMIP3)/CMIP5 models over 1800–2005 in the three parts of North Africa. Significant test results (p -value ≤ 0.05) shown in red.

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1950–2005			
Region	Time series	P-value	Slope (mm/day/year)
Western	Observation	0.017	-0.008
	Reanalysis	0.003	-0.001
Central	Observation	0.015	-0.003
	Reanalysis	0.003	-0.007
Eastern	Observation	0.02	-0.001
	Reanalysis	0.04	-0.004

Table 2.4: Results of modified Mann–Kendall test for precipitation trend detection for observations and reanalysis over 1950–2005 in three parts of North Africa.

Note: The red color in the P-value column is to highlight in which dataset the modified Mann-Kendall test for trend detection is significant with a p -value < 0.05 .

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Winter U850 wind speed over the last millennium is characterized by a significant decreasing trend in most PMIP3/CMIP5 simulations (CCSM4, MPI-ESM-P, BCC-CSM1, and CESM1) (Table 2.5, Figure 2.7), as identified by the modified Mann–Kendall test. These decreasing trends start around 1800, which is consistent with the decreasing trend observed in winter precipitation shown in the six PMIP3/CMIP5 simulations (Table 2.3, Figure 2.6).

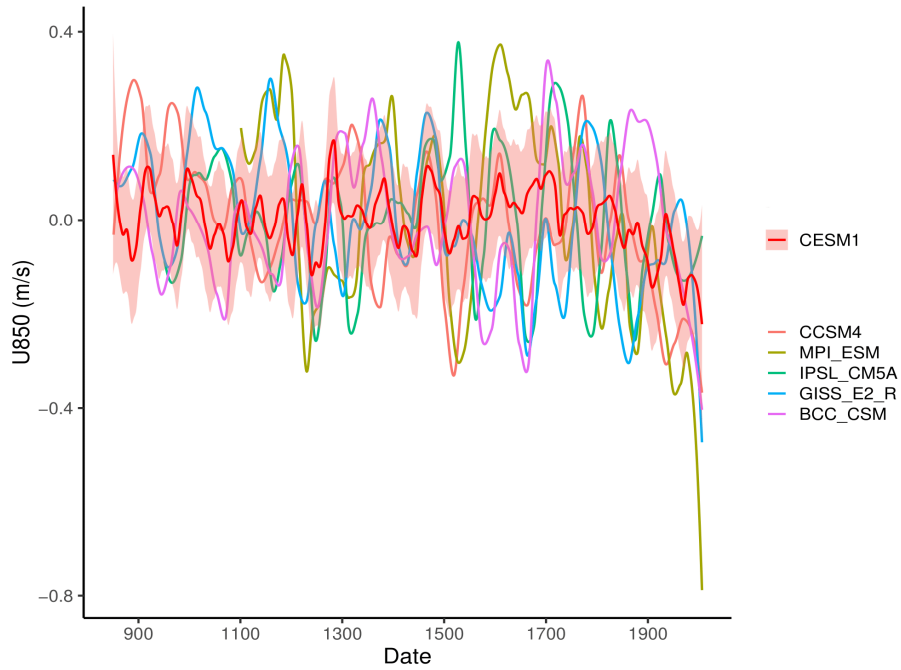


Figure 2.7: Loess-smoothed time series (30-year window) of simulated U850 (PMIP3/CMIP5 models) anomalies over 850–2005. Red line represents ensemble mean for 10 CESM1 simulations and shaded red corresponds to standard deviation of the ensemble.

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Period	1800–2005		
Location	Model Time Series	P-Value	Slope (m/s/yr)
North Africa (25–33°N, 10°W–23°E)	CCSM4	0.04	-0.001
	MPI-ESM-P	0.01	-0.001
	IPSL-CM5A-LR	0.8	
	GISS-E2-R	0.5	
	BCC-CSM1	0.03	-0.001
	CESM1 (ensemble mean)	0.01	-0.001

Table 2.5: Results of modified Mann–Kendall test of U850 trend detection for six PMIP3/CMIP5 model simulations over 850–1800 and 1800–2005 in North Africa.

Note: The red color in the P-value column is to highlight in which dataset the modified Mann–Kendall test for trend detection is significant with a p -value < 0.05

This decrease in precipitation could be explained by a higher pressure gradient across the North Atlantic channeling strong westerlies to central and Northern Europe, with little moisture reaching the Mediterranean basin and North Africa [26, 33]. This atmospheric pattern appears to have been more predominant during the 20th century, likely due to anthropogenic forcing [142]. However, this hypothesis is still debated and no formal attribution to human activities has been achieved yet [143].

Both results shown in Figures 2.6 and 2.7 are in accordance with [40], which demonstrated that the departure of precipitation in the Mediterranean wet season in the future (2070–2099) is linearly related to 850 hPa wind. The findings of [40] for the future also appear to be valid for the 20th century and the last millennium in the simulations.

It is not possible to compare the decreasing trend with observed precipitation before the 20th century, but time series of the summer (June–August)

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PDSI are available over the last millennium in Morocco and Algeria (Figure 2.8). These locations correspond to those of the tree ring chronologies in the Mediterranean region used for the OWDA (Supplementary Figure S1) [109].

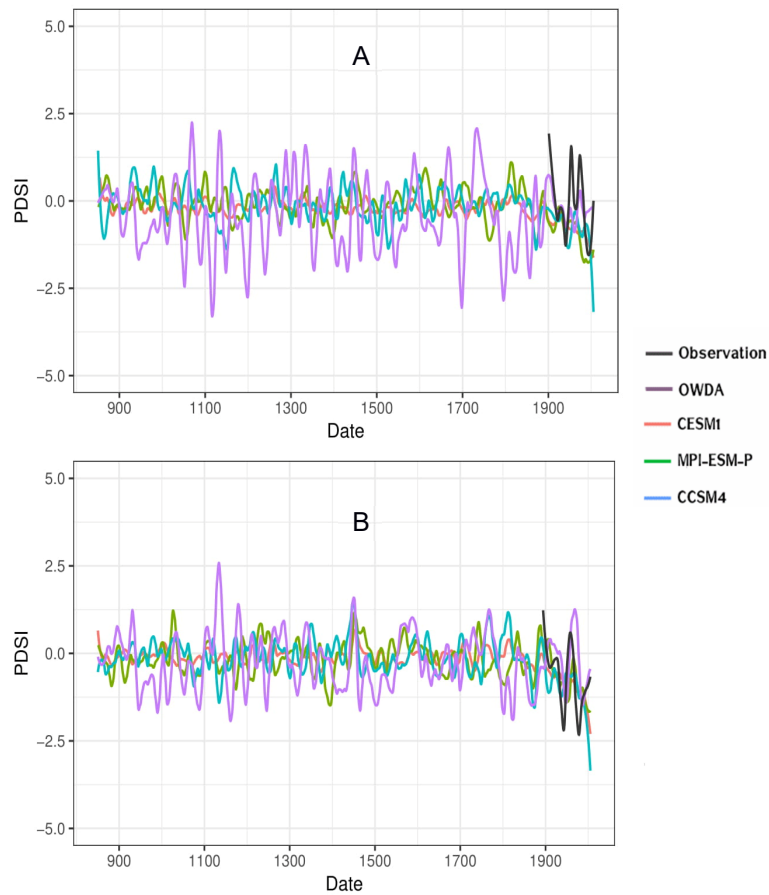


Figure 2.8: Loess-smoothed time series (30-year window) of summer (June, July, August (JJA)) PDSI in (A) Morocco and (B) Algeria in 850–2005 for CCSM4, MPI-ESM, and ensemble mean of CESM1 (mean of 10 members), and reconstructed and instrumental dataset (observation) from OWDA.

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Model results, proxy-based reconstruction, and observations consistently show a significant decreasing trend of the PDSI over the last century (Figure 2.8, Table 2.6), according to the modified Mann–Kendall test [111], while no trend was found during the last millennium (850–1800). The trend started at the end of the 19th century for both PMIP3/CMIP5 models and reconstructed datasets. The instrumental data also depict a trend starting in the beginning of the 20th century (Table 2.7), which is in agreement with [144], which showed a clear trend toward drier conditions during the 20th century in most western and central Mediterranean regions. The decreasing trend in the instrument-based PDSI is larger than the one computed from the CMIP5/PMIP3 model output or OWDA reconstruction (Table 2.7).

Period	1850–200		
Location	PDSI Time Series	P-Value	Slope (m/s/yr)
Algeria	CCSM4	0.007	-0.006
	MPI-ESM	0.002	-0.008
	CESM1 (mean of 10 members)	1.9×10^{-07}	-0.006
	Reconstructed OWDA	0.002	-0.006
Morocco	CCSM4	0.001	-0.008
	MPI-ESM	0.0004	-0.009
	CESM1 (mean of 10 members)	1.2×10^{-06}	-0.008
	Reconstructed OWDA	0.003	-0.005

Table 2.6: Results of modified Mann–Kendall test of trend detection applied to Palmer Drought Severity Index PDSI simulated by CMIP5 models, reconstructed from Old World Drought Atlas (OWDA) over 1850–2005 in Morocco and Algeria.

Note: The red color in the P-value column is to highlight in which dataset the modified Mann-Kendall test for trend detection is significant with a p -value < 0.05 .)

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Period	1901-2005		
Location	PDSI Time series	P-value	Slope
Algeria	Instrumental data	0.004	-0.01
Morocco	Instrumental data	0.005	-0.01

Table 2.7: Results of modified Mann–Kendall test of trend detection applied to observed PDSI over 1901–2005 in Morocco and Algeria.

Note: The red color in the P-value column is to highlight in which dataset the modified Mann-Kendall test for trend detection is significant with a p -value < 0.05 .)

The results show weak correlation coefficients between observed and simulated PDSI as well as between the PDSI simulated by the various models (Supplementary Table S1), which leads us to conclude that natural climate variability dominated the precipitation changes in the North African region over the last millennium, as observed in several other regions for the same period, e.g., [145, 146]. This is confirmed by the wide range of ensemble simulations performed with CESM1 over that period.

2.4 Conclusions

Our analysis of the links between North African winter precipitation changes and the atmospheric patterns NAO, WeMO, MO, and U850 over the last millennium is based on recent observational data, 20th century reanalysis, and the results of six PMIP3/CMIP5 models. Specifically, we analyze the winter (November–April) precipitation variability observed during three periods (recent, 1979–2005; last century, 1850–2005; and last millennium, 850–2005) in western, central, and eastern regions of North Africa. Furthermore, we compared the results with reconstructed PDSI.

The link between NAO, WeMO, and MO atmospheric circulation indices and rainfall variability over the three periods of the last millennium is

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stronger and more stationary in the western region of North Africa than in the central and eastern parts, where it is overall weak. The NAO and MO are more likely to affect precipitation variability in the western part of North Africa (Morocco coast and West Algeria) associated with highly negatively significant correlation values, compared to the eastern part with weak positive correlation results. Additionally, the link between WeMO and precipitation is spatially and temporally not stationary over the period investigated, as it depends on the length of the time series analyzed and the geographic location.

The correlation coefficients obtained between U850 and winter precipitation in the three parts of North Africa over the last millennium are stronger than those calculated with NAO, WeMO, and MO. The U850 correlation coefficients are uniformly distributed throughout the whole region, with significant high positive values for the different data sources. The link between winter U850 and precipitation variability is consequently more robust over the last millennium and well describes the interannual changes of precipitation in North Africa. The weak correlation coefficients of PDSI in the climate model simulations over the last millennium, notably between members of the same climate model (CESM1), but also between simulated and reconstructed PDSI, lead us to conclude that internal climate variability dominated the precipitation changes in North Africa over this period. In addition, the variability of the long-term PMIP3/CMIP5 winter precipitation, U850 time series and PDSI (simulated and reconstructed from OWDA) presents a significant decreasing trend toward drier conditions over the recent past. This negative trend started in the beginning of the 19th century, except for the instrumental PDSI dataset, for which the trend started in the mid-20th century.

In conclusion, we find that precipitation variability associated with U850 represents an important contribution to our understanding of potential causes of the long-term hydroclimate variability in North Africa during winter. These results will benefit national governments and private

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institutions in the planning and management of water resources in North Africa.

CHAPTER 3

The role of atmospheric circulation in precipitation decline in North Africa in present and future climate changes

3.1 Introduction

As stated in the general introduction, the goal of this chapter is to better understand the causes of droughts and ongoing hydroclimate aridification in North Africa region. The aim is to better understand and quantify the role of atmospheric circulation in precipitation response to the present climate as well as in future climate projections. [57] have applied an atmospheric moisture budget (hereafter MB) decomposition in order to explain the physical mechanisms by which the mean and transient circulation drive winter precipitation changes across the Mediterranean region. Based on ERA-Interim and CMIP5 model data results, they showed that over the Mediterranean land regions, the difference of Precipitation minus Evapotranspiration is positive in winter and negative in summer. They argued that in winter over land, P-E is sustained by transient eddy moisture convergence and opposed by mean flow moisture divergence. [57] have pointed out that the increase in the moisture divergence term in the mean flow under the future CMIP5

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climate projections, is the cause of drying over the Mediterranean region. This key role of the mean flow moisture divergence suggest that the circulation changes are a dominant contributor to future hydrological change in Mediterranean regions. [40] argued that Mediterranean precipitation change can not be exclusively explained by changes in thermodynamics, “wet get wetter, dry get drier” argument. This argument results in rising atmospheric moisture and transport in a warming climate, that drive a wetting of the tropics and extra-tropics and a drying of the subtropics [62].

Both westerly zonal wind and winter precipitation in North Africa present a decreasing trend over the last century [147]. As a result, Mediterranean region has received a great attention, but further investigation remains to be done to better explain winter precipitation decrease over this region. In this study, we will analyze whether the dominant processes responsible for precipitation changes in Mediterranean are also dominant in North Africa or if there are differences between these regions? We divide the region of North Africa into three sub-regions, West (10W-2W, 32N-35N), Center(2W-8E, 32N-36N) and East (8E-20E, 32N-36N), as in Chapter 3 (Figure 3.1), to understand better how the changes in circulation (anticyclone), U850 and the role of divergence in the future, are represented over each part. The outline of this chapter is as follows: Section 1 describes the atmospheric moisture decomposition on ERA-Interim reanalysis, CESM1-CAMS5 and CESM2 models over the Mediterranean basin, during the present (1990-2005) and in future (2071-2080) climate scenario (RCP85 and SSP585). We also show the change between these two periods to identify the contribution of each moisture budget terms (evaporation, mean flow, storm tracks and the role of moisture convergence and divergence) to North Africa precipitation variability. Section 2 discusses the correlation of each calculated MB terms with the atmospheric circulation U850 index, since it have been argued that this index can be linked to precipitation trends under the future climate warming. This will enable us to evaluate the atmospheric

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circulation dynamics within the MB terms that may contribute to the precipitation decline over North Africa region under the future climate change.

3.2 Data and Methodology

3.2.1 Contribution of the moisture budget terms in precipitation variability in North Africa in present and future conditions

To better understand the main drivers of precipitation change in North Africa for present conditions and under the future climate warming, we have performed a moisture budget analysis using the same methodology as in [57]. To compute the MB terms, we break down the atmospheric moisture budget (MB) equation into two main components, the dynamic and thermodynamic terms. The MB scripts rely on four main model inputs; zonal and meridional wind speed (ua , va), specific humidity (hus) at different vertical pressure levels and the surface pressure (p_s). From the analysis, we obtain five moisture budget terms that will be described in section 3.2.3 (simulated P-E, transient eddy-moisture convergence, mean flow convergence term and its components: divergence and convergence in the mean flow). A land-sea mask is applied on ERA-Interim reanalysis as well as CESM1-CAMS5 and CESM2 model data, to compute the averaged MB terms in the grid cells marked as land, over the three sub-regions. We apply the MB decomposition method to CESM1-CAMS5 and CESM2 models as well as to the ERA-Interim reanalysis dataset (the only datasets that include the required parameters for the MB calculation scripts at the time of this analysis) as described below:

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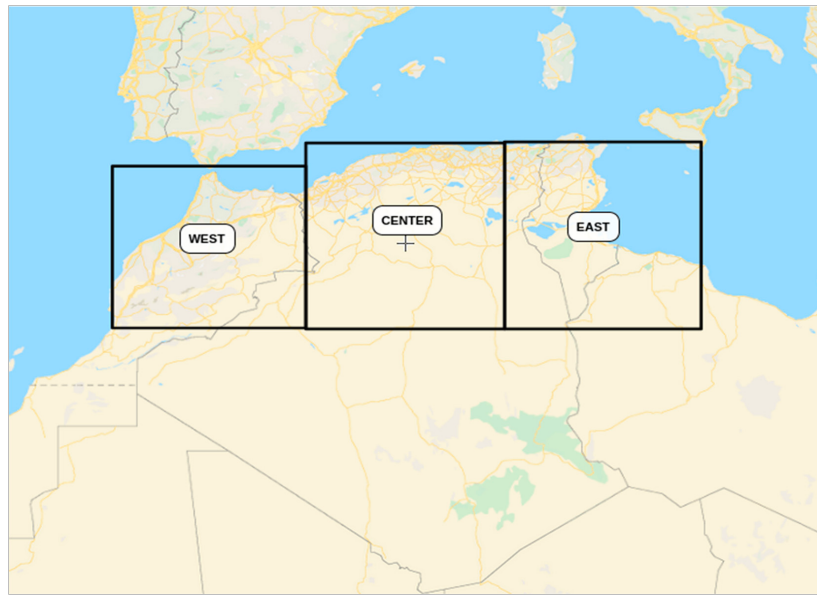


Figure 3.1: North Africa domain study. The black boxes represent the three sub-regions of North Africa (same figure as in Chapter 2).

3.2.2 Data

Reanalyses dataset

The ERA-Interim data set [148] is a reanalysis of the global atmosphere covering the period since 1979 and proceeds in real time. Updates of the ERA-Interim archive is done monthly. ERA-Interim includes several improvements, such as a refined data assimilation scheme (four-dimensional variational assimilation, 4D-Var, with variational bias correction method) and improved forecast model, as compared to its previous ECMWF reanalysis, the 40-yr ECMWF Re-Analysis (ERA-40). A large amount of observations have been assimilated, after initial, background and redundancy quality controls and also with time the number of observation assimilated increases. It has a greatly improved

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representation of the hydrological cycle comparative to ERA-40, which makes it suitable for our purpose. ERA-interim was prepared using cycle 31r2 of the Integrated Forecasting System (IFS – CY31R2, 2006), with 60 vertical levels, with the upper level at 0.01 hpa. Atmospheric variables are available on these levels and they are also interpolated to 37 pressure, 16 potential temperature and 1 potential vorticity level(s). Surface data are also use here, including 2D parameters e.g. precipitation, 2m temperature, top of atmosphere radiation and vertical integrals over the whole atmosphere [148]. The moisture budget calculations used here were performed on data that were archived by ECMWF on a regular 0.75° grid with 37 model levels and at 6 hourly resolution as in [57], covering the period (1990-2005).

CESM data

CESM designates the current generation of the coupled climate/Earth system models developed at the National Center for Atmospheric Research (NCAR), universities and other research institutions, as a collaborative effort between different groups of scientists, software engineers and students. The Community Earth System Model version 1 (CESM1) that includes the Community Atmospheric Model version 5 (CAM5), referred as CESM1-CAM5 model [149], is a model from phase 5 of the Coupled Model Intercomparison Project (CMIP5, [150]). The CESM1-CAM5 model has the identical land, ocean and sea ice constituents as in CCSM4 [151], with the main change included in the atmosphere. Characteristics of the model formulation are presented by [149]. The CAM5 is an atmospheric model with improved and more realistic formulations of radiation, boundary layer and aerosols [152]. The aerosol scheme [153] in particular is prognostic and the cloud microphysics [154] considers both the direct and indirect effect for sulfate black and organic carbon. Therefore, the CESM1-CAM5 displays more realism and complexity and makes an improved simulation of twentieth-century climate [149]. The

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CESM1-CAMS5 model is forced by RCP8.5 scenario. The Representative Concentration Pathways (RCPs) describe a series of different greenhouse gas concentration and emission targets, which aim's to assist research into the effects and possible responses to climate change [155]. The representative concentration path RCP8.5 fits to a target with high greenhouse gas emissions in comparison to other scenarios [156] and therefore the greenhouse gas emissions and concentrations in this scenario increase substantially during time, which drives to a radiative forcing of 8.5 W/m^2 at the end of the century.

The Community Earth System Model Version 2 (CESM2) builds on numerous successes of its predecessors. CESM2 was made available to the community in June 2018 (accessible at www.cesm.ucar.edu/models/cesm2/). Simulations performed with this model have contributed to Phase 6 of the CMIP (CMIP6; [157]), using the nominal horizontal resolution setting of 1° . CESM2 participates in approximately 20 Projects Model Intercomparison (MIP) within CMIP6, including ScenarioMIP, which calls for forward-looking simulations driven by plausible alternative futures of emissions and land use [155]. CMIP6 climate projections are driven by a new set of emissions and land use scenarios [155] made with integrated assessment models (IAMs) based on new future pathways of societal development, the Shared Socio-economic Pathways (SSPs), and related to the RCPs. The SSP585 scenario corresponds to the high end of plausible future forcing pathways and is comparable to the RCP8.5 [158, 157, 155]. CESM1-CAMS5 and CESM2 are the models selected in this study, because they are the only ones at the time of this analysis that include the required parameters for the MB calculation scripts (as described in the section 3-2-3). Results are presented for simulations of the present period (1990-2005) and the representative concentration pathway CESM1-CAMS5 (RCP8.5) scenarios (2071-2080) and the equivalent CESM2 ssp585 scenario [159].

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3.2.3 MB decomposition methodology

The CESM1-CAM5 as well as CESM2 data archive provides model data on native hybrid vertical levels rather than the pressure vertical levels of the model. [160] showed that the use of pressure coordinates rather than the model coordinates on which the data are analyzed produces acceptably small differences. In this study, we evaluate the ability of both CESM1-CAMS5 and CESM2 models compared to ERA-Interim reanalysis, to simulate the MB changes during winter seasons (November to April) in present (1990-2005) and future climate change (2071-2080). The MB scripts used in this present study are based on the work of [57]. [161] pointed out the importance of the computation of the moisture budget at the highest vertical resolution possible on to reduce calculation biases. Prior to the computation of the MB budget terms, we interpolate the model native hybrid levels to pressure levels (32 pressure levels).

The script input data includes velocities (ua , va) and specific humidity (hus) on a three-dimensional spatial grid as well as surface pressure (ps). The data is saved at 6-hourly temporal resolutions. The moisture budget is computed within each of the ten CESM1-CAMS5 model member simulations, before computing the model ensemble mean. The moisture budget at equilibrium, in pressure coordinates is given by:

$$P - E = -\frac{1}{g\rho_w} \nabla \cdot \int_0^{ps} \mathbf{u}qdp \quad (3.1)$$

where P is precipitation, E is evaporation, g is gravity acceleration, ρ_w is water density, p is pressure and ps is the surface value, q is specific humidity and u is the horizontal vector velocity. The notation follows here and thereafter that of [162]. To calculate the vertical integral, we sum over pressure levels, so Eq.3.1 leads to

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$$P - E = -\frac{1}{g\rho_w} \nabla \cdot \sum_{k=1}^K \mathbf{u}_k q_k dp_k \quad (3.2)$$

where k refers to the vertical level and dp_k refers to the pressure thickness of each level of which the lowest level is p_s . The monthly budget is computed by separating all amounts into different parts: Monthly means are represented by 'overbars', 'primes' correspond to deviations from monthly means and 'double overbars' represent monthly climatological means. The Eq.3.2 can be rewritten as

$$\overline{\overline{P}} - \overline{\overline{E}} \approx -\frac{1}{g\rho_w} \nabla \cdot \sum_{k=1}^K \overline{(\overline{\mathbf{u}}_k \overline{q}_k + \overline{\mathbf{u}'_k q'_k})} dp_k \quad (3.3)$$

The moisture convergence by the mean flow is shown as the first term on the right part of the Eq.3.3 and the moisture convergence by the transient eddies is shown as the second term. As an approximation, the term involving dp_k are ignored here. This is an acceptable simplification as in [163]. Finally, we split the mean flow contribution into terms related to advection and mass divergence. The divergence operator is therefore taken within a vertical summation and this leads to an additional surface term :

$$\overline{\overline{P}} - \overline{\overline{E}} = -\frac{1}{g\rho_w} \left[\sum_{k=1}^K \overline{(\overline{\mathbf{u}}_k \cdot \nabla \overline{q}_k + \overline{q}_k \nabla \cdot \overline{\mathbf{u}}_k)} dp_k + \nabla \cdot \sum_{k=1}^K \overline{\mathbf{u}'_k q'_k} dp_k \right] - \frac{1}{g\rho_w} \overline{\overline{q}_s \mathbf{u}_s \cdot \nabla \mathbf{p}_s} \quad (3.4)$$

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The climatological change between 21st century and 20st century is represented by the use of :

$$\Delta(\cdot) = (\cdot)_{21} - (\cdot)_{20} \quad (3.5)$$

which can be written as: $(P-E)_{21} - (P-E)_{20} =$

$$\begin{aligned} \Delta \bar{\bar{P}} - \Delta \bar{\bar{E}} = & -\frac{1}{g\rho_w} \sum_{k=1}^K \Delta \left[\overline{(\bar{\mathbf{u}}_k \cdot \nabla \bar{q}_k) dp_k} \right] - \\ & \frac{1}{g\rho_w} \sum_{k=1}^K \Delta \left(\overline{\bar{q}_k \nabla \cdot \bar{\mathbf{u}}_k dp_k} \right) - \frac{1}{g\rho_w} \nabla \cdot \sum_{k=1}^K \Delta \left(\overline{\bar{\mathbf{u}}'_k \bar{q}'_k dp_k} \right) - \frac{1}{g\rho_w} \Delta \left(\overline{q_s \mathbf{u}_s \cdot \nabla p_s} \right) \end{aligned} \quad (3.6)$$

Variations within the first and second terms of Eq.3.6 may be due to changes of moisture or circulation, the former being a thermodynamic mechanism and the latter being a dynamic mechanism for hydrological changes [163]. Those mechanisms can be divided diagnostically, by assessing these terms holding first the circulation and second the humidity, based on their climatological amounts over the 20th century. As will be shown, the term related to the divergent flow [the second term in Eq.3.6] is critical. We can approximate it as

$$\begin{aligned} -\frac{1}{g\rho_w} \sum_{k=1}^K \Delta \left(\overline{\bar{q}_k \nabla \cdot \bar{\mathbf{u}}_k dp_k} \right) \approx & -\frac{1}{g\rho_w} \sum_{k=1}^K \Delta \left(\overline{\bar{q}_k dp_k} \right) \nabla \cdot \bar{\mathbf{u}}_{k,20} \\ & -\frac{1}{g\rho_w} \sum_{k=1}^K \bar{\bar{q}}_{k,20} \Delta \left(\overline{\nabla \cdot \bar{\mathbf{u}}_k dp_k} \right) \end{aligned} \quad (3.7)$$

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We get this approximation by ignoring terms quadratic in the Δ and covariance of deviations of the monthly mean values from the climatological monthly means and from using the values of the 20th century for dp_k . The first term on the right in the Eq.3.7, includes humidity changes when circulation is fixed and represents the thermodynamic term. The second term includes circulation changes when humidity is fixed and represents the term "mean circulation dynamics" [164].

3.2.4 Link between U850 index and the moisture budget terms

As shown before, the link found between U850 index and precipitation over North Africa is robust over the last millennium as well as under the future climate change [147, 40]. In order to understand better this relation, we analyze the link between U850 index and each term within the moisture budget in both ERA-interim and global climate model data. We calculated the correlation coefficients between U850 index and each terms in ERA-Interim MB, over the present period (1990-2005). We do the same computation with the CESM1-CAM5 model (individual simulations as well as the model ensemble mean) and CESM2 experiment, during the period (1990-2005) and future (2071-2080) climate changes.

3.3 Results and discussion

First, we evaluate how CESM1-CAMS5 as well as CESM2 models represent the annual precipitation cycle in North Africa compared to ERA-Interim reanalysis (Figure 3.2). The result shows that both models reproduce well the precipitation annual cycle in North Africa (0E-8E, 34N-36N) in comparison to ERA-Interim reanalysis. However, it's worth to note, that CESM2 overestimate the winter season precipitation amounts whereas CESM1-CAMS5 underestimate it, compared to ERA-Interim reanalysis. The precipitation means vary from 0.79 mm/day in CESM2, 0.43 mm/day in CESM1-CAMS5 and 0.64 mm/day in ERA-

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Interim. Compared to summer season, precipitation variability between these three data set are small, changing between 0.12 mm/day to 0.19 mm/day.

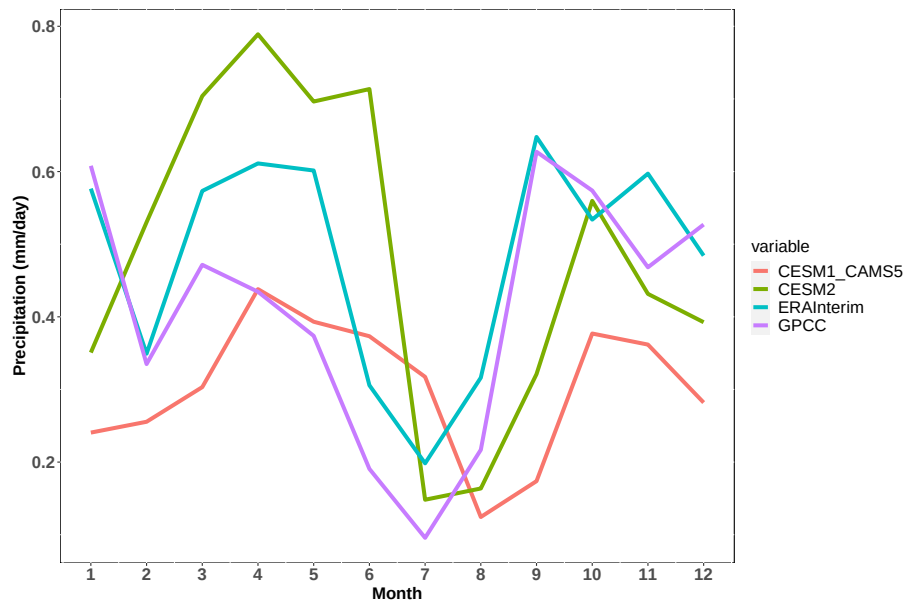


Figure 3.2: Precipitation annual changes in ERA-Interim reanalysis, GPCC station data, CESM1-CAMS5 and CESM2 climate models, in North Africa (0E-8E, 34N-36N) over 1990-2005.

3.3.1 Contribution of the moisture budget to precipitation changes in North Africa in the present and future climate

North Africa moisture budget decomposition in ERA-Interim, CESM1-CAMS5 and CESM2 models over winter 1990-2005

Figures 3.3 to 3.5 show ERA-Interim reanalysis, CESM1-CAMS5 and CESM2 model MB decomposition terms, over the winter (November-April) 1990-2005. In comparison to Northern Europe where P-E is strongly positive, most of the coastal regions of North Africa as well as

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across the Mediterranean land regions have small P-E values (Figure 3.3, c and d). Table 3.1 shows the winter (November-April) mean MB terms, averaged over three North Africa sub-regions (West: 10W-2W, 32N-35N, Center: 2W-8E, 32N-36N and East: 8E-11E, 32N-36N) as in [147]. ERA-Interim winter (November-April) precipitation mean (P), varies from 0.90 mm/day in the western part to 0.53 mm/day in central part and 0.44 mm/day in eastern part. We note that in the three data sets (ERA-Interim reanalysis, CESM1-CAMS5 and CESM2 models), winter precipitation means decrease from the West to the East of North Africa. P-E calculated from ERA-Interim and climate models, is negative except in CESM2 model, where the values are positive. P-E simulated from scripts is also negative in both ERA-Interim and climate models (CESM1-CAMS5 and CESM2). It is worth to note that both P-E calculated from reanalysis (models) and the one simulated from MB scripts are higher over the western part of North Africa in all data sets.

The differences between P-E calculated from reanalysis (models) data and the one simulated from scripts are small, less than 0.10 mm/day. This means that the used scripts reproduce well the MB terms. The winter mean contribution of transient eddies is positive all over the three North Africa sub-regions and in both reanalysis and models data (Table 3.1). The contribution of transient eddies changes between 0.17 to 0.48 mm/day in the different data sets. In contrast, the contribution of the mean flow, shows a negative values, over the three sub-regions in all data sets (Table 3.1), with high values over the West decreasing going to the East in all data sets. Its values vary between -0.29 mm/day to -0.73 mm/day. The divergence term in mean flow shows also negative values varying from -0.48 to -1.08 mm/day over the three parts of North Africa, in all data set. By contrast, the contribution of advection term in the mean flow, is generally positive with values varying between 0.10 to 0.55 mm/day, except over the western part in the three data sets. Overall, the contribution of the advection term is more important in CESM1-CAMS5 and CESM2 climate models than are the ERA-Interim reanalysis, except

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over the western part in both models. The results show that the transient eddies flux tends to drive moisture convergence (origin of precipitation) across the three sub-region of North Africa, whereas the mean flow divergence term tends to transport moisture away from the region. Our results agree with [57], who showed that over Mediterranean, transient eddies contributes to maintain positive P-E in the ERA-Interim and CMIP5 models, so it is expected that potential future drying may be linked to a weakening of the transient eddy moisture convergence.

MB terms (mm/day) / sub-region	ERA-Interim			CESM1-CAMS5			CESM2		
	West	Center	East	West	Center	East	West	Center	East
P	0.90	0.53	0.44	0.69	0.32	0.37	0.93	0.68	0.47
P-E (model)	-0.47	-0.13	-0.16	-0.35	-0.12	-0.28	0.18	0.16	0.14
P-E (script)	-0.41	-0.16	-0.12	-0.29	-0.16	-0.26	-0.15	-0.10	-0.12
Transient Eddies	0.32	0.24	0.17	0.41	0.26	0.32	0.48	0.23	0.20
Mean Flow	-0.73	-0.40	-0.29	-0.70	-0.42	-0.58	-0.63	-0.33	-0.32
Divergence in Mean Flow	-0.51	-0.50	-0.48	-0.56	-0.95	-1.08	-0.53	-0.80	-0.87
Advection by Mean Flow	-0.22	0.10	0.19	-0.14	0.53	0.50	-0.10	0.47	0.55

Table 3.1: Winter (November-April) 1990-2005 MB means, averaged over three sub-regions of North Africa: (West: 10W-2W, 32N-35N, Center: 2W-8E, 32N-36N and East: 8E-11E, 32N-36N).

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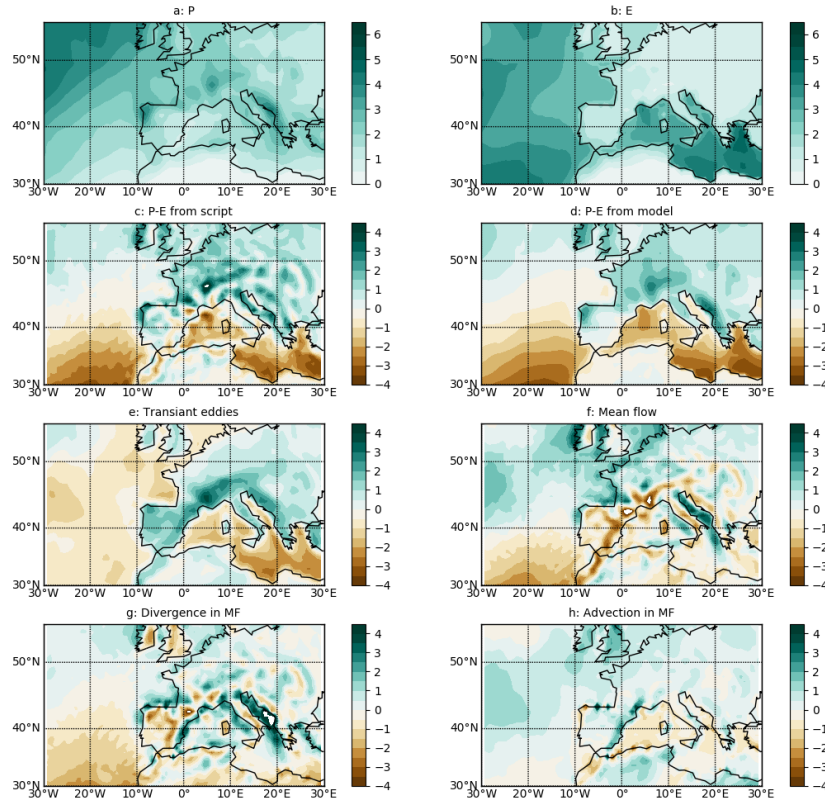


Figure 3.3: The November-April (1950-2005) moisture budget from ERA-Interim: (a) P, (b) E, and (c) P-E from scripts (d) P-E from ERA-Interim (e) transient eddy moisture convergence (f) The moisture convergence by the mean flow (MF) and its components due to (g) mass divergence and (h) advection. Units are mm/day.

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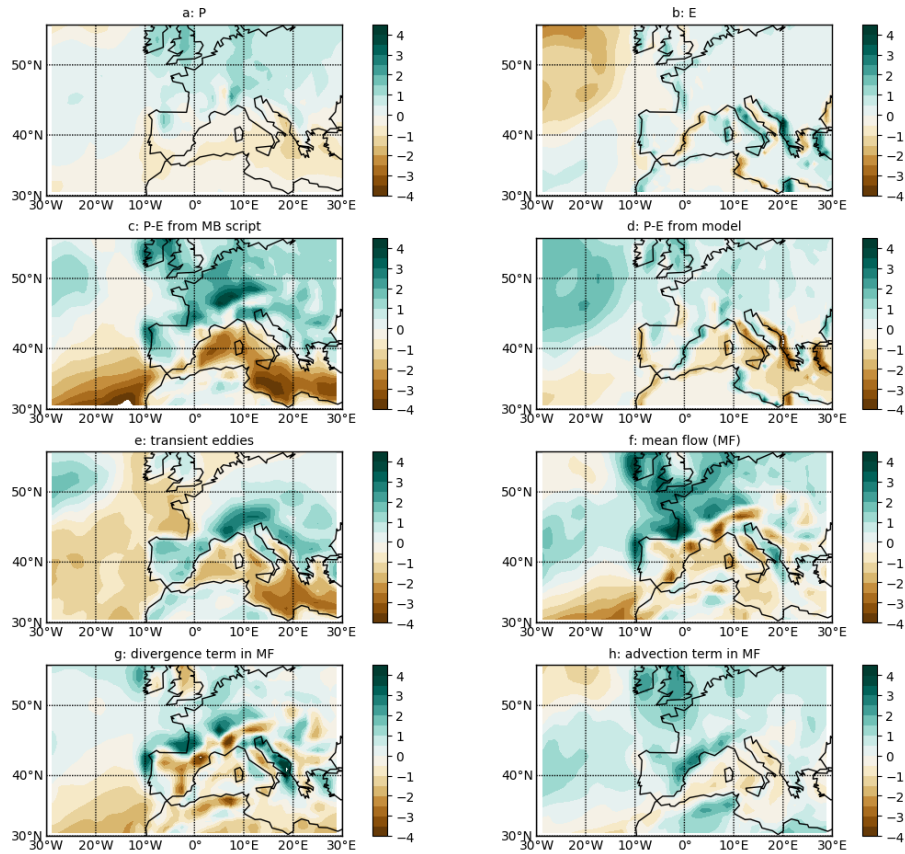


Figure 3.4: The November–April moisture budget from CESM1-CAM5 model (r1i1p1 experiment), over 1990–2005: (a) P, (b) E, and (c) P-E from MB script (d) P-E from model (e) transient eddy moisture convergence, (f) the moisture convergence by the mean flow (MF) and its components due to (g) mass divergence and (h) advection. Units are mm/day.

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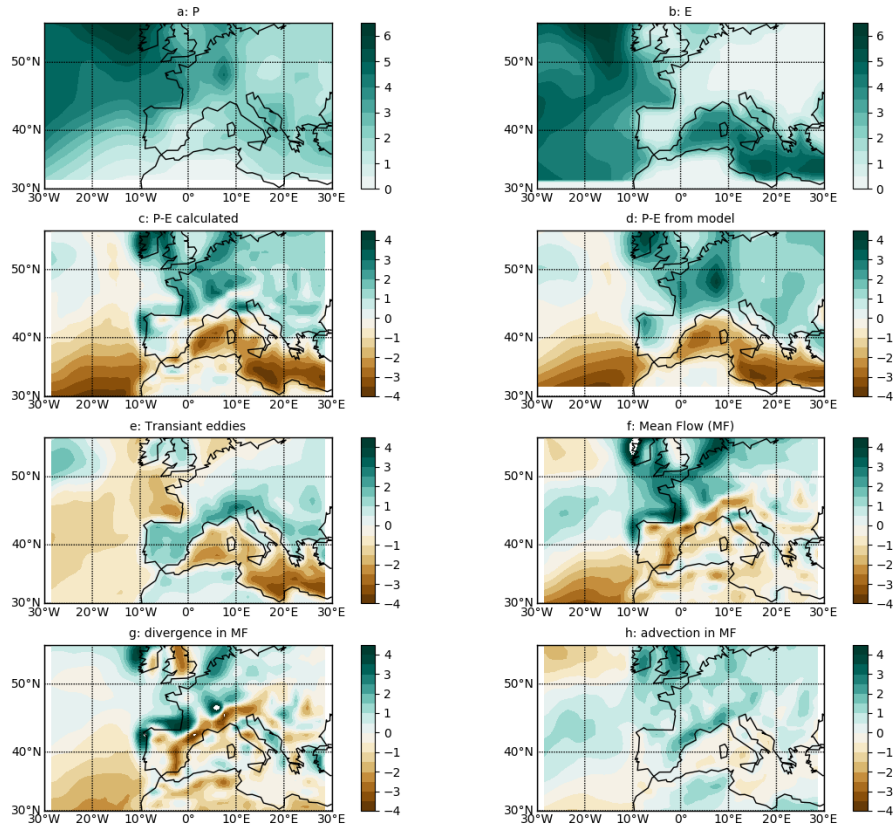


Figure 3.5: The November–April moisture budget from CESM2 model over 1990–2005: (a) P, (b) E, and (c) P-E from script, (d) P-E from model, (e) transient eddy moisture convergence, (f) the moisture convergence by the mean flow (MF) and its components due to (g) mass divergence and (h) advection. Units are mm/day.

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Model-projected precipitation changes in North Africa, between winter (2071-2080–1990-2005)

Figures 3.6 and 3.7 show climatological changes in CESM1-CAMS5 and CESM2 models precipitation and moisture budget decomposed terms between the two winter (November-April) periods (2071-2080–1990-2005). The North Africa decline in winter (November-April) mean precipitation over these two periods, varies from the West to the East, between -0.40 mm/day to -0.13 mm/day in CESM1-CAMS5 model and between -0.41 mm/day to -0.13 mm/day in CESM2 model. At the same time, both winter P-E simulated by the scripts and the one calculated from models are projected to decrease, all over the North Africa sub-regions from -0.09 mm/day to -0.37 mm/day in both cases (Figure 3.6, c and d and Table 3.2). It is important to note that the differences between the P-E simulated by MB scripts and the one calculated from models are small, less than 0.10 mm/day in all data sets.

The change in transient eddy moisture (Figure 3.6e and 3.7e, Table 3.2) is positive over the three sub-regions (West-Center-East), varying between 0.07 mm/day to 0.24 mm/day in both models except over the western part in CESM1-CAMS5 model. The difference in the mean flow convergence for the two climate models, CESM1-CAMS5 and CESM2 is negative, varying from -0.20 mm/day to -0.44 mm/day, over the three sub-regions. The changes in the divergence term in mean flow (Figure 3.6g, Figure 3.7g and Table 3.2) is also negative in all data sets and in all sub-regions of North Africa, varying from -0.24 mm/day to -1.08 mm/day (Table 3.1). The change in advection term in the mean flow is positive over the three sub-regions, except over the western part for CESM2 models with negative value (Table 3.1). The CESM1-CAMS5 ensemble mean (6.3 in Supplementary materials) gives similar MB decomposition results (not shown here) as in the present experiment (r1i1p1) used in this study. It is important to note that the same MB pattern as in Figure 3.7 is observed using 30 years of CESM2 model data, with small variations in

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the different terms (Figure 6.6 in Supplementary materials).

MB terms (mm/day) / sub-region	CESM1-CAMS5			CESM2		
	West	Center	East	West	Center	East
P	-0.40	-0.25	-0.13	-0.41	-0.17	-0.13
P-E (model)	-0.28	-0.25	-0.27	-0.13	-0.10	-0.11
P-E (script)	-0.37	-0.33	-0.30	-0.10	-0.05	-0.09
Transient Eddies	-0.03	0.11	0.07	0.24	0.15	0.16
Mean Flow	-0.34	-0.44	-0.37	-0.34	-0.20	-0.25
Divergence in Mean Flow	-0.73	-0.68	-0.65	-0.24	-1.01	-1.08
Advection in Mean Flow	0.39	0.24	0.28	-0.10	0.81	0.83

Table 3.2: Changes between 2071-2080 – 1990-2005 in MB winter (November-April) means, averaged over three sub-regions of North Africa: (West: 10W-2W, 32N-35N, Center: 2W-8E, 32N-36N and East: 8E-11E, 32N-36N)

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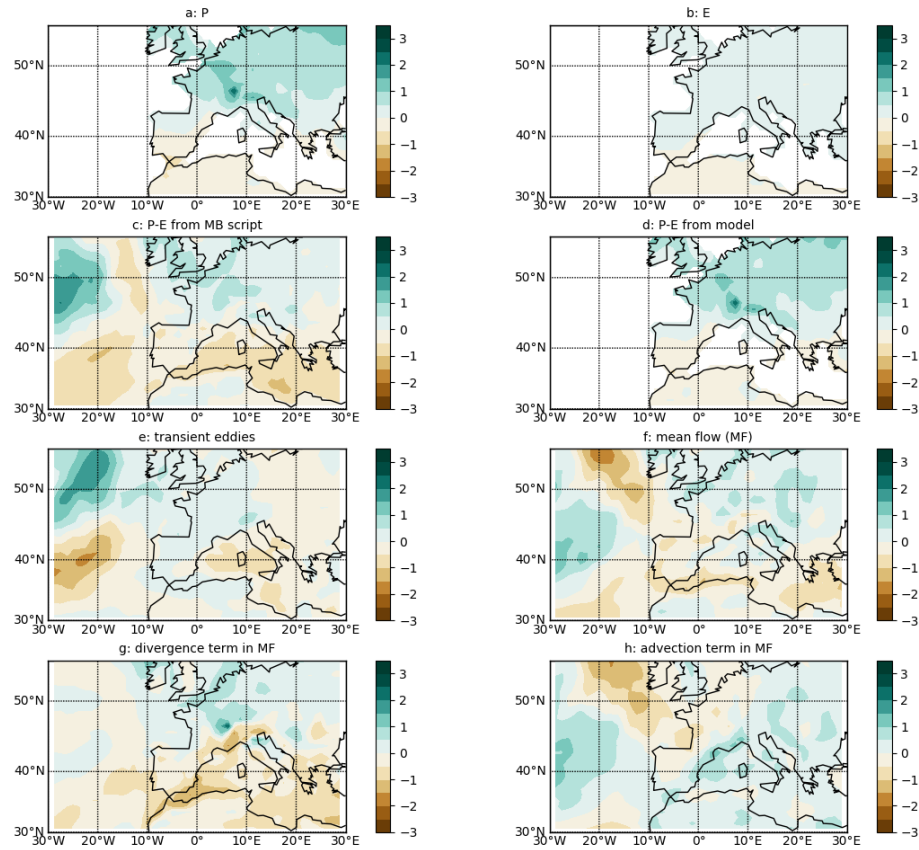


Figure 3.6: Changes from winter 1990-2005 – 2071-2080 period in the component of the moisture budget in CESM1-CAMS5 climate model. The changes in (a) P, (b) E, (c) P-E from scripts, (d) P-E from model, (e) change in transient eddy moisture convergence. (f) The change in moisture convergence by the mean flow and its component changes due to (g) the advection term, (h) mass divergence. Units are mm/day.

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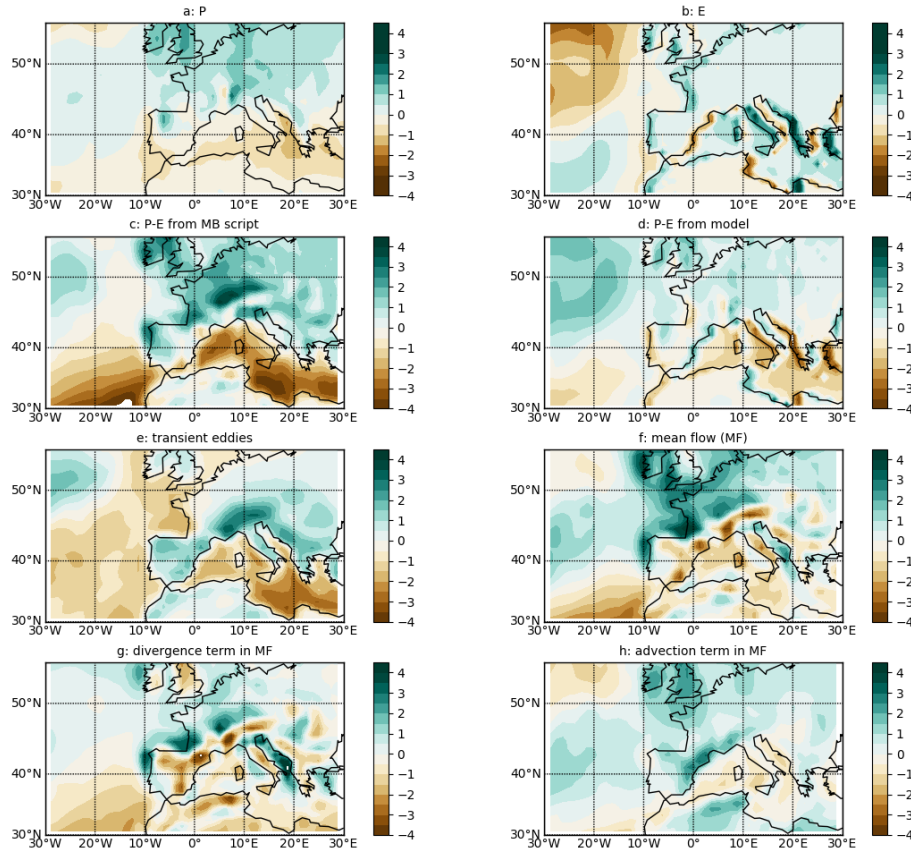


Figure 3.7: Changes from winter 1990-2005 to 2071-2080 period in the component of the moisture budget in CESM2 climate model. The changes in (a) P, (b) E, (c) P-E from scripts, (d) P-E from model, (e) change in transient eddy moisture convergence. (f) The change in moisture convergence by the mean flow and its component changes due to (g) the advection term, (h) mass divergence. Units are mm/day.

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We analyze the atmospheric circulation variability over these two periods to better understand moisture budget changes in North Africa sub-regions (West-Center-East) in present and future climate changes (Table 3.1 and 3.2). The Figure 3.8 (a, b and c) show Sea Level Pressure (SLP in millibars), wind speed (in knots, $1kt \approx 0.5m/s$) and wind directions (barbs) at the 850 pressure level, in ERA-Interim reanalysis, CESM1-CAMS5 and CESM2 models over the present period (winter 1990-2005), more details about wind barbs representation is provided in the Annex section. The ERA-Interim data set as well as CESM1-CAMS5 and CESM2 models reproduce well the Azores anticyclonic circulation, during winter 1990-2005 (Figure 3.8). An extension of the Azores anticyclonic circulation is detected in all data sets over the western part of North Africa (Morocco). This anticyclone is associated with high SLP values, varying between 1022 mb in ERA-Interim to 1024 hpa in CESM1-CAMS5 and 1025 mb in CESM2 model. The cause of this high pressure circulation centered over Morocco may be related to the subsidence induced by the high mountain ranges of Moroccan Atlas, acting like a barrier and thus deviate the atmospheric circulation to form the anticyclonic high pressure anomaly. Even though the regional topography is not well resolved by the global climate models (CESM1-CAMS5 and CESM2), it is important to consider the complex orographic features of North Africa region, that play a critical role in the regional precipitation patterns. [165] showed that under future changes, the drying over southwestern Mediterranean (Morocco) will be strongly influenced by the complex Atlas mountains, and can locally be of higher magnitude, due to subsidence induced by the topography. The anticyclonic circulation over Morocco, causes a change in 850 mb wind direction, from west over Morocco to North-West in Tunisia and Libya, with a transitional zone over Algeria. This particular North Africa anticyclonic circulation advects moisture from the Mediterranean Sea to water the central and eastern parts of our region of interest (Algeria to Libya), from North Atlantic Ocean to water Morocco coasts. However, it drives dry air from the Est towards

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Morocco lands. Thus, the advection term in the circulation mean flow tends to oppose the transient eddies moisture supply and in contrast, advects a dry air from the Sahara desert to North Africa countries and cause precipitation decline and broad aridification. Figure 3.9 shows the atmospheric circulation over North Africa in the period of winter 2071-2080, in CMIP5-CESM1 as well as CESM2 models. Both models simulate a North Africa anticyclonic circulation as seen in the present period, with small expansion towards the eastern part and pressure increase over North Africa.

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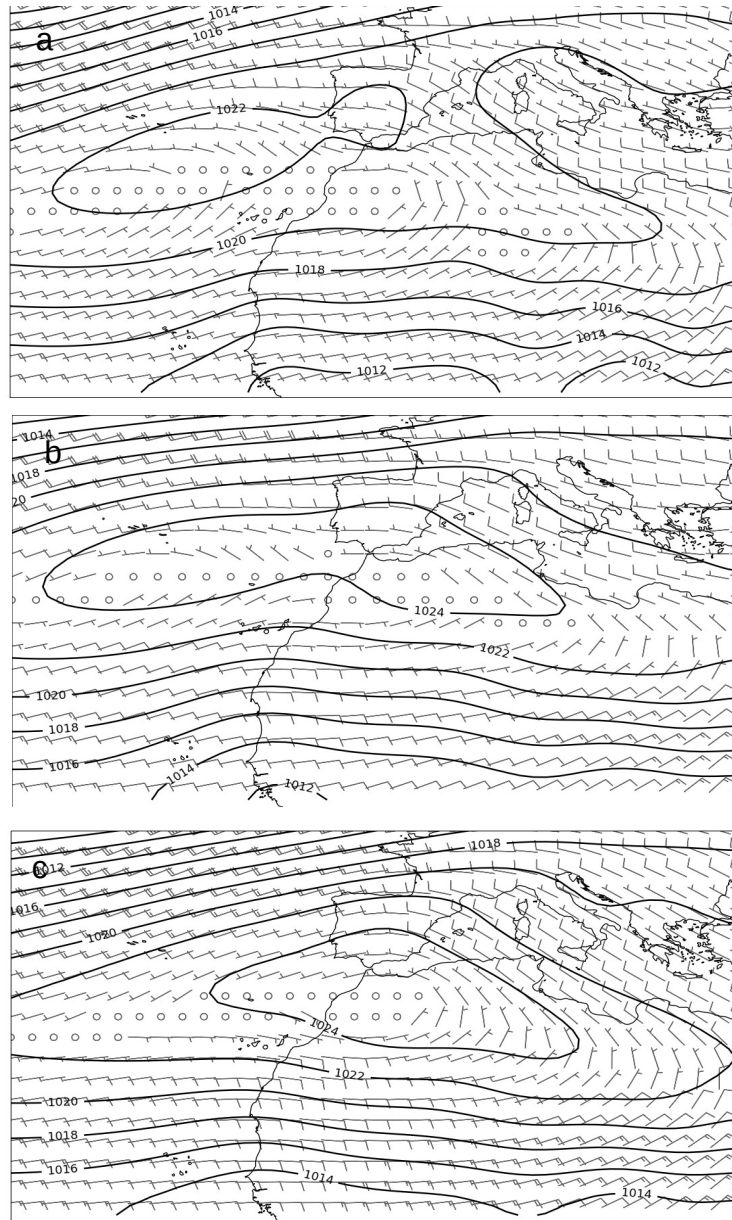


Figure 3.8: (a) ERA-Interim (b) CESM1-CAMS5 (c) CESM2 models, winter (November-April 1990-2005), Sea level pressure (black contours, from 1010 mb to 1030 mb, every 2mb), wind speed and direction (black bars, from 0-15m/s) at 850 pressure level.

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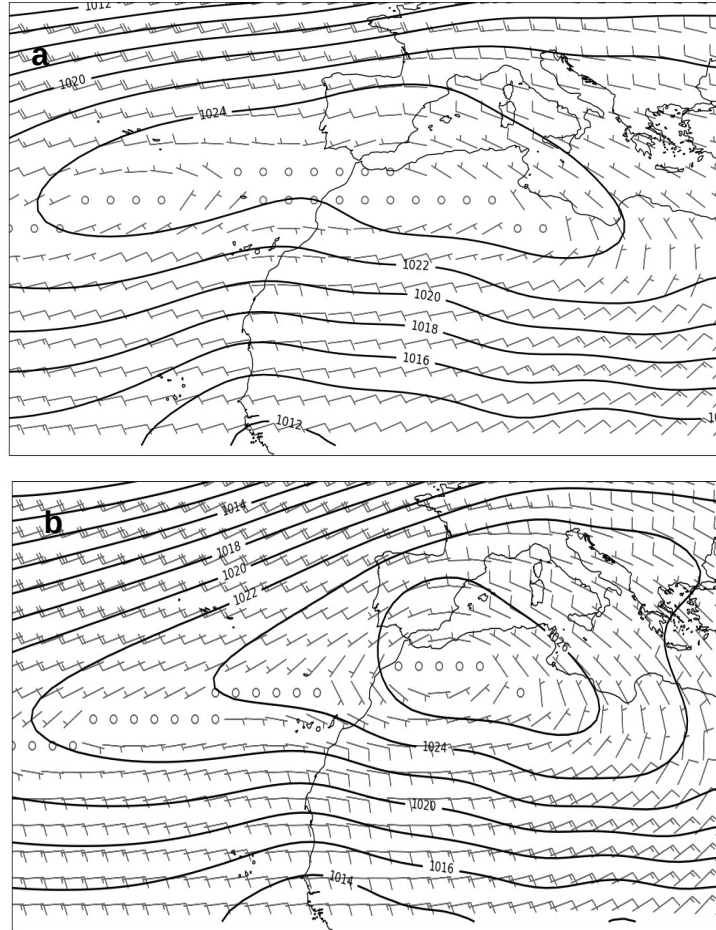


Figure 3.9: (a) CESM1-CAMS5 (b) CESM2 models, winter (November-April) 2071-2080, Sea level pressure (black contours, from 1010 mb to 1030 mb, every 2mb), wind speed and direction (black barbs, from 0-15m/s) at 850 pressure level.

Figure 3.10 (a and b) shows the projected winter (November-April) atmospheric circulation changes between winter 2071-2080 – 1990-2005, in both CESM1-CAMS5 and CESM2 models. We observe an anticyclonic circulation anomaly centered over eastern Mediterranean in both climate models. This particular high pressure anomaly is a

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distinct from the North Africa anticyclonic circulation, shown previously (Figure 3.8 and Figure 3.9). The changes in the eastern Mediterranean anticyclone (Figure 3.10) over these two periods, is associated with an increase in SLP between 1 and 3 mb in both data sets. This change in SLP is statistically not significant over the period of study (2071-2080, 1990-2005) if we compare it to the standard deviation of the interannual variations, but this is a robust feature already identified in previous studies [68]. This eastern Mediterranean pressure anomaly moves away the westerlies flux towards the Northern Europe and thus, prevent moisture to reach North Africa, which induces a decrease in moisture supply. In addition, changes in the mean flow divergent term tends to decrease and become negative over the three sub-regions of North Africa (Table 3.4), which in fact will drive more aridification and precipitation reduction over (Morocco-Algeria-Tunisia) under future changes (Table 3.3 and Figure 3.10). However, over North Libya to Egypt, this modified circulation may advect moisture from East-Mediterranean sea and water these regions (Figure 3.10 a and b).

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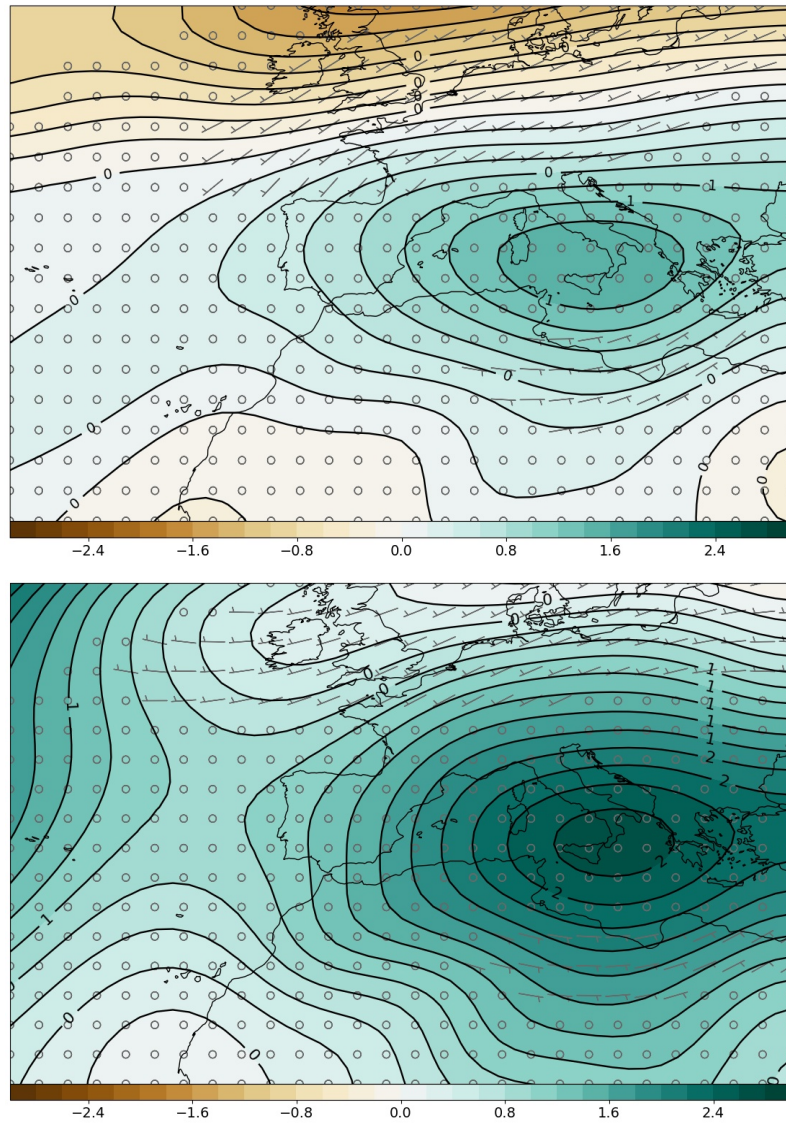


Figure 3.10: Winter 2071-2080 – 1990-2005 changes in (a) CESM1-CAMS5 (b) CESM2 models, sea level pressure (black contours), wind speed and direction (black barbs) at 850 pressure level.

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Correlation results between U850 index and ERA-Interim, CMIP5 and CMIP6 moisture budget data

In order to explain the robustness of the link between the wind index U850 and the precipitation variability over North Africa found over the last millennium as well as under changing climate (Figure 6.1 in Supplementary materials), we analyze the correlations between the U850 index and the MB decomposed terms, over the recent winter (November-April) period (1990-2005) as well as in the future winter (November-April) period (2071-2080). The Figures 3.11 to 3.13 and Table 3.3 show winter (November-April) correlation coefficients between U850 index and the different terms for ERA-Interim, CESM1-CAMS5 and CESM2 MB, during winter 1990-2005, averaged (Table 3.3) over the three North Africa sub-regions (West-Center-East). It is worth to note the high positive mean correlation values between U850 index and precipitation except over eastern part in CESM1-CAMS5 model. The correlations are also significant and positive in all data set between U850 and P-E budget. The correlations between the transient eddies and U850 index are negative, except over western and eastern part in ERA-Interim. The correlations are positive between U850 index and the mean flow term over the three sub-regions, except over the central part in CESM2 model. The correlations between U850 index and divergence term in mean flow, are positive except over the western parts in both ERA-Interim and CESM1-CAMS5 model. As for the divergence term, the correlations are positive between U850 index and the advection term in mean flow except over the western part in all data sets and over the central part in CESM2 models (Table 3.3).

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U850-MB correlations / sub-region	ERA-Interim			CESM1-CAMS5			CESM2		
	West	Center	East	West	Center	East	West	Center	East
U850 ~P	0.61	0.49	0.35	0.68	0.30	-0.10	0.69	0.51	0.42
U850 ~P-E (script)	0.60	0.46	0.40	0.60	0.58	0.30	0.25	0.20	0.21
U850 ~Transient Eddies	0.03	-0.08	0.05	-0.16	-0.40	-0.13	-0.16	-0.01	-0.19
U850 ~Mean Flow	0.45	0.31	0.08	0.53	0.48	0.10	0.38	-0.05	0.30
U850~Divergence in Mean Flow	-0.10	0.43	0.30	-0.07	0.27	0.21	0.04	0.35	0.38
U850 ~Advection in Mean Flow	-0.27	0.31	0.35	-0.38	0.29	0.23	-0.15	-0.10	0.30

Table 3.3: Mean correlation results between U850 index and MB terms over North Africa sub-regions (West-Center-East), in ERA-Interim, CESM1-CAMS5 and CESM2 models, during winter (November-April) 1990-2005.

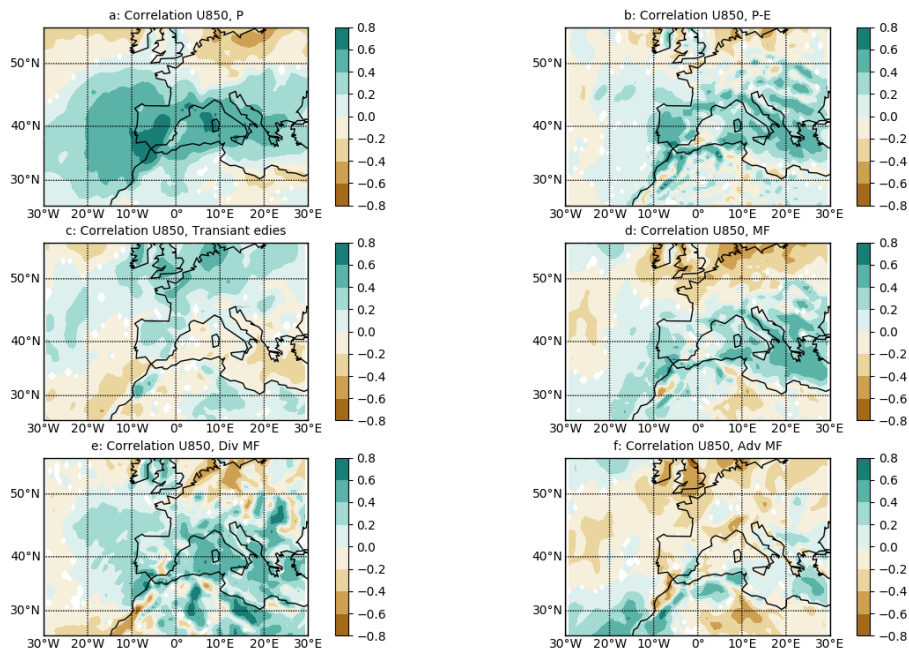


Figure 3.11: Significant correlations results between ERA-Interim U850 index and moisture budget terms over winter (November-April) 1990-2005 in North Africa (according to Pearson correlation test at 95%).

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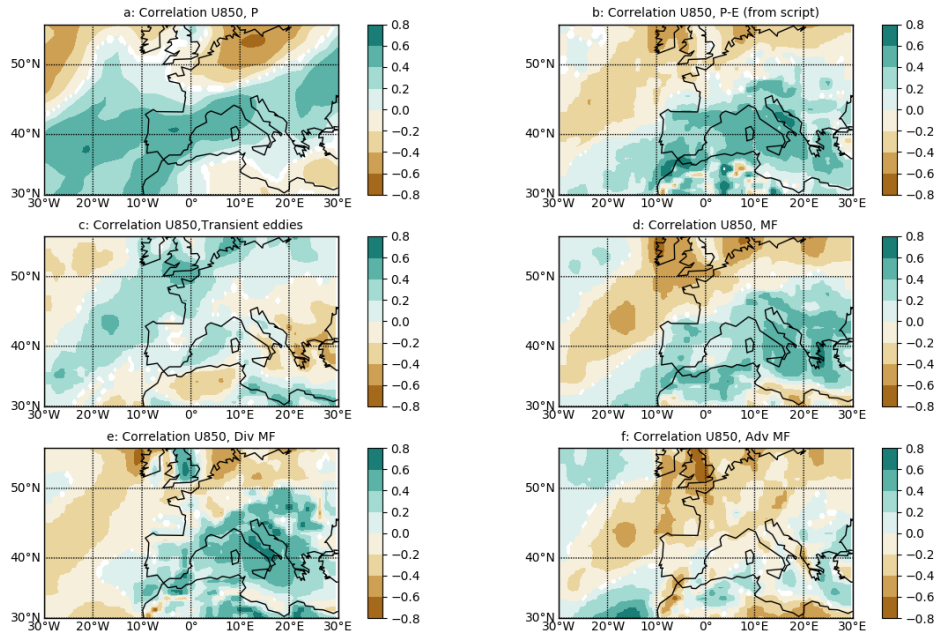


Figure 3.12: Significant correlations between CESM1-CAMS5 U850 index and moisture budget terms over winter (November-April) 1990-2005 in North Africa. (according to Pearson correlation test at 95%).

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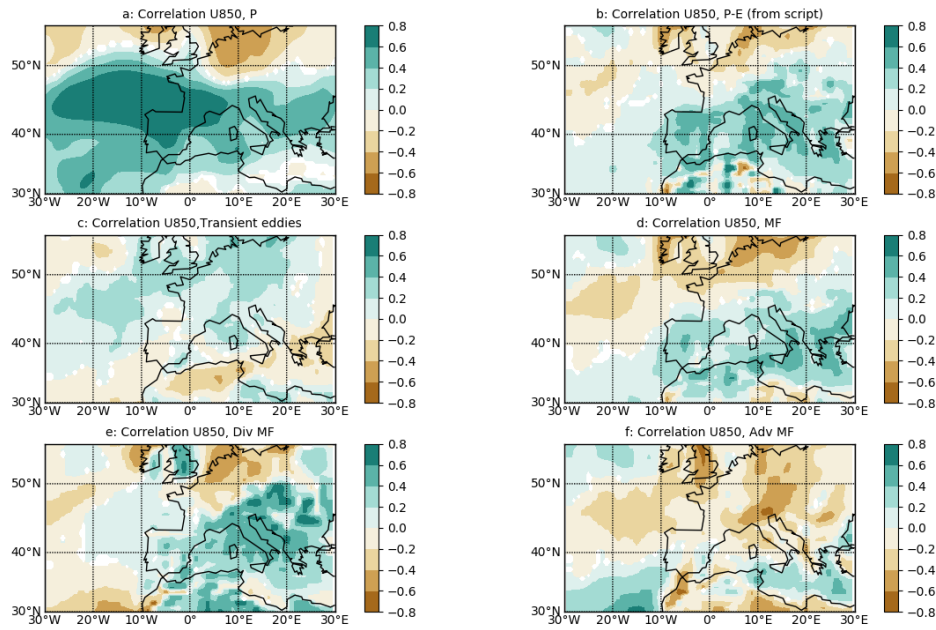


Figure 3.13: Significant correlations between CESM2 U850 index and moisture budget terms over winter (November-April) 1990-2005 in North Africa (according to Pearson correlation test at 95%).

Both CESM1-CAMS5 and CESM2 models agree with ERA-Interim re-analysis data, that in winter (1990-2005), the U850 index is more linked to the MB mean flow, over the three sub-regions of North Africa, except for CESM2 over the central part, where the link is weak.

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Hereafter, we analyze the winter (November-April) link between MB terms and U850 index in a warmer future climate, in both CESM1-CAMS5 and CESM2 model scenarios (RCP8.5 and SSP585). Overall, the correlation coefficients show the same behavior as in the present period, between P and U850 and between the MB terms and U850 index, with small differences between the three sub-regions.

The high pressure anomaly observed over eastern Mediterranean (Figure 3.10) under future changes, advects an eastern to south-eastern dry air over the three sub-regions of North Africa. As a consequence, the North Africa precipitation will be projected to decline, mainly by the divergence term in mean flow that oppose moisture flux from North Atlantic. In addition, the U850 index (defined between 10W-20E and 25N-32N) will be weakened by the influence of the subsidence effect of the eastern Mediterranean high pressures anomaly (anticyclone).

U850-MB correlations / sub-region	CESM1-CAMS5			CESM2		
	West	Center	East	West	Center	East
U850 ~P	0.15	0.68	0.45	0.35	0.05	0.40
U850 ~P-E (script)	0.20	0.34	0.26	0.32	0.36	0.28
U850 ~Transient Eddies	-0.25	-0.43	-0.38	-0.35	-0.36	-0.28
U850 ~Mean Flow	0.38	0.50	0.43	0.36	0.42	0.36
U850 ~Divergence in Mean Flow	-0.28	0.36	0.39	-0.15	0.30	0.34
U850 ~Advection in Mean Flow	-0.30	0.08	0.12	-0.40	0.10	0.11

Table 3.4: Mean correlation results between U850 index and MB terms over North Africa sub-regions (West-Center-East), in CESM1-CAMS5 and CESM2 models, during 2071-2080.

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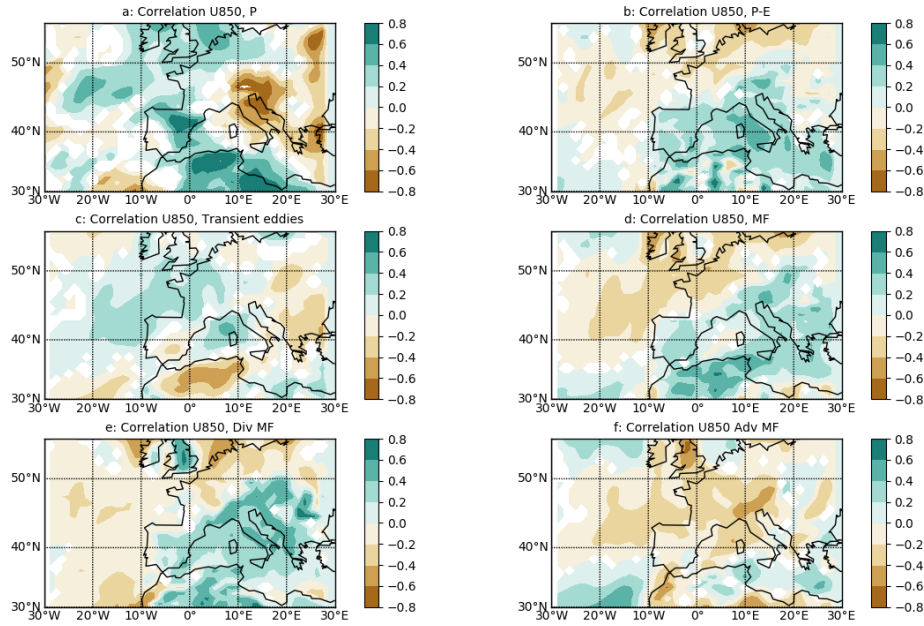


Figure 3.14: Significant correlations between CESM1-CAMS5: U850 index and moisture budget terms over North Africa, winter (November-April) 2071-2080, (according to Pearson correlation test at 95%).

3.4 Discussion

For the precipitation changes at the scale of the Mediterranean basin, our results confirm those showed by [57]. As for the whole Mediterranean region drying, the three North Africa sub-regions analyzed here, are linked to changes in moisture convergence by the mean flow in CESM1-CAMS5 and CESM2 models, however the drivers of those changes vary from the western to the eastern parts. To the West, the moisture convergence decreases, as the Atlantic moisture supply weakens due to the anticyclonic circulation over Morocco, which increases dry air advection from the Sahara Desert. The cause of this particular high pressure anomaly over Morocco may be due to interaction of Atlas

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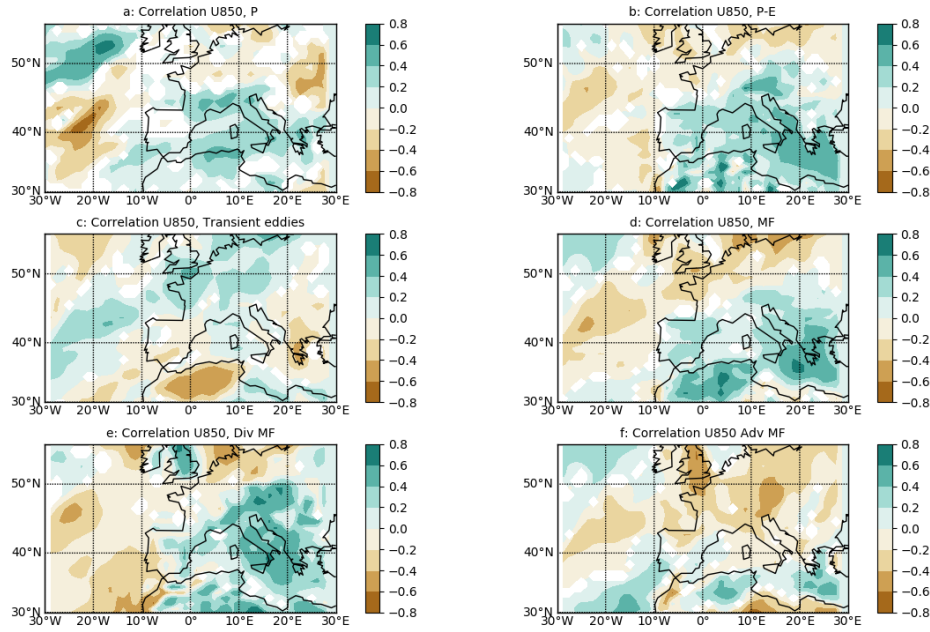


Figure 3.15: Significant correlations between CESM2: U850 index and moisture budget terms over North Africa, winter (November-April) 2071-2080, (according to Pearson correlation test at 95%).

mountains with the large-scale atmospheric circulation over the western part of North Africa. To the East, the potential cause of the precipitation decline is the subsidence movement of the eastern Mediterranean anticyclonic circulation, showed by [68]. They argued that the possible causes of this Mediterranean Sea anticyclonic circulation, is related to the relative decrease in the Mediterranean Sea temperature compared to the surrounding land areas, as a response to the projected climate warming over lands. The Mediterranean anticyclone circulation will also cause a weakening of the U850 index, under future climate changes. We noted that the correlations between U850 index and the divergence and advection terms in the mean flow are highly sensitive to the length

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of the time period considered in the analysis. So, it is important to take as long periods as possible. In Figure 6.6 (Supplementary materials) we showed a correlation results between U850 and MB terms in CESM2 model over 30 years (winter 2071-2100). We see a same pattern as in our results presented in Figure 3.15, but there is notable variation in the correlation results in the different terms.

Even though the complexity of processes involved in precipitation changes between the present and future climate over North Africa, the link found between U850 index and the different terms in MB, mainly the advection and divergence terms in mean flow is significant and thus, constitute a practical tool to monitor precipitation changes at regional and national scales.

3.5 Conclusion

By analyzing ERA-Interim reanalysis data set and global climate models (CESM1-CAM5 as well as CESM2), this study aims to understand better and quantify the role of atmospheric circulation in North Africa precipitation response to the present (1990-2005) climate as well as in future (2071-2080) climate changes during the cold season (November–April). The main findings of this study are the following:

the atmospheric moisture budget analysis in ERA-Interim reanalysis, CESM1-CAM5 and CESM2 models, revealed that not only the transient eddies moisture flux contribute to North Africa precipitation variability in present climate. Nonetheless, precipitation changes also depend on the atmospheric mean flow response, as described by U850 index over North Africa. Under the future changes, the atmospheric moisture budget is also dependent on both the transient eddies and the mean flow convergence, as in the present period.

Both CESM1-CAM5 and CESM2 climate models agree on the pattern of Northern Hemisphere atmospheric circulation changes under continued anthropogenic forcing: we showed a change in the atmospheric

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circulation over North Africa and Mediterranean Sea, by the mean of SLP and winds at 850 mb, as well as a change in the link between the MB terms and the U850 index.

North Africa region is impacted differently from West to East and have distinct behavior than the whole Mediterranean region. The anticyclonic circulation anomaly, observed over the western part of North Africa (Morocco) in present climate, in both ERA-Interim reanalysis as well as climate models (CEMS1-CAMS5 and CESM2), impacts differently, the pattern of the westerlies (and winds at 850 pressure level), going from the West to the East of North Africa. It deviates the atmospheric flux from North-West in the central and eastern parts which drives moisture and precipitation, to East in western parts, which drive aridification and precipitation decrease.

Under future changes, both North African anticyclone as well as the eastern Mediterranean high pressure anomaly, will impact not only North Africa western part, but all the three sub-regions. It will advect dry air from the Sahara and deviates the westerlies circulation towards North Europe and thus blocks moisture supply and causes more aridification and precipitation reduction over the three sub-regions of North Africa. As consequence of the changes in the atmospheric circulation, caused by the two anticyclones over North Africa and over the eastern Mediterranean Sea, the divergence term in the mean flow as well as the U850 index are projected to decrease under the future changes over the three sub-regions of North Africa. The complexity of processes involved in precipitation changes in North Africa leads us to conduct more analyses and simulations with further climate models at high resolutions, in order to help national and regional decision makers in water planning and management.

CHAPTER 4

Droughts assessment and monitoring in Algeria

4.1 Introduction

Dry areas, either semi-arid, arid, or hyper-arid, occupy a significant portion of North Africa. As shown in Chapter 1, drought is a complex natural phenomenon that affects many aspects of the economy and occurs over a variety of time scales. In addition, we showed that drought is divided into four categories: meteorological, hydrological, agricultural, and socioeconomic depending on approaches to drought measurement. North Africa region is not only highly prone to drought but also one of the most water-scarce areas in the world and approximately three quarters of the region consists of a desert [166]. Drought is a regular feature of North African life. These countries all have a Mediterranean climate around their coasts, where the majority of the population lives, but they all face the arid conditions of the Sahara desert in the south. Drought have become more widespread, prolonged, and frequent over the past decades, likely due to climate change [166] (See also Chapter 1). Agriculture, including livestock, forestry, and fisheries, is particularly vulnerable to drought with severe effects on food supplies and livelihoods, particularly for smallholders and rural societies' poor strata. Agriculture is the first sector to be affected when there is drought and also the most impacted

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of all economic sectors [166]. A paradigm shift to a more proactive approach, based on risk management concepts, is highly needed to build greater societal resilience to the impacts of drought.

In light of this issue, the high-Level Meeting on National Drought Management Policy (HMNDP), took place in Geneva in March 2013, bringing together more than 430 participants from over 90 nations [166]. The meeting's goal was to undertake fundamental change in drought awareness and management, as well as to encourage governments to establish and enforce national drought management policies that are in line with their development goals.

Algeria has minimal, volatile, and unevenly distributed natural resources. Natural water supplies are primarily located in the north of the country, except for fossil water aquifer (underground water reservoirs that have been geologically sealed) in the Sahara [166]. Around 90 percent of the population live on 13% of the country's land area along the North country coast. This is the wettest area with an average annual rainfall ranging from 400 to 1000 mm. The rainfall is between 100 and 400 mm in south region, for instance in the Saharan Atlas. These are high plains of cereals and steppe pastures from prehistorical-Saharan times, where consequences of drought are greatest. In the south, the arid desert of the Sahara occupies the remaining 87 % of the land [166]. 62 % of water resources are consumed by irrigation. Drinking water demand, which has significantly increased since the 1970s, accounts for 35 per cent of total demand. The industry's share of water is just 3 % [166]. In addition to the imbalances in resource availability between regions, there is strong competition for water among the major water users, which makes distribution and trade-offs difficult [166]. In terms of volume and proportion, domestic water use has increased significantly from 16 percent of the 1975 total to 35 percent today. In the same period, the share of agricultural water dropped from 80 to 62 % [167]. Agriculture is the sector most vulnerable to weather and also the most prone to drought [166].

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4.2 Droughts history and impacts in Algeria

Since the fifteenth century, significant and prolonged dry conditions occurred in Algeria (Table 4.1), in the 1540s, the 1860s, the 1890s, the 1922s and the 1940s. Over the nineteenth century (1850-1950), North Algeria, periodically experienced difficult years linked to droughts, agricultural crop failure, shortage, famines and health problems [166]. Drought, famine, cholera occurred during these drought events, according to historical sources, in particular a prolonged drought at Belesma (Batna) that caused significant damage to the cedar forests, according to [168]. [169] noted that the 1940s drought was primarily related to poor spring rains, while the 1981s drought was primarily due to poor winter rains. Official sources estimated that about 3 000 people died of starvation in a population of 80 000 in Ain Sefra (region of Oran), and 900 000 sheep were lost, totaling nearly 90% of the livestock. Algeria recorded the lowest rainfall since 1945, 20 years later in early 1966, resulting in poor crop germination and production loss [170].

Around 1922 and 1992, the north country experienced many dry periods: 1910-1920, 1939-1948, and 1977-1992. Famine, wildfires and significant social upheaval characterized these periods. The dry period from 1973 to 1997 had a negative effect on water supplies, with reduced inflows into dam storage, lower groundwater levels and shallow wells and springs running dry [171] (Table 4.1). In Algeria, the drought event during 1999–2002 appears to be the worst since the middle of the fifteenth century, and is probably due to anthropogenic climate change [172, 173]. Wide urban areas, especially in the Wilaya of Algiers, experienced critical water shortages as a result of low rainfall in 2001 and the winter season of 2002. Droughts have disrupted planting and germination in recent years, adversely affecting plant growth and crop production for arable and livestock farmers. Over the last 30 years, Algeria has experienced severe and prolonged drought, which has resulted in a rainfall shortfall of 30% of normal annual rainfall with negative effect on river flow regimes, with

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significant implications for all socioeconomic activities in the country.

Drought history in Algeria 1456–2002			
Period/year	Events	Duration	Observations
1456	1 drought event	1 year	The driest single years over 1456-2002
1548-1553		6 years periods	
1737-1742		6 years periods	
1765-1768	1 drought event	4 years	
1770-1775	1 drought event	6 years periods	
1867	1 drought event	1 year	
1890	1 drought event		Data from Oran station
1910-1920	Period of drought		Similar to droughts in Sahel
1922	1 drought event		
1937	1 drought event		Severe and generalised
1939 – 1948	1 drought event	2 years	Time of particularly severe dry conditions
1945-47	1 drought event	3 years	'year of hunger' described by Albert Camus, 3000 death south Oran
1961	1 drought event		Severe and generalised
1970	1 drought event		Severe and generalised
1977-1978	1 drought event	2 years	Second severe drought period preceding dieback of Cedrus atlantica
1978	1 drought event	1 year	The driest observed year (138.4 mm)
1981-84	2 drought events	4 year	
1997	1 drought event	1 year	The driest reconstructed (148.6 mm)
1999–2002	1 drought event	4 years	The longest period of consecutive drought years for AD 1456–2002

Table 4.1: Drought history in Algeria 1456–2002.

source: (<http://www.fao.org/3/CA0034EN/ca0034en.pdf>)

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The long drought periods observed in Algeria (Table 4.1) have clearly contributed to quantitative and qualitative water reserve depletion [174]. These droughts have had a negative impact on meeting the water needs of all socio-economic sectors, particularly agriculture, and the protection of terrestrial and aquatic ecosystems [174]. In 2007, only 43 000 hectares of the 150 000 hectares that had to be irrigated were actually irrigated due to drought [175]. Irrigation water was reallocated to the population's drinking water supply, particularly in the country's west [174]. Reduced rainfall due to drought since the early 1970s has resulted in a steady decline in groundwater supplies in major northern aquifers.

4.3 Drought risk management policies

In Algeria, there is no framework specifically dedicated to drought adaptation, and the existing early warning system for natural disaster management is mainly focused on floods and earthquakes. Among the national structures dedicated to natural disaster management, we distinguish:

- A specialized department under the Ministry of Environment that prevents and manages major risks and engages in national and worldwide media planning, coordination, and distribution.
- The Ministry of the Interior has established a National Center for Warning and Decision Support, which is responsible for collecting, disseminating, and leveraging real-time information and data of all kinds, including the first network built by the National Meteorological Office with the help of the World Bank in the aftermath of the floods in Algiers in 2001.
- Ministry of Water Resources (MRE) is the central authority responsible for the development and implementation of the National Water Policy.

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4.4 Drought Indices

An objective evaluation of drought condition in a particular area is the first step for planning of the use of water resources. For this purpose, along the years, several indices have been developed to evaluate the water supply deficit in relation to the time duration of precipitation shortage (see [176, 177] and references therein). The World Meteorological Organization (WMO) defines a drought index as “an index which is related to some of the cumulative effects of a prolonged and abnormal moisture deficiency”. Despite the number of indices developed, none of them is universal and complete in capturing all the aspects of droughts. [178, 177] identified four basic characteristics that drought indices should possess:

- the timescale of the index must be appropriate.
- the index must also quantify, both temporally and spatially, drought conditions (be able to cover the entire drought period over the studied region).
- the index should be relevant for the investigated drought event (use the appropriate index depending on the type of drought).
- there must be a long record for the index available for referencing. Most drought indices subtract normal seasonal fluctuations in water availability from climatological variations, estimating severe events and therefore drought. [177].

[172] have identified four weaknesses associated with the application of those indices.

- The index used determines the start, end, and cumulative drought stress of drought events. When using different indices to describe the same drought, this may lead to inaccuracy and conflicting results.

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- The impact of surface hydrology (runoff) and evapotranspiration is not always well known, as it is difficult to put these processes into equation.
- Most indices use a monthly (or longer) timescale, which makes it difficult to monitor droughts in real time. As a result, the fight against droughts will be ineffective.
- The difference in the drought calculation for the different drought indices (e.g. [179]) results in uncertainty of which indices could be used under a given circumstance and for what affected sector. To track drought with the right index, we need expert judgement.

Regional micro-elements (topography, vegetation, etc.) play a significant role in the local drought conditions. Consequently, caution must be taken when interpreting (maps of) drought indices. Among the panoply of the developed indices, we can mention: Standardized Precipitation Index (SPI), Palmer Drought Severity Index (PDSI), Surface Water Supply Index (SWSI), Percent of Normal/Departure from Normal Precipitation Deciles, Standardized Precipitation Evapotranspiration Index (SPEI), Effective Drought Index (EDI) and many others.

Here we describe only the most used drought indices (Table 4.2):

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Index	Type	Input data	Period (months)	Abbr	Reference
Standardized Precipitation Index	Meteorological	P	1,3,6,12,24	SPI	[180]
Standardized Precipitation Evaporation Index	Meteorological	P, PET	1,3,6,12,24	SPEI	[181]
Palmer Drought Severity Index	Soil moisture	P,T,AWC	-	PDSI	[182]

Table 4.2: Types of drought indices, Input data: Precipitation (P), Potential evapotranspiration (PET), Temperature (T), Available Water Content (AWC).

4.4.1 Standardized precipitation index (SPI)

The Standardized Precipitation Index (SPI) is the most commonly used index for detecting and characterizing meteorological droughts. The standardized precipitation index (SPI) was established by [183, 184] and described in detail by [185]. The SPI measures drought or excess in precipitation based on precipitation data recorded over a given period and precipitation climatology [186]. Since the SPI can be calculated over different precipitation accumulation periods (typically ranging from 1 to 48 months), the resulting different SPI allow for estimating different potential impacts of a meteorological drought [187].

For instance, the SPI-3 compares precipitation totals from a particular 3-month period with precipitation totals from the same 3-month period for all years in the climatology. Consequently, a calculated SPI-3 at the end of May compares the total precipitation in March, April and May of that year to the cumulative precipitation in March, April, and May of all previous years [188].

The computation of SPI does not need a specific number of precipitation time series length [189]. [190] argued that the effects of the probability distribution will be more consistent over time if more samples of time series of wet and dry events are used. The SPI can be estimated from

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as little as 20 years of data, but even though missing data is taken into consideration, it is better that the time series include at least 30 years of data [189].

Positive SPI values represent wet conditions, the higher the SPI, the more unusually a wet period of time is. Negative SPI values represent dry conditions; the lower the SPI, the more unusually dry a period of time is. The observed (monthly, 3-monthly,...) cumulative precipitation data is fitted to a probability distribution, for example a gamma distribution, which in turn is transformed to a normal distribution with a mean of zero (the steps are shown in the next paragraphs). The primary strength of the SPI normalization is that it can be compared between different climate regions [186, 191]. Implementations of the SPI involve drought monitoring and forecasting, climate impact studies, etc [186, 191]. Uncertainty in the SPI comes from the fact that the precipitation data must be fitted to the probability distribution function, since precipitation accumulated over a period of less than 12 months is generally not normally distributed. This leads to some confusions in the algorithm, as there are various probability distribution functions that can be used (different authors use different distribution functions). However, the gamma distribution is shown to fit the precipitation data quite well in several studies that utilize the SPI ([192]) and to be the best model for characterizing monthly precipitation.

Another shortcoming of the SPI is that it does not consider the loss of moisture via evapotranspiration [193]. Despite some drawbacks inherent to the SPI, this index is widely used. However, a complete drought monitoring system should not be based on one index alone [190, 194]. The combination of this index with other drought products (e.g. satellite data) and real-time monitoring of stream flow and soil moisture levels is effective for drought monitoring systems.

Drought events are identified when the results of SPI, for whichever timescale is being investigated, become negative. Before the SPI exceeds

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a value of 0, the drought is considered to be continuing. However, [184] showed that a drought period begins when the SPI is equal to -1 or less, but there is no rule in place, as some researchers will prefer a threshold that is less than 0, but not equal to -1, while others will initially classify drought at values less than -1 [189]. SPI's ability to be computed at different timescales enables it to be used in a variety of situations. SPI values of 3 months or less can be useful for simple drought tracking, 6 months or less for agricultural impacts, and 12 months or longer for hydrological impacts, depending on the drought effect under investigation. SPI can also be measured on gridded precipitation datasets, allowing users who aren't dealing with station-based data to apply it too. A classification of droughts (and wet periods) is given in Table 4.3.

SPI value	Category	Cumulative probability (matching quantiles)
$SPI \geq 2.0$	Extreme wet	0.89 - 1.00
$1.5 \leq SPI < 2.0$	Sever wet	0.93 - 0.89
$1.0 \leq SPI < 1.5$	Moderate wet	0.84 - 0.93
$-1.0 \leq SPI < 1.0$	Near normal	0.16 - 0.84
$-1.5 \leq SPI < -1.0$	Moderate dry	0.07 - 0.16
$-2.0 \leq SPI < -1.5$	Severe dry	0.02 - 0.07
$SPI < -2.0$	Extreme dry	0.00 - 0.02

Table 4.3: SPI drought categories [184]

Another advantage of SPI is that the SPI-algorithm is relatively easy to implement. The SPI is the standard deviation from this standardized normal variable (Figure 4.1).

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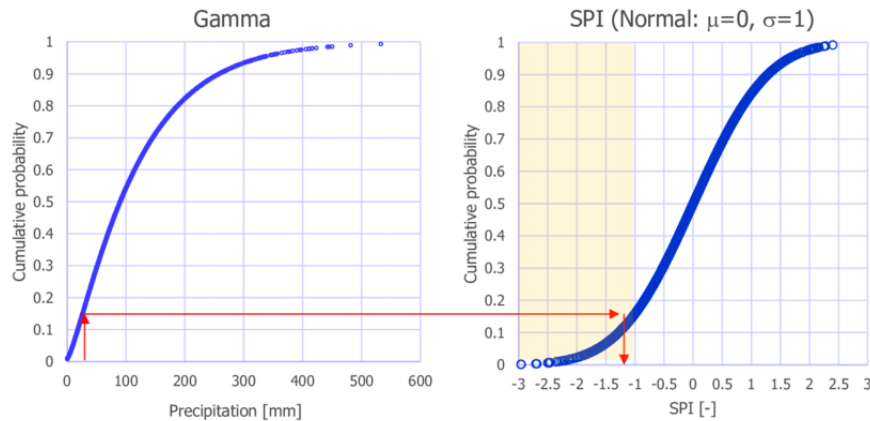


Figure 4.1: SPI calculation scheme. First, precipitation is fitted by gamma distribution (left). Second, cumulative probability of fitted precipitation is used to compute SPI value with the normal distribution (right) [195].

Procedure and formula for Computation of SPI:

The transformation of the precipitation value in to standardized precipitation index has the purpose of:

- a. Transforming the mean of the precipitation value adjusted to 0
- b. The standard deviation of the precipitation is adjusted to 1.0
- c. Skewness of the existing data has to be readjusted to zero

When these goals have been achieved, the standardized precipitation index can be interpreted as mean 0 and standard deviation of 1.0

The calculation steps are as follows:

- Mean of the precipitation can be computed as:

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$$Mean = \bar{X} = \frac{\sum_{i=1}^n X}{N} \quad (4.1)$$

Where N is the number of precipitation observations and X_i is the precipitation timeseries.

- The standard deviation for the precipitation is computed as

$$s = \sqrt{\frac{\sum_{i=1}^n (X_i - \bar{X})^2}{N}} \quad (4.2)$$

- The skewness of the given precipitation is computed as

$$skew = \frac{N}{(N-1)(N-2)} \sum_{i=1}^n \left(\frac{X - \bar{X}}{s} \right)^3 \quad (4.3)$$

- The precipitation is converted to log normal values and the statistics U , shape and scale parameters of Gamma distribution (Γ) are computed:

$$\log mean = \bar{X}_{\ln} = \ln(\bar{X}) \quad (4.4)$$

$$U = \bar{X}_{\ln} - \frac{\sum_{j=1}^n \ln(X)}{N} \quad (4.5)$$

$$shape\ parameter = \beta = \frac{1 + \sqrt{1 + \frac{4U}{3}}}{4U} \quad (4.6)$$

$$scale\ parameter = \alpha = \frac{\bar{X}}{\beta} \quad (4.7)$$

After that, the parameters are utilized to calculate the cumulative probability distribution, which is provided by:

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$$G(x) = \frac{\int_0^x x^{a-1} e^{\frac{-x}{\beta}} dx}{\beta^a \Gamma(a)} \quad (4.8)$$

Because the Gamma function is undefined at $x = 0$ and a precipitation distribution can contain zeros, the cumulative probability is given by:

$$H(x) = q + (1-q) G(x) \quad (4.9)$$

q is the probability of zero. The cumulative probability $H(x)$ is then transformed to the standard normal random variable Z with mean zero and variance of one, which is the value of the SPI [185]. We employ the approximate conversion provided by [196] as an alternative

$$Z = SPI = -\left(t - \frac{c_0 + c_1 t + c_2 t^2}{1 + d_1 t + d_2 t^2 + d_3 t^3}\right) \quad 0 < H(x) \leq 0.5 \quad (4.10)$$

$$Z = SPI = +\left(t - \frac{c_0 + c_1 t + c_2 t^2}{1 + d_1 t + d_2 t^2 + d_3 t^3}\right) \quad 0.5 < H(x) \leq 1 \quad (4.11)$$

where:

$$t = \sqrt{\ln\left(\frac{1}{H(x)^2}\right)} \quad 0 < H(x) \leq 0.5 \quad (4.12)$$

$$t = \sqrt{\ln\left(\frac{1}{(1.0 - H(x))^2}\right)} \quad 0.5 < H(x) \leq 1.0 \quad (4.13)$$

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$$\begin{aligned}c_0 &= 2.515517 \\c_1 &= 0.802583 \\c_2 &= 0.010328 \\d_1 &= 1.432788 \\d_2 &= 0.189269 \\d_3 &= 0.001308\end{aligned}\tag{4.14}$$

The values of $c_0, c_1, c_2, d_1, d_2, d_3$ are constants being widely employed for SPI computation [196]

4.4.2 The Standardized Precipitation Evapotranspiration Index (SPEI)

SPEI is based on the original SPI. It is obtained in the same way as the SPI. However, it takes into account the "climatic water balance", which is characterized as the standardized water balance between precipitation (P_i) and potential evapotranspiration (PET_i) as follows [197]:

$$Di = P_i - PET_i\tag{4.15}$$

the monthly $PET(mm)$ is calculated by

$$PET = 16K\left(\frac{10T}{I}\right)^m\tag{4.16}$$

T is the monthly mean temperature given in (C); I is a heat index, computed as the sum of twelve monthly index values i . i is computed from monthly mean temperature as follows:

$$i = \left(\frac{T}{5}\right)^{1.514}\tag{4.17}$$

The coefficient m is a function of I : $m = 6.75 \times 10^{-7} I^3 - 7.71 \times 10^{-5} I^2 + 1.79$

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$\times 102 I + 0.492$. K is a correction coefficient that is calculated based on the latitude and month,

$$K = \left(\frac{N}{12} \right) \left(\frac{NDM}{30} \right) \quad (4.18)$$

NDM represents the number of days of the month and N is the maximum number of sun hours, which is calculated as:

$$N = \left(\frac{24}{\pi} \right) \omega_s \quad (4.19)$$

where ω_s is the hourly angle of sun rising, given as

$$\omega_s = \arccos(-\tan \varphi \tan \delta) \quad (4.20)$$

φ is the latitude in radians and δ is the solar declination in radians, calculated as:

$$\delta = 0.4093 \sin\left(\frac{2\pi J}{365} - 1.405\right) \quad (4.21)$$

J is the mean Julian day of the month.

The Log-Logistic function, defined by the equation (4.22), is the probability density function used to model the D series, where α , β and γ are the size, shape, and origin parameters, respectively, for D values in the $(\gamma > D < \infty)$ range. The L-moments approach was used to derive these parameters.

$$f(D) = \frac{\beta}{\alpha} \left(\frac{D - \gamma}{\alpha} \right)^{\beta-1} \left(1 + \left(\frac{D - \gamma}{\alpha} \right)^{\beta} \right)^{-2} \quad (4.22)$$

L-moments method was developed by [198, 199]. This method is widely

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used for regionalization, estimation parameters and determine the quantiles. L-moments method is the linear function of the probability weighted moments (PWM) [200, 201]

The SPEI is calculated using the log-logistic distribution as standardized values of (D) as follows:

$$F(D) = \left[1 + \left(\frac{\alpha}{D - \gamma} \right)^\beta \right]^{-1} \quad (4.23)$$

Following the approximation of [196], we get

$$SPEI = W - \frac{C_0 + C_1 W + C_2 W^2}{1 + d_1 W + d_2 W^2 + d_3 W^3} \quad (4.24)$$

where

$$W = \sqrt{-2 \ln(P)} \text{ for } P \leq 0.5 \quad (4.25)$$

P is the probability of exceeding a given D value, $P = 1 - F(x)$. If $P > 0.5$, then P is substituted by $1 - P$ and the sign of the resultant SPEI is reversed. The constants are $C_0 = 2.515517$, $C_1 = 0.802853$, $C_2 = 0.010328$, $d_1 = 1.432788$, $d_2 = 0.189269$, and $d_3 = 0.001308$.

4.4.3 The Palmer Drought Severity Index (PDSI)

Palmer defines drought as a period of months or years during which the actual water supply is less than the predicted water supply [202]. The Palmer Drought Severity Index (PDSI) uses temperature and precipitation data to estimate relative dryness [202]. The PDSI estimates the departure relative to normal conditions in the surface water balance by using an hydrological accounting system [172, 177]. The PDSI is primar-

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ily considered as a meteorological drought indicator and sometimes, as an agricultural drought indicator [203].

This index is determined using a simple two-layer soil model based on the following assumptions: (1) Runoff occurs only if both the surface and deeper soil layers are saturated; (2) evapotranspiration removal from the surface soil layer occurs at a potential rate and (3) evapotranspiration loss from the deeper soil layer is dependent on initial moisture content [204]. The amount of water moved between the two soil layers is estimated using a soil–water balance equation, and the precipitation ($\hat{P}$) is computed as follows:

$$\hat{P} = \alpha_i PE + \beta_i PR + \gamma_i PRO - \delta_i PL \quad (4.26)$$

where $\hat{P}$ is the monthly amount of precipitation required to maintain a normal soil moisture level. PE , PR , PRO , and PL represent potential evapotranspiration, potential soil water recharge, potential runoff, and loss, respectively [205]. α , β , γ , σ represent their respective coefficients. These coefficients are calculated as follows:

$$\alpha_i = \frac{\overline{ET_i}}{\overline{PE_i}}, \beta_i = \frac{\overline{R_i}}{\overline{PR_i}}, \gamma_i = \frac{\overline{RO_i}}{\overline{PRO_i}}, \delta_i = \frac{\overline{L_i}}{\overline{PL_i}} \quad (4.27)$$

The average value over several months is shown by the bar above each parameter in the equation. ET , R , RO , and L represent the evapotranspiration, the soil water recharge, runoff and loss respectively.

The difference between the precipitation (P) in a particular month and the calculated precipitation ($\hat{P}$) under the actual conditions is called the moisture departure (d). This parameter is computed as follows:

$$d = P - \bar{P} = P - (\alpha_i PE + \beta_i PR + \gamma_i PRO - \delta_i PL) \quad (4.28)$$

At various locations and periods, moisture departure might have dif-

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ferent significance for the surface water balance. To account for this, the moisture departure was multiplied by the corresponding month's climate characteristic coefficient (K), where K for month i was determined by Palmer as follows:

$$K_i = \frac{17.67}{\sum_{j=1}^{12} \overline{D_j} K'_j} K'_i \quad (4.29)$$

$$K'_i = 1.5 \log_{10} \left[\frac{\frac{\overline{PE_i} + \overline{R_i} + \overline{RO_i}}{\overline{P_i} + \overline{L_i}} + 2.8}{\overline{D_i}} \right] + 0.5 \quad (4.30)$$

K_i is Palmer's estimation for the climate characteristic of a region, $\overline{D_i}$ is the average moisture departure for the corresponding month and 17.67 is an empirical constant. Multiplying d and K gives the moisture anomaly index (Z), or Palmer's Z , as given in the following equation:

$$Z = dk \quad (4.31)$$

The Z index is used to show how wet or dry a month was without taking into account recent precipitation trends [202]. The Z index is used to calculate the PDSI value for a given month using the formula

$$X_i = 0.897X_{i-1} + \left(\frac{1}{3}\right)Z_i \quad (4.32)$$

where X_i is the PDSI value of the current month and X_{i-1} is the PDSI value of the previous month. The quantities 0.897 and $1/3$ are called the duration factors.

Theoretically, the PDSI ranging from about -10 to $+10$, is a standardized measure that can be compared between different regions and time [172]. However, some complaints about the PDSI still exist as follows:

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- Even though the PDSI has been reasonably successful at quantifying long-term drought, it does not capture droughts on time scales less than about 12 months. The PDSI responds slowly to recognize short dry spells [206];
- The spatial comparability of the PDSI across regions is not satisfactory as the Standardized Precipitation Index (SPI). The reason is that the standardization and the calibration coefficients used by the PDSI depended on limited sample data [207];
- The calculation of potential evapotranspiration in the original Palmer model applying the Thornthwaite equation [208] based merely on latitude, month and temperature may lead to errors;
- The PDSI was criticized for its treatment of all precipitation (including snow) as rainfall, and the effects of frozen soils or vegetation on evaporation are not taken into account, in addition, it is more complex to calculate than the SPI [172, 207, 177].

We calculated the correlations between the PDSI and SPI (3, 6 and 12) in North Algeria over 1950-2012. The SPI and PDSI data sets are retrieved from (source: climatedataguide.ucar.edu/climate-data/) as netcdf files. We extracted both (SPI, PDSI) averaged values over Algeria domain (2W-10E, 32N-36N). The results showed that the correlation between the two indices has a maximum for the timescale of 12 months with a correlation coefficient of 0.80 suggesting that the PDSI does indeed have an inherent timescale even if it is not explicitly defined.

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4.5 International recommendations on the use of drought indices

SPI was recommended by the World Meteorological Organization (WMO) in 2009 as the key meteorological drought index that countries can use to track and follow drought conditions [189]. WMO offered guidance for countries seeking to create a standard of drought early warning systems by identifying SPI as an index for general use. Consequently, we have focused particularly in this study, on the SPI drought index.

4.6 Drought monitoring web application

A need has been expressed by the national meteorological service in Algeria to have an easy-to-use tool that would facilitate the monitoring of drought events in the northern part. So, we decided to design a web application to help and contribute in the fight against impacts of droughts, using as a basis the tool developed at the Royal Institute of Meteorology, Belgium (<https://www.meteo.be/en/weather/forecasts/drought>). The aim of this drought web application is to provide nationwide a drought early warning system. The goal is to deliver timely information prior to, or during, the early onset of drought, so that measures can be taken (via threshold triggers) within a drought risk management strategy to mitigate likely drought impacts. Monitoring such a threat involves a comprehensive and structured approach. This application is developed in R programming language with 'shiny' package that makes it easy to build interactive web applications.

National meteorological organizations are gathering the best quality in-situ measurements according to World Meteorological Organization (WMO) standards. At the same time, they are responsible for the preservation and distribution of their archived databases. Although most of this information is restricted to the main synoptic stations and

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only covers basic meteorological parameters, the data itself is typically more reliable than the widely used coarse grid reanalysis products [209]. Due to the sharing of weather information by national meteorological centres, meteorological observations from synoptic stations all over the world are freely accessible to the public (example: FM-12 code established by the WMO). Ogimet.com is one of the well known online databases that archives meteorological data sets based on the National Oceanic and Atmospheric Administration (NOAA) archives. Most of the Ogimet archived data set start around the second part of 1999 and is being updated immediately after new reports are available.

A standard workflow for downloading meteorological data from a server (like ogimet) conventionally using a web browser is to (1) pick a country or station (2) for the specified time period, and (3) measurement period (e.g. hourly/day). As one query is limited to a few tens of rows per request, it could involve manual and repetitive routines to produce a complete data set for a particular region. To overcome the problem of manual data downloading, an automated script is developed to retrieve daily (24h) accumulated precipitation at 06h GMT. The script is written in bash shell and scheduled with a 'cron' Linux service to retrieve data from the server at regular time-steps without the need of manual intervention. However, from 1999-2017, the SYNOP precipitation observations have been extracted from the Ogimet server with a maximum allowed time-step of 50 days at once. So we were forced to run the scripts for a couple of days before we constitute all the needed database.

The script then communicates with 3 other 'R scripts' that compute the drought SPI application. The R scripts are based on the one made by Mr Michel Journee from the RMI Belgium. In contrast to RMI, where the precipitation data are averaged over the whole country to constitute only one data set, in our case, we have modified the scripts to include each national meteorological station data set. The reason is the large differences in precipitation variability (micro-climates) between the different region in Algeria compared to Belgium. Additionally, due

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to the sparse of observation records, spatial data interpolation will be biased.

To achieve this task, we have interpolated the Global Precipitation Climatology Center, daily global land-surface precipitation (GPCC_daily) (Table 4.4) on the weather stations grid points. GPCC_daily precipitation data set are based on SYNOP data shared by national meteorological and hydrological services worldwide and collected by WMO GTS-data centers.

GPCC_daily well cover the missing period (from 1988-1999), in order to have the standard 30 years period of reference. Thereafter, we combined the GPCC_daily data with the SYNOP-Ogimet precipitation to construct the precipitation time series needed for development of the SPI web application. Despite these precipitation data are only available from 1988 to 2013, they are an important source of information for drought analysis at a global scale [210].

4.6.1 The daily SPI

A primary step in the algorithm as explained in the theoretical framework, is the computation of the climatological reference period (precipitation). The reference period that is considered in this application spans from 01/01/1988 to 31/12/2018. The SPI-3 uses a 90-day moving window, SPI-6 uses a 180-day moving window. The web application is made up of two parts:

- The first part, consists of the computation of the SPI as a graphical representation at a daily time step for different periods: 3 months (SPI-3), 6 months (SPI-6) and 12 months (SPI-12) (Figure 4.2). Additionally, forecasting future dry events in a region is very important for water management and risk assessment of drought periods. There are an increasing number of institutions which monitor and forecast droughts using SPI and numerical weather predictions (NWP) deterministic or ensemble output. For instance, the Croatian national meteorological

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service DMHZ combines 2 months of observation with 1 month of European Center for Medium-Range Weather Forecasts (ECMWF) data to predict SPI-3. For the Royal Meteorological Institute (RMI) the SPI forecast is calculated by the cumulative observed precipitation for the last 80 days and the cumulative forecast for the next 10 days. The selected forecasts also come from the European forecast center (ECMWF).

In the present SPI web application, we have followed the same methodology as done at RMI, but with different input parameters. The forecast of the SPI is also calculated with the cumulative rainfall of the last 80 days (synop) and the cumulative precipitation forecasts from the Global Ensemble Forecasting System (GEFS) [211]. The choice of the model comes from the fact that it is freely available. Here (also at the RMI) the SPI is shown as a graphical representation, an easy way to understand the SPI index and its evolution over time. This method allows tracking drought events at a daily time-step as soon as they occur. This constitutes an early warning system for droughts management and monitoring.

The SPI thresholds considered here differ from those used in the definition [184]. Instead of using SPI standard range (-2 to 2) [184], we defined 6 classes (wet/dry, very dry/very wet, extremely dry/extremely wet) as done at RMI. These 6 classes are established with respect to a climatological reference period from 1989 to 2018 and each refers respectively to return periods from 10 to 30 years, 30 to 50, 50 to 100 years (source: <https://www.meteo.dz/>). After that, we compute precipitation anomalies (90-daily moving average anomalies in case of SPI-3) which are then compared to the class they belong to. The curves are given as an average for each of the chosen weather stations of northern Algeria. The Figure (4.2) shows the evolution of possible drought or excess rainfall over time. The red line indicates the position of the total precipitation for the last 90/180/365 days (in percent), compared to normal climatological values (1988-2018). The green and cyan (light blue) lines indicate the evolution of the 90/180/365 days wettest/driest period respectively, at the same date.

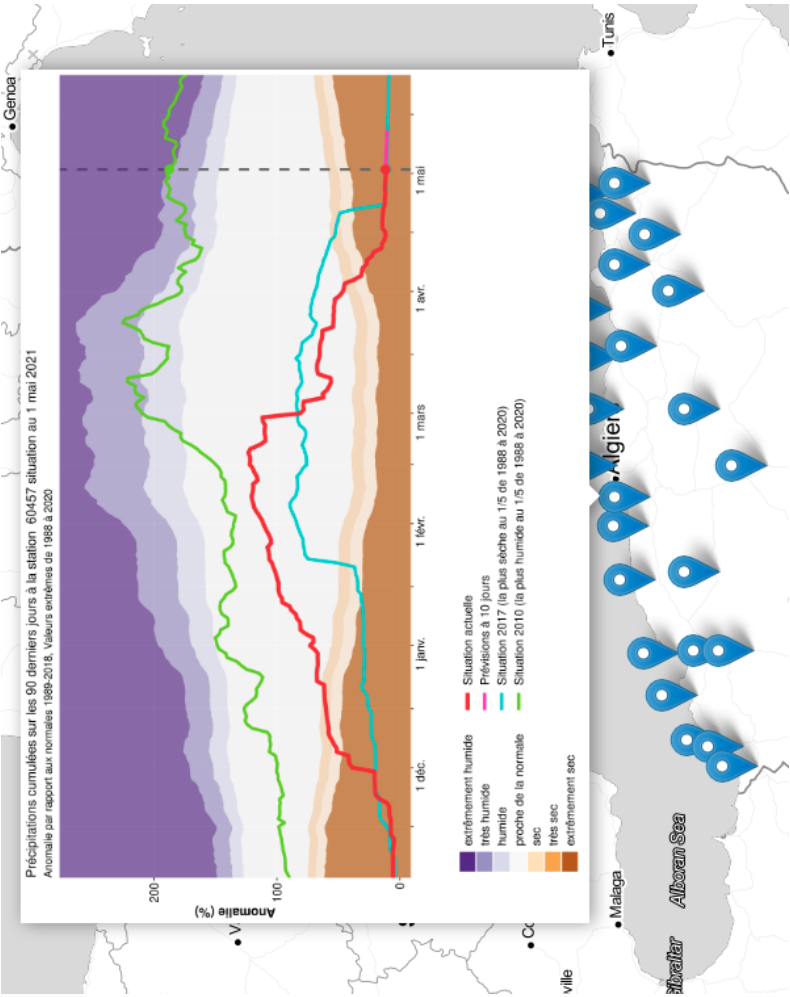


Figure 4.2: Drought web application.
The blue marks represent meteorological land stations in Algeria.

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Data source	Description	Spatial scale (LatxLon)	Temporal scale	Frequency
SYNOP	Daily precipitation totals collected from national SYNOP stations via the WMO's Global Telecommunication System(GTS) [189]	Irregular grid	Since 1999	Daily
GPCC	GPCC Full Data Daily Product: retrieved from daily global land-surface precipitation based on data provided by national meteorological and hydro-logical centers as well as WMO GTS-data [210].	1.0 x 1.0	01-1988 to 12-2013.	1-month delay.
GEFS	GEFS is a weather forecast model made up of 21 separate forecasts, or ensemble members [212]	1.0x1.0	since 2015	Four times a day
CRU TS v4.04	CRU TS (Climatic Research Unit gridded Time Series) is climate data set used worldwide. It is made by interpolation of monthly climate anomalies from large number of meteorological stations observations.	0.5°x 0.5°.	span 1901–2019	updated annually

Table 4.4: Summary of meteorological data sets used in the drought and monitoring application.

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4.6.2 Remote sensing drought monitoring and forecasting

The last part of the drought monitoring and forecasting web application, consist of additional products based on merged satellite, climate models and in-situ data set in order to deliver complete, reliable, up-to-date products for trend analysis and seasonal drought monitoring.

Remote-sensing technology has made it possible to monitor soil moisture and the condition of vegetation across large areas. In locations with a limited number of sampling gauges as in Algeria, remote sensing data may be the only available information source for drought monitoring [213, 214]. Satellite sensors provide a top-down picture by estimating the quantity of electromagnetic radiation (or energy) emitted or reflected from the tops of clouds or from the rain drops themselves. Using satellite measurements, scientists can distinguish between rain, snow, and other forms of precipitation, as well as examine the structure, intensity, and dynamics of storms. Figures 4.3 and 4.4 show precipitation totals and anomalies forecasting over 15 days, based on CHIRPS-GEFS combined observation data.

CHIRPS precipitation data

The United States Geological Survey (USGS) and Climate Hazards Center (CHC) scientists and National Aeronautics and Space Administration (NASA) have developed solutions for producing rainfall maps, especially in areas where surface data is sparse (as in Algeria) [215].

It is important that an evolving drier season must be placed in a historical context so that the severity of rainfall deficits can be quickly evaluated. However, estimates derived from satellite data provide precipitation averages that suffer from biases due to complex terrain and often underestimate the intensity of extreme precipitation events. In contrast, precipitation data produced from weather stations are sparse in rural regions where there are less rain-gauge stations.

CHIRPS data product is a more than 30 year quasi-global rainfall data

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set. It extends latitudes from 50°S-50°N (all longitudes), starting from 1981 to near-present. It incorporates 0.05° resolution satellite imagery with in-situ station data to create gridded rainfall time series for trend analysis and seasonal drought monitoring [215].

CHIRPS-GEFS:

CHIRPS-GEFS is a downscaled and bias-corrected precipitation forecasts that is spatially consistent with CHIRPS products [215]. These data are made up of 5 ensemble members. The CHIRPS-GEFS forecast data is particularly useful for CHIRPS users because it offers 5-day to 15-day GEFS forecast precipitation totals and trends that are consistent with historical CHIRPS data. This function enables a quick assessment of how the most recent forecast could affect the current agro-climatological situation [215].

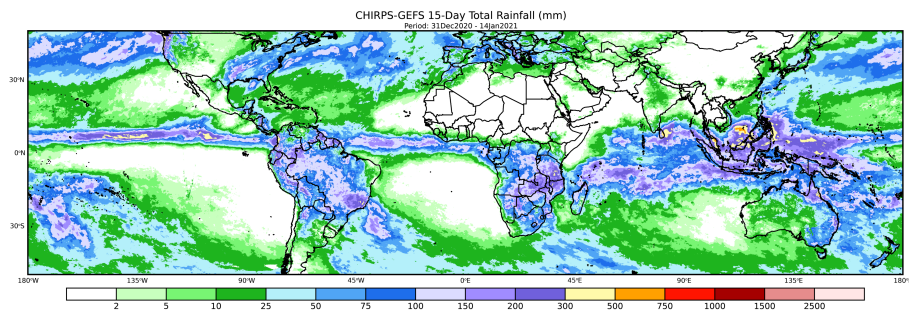


Figure 4.3: CHIRPS-GEFS 15 days total rainfall forecasting. Unit is mm/day [215].

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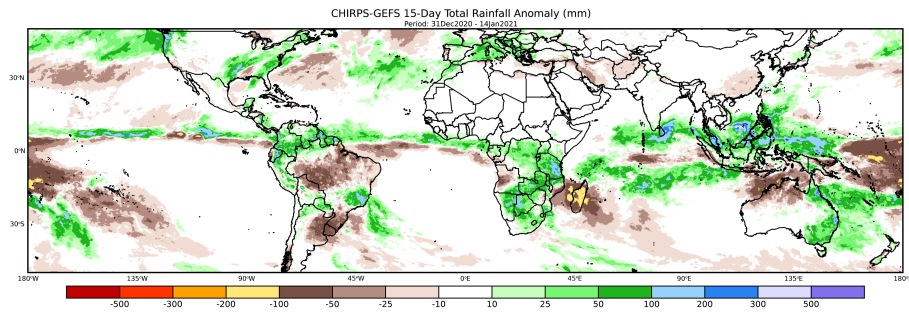


Figure 4.4: CHIRPS-GEFS 15 days rainfall anomalies forecasting. Unit is mm/day [215].

4.6.3 Case study: Drought period in Algeria during 2020-2021

In this section, we apply the developed SPI web application to analyse the drought event occurred in Algeria during 2020-2021.

Since the mid-2020, Algeria has experienced an abnormal decrease in the precipitation amounts. This abnormal meteorological situation suggests a risk of a drought period, especially in the northern country. This event has been also reported in different Media (<https://www.elwatan.com/edition/actualite/vers-une-crise-majeure-de-leau-10-03-2021>) and (<https://www.radioalgerie.dz/news/fr/article/20210306/208148.html>).

The aim of this study is to show the ability of our drought web application to track and monitor drought events as soon as they occur. To do so, we selected three distinct meteorological stations in North Algeria (West, Center and East) to depict the drought's spread and evolution. Figure 4.5 shows the computed SPI indices (3,6 and 12 months) over 2020 to winter 2021. The red line represents the average precipitation total of the last (90, 180, or 360 days). Additionally, they show (in pink color) 10 days SPI forecasts. The colored bands represent the severity

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of the droughts (or the excess of precipitation), ranging from purple (extremely wet) to dark orange (extremely dry).

This particular dry period is well detected in our drought web application (SPI indices, 3,6,12 months)

For the SPI-3 (Figure 4.5), we observe a drought period over the three regions that spans from August 2020 to the mid-December.

In the case of SPI-6, we denote also a drought period with a trend towards normal conditions starting at the same time as for SPI3. however, the situation is still under extreme dry conditions until mid-February. In Figure 4.7, based on SPI-12, the extreme dry period is continuing after February.

These results point out that the SPI-3 (SPI-6) indicates a trend towards normal weather conditions, the SPI-12 in contrast, still warn about the continuous of the drying period. These results are confirmed by comparing with the CHIRPS products, as shows in Figures 4.8 to 4.10.

SPI drought index web app:

Chapter 4. Droughts assessment and monitoring in Algeria

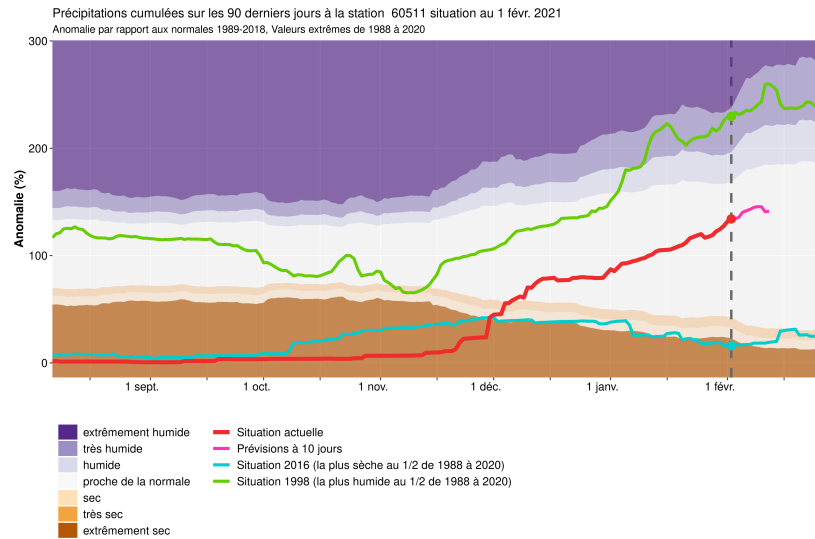


Figure 4.5: Graphical representation of SPI-3 over West Algeria Mostaganem, station (WMO ID: 60511).

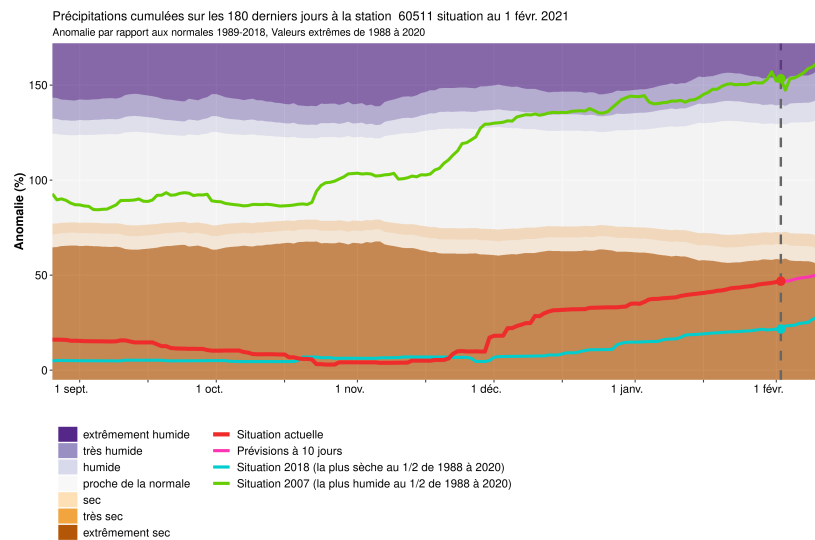


Figure 4.6: Graphical representation of SPI-6 over West Algeria Mostaganem, station (WMO ID: 60535).

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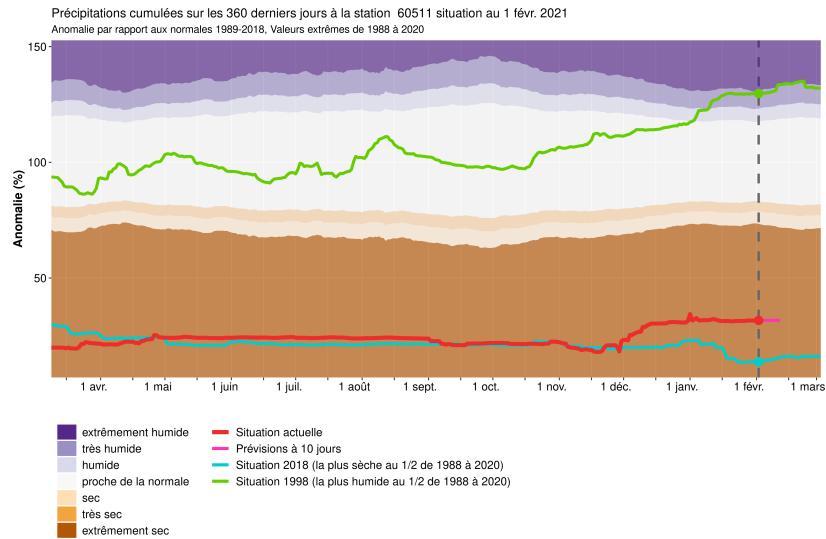


Figure 4.7: Graphical representation of SPI-12 over West Algeria Mostaganem, station (WMO ID: 60535)

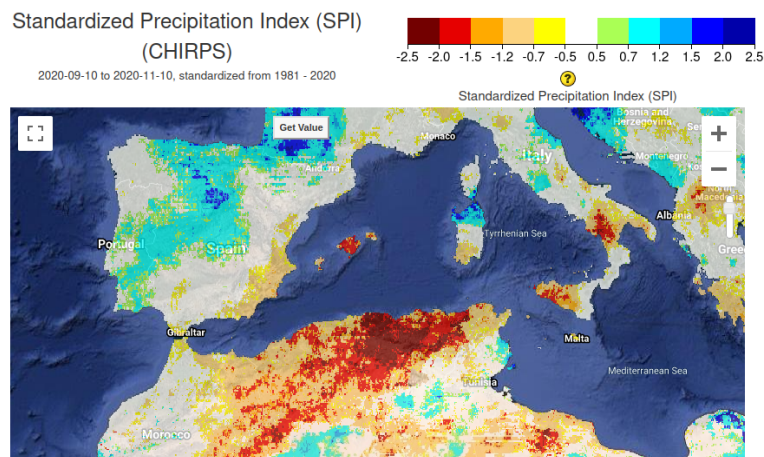


Figure 4.8: Chirps SPI-3 results over North Africa, from 10/09/2020 to 10/11/2020 relative to reference period 1981-2020. source: <https://app.climateengine.org/climateEngine>.

Chapter 4. Droughts assessment and monitoring in Algeria

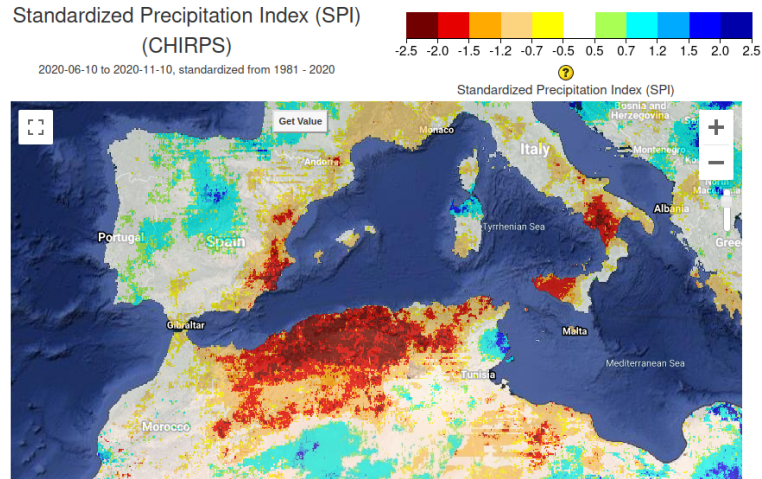


Figure 4.9: Chirps SPI-6 results over North Africa, from 10/06/2020 to 10/11/2020 relative to reference period 1981-2020. source: <https://app.climateengine.org/climateEngine>.

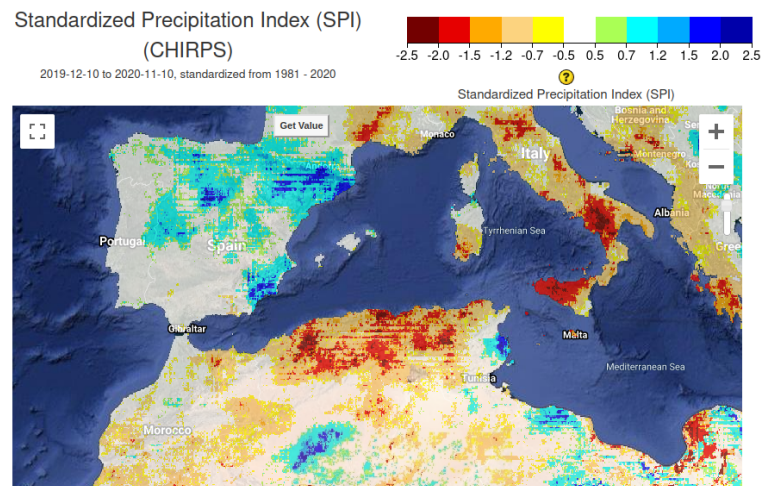


Figure 4.10: Chirps SPI-12 results over North Africa, from 10/12/2019 to 10/11/2020 relative to reference period 1981-2020. source: <https://app.climateengine.org/climateEngine>.

Chapter 4. Droughts assessment and monitoring in Algeria

4.7 Conclusion

In this chapter, we have focused on the study of droughts in Algeria. We have highlighted the importance of droughts management and monitoring systems for the protection and conservation of the natural and economic systems. Indeed, we have developed a web application, suited for droughts monitoring and forecasting. This R-Shiny-based web application goal is to give an easy-to-use tool to help authorities, weather forecasting centers and water supply managers to take rapid responses to monitor and manage extreme climate hazards (drought, floods, ...).

- The first part of the application consists of the computation of SPI drought index by the mean of WMO SYNOP precipitation data. This task is made to run automatically after each new inputs are available (daily precipitations at 06h00 GMT). In addition to SPI index, we have proposed additional tools that may help to better monitor and manage droughts:
- The second part is based on 3 data sets: remote sensing data, in-situ station observations and GEFS forecasts (CHIRPS-GEFS) to create gridded rainfall time series for trend analysis and seasonal precipitation (drought) monitoring and forecasting.

The web application is ready to use (https://droughtwebapp.shinyapps.io/CRU_data/), but could also be easily updated if needed by adding additional tools that may help stakeholders in drought monitoring and management.

CHAPTER 5

Conclusion

5.1 Main achievements

The main objective of the thesis dissertation was to understand the atmospheric mechanisms that underlie precipitation reduction in North Africa region and to make the connection with the processes that have been proposed to explain the changes over the whole Mediterranean basin in present climate and in future climate warming.

As shown in Chapter 1 of this manuscript, due to the particular position of North Africa in the transitional zone between wet European mid-latitude and a dry subtropical climate, its precipitation is characterized by high spatial and temporal variability. Variability of temperatures is more uniform than that of precipitation. It is over the winter periods that both precipitation and temperature vary the most between North Africa countries. This makes the climate of North Africa highly sensitive to atmospheric circulation changes. North Africa is highly vulnerable and has a low adaptive capability. Overall, it is considered one of the most vulnerable regions to global warming [34]. Drought constitutes a recurring climate phenomenon in the region. We have showed that since the past years, North Africa (for instance Algeria) has experienced extended dry periods that had caused serious impact on its economic and social structures, as well as the natural environment (see Chapter 2). Most

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of the studies analyzing precipitation change in the Mediterranean are focused on the whole basin or its Northern (North-western) parts. There are few studies responding to the concerns of North Africa countries, to face the challenges of ongoing and future climate change impacts. With regard to both scientific advance and practical tools for monitoring and forecasting climate extremes, like droughts, inundations, there is a critical need for developing and improving drought warning systems in North Africa countries. To achieve our goal, we have divided the main research objective into three parts.

In the first step, we conducted analyses to evaluate the links between precipitation changes in North Africa and the standard atmospheric modes of variability. As there are only a few works that have addressed the issue of long-term precipitation variability in this region and most were limited to the last 20th century period, this study covers the last millennium (850-2005). We analyzed four atmospheric indices: the North Atlantic Oscillation (NAO), the Western Mediterranean Oscillation (WeMO), the Mediterranean Oscillation (MO) and the zonal wind index (U850), (Chapter 3).

In the second step, we have focused on the analysis of the atmospheric circulation changes, to understand better and quantify the role of atmospheric circulation in winter precipitation decline over North Africa region. The last part was devoted to provide adequate information to help stakeholders and water managers to fight the threat of drought impacts over North Algeria by developing a drought web application. The following parts summarize our conclusions, focusing on our findings, emphasizing in turns the originality, relevance and limitations of our approach.

Previous studies have shown that a large part of the spatial and temporal variability of precipitation in the Mediterranean is closely related to a few large-scale atmospheric modes of variability, mainly the NAO, MO and WeMO. In addition to these standard atmospheric teleconnec-

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tions, we tested the U850 zonal wind index introduced by [40]. Other atmospheric teleconnections have been proposed to have an impact on Mediterranean precipitation variability [38, 39]. However, over our domain of study, the link is found to be weak. Additionally, we showed in Chapter 2 that precipitation variability in the western part of North Africa is the most impacted by the atmospheric teleconnections (NAO, MO and WeMO) than the central and eastern parts, with no significant changes over the three periods considered. In contrast, we found the link between precipitation and U850 to be more robust spatially and temporally than the previous patterns. Our conclusion is consistent with [40], who analyzed CMIP5 climate model projections (2070–2099) to quantify the role of atmospheric circulation in the Mediterranean precipitation response to climate change in the winter season. Among all the indices proposed at the Mediterranean scale, we found that U850 seems the best to explain the variability of precipitation over the last millennium, in North Africa.

By analyzing precipitation and zonal wind time series over the last millennium (850–2005), we found a significant decreasing trend at the end of the period, in the three North African regions. The change point is found around 1800–1850. Comparing with observations and reanalyses, we have found a decreasing trend starting around 1950 depending on the regions of North Africa. The decreasing trends found in precipitation and U850 over 1800–1850 are consistent with the reconstructed PDSI. We also found that the decreasing trend in the PDSI observations is larger than the one computed from the CMIP5/PMIP3 model output or OWDA reconstruction. Additionally, we pointed out, that the correlations results are weak between observed and simulated PDSI as well as between the PDSI simulated by the PMIP3/CMIP5 models, which leads to conclude that internal climate variability has a large influence on the precipitation changes in the North African region before the 20th century.

In Chapter 3, we have analysed the atmospheric moisture budget (P-E) in North Africa over the present period (1990–2005) and in projected climate

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changes (2071-2080), to better understand the atmospheric mechanisms responsible of winter precipitation decline in North Africa, compared to the Mediterranean basin. We have followed the methodology proposed by [57]. They have applied an atmospheric moisture budget decomposition (P-E) in order to explain the physical mechanisms by which the mean and transient circulation drive winter precipitation changes across the Mediterranean region. We applied moisture budget decomposition on ERA-Interim reanalyses, CESM1 (10 members) and CESM2 model over three regions in North Africa. Our findings show that P-E in winter, is sustained by transient eddy moisture convergence and opposed by mean flow moisture divergence, which agrees with [57]. The key role of the mean flow moisture divergence implies that the atmospheric circulation changes are a dominant contributor to future hydrological change in Mediterranean region. Some studies argued that Mediterranean precipitation change can not be exclusively explained by changes in thermodynamics, “wet get wetter, dry get drier” argument [40, 216]. [40] have pointed out that up to 85% decline in Mediterranean averaged precipitation is projected under future climate change, which depends on the atmospheric circulation response as represented by U850 index in North Africa. Additionally, we have found that in 6 climate models (participating to CMIP6 intercomparison project), the link between North Africa precipitation and U850 is robust, under future climate change. [59, 216, 56] found that the impact of low-level circulation under the future climate change, at regional scale is related to the development of an anomalous surface anticyclone over the Mediterranean basin. The magnitude of the strength of the anticyclonic anomaly and of the associated wind field is strongly correlated to the regional precipitation decline across CMIP5 models [40]. [60] showed that the anticyclone circulation in the Mediterranean, may cause the projected drying by increasing atmospheric stability in the region and suppressing Mediterranean cyclones, as shown in [61, 40]. [57] showed that changes in the stationary waves under increasing greenhouse gases, is also connected

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with the high pressure anomalies projected by CMIP5 models above the Mediterranean basin. Another hypothesis suggested that modification of the position of the Hadley Cell may contribute to the Mediterranean drying in the future. However, other studies conclude that such a shift in the Hadley cell is probably not the reason why the Mediterranean would be particularly sensitive to climate projections [68].

In addition to the anticyclone detected over the Mediterranean, our analyses of the atmospheric circulation over North Africa, revealed a high pressure anomaly centered over Morocco in the present period and strengthen in the future. Consequently, these two regional atmospheric features, may be responsible of a large part in future precipitation decline. The high pressure anomalies found over Morocco may be due the high mountain range of the Atlas, which constitutes a barrier to block North Atlantic circulation and thus deviates the westerlies to form the anticyclonic circulation. The climate models used in this thesis study have coarse resolution, thus it is important to make analyses at high spatial resolution to detect the origin of this anticyclone over Morocco. In our results, we have found that, as for the whole Mediterranean drying, the three North Africa sub-regions are linked to changes in moisture convergence by the mean flow. However, the drivers of those changes vary from the western to the eastern parts. For the western part of North Africa, the moisture convergence decreases, as the Atlantic moisture supply weakens because of the anticyclonic circulation over Morocco which increases dry air advection from the Sahara Desert and drives more precipitation decline than in central and eastern parts. To the East, the potential cause of the precipitation decline is due to the eastern Mediterranean anticyclonic circulation, which also advects dry air from the desert to cause aridification in North Africa. In Chapter 4, we have developed a drought web application. This application is developed using the R-shiny package and is made of two distinct components. The first is a graphical representation of the drought SPI index. This is based on the same method as the one used at the Royal Meteorological

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Institute in Belgium (RMI) (see Chapter 4). Our scripts are automated to retrieve precipitation inputs at daily time step from “Ogimet web server” in order to compute the different SPI (3,6 and 12 months). The second component consists on remote sensing precipitation forecasting, using the CHIRPS-GEFS data products. This product is also retrieved automatically using the bash scripts developed in the framework of this thesis. In this application, we have been able to track and monitor a recent drought event in Algeria started from 2020 to the present time. This application covers only Algeria at this time. However, it is easy to implement it in other North Africa countries.

5.2 Perspectives

The complexity of mechanisms involved in precipitation changes in North Africa requires more analyses and simulations with regional climate models at high resolution (RCM). The example of the Coordinated Regional Climate Downscaling Experiment (CORDEX) over Africa is a suited solution for future evaluation of precipitation changes in North Africa. CORDEX program, supported by World Climate Research Program (WCRP), aims to provide downscaled high resolution regional climate change projections. Our goal is to analyse and quantify the contribution of the regional circulation and its impact on precipitation changes in North Africa. As we have seen in this manuscript, GCMs are unable to represent fine-scale and small details of regional climate variability, due to their coarse resolution. The dynamics of the observed anticyclone circulation over Morocco in present and projected changes remains to be understood. [68] have shown that under future changes, precipitation decline will be the highest there across the whole Mediterranean compared to other regions. However, in recent paper [165] argued that high resolution climate simulations indicated that precipitation in Morocco and Andalusia is highly variable in space and time, with large variability over the Rif and Atlas Range peaks. In addition, [56] have integrated

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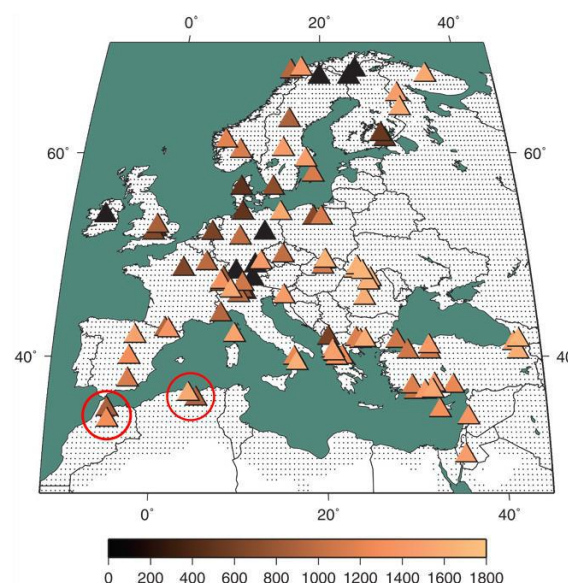
the global drivers of precipitation decline in the Mediterranean into the boundary conditions of their RCM simulations, by modifying the initial and boundary conditions of the climate simulations with the large-scale drivers that may cause the Mediterranean drying [217].

An interesting follow-up of the present work would be to replicate similar analyses as done by [165, 56] to study the impacts of the small scale features (orography, land cover. . .) in the regional atmospheric circulation during the present and projected precipitation decline in Algeria and Tunisia. Even though, winter is the most rainy season over North Africa region, convective precipitation during summer period will also be impacted by future climate warming. In their study, [56] showed that precipitation variability in the Mediterranean region follow different physical processes depending on the season (precipitation formation mechanisms differs from summer to winter). Thus, analysing the summer precipitation variability over North Africa will allow to understand the whole annual precipitation variability in North Africa under future changes and reduce uncertainties.

CHAPTER 6

Supplementary materials

6.1 Supplementary material of chapter 3



FigureS1: Locations of tree ring chronologies in the North African region of Old World Drought Atlas (OWDA). Colors and triangles indicate approximate start dates of various chronologies. Dark brown triangles circled in red represent two grid points (Algeria and Morocco), corresponding to sites where reconstructions of the OWDA are available in the region (map source: [109]).

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Correlation test	CCSM4		MPI-ESM-P		CESM1 (mean)		Reconstructed PDSI	
	Algeria	Morocco	Algeria	Morocco	Algeria	Morocco	Algeria	Morocco
CCSM4	1	1	0.04	0.03	0.08	0.13	0.05	0.07
MPI-ESM-P			1	1	0.16	0.11	-0.005	0.03
CESM1 (mean)					1	1	0.04	0.06
Reconstructed PDSI							1	1

Table S1: Correlations coefficient (none significant) between calculated PDSI from CMIP5 models and reconstructed from OWDA [109].

NAO		WEST		CENTER		EST	
	DATA	COR	P value	COR	P value	COR	P value
Recent Period	Dalware	-5.17E-01	1.47E-03	-4.54E-01	6.22E-03	1.23E-01	4.83E-01
	GPCC	-5.13E-01	1.64E-03	-4.62E-01	5.24E-03	2.06E-01	2.34E-01
	20c RE	-7.13E-01	1.54E-06	-6.43E-01	3.05E-05	-3.87E-01	2.17E-02
	CCSM4	-5.27E-01	2.31E-03	-9.92E-02	5.96E-01	2.72E-01	1.39E-01
	MPI ESM	-2.92E-01	1.10E-01	2.01E-01	2.79E-01	2.68E-01	1.45E-01
	IPSL LR	-3.74E-01	3.85E-02	-1.32E-02	9.44E-01	3.10E-01	9.01E-02
	GISS	3.05E-02	8.71E-01	1.73E-01	3.53E-01	2.18E-01	2.38E-01
	BCC	-4.49E-01	2.13E-02	-3.67E-01	6.49E-02	-2.61E-01	1.98E-01
Past century	CESM1	2.57E-02	8.91E-01	2.50E-02	8.94E-01	-1.45E-01	4.35E-01
	Dalware pc	-4.55E-01	1.94E-06	-1.29E-01	2.02E-01	2.99E-01	2.52E-03
	GPCC pc	-4.43E-01	4.02E-06	-1.29E-01	2.01E-01	3.38E-01	5.91E-04
	20c RE pc	-6.65E-01	4.63E-14	-4.66E-01	1.02E-06	-2.77E-01	5.28E-03
	CCSM4 pc	-2.16E-01	2.97E-02	1.00E-01	3.19E-01	3.38E-01	5.55E-04
	MPI ESM pc	-4.96E-01	1.32E-07	-1.71E-01	8.72E-02	1.72E-01	8.51E-02
	IPSL LR pc	-1.60E-01	1.10E-01	1.65E-01	9.85E-02	3.68E-01	1.51E-04
	GISS pc	-1.77E-01	7.71E-02	8.73E-02	3.85E-01	3.08E-01	1.75E-03
Past millennia	BCC pc	7.89E-02	4.33E-01	8.48E-02	3.99E-01	1.20E-01	2.31E-01
	CESM1 pc	-1.23E-01	2.22E-01	9.09E-04	9.93E-01	3.36E-02	7.40E-01
	CCSM4 pm	-2.60E-01	5.92E-17	-1.84E-02	5.60E-01	1.66E-01	1.30E-07
	MPI ESM pm	-4.64E-01	1.40E-54	-1.06E-01	8.09E-04	7.18E-03	2.32E-02
	IPSL LR pm	-3.30E-01	8.74E-27	2.86E-02	3.67E-01	3.12E-01	3.67E-02
	GISS pm	-1.10E-01	5.12E-04	1.38E-01	1.16E-05	1.38E-01	1.16E-05
	BCC pm	-1.99E-01	2.30E-10	3.35E-02	2.89E-01	3.35E-02	2.89E-01
	CESM1 pm	-3.95E-02	2.13E-01	1.60E-02	6.14E-01	-2.48E-04	9.94E-01

Table S2: Correlations between North Atlantic Oscillation (NAO) index and winter (DJFMA) precipitation for observations, reanalysis, and CMIP5 models over the three periods (1979–2005, 1850–1950, 850–1850). (Red indicates significance, with a p-value <0.05)).

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WeMOI	MODEL	WEST		CENTER		EST	
		COR	P value	COR	P value	COR	P value
Recent period	Dalware	-3.34E-01	4.98E-02	-1.22E-02	9.45E-01	-1.37E-01	4.33E-01
	GPCC	-3.19E-01	6.21E-02	-1.00E-01	5.66E-01	-1.45E-01	4.06E-01
	20c RE	-6.51E-01	2.30E-05	-5.16E-01	1.50E-03	-3.85E-01	2.22E-02
	CCSM4	-4.41E-01	1.29E-02	-1.96E-01	2.91E-01	5.71E-03	9.76E-01
	MPI ESM	-5.30E-01	2.19E-03	-2.66E-01	1.48E-01	-2.60E-01	1.58E-01
	IPSL LR	-4.71E-01	7.45E-03	-2.39E-01	1.95E-01	-7.73E-02	6.79E-01
	GISS	1.77E-02	9.25E-01	-2.11E-01	2.53E-01	-5.32E-02	7.76E-01
	BCC	-6.83E-02	7.40E-01	8.74E-02	6.71E-01	3.18E-01	1.14E-01
Past century	CESM1	-1.34E-01	4.71E-01	-2.49E-01	1.77E-01	-2.45E-01	1.85E-01
	Dalware pc	-4.88E-01	2.60E-07	-4.48E-02	6.58E-01	1.28E-01	2.06E-01
	GPCC pc	-5.04E-01	9.09E-08	-6.20E-02	5.40E-01	1.11E-01	2.70E-01
	20c RE pc	-4.41E-01	4.42E-06	-4.19E-01	1.40E-05	-2.36E-01	1.80E-02
	CCSM4 pc	3.70E-01	1.42E-04	1.81E-01	7.00E-02	6.84E-02	4.97E-01
	MPI ESM pc	-5.46E-01	3.63E-09	-4.72E-01	6.10E-07	-2.72E-01	5.87E-03
	IPSL LR pc	-2.21E-01	2.65E-02	1.53E-02	8.79E-01	-8.65E-02	3.90E-01
	GISS pc	3.91E-01	5.30E-05	2.36E-01	1.76E-02	1.39E-01	1.67E-01
Past millennia	BCC pc	-8.11E-02	4.20E-01	-1.30E-01	1.95E-01	-1.15E-01	2.54E-01
	CESM1 pc	7.61E-02	4.52E-01	-3.72E-04	9.97E-01	1.45E-01	1.51E-01
	CCSM4 pm	-2.88E-01	1.31E-20	-1.23E-01	9.10E-05	-7.76E-02	1.41E-02
	MPI ESM pm	-5.47E-01	5.89E-79	-4.08E-01	2.33E-41	-2.20E-01	1.79E-12
	IPSL LR pm	-3.39E-01	2.88E-28	-1.77E-01	1.67E-08	-1.55E-01	7.93E-07
	GISS pm	3.44E-01	3.03E-29	2.45E-01	3.39E-15	2.45E-01	3.39E-15
	BCC pm	5.23E-01	2.23E-71	4.31E-01	1.24E-46	4.31E-01	1.24E-46
	CESM1 pm	4.96E-02	1.17E-01	3.33E-02	2.93E-01	5.56E-02	7.92E-02

Table S3: Correlations between Mediterranean Oscillation (WeMO) index and winter (DJFMA) precipitation for observations, reanalysis, and CMIP5 models over the three periods (1979–2005, 1850–1950, 850–1850). (Red indicates significance, with a p-value <0.05).

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MOI		WEST		CENTER		EST	
	DATA	COR	P value	COR	P value	COR	P value
Recent period	Dalware	-3.11E-01	6.91E-02	-1.49E-02	9.32E-01	1.89E-01	2.77E-01
	GPCC	-3.33E-01	5.08E-02	-3.34E-02	8.49E-01	2.01E-01	2.46E-01
	20c RE	-6.34E-01	4.40E-05	-5.85E-01	2.27E-04	-2.51E-01	1.45E-01
	CCSM4	-6.42E-01	9.84E-05	-1.38E-01	4.61E-01	3.85E-01	1.69E-02
	MPI ESM	-6.38E-01	1.14E-04	-4.20E-01	1.86E-02	-5.93E-02	7.51E-01
	IPSL LR	-5.26E-01	2.35E-03	-1.62E-01	3.85E-01	3.38E-01	6.25E-02
	GISS	-2.42E-01	1.90E-01	-1.38E-01	1.90E-01	3.72E-01	4.60E-01
	BCC	2.84E-01	1.60E-01	1.48E-01	4.71E-01	-1.48E-02	9.43E-01
Past century	CESM1	-2.17E-01	2.41E-01	-2.72E-01	1.39E-01	-1.67E-02	9.29E-01
	Dalware pc	-5.01E-01	1.12E-07	-1.59E-01	1.13E-01	2.83E-01	4.28E-03
	GPCC pc	-5.01E-01	1.08E-07	-1.63E-01	1.04E-01	2.84E-01	4.14E-03
	20c RE pc	-5.07E-01	7.14E-08	-3.92E-01	5.43E-05	-1.45E-01	1.50E-01
	CCSM4 pc	-5.63E-01	8.62E-10	-9.10E-02	3.65E-01	4.88E-01	2.27E-07
	MPI ESM pc	-5.90E-01	8.59E-11	-5.44E-01	4.23E-09	-3.63E-02	7.18E-01
	IPSL LR pc	-5.00E-01	1.02E-07	-2.47E-01	1.26E-02	2.67E-01	7.02E-03
	GISS pc	-3.54E-01	2.76E-04	-1.01E-01	3.13E-01	2.71E-01	6.18E-03
Past millennia	BCC pc	3.97E-02	6.93E-01	-4.04E-02	6.88E-01	-6.57E-02	5.14E-01
	CESM1 pc	2.32E-02	8.19E-01	4.73E-02	6.40E-01	1.07E-01	2.88E-01
	CCSM4 pm	-4.00E-01	8.37E-40	-1.29E-01	4.26E-05	3.13E-01	3.43E-24
	MPI ESM pm	-5.82E-01	1.21E-91	-4.87E-01	8.65E-61	1.36E-01	1.57E-05
	IPSL LR pm	-4.83E-01	1.50E-59	-1.88E-01	2.08E-09	3.65E-01	6.33E-33
	GISS pm	-4.86E-01	1.59E-60	-2.75E-01	9.19E-19	-2.75E-01	9.19E-19
	BCC pm	-4.85E-01	4.38E-60	-2.42E-01	9.04E-15	-2.42E-01	9.04E-15
	CESM1 pm	-1.36E-02	6.68E-01	-2.64E-02	4.04E-01	-5.70E-02	7.19E-02

Table S4: Correlations between Mediterranean Oscillation (MO) index and winter (DJFMA) precipitation for observations, reanalysis, and CMIP5 models over the three periods (1979–2005, 1850–1950, 850–1850). (Red indicates significance, with a p-value <0.05).

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U850		WEST		CENTER		EST	
	DATA	COR	P value	COR	P value	COR	P value
	20c RE	3.84E-01	2.50E-02	1.82E-01	3.02E-01	7.53E-02	6.72E-01
	CCSM4	7.04E-01	9.91E-06	6.59E-01	5.63E-05	2.10E-01	2.56E-02
	MPI_ESM	7.25E-01	3.89E-06	5.95E-01	4.17E-04	1.84E-01	3.22E-02
Present period	IPSL_LR	4.08E-01	2.26E-02	5.00E-01	4.20E-03	9.48E-02	6.12E-01
	GISS	5.56E-01	1.15E-03	5.56E-01	5.41E-02	8.91E-02	6.34E-01
	BCC	1.26E-01	5.41E-01	1.26E-01	7.53E-01	-9.17E-03	9.65E-01
	CESM	6.50E-01	7.56E-05	5.21E-01	2.66E-03	3.09E-01	4.04E-02
	20c RE pc	1.81E-01	7.10E-02	7.98E-02	4.30E-01	7.47E-02	4.60E-01
	CCSM4 pc	6.80E-01	5.55E-15	5.53E-01	2.02E-09	1.12E-01	2.64E-01
Past century	MPI_ESM pc	7.16E-01	4.16E-17	6.82E-01	4.05E-15	2.67E-01	6.87E-03
	IPSL_LR pc	1.45E-01	1.47E-01	1.79E-01	5.27E-02	1.04E-02	9.17E-01
	GISS pc	5.42E-01	4.94E-09	3.45E-01	4.14E-04	2.11E-01	3.39E-02
	BCC pc	-1.98E-02	8.44E-01	-6.07E-02	5.46E-01	-7.17E-02	4.76E-01
	CESM pc	7.07E-01	1.48E-16	6.11E-01	1.11E-11	2.78E-01	4.80E-03
	CCSM4 pm	6.54E-01	2.21E-123	5.97E-01	9.70E-98	1.41E-01	8.01E-06
	MPI_ESM pm	2.63E-01	2.48E-13	2.45E-01	1.06E-11	6.75E-02	6.72E-01
Past millennia	IPSL_LR pm	7.31E-01	2.53E-151	7.31E-01	2.49E-69	1.31E-01	7.66E-05
	GISS_pm	6.27E-01	3.00E-110	6.27E-01	9.87E-21	1.92E-01	9.17E-10
	BCC pm	6.70E-01	3.29E-131	5.04E-01	1.09E-65	3.51E-01	2.27E-30
	CESM pm	2.53E-02	1.47E-16	2.71E-03	9.32E-01	4.82E-02	1.28E-01

Table S5: Correlations between U850 index and winter (DJFMA) precipitation for observations, reanalysis and CMIP5 models over the three period (1979-2005, 1850-1950, 850-1850). pc: past century, pm: past millennium.

Chapter 6. Supplementary materials

6.2 Supplementary material of chapter 4

1. Link between U850 index and winter precipitation in North Africa, in the period 2015-2100.

We analyzed the correlations between U850 index and CMIP6 model precipitation data, under the future climate change (2015-2100): The Table 6.1 shows Pearson correlations results between U850 index and precipitations data for 9 CMIP6 models:

Model name	Institution	Period covered	Atmospheric Resolution Horizontal/Vertical	Reference
CESM2	National Center for Atmospheric Research (NCAR), USA	2015-2100	192x288 32	[218]
BCC-CSM2-MIR	Beijing Climate Center, Beijing, China		360 x 232 40	[219]
IPSL-CM6A-LR	Institut Pierre Simon Laplace, Paris 75252, France		144 x 143 79	[220]
GISS-E2-1-G	Goddard Institute for Space Studies, New York, USA.		144x90 40	[221]
ACCESS-CM2	Commonwealth Scientific and Industrial Research Organisation, Australia; ARCCSS (Australian Research Council Centre of Excellence for Climate System Science)		144x192 50	[222]
MIROC-ES2L	Japan Agency for Marine-Earth Science and Technology, Japan; AORI (Atmosphere and Ocean Research Institute, The University of Tokyo), NIES (National Institute for Environmental Studies, Japan), and R-CCS (RIKEN Center for Computational Science, Japan)		128 x 64 40	[223]
GFDL-ESM4	National Oceanic and Atmospheric Administration, Geophysical Fluid Dynamics Laboratory, Princeton, USA		720 x 576 75	[224]
HadGEM3-GC31-LL	Met Office Hadley Centre, Fitzroy Road, Exeter, Devon, UK		192 x 144 85	[225]
NorESM2-LM	NorESM Climate modeling Consortium consisting of CICERO (Center for International Climate and Environmental Research, Oslo);		144 x 96 32	[226]
	MET-Norway (Norwegian Meteorological Institute, Oslo), NERSC (Nansen Environmental and Remote Sensing Center, Bergen); NILU (Norwegian Institute for Air Research, Kjeller), UiB (University of Bergen, Bergen 5007); UiO (University of Oslo) and UNI (Uni Research, Bergen)			

Table 6.1: Modelling centers, characteristics, and references of Coupled Model Intercomparison Project(CMIP6) models used in this study.

Chapter 6. Supplementary materials

1.1. Correlation between U850 index and winter precipitation data under the future climate change (2015-2100)

The Figure 6.1 shows the significant correlation's coefficients between U850 index and 9 CMIP6 model data (CESM2, BCC-CSM2-MR, IPSL-CM6A-LR, GISS-E2-1-G, ACCESS-CM2, MIROC-ES2L, GFDL-ESM4, HadGEM3-GC31-LL and NorESM2-LM). The precipitation data are averaged over North Africa region (10W-10E and 32N-36.5N). These correlation results are highly positive (> 0.5) and significant at 95% (Pearson's confidence interval). This result confirms the robustness of the link found by [147] between the U850 index and CMIP5 precipitation data over the last millennium.

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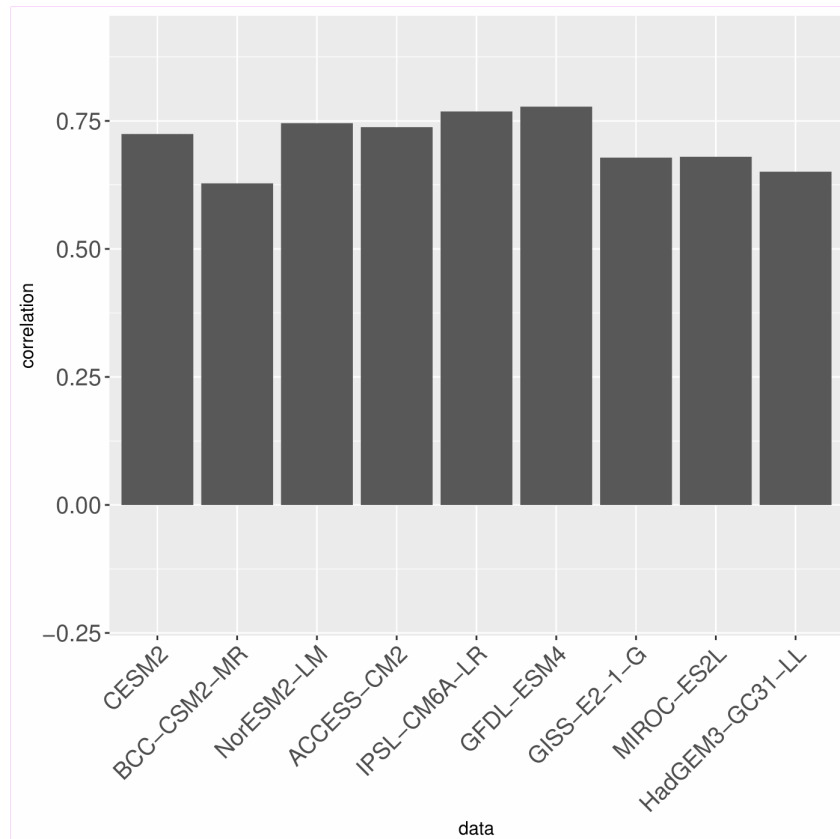


Figure 6.1: Pearson's correlation coefficients between U850 index and winter precipitation data, from 9 models participating in CMIP6 (CESM2, BCC-CSM2-MR, IPSL-CM6A-LR, GISS-E2-1-G, ACCESS-CM2, MIROC-ES2L, GFDL-ESM4, HadGEM3-GC31-LL and NorESM2-LM), in North Africa (-10W-10E and 32N-36.5N) during 2015-2100 (under the SSP585 scenario). These correlations are significant at 95% confident interval).

1.2. Changes in both CMIP6 precipitation and U850 index under future climate:

The Figure 6.2 shows a future wintertime precipitation and U850 index reduction under the scenario SSP585 in North Africa region, using

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CMIP6 output data. The decreasing trend observed both in winter precipitation and U850 index (Figure 6.2.a and b) is significant according to the modified Mann–Kendall [111] test for trend detection. These results confirm the future aridification of the North Africa climate, pointed out by numbers of scientists [68, 54, 56, 57], notably the results found by [40] who showed decreasing trend in both U850 index and winter precipitation over the Mediterranean region, using CMIP5 RCP8.5 climate projections. The decreasing trends in U850 index and North Africa wintertime precipitation confirm and extend that found during last century (1950-2005) by [147] over North Africa.

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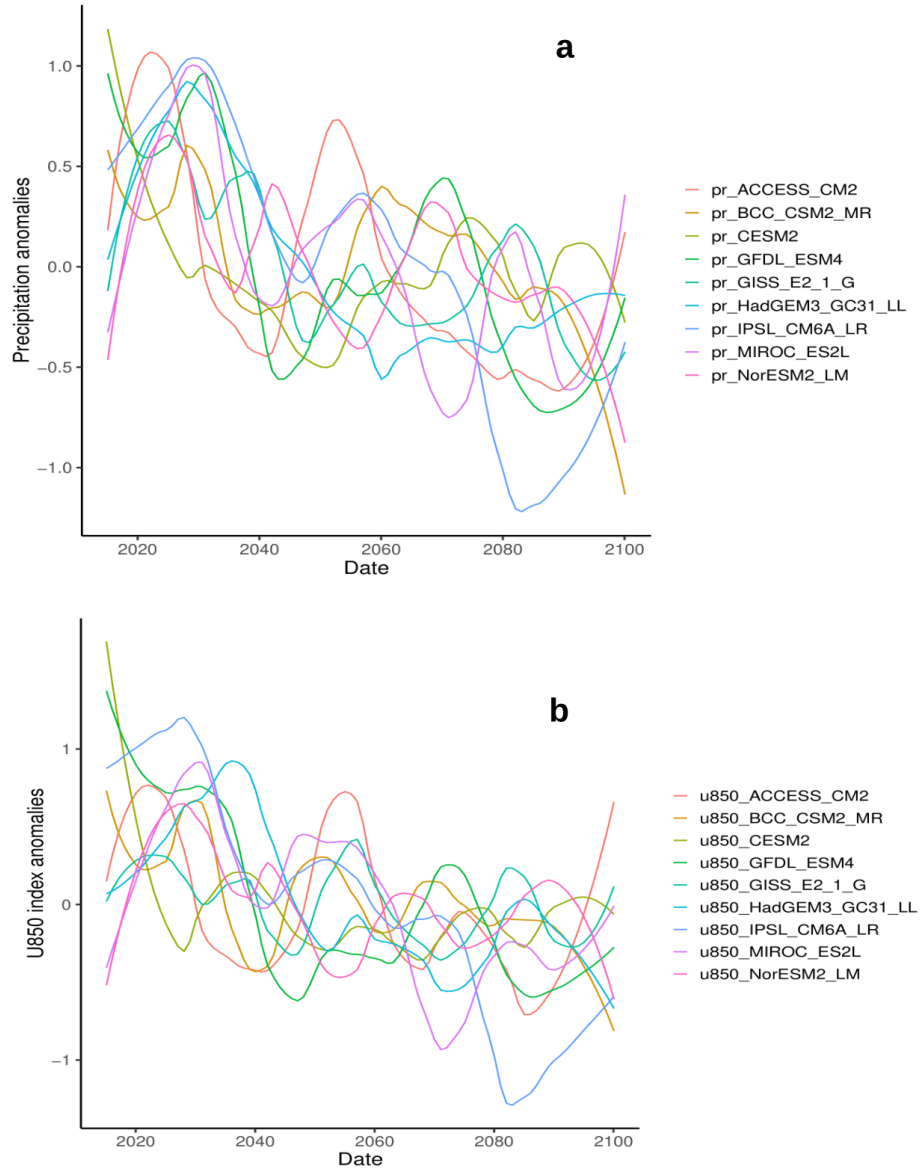


Figure 6.2: Loess smoothed (30 years) of 9 models participating in CMIP6 (CESM2, BCC-CSM2-MR, IPSL-CM6A-LR, GISS-E2-1-G, ACCESS-CM2, MIROC-ES2L, GFDL-ESM4, HadGEM3-GC31-LL and NorESM2-LM) precipitation time series anomalies and calculated U850 index in North Africa, over 2015-2100 period under the SSP585 scenario.).

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2- Moisture budget decomposition terms:

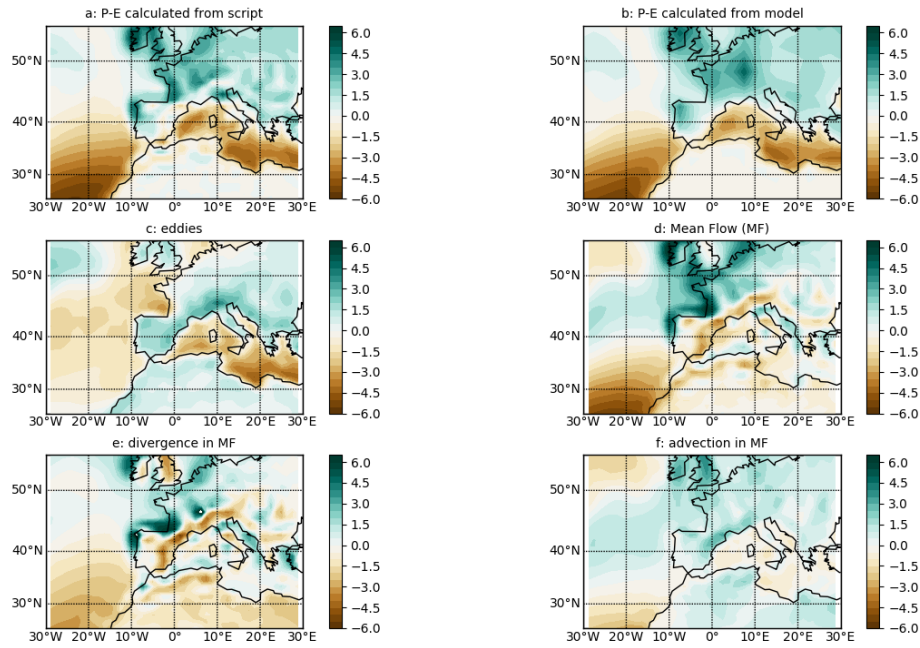


Figure 6.3: The November-April climatological moisture budget from the ensemble mean CESM1-CAM5 model over 1990-2005: (a) P-E (from MB script) (b) P-E from model (c) transient eddy moisture convergence (d) the moisture convergence by the mean flow and its components due to (e) mass divergence and (f) advection. Units are mm/day.).

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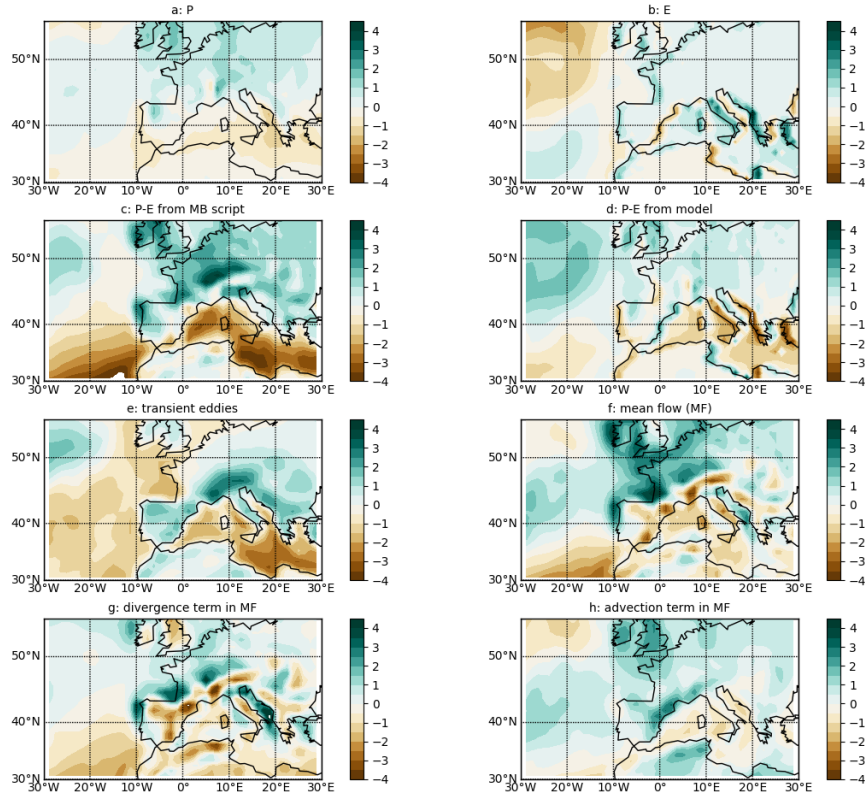


Figure 6.4: Changes from winter (November-April) 1990-2005 to 2071-2100 period in the component of the moisture budget in CESM2 climate model. The changes in (a) P, (b) E, (c) P-E from scripts, (d) P-E from model, (e) change in transient eddy moisture convergence (f) The change in moisture convergence by the mean flow and its component changes due to (g) the advection term, (h) mass divergence. Units are mm/day.

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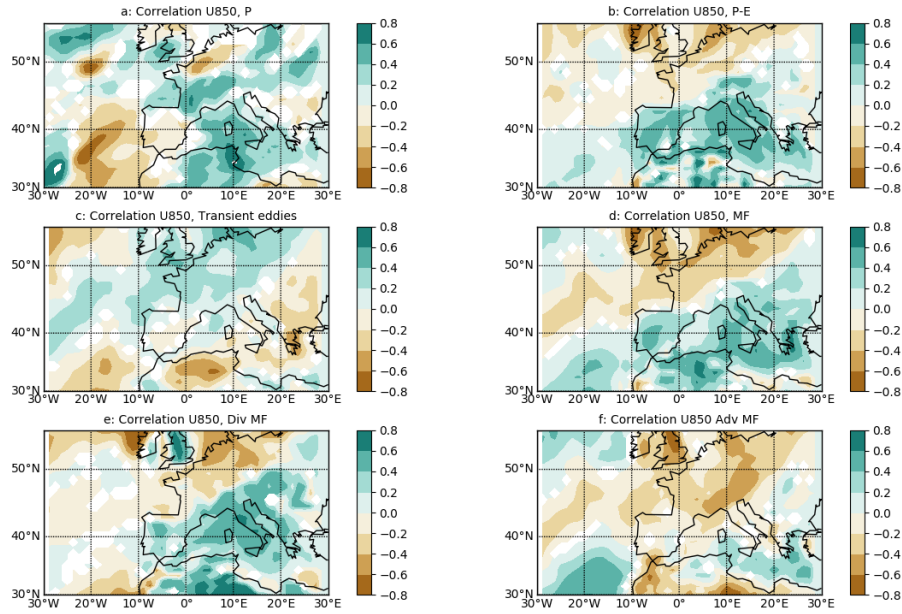


Figure 6.5: Significant correlations between CESM2: U850 index and moisture budget terms over North Africa, winter (November-April) 2071-2100, according to Pearson correlation test at 95%).)

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6.3 Annex

Definition of wind barbs

On weather maps, wind barbs are represented as a graphical device used to depict the intensity and direction of the wind at a specific location.

How to read wind barbs?

The barbs are made up of a bare point and a tail made up of lines and/or flags. The number of lines and/or flags on the tail denote the strength of the wind. A complete line is equal to 10 knots ($1kt \approx 0.5m/s$), while a half-line is equal to 5 kts and 50 kts equals a flag (triangle). To calculate the wind's magnitude, add all the items together. A circle represents very low wind. The barb's orientation on a 360-degree compass-like circle determines the wind direction. The wind blows from the tail to the bare point, (source:<https://a.atmos.washington.edu/wrfrt/descript/definitions/windbarbs.html>, accessed on 20/03/2021).

Example:

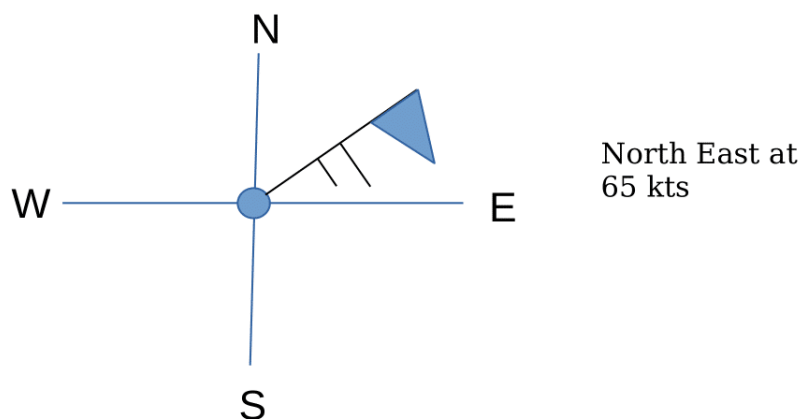


Figure 6.6: Wind barbs direction and speed example.

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