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**Evolution of extreme precipitation and associated
hydrological responses in Belgium using the MAR
regional climate model and the WOLFHydro
modelling framework**

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Abstract (EN)

Five years later, parts of Belgium are still recovering from the impacts of the July 2021 floods. These have been associated with record-breaking precipitation totals at various locations, as well as anomalous river discharges. In a context of global warming, this event raises several questions concerning the future of hydroclimatic risk in Belgium. This thesis investigates how Regional Climate Models and hydrological models can be used to assess expected changes in the intensity and frequency of extreme precipitation events and subsequent peak discharges. To address these questions, it combines statistical tools such as Extreme Value Analysis, bias adjustment of climatic inputs, and large-scale atmospheric analyses.

As a first step, Extreme Value Analysis is applied to precipitation outputs from the Regional Climate Model MAR to study the evolution of extreme precipitation across accumulation durations ranging from 1 to 72 hours. MAR is one of the model members of the CORDEX-Be.2 project (Coordinating Downscaling Model expertise in Belgium), which aims to inform Belgian stakeholders about the evolution of climate risks in Belgium. This analysis also provides an evaluation of MAR's ability to represent extreme precipitation events and thereby informs its future users about its performance.

Using future IPCC socio-economic scenarios and a set of six CMIP6 Earth System Models (ESMs), our results show that extreme precipitation events broadly follow Clausius–Clapeyron scaling, with a 5–7% increase in intensity per additional degree of global warming across all accumulation durations. However, regional variations emerge, with some locations exhibiting scaling rates exceeding the Clausius–Clapeyron relationship. Moreover, depending on the ESM used to force MAR, the intensity and timing of the most severe projected extreme precipitation events can vary substantially.

This variability in the timing and intensity of extreme events, reflecting the natural variability of the climate system, further motivates the study of a particular class of large-scale weather patterns: cut-off lows (COLs). These high-altitude depressions can lead to severe weather and have been associated with the July 2021 floods in Belgium, as well as other high-impact events across Europe, such as the October 2024 Valencia

floods. Despite their importance in generating severe weather, studies investigating the contribution of these systems to extreme precipitation events remain scarce.

Therefore, as a second step, using climate reanalyses, ESM outputs, and MAR simulations, a detection and association algorithm is developed to investigate the evolution of COL-associated precipitation events in the past and future at both European and Belgian scales. The results show that the timing of these events exhibits substantial natural variability and that COLs contribute to a large fraction of the most severe extreme precipitation events in both Europe and Belgium. Moreover, changes in the number of COL-associated events become significant only when high precipitation thresholds are considered. This suggests that their evolution is not primarily driven by an increase in the absolute number of COLs, but rather by changes in the intensity of precipitation associated with these systems and other processes discussed throughout this thesis.

Finally, to assess how these climatic evolutions translate into the hydroclimatic system, the MAR-WOLFH_{ydro} modelling chain is used. WOLFH_{ydro} is a hydrological modelling package allowing the use of various hydrological models. Four hydrological models are forced by future MAR outputs to assess the sensitivity of the hydroclimatic response to the structure of the hydrological models themselves. Tested over the Ourthe catchment, these hydrological model simulations reveal that the scaling of peak discharges with global warming can exceed the scaling of extreme precipitation. They also highlight the influence of natural climate variability, with the occurrence of the most extreme flood events remaining largely stochastic.

Therefore, this thesis contributes to improving our understanding of future hydroclimatic risks by combining expertise in regional climate and hydrological modelling. At the same time, it highlights the uncertainties inherent in both the natural climate system and the modelling frameworks used to investigate its future evolution.

Résumé (FR)

Cinq ans après les inondations de juillet 2021, certaines régions de Belgique se remettent encore de leurs impacts. Ces inondations ont été associées à des cumuls de précipitations records à différents endroits, ainsi qu'à des débits fluviaux exceptionnellement élevés. Dans un contexte de réchauffement climatique, cet événement soulève plusieurs questions concernant l'évolution future des risques hydroclimatiques en Belgique. Cette thèse étudie comment les modèles climatiques régionaux et les modèles hydrologiques peuvent être utilisés pour évaluer les changements attendus dans l'intensité et la fréquence des précipitations extrêmes et des débits de pointe associés. Pour répondre à ces questions, elle combine des outils statistiques tels que l'analyse des valeurs extrêmes, la correction des biais des données climatiques et l'analyse des circulations atmosphériques à grande échelle.

Dans un premier temps, une analyse des valeurs extrêmes est appliquée aux précipitations simulées par le modèle climatique régional MAR afin d'étudier l'évolution des distributions des précipitations extrêmes pour des durées d'accumulation allant de 1 à 72 heures. MAR fait partie des modèles du projet CORDEX-Be.2 (*Coordinating Downscaling Model expertise in Belgium*), qui vise à informer les parties prenantes belges sur l'évolution des risques climatiques en Belgique. Cette analyse permet également d'évaluer la capacité de MAR à représenter les précipitations extrêmes et ainsi d'informer ses futurs utilisateurs sur ses performances.

À l'aide de scénarios socio-économiques futurs du GIEC et d'un ensemble de six modèles du système Terre (ESMs) issus de CMIP6, nos résultats montrent que les précipitations extrêmes suivent globalement une relation de type Clausius–Clapeyron, avec une augmentation de leur intensité de 5–7 % par degré supplémentaire de réchauffement climatique, pour l'ensemble des durées d'accumulation considérées. Toutefois, des variations régionales apparaissent, certains endroits présentant des taux d'augmentation supérieurs à la relation de Clausius–Clapeyron. De plus, selon l'ESM utilisé pour forcer MAR, l'intensité et le moment d'occurrence des événements de précipitations extrêmes les plus sévères projetés peuvent varier considérablement.

Cette variabilité du timing et de l'intensité des événements extrêmes, reflétant la

variabilité naturelle du système climatique, motive l'étude d'une classe particulière de configurations météorologiques à grande échelle : les gouttes froides (*cut-off lows*, COLs). Ces dépressions d'altitude peuvent engendrer des conditions météorologiques sévères et ont notamment été associées aux inondations de juillet 2021 en Belgique, ainsi qu'à d'autres événements majeurs en Europe, tels que les inondations de Valence en octobre 2024. Malgré leur importance dans la génération de phénomènes météorologiques sévères, les études portant sur la contribution de ces systèmes aux événements de précipitations extrêmes restent rares.

Dans un deuxième temps, à l'aide de réanalyses climatiques, de sorties d'ESMs et de simulations MAR, un algorithme de détection et d'association est développé afin d'étudier l'évolution passée et future des événements de précipitations associés aux COLs, à l'échelle européenne et belge. Les résultats montrent que le moment d'occurrence de ces événements présente une variabilité naturelle importante et que les COLs contribuent à une part importante des événements de précipitations extrêmes les plus sévères, tant en Europe qu'en Belgique. De plus, les changements dans le nombre d'événements associés aux COLs deviennent significatifs uniquement lorsque des seuils élevés de précipitations sont considérés. Cela suggère que leur évolution n'est pas principalement liée à une augmentation du nombre absolu de COLs, mais plutôt à des changements dans l'intensité des précipitations associées à ces systèmes et à d'autres processus discutés dans cette thèse.

Enfin, afin d'évaluer comment ces évolutions climatiques se traduisent dans le système hydroclimatique, la chaîne de modélisation MAR-WOLFH_{ydro} est utilisée. Quatre modèles hydrologiques sont forcés par les sorties futures de MAR afin d'évaluer la sensibilité de la réponse hydroclimatique à la structure des modèles hydrologiques eux-mêmes. Testés sur le bassin de l'Ourthe, ces modèles hydrologiques montrent que l'évolution des débits de pointe avec le réchauffement climatique peut être supérieure à celle des précipitations extrêmes. Ils mettent également en exergue l'influence de la variabilité naturelle du climat, l'occurrence des événements de crue les plus extrêmes demeurant largement stochastique.

Ainsi, cette thèse contribue à améliorer notre compréhension des risques hydroclimatiques futurs en combinant les approches de modélisation climatique régionale et de modélisation hydrologique. Elle met également en évidence les incertitudes inhérentes à la fois au système climatique naturel et aux modèles utilisés.

Abbreviations

AM	Annual Maxima
CC	Climate Change
CMIP	Coupled Model Intercomparison Project
COL	Cut-Off Low
CORDEX	Coordinated Regional Climate Downscaling Experiment
DJF	December–January–February
ESM	Earth System Model
EVA	Extreme Value Analysis
ERA5	Fifth-generation ECMWF Reanalysis
GEV	Generalized Extreme Value
GMST	Global Mean Surface Temperature
GPD	Generalized Pareto Distribution
HBV	Hydrologiska Byråns Vattenbalansavdelning
JJA	June–July–August
MAR	Modèle Atmosphérique Régional
NSE	Nash–Sutcliffe Efficiency
POT	Peaks Over Threshold
RCM	Regional Climate Model
RCP	Representative Concentration Pathway

RMI	Royal Meteorological Institute
SAC-SMA	Sacramento Soil Moisture Accounting
SPW	Service public de Wallonie
SSP	Shared Socioeconomic Pathway
VHM	Vergemeend Conceptueel Hydrologisch Model
WOLFHydro	WOLFHECE hydrological modelling component

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CHAPTER 1

Introduction and Background

1.1 Motivation and context

Recent high-impact precipitation events have raised several questions in the context of global warming. In July 2021, eastern Belgium, western Germany, and the southern Netherlands experienced severe human and material losses (Dewals et al., 2021). In Belgium, three-day rainfall totals accounted for 20-25% of the annual average (Erpicum et al., 2017; Journée et al., 2023), reaching 300 mm at some locations. The catchments of the Vesdre and the Ourthe rivers were the most impacted in Belgium (Dewals et al., 2021) and received respective average cumulated precipitation of 178 and 123 mm between the 13th and the 15th of July 2021 (Fig. 1.1b).

As a result, the associated flooding led to water depths exceeding 4 m in some areas and caused 37 deaths (Dewals et al., 2021). In the Ourthe and Vesdre catchments, record-breaking peak discharges of 542 and 677 m³ s⁻¹ were recorded at the Tabreux and Chaudfontaine stations (Service public de Wallonie (SPW), 2026), respectively. These values correspond to 3.38 and 5.63 times the respective bankfull discharges (Van Campenhout et al., 2020).

The return period of the associated precipitation event was estimated at approximately 400 years based on historical precipitation time series (Tradowsky et al., 2023). At the synoptic scale, this precipitation event was driven by a cut-off low (COL), i.e., a high-altitude depression that becomes detached from the main westerly flow as a result of the elongation of jet-stream meanders (Palmén, 1949; Palmén and Newton, 1969; ECMWF, 2021) (Fig. 1.1a).

These systems have also been associated with other recent record-breaking floods across Europe. Notably, the Valencia region in 2024 (Fekete et al., 2025) and Central Europe during Storm Boris (Athanasé et al., 2024) experienced severe impacts. In a

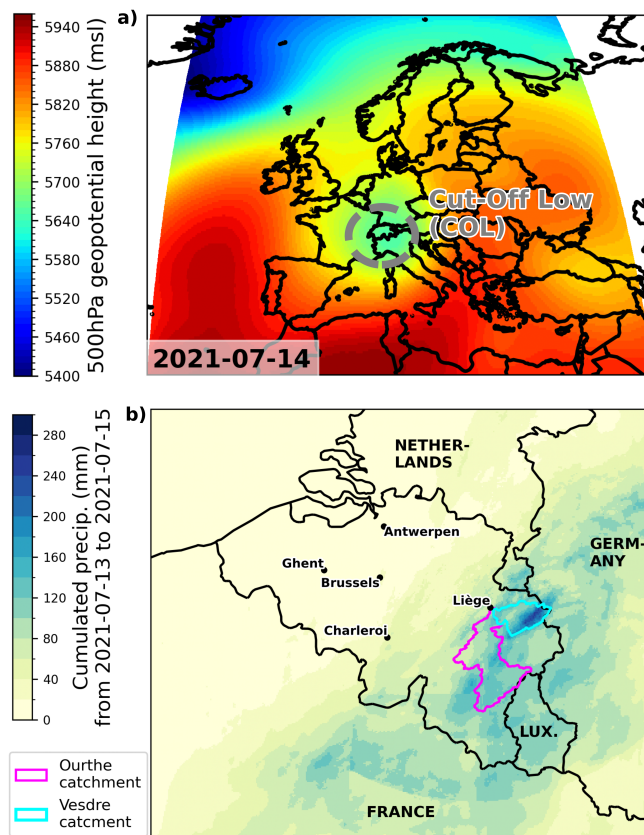


Figure 1.1: (a) Synoptic situation on 13 July 2021 derived from the ERA5 reanalysis and (b) associated precipitation observed by radar from 13 to 15 July 2021 (Journée et al., 2023).

warming climate, extreme precipitation events and associated flooding are projected to become more frequent and more intense globally (Seneviratne et al., 2021). Nevertheless, the magnitude and direction of these changes can vary considerably across regions (Bertola et al., 2019; Coppola et al., 2021; Rajczak and Schär, 2017).

Hence, the drivers of extreme precipitation differ widely depending on the geographical and climatic context (Aalbers et al., 2017). Therefore, we first place Belgium in the broader European context of global warming by describing the expected changes in temperature, precipitation, and extreme precipitation at both the European and Belgian scales. In a second step, we review the existing literature on the associated hydrological impacts.

1.2 Global warming in Europe and Belgium

Global temperatures are rising due to anthropogenic greenhouse gas emissions. Although water vapor and CO₂ are the main contributors to the greenhouse effect, human-induced warming is dominated by CO₂ emissions (IPCC, 2021).

At the Mauna Loa Observatory, atmospheric CO₂ concentrations have been measured continuously since the 1950s, yielding the well-known Keeling curve. From 2022, concentrations exceed 420 ppm, representing an increase of about 100 ppm since the start of the record. Preindustrial concentrations, inferred from polar ice cores, were close to 280 ppm (Scripps Institution of Oceanography, 2025). This increase in greenhouse gases

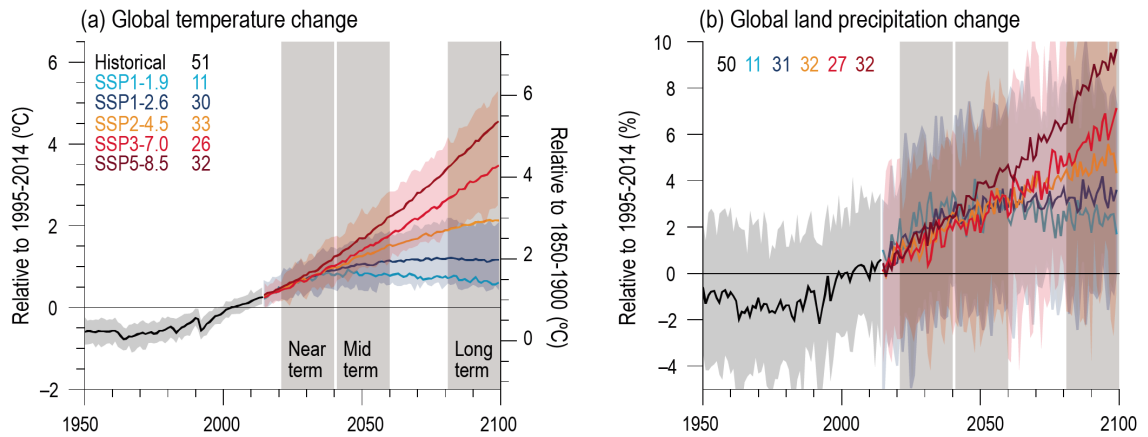


Figure 1.2: CMIP6 projected changes in (a) global mean temperature and (b) global land precipitation, based on historical simulations and future projections under various emission scenarios (Lee et al., 2021).

has already driven a global mean warming exceeding 1 K relative to the IPCC reference period 1850-1900 (IPCC, 2021; Copernicus, 2025).

Future global warming will therefore depend strongly on anthropogenic emissions. The latest generation of Earth System Models (ESMs), developed within the sixth phase of the Coupled Model Intercomparison Project (CMIP6), are widely used in IPCC assessments to simulate the evolution of Earth's climate at the global scale. With each new generation, the range of processes represented has expanded, reflecting both scientific progress and advances in computational performance (Meehl et al., 2020). These models are now capable of simulating complex feedbacks, including land-atmosphere and ocean-atmosphere interactions.

Using a range of socioeconomic scenarios, known as the Shared Socioeconomic Pathways (SSPs), CMIP6 models project a global mean warming between 2 K and 5.5 K by the end of the 21st century, under the most optimistic (SSP1-1.9) and most pessimistic (SSP5-8.5) scenarios, respectively (Fig. 1.2a). However, substantial inter-model variability remains, primarily due to differences in climate sensitivity among models (Lee et al., 2021).

For precipitation, inter-model variability is even higher. Under the SSP5-8.5 scenario, the average global land precipitation is projected to increase by about 10% (Fig. 1.2b), but substantial regional differences exist (Tebaldi et al., 2021). In the following sections, we first summarize current knowledge on the impacts of global warming on temperature and precipitation at the European scale, and then focus on the current understanding of climate change impacts in Belgium.

1.2.1 Europe

Europe is warming at a faster rate than the global average (Copernicus Climate Change Service, 2025), for three main reasons. First, changes in the frequency of large-scale weather patterns have increased the occurrence and intensity of summer heatwaves (Copernicus Climate Change Service, 2025; Beobide-Arsuaga et al., 2025). Second, a reduction in air pollution has led to lower aerosol concentrations, allowing a larger

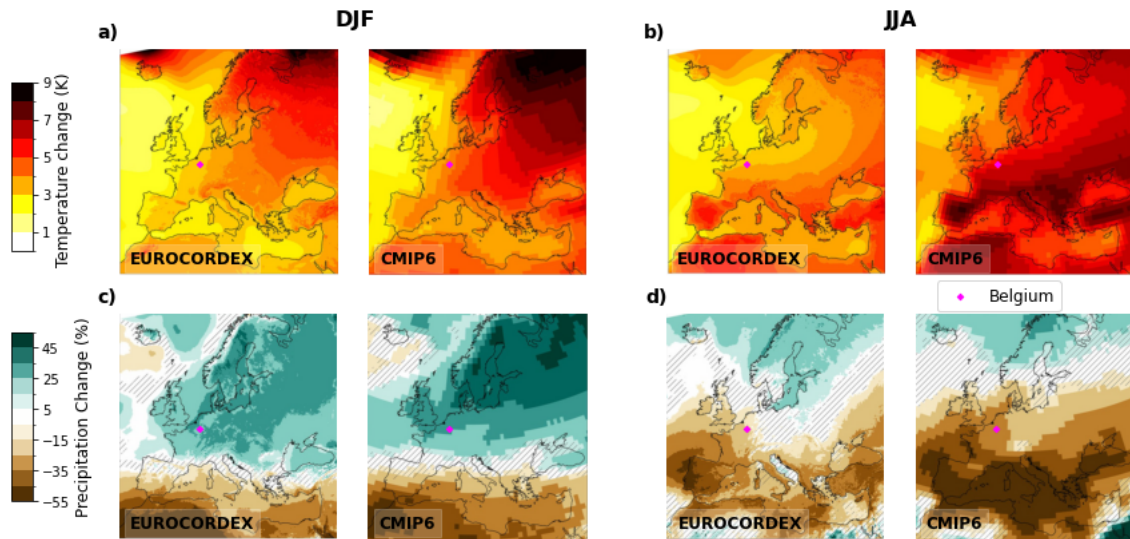


Figure 1.3: Projected changes in (a, b) mean temperature (K) and (c, d) total precipitation (%) for 2071–2100 relative to 1981–2010, shown for DJF (a, c) and JJA (b, d). Results are based on the Euro-CORDEX dataset under RCP8.5 (left panels) and the CMIP6 dataset under SSP5-8.5 (right panels) (modified after Coppola et al. (2021)). Belgium’s location is highlighted in pink.

fraction of solar radiation to reach the surface (Copernicus Climate Change Service, 2025). Third, parts of Europe extend into the Arctic, which is the fastest-warming region on Earth (Copernicus Climate Change Service, 2025). This amplified warming has important implications for precipitation regimes and extreme weather events, with pronounced regional differences that will be examined in the following sections.

1.2.1.1 Temperature and precipitation changes

The state-of-the-art datasets to study climatic impacts of global warming at the European scale are the Euro-CORDEX and CMIP6 datasets. Euro-CORDEX provides outputs at a finer spatial resolution, as they are generated by Regional Climate Models (RCMs), which are forced at their boundaries by CMIP5-generation models, i.e., the previous generation of global climate models used in the IPCC assessments (Fig. 1.3). These models therefore still use RCP scenarios (Representative Concentration Pathways) (Coppola et al., 2021). In contrast, CMIP6 datasets are available at a coarser resolution and rely on SSP scenarios. These two generations also differ in their climate sensitivity, with CMIP6 models generally exhibiting higher sensitivity to CO_2 concentration increases (Tebaldi et al., 2021). To highlight these differences, we describe the regional patterns of global warming impacts using the highest-emission scenarios, RCP8.5 and SSP5-8.5, respectively, for the Euro-CORDEX and CMIP6 outputs over 2071-2100.

Concerning temperature changes, in winter, Northern Europe is warming faster than Southern Europe (Fig. 1.3a), essentially because of the albedo feedback caused by reduced snow extent in polar environments. During summer, the Mediterranean is the fastest-warming region, which is even more pronounced in CMIP6 outputs (Fig. 1.3b). Hence, mean temperature anomalies reach 8 K in Spain, while they only reach 5 K on average within the Euro-CORDEX ensemble under the same RCP8.5 scenario. This spatial contrast can be explained by more frequent anticyclonic patterns over Southern Europe (Cresswell-Clay et al., 2022; Faranda et al., 2023), which can be linked to the expansion of

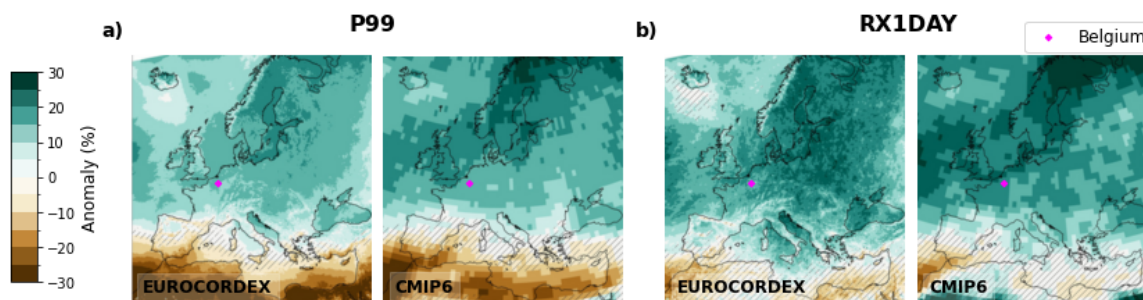


Figure 1.4: Projected changes (%) in (a) 99th precipitation percentile and (b) annual daily precipitation maximum (RX1DAY) for 2071–2100 relative to 1981–2010. Results are based on the Euro-CORDEX dataset under RCP8.5 (left panels) and the CMIP6 dataset under SSP5-8.5 (right panels) (modified after Coppola et al. (2021)). Belgium's location is highlighted in pink.

the general circulation Hadley cell (Lionello et al., 2024), making the Azores anticyclone migrate northward. Moreover, the Mediterranean's buffering effect on temperatures is less efficient than that of open oceans, since it is a closed basin (Lionello and Scarascia, 2018). For all these reasons, the Mediterranean region is often referred to as a global warming hotspot (Giorgi, 2006; Coppola et al., 2021).

Changes in precipitation can be summarized by the well-known sentence: "South getting drier and North getting wetter" (Coppola et al., 2021). This is observed for both summer and winter seasons (Fig. 1.3c and d). However, the limit between southern drying and northern wetting is located further north in summer (Fig. 1.3d). This southern drying is linked to the evolution of temperatures in Southern Europe, and thus shares similar underlying causes. The expansion of the Azores anticyclone is clearly seen in summer, with even Western France drying at a fast rate. This southern drying further accelerates southern warming by limiting the cooling effect of vegetation, which transpires less (Rehseh and Ruehr, 2021). Northern Europe is getting wetter mainly because the atmosphere's water retention capacity is increasing due to atmospheric warming (Trenberth et al., 2003). In addition, CMIP6 models suggest, with low confidence, that circulation changes might lead to more frequent negative NAO (North Atlantic Oscillation) phases, making cyclones migrate poleward (Tamarin and Kaspi, 2017; IPCC, 2021; Fuentes-Franco et al., 2023). In contrast, Central Europe's future is much more uncertain, since models do not necessarily agree on the sign and intensity of the changes (Coppola et al., 2021) (Fig. 1.3d).

1.2.1.2 Extreme precipitation events

Extreme precipitation events are projected to become both more frequent and more intense across Europe (Rajczak and Schär, 2017; Coppola et al., 2021; Cardell et al., 2020; André et al., 2024). While the continent is also experiencing longer and more persistent heatwaves due to rising temperatures (Coppola et al., 2021; Cardell et al., 2020; Boboc et al., 2025), this section focuses specifically on projected changes in extreme precipitation.

In general, extreme precipitation intensity is expected to follow the Clausius–Clapeyron (CC) scaling relationship, corresponding to an increase of approximately 7% per degree of global warming (Trenberth et al., 2003) due to the increased atmospheric water retention capacity. However, recent studies suggest that intense, short-duration events, such as mesocyclonic rainfall, may exhibit super-CC scaling, leading to even larger relative

increases (Da Silva and Haerter, 2025). Another important factor is topography. Regions with high topographic gradients tend to concentrate precipitation production, which explains why, for instance, precipitation associated with atmospheric rivers is expected to intensify particularly in these areas due to the topographic barrier effect (Payne et al., 2020).

Scandinavian countries are projected to experience the strongest relative increases in RX1DAY, i.e., the maximal annual daily precipitation, with changes reaching up to 30% by 2071–2100 in Northern Norway (Fig. 1.4b). This can potentially be attributed to the poleward displacement of storm tracks (Tamarin and Kaspi, 2017) and the shortening of the snow season at high latitudes (Buitink et al., 2021), which enhances evapotranspiration and consequently increases atmospheric water vapor content. As a result, the number of wet days at high latitudes is also expected to increase (André et al., 2024; Lhotka et al., 2024). These mechanisms highlight the pronounced spatial variability of extreme precipitation responses across Europe, which is influenced by changes in the frequency of weather regimes (Wood, 2023; Faranda et al., 2023; Tamarin and Kaspi, 2017), local geographical factors (Payne et al., 2020), and the non-linear scaling of precipitation with temperature increase (Lenderink, 2017; Da Silva and Haerter, 2025).

Among the other expected changes, modifications of the large-scale atmospheric circulation may also influence the characteristics of extreme precipitation. In particular, cut-off lows (COLs), i.e., cold-core upper-level low-pressure systems detached from the mid-latitude westerlies, are frequently associated with heavy precipitation over Europe (Palmén, 1949; Nieto et al., 2005). These systems were involved in several recent high-impact flood events, including the July 2021 floods in Belgium (Fig. 1.1a), the October 2024 Valencia floods, and the September 2024 floods in Central Europe. Consequently, understanding how their occurrence and impacts may evolve under climate change has become an active area of research. Some recent studies have suggested that Arctic amplification, i.e. the faster warming of the Arctic relative to lower latitudes, may modify the characteristics of the mid-latitude jet stream by reducing the meridional temperature gradient (Preece et al., 2023), thereby influencing the occurrence and persistence of cut-off lows (Muñoz et al., 2020; Mishra et al., 2025; Thompson et al., 2024a). However, other studies suggest that the future intensification of extreme precipitation over Europe will mainly be driven by thermodynamic processes, through changes in atmospheric moisture availability and convective available potential energy, rather than by changes in large-scale atmospheric dynamics (Pfahl et al., 2017; Crawford et al., 2025).

These expected changes in extreme precipitation intensity are already being confirmed across various locations in Europe, where observational records reveal significant trends (Da Silva and Haerter, 2025; Spanjers et al., 2025). However, due to the relatively recent start of systematic data collection, future model-projected trends are not always clearly detected in past observations (Da Silva and Haerter, 2025). This partly results from the low signal-to-noise ratio in precipitation time series (Serinaldi and Kilsby, 2015; Slater et al., 2021; Aalbers et al., 2017), whereas significant trends are more easily identified in temperature records (Spanjers et al., 2025). Nevertheless, ongoing efforts within the climate modelling community are providing increasingly detailed insights into what Europe's future climate may look like.

Extreme scenarios are particularly informative because they tend to exacerbate regional differences. However, beyond the previously mentioned climate change hotspots, namely the Mediterranean and Scandinavian regions, the impacts of climate change may remain

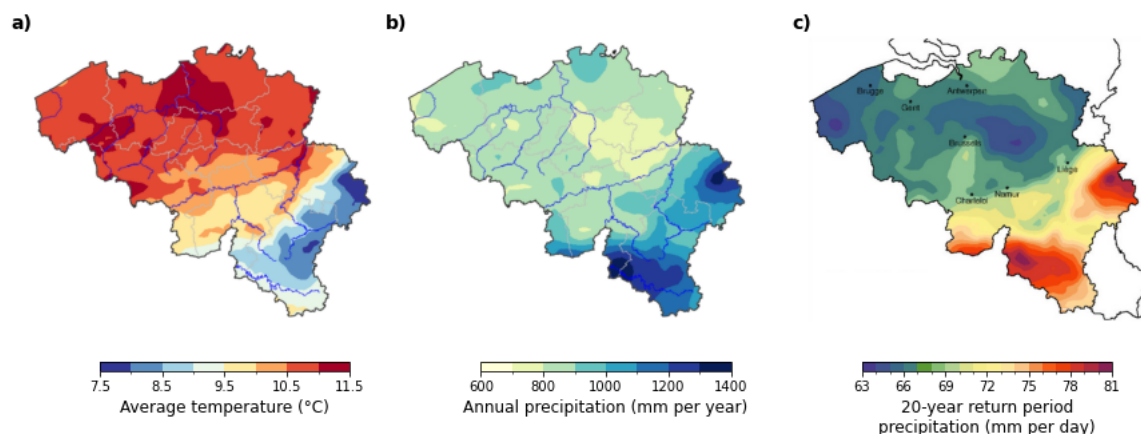


Figure 1.5: (a) Average temperature (°C) and (b) annual precipitation (mm per year) in Belgium for the reference period 1991–2020 (modified after RMI (2025)). (c) 20-year return level of daily precipitation (mm per day) (modified after de Vyver (2013)).

relatively modest in Central Western Europe under a +2 K warming scenario (Donnelly et al., 2017).

1.2.2 Belgium

1.2.2.1 Belgium's climate

Belgium's topography follows a general northwestern–southeastern gradient, with the highest elevations located in the Ardennian Massif near the German border. Three main topographic regions are commonly distinguished: Low Belgium (0–100 m), Medium Belgium (100–300 m), and High Belgium (above 300 m) (see Figure 2.1 for the spatial distribution of these topographic regions). This altitudinal pattern largely explains the spatial distribution of mean temperature across the country. Near the coast, annual mean temperatures reach approximately 11 °C, while they drop to around 7 °C in the highest areas of the Ardennes (Fig. 1.5a). In July, the warmest month of the year, average temperatures range from 15.5 to 19.5 °C, whereas in January, the coldest month, they vary between -0.5 °C and 4.5 °C (RMI, 2025).

Annual precipitation totals reflect a similar topographic gradient. In Low and Medium Belgium, mean annual precipitation ranges between 600 mm and 800 mm, while amounts reach up to 1300 mm in the highest parts of the country (Fig. 1.5b). This pattern is primarily governed by the prevailing westerly circulation, which brings moist air masses from the Atlantic Ocean. When these air masses encounter the Ardennian relief, orographic lifting enhances rainfall, leading to higher totals in southeastern Belgium (Erpicum et al., 2017). Despite this strong maritime influence, annual precipitation amounts remain relatively moderate compared to other oceanic climates.

Extreme precipitation indicators follow a similar northwestern–southeastern gradient. The 20-year return level of daily precipitation reaches about 80 mm per day in the highest locations, compared to roughly 60–65 mm per day in flatter regions (de Vyver, 2013) (Fig. 1.5c).

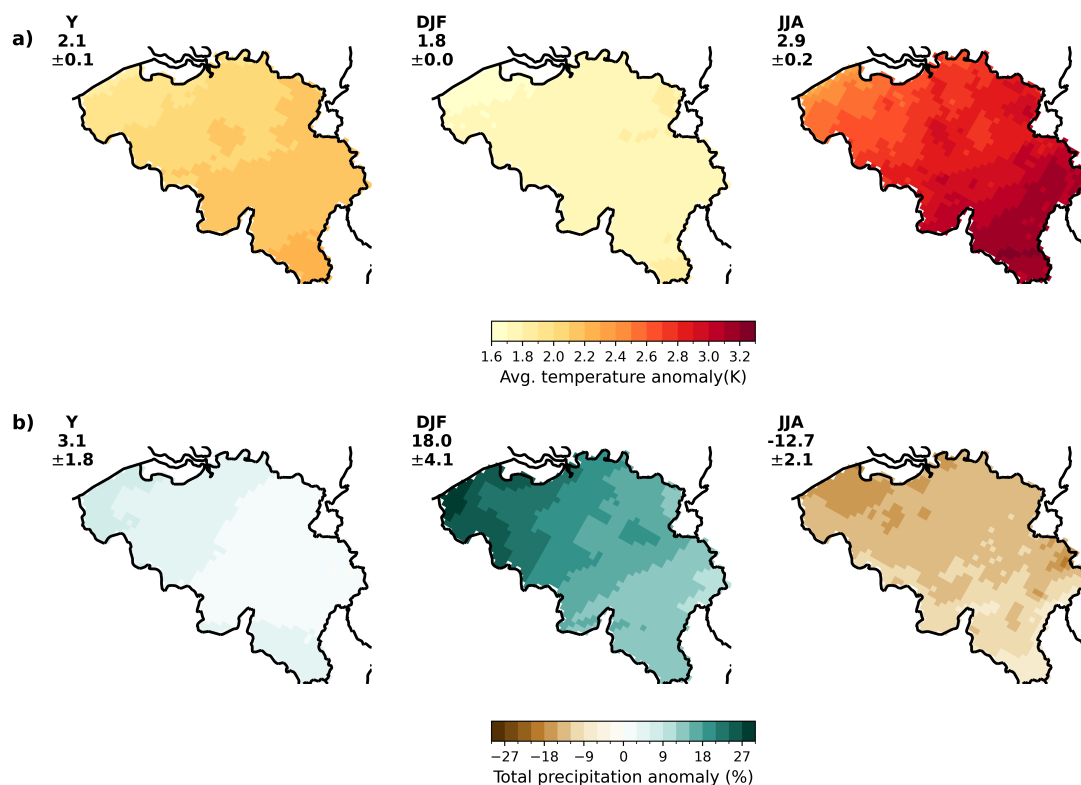


Figure 1.6: (a) Mean temperature and (b) total precipitation anomalies over Belgium, showing the difference between the +4 K and +2 K global warming levels, derived from CMIP6–MAR simulations. Results are shown for the annual mean (Y), winter (DJF), and summer (JJA). The average value $\pm$ the standard deviation are shown on the top left of each figure (adapted from Agence Wallonne de l’Air et du Climat (2024)).

1.2.2.2 Global warming in Belgium

Belgium is located at the interface between northern and southern Europe in terms of global warming impacts. In terms of temperatures, it is expected to warm more slowly than Scandinavian countries in winter (Fig. 1.3a), with an average temperature increase of +3 °C under the CMIP6 SSP5-8.5 scenario by 2071-2100 compared to preindustrial conditions. Summer mean temperatures may increase by up to 5 °C under the same scenario (Fig. 1.3b).

For precipitation, the two seasons show opposite changes, with an increase of up to 30% in winter and a decrease of up to 30% in summer (Fig. 1.3c and d).

These future temperature trends are supported by historical observations. At the Uccle station, the annual average temperature for 1991–2020 is already 1.2 °C above the 1961–1990 average. At the same station, the July mean temperature has increased by 1.4 °C over the same periods. In contrast, clear trends have not yet appeared for precipitation at Uccle, at least for annual values (RMI, 2025), with the signal being dominated by strong interannual variability.

Despite this, for some specific months, slight changes are being observed. For example, January totals were 67 mm in 1961–1990 and 76 mm in 1991–2020, corresponding to a 13% increase (RMI, 2025).

Recently, the CORDEX-BE.2 project was launched to combine regional downscaling

expertise from multiple Belgian institutes (RMI, 2025). This project aims to assess the impacts of global change at high resolution over Belgium using state-of-the-art RCMs. Among these, MAR (Modèle Atmosphérique Régional) provides robust projections for Belgium's future climate and is also used in this thesis to evaluate the impacts of global warming on extreme precipitation and resulting flooding (see Methodological Section 2.1 for details).

MAR outputs have been incorporated into technical reports and GIS-based web platforms (Agence Wallonne de l'Air et du Climat, 2024; Fettweis et al., 2024). These assessments are based on global warming levels, which select periods corresponding to a given warming threshold to evaluate changes in Belgium (Fig. 1.6).

According to MAR, average annual temperatures in Belgium are expected to track the global trajectory (Fig. 1.6a). For instance, a +2 K global warming is projected to correspond to a +2 K increase in Belgium. However, summer warming is likely to be stronger, with a corresponding rise of approximately 3 K, further confirming the trends observed in Euro-CORDEX projections (Fig. 1.3). Spatially, both annual and summer temperatures exhibit a northwestern–southeastern gradient, with coastal areas less affected by global warming than southern regions, due to the sea-buffering effect.

These projected future changes differ from the warming observed over the historical period, during which Belgium has warmed faster than the global average. For example, the annual mean temperature at the Uccle reference station has increased by approximately 1.9 K since the beginning of the instrumental record, compared with an estimated global mean warming of about 1.1 K (Climat.be, n.d.).

For precipitation, a +2 K global temperature increase leads to only a slight 3.1% change in annual rainfall over Belgium (Fig. 1.6b). The main impact of global warming is on the seasonal cycle of precipitation: winter precipitation is projected to increase on average by 18%, while summer precipitation is expected to decrease by approximately 13% (Fig. 1.6b). Interannual variability is also expected to rise, with very dry winters occasionally followed by very wet ones (Fettweis et al., 2024).

Concerning impacts on precipitation extremes, scientific literature remains relatively scarce. Hosseinzadehtalaei et al. (2018) analysed the outputs of the Euro-CORDEX community members to explore the response of Intensity Duration Frequency curves (IDFs) to Global Warming at specific locations. Their results suggest that the 10-year return level of daily rainfall might increase by 50 to 70% under a RCP8.5 scenario for Central Belgium.

Kreienkamp et al. (2021) investigated the attribution of the July 2021 event by analysing trends in the shape of extreme precipitation distributions. Their results suggest that the occurrence of the 2021 large scale precipitation event has become 1.2 to 9 times more likely due to climate change. Although the return period is estimated at 400 years based on observations, its frequency is expected to further increase with additional global warming. The findings of Mohr et al. (2023) and Ludwig et al. (2023) show that the July 2021 rainfall event could have even been 7 to 9 % more intense per degree of added global warming. Using a storyline approach, they concluded that this event was 11% more intense than it would have been under pre-industrial conditions.

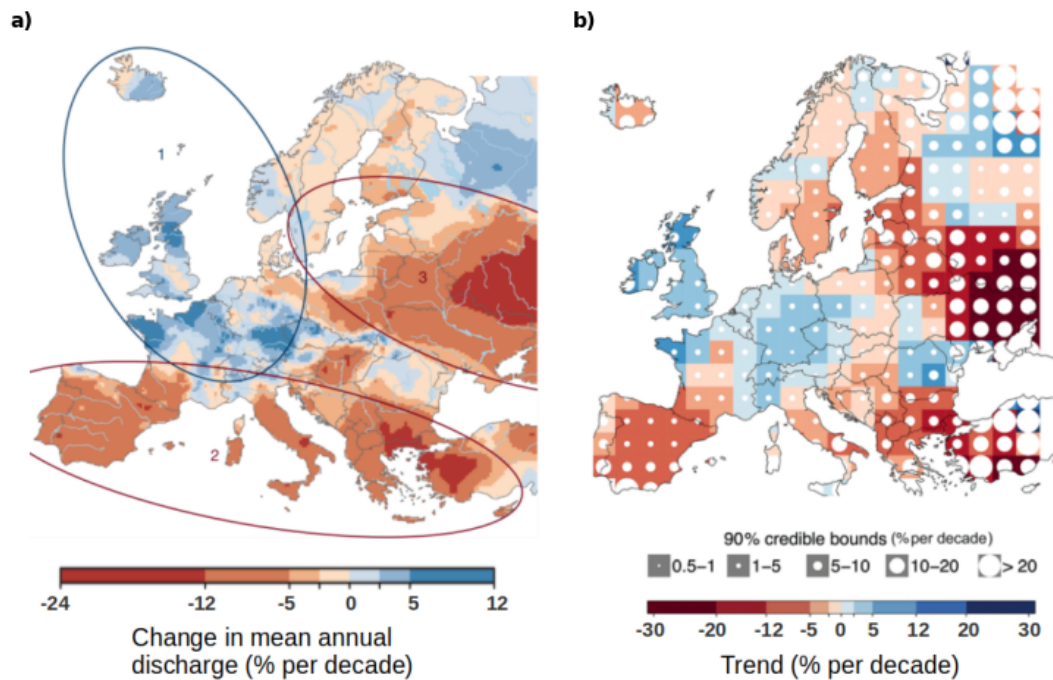


Figure 1.7: Observed trends in (a) annual mean river discharge (1960–2010) (modified after Blöschl et al. (2019)) and (b) 100-year return levels of river discharge for small catchments (100 km²) over the same period (modified after Bertola et al. (2019)).

1.3 Hydrological impacts

1.3.1 Hydrological impacts in Europe

In recent decades, significant trends in river discharge have been observed across Europe. Blöschl et al. (2019) analysed discharge records from 3,738 gauging stations between 1960 and 2010, covering catchments ranging from 5 to 100,000 km² (Fig. 1.7a). Their analysis revealed three distinct regions across Europe, each characterized by specific driving mechanisms.

In northwestern Europe, about 69% of stations exhibit increasing discharge trends (Fig. 1.7a). In this region, floods mainly originate from winter rainfall occurring when soils are already saturated. These observed increases are consistent with the intensification of extreme precipitation and the poleward displacement of North Atlantic storm tracks (Tamarin and Kaspi, 2017; André et al., 2024).

Southern Europe forms the second zone, where the northward expansion of the Hadley cell results in reduced precipitation and enhanced evapotranspiration, leading to drier soils and decreasing discharge trends. However, available data for this region mostly correspond to large catchments, where short-duration convective rainfall events have a comparatively smaller hydrological impact than in smaller basins.

Finally, in Eastern Europe, predominantly negative trends are observed, primarily linked to a shortened snow season and the reduced contribution of snowmelt to river flows. Overall, these spatially heterogeneous trends highlight the complex interplay between physical processes that govern Europe's hydrological response to climate change.

Bertola et al. (2019) used the same dataset as Blöschl et al. (2019), pooling the data within spatial windows of $600 \text{ km} \times 600 \text{ km}$ (Fig. 1.7b). The goal of this study was to compare trends between rare and frequent flood events, as well as between large and small catchments. Overall, their findings suggest that trends are more pronounced in smaller catchments than in larger ones. For these smaller catchments (up to 100 km^2), trends are more visible for rarer events than for median floods, while the opposite is observed for larger catchments. These observations might result from convective events becoming more intense and frequent, which more easily impact smaller catchments.

Simulated trends from CMIP-driven hydrological models generally support these observations, indicating that projected changes in river discharge broadly align with the historical patterns documented across Europe (Donnelly et al., 2017; Roudier et al., 2015; Schneider et al., 2013; Thober et al., 2018; Lobanova et al., 2018).

1.3.2 Hydrological change impacts in Belgium

Belgium represents a challenging case study due to its small size and its location within a climatic transition zone, situated between regions of decreasing discharge in Eastern Europe and increasing discharge in Northern–Western Europe (see Fig. 1.7). Despite its limited extent, the country exhibits a wide range of geological and geographical contexts, leading to diverse river regimes (Van Campenhout et al., 2020).

Studies addressing the hydrological impacts of climate change in Belgium date back to the 1990s. Gellens and Roulin (1998) used outputs from IPCC GCMs to force hydrological models over several Belgian river basins characterized by contrasting geological settings. Their results showed that clear trends in seasonal streamflow were difficult to identify and strongly depended on the choice of the driving GCM. Similarly, Leander et al. (2008) applied two regional climate models (RCMs) driven by three GCMs in combination with the hydrological model HBV (Hydrologiska Byråns Vattenavdelning) (Bergström and Forsman, 1973) to analyse extreme winter streamflows of the Meuse River, reporting an increase in winter precipitation but no clear signal in extreme discharge quantiles.

Subsequent studies focused on individual catchments using larger climate model ensembles. Driessen et al. (2010b) applied HBV simulations over the Ourthe River, showing overall increases in river discharge but decreases in summer flows toward the end of the 21st century due to the diminishing buffering effect of snowmelt. Baguis et al. (2010) extended this approach using 11 RCMs for the Ourthe and Gete, finding an overall decrease in mean streamflow and a pronounced reduction in summer water availability, while winter discharges could increase. Low-flow periods were amplified in the Ourthe, whereas no robust signal was identified for high-flow extremes.

Other sub-catchments of the Meuse also exhibit significant projected changes. Bauwens et al. (2011) highlighted that 2-year and 50-year return period 7-day average discharges could decrease by up to 20% and 35% in the Lesse and Vesdre, respectively, indicating potential summer water scarcity. Tavakoli et al. (2014) analysed the Grote Nete in Flanders, showing that climate change combined with urban growth increases peak flows while severely reducing low flows.

Hydraulic modelling has further quantified flood risks. Dewals et al. (2015) showed that a 30% increase in the 100-year Meuse discharge could raise water levels by up to 1.8 m between Liège and the Belgian–Dutch border. Wyard et al. (2017), using the MAR

1. Introduction

Table 1.1: Summary of hydrological change studies in Belgian catchments. These studies are summarized for quick reference of catchment, model, period, and main hydrological changes observed or projected.

Catchment Area	/	Model / Method	Period / Scenario	Main Observed / Projected Changes
Various Belgian basins		Hydrological models forced by IPCC GCMs	1990s, baseline	Seasonal streamflow trends unclear, strongly dependent on GCM choice (Gellens and Roulin, 1998)
Meuse River		HBV + 2 RCMs / 3 GCMs	Future climate scenarios	Extreme winter flows uncertain, winter precipitation increased (Leander et al., 2008)
Ourthe River		HBV	3 climate scenarios	Overall discharge increase, summer flow decrease (Driessen et al., 2010b)
Ourthe & Gete		SCHEME + 11 RCMs	Ensemble future scenarios	Mean flow decrease, summer low flows reduced, winter flows may increase, low-flow amplification; no robust high-flow signal (Baguis et al., 2010)
Lesse & Vesdre		Hydrological modelling	Future scenarios	2-year discharge down 20%, 50-year down 35%; potential summer water scarcity (Bauwens et al., 2011)
Grote Nete		WetSpa + 3 climate scenarios	Future climate + urban growth	Winter precipitation increase, summer decrease; high urbanization + warming → higher peak flows, severe low-flow reduction (Tavakoli et al., 2014)
Meuse River (Liège stretch)		2D hydraulic model	100-year flood scenario	30% increase in 100-year discharge → water level rise up to 1.8 m, higher flood frequency (Dewals et al., 2015)
Ourthe River		MAR RCM	Future winter scenarios	Winter flows reduced due to shorter snow season (Wynd et al., 2017)
Ourthe River		Hydraulic modelling	Long-term scenarios	Short overflow events decrease, long-duration events increase (Grelrier et al., 2019)
Wallonia (84 catchments)		Observations	1970–2017	High-flow trends highly variable; some sites increase, others decrease (Grandry et al., 2020)
Doode Bemde wetland		Hydrological analysis	Observed droughts	Prolonged dry periods → significant groundwater and wetland storage reduction (Wossenyeleh et al., 2021)
Lesse catchment		HBV & GR6J	Climate projections	High and low flows sensitive to calibration; extremes not robust (Ten Berge et al., 2025)
Ahr catchment (Germany, near Belgium)		PGW experiments + RCM ensemble	±2–3 K scenarios	+2 K → 18% rainfall increase → 39% flood peak increase; CC scaling 7–9%/K; super-CC up to 10%; extreme precipitation extent increases (Ludwig et al., 2023)

RCM, indicated that evolving winter hydroclimatic conditions would reduce winter flows in the Ourthe due to a shortened snow season. Grelrier et al. (2019) explored overflow scenarios for the Ourthe, showing that short-duration events are projected to decrease while long-duration overflows become more frequent, valuable information for flood risk management.

Observational and attribution studies provide further context. De Niel and Willems (2019) decomposed peak-flow trends in 29 Flanders catchments, showing that changes were mostly driven by catchment characteristics, though climate and land cover contributed up to 32% of observed increases. Grandry et al. (2020) analysed 84 catchments across Wallonia, finding strong spatial variability in high-flow trends, with some sites increasing and others decreasing. Wossenyeleh et al. (2021) examined drought propagation in the

Doode Bemde wetland, demonstrating significant reductions in groundwater levels and wetland storage during prolonged dry periods.

Model robustness and extreme-event analysis further inform projections. Ten Berge et al. (2025) evaluated HBV and GR6J simulations in the Lesse catchment, showing that projected changes in high and low flows are highly sensitive to calibration strategies and baseline selection, highlighting the importance of careful model evaluation. Ludwig et al. (2023) studied the July 2021 floods in the Ahr catchment, just across the border from Belgium. Pseudo-global warming experiments and a high-resolution RCM ensemble revealed that spatial mean precipitation scales with the Clausius–Clapeyron relation (7–9% per degree warming), while hydrological modelling showed non-linear amplification of flood peaks: under a +2 K scenario, an 18% increase in rainfall led to a 39% increase in peak discharge at Altenahr. Convection-permitting RCMs confirmed CC scaling for moderate precipitation but indicated super-CC scaling up to 10% for higher intensities, with an increased spatial extent of extreme precipitation events, highlighting potential implications for Belgian catchments near the border.

These studies are summarized in Table 1.1, providing an overview of the catchments, models, periods, and main hydrological changes observed or projected.

1.4 Research objectives

1.4.1 Addressed research gaps

Despite considerable progress in understanding the impacts of climate change on extreme precipitation, several important research gaps remain regarding hydroclimatic extremes in Belgium.

First, although numerous studies project an intensification of extreme precipitation following the Clausius-Clapeyron relationship, relatively few have investigated how these changes depend simultaneously on event rarity, accumulation duration, and spatial scale at high resolution over Belgium. The availability of the CORDEX-Be.2 simulations, performed at 5-km resolution, provides the opportunity to address these questions using state-of-the-art regional climate projections.

Second, the synoptic mechanisms responsible for extreme precipitation remain insufficiently quantified. In particular, cut-off lows (COLs) have been associated with several recent high-impact flood events, including the July 2021 floods in Belgium, the October 2024 Valencia floods, and the September 2024 floods in Central Europe. Although some studies suggest that the frequency, persistence, or impacts of COLs may evolve under climate change (Muñoz et al., 2020; Thompson et al., 2024a; Mishra et al., 2025), their contribution to extreme precipitation has rarely been quantified objectively over long time periods and remains largely unexplored over Belgium. This thesis therefore investigates the climatology of COLs and their contribution to extreme precipitation at both the European and Belgian scales.

Finally, considerable uncertainty remains regarding the propagation of projected climatic changes through the hydrological system. Previous studies conducted in Belgium have reported contrasting projections of future flood hazards (Section 1.3.2), partly because of differences in climate forcings, bias-adjustment methods, and hydrological

model structures. This thesis therefore evaluates a complete regional climate-hydrological modelling chain, investigates the sensitivity of hydrological projections to the choice of rainfall-runoff model, and assesses how projected changes in extreme precipitation translate into future flood hazards in Belgium.

1.4.2 Research questions and thesis structure

To address these research gaps, this thesis combines regional climate modelling, synoptic meteorology, and hydrological modelling. It is organised around three complementary research axes: (i) the evolution of extreme precipitation over Belgium under climate change, (ii) the role of cut-off lows in generating precipitation extremes, and (iii) the hydrological consequences of projected climatic changes for Belgian river basins.

First, we investigate how the frequency and intensity of extreme precipitation events are projected to evolve based on simulations from the *Modèle Atmosphérique Régional* (MAR). Particular attention is given to changes in statistical distributions and high-percentile rainfall, in order to quantify the potential intensification of extremes under future climate conditions. Chapter 3 presents the results related to this question. It is based on a published paper in *Journal of Hydrology: Regional Studies*, which analyzes the evolution of daily precipitation extremes, complemented by an extension to other temporal accumulations.

Second, the research examines the role of cut-off lows (COLs) in modulating extreme precipitation events at both European and Belgian scales. The objective is to better understand their occurrence, characteristics, and relationship with extreme precipitation under a changing climate. Chapter 4 first summarizes a submitted manuscript to *Weather and Climate Dynamics* (Brajkovic et al., 2026), which investigates the past and future evolution of COL-related precipitation at the European scale. It then extends this analysis to Belgium.

Third, this work explores how MAR climate simulations can be used as input for hydrological modelling. In particular, it assesses the integration of bias-corrected climate data into the *WOLFHydro* modelling framework. This includes addressing methodological challenges such as bias correction (e.g., quantile mapping), temporal consistency, and the suitability of climate model outputs for hydrological applications.

Finally, the study evaluates the hydrological response of a catchment to projected changes in precipitation extremes. This involves analyzing how variations in rainfall intensity and frequency affect river flood peaks. The objective is to understand how climate-driven changes propagate through the hydrological system and influence flood risk at the basin scale.

Chapter 5 presents these hydrological analyses. It focuses on the Ourthe catchment, which serves as a test case for the MAR-WOLFHydro modelling chain due to its extensive documentation in the literature. Simulated discharge time series at the Tabreux station are analysed to assess the hydroclimatic response at the catchment scale.

Chapter 6 discusses the main findings, limitations, and perspectives of the thesis.

CHAPTER 2

Tools and methods

To address the main research questions of this thesis, several modelling and statistical tools are employed.

To investigate the evolution of the Belgian climate under a warming scenario, the regional climate model MAR (Modèle Atmosphérique Régional), developed at the University of Liège, is used. MAR is one of the three Belgian regional climate models integrated within the CORDEX-Be.2 project, which aims to provide high-resolution climate projections over Belgium to inform stakeholders about the impacts of global warming (RMI, 2025). We first describe the main characteristics of this model and evaluate its performance with respect to the two key variables considered in this thesis, namely near-surface temperature and precipitation.

In a second step, we describe the hydrological modelling framework WOLFHydro, a Python-based tool also developed at the University of Liège, which is used to simulate river discharge in Belgian catchments.

Third, we detail the preprocessing steps applied to MAR outputs in order to generate suitable inputs for the hydrological models implemented within the WOLFHydro package.

Finally, we provide an overview of the statistical methods used throughout this thesis to analyse the different model outputs.

Within this thesis, several observational datasets are used to evaluate the climate and hydrological models and to preprocess MAR outputs. For climate variables, gridded observational data provided by the Royal Meteorological Institute (RMI) are used over Belgium at a daily time step for the period 1950-2021 (Journée et al., 2023). RMI station observations are additionally used for the evaluation and correction of hourly precipitation. For hydrological variables, observed discharge time series are obtained from the Walloon Public Service (SPW) (Service public de Wallonie (SPW), 2026).

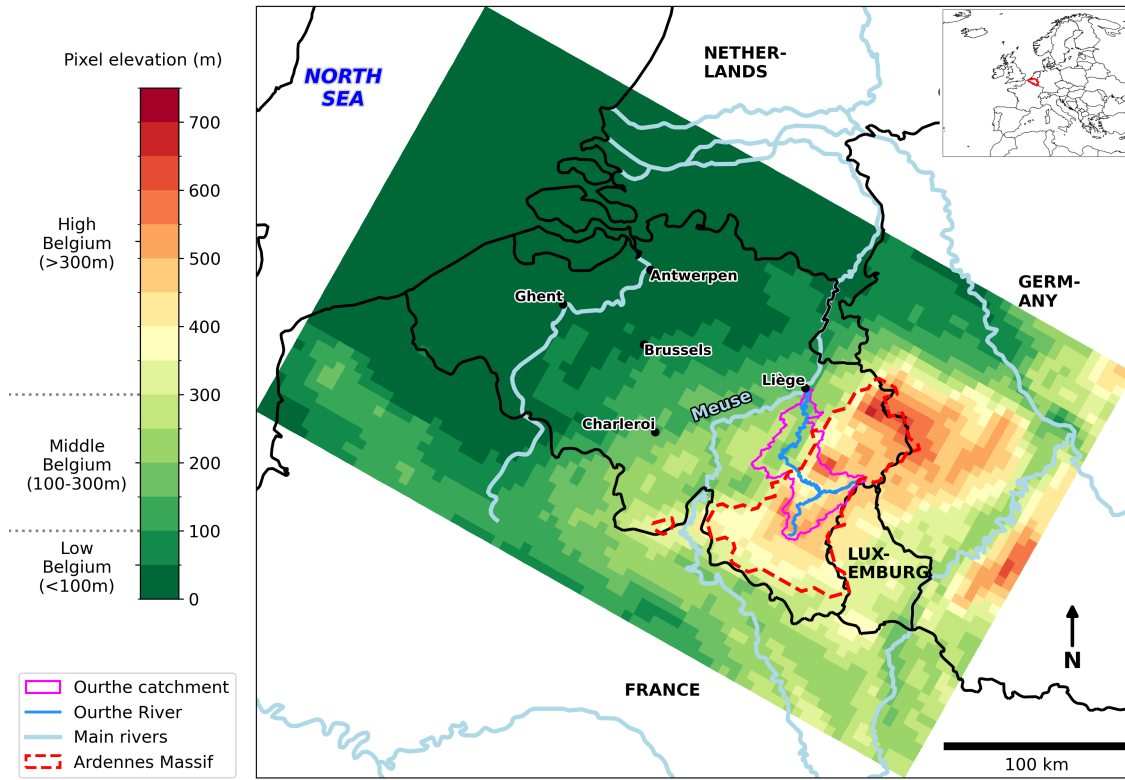


Figure 2.1: MAR inner simulation domain.

2.1 Regional climate model MAR

MAR is a hydrostatic regional climate model (RCM) that solves the primitive equations (Gallée and Schayes, 1994). The model has been thoroughly evaluated over Belgium for both temperature (Doutreloup et al., 2022b,a) and precipitation (Doutreloup et al., 2019). In the past, particular attention has been given to precipitation events associated with flooding in the Ourthe catchment (Wyart et al., 2021), as well as to specific meteorological phenomena such as freezing rain (Fettweis et al., 2017).

Deep convection in MAR is parameterized following Bechtold et al. (2001) (Doutreloup et al., 2019). The atmospheric component of the model is described in detail in Gallée and Schayes (1994) and Gallée (1995). MAR is coupled with the Soil Ice Snow Vegetation Atmosphere Transfer (SISVAT) scheme, which represents land–atmosphere interactions, including soil hydrology, snow processes, and surface runoff (Wyart et al., 2017).

In this study, we use the latest version of MAR (v3.14) (Grailet et al., 2025). Over Belgium, MAR is run at a 5-km horizontal resolution (Fig. 2.1). The MAR domain is rotated for computational efficiency purposes and to cover Belgium, parts of Western Germany, Northern France, the Netherlands and Luxembourg within the same grid.

2.1.1 Model forcings

MAR is driven at its lateral and upper boundaries by large-scale atmospheric fields provided by external datasets through a single nesting procedure. In contrast to multiple-nesting approaches, where an intermediate MAR simulation over a larger domain provides

CMIP6 ESM	Member	TCR (K)	Atm. horiz. res.	Associated MAR name
MIROC6	r1ip1f1	1.55	$\sim 1.4^\circ \times 1.4^\circ$	MAR-MIR
NorESM2-MM	r1ip1f1	1.33	$\sim 1.0^\circ \times 1.0^\circ$	MAR-NS2
MPI-ESM-2-HR	r1ip1f1	1.66	$\sim 0.9^\circ \times 0.9^\circ$	MAR-MPI
EC-Earth3-Veg	r1ip1f1	2.62	$\sim 0.7^\circ \times 0.7^\circ$	MAR-EC3
CMCC-CM2-SR5	r1ip1f1	2.09	$\sim 1.0^\circ \times 1.0^\circ$	MAR-CR5
IPSL-CM6A-LR	r1ip1f1	2.32	$\sim 2.5^\circ \times 1.3^\circ$	MAR-IPSL

Table 2.1: ESMs used to force MAR at its lateral boundaries (modified after Sobolowski et al. (2023)). TCR = Transient Climate Response.

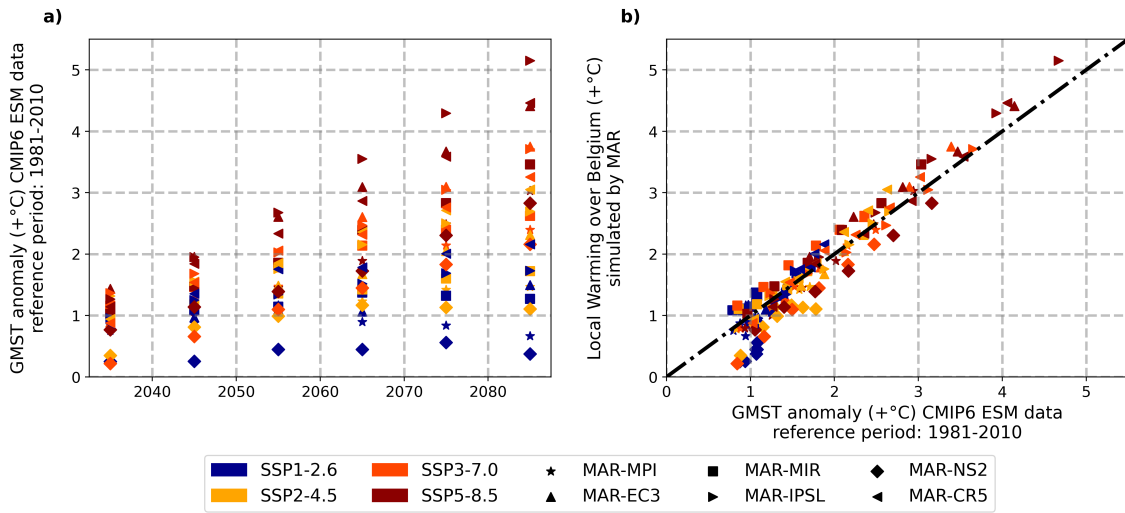


Figure 2.2: (a) 30-year centered window GMST anomaly for the different ESMs forcing MAR and (b) GMST anomaly computed by the ESMs vs. the average local warming computed by MAR over Belgium

the boundary conditions for a higher-resolution simulation, the Belgian MAR configuration used in this thesis is forced directly by the external datasets. Over the historical period (1940-2023), MAR is forced by the ERA5 reanalysis (0.25° resolution) (Hersbach et al., 2020) in order to reconstruct Belgium's climate over the past. Throughout this thesis, we refer to this MAR configuration as MAR-ERA5.

For future projections, MAR is forced by outputs from Earth System Models (ESMs) (Table 2.1). To assess the evolution of the Belgian climate, six ESMs were selected. These models were recommended by the EURO-CORDEX community based on their ability to represent the European climate (Sobolowski et al., 2023). For each ESM, one historical simulation (1981-2014) and four Shared Socioeconomic Pathways (SSPs) are available, namely SSP1-2.6, SSP2-4.5, SSP3-7.0, and SSP5-8.5. These scenarios represent increasingly high greenhouse gas emission trajectories.

As a result, global mean surface temperature (GMST) evolves differently depending on both the selected ESM and the emission scenario. Relative to the historical reference period (1981-2014), IPSL-CM6A-LR exhibits the largest GMST increase, reaching up to +5.1 K under SSP5-8.5, while NorESM2-MM shows the lowest increase, with approximately +0.3 K by 2071-2100 under SSP1-2.6 (Fig. 2.2a).

These differences arise from variations in the transient climate response (TCR) of each ESM, defined as the global mean surface temperature increase at the time of CO_2 doubling

under a $1\% \text{ yr}^{-1}$ increase in CO_2 concentration (IPCC, 2023). The six ESMs used in this study span a range of TCR values, from 1.33 K for NorESM2-MM to 2.62 K for EC-Earth3-Veg. This diversity allows exploration of a wide range of climate sensitivities.

For clarity, simplified names are used throughout this thesis when referring to MAR simulations forced by individual ESMs. These abbreviations are listed in Table 2.1.

As shown earlier (Fig. 1.6), Belgium's mean temperature closely follows global temperature evolution as simulated by MAR. Consequently, the warming simulated over Belgium by MAR is strongly correlated with GMST changes, with local temperature increases tracking global trends (Fig. 2.2b).

2.1.2 General evaluation of key climate variables

The performance of MAR v3.14 has already been evaluated by Grailet et al. (2025) for temperature, radiative fluxes, sea level pressure, and precipitation when forced by ERA5. Here, we provide a concise evaluation of the model's ability to reproduce daily near-surface temperature and precipitation over Belgium for the reference period 1981-2010.

These two variables are of primary importance for this thesis, as they directly influence both extreme value analyses and hydrological modelling. The evaluation focuses on spatial statistics computed over Belgium. The observational dataset used in this analysis consists of gridded daily precipitation data provided by the RMI (Journée et al., 2023).

The representation of sub-daily precipitation is discussed separately in Section 2.2, in the context of bias-adjustment for hydrological applications.

2.1.2.1 Temperature

For daily mean temperature, MAR exhibits a systematic seasonal bias, characterized by an underestimation in winter and an overestimation in summer.

On average over Belgium, MAR-ERA5 shows a mean bias of +0.9 K on the annual scale. This bias is strongly season-dependent, reaching +2.3 K in summer (JJA) and -0.1 K in winter (DJF) (Table 2.2).

When MAR is forced by ESMs, these biases can be further amplified. For example, a mean bias of up to +3.2 K is found in summer for the MAR-MIR configuration.

In terms of spatial variability, MAR generally reproduces the observed temperature patterns well, as indicated by generally high coefficients of determination for most seasons. However, the model performance degrades in summer, with $r^2 = 0.59$ for MAR-ERA5, compared to values close to or exceeding 0.8 in the other seasons.

2.1.2.2 Precipitation

For daily precipitation, MAR exhibits moderate biases that vary with season. On average over Belgium, MAR-ERA5 slightly underestimates annual precipitation, with a mean bias of -0.2 mm day^{-1} (Table 2.3). This underestimation is more pronounced during winter (DJF) and spring (MAM), where mean biases reach approximately -0.5 mm day^{-1} , while a slight overestimation is observed in summer (JJA) with a mean bias of $+0.2 \text{ mm day}^{-1}$.

Model	Season	Mean	std	MB	RMSE	r^2
Obs	Y	9.9	0.9	—	—	—
	DJF	4.3	1.0	—	—	—
	MAM	5.9	0.9	—	—	—
	JJA	15.4	0.8	—	—	—
	SON	14.1	0.8	—	—	—
MAR-ERA5	Y	10.8	1.1	0.9	1.1	0.83
	DJF	4.2	0.9	-0.1	0.3	0.97
	MAM	6.3	1.1	0.4	0.6	0.90
	JJA	17.7	1.6	2.3	2.7	0.59
	SON	15.2	1.1	1.1	1.3	0.78
MAR-MPI	Y	10.7	1.1	0.8	1.0	0.84
	DJF	4.1	1.0	-0.2	0.3	0.97
	MAM	5.9	1.0	-0.1	0.4	0.91
	JJA	17.6	1.6	2.2	2.6	0.63
	SON	15.2	1.1	1.1	1.3	0.77
MAR-MIR	Y	11.1	1.1	1.2	1.4	0.80
	DJF	3.7	0.8	-0.6	0.7	0.95
	MAM	6.4	1.1	0.4	0.7	0.90
	JJA	18.6	1.7	3.2	3.5	0.55
	SON	15.6	1.1	1.5	1.7	0.71
MAR-EC3	Y	10.9	1.1	1.1	1.2	0.82
	DJF	4.4	1.0	0.1	0.3	0.97
	MAM	6.3	1.1	0.4	0.6	0.92
	JJA	17.6	1.6	2.2	2.6	0.57
	SON	15.5	1.1	1.5	1.6	0.72
MAR-NS2	Y	11.0	1.1	1.1	1.3	0.85
	DJF	4.2	0.9	-0.1	0.3	0.96
	MAM	7.0	1.2	1.1	1.2	0.93
	JJA	17.5	1.6	2.1	2.5	0.63
	SON	15.5	1.1	1.4	1.6	0.75
MAR-CR5	Y	11.0	1.1	1.1	1.3	0.85
	DJF	4.0	1.0	-0.3	0.4	0.96
	MAM	6.9	1.2	0.9	1.0	0.93
	JJA	17.6	1.6	2.2	2.5	0.59
	SON	15.6	1.1	1.5	1.7	0.74
MAR-IPSL	Y	10.9	1.1	1.1	1.2	0.83
	DJF	4.0	0.9	-0.3	0.4	0.96
	MAM	6.2	1.1	0.2	0.5	0.92
	JJA	18.1	1.7	2.7	3.0	0.60
	SON	15.4	1.1	1.3	1.5	0.76

Table 2.2: Evaluation of MAR daily mean temperature over Belgium. Spatial statistics are computed by comparing MAR outputs to the IRM gridded observational dataset for all grid cells within Belgium. std denotes the standard deviation, MB the mean bias, RMSE the root mean square error, and r^2 the coefficient of determination.

Model	Season	Mean	std	MB	RMSE	r^2
Obs	Y	2.5	0.4	—	—	—
	DJF	2.7	0.6	—	—	—
	MAM	2.3	0.5	—	—	—
	JJA	2.5	0.2	—	—	—
	SON	2.6	0.3	—	—	—
MAR-ERA5	Y	2.4	0.4	-0.2	0.2	0.91
	DJF	2.2	0.6	-0.5	0.6	0.92
	MAM	1.8	0.4	-0.5	0.5	0.93
	JJA	2.7	0.4	0.2	0.3	0.90
	SON	2.6	0.3	0.1	0.2	0.81
MAR-MPI	Y	2.4	0.5	-0.1	0.3	0.90
	DJF	2.6	0.9	-0.1	0.5	0.88
	MAM	2.0	0.6	-0.2	0.3	0.94
	JJA	2.7	0.4	0.2	0.3	0.89
	SON	2.4	0.3	-0.2	0.3	0.81
MAR-MIR	Y	2.4	0.5	-0.2	0.3	0.85
	DJF	2.5	0.8	-0.3	0.5	0.85
	MAM	2.0	0.5	-0.3	0.3	0.91
	JJA	2.4	0.4	-0.1	0.3	0.83
	SON	2.4	0.3	-0.1	0.3	0.70
MAR-EC3	Y	2.3	0.4	-0.2	0.3	0.93
	DJF	2.5	0.7	-0.3	0.4	0.93
	MAM	2.0	0.5	-0.2	0.3	0.94
	JJA	2.5	0.3	0.0	0.2	0.88
	SON	2.3	0.3	-0.2	0.3	0.83
MAR-NS2	Y	2.3	0.5	-0.3	0.4	0.87
	DJF	2.4	0.9	-0.3	0.6	0.85
	MAM	1.9	0.6	-0.4	0.4	0.91
	JJA	2.7	0.4	0.2	0.3	0.88
	SON	2.1	0.4	-0.5	0.5	0.76
MAR-CR5	Y	2.3	0.5	-0.2	0.3	0.84
	DJF	2.5	0.8	-0.3	0.5	0.83
	MAM	2.1	0.6	-0.2	0.3	0.91
	JJA	2.3	0.4	-0.1	0.3	0.84
	SON	2.2	0.2	-0.3	0.4	0.62
MAR-IPSL	Y	2.5	0.5	-0.1	0.2	0.88
	DJF	2.5	0.8	-0.2	0.5	0.85
	MAM	2.0	0.5	-0.2	0.3	0.93
	JJA	2.7	0.3	0.2	0.3	0.83
	SON	2.6	0.2	0.1	0.2	0.78

Table 2.3: Evaluation of MAR daily precipitation over Belgium (units = mm per day). Spatial statistics are computed by comparing MAR outputs to the IRM gridded observational dataset for all grid cells within Belgium. std denotes the standard deviation, MB the mean bias, RMSE the root mean square error, and r^2 the coefficient of determination.

When MAR is forced by ESMs, the magnitude of these biases remains of the same order, although some configurations exhibit stronger seasonal deviations. For instance, MAR-NS2 shows a marked underestimation in autumn (SON), with a mean bias of -0.5 mm day^{-1} .

In terms of spatial variability, MAR generally reproduces the observed precipitation patterns well, with coefficients of determination typically ranging between 0.85 and 0.95 depending on the season. The model performs best in spring, with r^2 values exceeding 0.9 for most configurations, while slightly lower performances are found in autumn, particularly for MAR-CR5 ($r^2 = 0.62$).

Overall, these results indicate that MAR provides a reasonably realistic representation of daily precipitation and temperature over Belgium. However, the remaining biases can have important implications when MAR outputs are used to force hydrological models. Therefore, the simulated outputs are bias-adjusted before being used as hydrological model forcings, as further discussed in Section 2.2.

2.1.3 Hydrological modelling

As mentioned earlier, one of the objectives of Chapter 5 is to evaluate the ability of the MAR-WOLFHydro modelling chain to simulate hydrological responses under climate change conditions.

WOLFHydro enables the use of various hydrological models. In this thesis, four models available within the framework are used to assess the sensitivity of the simulated hydrological response to model structural uncertainty: GR4H (Mathevet, 2005), HBV (Bergström, 1995), VHM (Willems, 2014), and Sacramento (SAC-SMA) (Burnash et al., 1973). These are conceptual lumped rainfall-runoff models that are widely applied in hydrological studies. In a lumped modelling approach, the catchment is represented as a single spatial unit, and spatially aggregated meteorological inputs are used to reconstruct the discharge at the corresponding outlet. In this study, spatially averaged precipitation and potential evapotranspiration are therefore used to force the hydrological models.

In the following subsections, we briefly describe the selected hydrological models. These models differ in their conceptual complexity, which is mainly related to the representation of soil moisture processes and the number of parameters controlling water storage, partitioning, and flow routing between reservoirs. The models are presented in increasing order of conceptual complexity, from GR4H to SAC-SMA. This range of model structures allows us to assess the influence of structural uncertainty on the simulated hydrological response under climate change conditions. For a more detailed description of the implementation of each model within the WOLFHydro framework, the reader is referred to the documentation of the WOLFHydro framework (Hydraulics in Environmental and Civil Engineering (HECE) Group, 2025). All models are implemented at an hourly time step. This choice differs from many previous studies, which commonly rely on daily time steps, and allows for a more detailed representation of hydrological processes and extreme events.

Finally, we introduce the study area used for model calibration and evaluation, namely the Ourthe River basin.

2.1.3.1 GR4H

GR4H (*Génie Rural à 4 Paramètres Horaire*), derived from the GR4J model (Mathevet, 2005) is adapted to operate at an hourly time step. It is widely used due to its parsimonious structure and robust performance across a wide range of hydrological conditions.

The model consists of a production reservoir and a routing reservoir, which represent soil moisture dynamics and flow routing, respectively. GR4H relies on four parameters, each with a specific hydrological interpretation (Fig. 2.3a): (i) the maximum capacity of the production reservoir (mm), representing the maximum amount of water that can be stored in the production reservoir (X1 in Fig. 2.3a), (ii) the groundwater exchange coefficient (dimensionless), which controls subsurface water exchanges, including losses and gains (X2 in Fig. 2.3a), (iii) the maximum capacity of the routing reservoir (mm), which governs flow attenuation and the contribution of delayed runoff (X3 in Fig. 2.3a), and (iv) the base time of the routing function, which controls the transfer of water toward the outlet by modulating the diffusion and delay of the hydrograph (X4 in Fig. 2.3a).

Due to its relatively low complexity, GR4H is easy to calibrate and provides robust simulations.

2.1.3.2 HBV

The HBV model (Hydrologiska Byråns Vattenbalansavdelning) (Bergström and Forsman, 1973) has been widely applied in hydrological studies, notably in Belgium for the Ourthe River basin (Driessen et al., 2010b,a).

The model represents the catchment using three main components: a soil moisture reservoir, an upper response reservoir, and a lower response reservoir. The soil moisture reservoir represents water storage in the catchment and evapotranspiration processes and is controlled by several parameters governing soil water capacity and runoff generation.

The upper and lower reservoirs form the response routine of the model. The upper reservoir primarily represents faster runoff components, while the lower reservoir represents slower baseflow contributions. Their behaviour is governed by recession coefficients, a threshold parameter controlling the activation of fast runoff, and a percolation parameter regulating the transfer of water between the two reservoirs (Seibert, 2000).

In its original formulation, HBV includes a snow routine based on a degree-day approach to account for snow accumulation and melt processes. However, in the WOLFHydro implementation used in this study, the snow routine is not activated.

2.1.3.3 VHM

VHM (Veralgemeend Conceptueel Hydrologisch Model) is a Dutch abbreviation for a generalized lumped conceptual and parsimonious model structure identification and calibration approach (Willems, 2014).

VHM is composed of a soil moisture reservoir (Maroy et al., 2021), which controls the amount of water stored in the soil, an overland flow routing controlling the contribution of direct runoff, an interflow routing controlling the contribution of subsurface runoff, and a slow flow routing controlling the contribution of slow-response components (Fig. 2.3b).

The filling of the soil moisture reservoir is driven by incoming precipitation and outgoing potential evapotranspiration, and is controlled by two parameters representing the maximum evapotranspiration rate and the maximum storage capacity. Three additional parameters control the relationship between soil moisture and runoff fractions.

The overland flow is controlled by two parameters modulating the fraction of precipitation contributing to direct runoff. Two similar parameters control the interflow contribution as a function of soil moisture.

Each of the three flow components (overland flow, interflow, and baseflow) is associated with a time constant controlling the response time of each component.

2.1.3.4 Sacramento (SAC-SMA)

The Sacramento Soil Moisture Accounting model (SAC-SMA) (Burnash et al., 1973; eWater, 2018) is composed of two soil layers: an upper layer controlling the fast flow regime and a lower layer controlling evapotranspiration and the contribution of soil moisture to baseflow. Water can be stored and exchanged between these two soil zones (Fig. 2.3c).

Each layer is divided into tension water and free water reservoirs. Tension water represents water held in the soil that is not directly available for runoff, while free water contributes to runoff generation. In the lower layer, the free water reservoir is further divided into a primary and a supplemental component. The supplemental component drains over time scales of weeks to months following precipitation events, whereas the primary component drains more slowly, over months to years.

The SAC-SMA model represents several runoff components, including direct runoff from impervious areas, surface runoff from saturated soils, and interflow corresponding to subsurface flow. These components together form the fast flow response of the catchment.

In addition, baseflow is generated from the lower soil layer, with part of it potentially lost from the system through deep percolation (eWater, 2018).

This conceptual complexity is associated with a relatively large number of parameters controlling the partitioning between tension and free water, percolation processes, and the contribution of each runoff component. As a result, SAC-SMA is the most computationally demanding model to calibrate among the four models used in this thesis.

2.1.3.5 Required climatic inputs

All used conceptual rainfall-runoff models require precipitation and potential evapotranspiration as input variables.

In WOLFHydro, climatic input time series can be associated either with point locations (e.g., meteorological stations), which are spatially interpolated using the Thiessen polygon method, or directly with spatial polygons covering the entire modelled domain.

In this study, each MAR grid cell is treated as an individual polygon, with its geometry stored in ESRI shapefile format (Fig. 2.4). Hourly precipitation and daily minimum and maximum temperature time series are extracted for each polygon and provided as input to the WOLFHydro framework.

Potential evapotranspiration is then estimated internally within WOLFHydro using the Oudin formula (Oudin et al., 2005), based on daily minimum and maximum temperature data, which are converted to daily evapotranspiration values.

These inputs are subsequently aggregated to produce lumped precipitation and evapotranspiration time series for each modelled sub-basin.

However, as mentioned earlier, MAR outputs exhibit systematic biases. Therefore, both precipitation and temperature data are bias-adjusted prior to their use in the hydrological simulations.

2.1.3.6 Hydrological model calibration

Each of the aforementioned hydrological models relies on a set of parameters governing the model equations. These parameters are estimated over a calibration period using the corresponding observed hydrograph obtained from the Public Service of Wallonia (SPW) (Service public de Wallonie (SPW), 2026). During this period, an optimization procedure is applied to identify the parameter set that maximizes the Nash-Sutcliffe efficiency (NSE) (Nash and Sutcliffe, 1970). This efficiency criterion is defined as:

$$\text{NSE} = 1 - \frac{\sum_{t=1}^T \left(Q_{\text{obs}}^t - Q_{\text{sim}}^t \right)^2}{\sum_{t=1}^T \left(Q_{\text{obs}}^t - \bar{Q}_{\text{obs}} \right)^2} \quad (2.1)$$

where Q_{obs}^t and Q_{sim}^t are the observed and simulated discharges at time t , respectively, and $\bar{Q}_{\text{obs}}$ is the mean observed discharge over the calibration period.

The NSE ranges from $-\infty$ to 1, with values closer to 1 indicating better model performance.

Within the WOLFHydro framework, the optimization is performed using a simulated annealing algorithm (Corana et al., 1987; Dessers et al., 2025).

2.1.3.7 Study area: Ourthe River basin

The study area selected to evaluate the MAR-WOLFHydro modelling chain is the Ourthe River basin (Fig. 2.4).

The Ourthe basin is located in southeastern Belgium and is characterized by a marked south-north topographic gradient. The upstream part of the catchment lies in the Ardennes region, with elevations exceeding 500 m, and is mainly underlain by Devonian shales and sandstones. Further downstream, the basin crosses the limestone belt of the Calestienne before extending into the Famenne depression, which is mainly characterized by shales, and subsequently into the Condroz region, where alternating limestone and sandstone formations dominate (Demoulin, 2018). This geological and topographic gradient contributes to the general south-north flow direction of the river. The Ardennes region is particularly prone to extreme precipitation events (see Fig. 1.5c), making the basin highly relevant for hydroclimatic impact studies.

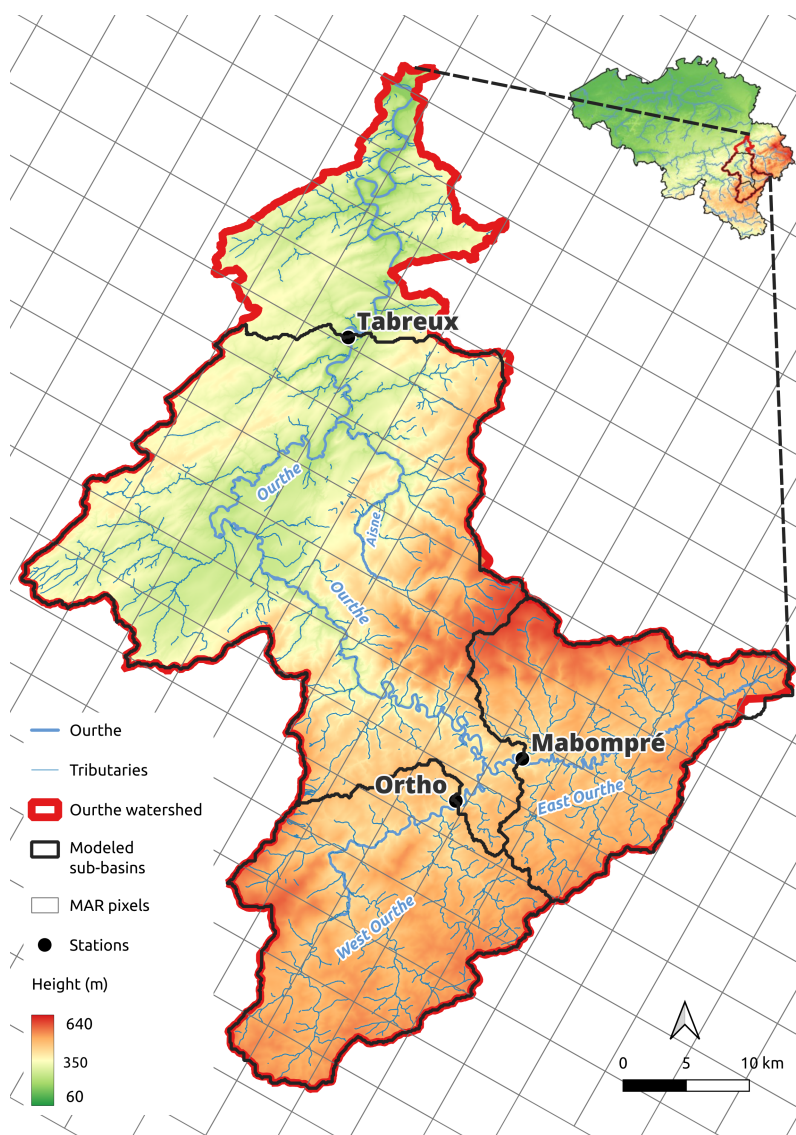


Figure 2.4: Ourthe watershed topography with the modelled sub-basins. Polygons representing MAR grid cells are also shown.

The Ourthe basin is also well documented in the hydroclimatic literature and has been widely used for hydrological modelling studies in Belgium (see Table 1.1). In addition, the basin is only weakly affected by large anthropogenic structures such as major dams, which makes it well-suited for conceptual rainfall-runoff modelling.

At the Tabreux station, the contributing drainage area is 1615.03 km² and is composed of 41.8% forest, 37.5% grassland, 10.7% cropland, 9.3% urban areas, and 0.8% river surface. These land-use fractions are derived from the preprocessing tools implemented within the WOLFHydro framework and are based on the most recent Walloon land-use dataset. However, in a lumped modelling approach, land-use characteristics are not represented explicitly in space. Instead, their influence on processes such as runoff partitioning and water storage is implicitly accounted for through the parameters of the hydrological models.

To better represent the internal dynamics of the catchment, the Ourthe basin is subdivided into three sub-basins corresponding to the selected gauging locations. Discharge simulations are evaluated at the Mabompré, Ortho, and Tabreux gauging stations. For each hydrological model, the sub-basins are calibrated independently, resulting in a specific parameter set for each sub-basin. The two upstream sub-basins, associated with Mabompré and Ortho, provide discharge contributions to the downstream Tabreux sub-basin. The resulting discharge contributions are then combined to reproduce the observed hydrograph at the final outlet, Tabreux, following a semi-distributed modelling approach.

2.2 Bias adjustment of climatic inputs

In this section, we describe the bias-adjustment procedures applied to MAR outputs in order to make them suitable for both extreme value analyses and hydrological modelling.

First, daily temperature and precipitation are corrected using quantile mapping techniques. The observational reference used for this correction is the gridded RMI dataset (Journée et al., 2023), ensuring consistency between simulated and observed statistical distributions.

For temperature, only the daily mean temperature is bias-adjusted in this study, while daily minimum and maximum temperatures are left unchanged. This represents a methodological simplification, as the latter are subsequently used to estimate potential evapotranspiration within WOLFHydro. The potential impact of separately bias-adjusting temperature extremes could be investigated in future work.

Second, the distribution of precipitation among the hours within each day is adjusted using an exaggeration function, allowing the generation of sub-daily precipitation suitable for forcing the hourly hydrological models implemented within WOLFHydro.

2.2.1 Quantile mapping over daily outputs

Daily precipitation and temperature simulated by MAR are bias-adjusted using a quantile mapping approach. More specifically, the Equidistant Cumulative Distribution Function matching (EDCDFm) method (Pierce et al., 2015) is applied.

In this framework, projected values V_{sim} are corrected using a multiplicative change factor CF . The method consists in locating the position of V_{sim} within the simulated

2. Tools and methods

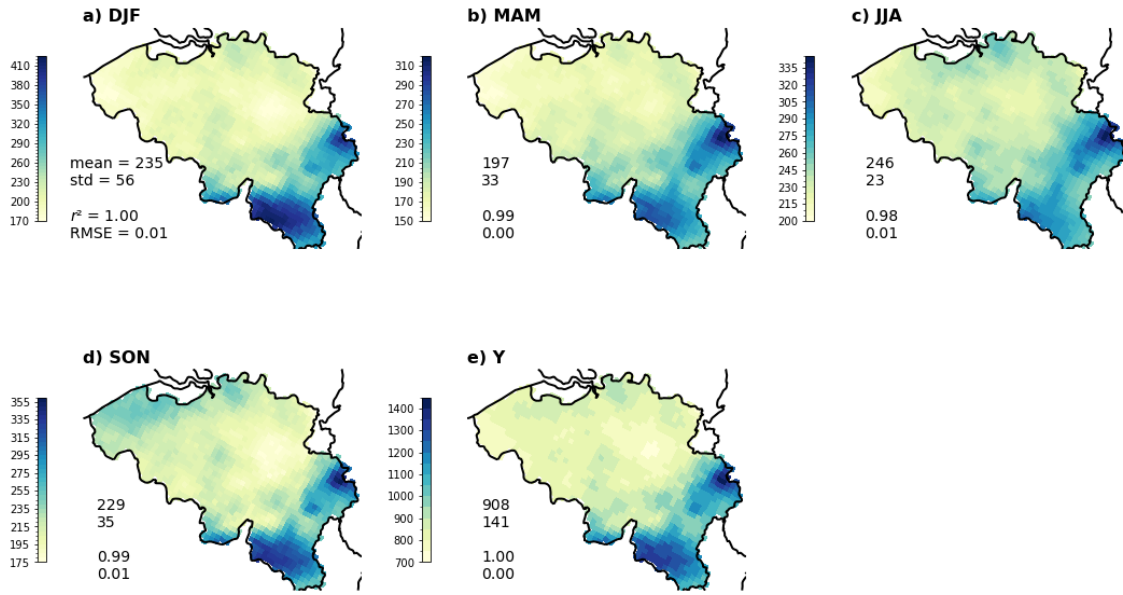


Figure 2.5: Seasonal accumulated precipitation (mm) simulated by bias-adjusted MAR-ERA5 over the period 1981-2010 for (a) DJF, (b) MAM, (c) JJA, (d) SON, and (e) annual (Y). For each panel, the spatial mean, standard deviation (std), coefficient of determination (r^2), and RMSE (mm day⁻¹) are indicated.

cumulative distribution function (CDF) computed over a 30-year period. This is achieved by identifying two consecutive empirical quantiles $q_{sim,i}$ and $q_{sim,i+1}$ such that:

$$q_{sim,i} < V_{sim} < q_{sim,i+1}.$$

The correction factor CF is then derived from the corresponding observed quantiles over the reference period (1981-2010), denoted $q_{obs,i}$ and $q_{obs,i+1}$, and the simulated quantiles over the same historical control period, denoted $q_{cont,i}$ and $q_{cont,i+1}$:

$$CF = \left[\frac{q_{obs,i}}{q_{cont,i}} \cdot W_i + \frac{q_{obs,i+1}}{q_{cont,i+1}} \cdot W_{i+1} \right] \quad (2.2)$$

where W_i and W_{i+1} are complementary interpolation weights defined as

$$W_i = 1 - \frac{V_{sim} - q_{sim,i}}{q_{sim,i+1} - q_{sim,i}}, \quad (2.3)$$

$$W_{i+1} = 1 - W_i. \quad (2.4)$$

The corrected value is then obtained as

$$V_{corr} = V_{sim} \cdot CF. \quad (2.5)$$

This formulation allows a smooth interpolation between quantiles, ensuring a continuous correction of the simulated values across the distribution.

The empirical distributions are computed on a seasonal basis. Specifically, they are

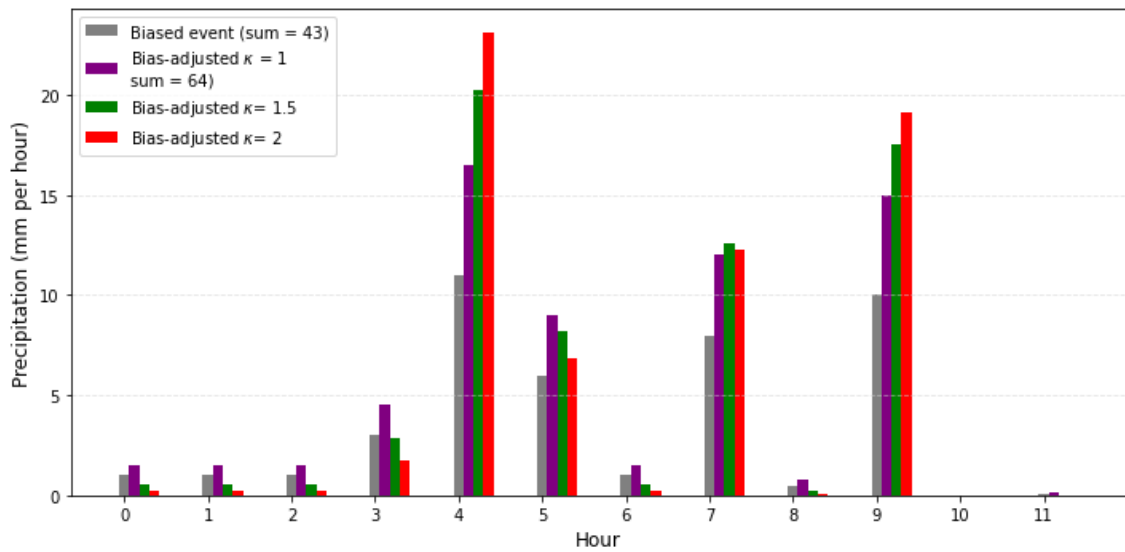


Figure 2.6: Illustration of the hourly exaggeration function used in the bias-adjustment procedure to correct the intra-day variability of hourly precipitation intensity.

derived over 30-year periods using a 3-month moving window centred on the day to be corrected. This approach accounts for the seasonal variability of MAR biases, which differ substantially throughout the year (see Tables 2.3 and 2.2).

In addition to the general correction, particular attention is given to the representation of extreme precipitation. To better capture the upper tail of the distribution, values exceeding the 97th percentile are treated separately. For these high precipitation days, the position of V_{sim} is evaluated within the empirical distribution constructed from values above this threshold.

This approach provides a more detailed representation of the upper tail, which is particularly important given the heavy-tailed nature of precipitation distributions.

The choice of the 97th percentile is empirical, but justified by the need to ensure sufficient sampling of extreme events. Hence, it allows the selection of approximately three events per year, providing a sufficiently large sample for characterizing the upper tail over the 30-year reference periods.

This bias-adjustment procedure is applied to all MAR outputs. Its performance in reproducing seasonal precipitation statistics is illustrated in Fig. 2.5. As expected, given that the correction is calibrated over the reference period, the bias-adjusted MAR-ERA5 seasonal precipitation closely matches the RMI gridded dataset over 1981-2010.

2.2.2 Exaggeration of the precipitation hourly signal

As mentioned in Section 2.1.3, one objective of this thesis is to assess the limits of the modelling chain composed of MAR and WOLFHydro. In this context, hydrological simulations are performed at the highest temporal resolution available in the MAR outputs, namely the hourly time scale.

However, MAR is known to perform poorly in representing sub-daily precipitation extremes. In particular, de Vyver et al. (2021) showed that hourly precipitation intensities

are strongly underestimated, with biases substantially larger than those found at the daily scale.

Consequently, although daily precipitation is first bias-adjusted using quantile mapping, an additional correction is required to obtain realistic sub-daily precipitation suitable for hydrological modelling.

The objective of this second step is therefore to reconstruct a more realistic hourly distribution of precipitation intensity within each day while preserving the corrected daily totals. This correction must also be applicable in a consistent manner to both MAR-ERA5 and MAR simulations forced by ESMs.

Rather than applying a full hourly quantile mapping, which would require combining heterogeneous observational datasets (e.g., RMI and SPW stations) and introduce additional complexity, we adopt a simpler approach. The method consists of exaggerating the hourly precipitation signal using a weighting function applied to hourly precipitation fractions.

Hence, for each day, once the daily precipitation has been bias-adjusted, we obtain a corrected daily total denoted $P_{c,d}$. The original hourly precipitation values within that day, denoted $P_{b,h}$ for hour h , are then redistributed in order to match this corrected daily total while modifying the intra-day variability.

The corrected hourly precipitation $P_{c,h}$ is defined as:

$$P_{c,h} = P_{c,d} \cdot \frac{P_{b,h}^\kappa}{\sum_{h=1}^{24} P_{b,h}^\kappa} \quad (2.6)$$

where:

- $P_{c,h}$ is the corrected precipitation at hour h ,
- $P_{c,d}$ is the bias-adjusted daily precipitation total,
- $P_{b,h}$ is the original (biased) hourly precipitation at hour h ,
- κ is an exponent controlling the exaggeration of the intra-day variability,

The influence of the parameter κ is illustrated in Fig. 2.6. When $\kappa = 1$, the original (biased) intra-day distribution of precipitation is preserved. For $\kappa > 1$, the contrast between low and high hourly precipitation intensities increases: high-intensity hours are amplified, while low-intensity values are further reduced. As a result, larger κ values lead to a progressive exaggeration of the intra-day precipitation variability and an enhancement of sub-daily extremes.

The optimal value of κ is determined by comparing the bias-adjusted hourly precipitation to RMI station observations. Figure 2.7 presents quantile-quantile (QQ) plots for 17 stations across Belgium. Three candidate values of κ were tested: 1.5, 1.75 and 2. No single value of κ provides a perfect fit at all locations: extreme quantiles may be overestimated at some stations and underestimated at others. However, $\kappa = 1.75$ is retained as a compromise, providing the best overall agreement between simulated and observed hourly precipitation across Belgium.

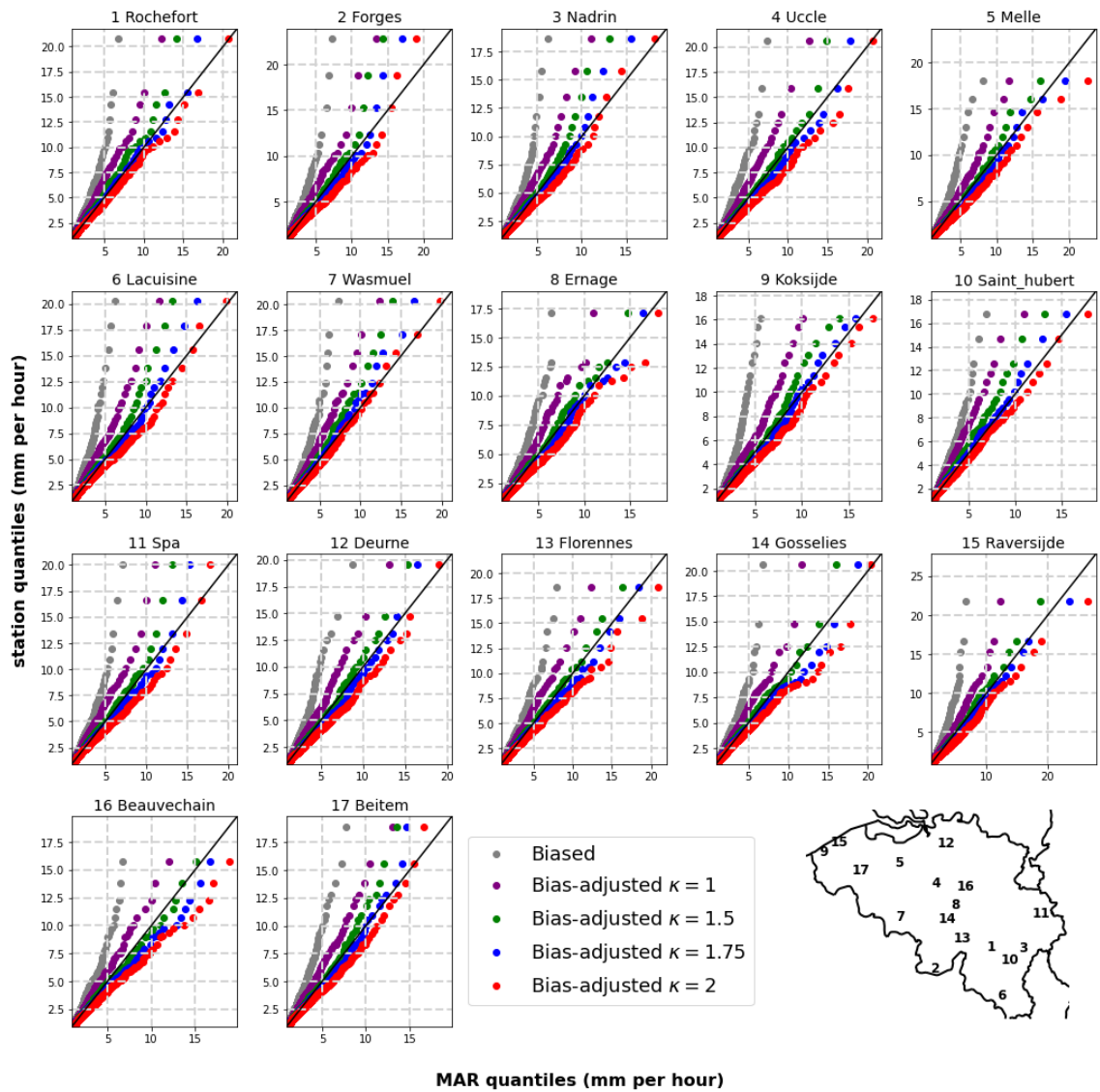


Figure 2.7: Quantile-quantile (QQ) plots comparing bias-adjusted MAR-ERA5 precipitation to RMI station observations for 17 stations. Only hourly precipitation exceeding 1 mm is considered. Results are shown for different values of the exaggeration exponent κ used in the bias-adjustment procedure. A map showing station locations and numbering is provided in the bottom-right panel.

2.3 Statistical tools

This section provides an overview of the statistical tools used throughout this manuscript for the analysis of the diverse outputs. We first present the extreme value analysis techniques employed in this thesis and then present the Poisson regression.

2.3.1 Extreme value analysis

Extreme value analysis (EVA) is used in this thesis to characterise the statistical behaviour of rare hydrometeorological events. In particular, EVA is applied to both precipitation outputs from MAR and discharge simulations obtained with the WOLFHydro modelling framework (Chapters 3 and 5).

Two classical EVA approaches are considered: the Annual Maxima (AM) method and the Peaks Over Threshold (POT) method. These approaches rely on fitting stationary statistical distributions to extreme values, namely the Generalized Extreme Value (GEV) distribution for the AM approach and the Generalized Pareto Distribution (GPD) for the POT approach.

In this thesis, the focus is placed on the stationary formulations of these methods. Non-stationary versions have also been tested, particularly in Chapter 3, but they do not significantly alter the main conclusions. A more detailed discussion of these non-stationary approaches is provided in Appendix A.

In the following, the theoretical background and implementation of each approach are briefly described.

2.3.1.1 Annual Maxima (AM) approach

The first approach is based on annual maxima (AM), for which a Generalized Extreme Value (GEV) distribution is fitted (Coles, 2001). The GEV distribution relies on three parameters: the location parameter μ , the scale parameter σ , and the shape parameter ξ (Fig. 2.8a-c). A variation of μ shifts the distribution (Fig. 2.8a), a variation of σ affects its spread (Fig. 2.8b), and ξ controls the weight of the distribution tail (Fig. 2.8c).

The return level R_T , associated with a return period T (in years), is obtained using (Coles, 2001):

$$R_T = \begin{cases} \mu - \frac{\sigma}{\xi} \left[1 - \{-\log(1-p)\}^{-\xi} \right] & , \text{ for } \xi \neq 0 \\ \mu - \sigma \log \{-\log(1-p)\} & , \text{ for } \xi = 0 \end{cases} \quad (2.7)$$

with $p = \frac{1}{T}$.

2.3.1.2 Peaks Over Threshold (POT) approach

The second approach is based on the Peaks Over Threshold (POT) method, where a Generalized Pareto Distribution (GPD) is fitted to threshold exceedances (Coles, 2001). Selecting values exceeding a given threshold u provides a larger sample size and therefore more information on the upper tail of the distribution.

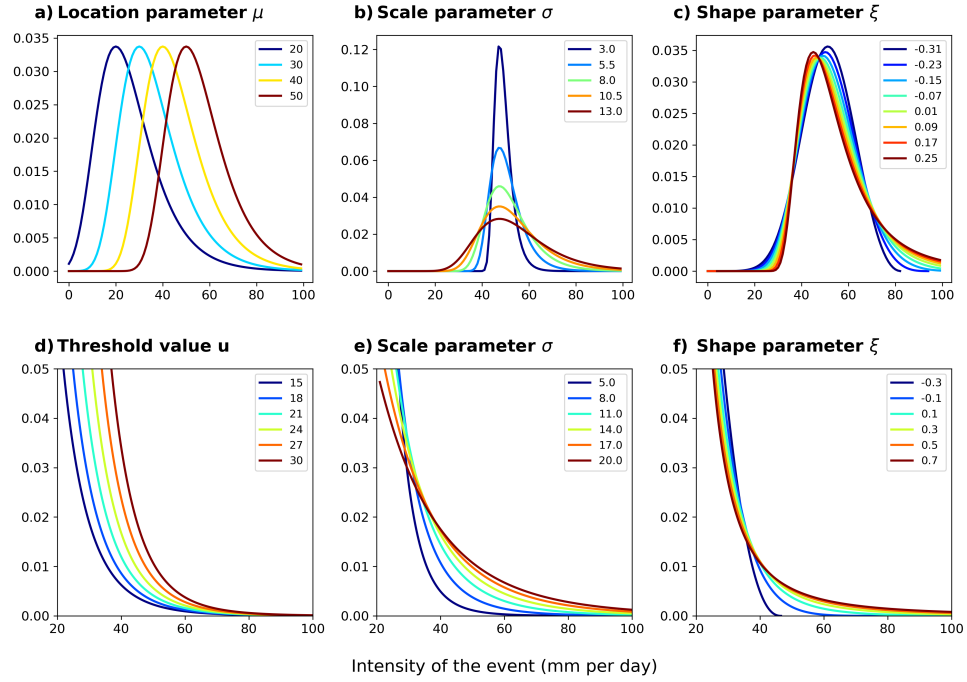


Figure 2.8: (a-c) Impact of the GEV parameters on its PDF (Probability Density Function). (d-f) same, but for the GPD.

To ensure that the selected events are statistically independent, exceedances are grouped into clusters. A cluster is defined as a sequence of threshold exceedances separated by less than a given number of days. For instance, in Chapter 3, a separation period of 7 days was used. Although this choice remains somewhat empirical, it represents a reasonable compromise for isolating events associated with independent synoptic situations.

Once clusters have been identified, only the maximum value within each cluster is retained. In this thesis, thresholds are defined based on empirical distribution quantiles and selected such that the mean number of independent cluster maxima per year $\lambda > 2$. This requirement keeps the main advantages of the POT approach compared to the AM method, which relies on a single event per year (Grandry et al., 2020; Cunnane, 1979; Taesombut, Viraphol and Yevjevich, Vujica M., 1978). The sensitivity to the threshold choice is further discussed in the paper associated with Chapter 3 (Appendix A).

The GPD is defined by two parameters: the scale parameter σ and the shape parameter ξ . Their influence on the probability density function is similar to that of the GEV distribution (Figs. 2.8e and f).

After fitting, the return level R_T is obtained as:

$$R_T = \begin{cases} u + \frac{\sigma}{\xi} [(\lambda T)^\xi - 1] & , \text{ for } \xi \neq 0 \\ u + \sigma \log(\lambda T) & , \text{ for } \xi = 0 \end{cases} \quad (2.8)$$

where u is the threshold and λ the mean number of exceedances per year.

The parameters of the GEV and GPD distributions are estimated using Maximum Likelihood Estimation (MLE), implemented with the *ismev* R package (Gilleland, 2003).

2.3.2 Poisson regression

In Chapter 4, the temporal evolution of count variables (e.g., the number of cut-off lows per year) is analysed. To model the evolution of the expected number of events γ , a Poisson distribution is assumed (Tramblay et al., 2013; Grandry et al., 2020; Brajkovic et al., 2025). The probability of observing n events is given by:

$$P(n) = \frac{\exp(-\gamma) \gamma^n}{n!}, \quad (2.9)$$

where γ denotes the expected annual number of events.

To account for potential temporal trends, γ is modeled as an exponential function of time:

$$\gamma = \exp(\gamma_0 + \gamma_1(y - y_0)), \quad (2.10)$$

where y is the calendar year, y_0 is a reference year corresponding to the beginning of the study period, and γ_0 and γ_1 are regression coefficients estimated from the data. The parameter γ_1 controls the temporal trend in the expected number of events.

The statistical significance of this trend is assessed using a Wald test applied to the slope parameter γ_1 . A 95% confidence level is adopted throughout this thesis, such that p -values below 0.05 indicate a statistically significant trend.

CHAPTER 3

Extreme value analysis applied to MAR outputs

This chapter investigates the evolution of extreme precipitation intensity and frequency in Belgium using outputs from the regional climate model MAR. It combines results from a published study and ongoing research, following the semi-article structure of this thesis.

Section 3.1 summarizes the main findings of a peer-reviewed paper included in Appendix A of this manuscript (Brajkovic et al., 2025). This study focuses on daily precipitation extremes, for which Extreme Value Analyses (EVA) are applied. The results show that 1-day accumulated precipitation intensity is projected to increase on average following the Clausius-Clapeyron (CC) scaling, corresponding to an increase of approximately 7% per degree of global warming. However, a substantial inter-model variability is identified, primarily originating from differences in the forcing Earth System Models (ESMs).

Section 3.2 extends this analysis by investigating extreme precipitation intensity anomalies across multiple accumulation timescales, ranging from 1 to 72 hours. This section explores how the identified trends in precipitation extremes translate into changes in Intensity-Duration-Frequency (IDF) curves for several locations across Belgium.

3.1 Impacts of global warming on daily precipitation

For conciseness, this summary focuses on the POT approach used in the published paper. However, the AM approach yields very similar results and conclusions, which can be found in the paper itself and in its Supplementary Material. In the present summary, we therefore also rely on the POT approach.

3. Extreme value analysis applied to MAR outputs

Models	2-year return level						20-year return level					
	Mean	std	MB	MB(%)	RMSE	r^2	Mean	std	MB	MB(%)	RMSE	r^2
Obs	1.7	0.2	—	—	—	—	2.9	0.4	—	—	—	—
MAR-ERA5	1.2	0.2	-0.5	-30.5	0.54	0.74	2.0	0.3	-0.8	-29.7	0.92	0.48
MAR-MPI	1.1	0.2	-0.6	-36.0	0.63	0.68	1.7	0.2	-1.1	-40.2	1.22	0.09
MAR-EC3	1.1	0.2	-0.6	-34.9	0.61	0.79	1.8	0.3	-1.1	-38.7	1.20	0.04
MAR-MIR	1.2	0.2	-0.5	-27.6	0.51	0.61	2.1	0.3	-0.7	-25.9	0.86	0.26
MAR-NS2	1.3	0.2	-0.4	-25.9	0.47	0.70	2.0	0.3	-0.8	-29.7	0.94	0.20
MAR-CR5	1.4	0.2	-0.4	-20.9	0.39	0.69	2.2	0.3	-0.6	-22.5	0.74	0.44
MAR-IPSL	1.1	0.2	-0.6	-37.7	0.66	0.68	1.7	0.3	-1.1	-39.0	1.22	0.03

Table 3.1: Statistical parameters computed spatially over Belgium for the control period 1981–2010: mean value (mm h^{-1}), standard deviation (std) (mm h^{-1}), mean bias (MB) (mm h^{-1} and %), root mean square error (RMSE) (mm h^{-1}), and coefficient of determination (r^2) for the 2-year and 20-year return levels (precipitation accumulation duration = 24 h).

3.1.0.1 Representation of extreme precipitation by MAR

Applying an Extreme Value Analysis (EVA) to MAR outputs highlights the model's weaknesses in representing extreme precipitation events. When compared to the observational gridded dataset over the 1981–2010 period, MAR-ERA5 exhibits mean biases of approximately -30% for both the 2-year and 20-year daily precipitation return levels, averaged over Belgium (Table 3.1).

When MAR is forced by Earth System Models (ESMs), these biases are generally more pronounced for higher return levels, for which they can reach up to 40% on average, as illustrated by the MAR-MPI simulation for the 20-year return level. The main reason identified in this study is that, for higher return periods, MAR's ability to represent extreme precipitation strongly depends on the capacity of the driving ESMs to reproduce extreme large-scale weather patterns at the boundaries of the regional model domain.

The frequency of such weather patterns depends both on the physical parameterizations of each ESM and on its internal variability. This internal variability, which reflects the natural variability observed in the climate system, leads to periods that are more or less dominated by extreme weather types. In the historical control simulations covering 1981–2014, this variability can strongly affect the estimation of return levels, resulting in very low spatial correlation coefficients in some cases (Table 3.1).

These biases further motivated the use of bias-adjustment methods for MAR outputs in hydrological applications and extreme value analyses.

3.1.0.2 Evolution of extreme precipitation intensity

The evolution of the statistical parameters derived from the POT approach suggests an increase in both the frequency and intensity of daily precipitation extremes. In this study, the EVA is applied independently to each MAR grid cell using sliding 30-year windows, ranging from 2021–2050 to 2071–2100 with a 10-year increment. Considering the six forcing ESMs and four emission scenarios, this results in a total of 144 analysed periods.

On average, the POT threshold exhibits an increase of approximately 5.5% per additional degree of global warming (Fig. 3.1a). The scale parameter σ shows an even stronger increase, reaching about 11% per degree of global warming. These parallel

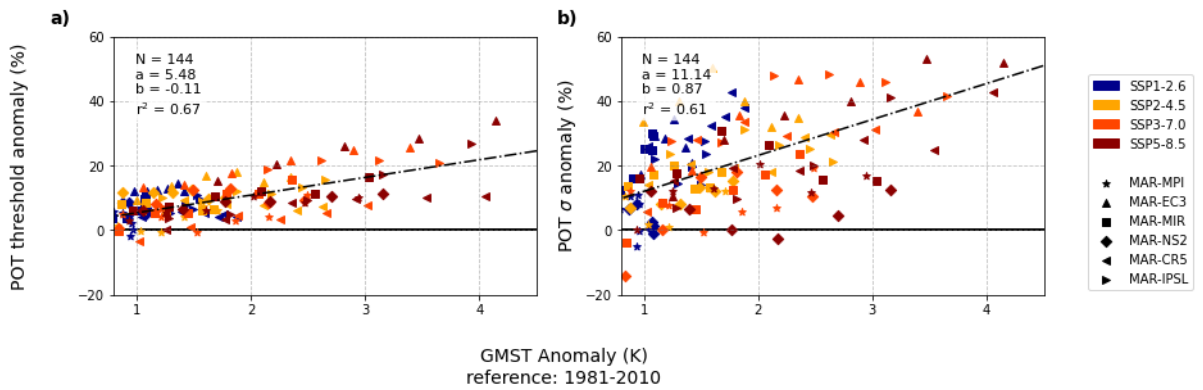


Figure 3.1: Anomalies relative to 1981-2010 averaged over 30-year moving windows (from 2021-2050 to 2071-2100 with a 10-year time-step) for (a) POT threshold, (b) GPD-derived σ . A linear regression is applied to the data clouds (black dotted line of slope a and intercept b). The impact of those parameters on the PDFs is illustrated in Fig. 2.8. Reproduced from Brajkovic et al. (2025).

evolutions indicate a general shift of extreme precipitation distributions towards higher intensities.

The evolution of the shape parameter ξ is not discussed further, as it is strongly affected by the internal variability of the models, leading to a highly irregular behaviour with respect to global warming.

These distributional changes translate into an increase in the computed return level intensities (Fig. 3.2). The evolution of daily extreme precipitation generally follows the Clausius-Clapeyron scaling, corresponding to an increase of about 7% per additional degree of global warming (Fig. 3.2a).

The impact of each model's internal variability is clearly visible at this stage, as the relationship between intensity and global warming becomes less robust with increasing return period. For instance, the coefficient of determination drops to $r^2 = 0.47$ for the 20-year return levels, reflecting the growing influence of internal variability on rarer extremes (Fig. 3.2a).

This 7% increase represents an average value over Belgium. However, the results suggest that this scaling can vary spatially, even at the scale of a relatively small country such as Belgium. The pixel-wise estimation of the scaling coefficient reveals that the southern border of the Ardennian Massif, as well as the Campine Plateau and the coastal region, may experience super-Clausius-Clapeyron scaling, a feature that becomes more pronounced for higher return periods.

Locally, in these identified regions, the scaling reaches values of up to 15% per degree of global warming (Fig. 3.2b).

The fact that higher return levels are strongly affected by the internal variability of each ESM motivates an alternative estimation of the 20-year return levels, computed over the entire available simulation period (2015-2100) and subsequently averaged across the six MAR configurations (Fig. 3.3a). This approach confirms that, under the SSP5-8.5 scenario, the southern Ardennes exhibit levels of extreme precipitation activity comparable to those of the northern Ardennes under the SSP1-2.6 scenario (Fig. 3.3a).

For identified extreme-precipitation hot spots, namely the northern and southern elevated regions of the Ardennian Massif, future 20-year return levels reach at least

3. Extreme value analysis applied to MAR outputs

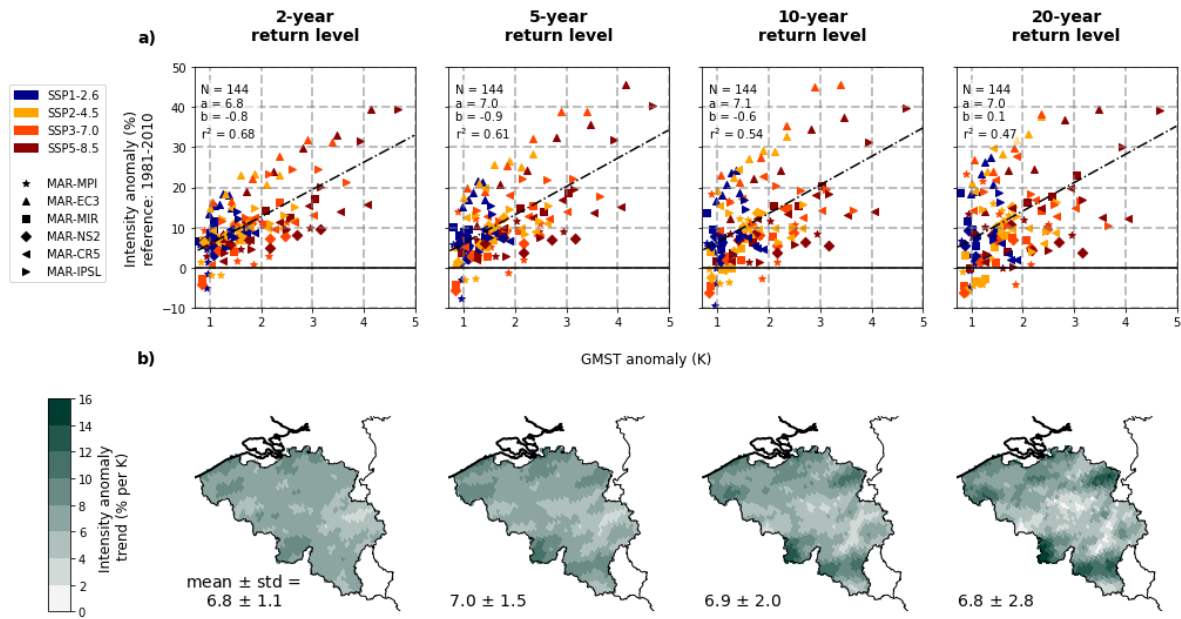


Figure 3.2: (a) Intensity anomalies over 30-year moving windows relative to 1981-2010 for different return levels using the POT approach. A linear regression is applied to the data clouds (black dotted line of slope a and intercept b). (b) Same regression as in (a) but computed at each pixel location. The written values represent the map average $\pm$ the standard deviation. The results using the AM approach are very close and displayed in Supplementary Figure S4 in Appendix B. The trends for each model are plotted on Supplementary Figure S6 in Appendix B. Reproduced from Brajkovic et al. (2025).

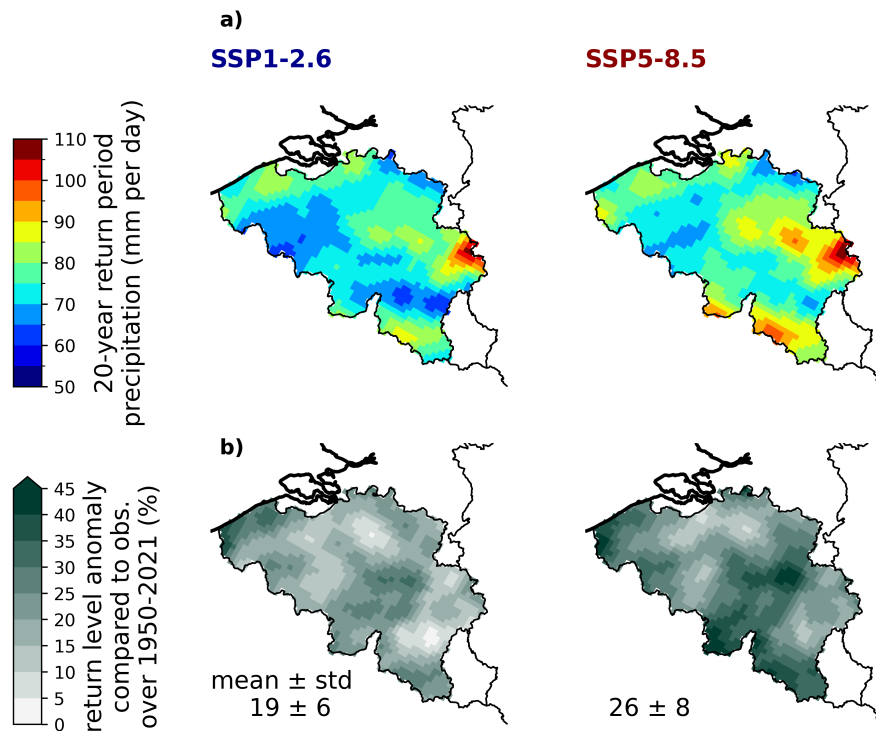


Figure 3.3: (a) 20-year return level computed over 2015-2100 under SSP1-2.6 and SSP5-8.5 scenarios. (b) 20-year return anomaly as compared to 1951-2021. Return levels are calculated as the mean return levels of the six used models. The written values represent the map average $\pm$ the standard deviation. Reproduced from Brajkovic et al. (2025).

100 mm day⁻¹ under the SSP5-8.5 scenario. These changes correspond to relative increases exceeding 40% in some areas (Fig. 3.3b). In this framework, the spatial pattern of anomalies becomes less geographically structured. This likely results from the fact that all simulated events over the 2015-2100 period are considered, and that, within each ESM pseudo-reality, extreme precipitation events may occur at different locations and times.

3.2 Extension to other accumulation timescales

Although MAR underestimates sub-daily precipitation accumulations even more (de Vyver et al., 2021), the EVA is extended to shorter accumulation durations in order to assess whether similar scaling behaviours are obtained across timescales.

In a first step, the same POT framework described previously is applied to 1-, 3-, 6-, 12-, 48-, and 72-hour accumulations. Thresholds are selected following the same principle of retaining approximately two exceedances per year, thereby preserving the advantages of the POT approach over the AM approach (Cunnane, 1979; Brajkovic et al., 2025). The corresponding thresholds are the 99.95th, 99.9th, 99.8th, 99.7th, 99th, and 98.5th percentiles, respectively.

In a second step, the scaling coefficients derived from the regression against global mean surface temperature are applied to IDF curves constructed from bias-adjusted MAR-ERA5 outputs over the 1940-2021 period. This procedure allows us to illustrate the hydrological implications of the projected changes in extreme precipitation intensity.

3.2.1 Extreme precipitation anomaly for other time accumulations

The scaling of extreme precipitation with the GMST anomaly depends on the considered precipitation duration. For 1-hour and 72-hour accumulations, precipitation intensity increases by approximately 5.5% per additional degree of global warming, regardless of the return period considered (Fig. 3.4a). For 6-hour precipitation, the scaling is closer to that obtained for daily precipitation, reaching values around 7% per degree of warming (Fig. 3.4a). As previously observed, a larger spread between models is found for higher return periods, reflecting the influence of each ESM's internal variability. Spatially, and consistently with the 24-hour results, the pixel-wise estimation of scaling values exhibits greater variability for higher return periods (Fig. 3.4b).

Therefore, the average scaling values shown in Fig. 3.4a should be interpreted as spatial means over Belgium. At the pixel level, however, super-Clausius-Clapeyron scaling can emerge, particularly for higher return periods. In the most extreme cases, scaling values reach up to 15% per degree of warming over the elevated regions of the Ardennian Massif.

The spatial pattern identified for 24-hour precipitation is also apparent for 6- and 72-hour accumulations, with the Ardennian Massif and certain flatter coastal areas exhibiting super-CC scaling. This spatial coherence is less pronounced for 1-hour precipitation, for which the role of topographic gradients appears to be less dominant in modulating precipitation intensities. In this case, other topoclimatological factors, such as land use or surface roughness, may contribute more substantially.

The increasing spread in scaling factors for higher return periods once again illustrates the difficulty of deriving robust scaling estimates for rare extremes. Since these analyses

3. Extreme value analysis applied to MAR outputs

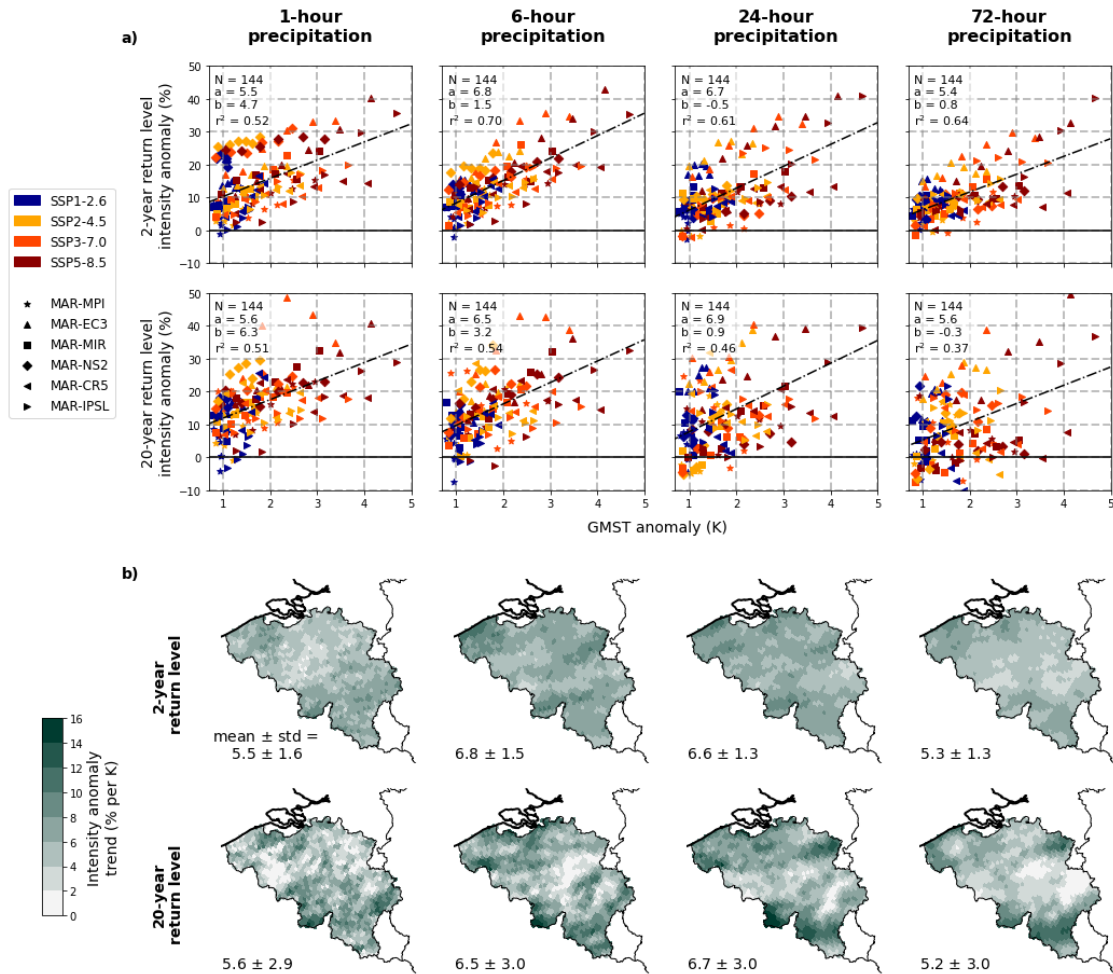


Figure 3.4: Extreme precipitation anomalies as a function of the GMST anomaly. (a) Anomalies averaged over Belgium with a fitted linear regression (slope a , intercept b); r^2 denotes the coefficient of determination and N the number of data points. Four accumulation periods are considered (columns) and two return periods (rows). (b) Pixel-wise values of the regression slope a . Reproduced from Brajkovic et al. (2025).

rely on sliding 30-year windows, return periods higher than 20 years are not investigated in this framework. Nevertheless, in the next subsection, for the purpose of illustrating hydroclimatic impacts of climate change, a simplified uniform scaling of 7% per degree of warming is adopted across all durations and return periods.

3.2.2 Impact of global warming on IDF curves

As an illustrative example, IDF curves are constructed for eight representative locations across Belgium (Fig. 3.5): Brussels (a), Liège (b), Koksijde (c), Ghent (d), Antwerp (e), Spa (f), Saint-Hubert (g), and Mons (h). These sites were selected to capture a broad range of topographic settings and hydroclimatic contexts.

Return levels are derived from the bias-adjusted MAR-ERA5 product over the 1940-2021 period. However, as sub-daily precipitation is less reliably represented in MAR and the intra-day variability is partially corrected using a weighting function, caution is required when interpreting absolute intensity values for durations shorter than 6 hours. Hence, these sub-daily return levels are generally lower than those reported in the reference RMI IDFs (de Vyver, 2013), even after applying the scaling associated with global warming.

Given that the observational reference period is centred around 1961-1990, the baseline IDFs (dotted lines) are interpreted as representative of a +0.5 K global warming level relative to the IPCC pre-industrial reference period of 1850-1899.

To illustrate the impact of climate change on design precipitation intensities, a uniform scaling of 7% per degree of global warming is applied to all return levels. This yields the +3 K IDFs (solid lines), corresponding to a scaling consistent with Clausius-Clapeyron expectations.

As expected, Spa exhibits the highest return levels for durations longer than 6 hours because of the topographic barrier of the Ardennes (Fig. 3.5f), while Koksijde, located on the coast (Fig. 3.5c), exhibits the least intense IDFs. Overall, applying a uniform Clausius-Clapeyron scaling to all return levels substantially reduces the return periods associated with a given precipitation intensity. In practical terms, for all locations considered, the present-day (+0.5 K) 200-year return level corresponds roughly to the future (+3 K) 100-year return level. In some cases, this shift is even more pronounced. For example, in Mons (Fig. 3.5h), the projected 20-year return level under +3 K warming reaches intensities comparable to the historical 50-year return level.

3.3 Implications and limitations

Over Belgium, extreme precipitation intensity generally follows Clausius-Clapeyron (CC) scaling, corresponding to an increase of approximately 5-7% per additional degree of global warming depending on the accumulation duration considered. Nevertheless, caution is required when interpreting the absolute values of sub-daily precipitation extremes. Hourly precipitation peaks are more strongly underestimated by MAR than daily extremes, and the hourly correction procedure does not provide equivalent performances across all locations (see Fig. 2.7).

3. Extreme value analysis applied to MAR outputs

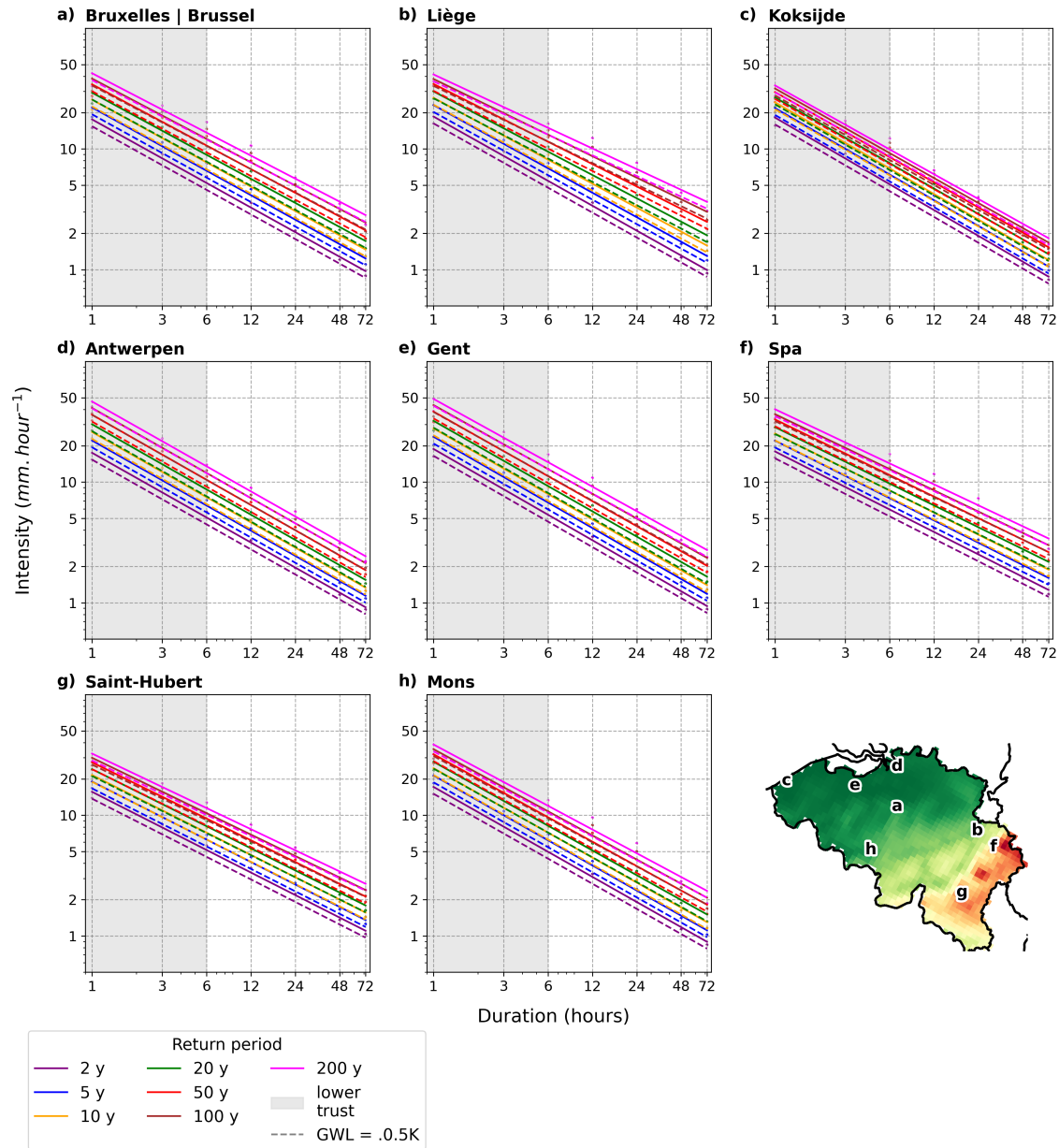


Figure 3.5: Illustrative Intensity-Duration-Frequency (IDF) curves in a +3 K world for eight Belgian locations: (a) Brussels, (b) Liège, (c) Koksijde, (d) Ghent, (e) Antwerp, (f) Spa, (g) Saint-Hubert, and (h) Mons. Points represent the intensities associated with different return periods (2 to 200 years), obtained using the POT approach. Solid coloured lines represent the corresponding exponential interpolation. Dashed lines illustrate the corresponding intensities under a +0.5 K global warming level (GWL), derived from the regression of return levels against GMST anomalies. The locations are shown on the topographic map in the bottom-right corner (same legend as Fig. 2.1).

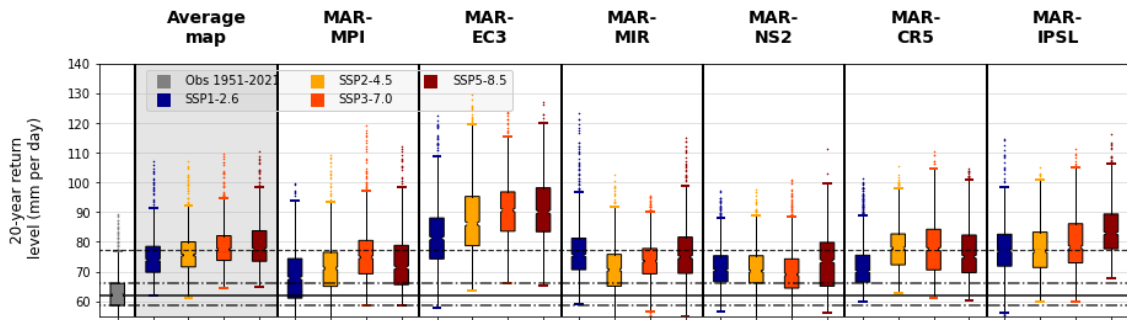


Figure 3.6: Boxplots associated to the return level maps of Belgium for a return period of 20 years using all the models over 2015-2100. The POT approach is used for the computation of the return levels. The grey box refers to the return level map computed using observational data over 1951-2021 (Journée et al., 2023). Visual marks (horizontal black lines) have been added so each box can be compared to the observational box. Reproduced from Brajkovic et al. (2025).

However, the spatial mapping of these scaling rates suggests that local factors, such as topographic gradients or proximity to the sea, may modulate this relationship.

The observed CC scaling highlights the dominant thermodynamic contribution of global warming to the intensification of extreme precipitation. This interpretation is further supported by Schoofs et al. (2025), who separated thermodynamic and dynamic contributions and reached similar conclusions regarding the dominant role of atmospheric moisture increase.

Although this study does not investigate return periods exceeding 20 years because of the increasing inter-model spread and internal variability affecting rarer events, previous studies suggest that thermodynamic amplification also affects exceptional extremes. For example, Mohr et al. (2023) showed that the July 2021 event, whose return period has been estimated at approximately 400 years based on observations (Kreienkamp et al., 2021), would likely have been around 7% less intense under pre-industrial climate conditions.

However, the six climate models used in this study exhibit substantial inter-model variability when estimating absolute return levels over the 2015-2100 period (Fig. 3.6). For instance, the MAR-EC3 boxplot associated with 20-year daily return levels over Belgium ranges approximately from 67 to 120 mm day^{-1} , whereas MAR-IPSL, despite exhibiting an even higher climate sensitivity, ranges from approximately 68 to 107 mm day^{-1} .

Moreover, the timing of periods characterized by enhanced extreme precipitation activity differs substantially between models. This further illustrates the non-linear response of extreme precipitation to global warming, with the occurrence of the most intense events remaining strongly influenced by internal climate variability and stochastic atmospheric dynamics.

As observed during the July 2021 floods in northwestern Europe and the October 2024 event in Valencia, cut-off lows (COLs) can generate exceptionally intense precipitation events and therefore strongly influence the upper tails of precipitation distributions. Consequently, Chapter 4 investigates their role in modulating the occurrence of extreme precipitation events at both European and Belgian scales. In addition, it evaluates how well the selected ESMs reproduce COL frequency and the associated precipitation intensity.

Take-home messages

- MAR systematically underestimates daily and sub-daily precipitation extremes over Belgium, motivating the use of bias-adjustment procedures.
- Daily extreme precipitation intensities generally increase following Clausius-Clapeyron scaling, corresponding to approximately 7% per additional degree of global warming.
- Scaling rates vary spatially, with elevated Ardennian regions and coastal areas locally exhibiting super-CC scaling.
- Internal climate variability strongly affects the estimation of higher return levels and contributes substantially to inter-model spread.
- Future increases in extreme precipitation intensity suggest potentially major implications for hydrological risks in Belgium.

CHAPTER 4

Evolution of cut-off low frequency

4.1 Outline of the chapter

Chapter 3 showed that the occurrence of particularly active periods in terms of extreme precipitation is largely stochastic. As a result, the timing of these periods strongly depends on the Earth System Model (ESM) used to force MAR.

In recent years, several record-breaking extreme precipitation events in Europe have been associated with cut-off lows (COLs). In this context, this chapter aims to assess the extent to which the occurrence of COLs can explain recent changes in the frequency of extreme precipitation at both European and Belgian scales. It also investigates the role of COL occurrence in driving the natural variability observed in past records and future projections.

In Section 4.2, the main findings of a submitted manuscript are summarized (Brajkovic et al. (2026); latest version currently under review). This study focuses on the evolution of COL-related extreme precipitation events across Europe. To address this question, a detection and association algorithm was developed. The full manuscript and its supplementary material are provided in Appendixes C and D.

In Section 4.3, this algorithm is applied to extreme precipitation events over Belgium.

4.2 Evolution of COL-related extreme precipitation events in Europe

A cut-off low (COL) is a cold-core upper-level depression that becomes isolated from the main westerly flow as a result of large-amplitude jet stream meanders (Palmén, 1949;

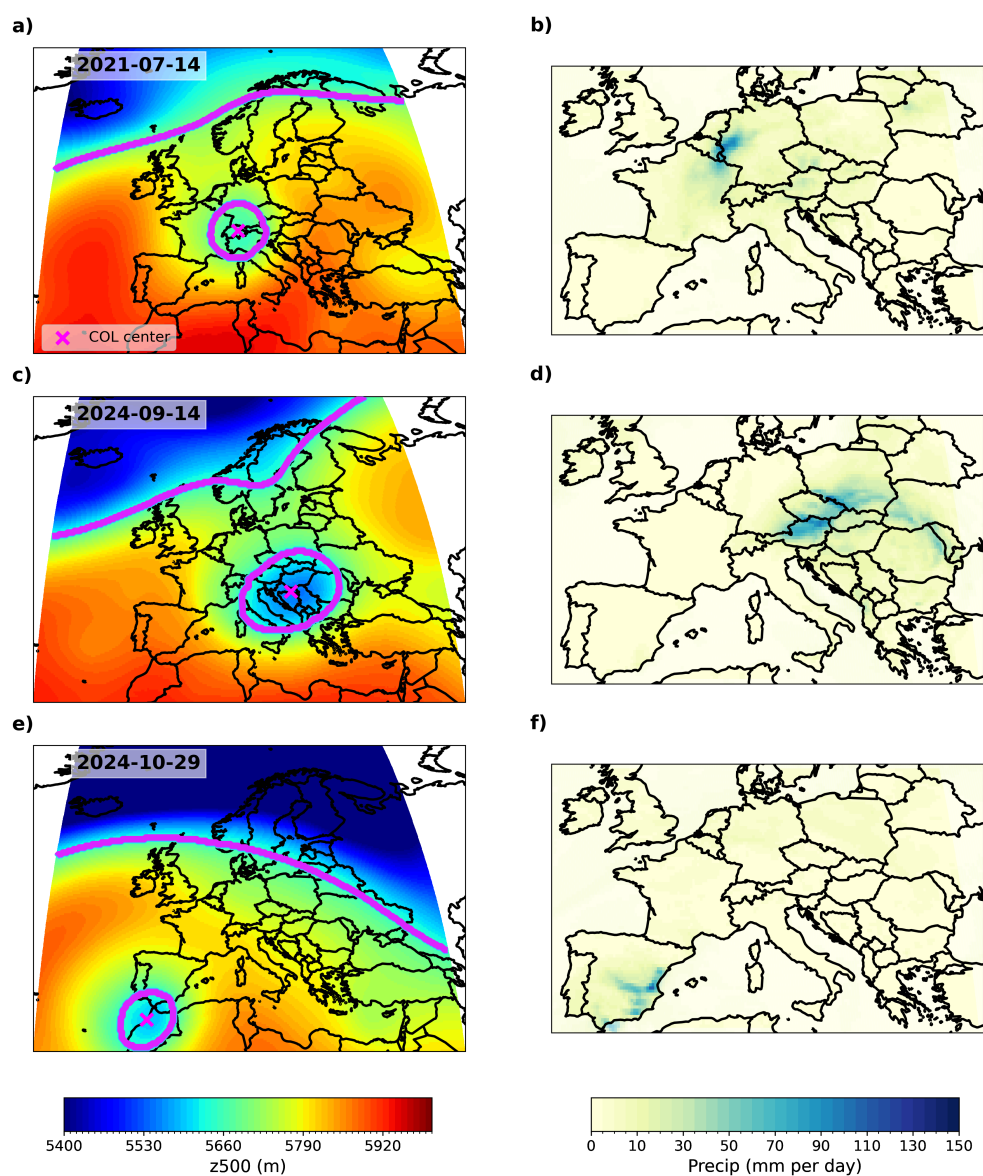


Figure 4.1: Daily Z500 fields (a, c, e) and associated daily precipitation fields (b, d, f) for 13 July 2021 (Storm Bernd), 14 September 2024 (Storm Boris), and 29 October 2024 (Storm Dana), respectively.

Eumetsat, 2024). The cold-core structure of these systems is associated with strong horizontal temperature gradients and enhanced vertical instability, both of which can substantially increase precipitation potential (Eumetsat, 2024). Assessing the evolution of their frequency in a context of global warming is therefore critical for understanding changes in extreme precipitation.

Recent extreme precipitation events in Europe have been associated with COLs, including the Bernd system (Fig. 4.1a and b), which resulted in the 2021 July floods in Belgium, the storm Boris (Fig. 4.1c and d), which affected central Europe in September 2024 (ECMWF, 2024) and the storm Dana (Fig. 4.1e and f), which affected the Valencia region in Spain in October of the same year (Rodero Astaburuaga, 2025).

However, research on the future evolution of COLs remains relatively limited, with only a few studies explicitly addressing potential changes in their occurrence. For example, Mishra et al. (2025) showed that the frequency of particularly long-lasting COLs may increase in the future, based on analyses of both historical reanalyses and Earth System Model (ESM) projections. Using NCEP (National Centers for Environmental Prediction) reanalysis data, Muñoz et al. (2020) reported significant positive trends in the annual number of COLs over the Northern Hemisphere and Europe, based on a detection algorithm relying on isolated low-pressure systems associated with eastward-propagating thermal fronts. These evolutions are believed to be linked with large-scale atmospheric circulation changes. Hence, some recent studies have suggested that Arctic amplification, i.e., the faster warming of the Arctic relative to lower latitudes, which reduces the meridional temperature gradient, may affect the characteristics of the mid-latitude jet stream (Preece et al., 2023) and, consequently, the occurrence and persistence of COLs (Mishra et al., 2025). For a more comprehensive overview of previous efforts to detect and characterize COLs, the reader is referred to the original associated paper in Appendix C

Moreover, no comprehensive climatology of COL-related extreme precipitation currently exists in the literature. To address this gap, we developed a detection and association algorithm based on image decomposition techniques, enabling the identification of COLs and their association with daily precipitation events.

The algorithm is presented in detail in the associated paper provided in Appendix C. Here, we briefly describe its main principles. Subsequently, the algorithm is applied to ERA5 reanalysis data over the period 1940-2024 to derive a climatology of COL-related precipitation. Finally, it is applied to the six CMIP6 Earth System Models (ESMs) used in this thesis (see Table 2.1), under both the historical scenario (1981-2014) and the SSP5-8.5 scenario (2015-2100).

4.2.1 Brief overview of the algorithm

The objective of the algorithm is to detect low-pressure systems that are detached from the main high-latitude westerly flow. To achieve this, each daily geopotential height field at 500 hPa (z_{500}) is analysed through a histogram-based decomposition.

More specifically, lower and upper boundary values of z_{500} are identified. Within this range, isohypses (i.e., lines of equal geopotential height) are iteratively drawn between the lower boundary, which represents the transition between polar cyclones (pink continuous line in Figures 4.2a and c), and the upper boundary, corresponding to the maximum extent beyond which anticyclonic conditions prevail.

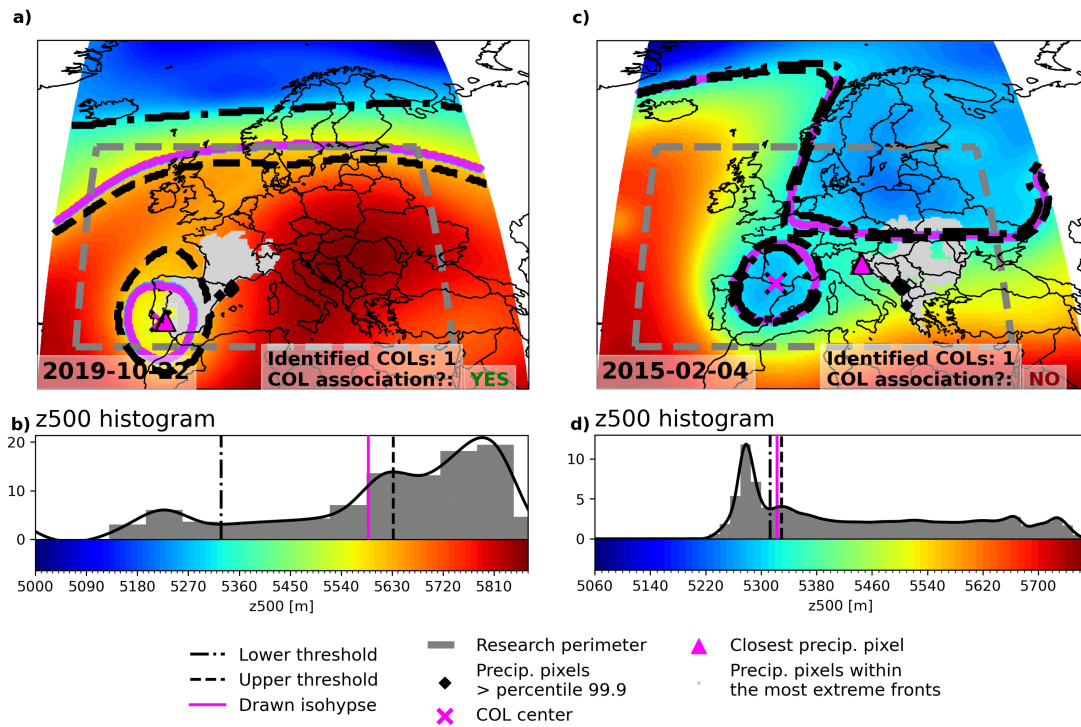


Figure 4.2: Illustration of the cut-off low (COL) detection and precipitation association methodology. (a) Example of a COL detected in the 500 hPa geopotential height (z_{500}) field on 22 October 2019. The magenta contour represents the closed isohypse retained by the iterative detection algorithm and the cross marks the centre of the identified COL. The most extreme precipitation front is associated with the detected COL because it intersects the COL area, resulting in a positive COL-precipitation association. (b) Histogram of z_{500} values showing the dynamically derived lower and upper thresholds used to guide the identification of closed contours. (c,d) Same as (a,b), but for 4 February 2015. Although a COL is successfully detected, the most extreme precipitation front does not intersect the detected COL. Consequently, no COL-precipitation association is assigned.

Through this iterative process, isolated low-pressure regions can be identified when present (enclosed pink contours in Figures 4.2a and c). To be classified as a COL, each detected region must satisfy two criteria. First, its area must exceed a critical threshold A_{crit} , set to 50,000 km² in the associated study. Second, its center must lie within the predefined study domain (Fig. 4.2a).

Once COLs are identified for a given day, the corresponding precipitation field is analysed to associate precipitation with these systems. A COL is considered to be associated with precipitation if at least one land grid cell within the most extreme precipitation front lies within its spatial extent. The most extreme precipitation fronts are defined as continuous surfaces of pixels which contain at least one grid cell exceeding the daily 99.9th spatial precipitation percentile over the European domain (Figs. 4.2a and b). This methodology ensures that the identified precipitation front contains the most extreme precipitation observed over Europe while allowing the remaining, less intense portions of the same continuous precipitation system to extend beyond the 99.9th percentile threshold.

An illustration of the detection and association procedure is provided in Figure 4.2.

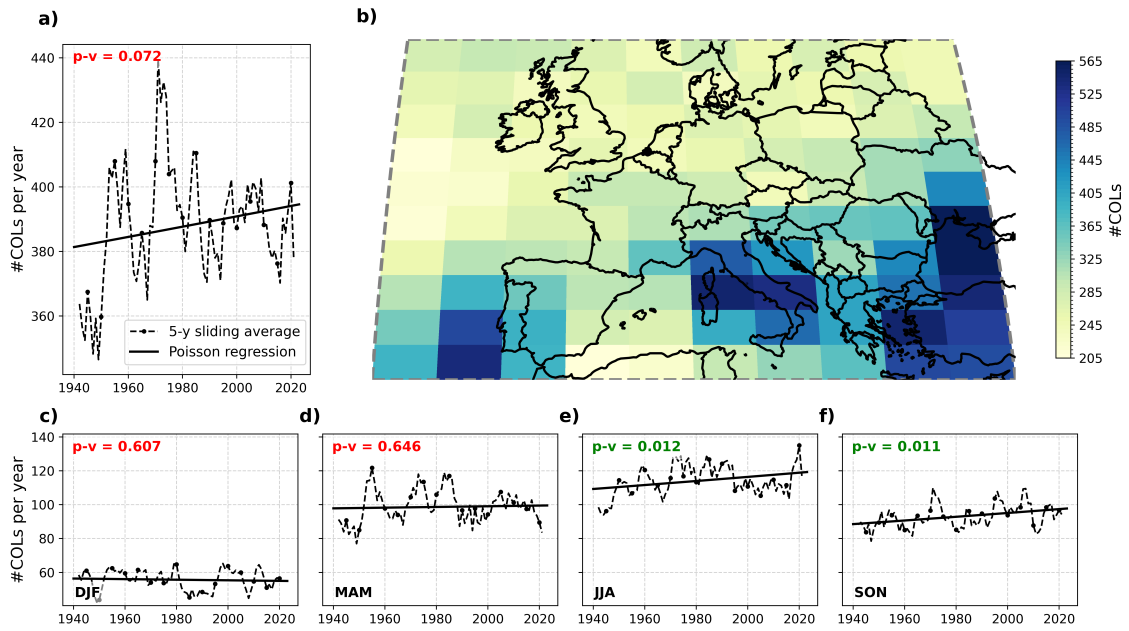


Figure 4.3: (a) Annual number of detected COLs over the European domain for the period 1940-2024. The dashed line shows the 5-year running mean and the solid line indicates the Poisson regression fit; the associated p -value is reported (green: statistically significant at the 95% confidence level; red: non-significant). (b) Spatial distribution of COL centres, expressed as the cumulative number of detected COLs over the same period. (c–f) Seasonal evolution of the annual number of COLs for winter (DJF), spring (MAM), summer (JJA), and autumn (SON). Dashed lines represent 5-year running means, solid lines correspond to Poisson regression fits, and reported p -values indicate the statistical significance of the temporal trends.

4.2.2 Historical COL counting over Europe

In a first step, we estimate the annual number of detected COLs using ERA5 reanalysis z_{500} data over the period 1940-2024 (Fig. 4.3a). On a yearly basis, the number of detected COLs can exceed 400 events per year. Elevated frequencies are particularly observed in the 1970s, where an isolated peak is apparent. To characterize the temporal evolution of COL occurrence, a Poisson regression is fitted to the time series, modelling the evolution of the mean annual count γ .

Overall, γ remains relatively stable around approximately 360 COLs per year, with a p -value of 0.266, indicating that no statistically significant trend is detected in the annual counts (Fig. 4.3a). Although this average count may appear high, it reflects the counting methodology: individual COLs may be counted multiple times if they persist over several days. This approach was adopted to avoid introducing additional methodological complexity related to spatial or temporal declustering. Nevertheless, applying a spatio-temporal declustering procedure does not alter the statistical significance of the observed trends; it only reduces the annual number of detected COLs (see Supplementary Fig. S1 in Appendix D).

From a spatial perspective, three main COL occurrence hotspots are identified: southwest of Portugal, over Corsica and Sardinia, and in a region extending from the western Black Sea to the Ionian Sea. In these areas, more than 500 COLs are detected over the study period (Fig. 4.3b).

From a seasonal perspective, significant trends are identified for summer (JJA) and

4. Evolution of cut-off low frequency

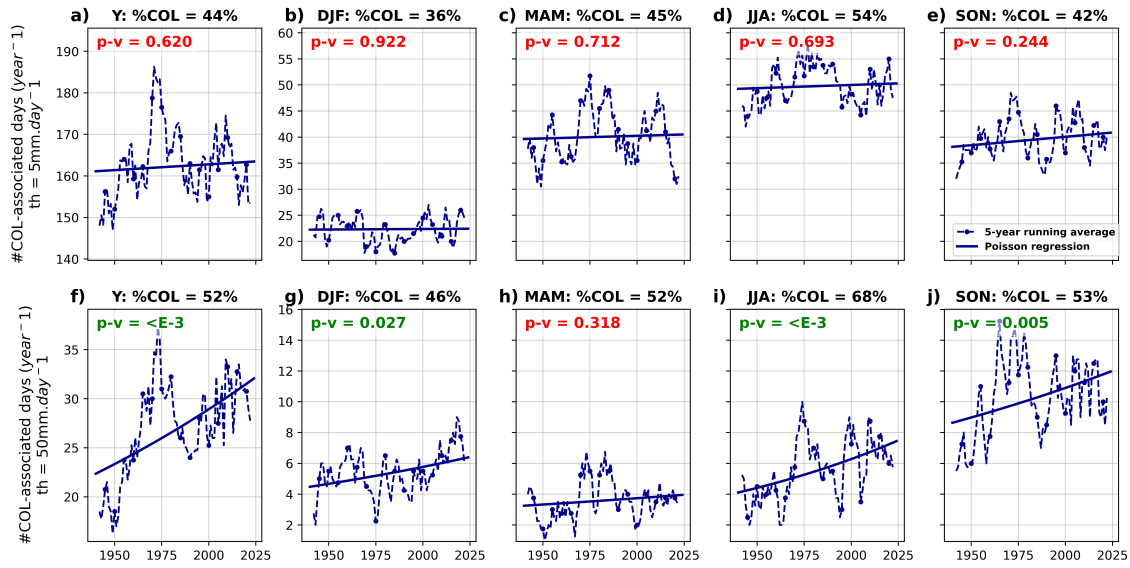


Figure 4.4: Annual and seasonal evolution of the number of extreme precipitation days associated with cut-off lows (COLs) in ERA5. Panels (a-e) show the number of days per year exceeding the spatial 99.9th percentile of daily precipitation and reaching at least 5 mm day^{-1} , associated with COLs, for the whole year (a), DJF (b), MAM (c), JJA (d), and SON (e). Panels (f-j) show the same diagnostics using a stricter precipitation threshold of 50 mm day^{-1} . Dashed lines represent 5-year running means, while solid lines correspond to Poisson regression fits. Reported p-values indicate the statistical significance of the temporal trends (green: statistically significant at the 95% confidence level; red: non-significant)

autumn (SON). In summer, γ increases to approximately 110 COLs per year by the end of the period (Fig. 4.3e), while in autumn it reaches approximately 90 COLs per year. No significant trends are detected for the other seasons.

These results suggest that, while no significant long-term trend is detected at the annual scale, seasonal changes may emerge, particularly during the warm season.

4.2.3 Association of past extreme precipitation events with COLs

In a second step, we associate each precipitation event in Europe with the presence of detected COLs. To analyse the evolution of COL-related precipitation, two thresholds are considered: a lower threshold of 5 mm day^{-1} and a higher threshold of 50 mm day^{-1} .

For each day, we identify extreme precipitation events based on the spatial 99.9th percentile of daily precipitation. We then count the number of days for which this percentile exceeds the selected thresholds and is associated with at least one detected COL.

Using the lower threshold of 5 mm day^{-1} , the number of COL-associated precipitation days fluctuates around a mean value γ of approximately 160 days per year (Fig. 4.4a). As in the total COL counts, the 1970s appear as a period of enhanced activity, with an isolated peak in the time series. However, no significant long-term trend is detected in γ , and the variability is primarily dominated by decadal fluctuations. From a seasonal perspective, no statistically significant trends are identified for any season (Figs. 4.4b-e).

Using the higher threshold of 50 mm day^{-1} , a significant increase in the number of COL-related precipitation days is detected, with γ reaching approximately 30 events per

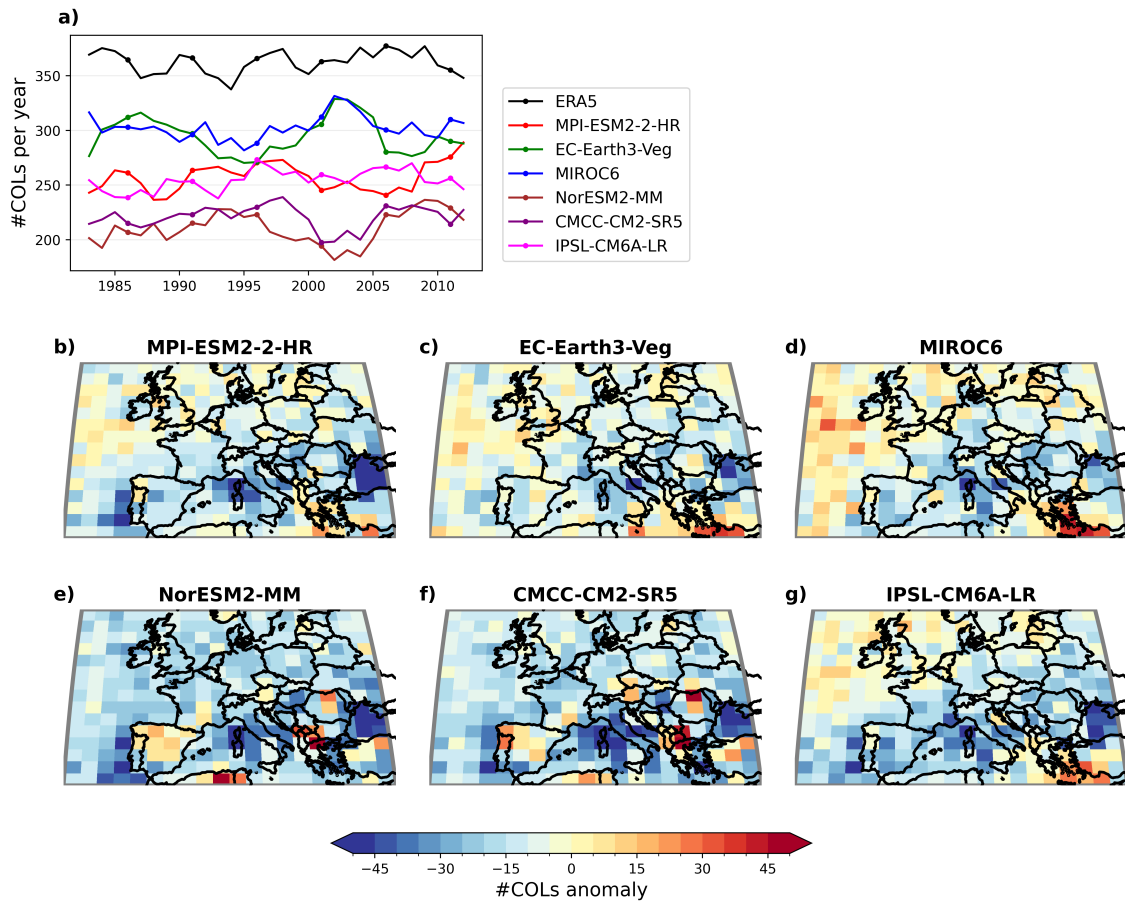


Figure 4.5: Evaluation of cut-off low (COL) yearly counts in CMIP6 ESMs relative to ERA5 over the historical period (1981–2014). (a) Annual number of detected COLs in ERA5 and in the selected ESMs (see Table 2.1), with a 20-year running mean applied. (b–g) Spatial anomalies in the cumulative number of detected COL centres in each ESM relative to ERA5 over the same period.

year by the end of the period (Fig. 4.4f). Significant trends are also identified for winter (DJF), summer (JJA) and autumn (SON) (Figs. 4.4g, i and j).

We interpret this dependence on the precipitation threshold as an indication that changes in COL-related extreme precipitation are primarily driven by thermodynamic or other dynamical processes rather than large-scale dynamical changes in COL frequency.

4.2.4 Modelled COL evolution by ESMs

To assess future trends in COL-related precipitation, the detection and association algorithm was applied to the six CMIP6 ESMs used in this study (see Table 2.1).

However, the ability of the ESMs to represent COL occurrences over Europe varies substantially, even before any precipitation association is performed. Using the historical scenario (1981–2014), all six ESMs underestimate the annual number of detected COLs compared to ERA5 (Fig. 4.5a). EC-Earth3-Veg and MIROC6 exhibit the highest COL counts among the models, with values fluctuating around 300 events per year.

These differences are also reflected spatially. Models such as CMCC-CM2-SR5 and NorESM2-MM, which show the lowest annual counts, strongly underestimate COL occurrences in the previously identified hotspot regions (Figs. 4.5e and f).

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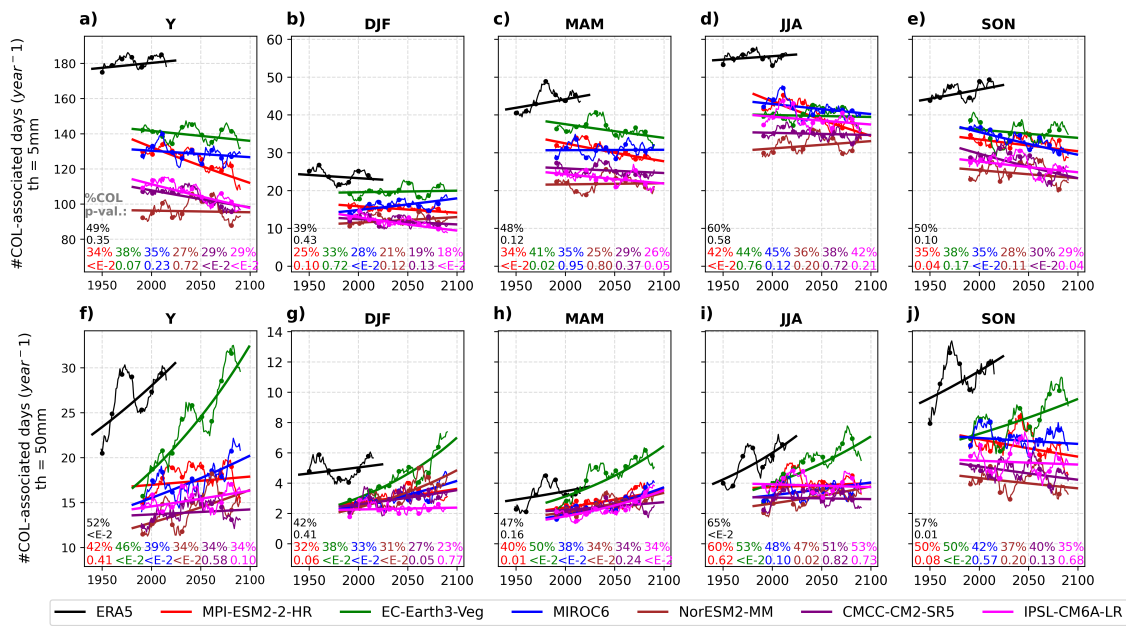


Figure 4.6: Evolution of the annual and seasonal number of extreme precipitation days associated with cut-off lows (COLs) in ERA5 and CMIP6 ESMs. Panels (a–e) show the annual number of days exceeding the spatial 99.9th percentile of daily precipitation and reaching at least 5 mm day^{-1} , associated with COLs, for the whole year (a), DJF (b), MAM (c), JJA (d), and SON (e). Panels (f–j) show the same diagnostics using a stricter precipitation threshold of 50 mm day^{-1} . Black curves correspond to ERA5 over the historical period, while coloured curves show CMIP6 ESM projections under the SSP5-8.5 scenario. Dashed curves represent 20-year running means, while solid lines indicate Poisson regression fits used to assess temporal trends. Associated p -values and percentage of COL-associated days (%COL) are displayed at the bottom of each panel in the colour corresponding to the respective curve.

Taking these limitations into account, future precipitation events are associated with detected COLs using bias-adjusted ESM data and the same precipitation thresholds defined previously, under the SSP5-8.5 scenario.

Using the lower threshold of 5 mm day^{-1} , the simulated annual number of COL-related events remains below the levels observed in ERA5 (Fig. 4.6a). EC-Earth3-Veg again exhibits the highest counts, fluctuating around 130 events per year, while NorESM2-MM shows the lowest values, around 100 events per year. This overall underestimation is consistent across seasons (Figs. 4.6b-e). No significant trends are detected in most models, with time series dominated by strong interdecadal variability and alternating periods of high and low COL activity.

Using the higher threshold, ERA5 levels are only reached by EC-Earth3-Veg by the end of the century, with the mean annual count γ approaching 30 events per year and showing a significant positive trend (Fig. 4.6f). Among all models, EC-Earth3-Veg exhibits the closest behaviour to ERA5 in terms of temporal evolution, with increasing trends in MAM, JJA, and SON (Figs. 4.6h-j). In MAM, it even exceeds historical ERA5 values despite the absence of a significant trend in observations.

Overall, strong interdecadal variability is observed across all models, indicating that the end of the century is not necessarily the most active period in terms of COL-related events. These results highlight the dominant role of internal climate variability in controlling the occurrence of high-impact COL-related precipitation events. They also suggest that, despite the increased sensitivity of CMIP6 models to extreme precipitation (see Fig. 1.4), simulated COL-related impacts remain delayed compared to historical observations.

4.2.5 Discussion summary

Several aspects of these results are discussed in detail in the associated paper (Appendix C). Here, we briefly summarize the main limitations of the detection and association algorithm, and discuss the implications of the results in the context of a warming climate.

4.2.5.1 Algorithm limitations and comparison with previous studies

The main advantage of our algorithm lies in its simple conceptual complexity, which makes it easily adaptable to various datasets. Nevertheless, it still relies on some arbitrary choices.

For instance, the A_{crit} parameter, which represents the minimum area that an enclosed region must reach to be classified as a COL, is set to $50,000 \text{ km}^2$ in this study. This value was chosen as a compromise between excluding small, short-lived systems associated with transient troughs and retaining sufficiently large, well-defined COL structures. Increasing A_{crit} systematically decreases the number of detected COLs and reduces the statistical significance of trends over the period 1940-2024 (see supplementary Fig. D.3 in Appendix D).

Based on these geometrical criteria, we identify an average of approximately 360 COLs per year over Europe, with temporal variability mainly controlled by interdecadal fluctuations. Using a different detection methodology, Muñoz et al. (2020) identified approximately 50 COLs per year over a different European domain and reported a statistically significant increasing trend over the period 1960-2017. When restricting

our analysis to the same period, no significant trend is detected, with variability again dominated by pronounced interdecadal fluctuations (see Appendix C, Fig. D.4). Similarly, Mishra et al. (2025) identified approximately 50 COLs per year over Europe using the TRACK algorithm.

These discrepancies are likely explained by methodological differences. First, Muñoz et al. (2020) used the NCEP/NCAR reanalysis, whereas the present study relies on ERA5. Their detection procedure also requires the presence of an eastward-propagating thermal front for a feature to be classified as a COL. In contrast, our algorithm is considerably more permissive, as it does not exclude systems based on the relative position of the associated front. Moreover, the precipitation front associated with a COL is not necessarily located on its eastern flank (Fig. 4.1d). Finally, we count COLs on a daily basis, whereas both Muñoz et al. (2020) and Mishra et al. (2025) apply a spatio-temporal declustering procedure, resulting in each COL being counted only once. Consequently, the larger number of detected COLs reported here should not be interpreted as evidence of increased COL activity, but rather as a direct consequence of the adopted counting strategy and detection criteria.

Despite these methodological differences, we consider the main conclusions to be robust. Indeed, the manual evaluation of randomly selected cases shows that the algorithm successfully identifies the targeted features (see the Zenodo repository at <https://doi.org/10.5281/zenodo.19604862> and the evaluation presented in Appendix C). Moreover, the spatial distribution of COL centres over Europe closely corresponds to that reported by Nieto et al. (2005), with three nearly identical Mediterranean hot-spots clearly identified.

Concerning the precipitation association procedure, the choice was made to consider continuous precipitation pixel surfaces containing the most extreme precipitation pixels, i.e., pixels exceeding the spatial 99.9th percentile of daily precipitation. The values of the connected pixels are required to exceed 1 mm day^{-1} (see Figs. 4.2a and b). This methodological criterion is intentionally permissive in order to associate as many extreme precipitation events as possible with potentially present COLs. However, a stricter criterion could be adopted by requiring the most extreme precipitation pixels to be more spatially concentrated around the COL centre. To assess whether this methodological choice affects the main conclusions of the study, a sensitivity analysis is performed using alternative definitions of the most extreme precipitation fronts.

In this sensitivity analysis, the pixels forming the most extreme precipitation fronts are required to exceed a given spatial precipitation percentile, denoted q_{rain} . By progressively increasing this threshold, increasingly restrictive definitions of the precipitation fronts are obtained. The statistical significance of the temporal trends exhibits the same threshold dependence as described previously, with no qualitative change in the conclusions. Increasingly strict definitions of the precipitation fronts result in a temporally consistent downward shift in the annual number of COL-associated extreme precipitation events (Fig. 4.7).

For a precipitation threshold of 50 mm day^{-1} , the original definition associates up to 52% of extreme precipitation events with COLs (Fig. 4.7f). Using the strictest definition tested, based on the 95th spatial precipitation percentile ($q_{\text{rain}} = 95$), this proportion decreases to 33% (Fig. 4.7j). Thus, although the absolute number of COL-associated events decreases when applying stricter association criteria, COLs still account for a substantial fraction of the most extreme precipitation events.

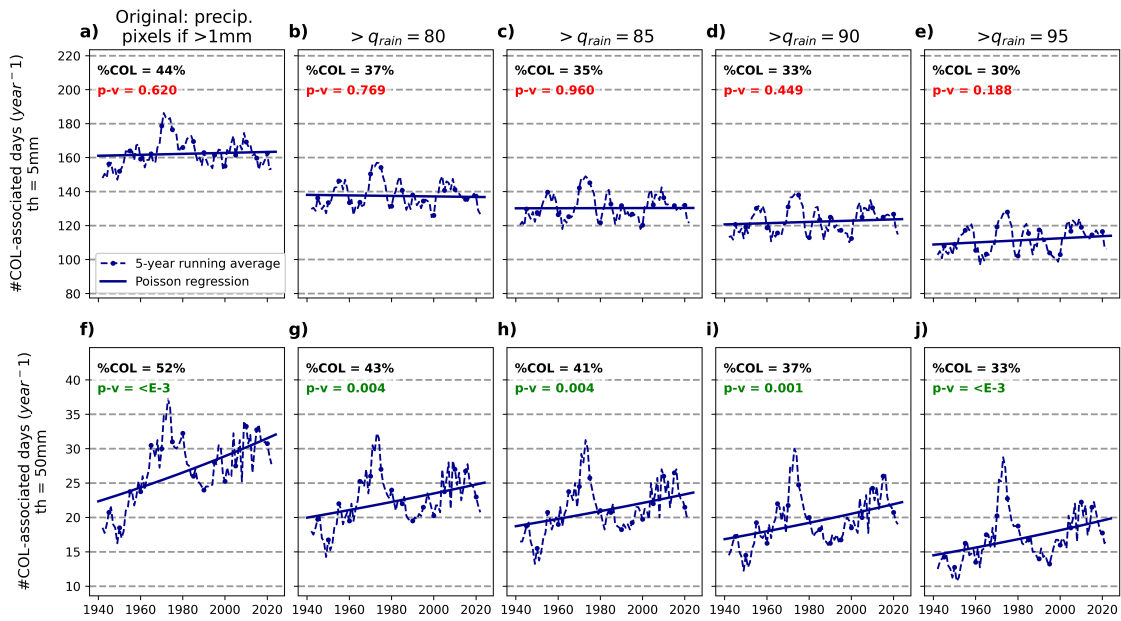


Figure 4.7: Evolution of the annual number of extreme precipitation days associated with cut-off lows (COLs) over 1940-2024 using different definitions of the most extreme precipitation fronts. Panels (a-e) show the annual number of days exceeding the spatial 99.9th percentile of daily precipitation and reaching at least 5 mm day⁻¹. Panels (f-j) show the same diagnostics using a stricter precipitation threshold of 50 mm day⁻¹. Dashed lines represent 5-year running means, while solid lines correspond to Poisson regression fits. Reported p-values indicate the statistical significance of the temporal trends (green: statistically significant at the 95% confidence level; red: non-significant). The corresponding figures for the four seasons are provided in Appendix E (Figs. E.1- E.4).

Taken together, these sensitivity analyses indicate that the main conclusions regarding the temporal evolution of COL-associated extreme precipitation are robust to the methodological choices tested.

4.2.5.2 Toward a Better Understanding of Physical Drivers Behind COL Impacts

Our results indicate that significant increasing trends in COL-related precipitation days in Europe emerge only when high precipitation thresholds are used to define extreme events. This raises the question of why COL-related extreme events intensify under warming conditions.

We argue that this behaviour results from the interplay between multiple aggravating factors and feedbacks, including land-atmosphere interactions. For instance, Insua-Costa et al. (2022) showed that approximately 50% of the daily precipitation during the 13 July 2021 event originated from local moisture recycling. Using the WRF model coupled with a Lagrangian moisture tracking framework, they demonstrated that a substantial fraction of the moisture feeding the precipitation system originated from evapotranspiration over central European forests. In the discussion of the associated paper, we further show that several European regions exhibit significant seasonal trends in evapotranspiration, which may enhance the moisture available for COL-related precipitation. Insua-Costa et al. (2022) also suggest that such feedbacks could increasingly modulate the impact of COLs in a changing climate.

Moreover, Ludwig et al. (2023) showed that the July 2021 COL event would have produced approximately 9% more cumulative precipitation over a three-day period per

additional degree of global warming. Using a pseudo-global-warming framework, they highlight how thermodynamic scaling of precipitation with temperature can amplify the impact of such events.

More recently, Calvo-Sancho et al. (2026) applied a similar approach to the Dana COL event, which affected Valencia in October 2024 (see Fig. 4.1e and f), and found that it produced approximately 20% more heavy precipitation compared to pre-industrial conditions. This corresponds to an increase of about 55% in the spatial extent of the precipitation event. Their analysis attributes this intensification to enhanced moisture fluxes from the Mediterranean Sea (approximately +9%) and increased updraft velocities within convective systems.

In this context, COLs may primarily act as synoptic-scale facilitators, while the severity of extreme precipitation events is likely controlled by a combination of thermodynamic conditions, local dynamics, and land-atmosphere feedbacks. These interactions remain insufficiently constrained and represent an important avenue for future research aimed at improving our understanding and prediction of extreme precipitation events in Europe.

4.3 Focus on Belgium

Although Belgium is not among the European regions with the highest COL frequencies (see Fig. 4.3b), it was recently affected by the July 2021 cut-off low, which led to record-breaking rainfall (Kreienkamp et al., 2021). Therefore, we assess the evolution of COL-related precipitation in Belgium using the same framework as applied at the European scale.

4.3.1 Historical evolution of COL-related precipitation events in Belgium

Northwestern Europe exhibits a lower frequency of detected COLs compared to previously identified hotspot regions (see Fig. 4.3b). As a result, Belgium experiences on average 25 COL-related precipitation days per year (Fig. 4.8a).

Using a 1 mm day⁻¹ threshold, the spatial 99th percentile of daily precipitation in bias-adjusted MAR-ERA5 outputs exceeds this value approximately 25 times per year on average, with no significant temporal trend. The corresponding time series is dominated by pronounced interdecadal variability, with periods of enhanced COL activity, such as in the late 1970s.

From a seasonal perspective, summer (JJA) is the most active season, with around 8 events per year on average, but no significant temporal trend is detected. No other season exhibits a significant trend either.

When increasing the precipitation threshold from 1 to 30 mm day⁻¹, approximately 14% of the selected extreme precipitation events are associated with COLs (Fig. 4.9a). Interestingly, among the historical COL-associated events, the 13, 14 and 15 July appear among the most extreme ever recorded. This threshold is chosen as a compromise between retaining sufficiently extreme precipitation events and obtaining a sufficient number of events per year for the analysis of temporal trends in annual counts, yielding on average approximately two events per year.

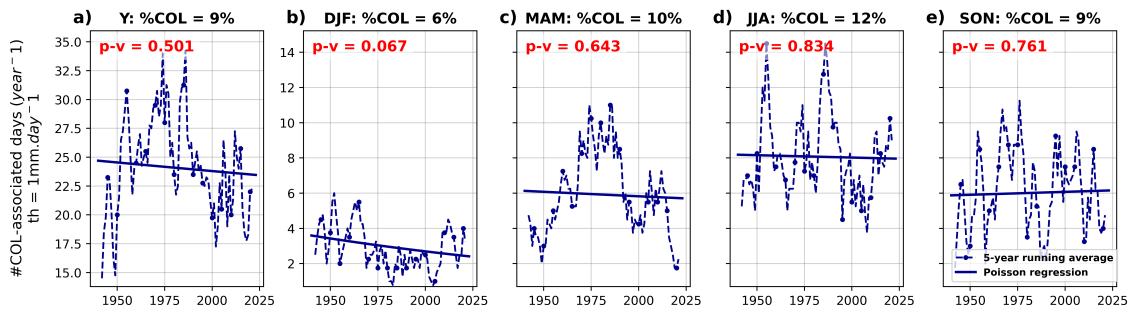


Figure 4.8: (a) Annual number of COL-associated precipitation days for which the spatial 99th percentile exceeds 1 mm per day. Panels (b), (c), (d) and (e) show the corresponding counts for DJF, MAM, JJA, and SON, respectively. Reported p-values indicate the statistical significance of the temporal trends (green: statistically significant at the 95% confidence level; red: non-significant)

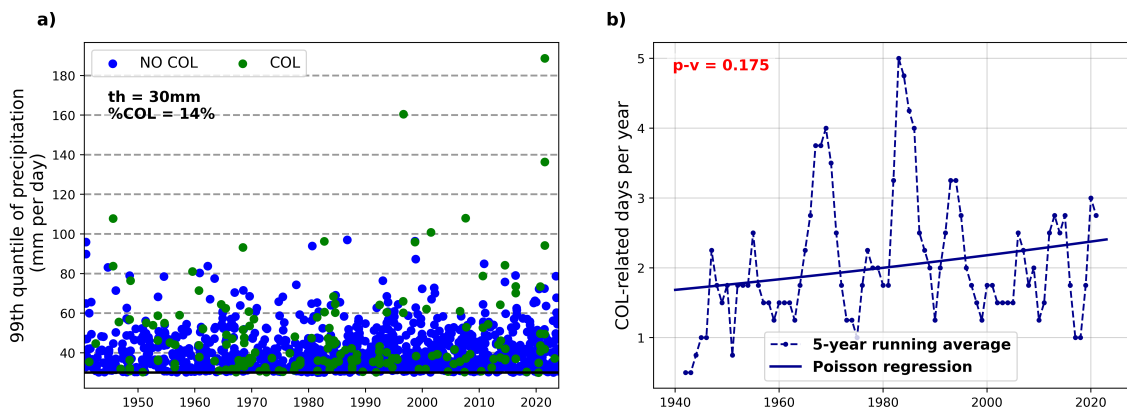


Figure 4.9: (a) Precipitation days where the 99th spatial percentile is higher than 30 mm per day and (b) associated annual evolution of the number of precipitation days associated with cut-off lows (COLs) in Belgium using MAR-ERA5 bias-adjusted precipitation. Reported p-values indicate the statistical significance of the temporal trends (green: statistically significant at the 95% confidence level; red: non-significant)

The Poisson parameter γ increases from approximately 1.7 to 2.4 COL-related events per year over the period 1940-2023, although this trend remains statistically non-significant (Fig. 4.9b). At the Belgian scale, the early 1980s stand out as a particularly active period, with annual frequencies exceeding 4 COL-related events per year.

Nevertheless, 7 out of the 10 most intense precipitation events over the period 1940-2023 are associated with COLs, with the most extreme event corresponding to the July 2021 floods (Fig. 4.9a).

The synoptic configurations associated with the six most impactful COLs, excluding July 2021, are shown in Fig. 4.10a-f. All these events occurred between July and October. As illustrated, these COLs are not all at the same stage of development and are not equally detached from the main westerly flow. For instance, the COL of 7 July 2007 (Fig. 4.10e) appears less isolated than the one observed on 29 August 1996.

In terms of precipitation patterns, the Ardennes topography plays a key role in enhancing precipitation for several events, as observed for 7 October 1982 and 9 August 2007 (Figs. 4.10h and i), similarly to the July 2021 event. However, COL-related precipitation is not systematically confined to high-relief regions. For example, on 29 August 1996, central Belgium and Flanders experienced daily precipitation totals exceeding 150 mm at some locations (Fig. 4.10i). In other cases, such as August 1985

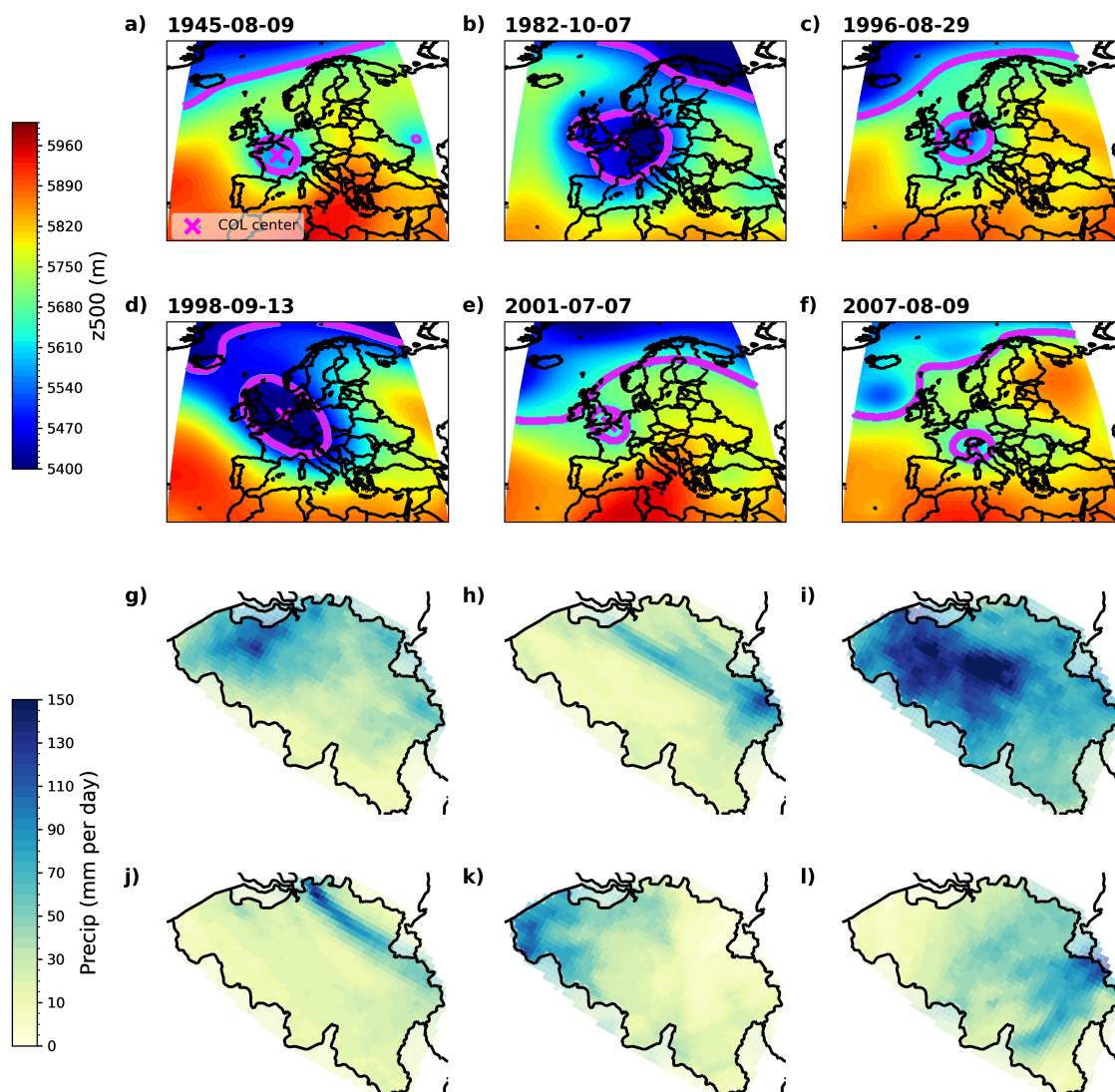


Figure 4.10: (a-f) 500 hPa geopotential height fields associated with the six most impactful cut-off lows (COLs) identified in Belgium over the period 1940-2023 (excluding July 2021). (g-l) Corresponding daily precipitation fields from bias-adjusted MAR-ERA5 outputs.

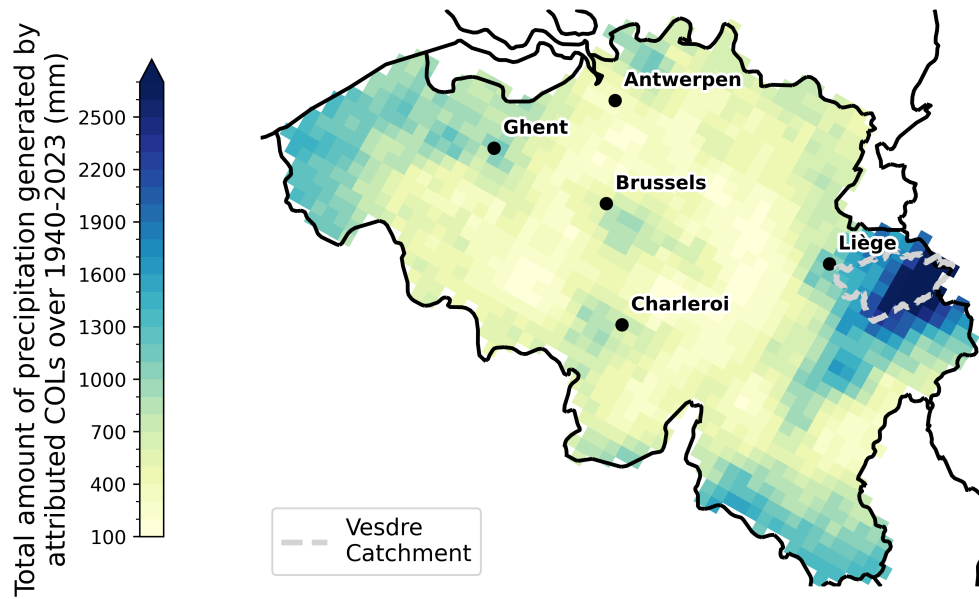


Figure 4.11: Total amount of precipitation for COL-related days identified over 1940-2023.

and July 2021 (Figs. 4.10g and k), the coastal regions also recorded daily precipitation amounts above 130 mm.

This topography enhancement role clearly appears when we plot the total amount of rainfall associated with COLs over 1940-2023 (Fig. 4.11). Over the 1940–2023 period, the region receiving the largest cumulative amount of COL-related precipitation is located around the Vesdre catchment, where totals exceed 2700 mm. This catchment was among the most severely affected areas during the July 2021 floods (Dewals et al., 2021). The cyclonic rotation associated with COLs leads to moisture fluxes that are mainly oriented from north to south. When the precipitation front crosses the relatively elevated Ardennes topography, convection is enhanced, resulting in higher precipitation amounts.

Although less pronounced, the southern Ardennes relief can also enhance precipitation when crossed by precipitation systems. As a result, this region exhibits cumulative COL-related precipitation totals close to 1500 mm over the study period (Fig. 4.11).

Beyond these areas of steep topography, other regions of enhanced COL-related precipitation can be identified. In particular, localized maxima appear in the vicinity of major urban areas such as Brussels, Antwerp, Ghent, and Charleroi. In addition, the Belgian coastal region stands out, with cumulative totals reaching approximately 1300 mm.

This spatial distribution suggests that, in addition to topographic effects, other local or regional factors may contribute to amplifying COL impacts, including proximity to moisture sources such as the North Sea and potentially urban or other land-surface effects.

4.3.2 Projected evolution of COL-related precipitation events in Belgium

The projected evolution of COL-associated precipitation days over Belgium leads to conclusions similar to those obtained at the European scale. For low precipitation

4. Evolution of cut-off low frequency

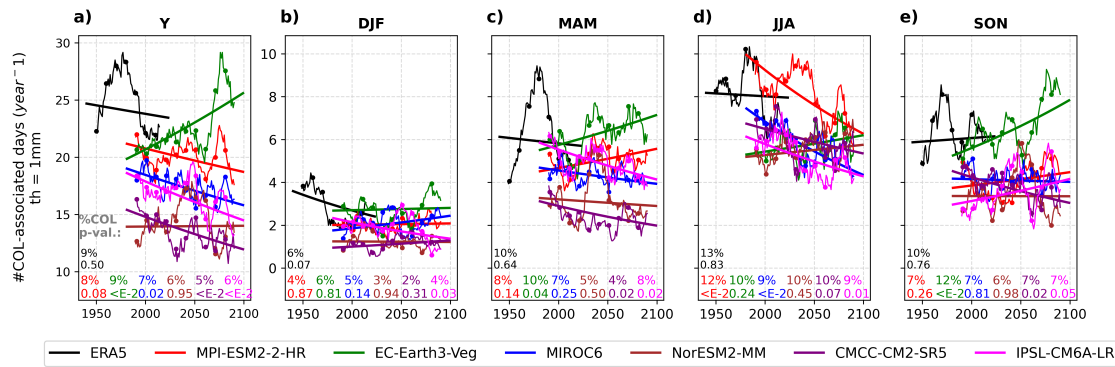


Figure 4.12: Evolution of the annual and seasonal number of extreme precipitation days associated with cut-off lows (COLs) in ERA5 and CMIP6 ESMs under the SSP5-8.5 scenario for Belgium. Panels (a–e) show the annual number of days exceeding the spatial 99th percentile of daily precipitation and reaching at least 1 mm day⁻¹, associated with COLs, for the whole year (a), DJF (b), MAM (c), JJA (d), and SON (e). Black curves correspond to ERA5 over the historical period, while coloured curves show CMIP6 ESM projections under the SSP5-8.5 scenario. Dotted curves represent 20-year running means, while solid lines indicate Poisson regression fits used to assess temporal trends. Associated *p*-values and percentages of COL-associated days (%COL) are displayed at the bottom of each panel in the colour corresponding to the respective curve.

thresholds (1 mm day⁻¹), only EC-Earth3-Veg reproduces frequencies comparable to those derived from ERA5 over the historical period (Fig. 4.12a).

By the end of the century, for EC-Earth3-Veg, the average number of COL-related precipitation days significantly increases, reaching approximately 25 days per year under SSP5-8.5 (Fig. 4.12a). However, similar levels were already observed during the 1970s in ERA5, highlighting the strong influence of internal climate variability on COL occurrence.

A pronounced interdecadal variability is found for all models. On average, most simulations still underestimate the observed number of COL-related precipitation days over Belgium. Among the remaining models, MPI-ESM1-2-HR produces results closest to EC-Earth3-Veg, although substantial year-to-year fluctuations remain.

From a seasonal perspective, similar conclusions are obtained for MAM and SON (Fig. 4.12c and e). However, during JJA, MPI-ESM1-2-HR is the only model whose simulated frequencies approach those derived from ERA5, with seasonal counts oscillating around 8 COL-associated extreme precipitation days per year. The other models systematically underestimate the observed frequencies, although all simulations exhibit pronounced interannual and interdecadal variability.

For more extreme events, EC-Earth3-Veg and MPI-ESM1-2-HR both reproduce frequencies comparable to those derived from ERA5. In both simulations, the annual number of COL-associated extreme precipitation days increases throughout the 21st century, reaching more than 3 events per year by the end of the century under SSP5-8.5. This increase is statistically significant for EC-Earth3-Veg (Fig. 4.13b).

Moreover, both models simulate proportions of COL-related extreme precipitation events close to those observed in ERA5, with approximately 12% of extreme events associated with COLs compared to 13% in the reanalysis.

The influence of internal climate variability is further highlighted by the strong differences between models, with some simulations even exhibiting negative trends, as illustrated by CMCC-CM2-SR5. In this configuration, the late 21st century does not correspond to a particularly active period in terms of COL-related extreme precipitation.

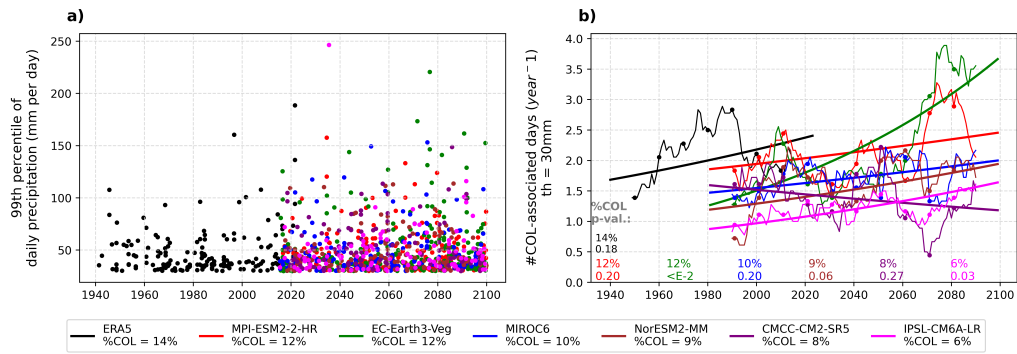


Figure 4.13: (a) Spatial distribution of precipitation days for which the spatial 99th percentile of daily precipitation exceeds 30 mm day^{-1} in Belgium, and (b) associated annual evolution of the number of extreme precipitation days associated with cut-off lows (COLs) using ERA5 and the studied CMIP6 ESMs under the SSP5-8.5 scenario. Associated p -values and percentage of COL-associated days (%COL) are displayed at the bottom of each panel in the colour corresponding to the respective curve.

The non-linearity and stochastic nature of extreme COL-related precipitation are particularly well illustrated when considering the absolute intensity of events exceeding the 30 mm day^{-1} threshold (Fig. 4.13a). For example, although IPSL-CM6A-LR underestimates the annual occurrence of COL-associated events, the corresponding down-scaled and bias-adjusted MAR-IPSL simulation produces an event reaching approximately 250 mm day^{-1} during the late 2030s over Belgium.

This result illustrates that the timing and intensity of the most impactful COL-related precipitation events will likely remain strongly stochastic under future climate conditions.

EC-Earth3-Veg is particularly noteworthy, as it is the only simulation producing several COL-related events with precipitation intensities comparable to those observed during the 14 July 2021 and 28 August 1996 events, which correspond to the most impactful observed COL-related precipitation extremes over Belgium.

4.4 Implications of the results

In this chapter, a new detection algorithm was introduced to identify cut-off lows (COLs) and associate extreme precipitation events with their potential presence. The flexibility of the algorithm was demonstrated through its application to both ERA5 reanalyses and CMIP6 Earth System Model outputs.

At the European scale, the analysis shows that the number of COL-related precipitation days associated with relatively low precipitation thresholds does not exhibit significant long-term trends. Instead, the occurrence of these events is largely controlled by natural climate variability, characterized by pronounced interdecadal fluctuations. The active COL period observed during the 1970s constitutes one example of such variability.

However, when focusing on the most extreme precipitation events, a substantially different picture emerges. Increasing the precipitation threshold reveals significant positive trends in both annual and seasonal COL-related event frequencies, particularly during JJA and SON. Although internal climate variability continues to modulate year-to-year variations, these results suggest that the recent increase in COL-related extreme precipitation is primarily associated with thermodynamic intensification and other local or dynamical processes, rather than with an increase in the absolute number of COLs.

Extending the analysis to CMIP6 ESM outputs further highlights the difficulty of reproducing realistic COL frequencies. Among the considered models, EC-Earth3-Veg is the only simulation producing frequencies approaching those derived from ERA5 under the SSP5-8.5 scenario. This result is particularly noteworthy given the strong warming associated with this scenario.

At the Belgian scale, similar conclusions are obtained. Significant long-term trends in COL-related precipitation frequencies remain limited, even for higher precipitation thresholds. The occurrence of the most extreme COL-related precipitation events remains strongly stochastic, with large differences between models and periods.

Nevertheless, this chapter also demonstrates the important role of geography and topography in modulating precipitation intensities associated with COLs. Elevated regions, particularly over the Ardennian Massif, appear especially prone to extreme COL-related precipitation accumulation.

Furthermore, the fact that EC-Earth3-Veg is the only model capable of reproducing both realistic COL frequencies and precipitation intensities comparable to the most severe observed Belgian events makes this simulation particularly compelling. Consistently, this model also produces some of the highest projected future return levels over Belgium (see Fig. 3.6).

Take-home messages

- A new flexible detection algorithm was developed to identify cut-off lows (COLs) and associate precipitation extremes with their occurrence in both reanalyses and CMIP6 model outputs.
- At the European scale, low-intensity COL-related precipitation events are mainly controlled by natural climate variability and do not exhibit robust long-term trends.
- For the most extreme precipitation events, significant increases in COL-related event frequencies emerge, particularly during summer and autumn, suggesting that these changes are primarily associated with thermodynamic and local dynamical processes rather than with changes in the absolute number of COLs.
- CMIP6 Earth System Models strongly differ in their ability to reproduce realistic COL frequencies and associated precipitation intensities, highlighting substantial inter-model uncertainty.
- EC-Earth3-Veg is the only model reproducing both realistic COL frequencies and precipitation intensities comparable to the most severe observed Belgian events.
- At the Belgian scale, the occurrence of the most extreme COL-related precipitation events remains highly stochastic and strongly modulated by internal climate variability.
- Topography plays a major role in amplifying COL-related precipitation extremes, particularly over the Ardennian Massif.

CHAPTER 5

Hydrological response to climate change in the Ourthe basin

In Chapters 3 and 4, we showed how extreme precipitation intensity and frequency are projected to evolve according to MAR outputs, and to what extent part of this non-linear response may be attributed to changes in cut-off low (COL) occurrence. In this chapter, we investigate how the Ourthe watershed may respond to these projected climatic changes.

To address this question, Section 5.1 first evaluates how well the four hydrological models used in this study (GR4H, HBV, VHM and SAC-SMA) reproduce observed discharges over the calibration period 1991-2010 when forced by bias-adjusted MAR-ERA5 outputs. Section 5.2 then assesses the ability of the calibrated models to reproduce observed discharges over the independent validation period 2011-2023.

Finally, Section 5.3 compares the responses of the different hydrological models to projected MAR climate forcings.

5.1 Hydrological model calibration over 1991-2010

At the Tabreux station, all four models reach NSE values close to 0.7 (Fig. 5.1a-d). However, GR4H appears to reproduce the most extreme discharge peaks more accurately. For instance, during January 1993, the peak discharge at Tabreux reached $433 \text{ m}^3 \text{ s}^{-1}$. GR4H underestimates this hourly peak by $40 \text{ m}^3 \text{ s}^{-1}$ (Fig. 5.1a), whereas HBV, VHM and SAC-SMA underestimate it by approximately $100 \text{ m}^3 \text{ s}^{-1}$ (Figs. 5.1b-d).

However, when considering all peak discharges associated with the selected events, the mean bias (MB_{peak}) remains relatively small for all models, ranging from -7.8% for GR4H to 0.7% for VHM, the latter showing the lowest absolute mean bias (Table 5.1).

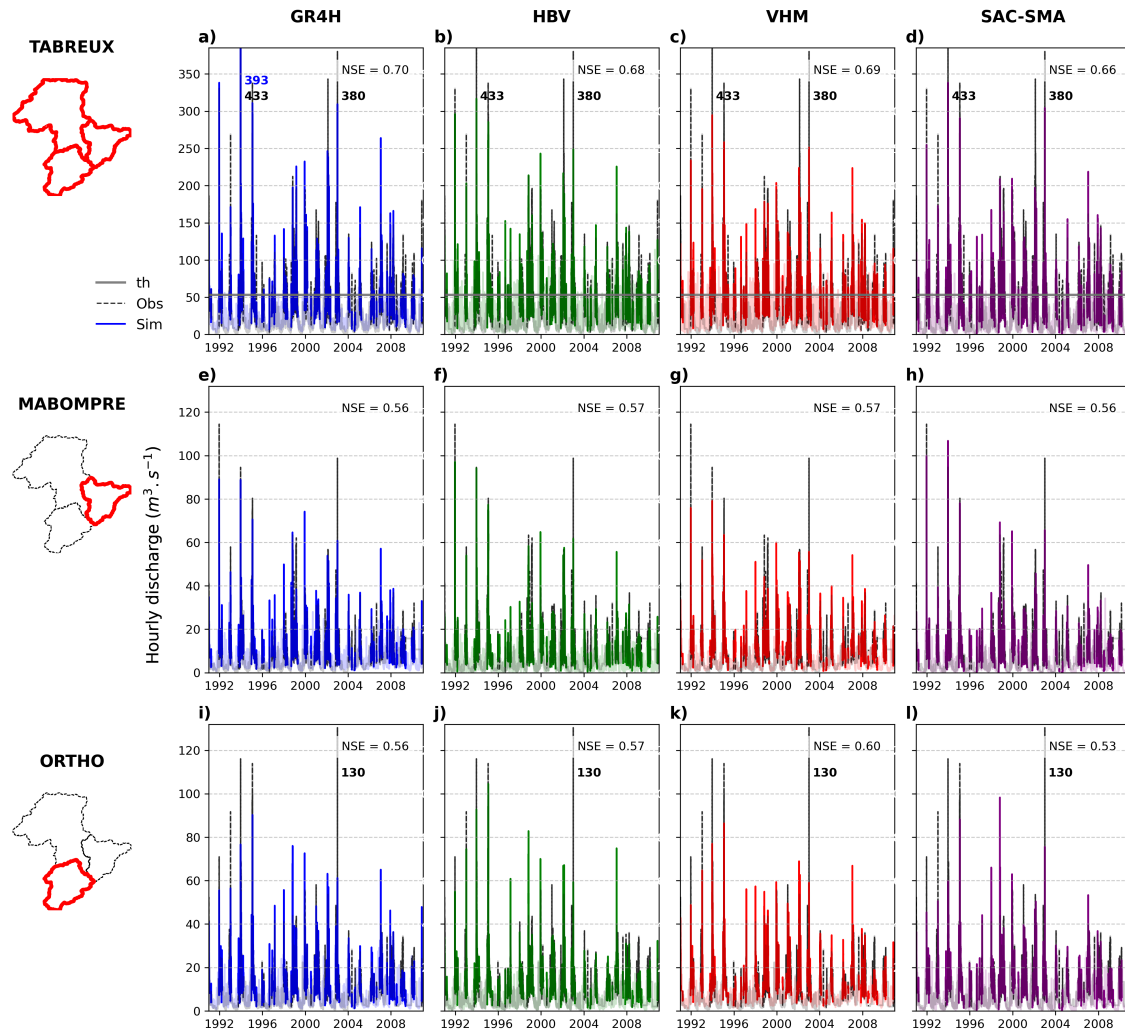


Figure 5.1: Evaluation of hydrological model calibrations over the 1991-2010 period for (a) GR4H, (b) HBV, (c) VHM and (d) SAC-SMA at the Tabreux station. Panels (e-h) show the corresponding results for the Mabompré sub-basin, while panels (i-l) present the results for the Ortho sub-basin. The contributing watersheds associated with the simulated discharges are highlighted on the maps displayed in the first column. For each model, the Nash-Sutcliffe Efficiency (NSE) is indicated. Time steps not selected for the optimization, i.e., those corresponding to events below the threshold, are shown with reduced opacity. Values exceeding the upper limit of the y-axis are indicated numerically.

models	TABREUX			MABOMPRE			ORTHO		
	NSE	MB _{peak}	RMSE _{peak}	NSE	MB _{peak}	RMSE _{peak}	NSE	MB _{peak}	RMSE _{peak}
GR4H	0.70	-7.8	35.2	0.56	6.9	44.5	0.56	3.1	40.9
HBV	0.68	-1.6	30.5	0.57	-6.4	35.4	0.57	-5.9	36.6
VHM	0.69	0.7	30.5	0.57	-1.4	42.4	0.60	-4.4	39.4
SAC-SMA	0.66	-5.3	35.9	0.56	-6.5	38.6	0.53	1.8	40.5

Table 5.1: Summary of hydrological model performance over 1991-2010. NSE = Nash-Sutcliffe efficiency, MB_{peak} = mean bias of event peak discharge (%), and RMSE_{peak} = root mean square error of event peak discharge (%).

These mean biases are substantially lower in absolute value than the corresponding root mean square errors ($RMSE_{peak}$), which exceed 30% for all four models and reach 36% for SAC-SMA (Table 5.1). This indicates that errors across individual peak discharges partly compensate for one another, resulting in relatively low mean biases despite larger errors for individual events.

Overall, considering that the forcing meteorological data are derived from bias-adjusted MAR-ERA5 outputs, all four models appear able to reproduce the interannual variability reasonably well, with satisfactory NSE values.

For the two upstream contributing catchments at the Mabompré and Ortho stations, NSE values decrease to around 0.55 for all models (Figs. 5.1e-l). For all models at both stations, the 2003 peak-flow event, which exceeded $120 \text{ m}^3 \text{ s}^{-1}$ at Ortho, is underestimated by approximately 50%. Here, the $RMSE_{peak}$ is overall higher than at Tabreux and reaches up to 45% (Table 5.1).

We believe that these lower performances over the smaller catchments may reflect the difficulty of calibrating hydrological models over limited drainage areas when the forcing meteorological data are available at a spatial resolution of 5 km. Nevertheless, the interannual variability of peak discharges remains reasonably well reproduced for both contributing sub-basins.

5.2 Model evaluation over the validation period (2011-2023)

To assess whether the hydrological model calibrations remain robust outside the calibration period, the four calibrated models are forced using MAR-ERA5 inputs over the 2011-2023 period (Fig. 5.2). In other words, the objective is to evaluate the ability of the MAR-ERA5-WOLFHydro modelling chain to reproduce the hydroclimatic variability over an independent validation period.

At the Tabreux station, NSE values decrease for all four models, particularly for GR4H and SAC-SMA, whose NSE drops to around 0.25 (Fig. 5.2a), whereas HBV and VHM retain NSE values close to 0.5 (Figs. 5.2b and c). As during the calibration period, some peak flows are underestimated while others are overestimated, such as the July 2021 peak discharge.

For this event, GR4H, HBV and SAC-SMA strongly overestimate the peak discharge, with simulated values reaching 1333, 926 and $1001 \text{ m}^3 \text{ s}^{-1}$, respectively, compared with

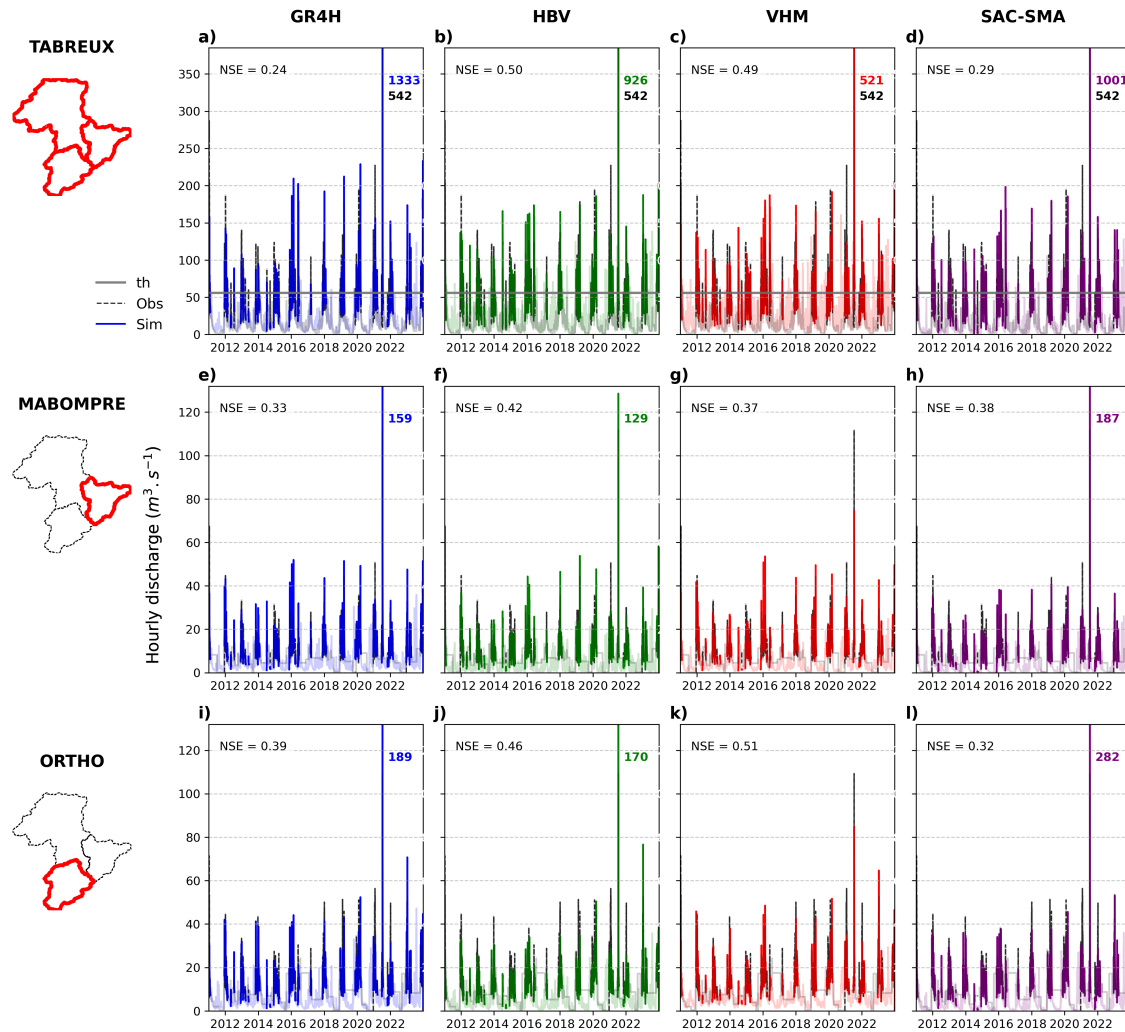


Figure 5.2: Evaluation of hydrological model performance over the independent validation period 2011-2023 for (a) GR4H, (b) HBV, (c) VHM and (d) SAC-SMA at the Tabreux station. Panels (e-h) show the corresponding results for the Mabompré sub-basin, while panels (i-l) present the results for the Ortho sub-basin. The contributing watersheds associated with the simulated discharges are highlighted on the maps displayed in the first column. For each model, the Nash-Sutcliffe Efficiency (NSE) and the mean bias of daily maximum discharge peaks (MB_{24h}) are indicated. Values exceeding the upper limit of the y-axis are indicated numerically.

5. Hydrological response to climate change in the Ourthe basin

models	TABREUX			MABOMPRE			ORTHO		
	NSE	MB _{peak}	RMSE _{peak}	NSE	MB _{peak}	RMSE _{peak}	NSE	MB _{peak}	RMSE _{peak}
GR4H	0.41	-11.9	34.7	0.30	15.0	42.6	0.44	-4.4	33.4
HBV	0.39	-3.8	33.2	0.35	-0.7	32.5	0.48	-16.7	28.4
VHM	0.39	-4.8	31.3	0.29	7.3	38.3	0.47	-10.4	27.2
SAC-SMA	0.41	-12.1	36.2	0.40	-2.6	31.3	0.51	-7.0	29.5

Table 5.2: Summary of hydrological model performance over 2011-2020, excluding the July 2021 event. NSE = Nash-Sutcliffe efficiency, MB_{peak} = mean bias of event peak discharge (%), and RMSE_{peak} = root mean square error of event peak discharge (%).

the observed peak of $542 \text{ m}^3 \text{ s}^{-1}$. In contrast, VHM reproduces this event much more realistically, with a simulated peak discharge of $521 \text{ m}^3 \text{ s}^{-1}$.

At Ortho and Mabompré, similar conclusions are obtained overall. GR4H and SAC-SMA again exhibit the strongest decrease in NSE, together with a substantial overestimation of the July 2021 flood event (Figs. 5.2d-i). For HBV and VHM, NSE values decrease to approximately 0.4 at Mabompré and 0.5 at Ortho.

Once again, the simulation of these smaller sub-basins appears more challenging, likely because of their more localized hydrological response combined with the 5-km spatial resolution of the forcing meteorological data. Nevertheless, the interannual variability and the overall timing of the main discharge peaks remain reasonably well reproduced.

Despite the overall better NSE values obtained for VHM, this does not necessarily imply that it is the most suitable model for simulating the most extreme flood events. The strong overestimations obtained for the July 2021 flood event at Tabreux in GR4H and HBV are not primarily related to poor hydrological calibration, but rather to uncertainties associated with the meteorological forcing data.

The forcing data used in this study consist of bias-adjusted hourly MAR-ERA5 precipitation fields. Although this bias-adjustment procedure generally improves the representation of extreme precipitation (see Chapter 3, Fig. 3.4), it may occasionally lead to an overcorrection of already well-simulated events.

This behaviour is particularly evident for the July 2021 flood event. Between 10 and 15 July 2021, the observed catchment-average precipitation over the Ourthe basin at Tabreux, derived from corrected radar observations, reached approximately 123 mm (Fig. 5.3a). Over the same period, the average precipitation simulated by MAR-ERA5 amounts to 124 mm (Fig. 5.3b), indicating that the event is already realistically reproduced before bias adjustment.

However, because summer extreme precipitation events are underestimated on average over this region, the applied correction factors substantially amplify the precipitation totals associated with this event. As a result, the bias-adjusted MAR-ERA5 forcing reaches a basin-average precipitation of approximately 183 mm over the same period (Fig. 5.3c).

To assess how this local overestimation of precipitation affects the simulated discharge response, the calibrated hydrological models are forced using alternative precipitation inputs. More specifically, the MAR-ERA5 precipitation fields for the year 2021 are replaced by corrected radar observations over the same period.

Using the bias-adjusted MAR-ERA5 forcing, the simulated hourly peak discharge associated with the July 2021 flood event reaches 1333, 926, 521 and $1001 \text{ m}^3 \text{ s}^{-1}$ for GR4H, HBV, VHM and SAC-SMA respectively (Figs. 5.4a-d).

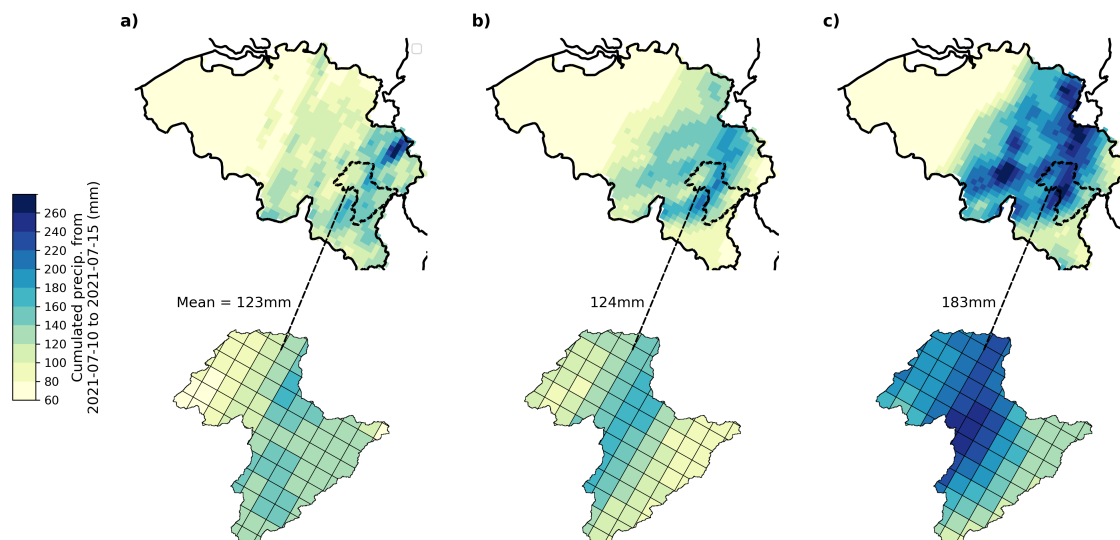


Figure 5.3: Accumulated precipitation between 10 and 15 July 2021 derived from (a) corrected radar observations (Journée et al., 2023), (b) MAR-ERA5 outputs, and (c) bias-adjusted MAR-ERA5 outputs. The lower panels show a zoom over the Ourthe watershed at Tabreux. Basin-average precipitation totals computed over the catchment are indicated above each panel.

When the forcing data are replaced by corrected radar observations, the simulated peak discharges substantially decrease, reaching 449, 361, 282 and 431 $\text{m}^3 \text{s}^{-1}$, compared to the observed peak discharge of 542 $\text{m}^3 \text{s}^{-1}$ (Figs. 5.4e-h).

These results indicate that the strong overestimations obtained with GR4H, HBV and SAC-SMA primarily originate from the meteorological forcing rather than from unrealistic hydrological parameterization. Once more realistic precipitation inputs are used, GR4H and SAC-SMA especially reproduce the July 2021 flood event much more realistically.

For these reasons, when we exclude the July 2021 event from the validation period and focus on the 2011-2020 period, NSE values are more uniform and reach around 0.40 for all four hydrological models at Tabreux (Table 5.2) with RMSE_{peak} values of the same order as for the calibration period.

To further assess the ability of the hydrological models to reproduce extreme discharges, extreme value analysis is used to characterize the distribution of peak-flow events. When extending the simulation period to 1981-2020, GR4H stands out in particular when comparing simulated discharge return levels with observations (Figs. 5.5a-c). Overall, hourly peak discharge return levels are underestimated by all four hydrological models at Tabreux, with GR4H providing the most realistic representation of the observed return levels (Fig. 5.5a). At Ortho and Mabompré, GR4H, VHM and SAC-SMA show similar performances, with underestimations being more pronounced at Ortho (Figs. 5.5b and c). In contrast, HBV appears more reactive in simulating rarer events, with simulated return levels exceeding the observed values for return periods longer than 15 years.

Overall, this analysis shows that, despite comparable performance scores during the calibration and validation periods, the distribution of peak discharges is more realistically reproduced by GR4H than by the other models

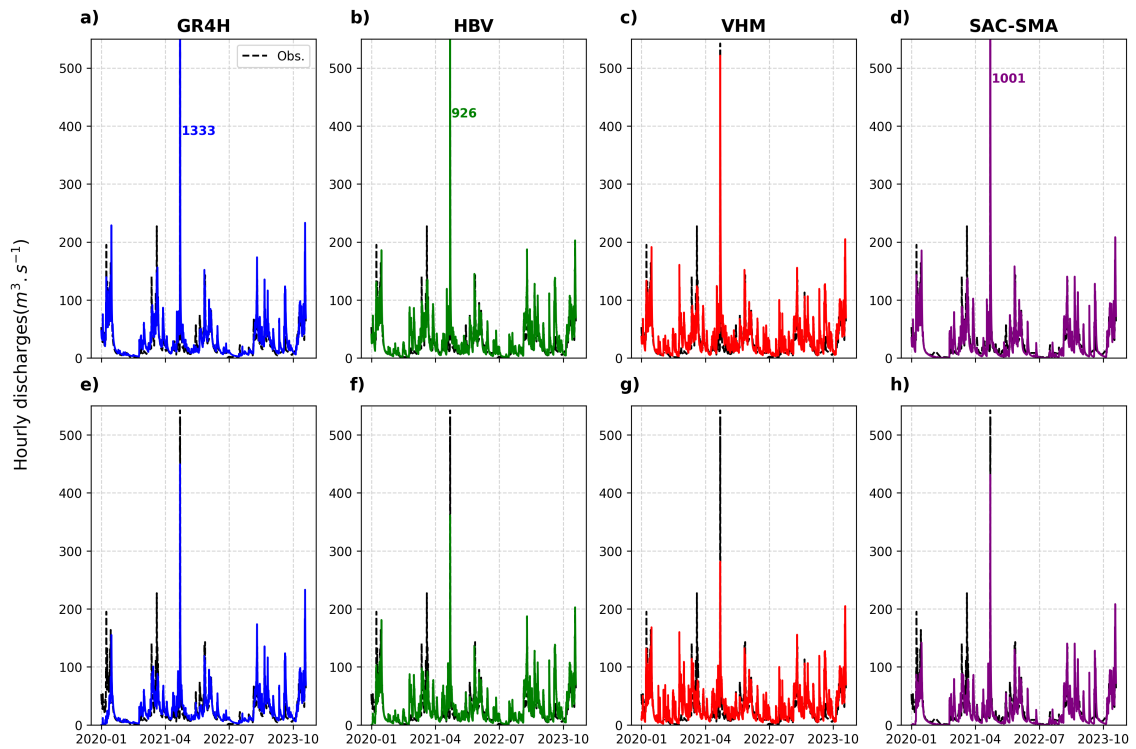


Figure 5.4: Simulated hourly discharges at the Tabreux station over the 2020-2023 period using bias-adjusted MAR-ERA5 precipitation forcing (a-d) and corrected radar precipitation forcing (e-h). Panels correspond respectively to GR4H, HBV, VHM and SAC-SMA simulations. The dashed black curve represents the observed discharge. Values exceeding the upper limit of the y-axis are indicated numerically.

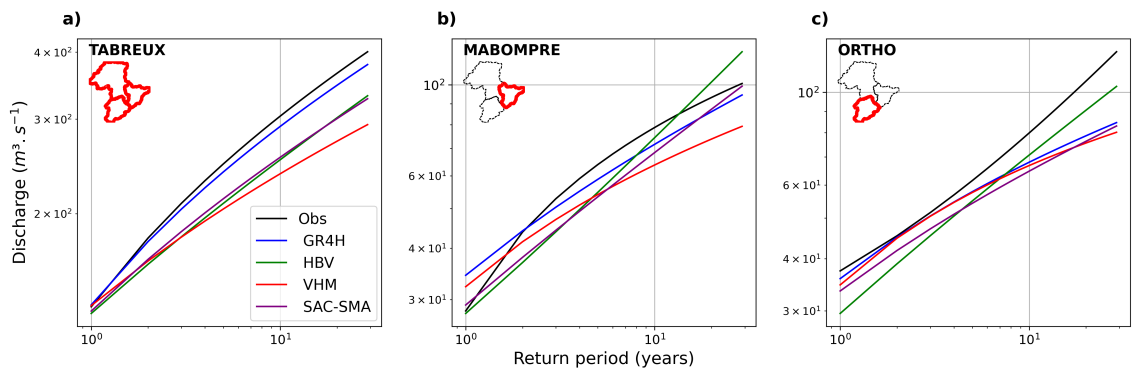


Figure 5.5: Modelled hourly discharge return levels at (a) Tabreux, (b) Mabompré, and (c) Ortho over the 1981-2020 period. A Peaks-over-Threshold (POT) approach is applied using the 95th percentile of observed discharges as threshold, resulting in the selection of approximately two to three extreme events per year. Observed return levels are shown for comparison.

5.3 Comparison of hydrological model responses to global warming

In this section, the four calibrated hydrological models are forced using future climate projections. Based on the results presented in Chapter 4, we focus on the four climate model configurations that provide the most coherent future evolution of COL activity over Belgium, namely MAR-MIR, MAR-EC3, MAR-MPI, and MAR-IPSL (see Fig. 4.12). To assess the sensitivity of the hydrological response to greenhouse gas forcing, we focus on the two contrasting scenarios SSP1-2.6 and SSP5-8.5. Historical simulations over 1981-2014 are also included as a reference period.

At the Tabreux station, unprecedented hydrological events are simulated over 2015-2100, especially using the GR4H, HBV and SAC-SMA models, which were previously shown to reproduce extreme peak discharges more realistically (Fig. 5.6). For instance, MAR-MIR under SSP1-2.6 produces a peak discharge close to $1000 \text{ m}^3 \text{ s}^{-1}$ in the early 2020s (Fig. 5.6a), while MAR-MPI under SSP5-8.5 exceeds $900 \text{ m}^3 \text{ s}^{-1}$ during the late 2030s (Fig. 5.6i). Overall, HBV appears to be the most reactive hydrological model. For the latter event, HBV simulates a peak discharge of almost $1900 \text{ m}^3 \text{ s}^{-1}$. As expected from the previous analyses, VHM produces a much more damped response, with peak discharges rarely exceeding $500 \text{ m}^3 \text{ s}^{-1}$ regardless of the forcing or scenario considered. SAC-SMA offers an intermediate response between HBV and VHM and also provides exceptional peak discharges over 2015-2100.

Once again, the timing of the most extreme hydrological events remains highly variable and further illustrates the non-linearity of the hydrological response to climate change. In some cases, particularly when using MAR-MIR forcings, more extreme events can even be found under SSP1-2.6 than under SSP5-8.5. This behaviour highlights the important role of internal climate variability on extreme hydrological events.

Nevertheless, despite the large spread between hydrological models and climate forcings, all simulations suggest that flood events exceeding those observed during July 2021 remain plausible over the 21st century. Even VHM, which underestimated the observed July 2021 peak discharge of approximately $542 \text{ m}^3 \text{ s}^{-1}$ by a large amount when forced by radar-derived precipitation, reaches comparable values under MAR-MPI SSP5-8.5 forcings (Fig. 5.6k).

Similar conclusions can be drawn for the two upstream contributing sub-basins at the Mabompré (Fig. 5.7) and Ortho (Fig. 5.8) stations. In both cases, future simulations produce flood events exceeding those observed during the historical period, although the magnitude of the response remains strongly dependent on both the hydrological model and the climate forcing considered. However, for these smaller sub-basins, SAC-SMA appears to be as reactive as HBV, which is especially marked after 2070 using MAR-EC3 as forcing under SSP5-8.5 (Fig. 5.7h and Fig. 5.8h).

Similarly to the analysis performed in Chapter 3, an extreme value analysis (EVA) is applied over 30-year moving windows ranging from 2021-2050 to 2071-2100, with a 10-year shift between consecutive windows. The analysis is performed on the simulated hydrological outputs.

At the Tabreux station, the evolution of the 2-year hourly discharge return levels with respect to the GMST anomaly is relatively consistent across GR4H, HBV, and SAC-SMA

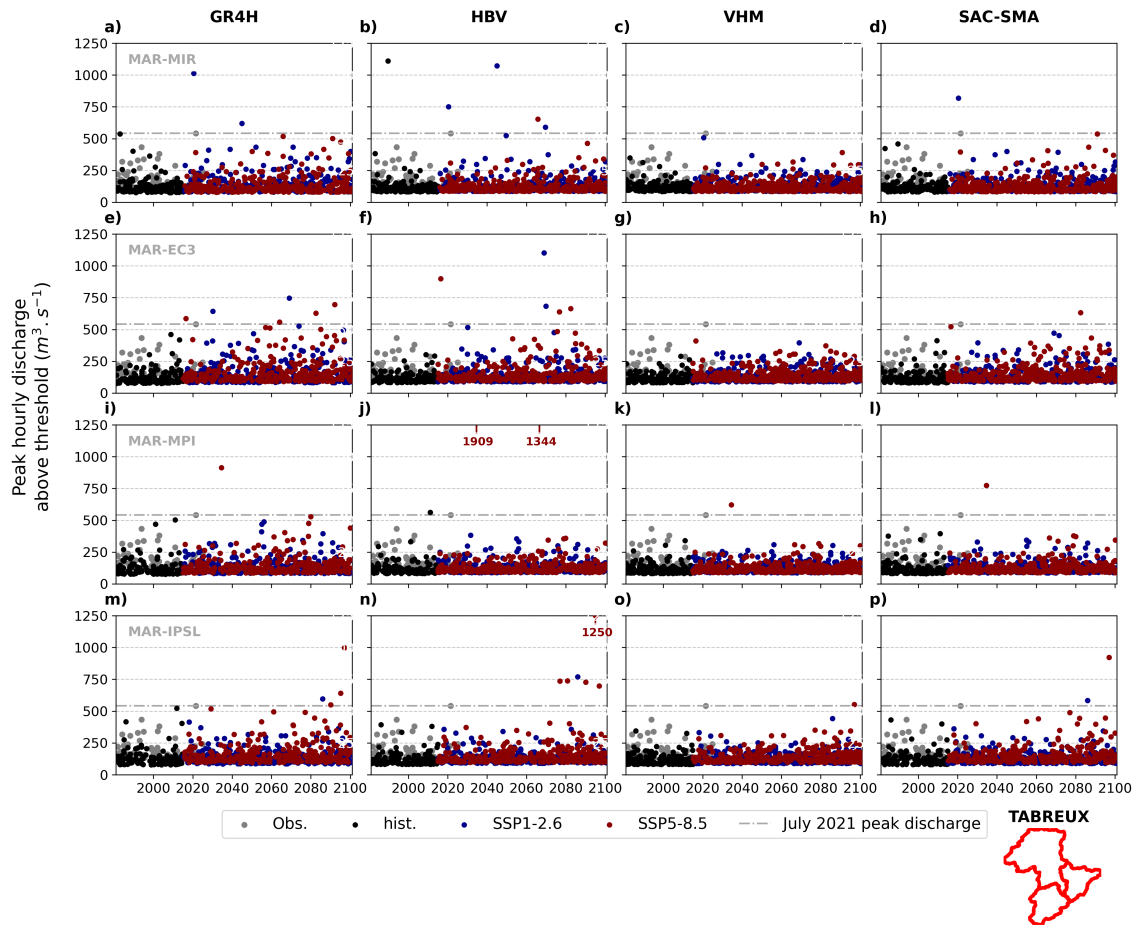


Figure 5.6: Peak hourly discharges exceeding the 95th percentile at the Tabreux station simulated using (a) GR4H, (b) HBV, (c) VHM and (d) SAC-SMA, forced by MAR-MIR under the historical, SSP1-2.6 and SSP5-8.5 scenarios. Panels (e-p) show the corresponding simulations using MAR-EC3, MAR-MPI, and MAR-IPSL as climate forcings. Hydrological events are declustered using a 7-day temporal separation criterion. Values exceeding the upper limit of the y-axis are indicated numerically.

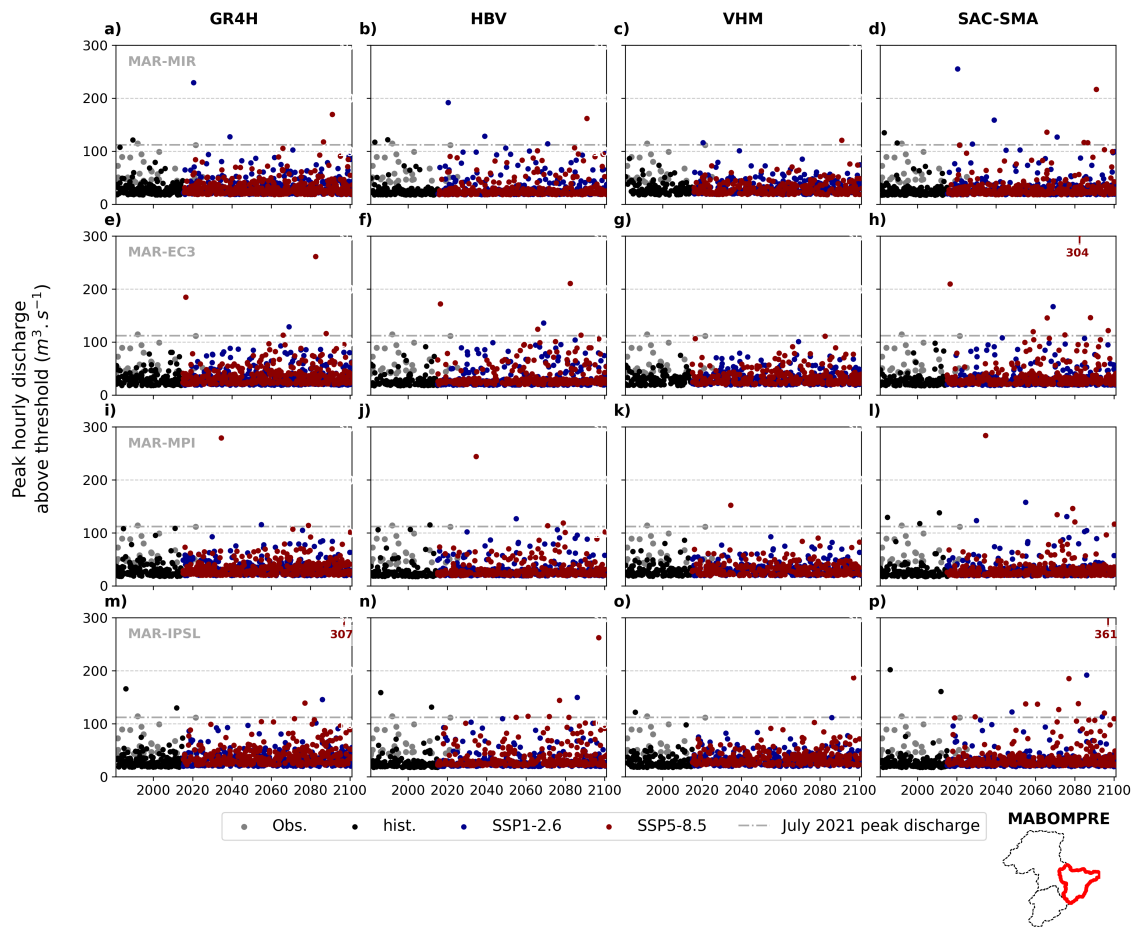


Figure 5.7: Peak hourly discharges exceeding the 95th percentile at the Mabompré station simulated by (a) GR4H, (b) HBV, (c) VHM and (d) SAC-SMA forced by MAR-MIR under the historical, SSP1-2.6, and SSP5-8.5 scenarios. Panels (e-p) show the corresponding simulations using MAR-EC3, MAR-MPI, and MAR-IPSL as climate forcings. Hydrological events are declustered using a 7-day temporal separation criterion. Values exceeding the upper limit of the y-axis are indicated numerically.

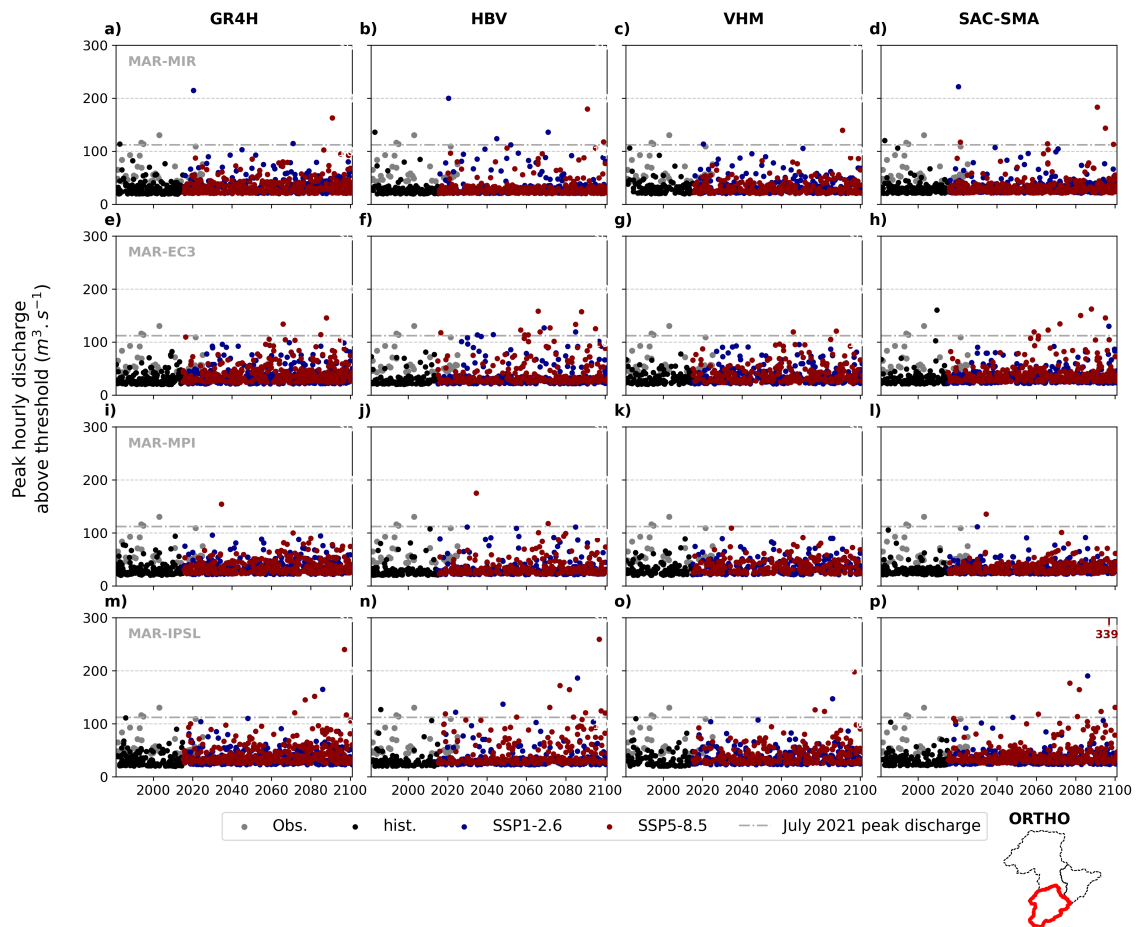


Figure 5.8: Peak hourly discharges exceeding the 95th percentile at the Ortho station simulated by (a) GR4H, (b) HBV, (c) VHM and (d) SAC-SMA forced by MAR-MIR under the historical, SSP1-2.6, and SSP5-8.5 scenarios. Panels (e-p) show the corresponding simulations using MAR-EC3, MAR-MPI, and MAR-IPSL as climate forcings. Hydrological events are declustered using a 7-day temporal separation criterion. Values exceeding the upper limit of the y-axis are indicated numerically.

5. Hydrological response to climate change in the Ourthe basin

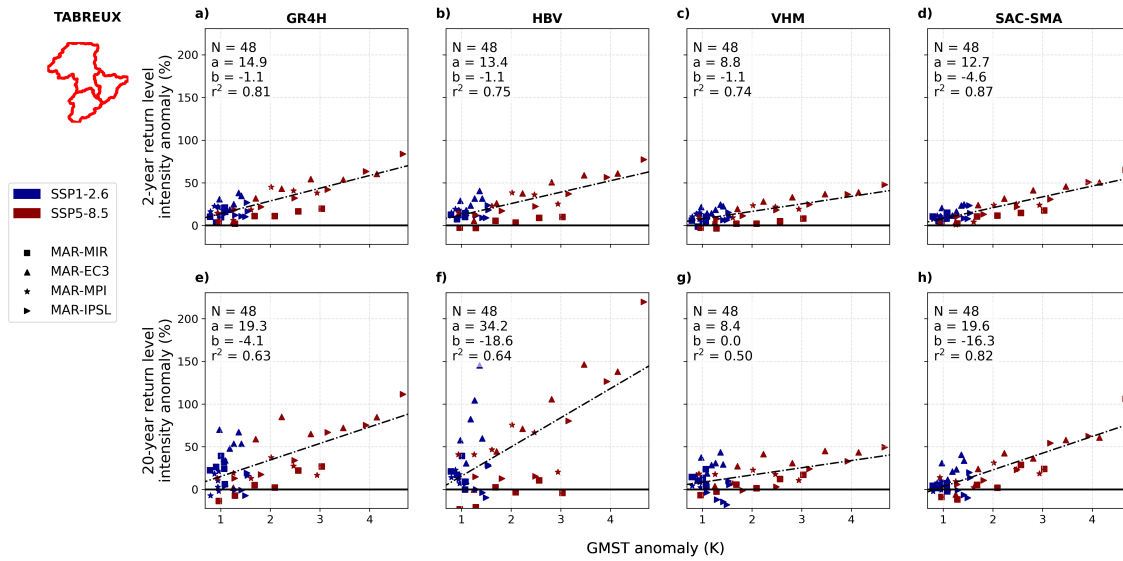


Figure 5.9: Anomalies of the 2-year hourly discharge return levels as a function of the Global Mean Surface Temperature (GMST) anomaly (K) at the Tabreux station using (a) GR4H, (b) HBV, (c) VHM and (d) SAC-SMA. Panels (e-h) show the corresponding results for the 20-year return levels. A Peaks-over-Threshold (POT) approach is applied for the computation of return levels, using the 95th percentile of hourly discharges as threshold. Linear regressions of slope a and intercept b are fitted to the N -point data clouds. The corresponding coefficient of determination (r^2) is also indicated.

(Figs. 5.9a, b and d), which exhibit intensity increases of 14.9, 13.4, and 12.7% K⁻¹, respectively. In contrast, VHM exhibits a lower scaling rate of only 8.8% K⁻¹, consistent with the damping behaviour observed over the historical period (Fig. 5.9c).

Nevertheless, the differences between models become more pronounced when considering the 20-year return levels (Figs. 5.9e-h). The stronger sensitivity of HBV results in an intensity increase of 34.2% K⁻¹, whereas GR4H, VHM, and SAC-SMA exhibit lower scaling rates of 19.3, 8.4, and 19.6% K⁻¹, respectively. While GR4H and SAC-SMA display relatively similar scaling rates across the two return periods, HBV stands out as particularly sensitive to warming when rare hydrological extremes are considered. In contrast, the damping behaviour of VHM is again confirmed by this analysis.

For the Mabompré and Ortho sub-basins (Figs. 5.10 and 5.11), similar behaviours are observed for the 2-year return levels. However, for rarer events, the scaling rates reached by HBV and SAC-SMA are much higher (Fig. 5.10f and h and Fig. 5.11f and h), where the 20-year return level increases by more than 30% K⁻¹.

Overall, these results suggest that uncertainties associated with hydrological model structure become increasingly important for the most extreme flood events. Although all four models indicate increasing discharge return levels under warmer climate conditions, the magnitude of this increase remains strongly model dependent, especially for rarer events and smaller contributing watersheds. In particular, the large differences obtained between HBV, SAC-SMA, GR4H, and VHM for the 20-year return levels highlight the substantial uncertainty associated with hydrological model choice when assessing future flood hazards.

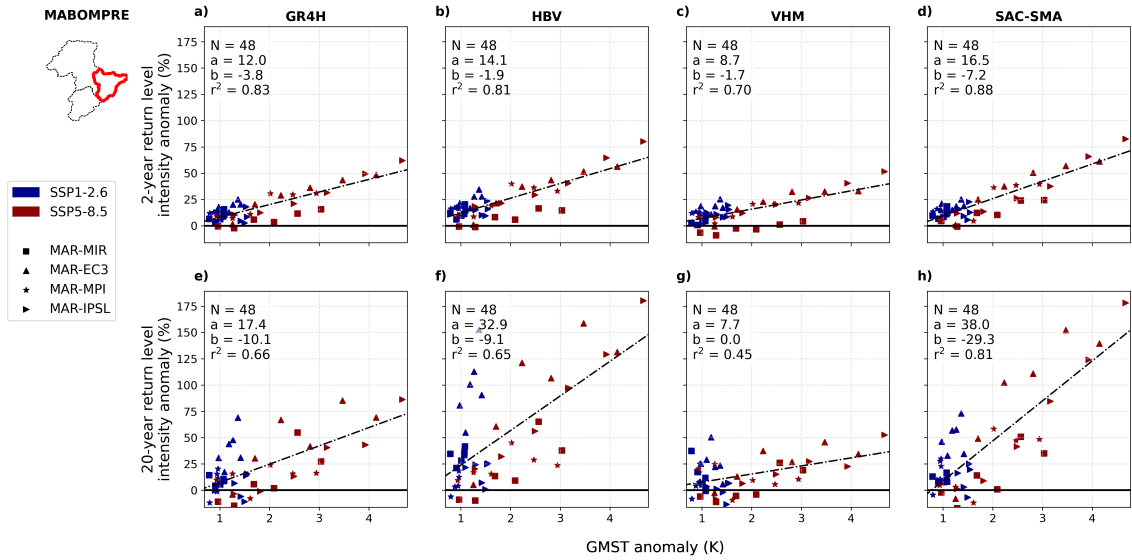


Figure 5.10: Anomalies of the 2-year hourly discharge return levels as a function of the Global Mean Surface Temperature (GMST) anomaly (K) at the Mabompré station using (a) GR4H, (b) HBV, (c) VHM and (d) SAC-SMA. Panels (e-h) show the corresponding results for the 20-year return levels. A Peaks-over-Threshold (POT) approach is applied for the computation of return levels, using the 95th percentile of hourly discharges as threshold. Linear regressions of slope a and intercept b are fitted to the N -point data clouds. The corresponding coefficient of determination (r^2) is also indicated.

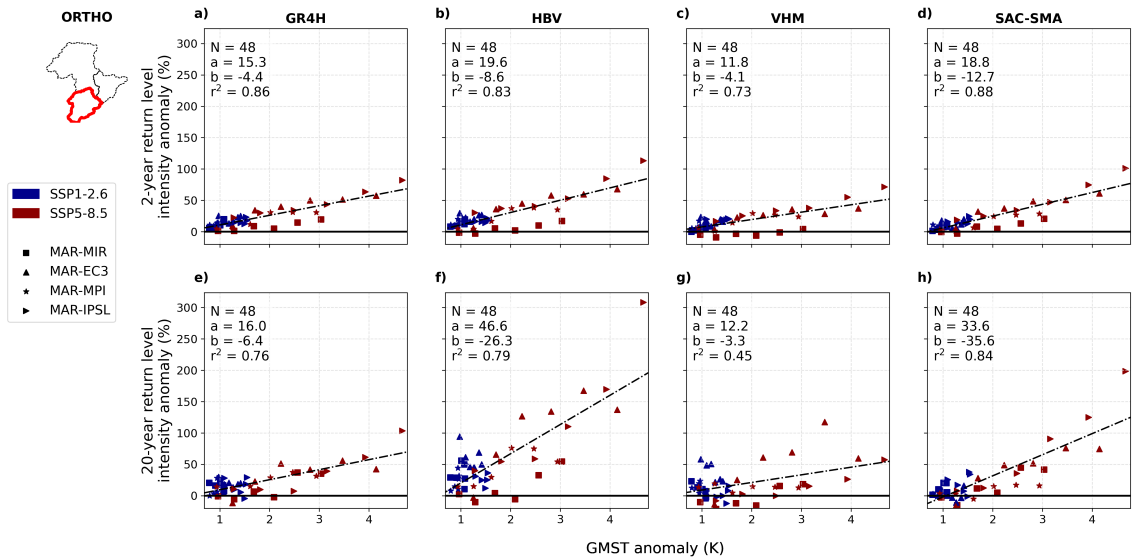


Figure 5.11: Anomalies of the 2-year hourly discharge return levels as a function of the Global Mean Surface Temperature (GMST) anomaly (K) at the Ortho station using (a) GR4H, (b) HBV, (c) VHM and (d) SAC-SMA. Panels (e-h) show the corresponding results for the 20-year return levels. A Peaks-over-Threshold (POT) approach is applied for the computation of return levels, using the 95th percentile of hourly discharges as threshold. Linear regressions of slope a and intercept b are fitted to the N -point data clouds. The corresponding coefficient of determination (r^2) is also indicated.

5.4 Implications and limitations

5.4.1 Historical performance of the MAR-WOLFHydro modelling chain

Our results show that unprecedented peak discharges can be expected within the 21st century despite the various hydrological model sensitivities. The four hydrological models used in this study provide an interesting combination, as their conceptual complexity varies from GR4H, which is the least complex, to SAC-SMA, which is the most complex. Over the calibration period, all four models reached similar NSE scores of around 0.7 at Tabreux. However, when computed over 1981-2020, the peak discharge return levels are better represented by GR4H. GR4H also appears to be the closest to the observed July 2021 peak discharge when forced by corrected radar observations. We believe that these satisfactory NSE scores and coherent behaviours over the validation period provide confidence in the framework's ability to reproduce broad hydrological variability, while substantial uncertainty remains for individual extreme events.

The exercise conducted here is particularly ambitious since bias-adjusted MAR-ERA5 outputs are used as observations for the calibration of hydrological models. This framework therefore provides the opportunity to directly apply the calibrated models to future bias-adjusted climate model outputs. However, these inputs are not perfect and, due to the bias-adjustment procedure, historical precipitation events can in some cases be underestimated or overestimated, as was the case for the July 2021 event (Fig. 5.3). Moreover, the bias-adjustment procedure itself is based on an amplification of the MAR hourly precipitation signal since the model fails to accurately represent hourly precipitation peaks (de Vyver et al., 2021).

These corrected hourly precipitation peaks are also not necessarily perfectly timed and can be shifted relative to the observed timing of precipitation peaks at stations. Based on precipitation days identified at RMI stations over the 1981-2004 period, we therefore compared the timing of observed and MAR-ERA5 simulated hourly precipitation peaks (Fig. 5.12). For each precipitation day identified in the station data ($>1 \text{ mm day}^{-1}$), we computed the hourly difference between the modelled and observed peaks within a 24-hour window centred on the observed peak. When fitting a normal distribution with location parameter μ and scale parameter σ to the observed distributions of hourly shifts, the distributions are generally centred around a 0-hour shift, with a spread of approximately 6 hours at all station locations. Nevertheless, in some cases, the shift can exceed 6 hours.

For all these reasons, the calibration of hydrological models remains challenging, as the optimisation procedure must account for temporal shifts in precipitation peaks while attempting to reproduce observed river discharges. Nevertheless, as mentioned earlier, the validation period confirms that the calibrations do not generate unrealistic hydrological behaviour within the studied catchment. However, the historical performance of the various hydrological models may primarily reflect their ability to adapt to imperfections in the observational products, rather than a genuinely superior ability to represent peak discharges.

These challenges also help explain why the application of hydrological models to smaller contributing sub-basins is more difficult. For these smaller catchments, the

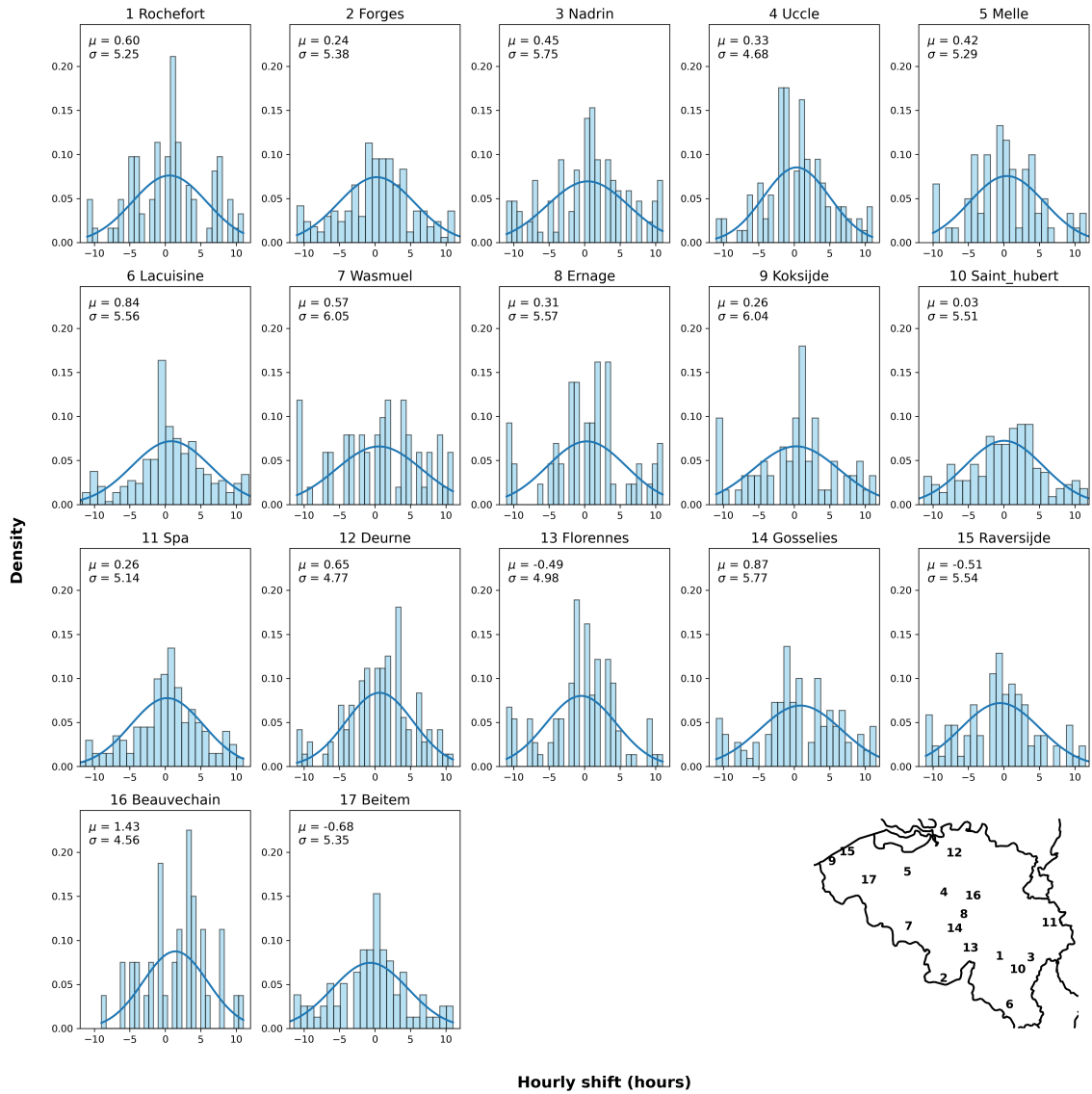


Figure 5.12: Hourly shifts between observed and MAR-ERA5 simulated precipitation peaks for precipitation days over the 1981-2004 period. For each precipitation day identified at the station locations ($>1 \text{ mm day}^{-1}$), the temporal difference between the observed and simulated hourly precipitation peak is computed within a 24-hour window centred on the observed peak. A normal distribution with location parameter μ and scale parameter σ is fitted to the resulting distributions. Station locations are shown on the map in the lower-right corner.

limitations of the bias-adjusted climatic inputs become even more important, as they are more reactive to precipitation forcing and therefore more sensitive to uncertainties associated with the 5-km spatial resolution of the climatic data.

5.4.2 Relevance of projected discharges

Our analysis reveals that, within the MAR-WOLFHydro framework, peak hourly discharges are generally best represented by the simplest conceptual model, GR4H, which achieves performances comparable to those of the more complex models in terms of NSE while providing the most realistic representation of discharge return levels. These results also demonstrate that increasing model complexity does not necessarily lead to a better representation of extreme discharges. For instance, VHM has an intermediate level of complexity between HBV and SAC-SMA, yet it is outperformed by both models in the simulation of extreme events.

The damping effect of VHM is also clearly visible in the projected future peak discharges (Fig. 5.6), where it rarely produces events exceeding the magnitude of the July 2021 flood. In contrast, GR4H, HBV, and SAC-SMA simulate several substantially larger flood events under future climate conditions.

Given its comparatively good representation of historical discharge return levels and its realistic response when forced with corrected radar precipitation, GR4H is selected as the reference hydrological model for the subsequent analysis of projected changes. In this chapter, the sensitivity of projected discharge changes was evaluated using two contrasting climate scenarios (SSP1-2.6 and SSP5-8.5) together with four climate models. By extending the analysis to all SSP scenarios and re-evaluating the GR4H response to global warming, a more robust estimate of future discharge changes can be obtained (Fig. 5.13). At the Tabreux station, the 2-year and 20-year return levels exhibit intensity increases of 14.5 and 18.7% K⁻¹, respectively.

Interestingly, these projected changes are strongly season-dependent. Most of the increase in discharge extremes originates from winter events (DJF), for which the 2-year and 20-year return levels increase by 14.0 and 17.2% K⁻¹, respectively (Figs. 5.13b,e). In contrast, summer (JJA) discharge extremes do not exhibit any robust increase with global warming and even show slightly negative scaling rates (Figs. 5.13c,f), although the very low coefficients of determination indicate that these relationships are not statistically meaningful.

We believe this behaviour highlights the dominant influence of internal climate variability and event-to-event stochasticity on summer flood-generating processes. It also reflects changes in mean precipitation simulated by the climate models. As discussed in Chapter 3, Section 1.2.2.2, mean precipitation in Belgium is expected to increase during winter, while a decrease is projected during summer. For the climate models considered in this chapter, winter mean precipitation increases by approximately 9% K⁻¹ of additional global warming, with a coefficient of determination of $R^2 = 0.69$ (see Fig. F.2b in Appendix F). In contrast, summer precipitation exhibits a weak negative relationship with global warming, with a coefficient of determination of $R^2 = 0.21$ (see Fig. F.2d in Appendix F). These changes in mean precipitation therefore provide more favourable average moisture conditions during winter than during summer as global warming increases. Consequently, the projected increase in annual discharge extremes appears to be primarily influenced by changes occurring during the cold season.

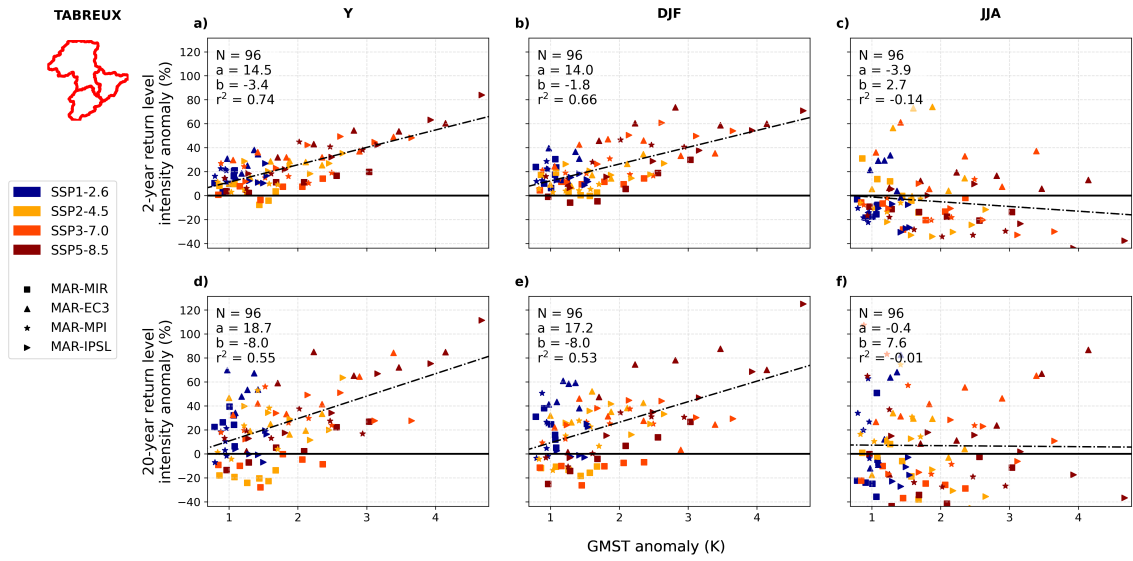


Figure 5.13: Anomalies of the 2-year (a-c) and 20-year (d-f) hourly discharge return levels as a function of the Global Mean Surface Temperature (GMST) anomaly at the Tabreux station. Results are shown for the whole year (Y; a,d), winter (DJF; b,e), and summer (JJA; c,f), using the GR4H hydrological model forced by the four selected MAR climate simulations and all SSP scenarios. Return levels are computed using a Peaks-over-Threshold (POT) approach with the 95th percentile of hourly discharge as threshold. Linear regressions of slope (a) and intercept (b) are fitted to the (N)-point data clouds, and the corresponding coefficient of determination (r^2) is reported. The same figure with all the seasons plotted is available in Fig. F.1 in Appendix F

In Chapter 3, we investigated the scaling of extreme precipitation with global warming and showed that daily precipitation extremes over Belgium generally follow a Clausius-Clapeyron scaling of approximately $7\% \text{ K}^{-1}$ regardless of the return period (Fig. 3.2). Here, we reproduce a similar analysis for the Ourthe watershed at Tabreux using the four climate models retained in this chapter and the four climate scenarios used in this thesis. The projected 2-year and 20-year daily precipitation return levels exhibit scaling rates of 8.4 and $5.7\% \text{ K}^{-1}$, respectively (Fig. 5.14a). As expected, internal climate variability introduces larger uncertainties for the rarer events, resulting in a decrease of the coefficient of determination from 0.68 to 0.35 for the 20-year return level. Spatially, the strongest 20-year precipitation scaling is found over the southern elevations of the watershed (Fig. 5.14b), where values locally reach up to $15\% \text{ K}^{-1}$.

Nevertheless, these results indicate that the scaling of peak discharge with global warming can exceed the scaling of extreme precipitation itself. This hydrological amplification was also reported by Ludwig et al. (2023). Using a pseudo-global warming experiment, they showed that the July 2021 precipitation event would intensify by approximately $9\% \text{ K}^{-1}$, whereas the associated peak discharge in the Ahr catchment would increase by approximately $21\% \text{ K}^{-1}$.

These results, obtained using the MAR-WOLFHhydro modelling framework, demonstrate the potential of combining high-resolution regional climate simulations with hydrological modelling to assess future flood hazards in Belgium. Although our analyses focus on return periods of up to 20 years, it is nevertheless noteworthy that several climate model simulations produce peak discharges comparable to or higher than the observed July 2021 peak at different times during the future period, including under relatively low global warming levels (see Fig. 5.6a, e, i and m). Therefore, despite the exceptional magnitude of the July 2021 event relative to the historical record, our results suggest that comparable

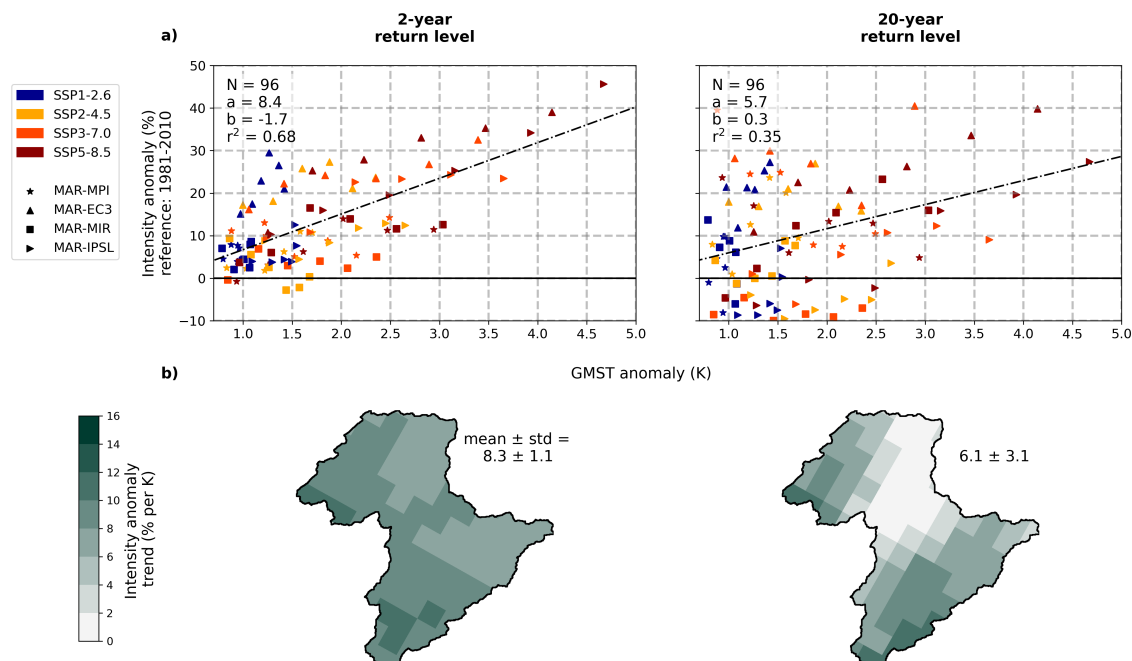


Figure 5.14: Anomalies of the 2-year and 20-year daily precipitation return levels as a function of the Global Mean Surface Temperature (GMST) anomaly over the Ourthe watershed at Tabreux. The analysis is performed using the four climate models retained in this chapter (MAR-EC3, MAR-MIR, MAR-IPSL, and MAR-MPI) and all available SSP scenarios. Panel (a) shows the scaling relationships derived from 30-year moving windows, while panel (b) presents the spatial distribution of the corresponding scaling rates over the watershed. Linear regressions of slope a and intercept b are fitted to the data clouds, and the corresponding coefficient of determination (r^2) is indicated.

or even larger peak discharges remain plausible within the 21st century.

As an extension of these results, similar analyses could be applied to other Belgian catchments and to additional CORDEX-Be.2 simulations. Nevertheless, the choice of hydrological model remains a critical source of uncertainty. The Ourthe catchment already represents a relatively complex hydrological system. In more regulated basins, including reservoirs, dams, or stronger anthropogenic influences, modelling assumptions could have an even larger impact on projected discharge changes.

Take-home messages

- The MAR-WOLFHydro framework is able to reproduce the main characteristics of historical flood events over the Ourthe catchment despite the uncertainties associated with bias-adjusted climatic forcings.
- Hydrological model performance depends on the evaluation metric considered. While VHM achieves the highest NSE values, GR4H provides a more realistic representation of extreme discharge return levels and of the July 2021 flood event when realistic precipitation inputs are used.
- Calibration of hydrological models becomes more challenging for the smaller contributing sub-basins, highlighting the limitations associated with the 5-km spatial resolution of the climatic forcing data.

- Future simulations consistently indicate that flood events exceeding the magnitude of the July 2021 flood remain plausible within the 21st century, although their timing and magnitude remain strongly dependent on both climate and hydrological model uncertainties.
- The projected intensification of extreme discharges exceeds the projected intensification of extreme precipitation, suggesting a non-linear amplification of the climatic signal by hydrological processes.
- Hydrological model structure remains a major source of uncertainty for future flood projections, particularly for the rarest events and for smaller sub-basins.

CHAPTER 6

Discussion and perspectives

This thesis investigated how global warming may affect extreme precipitation and flood hazards in Belgium through a multi-scale hydroclimatic framework.

First, we quantified the scaling of extreme precipitation with global warming using simulations from the regional climate model MAR (Chapter 3). Second, we examined the role of cut-off lows (COLs) in shaping the occurrence of extreme precipitation events at both European and Belgian scales. We also assessed how their stochastic behaviour contributes to the non-linear response of rare precipitation extremes to climate change (Chapter 4). Third, we forced calibrated conceptual rainfall-runoff models with MAR outputs over the Ourthe catchment to investigate the hydrological consequences of projected climatic changes and the future evolution of flood hazards (Chapter 5).

Taken together, these results provide new insights into the mechanisms controlling hydroclimatic extremes in Belgium and their potential evolution under future climate conditions. They also highlight the importance of considering interactions between atmospheric processes, internal climate variability, and catchment-scale hydrological responses when assessing future flood risks.

In this chapter, we first discuss the implications of the projected changes in extreme precipitation intensity and frequency for Belgium. We then examine the non-linear nature of the hydroclimatic response identified in this thesis and explore the mechanisms that may explain the amplification of extreme flood events. Subsequently, we discuss the physical drivers of COL-related extreme precipitation and their contribution to future hydroclimatic extremes. Finally, we summarize the main limitations of this work and outline perspectives for future research.

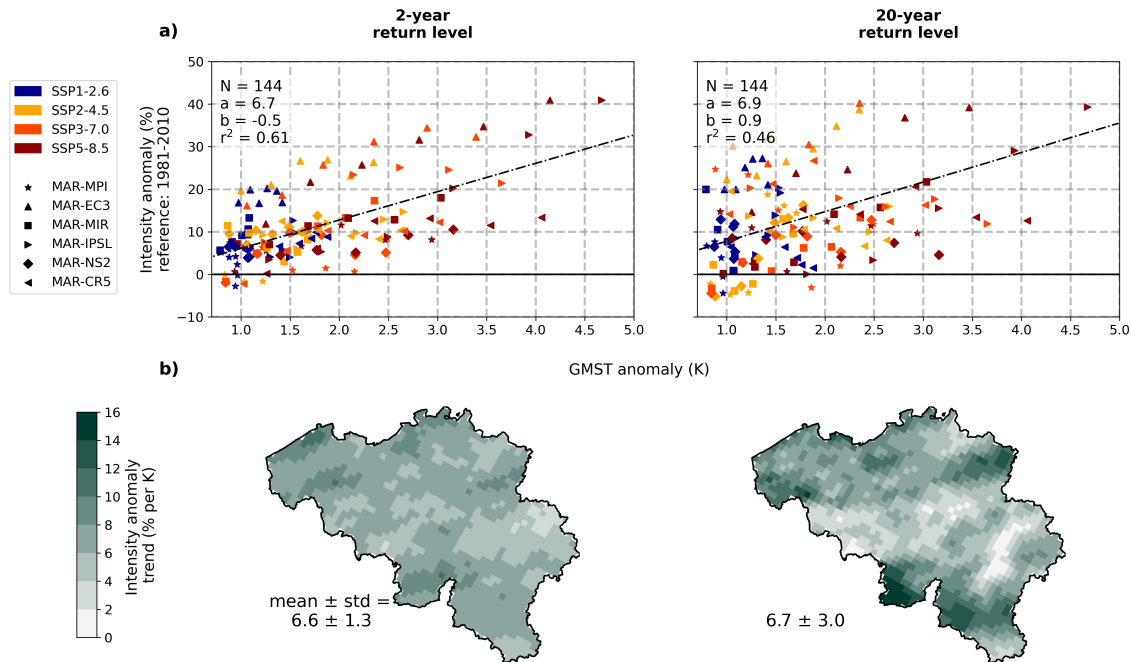


Figure 6.1: Same as Fig. 3.2, but restricted to the 2-year and 20-year daily precipitation return levels.

6.1 Implications of changes in extreme precipitation for Belgium

Extreme precipitation events are projected to become both more frequent and more intense in Belgium. In Chapter 3, we showed that daily precipitation extremes generally increase by approximately 7% per additional degree of global warming, regardless of the considered return period (Fig. 6.1a). This scaling is close to the Clausius-Clapeyron (CC) relationship, which predicts an increase in atmospheric moisture-holding capacity of about $7\% \text{ K}^{-1}$.

Although the average scaling remains relatively stable across return periods, the relationship between precipitation intensity and global mean surface temperature (GMST) becomes less robust for rarer events. The coefficient of determination decreases substantially for higher return levels (Fig. 6.1a), indicating a stronger influence of internal climate variability on the occurrence of rare precipitation extremes. Consequently, while the thermodynamic response of extreme precipitation appears robust at the regional scale, the timing and magnitude of individual high-impact events remain strongly stochastic.

The extension of the analysis to precipitation accumulations ranging from 1 to 72 hours revealed similar scaling behaviours across timescales. Depending on the duration considered, extreme precipitation intensities increase by approximately $5\text{--}7\% \text{ K}^{-1}$, suggesting that the thermodynamic response to warming remains the dominant driver of future changes in precipitation extremes. The robustness of this result is further supported by Dierickx et al. (2025), who reported comparable scaling rates for sub-daily precipitation extremes using an ensemble of Euro-CORDEX simulations over Belgium.

This interpretation is supported by previous studies conducted over Belgium. For example, Schoofs et al. (2025) showed that the frequency of the weather types associated with extreme precipitation is not projected to increase substantially in future climate

conditions. Nevertheless, precipitation intensities still increase within these weather types, highlighting the dominant role of thermodynamic amplification. Taken together, these results suggest that future changes in Belgian precipitation extremes are primarily controlled by the increasing moisture content of the atmosphere rather than by major shifts in large-scale circulation patterns.

Nevertheless, this apparent linear scaling masks important spatial heterogeneities in the response of extreme precipitation to global warming. MAR simulations suggest that the steepest regions of the Ardennian Massif, as well as parts of the Belgian coastal zone, may experience substantially higher scaling rates than the national average (Fig. 6.1b). In some locations, scaling values locally exceed $10\text{--}15\% \text{ K}^{-1}$, indicating the potential emergence of super-Clausius-Clapeyron behaviour.

Consequently, the Clausius-Clapeyron relationship should primarily be interpreted as a first-order estimate of future changes in precipitation extremes over Belgium. Local geographical factors, including topography and proximity to moisture sources, may substantially amplify the response in specific regions (Moustakis et al., 2020).

These regional differences are consistent with findings from previous studies showing that precipitation-temperature scaling can vary substantially across relatively small spatial domains due to differences in regional weather patterns and local geographical factors. For instance, Schroeer and Kirchengast (2018) found lower precipitation-temperature scaling rates in the mountainous regions of south-eastern Austria than in the eastern lowlands, while Moustakis et al. (2020) showed that orography, atmospheric convection, and large-scale dynamics can contribute to regional deviations from Clausius-Clapeyron scaling. Such results highlight that the Clausius-Clapeyron scaling should not be interpreted as a spatially uniform relationship, particularly for short-duration and high-percentile precipitation.

This spatial heterogeneity has important implications for climate adaptation strategies, as future changes in extreme precipitation are unlikely to be distributed uniformly across the country.

Moreover, depending on the selected future 30-year period, and particularly for longer return periods, the average anomaly can reach up to 40% (Fig. 6.1b). Hosseinzadehtalaei et al. (2018) similarly showed that, locally, 10-year sub-daily precipitation may increase by 50-70% at Uccle over 2071-2100 under the RCP4.5 and RCP8.5 scenarios using a Euro-CORDEX ensemble. By considering multiple future 30-year periods, our analysis shows that internal variability within the driving ESMs plays a key role in modulating the most extreme future periods. As a result, even under low-warming scenarios, large anomalies can occur during the early 21st century (Fig. 6.1b).

All these results demonstrate the need to update the official Belgian IDF curves (de Vyver, 2013) in order to better account for future changes in extreme precipitation and to support the adaptation of hydraulic infrastructure, stormwater management systems, and flood protection measures.

In the extension of Chapter 3, we illustrated this issue by constructing IDF curves representative of a +3 K global warming level for several locations across Belgium. These curves were derived from bias-adjusted MAR-ERA5 outputs over the 1940-2021 period and by applying a uniform increase of 7% per degree of global warming to the estimated return levels (Fig. 6.2).

6. Discussion and perspectives

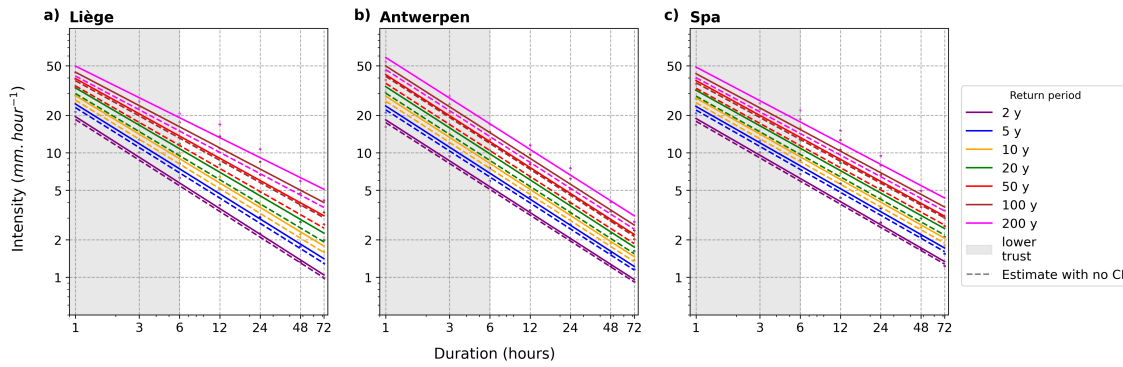


Figure 6.2: Projected +3 K Intensity-Duration-Frequency (IDF) curves for Liège, Antwerpen, and Spa. Solid coloured lines show the estimated return levels for return periods ranging from 2 to 200 years. Vertical bars indicate the associated 95% confidence intervals, while dashed lines represent the fitted return-level estimates. The widening uncertainty for short-duration and high-return-period events highlights the difficulty of constraining the most extreme precipitation intensities, even under present-day climate conditions.

Although this approach remains simplified and does not explicitly account for the spatial heterogeneity of projected scaling rates, it provides a first-order estimate of the expected changes in precipitation extremes. Moreover, the use of Clausius-Clapeyron-based scaling factors to update existing IDF relationships is increasingly considered as a practical first step for climate adaptation studies and engineering applications (Gimeno et al., 2022).

The resulting IDF curves suggest substantial increases in design precipitation intensities across all durations and return periods. In practical terms, events currently associated with return periods of several decades may become considerably more frequent under future warming levels, highlighting the need to reassess current design standards and flood-risk management strategies.

Furthermore, recent record-breaking precipitation events across Europe (Tradowsky et al., 2023; Rodero Astaburuaga, 2025) illustrate that high-impact extremes can occur even when their estimated probability remains low. Given that extreme precipitation scaling rates may vary geographically, the upper bounds of the confidence intervals associated with future IDF curves should not be disregarded when assessing flood hazards and infrastructure resilience (Fig. 6.2).

However, these implications should be interpreted in light of the limitations of the modelling framework. The scaling rates presented in this thesis are derived from MAR simulations, which are known to underestimate daily precipitation extremes and, even more markedly, sub-daily precipitation intensities (de Vyver et al., 2021). While MAR generally provides a rather realistic representation of the mean climate over Belgium, the simulation of individual extreme precipitation events remains more challenging.

Several factors contribute to these limitations, including the relatively coarse spatial resolution of the model (5 km), the use of a parameterized convection scheme (Doutreloup et al., 2019), and the hydrostatic assumption. Together, these modelling choices limit the ability of MAR to explicitly represent short-lived and highly localized convective precipitation events, which often contribute to the most severe rainfall extremes.

In this thesis, part of this limitation was addressed through the application of a bias-adjustment procedure designed to improve the representation of hourly precipitation

intensities. Nevertheless, this correction remains statistical in nature and cannot fully compensate for structural model deficiencies. Consequently, the sub-daily IDF curves presented here should be interpreted with caution and may still provide conservative estimates of the most extreme short-duration precipitation events.

6.2 Non-linearity of the hydroclimatic response

This thesis has demonstrated that, especially for the rarest events, the response of extreme precipitation to global warming is non-linear. Hence, for rare events, natural climate variability appears to play a key role in modulating the timing of the most extreme precipitation events (Kendon et al., 2023; Jain et al., 2023). Nevertheless, the intensity of the rarest events is also expected to increase more rapidly with global warming (Gründemann et al., 2022).

In the historical attribution, the 1970s appeared as a particularly favourable period for COL detachments, a feature previously linked to prolonged negative phases of the North Atlantic Oscillation (NAO) (Dai et al., 2018; Sarkar and Maity, 2021). Although Belgium is less affected by COLs than several other European regions, our results show that COLs contributed to some of the most extreme precipitation events recorded over the country during the last 80 years. Therefore, COLs play an important role in modulating the upper tails of extreme precipitation distributions.

This non-linearity in extreme precipitation is also reflected in the behaviour of extreme hydrological events. Peak hourly discharges are increasing at rates higher than the scaling of extreme precipitation over the Ourthe watershed. This hydrological amplification was recently observed by Ludwig et al. (2023), who showed that the peak discharge increase associated with the July 2021 precipitation would have been higher than the increase of the precipitation event itself under warmer conditions. Similarly, Quintero et al. (2022) and Munz et al. (2024) found that increases in extreme precipitation can lead to disproportionately larger increases in peak discharge, highlighting the nonlinear response of hydrological systems to climate change.

However, this scaling of extreme discharges with global warming appears to be much more evident during winter than during summer. In the Ourthe catchment, no robust relationship between summer discharge extremes and global warming could be identified (Chapter 5, Fig. 5.13). As discussed in the Introduction (Fig. 1.6), mean precipitation is projected to increase during winter and decrease during summer over Belgium, while annual precipitation totals remain relatively stable.

We therefore believe that the hydrological conditions favourable to flooding, particularly high antecedent soil moisture and catchment saturation, will be met more frequently during winter in future climate conditions. In contrast, the occurrence of extreme summer floods may remain strongly dependent on year-to-year climate variability. The July 2021 flood event illustrates this behaviour well, as the exceptional rainfall occurred after a period of already elevated precipitation, which contributed to increasing soil saturation and amplifying the hydrological response.

These interpretations are broadly consistent with previous studies conducted in Belgium. For example, Driessen et al. (2010b) projected increasing winter discharges in the Ourthe and Gete catchments together with a reduction of summer low flows. Similarly,

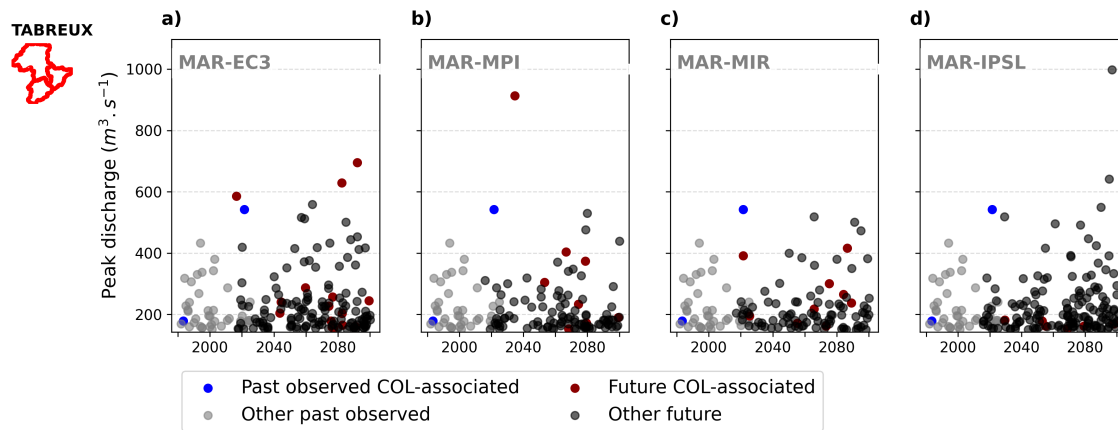


Figure 6.3: Hourly peak discharges exceeding $150 \text{ m}^3 \text{ s}^{-1}$ at the Tabreux station simulated under the SSP5-8.5 scenario using the GR4H hydrological model forced by (a) MAR-EC3, (b) MAR-MPI, (c) MAR-MIR, and (d) MAR-IPSL climate projections. Past observed COL-associated peak discharges are highlighted in blue. Future simulated COL-associated peak discharges are highlighted in red.

Wossenyeleh et al. (2021) showed that wetland storage is expected to decrease under warmer climate conditions in the Doode Bemde wetland. Such changes in summer hydrological conditions may reduce the contribution of deeper subsurface flow and limit antecedent soil saturation, thereby decreasing the efficiency with which extreme rainfall events are transformed into extreme discharge events.

Despite this dominant winter signal, some of the most damaging floods may still originate from rare summer COL-related events. In Chapter 5, four climate model configurations were used to force the hydrological models, namely MAR-EC3, MAR-MPI, MAR-MIR, and MAR-IPSL. These simulations were selected because they best reproduce the observed evolution of COL-related extreme precipitation frequencies over Belgium. Among them, MAR-EC3 and MAR-MPI provide the most realistic representation of both historical and projected COL-related extreme precipitation occurrences.

Interestingly, when considering future hourly peak discharges exceeding $300 \text{ m}^3 \text{ s}^{-1}$ at Tabreux under the SSP5-8.5 scenario (Fig. 6.3), only MAR-EC3 and MAR-MPI simulate COL-associated flood events exceeding the observed July 2021 peak discharge ($542 \text{ m}^3 \text{ s}^{-1}$; Figs. 6.3a,b). In contrast, MAR-MIR and MAR-IPSL do not produce comparable COL-related discharge extremes (Figs. 6.3c,d).

These results highlight the importance of accurately representing synoptic-scale circulation patterns conducive to extreme precipitation. Uncertainties in the simulation of COL activity directly propagate into projections of future flood hazards.

Finally, an important underlying assumption of the hydrological projections presented in Chapter 5 is the stationarity of the calibrated hydrological model parameters through time. In reality, catchment characteristics may evolve as a result of urbanization, land-use changes, river engineering, forest management, or other anthropogenic and natural processes. Such modifications can substantially alter runoff generation mechanisms and therefore influence the hydrological response to climate change.

For example, Tavakoli et al. (2014) showed for the Grote Nete catchment in Belgium that the contribution of urbanization to observed trends in extreme peak discharges is far from negligible and may even exceed the contribution associated with climatic changes. Consequently, future changes in catchment characteristics could locally amplify

Anthropogenic Climate Change differences with pre-industrial conditions for deep-moist convection

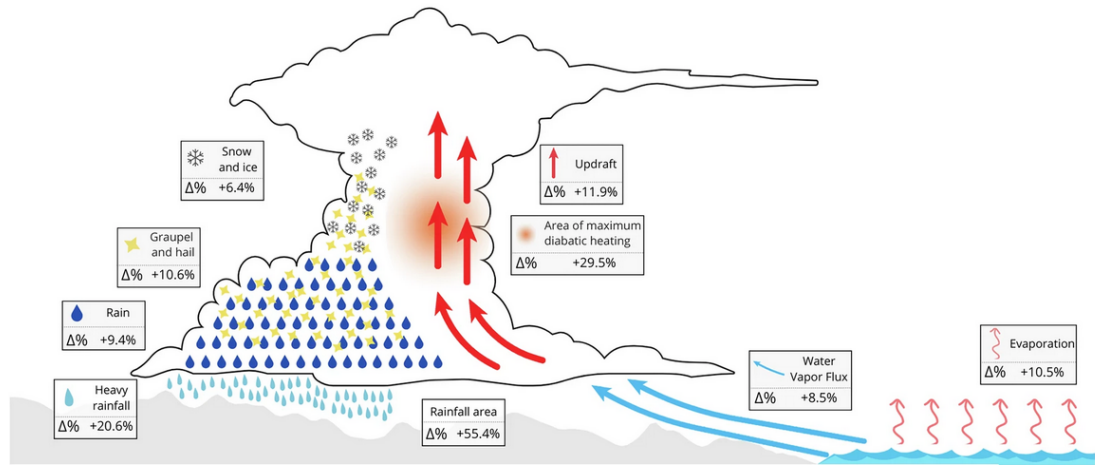


Figure 6.4: Projected changes in key atmospheric factors contributing to the precipitation intensity of the October 2024 Valencia DANA event relative to pre-industrial climate conditions. Results are derived from a pseudo-global warming experiment and illustrate the combined influence of increased atmospheric moisture availability, enhanced evapotranspiration, and greater convective instability on extreme precipitation generation (Calvo-Sancho et al., 2026).

or attenuate the projected evolution of flood hazards.

When combined with the hydrological model sensitivities identified in this thesis, these potential changes in watershed properties represent an additional source of uncertainty. They may further contribute to the non-linearity of the hydrological response to global warming.

6.3 Physical drivers of COL-related extreme precipitation

In Chapter 4, we showed that the significance of trends in COL-related extreme precipitation event frequencies strongly depends on the threshold used to define extreme events. While weak and moderate COL-related precipitation events generally do not exhibit significant long-term trends, the most extreme events display a clear increase in frequency under warmer climate conditions. We interpret this threshold dependence as evidence that thermodynamic and local dynamical processes contribute more strongly to the intensification of COL-related precipitation extremes than changes in the large-scale dynamics controlling COL occurrence itself.

Several mechanisms may explain the projected increase in COL-related precipitation extremes. These include enhanced evapotranspiration from both land and ocean surfaces, the increased moisture-holding capacity of a warmer atmosphere, and higher levels of convective instability, commonly quantified through the Convective Available Potential Energy (CAPE). The combined influence of these processes on extreme COL-related precipitation was recently illustrated by Calvo-Sancho et al. (2026), who investigated the October 2024 Valencia DANA event using a pseudo-global warming framework.

They showed that anthropogenic warming enhanced several key ingredients of the event,

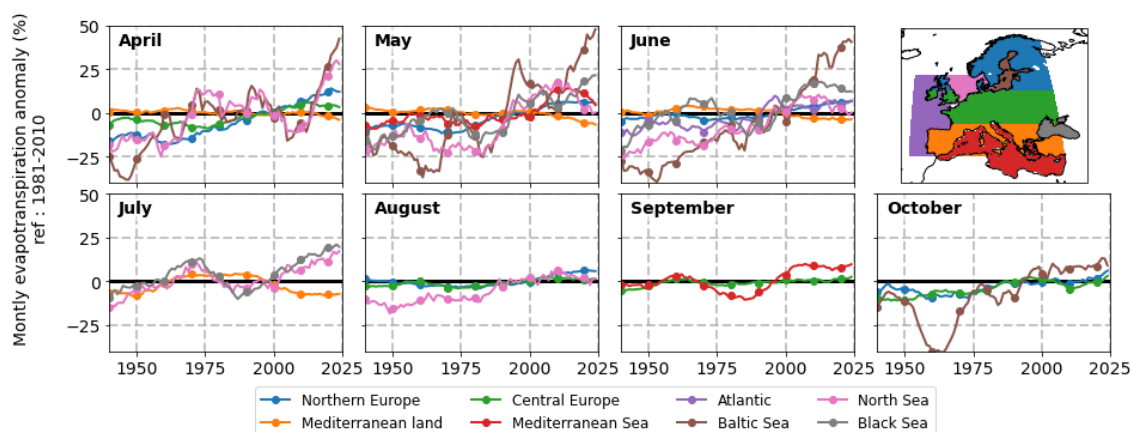


Figure 6.5: Monthly anomalies of total evapotranspiration (%) over selected European regions based on ERA5 data, expressed relative to the 1981–2010 climatological mean. Linear trends are shown only where they are statistically significant (p -value < 0.05), as determined using a Mann-Kendall test. A 15-year running mean is applied.

including a $\sim 11\%$ increase in western Mediterranean evaporation and a $\sim 12\%$ increase in Convective Available Potential Energy (CAPE). As a consequence, the simulated snow/ice, graupel/hail, and rain water contents increased by approximately $\sim 6\%$, $\sim 11\%$, and $\sim 9\%$, respectively. Together, these changes resulted in an increase of about $\sim 20\%$ in heavy rainfall intensity and a $\sim 55\%$ expansion of the affected rainfall area (Fig. 6.4).

The importance of land-atmosphere feedbacks was also highlighted by Insua-Costa et al. (2022), who showed that approximately 50% of the moisture contributing to the precipitation that occurred on 13 July 2021 originated from local moisture recycling. Their analysis identified evapotranspiration from central European forests as the main moisture source feeding the event. They further argued that the observed greening of European ecosystems may play an important role in modulating future moisture availability at the continental scale, potentially contributing to the intensification of extreme precipitation events.

To further explore this hypothesis, we computed monthly evapotranspiration anomalies over several European source regions using ERA5 reanalyses over the 1940–2023 period (Fig. 6.5). Several regions that may contribute moisture to western Europe exhibit statistically significant positive trends. In particular, central European land areas show increasing evapotranspiration during April, August, September, and October, while the North Sea exhibits significant positive trends from April to August. Northern European land regions also display increasing evapotranspiration during April, May, June, August, and October.

Although these results do not directly demonstrate an increase in moisture transport towards Belgium, they suggest that several potential moisture-source regions are releasing increasing amounts of water vapour into the atmosphere. Combined with the enhanced moisture-holding capacity of a warmer atmosphere, these changes may favour the intensification of extreme precipitation events associated with cut-off lows and other high-impact weather systems.

As shown in Chapter 4, precipitation associated with cut-off lows is not distributed uniformly across Belgium. Both the historical attribution and future projections indicate that COL-related precipitation preferentially affects the northern slopes of the Ardennian

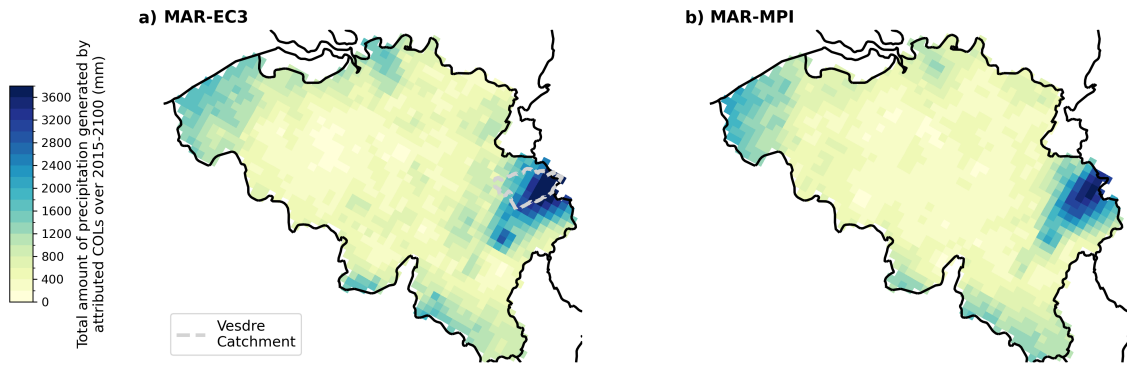


Figure 6.6: Cumulative precipitation associated with cut-off low (COL) days over the 2015-2100 period under the SSP5-8.5 scenario using bias-adjusted (a) MAR-EC3 and (b) MAR-MPI simulations. The dashed contour indicates the Vesdre catchment. The spatial distribution highlights the persistent concentration of COL-related precipitation over the northern Ardennian relief.

region. This spatial pattern highlights the important role of Belgian topography in enhancing precipitation generated by these synoptic systems.

This behaviour is projected to persist under future climate conditions. Considering the two climate model configurations that provide the most realistic representation of COL-related precipitation occurrence over Belgium, namely MAR-EC3 and MAR-MPI, the cumulative precipitation associated with COL days exceeds 3600 mm over the 2015-2100 period under the SSP5-8.5 scenario in parts of the Vesdre catchment (Fig. 6.6). These results suggest that the region most severely affected during the July 2021 floods is likely to remain one of the Belgian hotspots for COL-related precipitation in the future.

Nevertheless, other synoptic features may also become increasingly important in the future. For instance, Zhang et al. (2024) showed that atmospheric rivers are projected to transport increasing amounts of water vapour in a warming climate, thereby contributing more strongly to extreme precipitation events. Although this thesis focuses on cut-off lows, other synoptic-scale mechanisms should also be considered in future studies to better understand the full range of processes driving future hydroclimatic extremes.

6.4 Limitations and perspectives

This thesis provides new insights into the future evolution of extreme precipitation and flood hazards in Belgium, while also highlighting several limitations that offer perspectives for future research.

A first limitation lies in the use of a single regional climate model, namely MAR, to derive the scaling of extreme precipitation with global warming. Although the obtained scaling rates are consistent with previous studies conducted over Belgium (e.g., Dierickx et al., 2025), the representation of extreme precipitation events in MAR remains constrained by several model characteristics, regardless of the precipitation duration considered. These include the hydrostatic assumption, the parameterization of deep convection, and the 5-km spatial resolution, which limits the explicit representation of highly localized convective processes.

Fortunately, this thesis is part of the CORDEX.be II project, which aims to combine Belgian expertise in regional climate modelling. In addition to MAR, future high-

resolution simulations will become available using the ALARO model developed at the RMI and the COSMO-CLM model developed at KU Leuven (RMI, 2025). Future studies will therefore be able to compare the projected evolution of precipitation extremes across several regional climate models and assess the robustness of the scaling relationships identified in this thesis. Such inter-model comparisons will be particularly valuable for evaluating the representation of convective precipitation processes and for quantifying the uncertainty associated with regional climate projections.

Within this thesis, MAR biases were corrected using classical quantile-mapping techniques. It is important to acknowledge that no bias-adjustment procedure is perfect and that all methods rely on assumptions that may not hold under future climate conditions. Nevertheless, the univariate bias-adjustment approach adopted here is not necessarily less performant than more sophisticated multivariate techniques, as highlighted by Van de Velde et al. (2022). Furthermore, the bias-adjusted climate simulations were successfully integrated within the MAR-WOLFHydro framework and allowed a satisfactory representation of historical hydrological extremes, thereby providing confidence in their application for future climate impact studies.

The hydrological analysis conducted in this thesis focuses on a single watershed, namely the Ourthe catchment. Although this choice limits the direct transferability of the results to other Belgian river basins, the proposed framework could readily be applied to additional catchments. Extending the analysis to watersheds characterized by different climatic, geological, and anthropogenic conditions would provide valuable information on the spatial variability of future flood responses across Belgium.

Nevertheless, the strong dependence of the projected discharge changes on the selected hydrological model highlights the importance of methodological choices in climate impact studies. The comparison of several model structures demonstrated that hydrological uncertainty remains a major source of uncertainty when assessing future flood hazards. Within the MAR-WOLFHydro framework, GR4H appeared to provide a particularly suitable compromise, combining a realistic representation of extreme discharge return levels with a relatively simple model architecture that facilitates calibration.

In this thesis, six Earth System Models (ESMs) were used as the basis of the MAR-WOLFHydro framework. We demonstrated how the ability of these models to reproduce the synoptic-scale features associated with extreme precipitation events can influence the entire modelling chain, from projected precipitation changes to future flood hazards. For future studies aiming, for instance, to construct plausible extreme-discharge storylines for Belgium, the attribution framework developed in Chapter 4 could be used to identify and select ESMs that provide the most realistic representation of the atmospheric processes responsible for high-impact precipitation events.

Furthermore, the attribution analyses performed in this thesis suggest that some future changes in extreme precipitation occurrence may not be fully captured by considering cut-off lows alone. The most impactful precipitation events often result from a complex interplay between thermodynamic amplification, moisture transport, land-atmosphere feedbacks, and large-scale circulation patterns. Future research should therefore investigate the combined influence of multiple synoptic and mesoscale processes, including atmospheric rivers, convective systems, and cut-off lows, in order to obtain a more comprehensive understanding of future hydroclimatic extremes over Belgium.

CHAPTER 7

General Conclusions

7.1 Thesis outline

The devastating floods that affected western Europe in July 2021 raised important concerns regarding the future evolution of extreme precipitation and associated flood hazards under climate change. Within the framework of the CORDEX-BE.2 project, this thesis investigated how global warming may affect extreme precipitation and flooding in Belgium using simulations from the regional climate model MAR.

A first objective was to quantify the evolution of extreme precipitation intensity and frequency under future climate conditions. Particular attention was given to the scaling of precipitation extremes with global warming and to the associated uncertainties.

Because the response of extreme precipitation to climate change is not necessarily linear, a second objective was to investigate the role of cut-off lows (COLs), i.e. upper-level cyclonic systems that can trigger severe rainfall events such as those associated with the July 2021 Bernd event. The links between COL occurrence and extreme precipitation were analysed at both European and Belgian scales in order to better understand the atmospheric processes controlling the most impactful rainfall events.

Finally, the hydrological consequences of these projected climatic changes were investigated over the Ourthe catchment using the MAR-WOLFHydro framework. This modelling chain made it possible to assess how projected changes in precipitation may translate into future flood hazards while also quantifying the uncertainties associated with different hydrological model structures.

Taken together, the different components of this thesis provide a multi-scale assessment of hydroclimatic extremes in Belgium, spanning from synoptic-scale atmospheric processes to catchment-scale hydrological impacts.

7.2 Thesis contributions

A first contribution of this thesis is the quantification of the scaling of extreme precipitation with global warming over Belgium using simulations from the regional climate model MAR. By applying Extreme Value Analysis (EVA) techniques to precipitation extremes ranging from hourly to daily timescales, this work provides one of the first comprehensive assessments of the relationship between global warming and precipitation return levels derived from MAR simulations over Belgium.

The results show that, although extreme precipitation generally scales at rates close to the Clausius-Clapeyron relationship, the response becomes increasingly uncertain for rarer events. In particular, the scaling of high return levels exhibits substantial variability across climate model simulations, highlighting the combined influence of thermodynamic amplification and internal climate variability on extreme precipitation occurrences.

Despite these uncertainties, the projected scaling rates remain broadly consistent with recent studies conducted over Belgium. Consequently, the Clausius-Clapeyron relationship appears to provide a useful first-order approximation for updating Intensity-Duration-Frequency (IDF) curves under future climate conditions. Beyond its scientific relevance, this result has practical implications for the adaptation of infrastructures in Belgium.

The second major contribution of this thesis is the development of a COL-attribution framework, which enabled the construction of a comprehensive climatology of COL-related extreme precipitation events at both European and Belgian scales. This framework provides new insights into the role of synoptic-scale circulation patterns in shaping the occurrence of extreme precipitation and helps explain part of the variability observed in extreme precipitation time series.

Beyond the attribution exercise itself, the proposed methodology also offers a practical tool for evaluating the ability of climate models to reproduce the occurrence of COL-related precipitation extremes. This makes it possible to identify the most relevant climate model configurations for future hydroclimatic impact studies.

Our results confirm that cut-off lows represent an important source of natural variability and contribute disproportionately to the most extreme precipitation events. Although their future occurrence does not necessarily increase substantially, they are expected to continue modulating the evolution of the upper tails of precipitation distributions under warmer climate conditions through their interaction with thermodynamic amplification processes.

The third contribution of this thesis is the implementation of the MAR-WOLFH_{ydro} framework, which combines regional climate simulations, bias-adjustment procedures, and conceptual hydrological modelling within a coherent hydroclimatic modelling chain. This framework proved suitable for investigating both the future evolution of extreme discharges and the uncertainties associated with hydrological impact assessments.

The comparison of four conceptual hydrological models highlighted the substantial influence of model structure on projected flood hazards, particularly for the rarest events. These results provide a basis for future applications to additional Belgian watersheds and forthcoming CORDEX-BE.2 simulations.

Finally, this thesis contributes to the understanding of how hydrological processes may amplify the response of flood hazards to climate change. While extreme precipitation

generally scales at rates close to the Clausius-Clapeyron relationship, our results suggest that peak discharge return levels may increase at substantially higher rates. Despite the uncertainties associated with both climate and hydrological models, this hydrological amplification has important implications for future flood-risk assessments and climate adaptation strategies.

7.3 Closing remarks

Hydroclimatic extremes are expected to intensify under global warming in Belgium. Despite this general conclusion, the synoptic and hydrological processes shaping the evolution of these events deserve particular attention, as they largely determine the occurrence of the most severe and impactful extremes.

In this context, this thesis contributes to improving our understanding of future hydroclimatic risks by combining expertise in regional climate modelling and hydrological modelling. At the same time, it highlights the uncertainties inherent both to the natural climate system and to the modelling frameworks used to investigate its future evolution.

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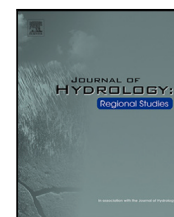
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Appendices

APPENDIX A

First paper: Increased intensity and frequency of extreme precipitation events in Belgium as simulated by the regional climate model MAR



Increased intensity and frequency of extreme precipitation events in Belgium as simulated by the regional climate model MAR

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ABSTRACT

Study region: Belgium

Study focus: In July 2021, Western Germany, the Netherlands and Belgium were hit by extreme rainfall events of unprecedented intensity, raising concerns about future trends. To assess future trends in extreme precipitation frequency and intensity, we use the regional climate model MAR at 5-km spatial resolution to conduct climate projections until 2100. MAR is forced by a set of six CMIP6 Earth System model (ESMs) with 4 IPCC scenarios ranging from low to high-end warming.

Study insights: First, an Extreme Value Analysis (EVA) is performed on bias-adjusted daily model outputs over 30-year moving windows. We find that, on average, extreme precipitation intensity rises following the Clausius-Clapeyron scaling, i.e., a 7% increase per additional degree of global warming. This trend results from a clear increase in the spread and in the modes of the extreme precipitation event distributions. Second, an EVA is performed over the complete simulated period (2015–2100). Our analyses reveal that, even under low-warming scenarios, return level maps, i.e., representing the precipitation values associated with a certain return period, are more intense than those obtained using observational data over 1951–2021. For the 21st century, the 20-year daily return level can locally reach 100 mm per day, which represents a 25%–30% increase relative to the present day.

1. Introduction

The rainfall event of July 2021 which caused major floods over western Germany, Belgium and the Netherlands was of unprecedented intensity. The amount of cumulated precipitation over 3 days (i.e., 14th to 16th of July) reached 290 mm in the basin of the Vesdre valley (purple contour in Fig. 1a) (Journée et al., 2023), accounting for approximately 20% of the annual mean precipitation over High Belgium (Erpicum et al., 2017) (Fig. 1). Belgium's topography generally follows a northwestern-southeastern gradient with its highest parts located in the Ardennian Massif near the German border (purple contour in Fig. 1b). Usually, we distinguish three topographic areas: Low Belgium (0–100 m), Medium Belgium (100–300 m) and High Belgium (above 300 m) (Fig. 1). Over Medium and Low Belgium, annual precipitation average ranges from 600 to 800 mm with a 1300 mm peak for the highest parts of Belgium. Dominant Westward Circulation patterns explain this precipitation pattern, exhibiting relatively low values for an oceanic type climate (Erpicum et al., 2017).

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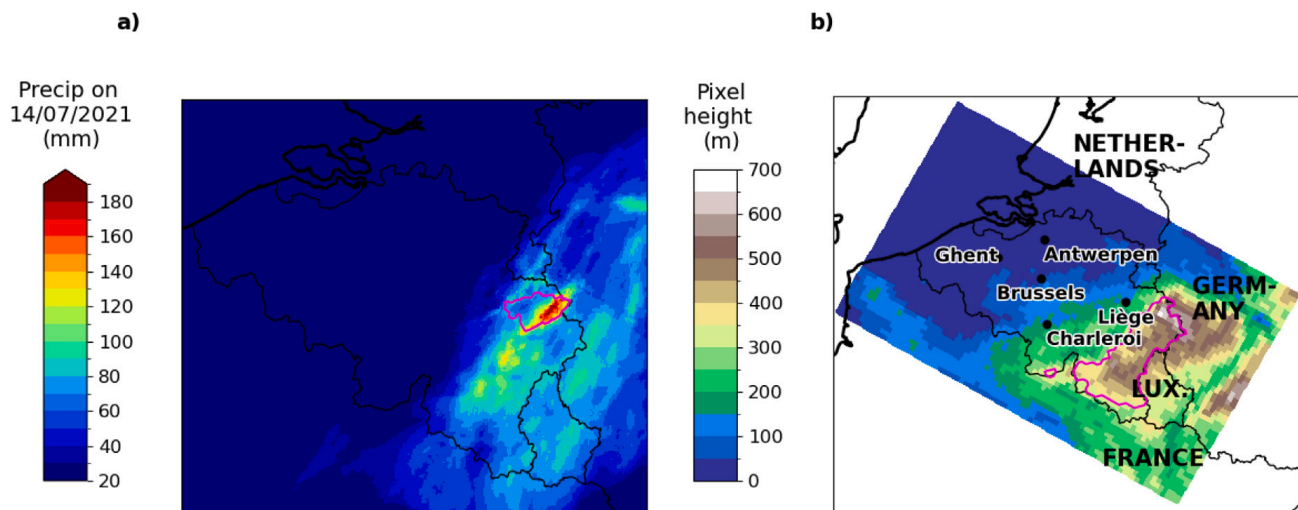


Fig. 1. (a) Precipitation on the 14th of July 2021 as recorded by the radar (Journée et al. 2021). The catchment of the Vesdre river is highlighted in purple. (b) MAR inner domain topography at 5 km. The Ardennian Massif is located in Belgium and highlighted in purple.

Table 1

ESMs used to force MAR at its lateral boundaries (modified after (Sobolowski et al., 2023)). TCR = Transient Climate Response.

CMIP6 ESM	Member	TCR Plausible range (1.2K-2.4K)	Name in this paper
MIROC6	r1ip1f1	1.55	MAR-MIR
NorESM2-MM	r1ip1f1	1.33	MAR-NS2
MPI-ESM-2-HR	r1ip1f1	1.66	MAR-MPI
EC-Earth3-Veg-HR	r1ip1f1	2.62	MAR-EC3
CMCC-CM2-SR5	r1ip1f1	2.09	MAR-CR5
IPSL-CM6A-LR	r1ip1f1	2.32	MAR-IPSL

In July 2021, the Vesdre valley experienced strong floods resulting from extreme precipitation, with major damages to human infrastructures (Dewals et al., 2021) (Fig. 1a). Recent analyses attributed a 400-year return period to this extreme precipitation event (Tradowsky et al., 2023), which was linked to a very singular meteorological situation, with at least three aggravating factors.

At the synoptic scale, the meteorological situation was driven by a depression centered over Germany with an associated North-East flux over Belgium. On top of this, a quasi-stationary rainy occluded front stagnated over the whole Vesdre valley for three consecutive days, constituting a first aggravating factor. A second factor was related to the depression typology: a cut-off low with an associated cold core. That cold air, which was at middle altitudes (about 5 kilometers), enhanced the precipitation production by allowing ascending convective movements into the air mass, hence leading to rapid moist condensation. The depression was also unstable which favored the formation of embedded convective precipitation. A third factor lies in the topography of the Ardennian Massif (Fig. 1b) which “barrier effect” played a key role in forcing the air mass to rise, in turn enhancing condensation and subsequent precipitation (ECMWF, 2021).

In a context of global warming, it is crucial to determine whether such events become more frequent and more intense in the near future. At the global scale, the predicted precipitation change varies spatially but the frequency of more intense events is expected to increase (IPCC, 2023b). The rarer the event the higher the increase in intensity (Gründemann et al., 2022; Pfahl et al., 2017). Hence, the 100-year return period for 1-day rainfall intensity is expected to raise by 13.5 to 38.3% for low and high emissions scenarios, respectively. However, for the Western Central Europe region, while the observed change in heavy precipitation shows an increase, the confidence in the anthropogenic contribution remains low (IPCC, 2023a). At the European scale, for Summer, EURO-CORDEX Regional Climate Models (RCMs) ensemble simulations show a drying of the Mediterranean area while Northern Europe experiences wetter conditions with trends being more pronounced for high emission scenarios (Rajczak and Schär, 2017). For extreme events with return periods ranging from 10 to 100 years, the increase in intensity varies from 15 to 35% in Northwestern Europe by 2070–2099 for the AR5 RCP8.5 scenario.

The impact of global warming on extreme precipitation has been studied in various countries (Babaousmail et al., 2022; Kendon et al., 2023; Luppichini et al., 2023; Arnbjerg-Nielsen, 2012). However, in Belgium, only a few studies have assessed these impacts of climate change (de Vyver et al., 2021; Muñoz Lopez et al., 2023). For instance, Hosseinzadehtalaei et al. (2018) analyzed the outputs of the Euro-Cordex community members to explore the response of Intensity Duration Frequency curves (IDFs) to Global Warming at specific locations. Their results suggest that the 10-year return level of daily rainfall might increase by 50 to 70% under a RCP8.5 scenario for Central Belgium.

Kreienkamp et al. (2021) attributed the July 2021 event by computing trends in extreme precipitation distributions shapes. Their results suggest that the occurrence of the 2021 large scale event has become 1.2 to 9 times more likely due to climate change.

Although the return period is estimated at 400 years based on observations, its frequency is expected to further increase with additional global warming. The findings of [Mohr et al. \(2023\)](#) and [Ludwig et al. \(2023\)](#) showed the July 2021 rainfall event could have even been 7 to 9% more intense per degree of added global warming. Using a storyline approach, they concluded that this event was 11% more intense than it would have been under pre-industrial conditions.

In this paper, we use climate projections from one high-resolution Regional Climate Model (RCM) called MAR (Modèle Atmosphérique Régional) to estimate future trends in extreme precipitation events intensity and frequency across Belgium, and quantify their intensity within the 21st century. First, using the Extreme Value Analysis (EVA) theory, we look at future daily extreme precipitation anomalies over 30-year periods (Section 3.1). Then, the whole simulated period (2015–2100) is used to compute absolute values of extreme precipitation associated to particular return periods, hereafter referred to as return levels (Section 3.2). Finally, we discuss the methodological choices and the major limitations of the conducted analyses (Section 4).

2. Material and method

2.1. The MAR model

Here we use the *Modèle Atmosphérique Régional* (MAR) at high spatial resolution (5-km, [Fig. 1](#)) to locally explore the impact of global warming on extreme precipitation frequency and intensity.

MAR is a hydrostatic regional climate model (RCM) solving primitive equations ([Gallée and Schayes, 1994a](#)), which has been thoroughly evaluated over Belgium for both temperatures ([Doutreloup et al., 2022b,a](#)) and precipitation ([Doutreloup et al., 2019](#)). For precipitation, special attention has been drawn to events linked to floods over the Ourthe catchment ([Wyard et al., 2021](#)) and to more specific meteorological events such as freezing rain [Fettweis et al. \(2017\)](#). MAR's convection is parameterized after ([Bechtold et al., 2001](#); [Doutreloup et al., 2019](#)). Its atmospheric part is described in [Gallée and Schayes \(1994b\)](#) and [Gallée \(1995\)](#). MAR is coupled with the Soil Ice Snow Vegetation Atmosphere Transfer model (SISVAT) which represents the interactions between the surface and the atmosphere, the percolation of water in the soil and surface run-off ([Wyard et al., 2017](#)). Here, we use the latest MAR version 3.14 ([Grailet et al., 2024](#)). MAR domain ([Fig. 1](#)) is rotated for computational efficiency purposes and to cover Belgium, parts of Western Germany, Northern France, the Netherlands and Luxembourg within a same grid.

Over the present climate (1940–2021), MAR was forced by ERA5 reanalyses (MAR-ERA5) and was evaluated using observational data from the Belgian Royal Meteorological Institute (RMI). For climate projections, MAR was forced by six CMIP6 community members ([Table 1](#)) using a single nesting procedure. Control runs (1980–2014) were first compared to MAR outputs and observations over their overlapping period. By doing so, we assessed the performance of MAR outputs, ensuring their ability to be used as input of Extreme Value Analyses (EVA) ([Table 2](#)). The six CMIP6 Earth System Models (ESMs) members ([Table 1](#)) forcing MAR were selected following the recommendations of the Euro-Cordex community ([Sobolowski et al., 2023](#)). The used ESMs have different climate sensitivities as highlighted by their Transient Climate Responses (TCR). The TCR can be defined as the Global Warming Level (GWL) reached by the ESM for a doubling of the initial CO₂ concentration, with the CO₂ being added in the atmosphere at a constant rate ([IPCC, 2023b](#)). Four Shared Socioeconomic Pathways baseline scenarios (SSP) are used ranging from low-end (SSP1-2.6) to high-end emissions (SSP5-8.5) (see [Fig. 3](#) for the Global Mean Surface Temperature (GMST) anomaly reached under specific scenarios for all the models).

2.2. Statistical methodology

Here, we use two common approaches in EVA, namely the Annual Maxima (AM) and Peaks Over Threshold (POT) to assess the robustness of our methodology. These two approaches fit stationary laws (GEV and GPD respectively) to the studied samples, i.e., where parameters do not vary with time. The results of this paper focus on the POT approach since both approaches yield close results (see discussion [Section 4.1.1](#)).

2.2.1. Peaks Over Threshold (POT) approach

The Peaks Over Threshold (POT) approach fits a General Pareto Distribution (GPD) to threshold exceedances. Selecting values exceeding a given precipitation threshold u yields a larger sample size with more information about the distribution of high precipitation events. To ensure that all the selected events are statistically independent, clusters must be identified in the sample. A cluster is defined as a group of events (one event or more) separated by a period of time shorter than a given number of days ([Coles, 2001](#)). For this study, it was set to 7 days (one week) which is an arbitrary choice. However, one week seems to be a reasonable value for selecting events that are caused by independent synoptic patterns.

When all the clusters have been identified, only the highest values of the clusters (i.e., exceeding threshold u) are conserved in the sample. Here, we set this threshold to the 99.5 th quantile of the daily rainfall (including dry days). This value was chosen because it fulfills the requirement of having an average value of selected events per year $\lambda > 2$. The latter requirement confers an advantage to the POT over the AM method, which is based on one selected event per year ([Grandry et al., 2020](#); [Cunnane, 1979](#); [Taesombut et al., 1978](#)). The impact of the threshold choice is further discussed in [Section 4.1.2](#). The GPD has two parameters σ and ξ which are respectively the scale parameter and the shape parameter. The σ parameter affects the spread of the distribution and ξ influences the weight of the distribution tail ([Supplementary Figure S1a](#) displays the impacts of those parameters on the GPD

Table 2

Statistical parameters computed spatially over Belgium over the control period 1981–2010 : Mean value, standard deviation (std), mean bias (MB), Random Mean Square Error (RMSE) and determination coefficient (r^2) for the 2-year and 20-year return levels (duration = 24 h) over Belgium (spatial statistics). Units for mean, std, MB and RMSE are mm.hour⁻¹. MB and RMSE are computed using the difference between the observed return levels and the simulated ones for all the pixels within Belgium.

models	2-year return level						20-year return level					
	Mean	std	MB	MB (%)	RMSE	r^2	Mean	std	MB	MB (%)	RMSE	r^2
Obs	1.7	0.2	0.0	0.0	0.00	1.00	2.9	0.4	0.0	0.0	0.00	1.00
MAR-ERA5	1.2	0.2	-0.5	-30.5	0.54	0.74	2.0	0.3	-0.8	-29.7	0.92	0.48
MAR-MPI	1.1	0.2	-0.6	-36.0	0.63	0.68	1.7	0.2	-1.1	-40.2	1.22	0.09
MAR-EC3	1.1	0.2	-0.6	-34.9	0.61	0.79	1.8	0.3	-1.1	-38.7	1.20	0.04
MAR-MIR	1.2	0.2	-0.5	-27.6	0.51	0.61	2.1	0.3	-0.7	-25.9	0.86	0.26
MAR-NS2	1.3	0.2	-0.4	-25.9	0.47	0.70	2.0	0.3	-0.8	-29.7	0.94	0.20
MAR-CR5	1.4	0.2	-0.4	-20.9	0.39	0.69	2.2	0.3	-0.6	-22.5	0.74	0.44
MAR-IPSL	1.1	0.2	-0.6	-37.7	0.66	0.68	1.7	0.3	-1.1	-39.0	1.22	0.03

Probability Density Function (PDF)). After the fitting, the return level associated to a return period of T years R_T can be obtained following (Coles, 2001):

$$R_T = \begin{cases} u + \frac{\sigma}{\xi}(\lambda T)^{\xi-1} & , \text{ for } \xi \neq 0 \\ u + \sigma \log(\lambda T) & , \text{ for } \xi = 0 \end{cases} \quad (1)$$

Here u and λ represent, respectively, the chosen threshold value and the mean number of clusters exceeding the threshold each year.

The fitting of the EVA statistical laws are performed by Maximum Likelihood Estimation (MLE) using the R package *ismev* (Gilleland, 2003).

2.2.2. Representation of extremes in MAR

Over Belgium, using observational data from the control period 1981–2010, the GPD-derived 24-hour 2 and 10-year return levels reach intensities of up to 1.7 and 2.8 mm.hour⁻¹ (Table 2). The mean bias of MAR-ERA5 for different return levels is on average around -30%. However, when MAR is forced by CMIP6 ESMs, the mean bias increases (about 35% for MAR-EC3, MAR-MPI and MAR-IPSL) or decreases in some cases (around 21% for MAR-CR5). The ability of MAR to correctly represent extreme precipitation events seems to rely on the forcing ESMs ability to create synoptic extreme weather patterns at the domain boundaries. Since the occurrence of such patterns is partly stochastic and depends on the internal variability of the ESMs over the study area (Jain et al., 2023; Kendon et al., 2023), for higher return levels (e.g. 20 years), spatial correlation coefficients (r^2) can drop to 0.03 (MAR-EC3).

Those biases illustrate the need for bias-adjustment of MAR outputs. Here, the observations used for MAR outputs correction consist in gridded daily precipitation data from the Royal Meteorological Institute (RMI) covering the period 1950–2021 at a resolution of 5 km (Journée et al., 2023).

The bias-adjustment procedure used to correct the sampled values is based on the distributions of the POT (i.e., $u = 99.5$ th quantile of the daily precipitation time-series) over the period 1981–2010. Bias-adjustment consists in matching the simulated quantiles to the observed ones. This technique is referred to as Equidistant quantile matching (EDCDFm) (Pierce et al., 2015).

Projected precipitation values V_{sim} are corrected using a multiplicative change factor CF . To calculate this factor, the position of V_{sim} in the simulated Cumulative Density Function (CDF) must be retrieved. This requires to identify the two empirical quantiles $q_{sim,i}$ and $q_{sim,i+1}$ with $q_{sim,i} < V_{sim} < q_{sim,i+1}$. Then, CF is computed using the observed quantiles of the control period (1981–2010) $q_{obs,i}$ and $q_{obs,i+1}$, and the quantiles of the simulated daily precipitation over the control period $q_{cont,i}$ and $q_{cont,i+1}$:

$$CF = \left[\frac{q_{obs,i}}{q_{cont,i}} * W_i + \frac{q_{obs,i+1}}{q_{cont,i+1}} * W_{i+1} \right] - 1 \quad (2)$$

where W_i and W_{i+1} represent weights which are computed as follows:

$$W_i = 1 - \frac{V_{sim} - q_{sim,i}}{q_{sim,i+1} - q_{sim,i}} > 0 \quad (3)$$

CF can reach 50% on the coast of Belgium where MAR struggles at representing extreme precipitation. Over regions with steeper topographic gradients, its performance increases with lower CF values down to 15% (Fig. 2).

To avoid abrupt changes over short distances in the return level maps, which might not be representative of the reality, all pixel values shown in maps hereafter have been smoothed using a 3 by 3 averaging kernel, which is an arbitrary choice.

3. Results

3.1. Linking extreme precipitation anomalies to global warming (working on 30-year periods)

The distributions of extreme events are overall shifting towards more extreme values. Firstly, the modes of the extremes are increasing, as shown by the rise of the threshold u (99.5th quantile) in the POT approach (Fig. 3a). This parameter displays an

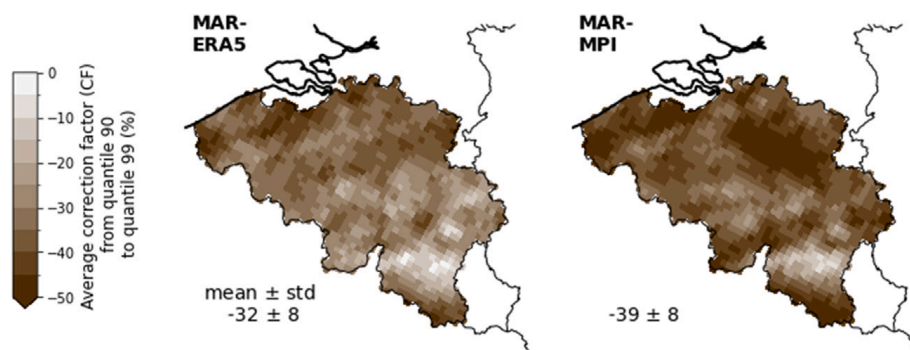


Fig. 2. Average Correction Factor (CF) from quantile 90 to quantile 99 of the POT (Peaks over Threshold) for two selected models ((a) MAR-ERA5 and (b) MAR-MPI). Other models are displayed in Supplementary Figure S3.

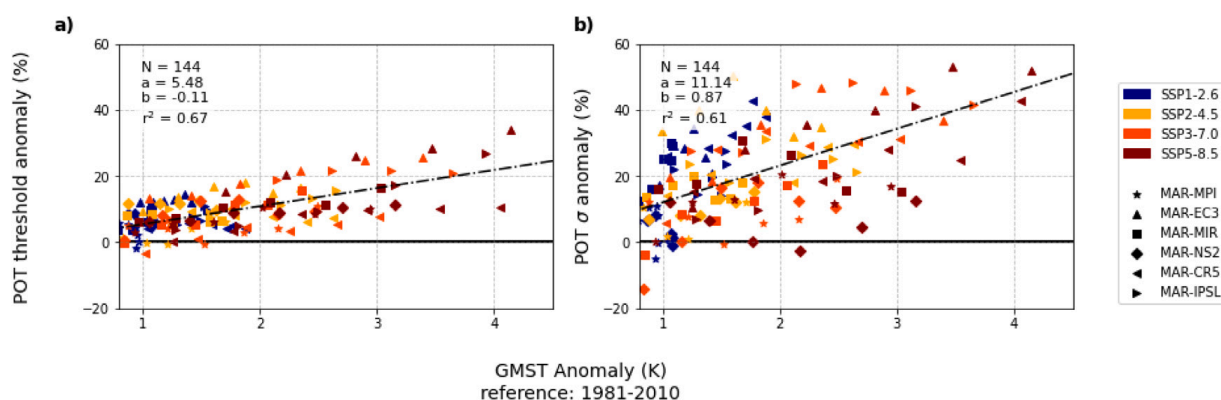


Fig. 3. Anomalies relative to 1981–2010 averaged over 30-year moving windows (from 2021–2050 to 2071–2100 with a 10-year time-step) for (a) POT threshold, (b) GPD-derived σ . A linear regression is applied to the data clouds (black dotted line of slope a and intercept b). The impact of those parameters on the PDFs is displayed in Supplementary Material Fig. S1. The GMST anomaly reached for each model is plotted on Supplementary Material Fig. S2.

increases of 5.5K^{-1} of additional global warming respectively. This trend value is obtained by fitting a linear regression on the spatially averaged future anomalies over 30-year periods (slope a on Fig. 3a). Those periods are obtained by moving a 30-year window from 2021–2050 to 2071–2090. This leads to point clouds of 6 (30-year periods) $\times$ 4 (scenarios) $\times$ 6 (models) = 144 points. Second, the spread of the extremes becomes larger. This is clearly shown by the increase of the σ parameter, with a trends reaching 11K^{-1} (Fig. 3b).

This results in an average increase of 7K^{-1} of additional global warming for various return levels, in line with the Clausius-Clapeyron scaling (CC-scaling) (Fig. 4a), i.e., a 7% raise in atmospheric moisture for each K of warming (Trenberth et al., 2003). However, we find that, despite having an overall averaged increasing trend of 7K^{-1} irrespective of the return period, the correlation to global warming is greater for smaller return periods (Fig. 4a). Hence, we assume that this fluctuation stems from the internal variability of the ESMs. Indeed, due to the natural variability within the Earth system, the periods with the most extreme and rarest events might not be the warmest ones (Kendon et al., 2023; Jain et al., 2023) and thus not the latest ones. For some ESMs under a particular scenario, most extreme events (above 150 mm per day after bias-adjustment) occur before the 2050s while for others, they occur by the end of the century. This is why for higher return periods e.g., 20 years, more spread is found in the data cloud of extreme intensity anomaly as a function of additional global warming than for shorter return periods, with lower associated determination coefficient values. Hence, the timing of most extreme and rarest precipitation events remains stochastic.

To assess the trends at the local scale, we also compute a linear regression for each grid cell within Belgium (Fig. 4b). For higher return periods, some spatial variability appears with more pronounced trends for some areas of the country. Hence, the Southern flank of the Ardennian Massif as well as the northern parts of the countries show trends exceeding the Clausius-Clapeyron scaling, i.e., reaching up to 15K^{-1} . Those areas also show the largest frequency trends (Fig. 5b). The combination of higher atmospheric moisture and increased frequency may lead to trends exceeding 7K^{-1} , also called Super Clausius-Clapeyron scaling. However, the mean value in Belgium remains close to 7K^{-1} for the studied return levels.

Besides, concerning the trends in frequencies, we find that, the rarer the event, the higher the change in frequency (Fig. 5a). When averaged over Belgium, we estimate that for each extra degree of global warming, the 20-year return level computed over the 1981–2010 period becomes +0.6 times more frequent. This means that the 20-year return level becomes 1.6 times more frequent in a 1 degree warmer world, and 2.2 times more frequent in a 2-degree warmer world. Nevertheless, the spread of frequency anomalies increases with longer return periods, the result of the internal variability in the forcing ESMs (Fig. 5a). As for the intensities, this

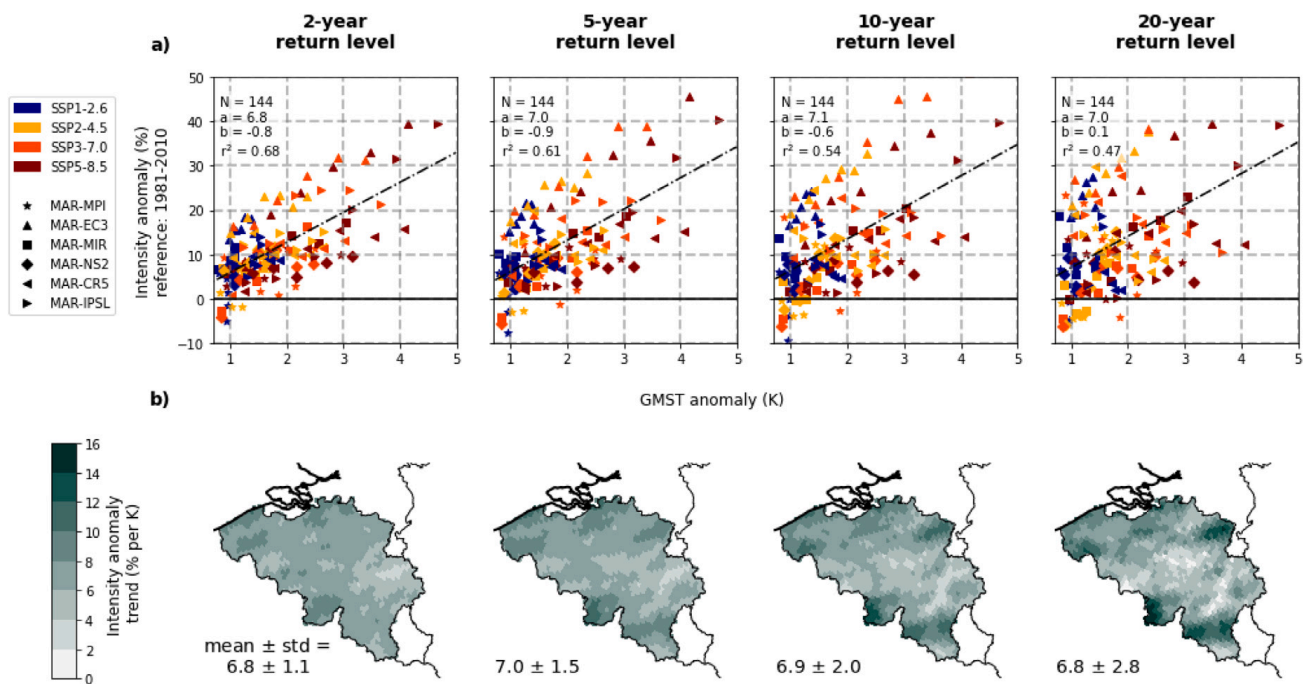


Fig. 4. (a) Intensity anomalies over 30-year moving windows relative to 1981–2010 for different return levels using the POT approach. A linear regression is applied to the data clouds (black dotted line of slope a and intercept b). (b) Same regression as in (a) but computed at each pixel location. The written values represent the map average $\pm$ the standard deviation. The results using the AM approach are very close and displayed in Supplementary Figure S4. The trends for each model are plotted on Supplementary Figure S6.

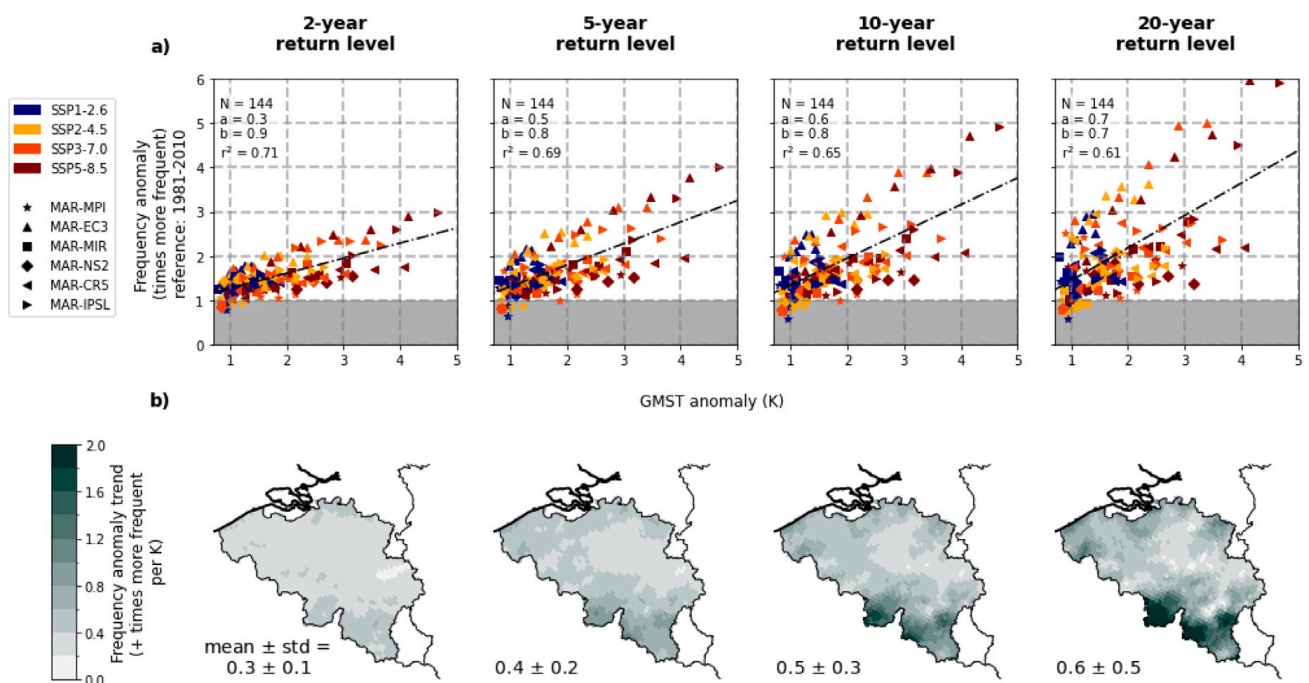


Fig. 5. (a) Frequency anomalies over 30-year moving windows relative to 1981–2010 for different return levels using the POT approach. A linear regression is applied to the data clouds (black dotted line of slope a and intercept b). (b) Same regression as in (a) but computed at each pixel location. The written values represent the map average $\pm$ the standard deviation.

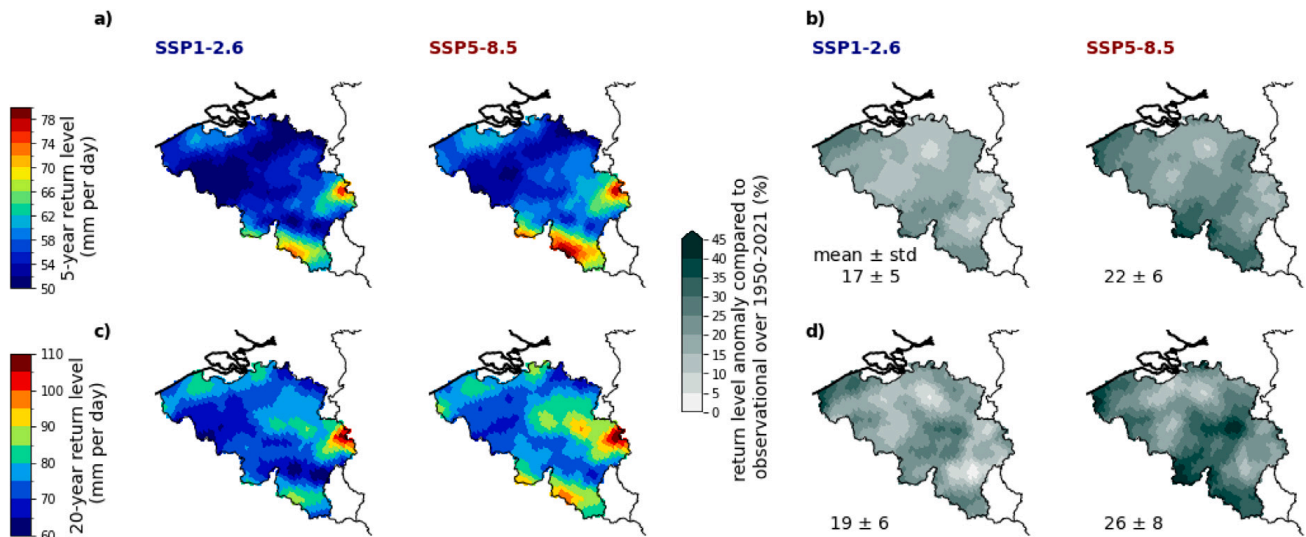


Fig. 6. (a) 5-year return level computed over 2015–2100 under SSP1-2.6 and SSP5-8.5 scenarios. (b) 5-year return anomaly as compared to 1951–2021. Same for the 20-year return level in (c) and (d). Return levels are calculated as the mean return levels of the six models (Table 1). The written values represent the map average $\pm$ the standard deviation. Widths of the 95% confidence intervals are shown in Supplementary Figure S7.

value of +0.6 times per K appears to be larger for the same specific areas, where it can easily reach + 1 times per K for the 20-year return level. This means that, in a 1-degree warmer world, the 20-year return level becomes twice as likely relative to present day.

3.2. Return levels over the whole simulated period (2015–2100)

As mentioned above, because of the internal variability of the forcing ESMs, the timing of extreme events is challenging to predict. This is why, trends are less clear for higher return periods (Figs. 4 and 5). To eliminate the effect of the internal variability, we apply an EVA to the whole future projection (2015–2100). By proceeding this way, all the simulated extreme events are accounted for in the analysis and we can estimate the absolute return level values expected within the 21st century.

When we average the obtained return level maps for the six models, and the SSP1-2.6 and 5–8.5 scenarios using the POT approach (Fig. 6), the spatial distribution of modeled extremes is equivalent to the observed one over 1951–2021. We identify two main hot-spots located on heights in the vicinity of Liège (see Fig. 1b for the location) and on the Southern flank of the Ardennian massif (Fig. 6).

Comparing the return level maps (Fig. 6) from recent observations with our future projections, the expected shift seen earlier in the anomaly analysis appears once again. On average the 20-year return level increases by 19 ± 6 and $26 \pm 8\%$ leading to return level values exceeding 100 mm per day even under a SSP1-2.6 on the precipitation hot-spot nearby the German border. This increase appears also for smaller return periods but is less pronounced, with relative increases of 17 ± 5 and $22 \pm 6\%$ for the two same scenarios. However, intensity increases are more pronounced on both Southern and Northern Flanks of The Ardennian Massif, i.e., where topographic gradients are higher. There, the increase in the 20-year return level exceeds 35% under SSP5-8.5. Low-elevated areas on the Sea-side experience similar increases.

Those more pronounced intensity shifts are linked to particularly high increases in frequency. Hence, for those areas which show the highest intensity increases, the new return period of the observational 20-year return level ranges from 4 to 7 years, even under low-emission scenarios. For the areas where the intensity increase is close to the Belgian mean, the new return period of the historical 20-year return level is of 12–15 and 8–12 years, respectively for SSP1-2.6 and SSP5-8.5 scenarios.

Fig. 6 shows maps of return levels averaged between the 6 models (Table 1). However, from one model to the other, differences can be large (Fig. 7). MAR-IPSL and MAR-EC3 display a scenario-linked logic. Hence, for those two models, the more pessimistic the emission scenario, the more extreme the return level map. Those two models also show the highest TCRs of all the ESMs forcing MAR. For the other models, the highest emission scenario does not systematically lead to the most extreme return level maps. MAR-MIR especially corroborates this statement by displaying the most extreme return level maps for the two ‘opposite’ scenarios SSP1-2.6 and SSP5-8.5 (Fig. 7), while the two intermediate scenarios display less extreme events. This implies that even under low-emission scenarios, intense extreme events will occur as well. Hence in MAR outputs, precipitation events exceeding 150 mm per day are found in all scenarios.

In terms of absolute values, MAR-EC3 yields the most intense return level-maps where the 20-year return level can reach 120 mm per day, while it only reaches 105 mm per day on the same hot-spots in MAR-NS2. However, even if MAR-NS2 return-level maps are less intense, the shift relative to observations from 1951 to 2021 remains clear, with values reaching 100 mm per day for the 20-year return level on hot-spots for all the models.

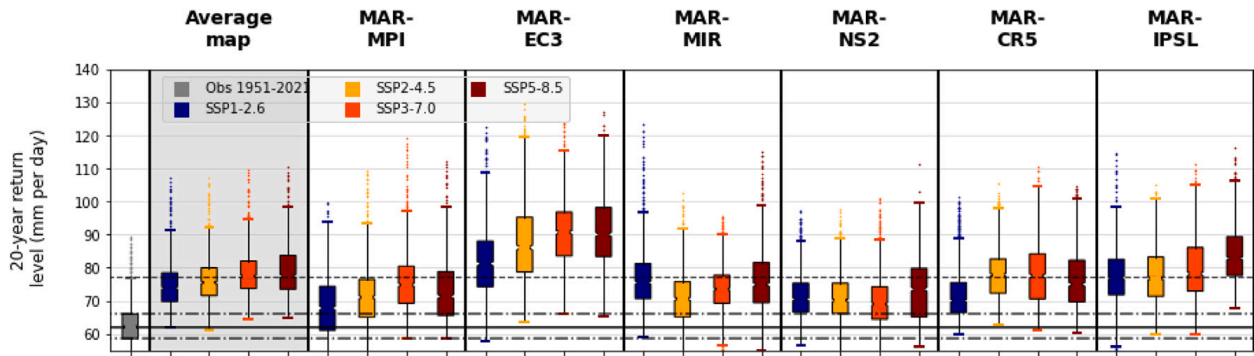


Fig. 7. Boxplots associated to the return level maps of Belgium for a return period of 20 years using all the models (Table 1) over 2015–2100. The POT approach is used for the computation of the return levels. The gray box refers to the return level map computed using observational data over 1951–2021 (Journée et al., 2023). Visual marks (horizontal black lines) have been added so each box can be compared to the observational box.

4. Discussion

4.1. Impact of the methodological choices over the EVA

Concerning the EVA, several methodological choices which might impact computed return levels are done. Hereafter, we first discuss the impact of the POT approach. Second, we look into the stationary assumption when dealing with the whole simulated period (2015–2100). Third, we discuss the bias-adjustment strategy.

4.1.1. POT approach

In this study, we also tried another common approach based on the daily precipitation annual maxima (AM), on which a Generalized Extreme Value (GEV) distribution can be fitted (Coles, 2001). This was achieved to estimate uncertainties in EVA. Concerning future anomalies in the intensity and frequency of extreme precipitation (Section 3.1.), the AM and POT approaches yield very close results (Figs. 4 and 5, and Supplementary Figure S4). This shows that our statistical methodology does not alter the effect of climate change on the intensity and frequency of extreme precipitation.

For the computation of return level absolute values over the whole simulated period (Section 3.2), AM and POT approaches show almost identical results (Fig. 8), highlighting the robustness of this study. However, in the POT approach, results would have been slightly different with lower threshold values (i.e., $u < 99.5$ th quantile). The threshold selection had to meet two main objectives. First, the chosen threshold needed to allow the POT distributions to be tailed enough to prevent underestimating higher return level values. Second, it also needed to preserve the advantage of the POT approach over the AM approach for shorter return periods, i.e., > 2 selected events per year. Hence, for smaller return periods (2 to 5 years), the POT approach predicts more intense return level values. The latter approach is more reliable since it selects more events, hence providing more information about return level values associated with low return periods.

To meet these 2 conditions, several threshold values were tested based on the observational data covering 1950–2021. The 99.5th quantile appeared to fulfill those two conditions the best, with on average over Belgium, 2.6 POT selected per year. With a lower threshold value, the risk would have been to slightly underestimate the 20-year return level intensity as compared to the AM approach which naturally leads to tailed distributions since it selects fewer events by definition (Supplementary Fig. S5).

4.1.2. Non-stationarity of the EVA

In Section 3.1, we identified trends in extreme precipitation intensity anomalies estimated using future 30-year periods. The stationary assumption usually does not hold in the presence of such trends, i.e., probability of extreme events occurrence varies in time. Here, we use a stationary approach to estimate return level values over 2015–2100 for two main reasons.

First, extreme precipitation intensity anomalies exhibit a linear trend of 7% per K, consistent with the Clausius-Clapeyron relationship, but only when data from all models are combined (Fig. 4). It is widely recognized that a multi-model ensemble mean generally provides a better estimate than each individual ensemble member (Gleckler et al., 2008; Tebaldi et al., 2011; Reichler and Kim, 2008). Hence, because of the internal variability within each ESM, trends can totally disappear when each model is considered individually, especially for higher return periods (Fig. S6).

Second, non-stationary laws were also tested for completeness, but these led to overall similar results to those of the stationary approach (Fig. 8). To incorporate non-stationarity in the POT approach, first, we make the σ of the GPD vary with the GMST anomaly. Moreover, the evolution of λ with time (Eq. (1)) can also be modeled by fitting a Poisson regression model to the distribution of the number of threshold exceedances per year (Grandry et al., 2020). By proceeding so, we model the potential non-stationarity in the number of times per year where the threshold value is exceeded. In the literature, such models are referred to as Poisson-GP models (Tramblay et al., 2013; Grandry et al., 2020). Non-stationarity can also be implemented using the AM approach by making the GEV parameters vary with the GMST anomaly. In our case, we tried a GEV model where μ and σ can vary linearly with the reached global warming level (Fig. S1).

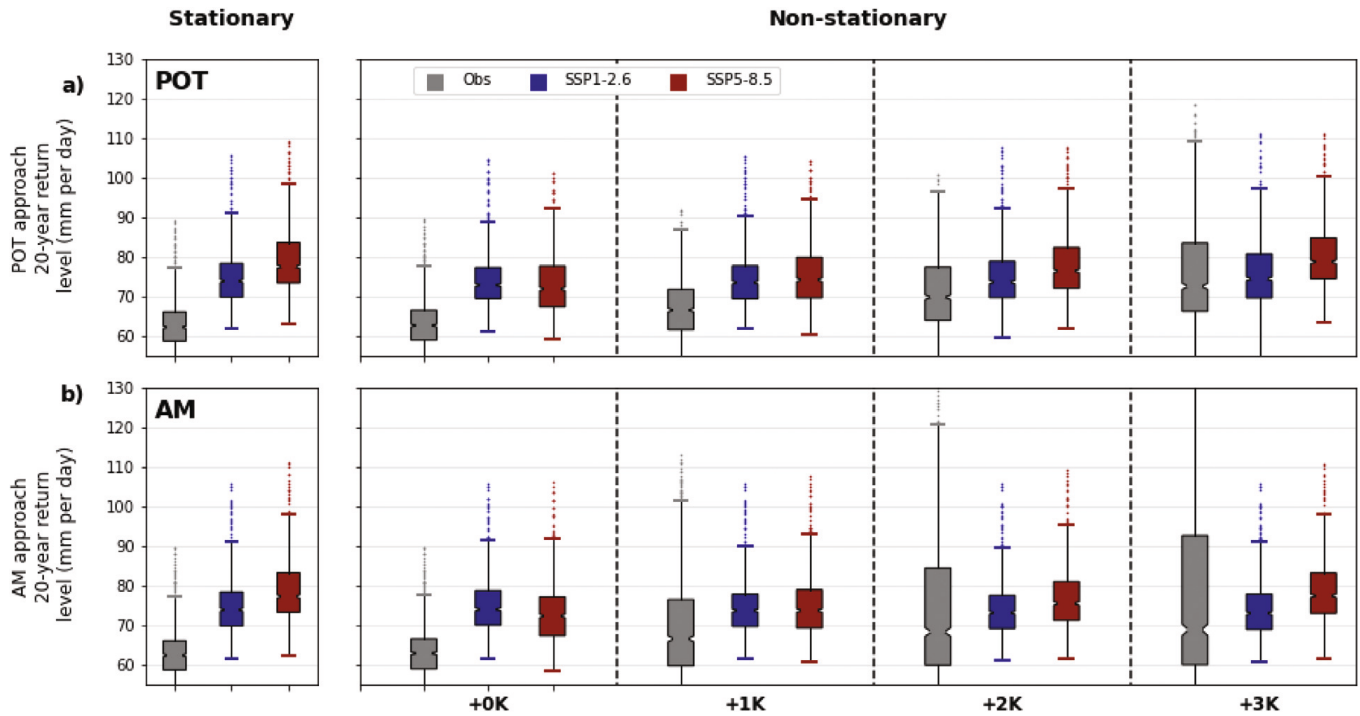


Fig. 8. Distributions of 20-year return levels using stationary and non-stationary approaches for both (a) POT and (b) AM approaches over 2015–2100. The probability of encountering extreme precipitation is time-evolving in the non-stationary approach while it is fixed in the stationary case. Return levels based on model outputs are computed as the mean of the six models (Table 1). The results of the log-ratio test, which checks whether the non-stationary approach performs better, are plotted on Supplementary Figure S8.

Non-stationary laws (GEV_n /Poisson-GP) were applied only if their performance was significantly better as compared to stationary models using the log-ratio test with a 95% confidence level (Coles, 2001).

Using non-stationary models, the difference between the 20-year return level maps under low-emission and high-emission scenario is negligible at GMST anomaly=0K, but becomes more pronounced for higher GMST anomaly values (Fig. 8). Interestingly, using observations and non-stationary approaches to extrapolate future return level values, return levels reach or even exceed the values obtained using bias-adjusted model data in some areas. This occurs when the GMST anomaly reaches 1K, specifically when using the AM approach.

However, we find that, overall, the stationary approach performs better than the non-stationary ones across most of Belgium, based on a log-ratio test (Fig S8), hence supporting our decision to base our analyses on stationary laws. We observe no significant geographical or scenario-linked patterns where and when non-stationary models perform better, and conclude that this is purely model dependent (Fig. S8). Slater et al. (2021) and Serinaldi and Kilsby (2015) have previously shown that estimating non-stationary models for precipitation comes with large uncertainties due to a small signal-to-noise ratio. Here, only one model successfully timed the most extreme precipitation events coincident with global warming, namely MAR-EC3, while for other models, predicting the timing of extreme events remains challenging.

For these reasons, fitting non-stationary laws to the data is challenging, and the stationary approach was therefore preferred in our analyses. Furthermore, the stationary approach allows us to compare future with historical return levels, since the latter were also estimated in a stationary fashion. In Section 3.2, we show average return-level maps under specific scenarios over the whole transient run. Hundhausen et al. (2024) recently suggested that proceeding this way is more appropriate due to the high internal variability within each forcing ESM. In our case, the shift in the 20-year return level values as compared to the observational ones is clear, irrespective of the model or scenario (Fig. 7). This result remains robust, particularly considering that the return periods analyzed are significantly shorter than the 85-year length of the future transient runs. However, given the variability between models, it cannot be excluded that return levels may exceed the values presented in average return level maps, especially for high-warming scenarios.

4.1.3. Bias-adjustment of extremes

To assess the potential impact of our bias-adjustment procedure on return level estimates, we computed the return levels using the POT approach over bias-adjusted MAR-ERA5 outputs (i.e., same bias-adjustment technique as presented in the paper). The return level maps are similar to the ones obtained using observational data over 1951–2021 (Fig. 9). Nevertheless, the interpolation procedure used to create the gridded observations smooths the precipitation fields, which means that associated return levels might be underestimated in some regions.

Another important result is that our bias-adjustment does not affect the impact of atmospheric warming on extremes precipitation. Hence, Figs. 3, 4 and 5 were also shown using non bias-adjusted MAR outputs, with no significant change (Supplementary Fig. S9).

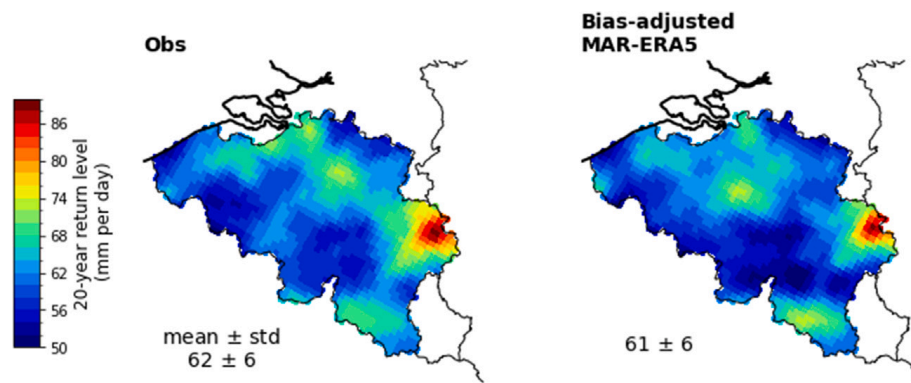


Fig. 9. 20-year return level computed using Observational data and bias-adjusted MAR-ERA5 data over 1951–2021. The POT approach is used here for the computation of the return levels. The written values represent the map average $\pm$ the standard deviation.

4.2. Implication of the results

In our results, we show that the average trend for extreme intensity in Belgium follows the Clausius-Clapeyron scaling of $7\% \cdot K^{-1}$. This increase is linked to a clear shift towards higher values of the distributions modes (μ of the GPD under the POT approach) and spreads (σ parameter) (Fig. 3). In areas showing higher topographic gradients in the South, and on the sea-side, for higher return levels, the intensity and frequency trends appear to be up to twice larger than the average over Belgium (Figs. 4 and 5). This suggests that even at small spatial scales, the impact of global warming on both the intensity and frequency of extreme precipitation might vary significantly, although statistical noise may hamper strong conclusions at such scales. This result is in line with Ludwig et al. (2023) and Hundhausen et al. (2024) who found extreme precipitation intensity increases following the Clausius-Clapeyron scaling or even beyond it in Western Germany.

Hot-spots of extreme precipitation have been clearly identified in our analysis. Sensitive areas appear in the vicinity of the city of Liège (Vesdre watershed) on the highest parts of the country, which were highly impacted by the July 2021 event (Fig. 1). There, 20-year return levels can reach 110 mm per day irrespective of the scenario. Additional hot-spots appear in the future (Fig. 6). The southern flank of the Ardennian Massif experiences high return level values and this hot-spot especially stands out when it comes to lower return periods (e.g. 2 or 5 years). This indicates that the associated extreme precipitation distributions at these locations are spread but with lighter tails, i.e., with smaller ξ parameter values than in the above mentioned hot-spot. Yet, some other extreme-prone areas appear also at lower elevations, notably around the city of Antwerpen (Fig. 1). There, extreme precipitation values are necessarily linked to other factors than topography given the flatness of the area. Compared to observational return levels computed over 1951–2021, intensity increases can reach up to 40% (Fig. 7), which is even higher than the projected change over North-Western Europe under high-emission scenarios (20%–35%) (Rajczak and Schär, 2017).

4.3. Limitations of the study

MAR underestimates extremes (Table 2, Fig. 2), especially for low-elevated Belgium where (Doutreloup et al., 2019) previously showed that MAR tends to underestimate convective precipitation, supporting the need for bias-adjustment. For bias-adjustment, we use a probabilistic approach based on distributions (Teutschbein and Seibert, 2012). However, in the coastal area north of the city of Ghent (Fig. 1), the underestimation of extreme quantiles in MAR (Fig. 2) is such that the estimated return levels here might not be realistic. Moreover, only one RCM is used, while Hosseinzadehtalaei et al. (2018) showed that the largest source of uncertainty lies in the RCMs themselves followed by the ESMs forcing at their limits.

Our analyses are conducted on daily basis because RCMs tend to underestimate hourly peaks of extreme precipitation (Pichelli et al., 2021; Ban et al., 2021). Extreme precipitation underestimation may possibly originate from the hydrostatic hypothesis in MAR, but Malardel (2014) showed that with an adequate parametrization of convection, RCMs perform well at spatial resolutions > 3 km. Moreover, MAR is run using a single nesting procedure over a small domain. Due to the small area of the domain and the relatively coarse resolution of the forcing ESMs (between 40 and 130 km), the underestimation of extreme precipitation might result from the fact that realistic convective cells do not develop (Brisson et al., 2016). However, Wyard et al. (2018), showed that the different resolutions of forcing fields do not have a major impact on RCMs outputs such as the simulated cloud cover in Belgium. Future sensitivity analyses should investigate the impact of spatial resolution on precipitation in Belgium.

Finally, the impact of climate change can largely vary depending on the future period and model selected (Figs. 4a and 7). Future work should assess the ability of the forcing ESMs to model extreme weather patterns over Western Europe. For instance, examining extreme weather types at a synoptic scale in CMIP6 ESMs outputs might be an option. Model differences may not only be explained by their different TCRs, but also by their ability to simulate the frequency of Cut-off lows and other variables as well. Using analogues, Thompson et al. (2024) suggested that the frequency and persistence of synoptic weather cut-off lows, similar to July 2021, show an increasing trend in Northern-Western Europe. Insua-Costa et al. (2022) showed that the July 2021 event itself was particularly intense due to the huge moisture content in the atmosphere caused by massive prior evapotranspiration from

Western European forests. Using MAR outputs and ESM data, we could identify which weather patterns lead to the most intense precipitation events in Belgium. Building on this, we could test whether the ESMs used in this study predict coherent extreme weather frequencies.

5. Conclusion

In this study, we estimate projected trends in extreme precipitation frequency and intensity over Belgium. The Regional Climate Model MAR forced by six CMIP6 ESMs at 5 km resolution shows that the intensity of future extreme precipitation will, on average over the country, increase following the Clausius Clapeyron scaling of 7% per added degree of global warming. However, MAR projects larger trends for frequency anomalies over specific areas, suggesting that intensity increase may locally exceed the Clausius Clapeyron scaling, i.e. super Clausius Clapeyron scaling, notably in high topographic gradient areas and in low-elevated coastal areas.

When bias-adjusted, MAR outputs show return level values associated to a 20-year return period reaching 100 mm per day or more in extreme precipitation hot-spots throughout the 21st century. Projected intensity increases can reach up to 40% under high-emission scenarios.

Our study identifies Belgian areas that are at risk of experiencing more frequent extreme precipitation events, notably the Ardennian Massif and low-elevated coastal areas. We do not cover hydrological extremes, such as floods, for which impact studies might be conducted using bias-adjusted MAR outputs and hydrological models. Future work should identify climate drivers that enhance the strength and frequency of extreme events.

CRedit authorship contribution statement

Josip Brajkovic: Writing – original draft, Methodology, Formal analysis, Conceptualization. **Xavier Fettweis:** Writing – original draft, Supervision, Data curation. **Brice Noël:** Writing – original draft. **Hans Van De Vyver:** Supervision. **Nicolas Ghilain:** Writing – original draft, Supervision, Data curation. **Pierre Archambeau:** Writing – original draft, Supervision. **Michel Piroton:** Supervision. **Sébastien Doutreloup:** Writing – original draft, Supervision.

Credits

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Declaration of competing interest

The authors declare that they have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.

Appendix A. Supplementary data

Supplementary material related to this article can be found online at <https://doi.org/10.1016/j.ejrh.2025.102399>.

Data availability

Data will be made available on request.

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APPENDIX **B**

First paper: supplementary material

Supplementary material

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March 2024

1 Supplementary material

The supplementary material includes 9 Supplementary Figures

References

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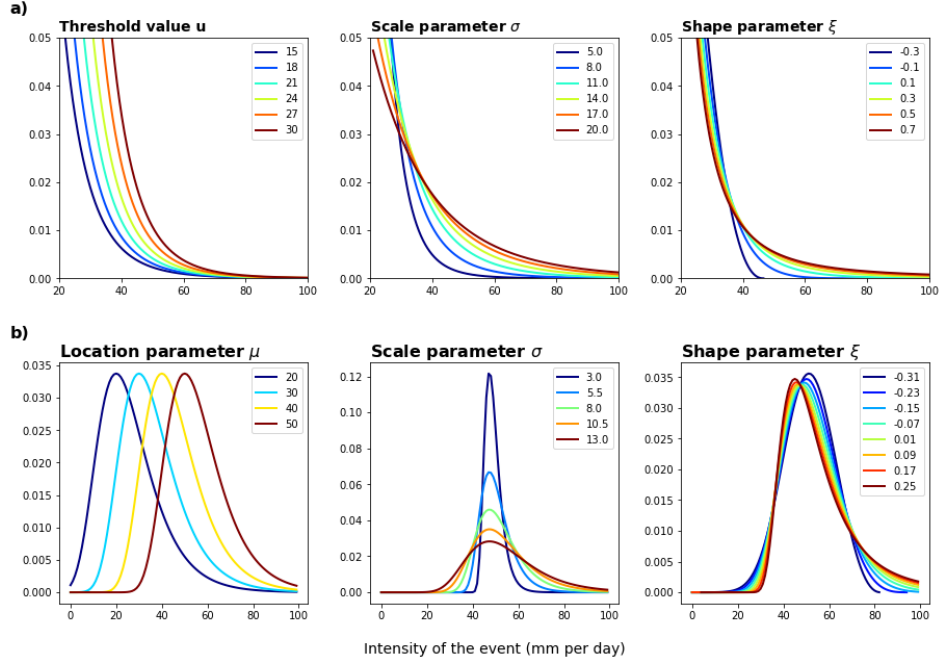


Figure S 1: (a) Impact of the GPD parameters on its PDF and (b) same for the GEV. GEV return levels associated to a return period T are obtained following (Coles, 2001): $R_T = \mu - \frac{\sigma}{\xi} [1 - \{-\log(1 - p)\}^{-\xi}]$

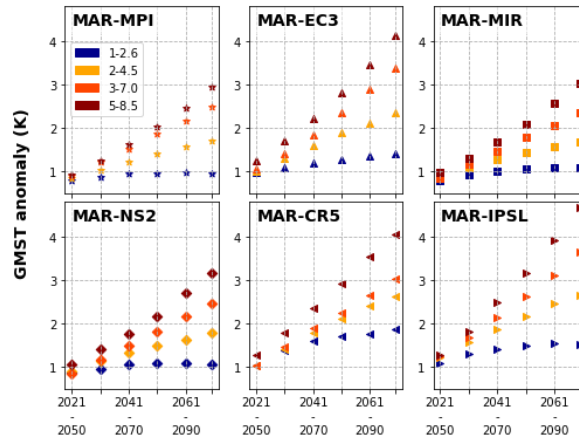


Figure S 2: GMST anomaly reached by the studied models (reference = 1981-2010)

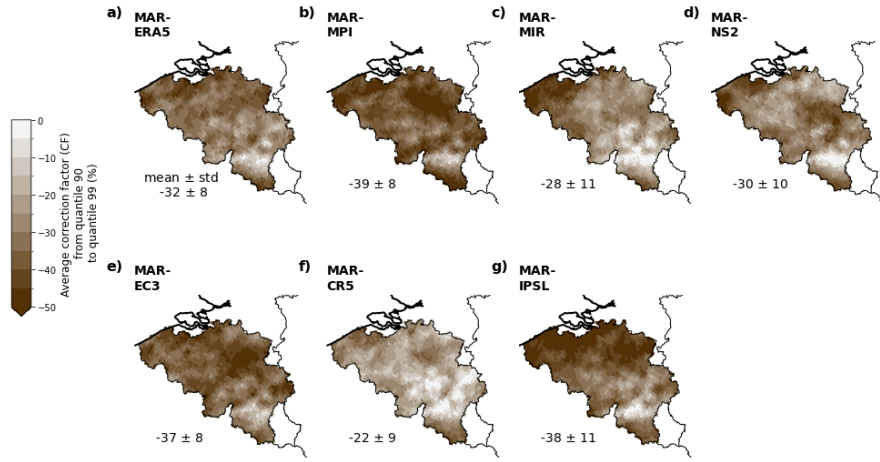


Figure S 3: Averaged Correction Factor (CF) between quantiles 90 and 99 for the distribution of independent POT (threshold being the 99.5th quantile of the precipitation time-series) for (a) MAR-ERA5, (b) MAR-MPI, (c) MAR-MIR, (d) MAR-NS2, (e) MAR-EC3, (f) MAR-CR5 and (g) MAR-IPSL. The written values represent the map average $\pm$ the standard deviation.

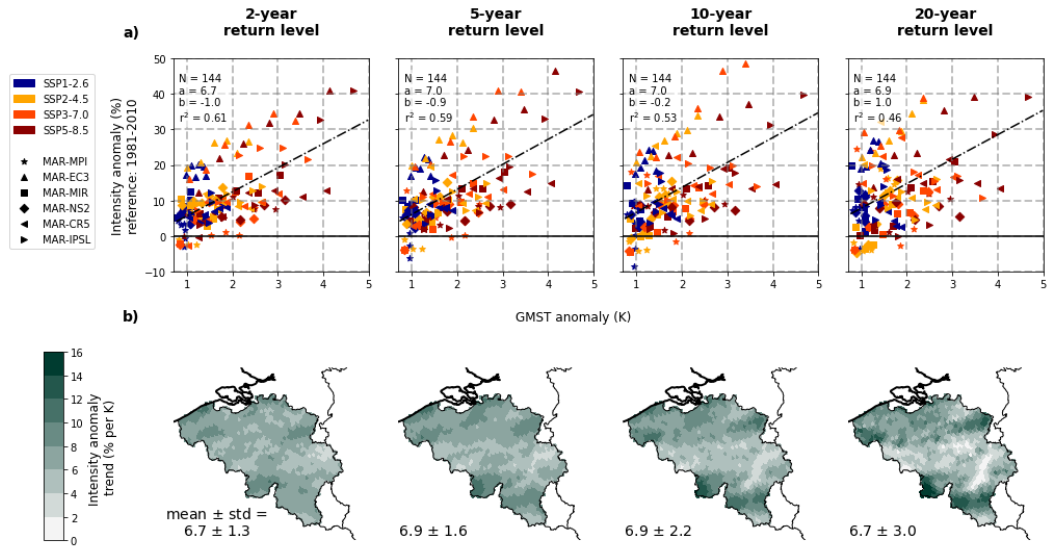


Figure S 4: (a) Intensity anomalies over 30-year moving windows relative to 1981-2010 for different return levels using the AM approach. A linear regression is applied to the data clouds (black dotted line of slope a and intercept b). (b) Same regression as in (a) but computed at each pixel location. The written values represent the map average $\pm$ the standard deviation.

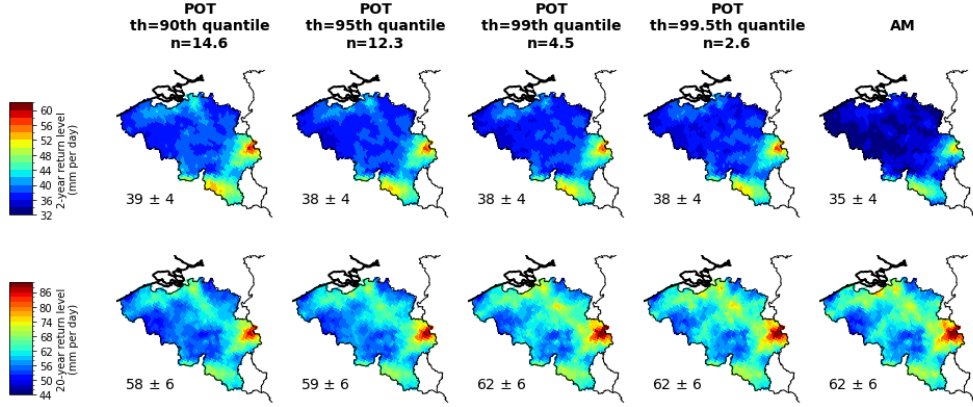


Figure S 5: Return level maps drawn using observation gridded data over 1951-2021 for various return periods (rows). The columns represent different methods. For the GPD several thresholds were tried (from the 90th quantile to the 99.5th quantile). Depending on the used threshold in the POT approach, The mean number n of Peaks Over Threshold selected per year varies. The written values represent the map average $\pm$ the standard deviation.

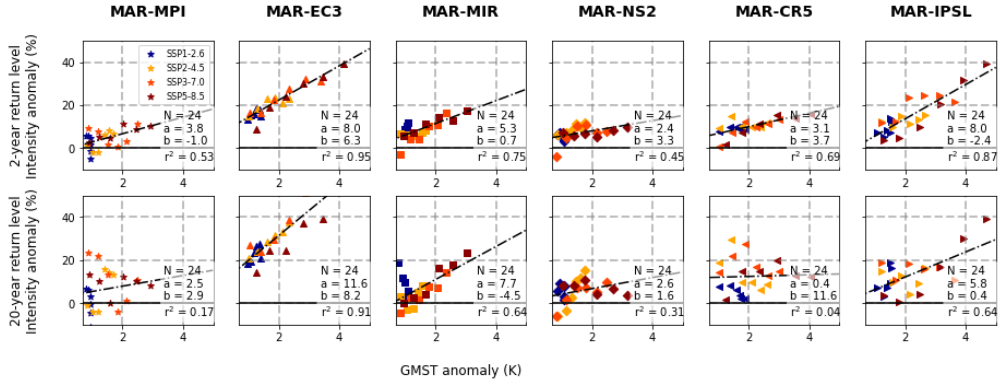


Figure S 6: 2-year and 20-year return level intensity anomalies relative to 1981-2010 averaged over 30-year moving windows (from 2021-2050 to 2071-2100 with a 10-year time-step). A linear regression is applied to the data clouds (black dotted line of slope a and intercept b).

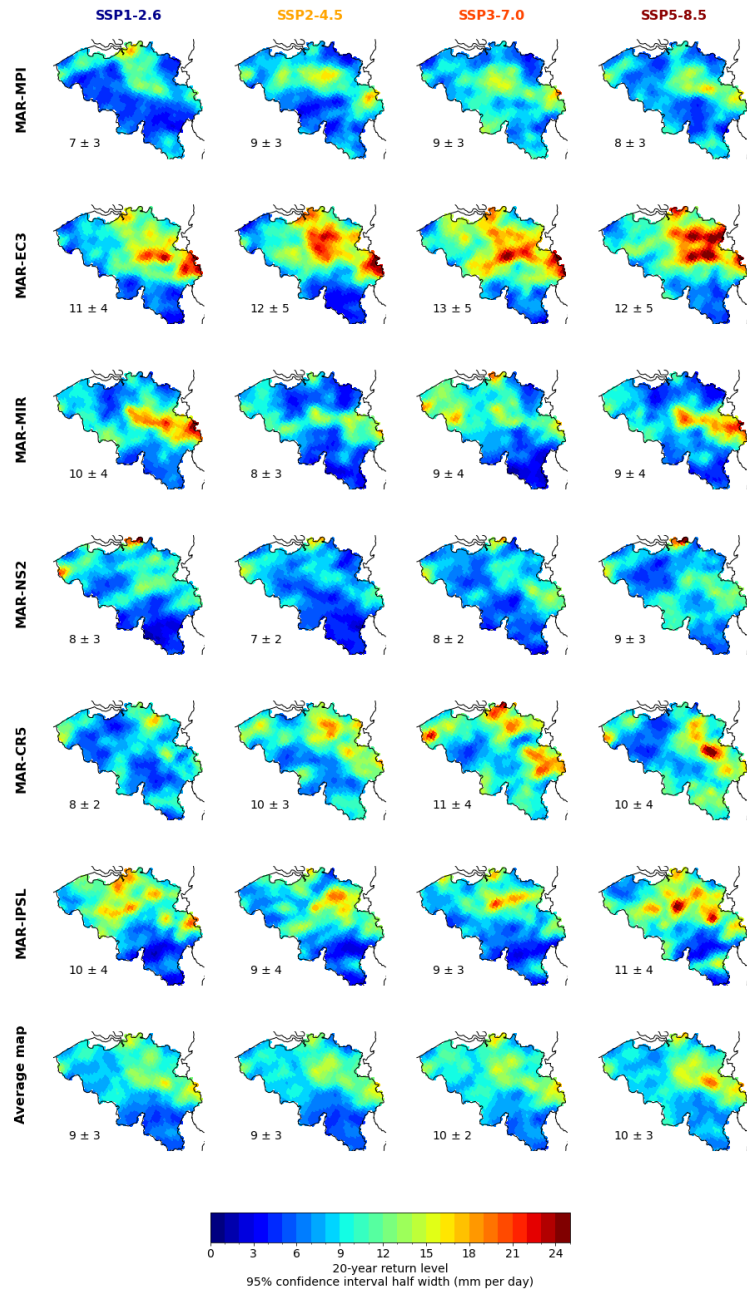


Figure S 7: 95% confidence interval half widths for all the models under all scenarios of the 20-year return level using the POT approach over 2015-2100. The written values represent the map average $\pm$ the standard deviation.

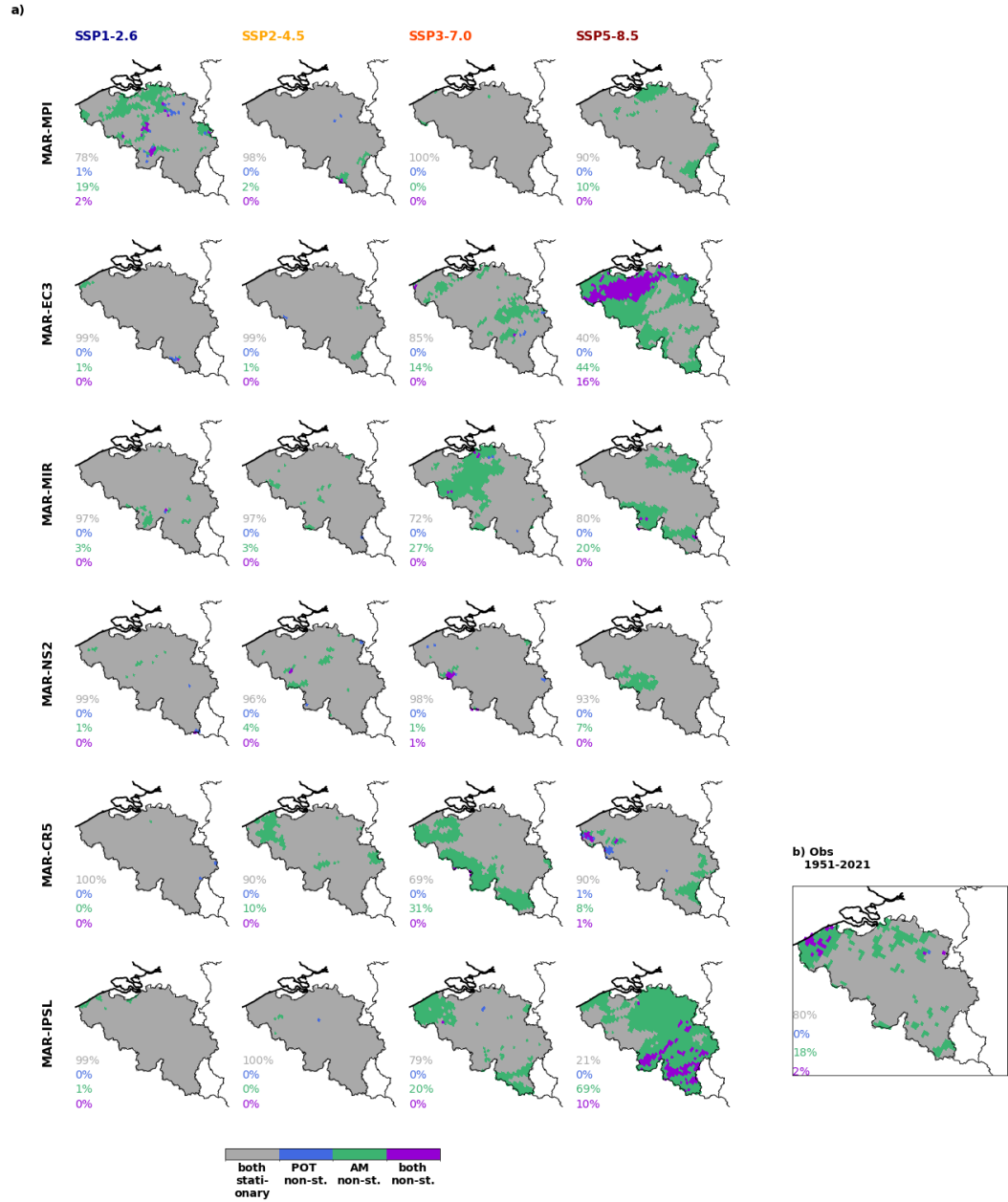


Figure S 8: Results of the log-ratio test. This checks whether non-stationary models perform better than stationary ones with a 95% confidence level. It is applied to corrected MAR outputs (2015-2100) (a) and to observations (1951-2021) (b). Blue color is plotted when the non-stationary model performs better in the POT (approach), Green color is plotted when it is the case for the AM approach and purple color is plotted if it is the case for both approaches. Written colored values correspond to the percentage of the map that is covered by the associated class.

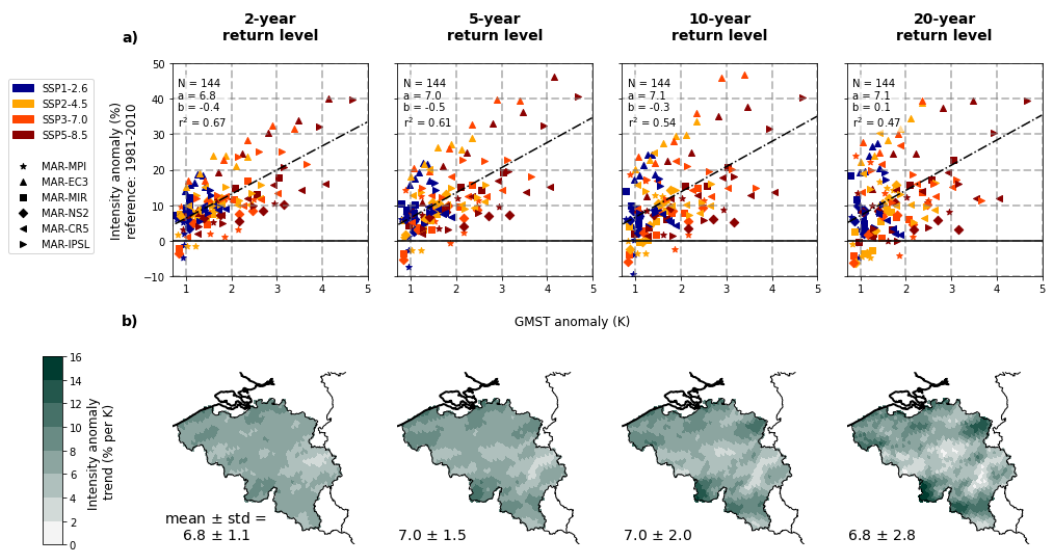


Figure S 9: (a) Intensity anomalies over 30-year moving windows for different return levels using the POT approach. A linear regression is applied to the data clouds (black dotted line of slope a and intercept b). (b) Same regression as in (a) but computed at each pixel location (the boxplots represents the distribution of the map values). The reference period is 1981-2010. Here MAR outputs are not bias-adjusted.

APPENDIX C

Second paper: Past and future evolution of cut-off-low-associated extreme precipitation in Europe

Abstract

Extreme precipitation events are projected to become more frequent and intense in Europe under climate change. In recent decades, cut-off lows (COLs), i.e., cold-core, upper-level low-pressure systems originating from polar/mid-latitude regions, have caused severe impacts across several European countries, leading to unprecedented rainfall totals. Understanding how the influence of COLs on extreme precipitation may evolve in a warming climate is therefore a key scientific challenge. To address this issue, we develop an objective detection algorithm based on 500 hPa geopotential height (Z500) fields to identify COLs.

Using the ERA5 reanalysis for both daily precipitation and Z500 fields, we construct a climatology of COLs affecting Europe over the period 1940–2024. We first show that the total number of COLs exhibits no significant long-term trend. We then associate daily precipitation events to potential COL influence. Our results indicate that approximately 52% of the most extreme European precipitation events are linked to a COL. Furthermore, we find that the statistical significance of trends in the annual number of COLs associated with precipitation extremes depends on both the precipitation threshold used to define extremes and the season considered.

In a second step, the same detection algorithm is applied to Z500 fields from six Earth System Models (ESMs) participating in the CMIP6 experiments recommended for EURO-CORDEX downscaling. We show that these state-of-the-art ESMs exhibit a negative

bias in the simulated number of COLs over the historical period (1981–2014), which is particularly pronounced over the Mediterranean basin. Using bias-adjusted precipitation fields, precipitation events in each ESM are then associated with COLs under the SSP5-8.5 scenario. As in the reanalysis, trends in the number of COL-associated events depend on the precipitation threshold. Our results suggest that current-generation ESMs struggle to represent both the dynamical characteristics of COLs and their associated precipitation impacts, which has important implications for the interpretation of future projections of extreme precipitation over Europe.

Overall, this study provides new insights into the past and future synoptic-scale impacts of COLs over Europe. Our results are consistent with the hypothesis that observed trends in extreme precipitation are primarily thermodynamically driven rather than dynamically. Because the proposed algorithm relies solely on Z500 fields, an accessible and widely available dataset, it can be readily applied to other reanalysis products or climate model outputs, offering a practical tool for assessing model performance in representing COL occurrence over specific regions.

C.1 Introduction

Extreme precipitation events are projected to become more frequent and intense in Europe under climate change (Coppola et al., 2021; Rajczak and Schär, 2017). The main underlying cause of these expected changes is the increased moisture retention capacity of the atmosphere resulting from its warming. This thermodynamic change leads, overall, to a 7% increase in precipitation intensity per additional degree of global warming, a relationship known as the Clausius–Clapeyron scaling (CC scaling) (Trenberth et al., 2003; Coppola et al., 2021; Rajczak and Schär, 2017).

While this thermodynamic intensification provides the background conditions for heavier precipitation, extreme precipitation events generally require favourable dynamical forcing at the synoptic scale. In recent years, cut-off lows (COLs) have been responsible for several major flooding events across Europe. In September 2024, Central Europe was affected by Storm Boris (ECMWF, 2024), which triggered widespread flooding. This was followed by the devastating floods in Valencia in October 2024 (Rodero Astaburuaga, 2025). Meanwhile, parts of Belgium and Germany are still recovering from the catastrophic floods of July 2021 (Kreienkamp et al., 2021; Dewals et al., 2021). A COL is a cold-core upper-level depression that becomes isolated from the main westerly flow as a result of large-amplitude jet-stream meanders (Palmén, 1949; Palmén and Newton, 1969; Nieto et al., 2005; Eumetsat, 2024). The cold-core nature of these systems can enhance atmospheric instability through the presence of cold air in the upper troposphere, which can enhance upward motion and the development of intense precipitation. Understanding how the frequency and impacts of COLs may evolve under future climate conditions is therefore an important research question.

Research on the future evolution of COLs remains relatively limited, with only a few studies having explicitly addressed potential changes in their occurrence. Recently, Muñoz et al. (2020) published COL climatologies for several regions across the globe. They also provided a review of previous efforts to generate COL climatologies. Most previous studies have used techniques based on geopotential height minima to identify COLs (Fuenzalida et al., 2005; Ndarana and Waugh, 2010; Favre et al., 2012; Pinheiro

et al., 2016), while other techniques are based on closed contours of potential vorticity (PV) anomalies identified on isentropic surfaces (Wernli and Sprenger, 2007) or on conceptual models of COLs (Nieto et al., 2005; Reboita et al., 2010; Muñoz et al., 2020). These PV cut-offs generally originate from Rossby-wave breaking, during which high-PV air becomes detached from the main westerly flow. Among these studies, only Muñoz et al. (2020) proposed an analysis of the evolution of annual COL counts over Europe for the period 1960–2017 using National Centers for Environmental Prediction (NCEP) reanalysis data, based on a detection algorithm relying on isolated low-pressure systems associated with eastward-propagating thermal fronts. Their results suggest significant positive trends over Europe.

Some recent studies have suggested that climate change may also influence the large-scale atmospheric circulation through Arctic amplification, i.e. the faster warming of the Arctic relative to lower latitudes, which reduces the meridional temperature gradient (Preece et al., 2023). Such changes may affect the characteristics of the mid-latitude jet stream and, consequently, the occurrence and persistence of cut-off lows. For example, Mishra et al. (2025), based on analyses of both the fifth-generation ECMWF atmospheric reanalysis (ERA5; Hersbach et al., 2020) and Earth System Model (ESM) projections, suggested that a weakening of the upper-level westerlies could favour longer-lasting COL systems. By identifying circulation analogues of the July 2021 event, Thompson et al. (2024b) similarly found that longer-lasting COLs may become more frequent in the future.

Despite these studies, very little is known about the evolution of COL-associated extreme precipitation. Among the existing studies, Barnes et al. (2023) examined slow-moving cyclones over Australia and found that approximately 72% of extreme precipitation events were associated with such systems. Focusing on southern Europe, Ferreira (2021) used a pseudo-global-warming framework under a high-emissions scenario and showed that precipitation intensities associated with COLs could increase by up to 88% over Spain, identifying the Valencia region as a potential hotspot. Despite these advances, no previous study has systematically associated extreme precipitation events across Europe to detected COLs over the historical period while simultaneously assessing their future evolution using Earth System Model projections. Consequently, the contributions of changes in COL-associated precipitation intensity to future hydroclimatic risk remain poorly constrained.

In this context, this article presents an objective algorithm based on image analysis techniques that enables the systematic association of precipitation events to COLs, as other techniques, including analogue-based approaches, may be ill-suited to detect systems that often exhibit highly variable spatial structures (Thompson et al., 2024b). The primary objective is to associate high-impact precipitation events over Europe, identified using ERA5 reanalysis data for the period 1940–2024, with dynamically detected COLs. In contrast to previous approaches that either focus solely on circulation patterns or directly analyse precipitation extremes, our framework explicitly separates the dynamical occurrence of COLs from their associated precipitation impacts. This allows us to discriminate changes in COL frequency from changes in precipitation intensity. Moreover, this framework provides the opportunity to construct a consistent climatology of COL occurrence over Europe during the historical period.

As a second application, the same methodology is applied to outputs from six ESMs, using bias-adjusted daily precipitation fields regridded to the ERA5 grid over a historical period (1981–2014) and a future period under the SSP5-8.5 scenario. The selected

models belong to the subset of ESMs recommended for EURO-CORDEX downscaling experiments (Sobolowski et al., 2025). Brajkovic et al. (2025) and Kendon et al. (2023) recently showed that, for rare extreme precipitation events, analyses based on Climate Model outputs suggest that natural variability may play a dominant role in determining the timing of the most extreme occurrences. Our study therefore provides an opportunity to investigate the role of natural variability in the occurrence and impacts of COLs during the historical period and in future simulations, using a consistent COL association framework.

This article is structured as follows. Section C.2 describes the detection algorithm and the datasets used in the study. In section C.3, we first analyse the evolution of COL occurrence in the ERA5 reanalysis, followed by a COL-association analysis using selected CMIP6 model simulations. For each results subsection, a trend analysis is performed to identify statistically significant changes in the annual number of COLs. In addition, regions most affected by COL-related precipitation are identified, allowing the characterization of COL-associated precipitation hotspots. Finally, Section C.4 discusses the limitations of the methodology and highlights potential extensions and applications for future research.

C.2 Material and methods

This section first describes the study domain and the input datasets used for the detection of cut-off lows (COLs) and the association of extreme precipitation events with these systems. Next, the objective COL detection algorithm is presented together with its manual validation. The methodology used to associate precipitation with detected COLs is then described. Finally, the trend analysis adopted to analyse temporal evolutions throughout the study is introduced. The complete workflow of the proposed methodology is summarized in Fig. C.2.

C.2.1 Studied domain and input data

C.2.1.1 European domain

The analysis focuses on a single European domain designed to capture the large-scale influence of COLs on extreme precipitation, while excluding high-latitude systems governed by distinct dynamical mechanisms (Tamarin and Kaspi, 2017). The domain extends from approximately 35°N to 60°N in latitude and from 20°W to 35°E in longitude (red contour in Fig. C.1). This region encompasses Western, Central, and Southern Europe, including the Mediterranean basin. The Baltic and northern Scandinavian regions are deliberately excluded, as the present study does not aim to investigate polar depressions.

C.2.1.2 Input data

The algorithm developed in this study aims to associate daily precipitation extremes with dynamically identified COLs over Europe. To ensure that the methodology can be readily applied to both reanalyses and ESM outputs, the detection algorithm relies

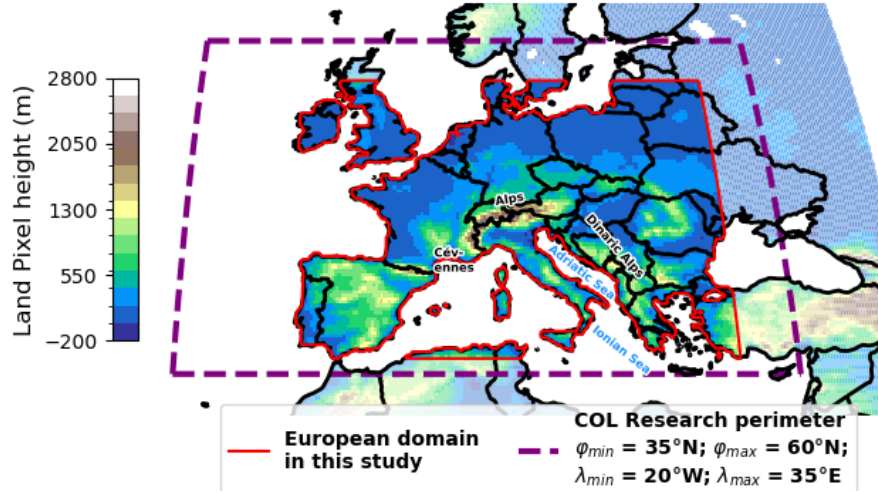


Figure C.1: Studied European domain and COL research perimeter. The red contour delineates the European domain used for the analysis of precipitation extremes. The background shading shows ERA5-Land pixel elevations. The dashed purple contour indicates the broader research perimeter within which COLs are dynamically detected ($\varphi_{\min} = 35^{\circ}\text{N}$, $\varphi_{\max} = 60^{\circ}\text{N}$, $\lambda_{\min} = 20^{\circ}\text{W}$, $\lambda_{\max} = 35^{\circ}\text{E}$). This enlarged perimeter allows the identification of dynamically relevant COLs whose centres may lie outside the precipitation impact domain.

CMIP6 ESM	Member	TCR (K)	Atm. horiz. res.
MIROC6	r1ip1f1	1.55	$\sim 1.4^{\circ} \times 1.4^{\circ}$
NorESM2-MM	r1ip1f1	1.33	$\sim 1.0^{\circ} \times 1.0^{\circ}$
MPI-ESM1-2-HR	r1ip1f1	1.66	$\sim 0.9^{\circ} \times 0.9^{\circ}$
EC-Earth3-Veg-HR	r1ip1f1	2.62	$\sim 0.7^{\circ} \times 0.7^{\circ}$
CMCC-CM2-SR5	r1ip1f1	2.09	$\sim 1.0^{\circ} \times 1.0^{\circ}$
IPSL-CM6A-LR	r1ip1f1	2.32	$\sim 2.5^{\circ} \times 1.3^{\circ}$

Table C.1: ESMs used in this study (modified after Sobolowski et al. (2025)). TCR = Transient Climate Response.

exclusively on daily 500 hPa geopotential height (z_{500}) fields together with daily precipitation fields. For the historical period (1940–2024), we use the ERA5 reanalysis, with a horizontal resolution of $0.25^{\circ} \times 0.25^{\circ}$ (Copernicus Climate Change Service (C3S), 2017; Hersbach et al., 2020).

For climate projections, six ESMs are analyzed under both historical (1981–2014) and SSP5-8.5 (2015–2100) scenarios. These models belong to the subset recommended for EURO-CORDEX dynamical downscaling experiments (Sobolowski et al., 2025) (Table C.1). Since the detection algorithm relies on the spatial extent of COLs, daily z_{500} fields from each ESM are regridded to the ERA5 grid using a distance-weighted average of the four nearest model grid points. This minimizes artificial differences in detected COL size resulting solely from differences in model horizontal resolution.

ESM precipitation fields are bias-adjusted using a quantile-mapping approach, specifically the Equidistant Quantile Mapping (EDCDFm) method (Pierce et al., 2015). Prior to bias adjustment, precipitation fields are regridded to the ERA5 grid. ERA5 precipitation is used as the reference dataset over the 1981–2010 calibration period to correct daily precipitation values at each grid cell. The correction factors are season-dependent and

differ between extreme and non-extreme precipitation days. This approach is similar to that employed by Brajkovic et al. (2025) and ensures consistent comparisons between COL-related precipitation in ERA5 and ESM datasets.

The selected ESMs exhibit a wide range of Transient Climate Responses (TCRs), defined as the global mean surface temperature increase at the time of CO₂ doubling under a 1% yr⁻¹ CO₂ increase scenario (IPCC, 2023). This diversity in climate sensitivity allows the exploration of a range of model responses. The SSP5-8.5 scenario is used because it provides the strongest climate change signal, facilitating the analysis of potential changes in COL occurrence and associated precipitation. Results obtained under SSP1-2.6 are discussed in Section C.4.3.

C.2.2 COL detection algorithm

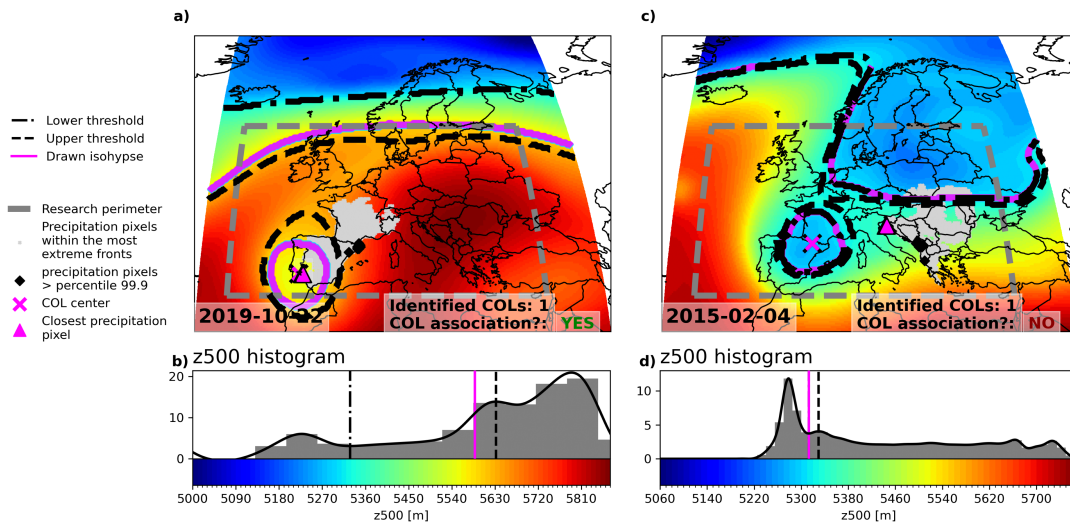


Figure C.2: Illustration of the cut-off low (COL) detection and precipitation association methodology. (a) Example of a COL detected in the 500 hPa geopotential height (z_{500}) field on 22 October 2019. The magenta contour represents the closed isohypse retained by the iterative detection algorithm and the cross marks the centre of the identified COL. The most extreme precipitation front is associated with the detected COL because it intersects the COL area, resulting in a positive COL–precipitation association. (b) Histogram of z_{500} values showing the dynamically derived lower and upper thresholds used to guide the identification of closed contours. (c,d) Same as (a,b), but for 4 February 2015. Although a COL is successfully detected, the most extreme precipitation front does not intersect the detected COL. Consequently, no COL–precipitation association is assigned.

The detection of COLs relies exclusively on the dynamical characteristics of isolated low-pressure systems in the 500 hPa geopotential height (z_{500}) field. While previous studies have also used either potential vorticity (PV) anomalies (Barnes et al., 2023) or temperature gradients to detect thermal fronts (Muñoz et al., 2020), we focus on the z_{500} field, in which COLs are classically represented as isolated closed geopotential minima. COLs generally develop when upper-level troughs become detached from the main westerly flow, forming isolated cold-core circulations in the mid-troposphere (Fig. C.2a; (Muñoz et al., 2020)). In the z_{500} field, such systems are characterized by closed isohypses forming isolated minima.

C.2.2.1 Detection of closed geopotential structures

For each synoptic configuration, a histogram of z_{500} values is constructed over the analysis domain (Fig. C.2b). A cubic spline is fitted to this distribution in order to identify local extrema, which are then used to define two threshold values delimiting the range of geopotential heights within which closed contours are most likely to occur. Specifically, the lower threshold corresponds to the first local minimum, and the upper threshold to the second local maximum of the fitted curve (Fig. C.2b). This approach enables the identification of isolated mid-tropospheric low-pressure systems embedded between the lower geopotential heights at higher latitudes and the higher geopotential values typically found further south.

Isohypsers are subsequently drawn at 10 m intervals within this range. Each closed contour is then evaluated against the following dynamical and geometrical criteria to determine whether it qualifies as a COL:

1. The area enclosed by the closed contour must exceed a critical threshold of $A_{\text{crit}} = 50,000 \text{ km}^2$. This value is chosen to exclude small-scale, short-lived geopotential anomalies that are not representative of mature synoptic-scale COLs, such as transient upper-level troughs embedded in the large-scale flow. The influence of the chosen value of A_{crit} on the climatological results is evaluated in Section C.4.1.
2. The centre of the enclosed area must lie within a predefined domain, bounded by latitude (φ) and longitude (λ):

$$\varphi_{\min} < \varphi < \varphi_{\max}, \quad \lambda_{\min} < \lambda < \lambda_{\max}.$$

At this stage, it is important to note that the detection of COLs relies solely on the structure of the z_{500} field. No precipitation-related information is used to identify closed geopotential systems. Because the analysis is based on daily averaged z_{500} fields, only systems that persist long enough over the day are detected, thereby excluding short-lived transient features. Furthermore, because the histogram decomposition is performed on the entire z_{500} field, the threshold values are derived from the large-scale synoptic configuration rather than from local properties of individual COLs. This provides a consistent statistical framework for identifying isolated upper-level circulations while preserving their relationship with the surrounding synoptic-scale flow.

To improve robustness, the detection algorithm is applied using multiple histogram class widths ranging from 10 m to 110 m in steps of 10 m. For each iteration, the thresholds are recalculated and the search for closed contours is repeated. This multi-resolution approach ensures that the detection is not sensitive to a specific binning choice and allows the algorithm to adapt to varying synoptic configurations.

C.2.2.2 Validation of the detection algorithm

To evaluate the robustness of the proposed detection algorithm, a manual validation was performed using 100 randomly selected daily synoptic situations, including 50 dates for which at least one COL was detected and 50 dates for which no

COL was identified. The complete validation database, together with the corresponding diagnostic figures illustrating the successive steps of the detection algorithm, is provided in the supplementary files `COL_algorithm_validation_part1.xlsx` and `COL_algorithm_validation_part2.xlsx`, available in the Zenodo repository at <https://doi.org/10.5281/zenodo.19604862>.

For the 50 randomly selected dates containing at least one detected COL, the algorithm correctly identified all visually identifiable COLs in the z_{500} field. A slight overestimation ($\sim 1\%$) was nevertheless observed because elongated COLs were occasionally partitioned into two adjacent closed systems by the contour extraction procedure (2 cases out of 50). Among the 50 randomly selected dates for which no COL was detected, the non-detection was confirmed in 98% of the cases. The one missed detections correspond to a COL located close to the domain boundaries, where the circulation is only partially represented.

Overall, this manual validation indicates that the proposed methodology reliably identifies synoptic-scale COLs at various stages of their lifecycle while providing a transparent assessment of its principal limitations. The proposed methodology identifies upper-level COLs solely from the z_{500} field and does not explicitly distinguish between the different types of surface cyclones that may be associated with them. Consequently, some detected COLs may also correspond to medicanes, provided that these systems are associated with a well-defined upper-level cut-off circulation. This behaviour is consistent with the objective of the algorithm, which aims to identify upper-level cut-off circulations rather than classify the associated surface weather system. Moreover, medicanes are commonly associated with upper-level COLs (Eumetsat, 2024; Saraceni et al., 2023).

The detection stage described above is entirely independent of precipitation. The association between detected COLs and extreme precipitation is performed in a second step, described below.

C.2.2.3 Association of precipitation with detected COLs

Once a COL has been identified from the z_{500} field, the association between the detected COL and extreme precipitation is performed in a second, independent step. The objective is not to associate all precipitation occurring over Europe with a COL, but rather to determine whether the most extreme precipitation event of the day is spatially connected to a detected COL.

To this end, precipitation fronts are first identified as spatially continuous regions composed of precipitation pixels exceeding 1 mm day^{-1} . A precipitation front is retained if it contains at least one grid cell exceeding the daily 99.9th spatial precipitation percentile over the European domain. A detected COL is then classified as precipitation-associated if at least one precipitation pixel belonging to one of these extreme precipitation fronts lies within the closed geopotential contour defining the COL.

Given that the European land domain contains 8,469 grid cells, the 99.9th percentile typically corresponds to approximately the nine most intense precipitation pixels of a given day. Consequently, this methodology ensures that the identified precipitation front contains the most extreme precipitation observed over Europe while allowing the remaining, less intense portions of the same continuous precipitation system to extend beyond the 99.9th percentile threshold (Fig. C.2a).

This methodological choice differs from previous studies, which generally associate precipitation with synoptic systems using fixed distance criteria (Hénin et al., 2018; Konstali et al., 2024; Rüdisühli et al., 2020). Instead, the proposed framework relies on the observed spatial intersection between identified precipitation fronts and detected COLs, thereby avoiding the introduction of an arbitrary radius of influence around each COL. It also prevents the association of precipitation systems that are spatially disconnected from the detected COL, as illustrated by the 4 February 2015 example shown in Fig. C.2c.

As with any objective association methodology, the proposed approach depends on the chosen definition of a precipitation front. However, by relying on the observed spatial organization of the precipitation field rather than on a prescribed distance threshold, it provides a physically consistent framework for associating extreme precipitation events with upper-level cut-off lows.

C.2.3 Application over Europe and trend identification

Extreme precipitation days are identified using spatial precipitation statistics computed over the European domain. For each day, spatial precipitation quantiles are calculated, and precipitation is subsequently associated with dynamically detected COLs following the methodology described above. The annual number of COL-associated precipitation events exceeding predefined intensity thresholds is then computed.

In hydrological and climate sciences, the temporal evolution of counting variables is commonly modeled using a Poisson regression framework (Brajkovic et al., 2025; Trambly et al., 2013; Grandry et al., 2020). In this context, the probability P_n of observing n COL occurrences in a given year is expressed as:

$$P_n = \frac{\exp(-\gamma) \gamma^n}{n!}, \quad (\text{C.1})$$

where γ denotes the expected annual number of events.

To account for potential temporal trends, γ is modeled as an exponential function of time:

$$\gamma = \exp(\gamma_0 + \gamma_1(y - y_0)), \quad (\text{C.2})$$

where y is the calendar year, y_0 is a reference year corresponding to the beginning of the study period, and γ_0 and γ_1 are regression coefficients estimated from the data. The parameter γ_1 represents the temporal trend in the expected number of COL occurrences.

The statistical significance of the temporal trend is assessed using a Wald test applied to the estimated slope parameter γ_1 . A 95% confidence level is adopted throughout the study, such that p -values below 0.05 indicate a statistically significant trend. A significant result implies that the observed temporal evolution in the number of COLs is unlikely to arise from random variability alone.

C.3 Results

The applied detection approach allows the systematic attribution of large-scale precipitation events to upper-level COLs in the ERA5 reanalysis. Over the period considered,

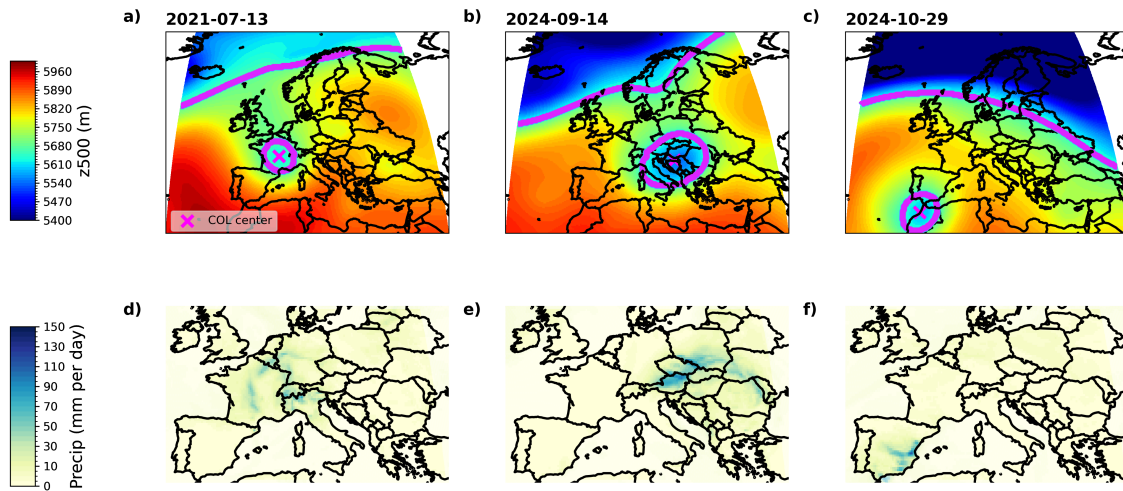


Figure C.3: Illustration of the cut-off low (COL) detection and precipitation attribution procedure for three representative events. 500 hPa geopotential height (z_{500}) fields showing dynamically detected closed geopotential structures on 13 July 2021 (a), 14 September 2024 (b) and 29 October 2024 (c). The magenta contours delineate the closed isohypses identified by the detection algorithm, and the magenta cross marks the corresponding COL center. (d-f) Daily precipitation fields for the same dates.

numerous high-impact precipitation events across Europe are associated with COLs detected by the algorithm (Fig. C.3). Depending on their location and structure, COLs can affect a wide range of European regions.

Three recent extreme events illustrate the ability of the algorithm to identify precipitation impacts associated with cut-off lows (COLs). The October 2024 Storm Dana event is clearly detected, with a well-defined upper-level closed low centered near the Strait of Gibraltar on 29 October (Fig. C.3c). This circulation pattern induced a predominantly east-to-west-oriented precipitation band affecting eastern Spain, where daily precipitation totals in ERA5 locally reached up to 120 mm day^{-1} near Valencia (Fig. C.3f).

Similarly, the July 2021 event that caused severe flooding in Belgium and western Germany is associated with a persistent COL detected by the algorithm (Fig. C.3a). In this case, the circulation produced a predominantly north-to-south-oriented precipitation band. Likewise, Storm Boris, centered over central Europe, generated intense rainfall organized along a predominantly east-to-west axis (Fig. C.3e).

These three recent COL-associated extreme events demonstrate how the algorithm links extreme precipitation to cut-off lows. They also illustrate the wide range of possible spatial configurations between the COL centre and the associated precipitation bands, highlighting that intense precipitation can occur in various locations relative to the upper-level circulation.

C.3.1 Historical COL impacts

While several recent flooding events were associated with exceptionally intense COL-related precipitation, not all COLs generate extreme rainfall. In the following, we therefore distinguish between the occurrence of COLs irrespective of their precipitation impacts and the subset of COLs associated with extreme precipitation events.

First, we analyse the long-term evolution and spatial distribution of COL occurrence

over the studied European domain, independently of precipitation. Second, we attribute historical daily precipitation extremes, defined using high spatial precipitation quantiles, to dynamically detected COLs in order to quantify their contribution to extreme precipitation events.

C.3.1.1 Evolution of the number of COLs

The historical record exhibits pronounced multi-decadal variability in COL occurrence (Fig. C.4a). In particular, the 1970s stand out as a phase of high activity, with annual counts exceeding 400 detected COLs in the daily geopotential height fields. This period is preceded and followed by alternating phases of higher and lower COL occurrence, with annual values generally fluctuating around the average value of 360 COLs per year.

Although these annual counts may appear high relative to the number of days in a year, COLs are counted on a daily basis. Persistent systems are therefore recorded once per day, and multiple COLs may simultaneously occur within the domain. This counting strategy was intentionally adopted because the objective of the study is to associate precipitation with COLs on a daily basis. Consequently, persistent COLs are counted once per day while they remain active. For completeness, the annual COL counts were also spatially and temporally declustered so that each persistent COL was counted only once. This declustering reduces the mean annual number of detected COLs to approximately 160 yr^{-1} but does not alter the statistical significance of the identified trends (Fig. D.2).

COL centres are preferentially concentrated over three regions (Fig. C.4b): (i) west of the Iberian Peninsula, (ii) central Italy east of Corsica, and (iii) the Black Sea region. In each of these areas, more than 200 COL centres are detected over the period 1940–2024, whereas regions of lowest activity typically record fewer than 170 events over the same period.

From a seasonal perspective, the highest COL activity is observed in spring (MAM; Fig. C.4d), summer (JJA; Fig. C.4e), and autumn (SON; Fig. C.4f). These three seasons exhibit pronounced inter-decadal variability, with a marked activity maximum during the 1970s. While significant positive tendencies are apparent in JJA and SON, no statistically significant trend is detected for MAM.

In contrast, winter (DJF; Fig. C.4c), despite being the least active season in terms of absolute COL counts, exhibits a slight decreasing trend over the study period. Consequently, the annual series is dominated by inter-decadal variability and exhibits no statistically significant long-term trend (Fig. C.4a).

C.3.1.2 Association of historical precipitation events with COLs

Across the European domain, we quantify the number and fraction of extreme precipitation events associated with COLs. Two precipitation thresholds (5 and 50 mm day^{-1}) are considered in order to distinguish between moderate and high-impact precipitation events. For each day, events are identified when the spatial 99.9th percentile of daily precipitation over the European domain exceeds the selected threshold, after which they are associated with detected COLs (Fig. C.5). The analysis is performed for the whole year (Y) as well as for the DJF, MAM, JJA, and SON seasons.

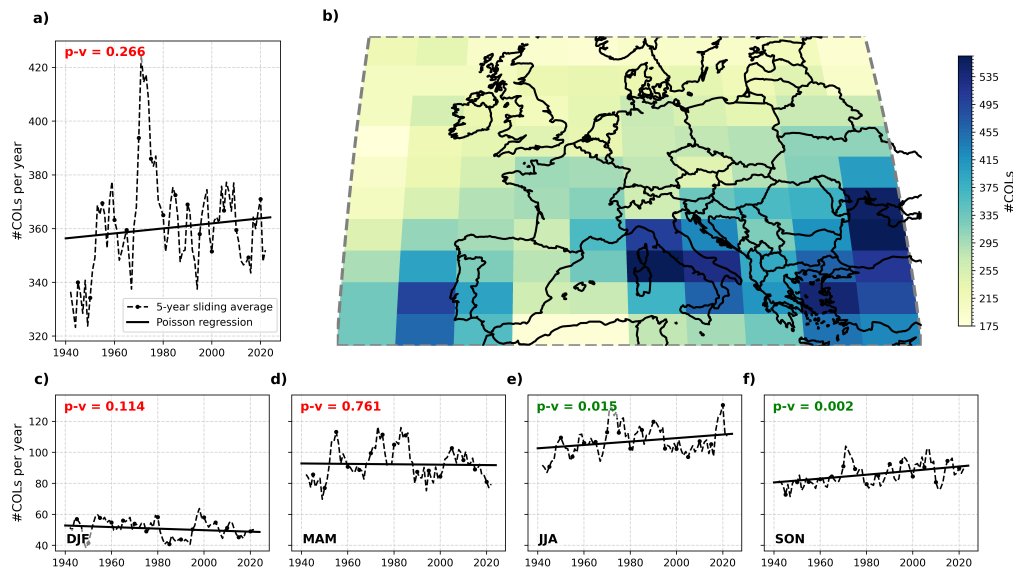


Figure C.4: Evolution of cut-off low (COL) occurrence over Europe in ERA5. (a) Annual number of detected COLs over the European domain for the period 1940–2024. The dashed line shows the 5-year running mean and the solid line indicates the Poisson regression fit; the associated p -value is reported (green: statistically significant at the 95% confidence level; red: non-significant). (b) Spatial distribution of COL centres, expressed as the cumulative number of detected COLs over the same period. (c–f) Seasonal evolution of the annual number of COLs for winter (DJF), spring (MAM), summer (JJA), and autumn (SON). Dashed lines represent 5-year running means, solid lines correspond to Poisson regression fits, and reported p -values indicate the statistical significance of the temporal trends. Seasonal spatial distributions are shown in Appendix Fig. D.1. An algorithm sensitivity test for the A_{crit} parameter is presented in Appendix Figure D.3.

On an annual basis, approximately 44% of annual extreme precipitation events are associated with a COL when using the 5 mm day^{-1} threshold (Fig. C.5a). Despite pronounced interannual and inter-decadal variability, no statistically significant long-term trend is detected for the 5 mm day^{-1} threshold (Fig. C.5a). Periods of enhanced COL activity nevertheless occur during the 1970s and the 2000s, reaching up to about 170 COL-associated events per year.

In contrast, applying the higher threshold of 50 mm day^{-1} yields a markedly different behaviour. In this case, the temporal evolution shows a statistically significant positive trend (Fig. C.5f). The 1970s again emerge as a period of elevated COL activity, with up to 40 COL-associated events per year exceeding the threshold. Over the full 1940–2024 period, the Poisson regression indicates an increase in the expected annual number of COL-associated extreme events from approximately 24 to 30. These results indicate that, whereas moderate COL-associated precipitation has remained approximately stationary, the most intense COL-associated precipitation events have become significantly more frequent over the study period.

From a seasonal perspective, summer (JJA) emerges as the season most affected by COL activity, with 54% and 68% of events associated with COLs for the lower and higher thresholds, respectively (Fig. C.5d and i). Using the lower threshold of 5 mm day^{-1} , no statistically significant trend is detected, with annual counts remaining close to 50 COL-associated events per year (Fig. C.5d). When applying the higher threshold of 50 mm day^{-1} , a statistically significant positive trend is identified. In this case, the expected annual number of COL-associated extreme precipitation events rises to

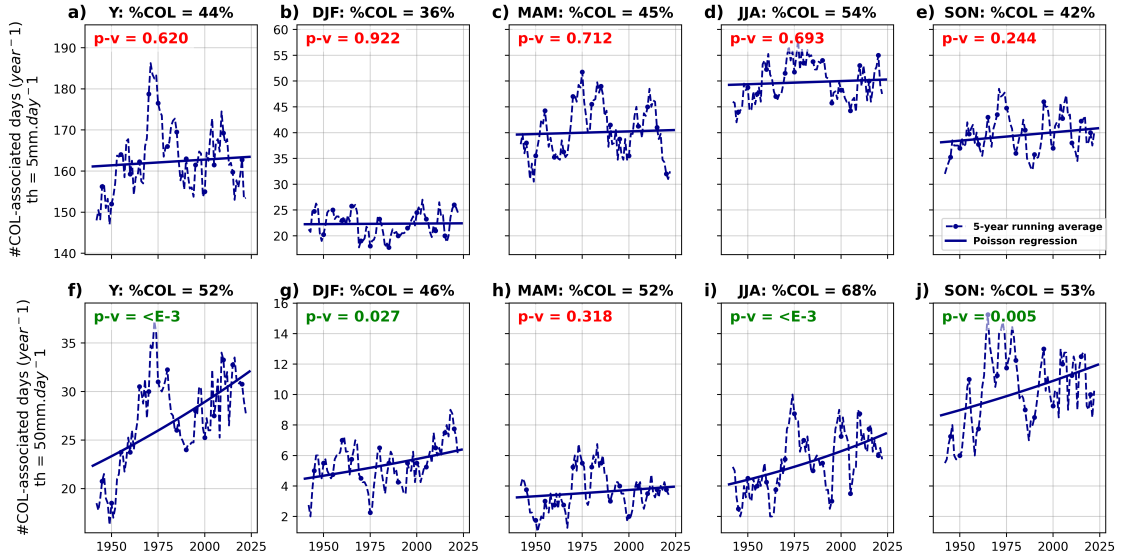


Figure C.5: Annual and seasonal evolution of the number of extreme precipitation days associated with cut-off lows (COLs) in ERA5. Panels (a–e) show the number of days per year exceeding the spatial 99.9th percentile of daily precipitation and reaching at least 5 mm day⁻¹, associated with COLs, for the whole year (a), DJF (b), MAM (c), JJA (d), and SON (e). Panels (f–j) show the same diagnostics using a stricter precipitation threshold of 50 mm day⁻¹. Dashed lines represent 5-year running means, while solid lines correspond to Poisson regression fits. Reported p-values indicate the statistical significance of the temporal trends (green: statistically significant at the 95% confidence level; red: non-significant). The same figure with other applied thresholds is available in Supplementary Fig. D.6

approximately 7 events per year in recent decades (Fig. C.5i).

With this higher threshold, more events are found in SON, which also shows a statistically significant positive trend, rising to approximately 12 events per year in recent decades (Fig. C.5j). A statistically significant positive trend is also identified in DJF (C.5b and g). For MAM, the number of COL-associated events exhibits no trend for both thresholds (C.5c and h).

Overall, these results indicate that statistically significant long-term increases are confined to the most intense COL-associated precipitation events.

C.3.1.3 Historical COL impact locations

COLs have contributed to a large proportion of extreme precipitation events across Europe. To identify the regions most frequently affected, we mapped the number of days on which COL-associated precipitation exceeded 15 mm day⁻¹. This threshold was selected to ensure that at least one COL-related precipitation event affected each grid cell within the European domain.

The region most frequently impacted by COL-related precipitation is located in the southern Dinaric Alps, in the vicinity of Montenegro (Fig. C.6a). Over the period 1940–2024, this area experienced more than 500 COL-associated precipitation days. Additional regions with similarly high frequencies are found in parts of the European Alps.

These regions are characterized by pronounced topographic gradients (Fig. C.1), which favour orographic uplift. In coastal mountain ranges such as the Dinaric Alps, moist maritime air masses from the Adriatic Sea as well as continental air masses can be

forced upslope, enhancing precipitation during COL events. A similar enhancement is observed over the mountainous regions of western Greece bordering the Ionian Sea. When examining the mean precipitation associated with COLs (Fig. C.6b), the southern Dinaric Alps again emerge as a major hotspot, exhibiting both high frequencies and high intensities of COL-related precipitation. Other prominent topographic features are also visible, including the Apennines, the Cévennes Massif, and the Iberian mountain ranges. These areas consistently coincide with enhanced COL-associated precipitation.

However, elevated COL-related precipitation is not restricted to mountainous regions. Flatter areas such as eastern Germany also exhibit relatively high impacts, indicating that topography alone does not fully explain the spatial distribution of COL-associated precipitation. The identified precipitation hotspots are consistent with the preferred regions of COL occurrence shown in Fig. C.4b, suggesting that the spatial distribution of COL-associated precipitation likely reflects the combined influence of preferential COL trajectories and regional topographic enhancement.

Overall, southern European mountain ranges emerge as the principal hotspots of COL-associated precipitation, although significant impacts are also observed over several lowland regions, highlighting the combined influence of synoptic circulation and regional topography.

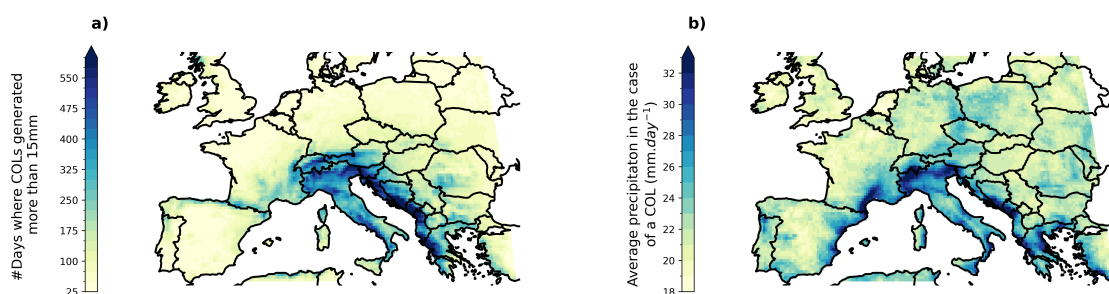


Figure C.6: Spatial distribution of cut-off-low (COL)–associated precipitation impacts over Europe for the period 1940–2024. (a) Number of days on which COL-associated precipitation exceeds 15 mm day^{-1} . (b) Mean daily precipitation amount associated with COLs (mm day^{-1}). Both panels highlight regions frequently affected by COL-associated precipitation as well as areas where COLs produce higher average rainfall intensities.

C.3.2 Evolution of COL impacts in Earth System Models

The historical evolution of COL-induced precipitation events exhibits pronounced natural variability. In this section, we first assess how well the selected ESMs reproduce the frequency and spatial distribution of COLs over the historical period. We then examine the projected evolution of COL-related precipitation impacts as simulated by these models.

C.3.2.1 Historical representation of COL occurrence by ESMs

The selected ESMs exhibit varying skill in representing COL occurrence independently of any precipitation attribution. When analysing the evolution of the total number of detected COLs over the historical period, a pronounced inter-decadal variability is evident in all models, similar to that observed in ERA5 (Fig. C.7a).

However, the absolute number of simulated COLs is systematically underestimated by the ESMs compared to ERA5. MIROC6 and EC-Earth3-Veg provide the closest agreement with ERA5, with annual counts fluctuating around 300 COLs per year (Fig. C.7a). In contrast, CMCC-CM2-SR5 and NorESM2-MM exhibit the strongest underestimation, with approximately 210 identified COLs per year. Finally, MPI-ESM1-2-HR and IPSL-CM6A-LR fall between these two groups, with annual counts around 250 COLs per year.

This overall underestimation is also reflected in the spatial distribution of COL occurrence. All ESMs exhibit their largest negative biases over the Mediterranean basin, a feature that is particularly pronounced in NorESM2-MM and CMCC-CM2-SR5 (Fig. C.7e and f). As a result, the three main regions of enhanced COL activity identified in ERA5 are not reproduced by the models.

The spatial anomaly patterns (Figs. C.7b–g) indicate that COL occurrence is preferentially shifted northwestward in the ESMs compared to ERA5 (Fig. C.4b). For example, MIROC6, despite providing the closest agreement with ERA5 in terms of total COL counts, shows a positive bias over the North Atlantic while underestimating COL activity over the Mediterranean basin and the Black Sea region.

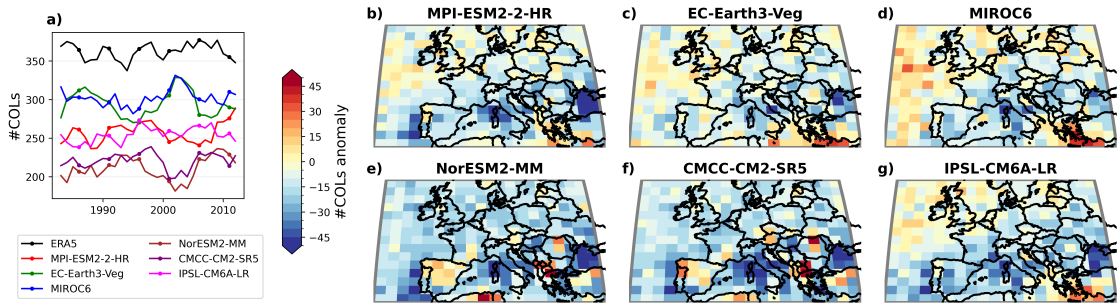


Figure C.7: Evaluation of cut-off low (COL) occurrence in CMIP6 ESMs relative to ERA5 over the historical period (1981–2014). (a) Annual number of detected COLs in ERA5 and in the selected ESMs (see Table C.1), with a 20-year running mean applied. (b–g) Spatial anomalies in the cumulative number of detected COL centres in each ESM relative to ERA5 over the same period.

C.3.2.2 Association of projected extreme precipitation events by ESMs with COLs

Using the same methodology as for ERA5 (Section C.3.1.2), future extreme precipitation events simulated by the ESMs under the SSP5-8.5 scenario are associated with potential COLs. When applying the lower 5 mm day^{-1} threshold, all ESMs exhibit a decrease in the annual number of COL-associated precipitation events, in contrast to the absence of a significant trend in ERA5 over the historical period (Fig. C.8a).

In addition to these differing trends, the absolute number of COL-related events is systematically underestimated by the models. Two distinct groups emerge: one characterized by stronger underestimation (NorESM2-MM, CMCC-CM2-SR5, and IPSL-CM6A-LR) and another showing closer agreement with historical ERA5 counts (EC-Earth3-Veg, MIROC6, and MPI-ESM1-2-HR). This grouping remains consistent across precipitation seasons regarding this lower threshold (Fig. C.8).

When applying the higher threshold of 50 mm day^{-1} , most ESMs reproduce the behaviour observed in ERA5, with statistically significant increases in the number of

COL-associated extreme precipitation days (Fig. C.8f). CMCC-CM2-SR5 is the only exception, showing no significant positive trend. Despite these increases, most models continue to underestimate the historical ERA5 event counts throughout the 21st century. EC-Earth3-Veg constitutes the main exception, reaching values comparable to those observed in ERA5 by the end of the century. It also provides the closest agreement with the ERA5-based attribution, with 49% of extreme precipitation days associated with a COL compared with 53% in ERA5 (Fig. C.8f).

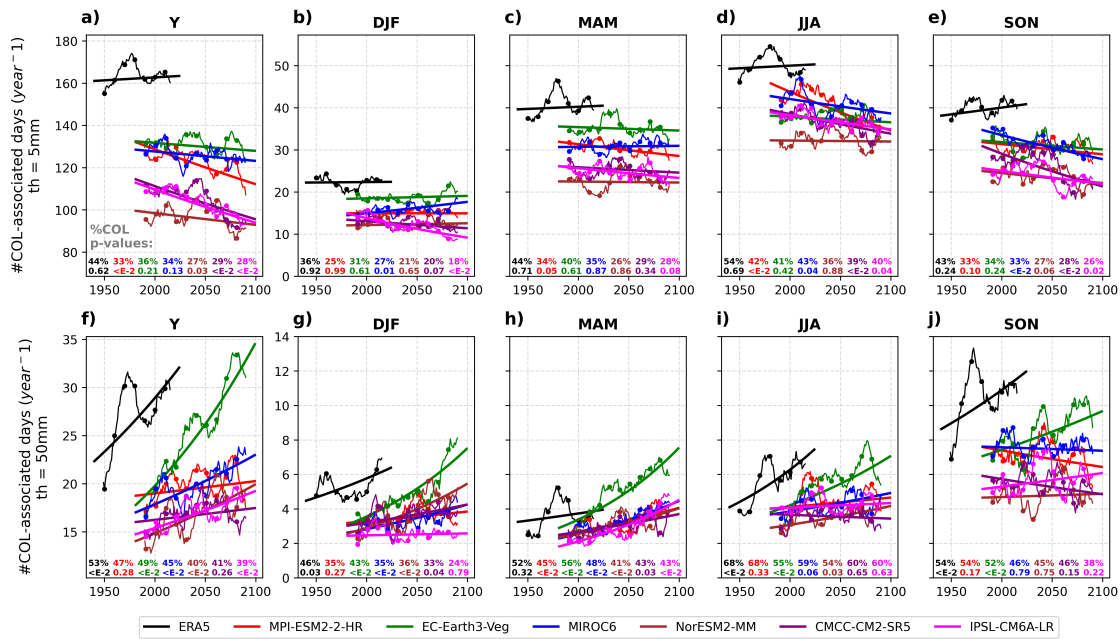


Figure C.8: Annual and seasonal evolution of the number of COL-associated extreme precipitation days in ERA5 and CMIP6 ESMs. Panels (a–e) show the annual number of days exceeding the spatial 99.9th percentile of daily precipitation and a precipitation threshold of 5 mm day⁻¹, attributed to cut-off lows (COLs), for the whole year (a), DJF (b), MAM (c), JJA (d), and SON (e). Panels (f–j) show the same diagnostics using a higher precipitation threshold of 50 mm day⁻¹. Black curves correspond to ERA5 over the historical period, whereas coloured curves correspond to continuous ESM simulations covering the historical period (1981–2014) followed by the SSP5-8.5 scenario (2015–2100). Dashed curves denote 20-year running means, while solid lines represent Poisson regression fits used to assess temporal trends. The percentage of extreme precipitation days associated with COLs and the corresponding *p*-values are shown at the bottom of each panel using the colour of the corresponding dataset.

From a seasonal perspective, spring (MAM) shows the strongest inter-model agreement in future trend behaviour under the higher 50 mm day⁻¹ threshold. All ESMs exhibit increasing tendencies in the number of COL-associated extreme precipitation events (Fig. C.8h). EC-Earth3-Veg stands out in particular, exceeding historical ERA5 levels by the end of the century, with more than six events per year during some periods.

In contrast, the systematic underestimation of COL-related extreme precipitation by ESMs is particularly pronounced in autumn (SON) and Summer (JJA) (Figs. C.8i and j). For these two seasons, simulated event counts under the higher threshold never reach the levels observed in ERA5. EC-Earth3-Veg is the only model exhibiting increasing trends comparable in sign to those derived from ERA5, although with substantially lower absolute magnitudes.

Beyond the long-term trends, each ESM exhibits its own internal variability, reflecting the natural variability also observed in the reanalysis. In summer (JJA), for example,

MPI-ESM1-2-HR displays a period of enhanced COL-related activity at the beginning of the simulation, followed by extended phases of lower activity (Fig. C.8i). Similar inter-decadal fluctuations are apparent across all seasons and models, highlighting the strong role of internal variability in modulating COL-related precipitation at multi-decadal timescales.

C.4 Discussion

This study shows that trends in COL-associated precipitation extremes depend strongly on the precipitation threshold used to define extreme events, and that ESMs systematically underestimate both the occurrence and impacts of COLs over Europe. In this section, we first discuss the main limits of the presented framework and place our work with regard to recent studies. We then briefly review additional physical mechanisms that may contribute to the observed trends in COL-associated extreme precipitation beyond changes in COL occurrence alone. Finally, we discuss the broader implications of these findings for the interpretation of past trends and future projections of COL-related extreme precipitation.

C.4.1 Methodological limitations and comparison with previous studies

Although the proposed algorithm is readily applicable to different datasets, it requires the definition of a small number of user-defined parameters. In the present study, sensitivity analyses indicate that these parameters primarily influence the absolute number of detected COLs rather than the main climatological conclusions. For instance, one parameter that influences the total number of detected COLs is A_{crit} , the critical area that an isolated low-pressure system must reach in order to be classified as a COL. In this study, A_{crit} is set to $5 \times 10^4 \text{ km}^2$. Increasing A_{crit} to 10^6 km^2 results in a systematic downward shift in the annual counts shown in Fig. C.4, without affecting the temporal behaviour or the detected trends (See supplementary figure D.3).

Over the historical period, the proposed algorithm identifies on average approximately 360 COLs yr^{-1} over the European domain. At first glance, this value may appear relatively large compared with previous climatologies. These reported counts correspond to daily detections and are therefore not spatio-temporally declustered. As shown by the additional declustering analysis (Appendix Fig. D.2), counting each persistent COL only once substantially reduces the annual number of detected systems while preserving the temporal evolution and the main conclusions of the study. Together with the manual validation presented in Section C.2.2.2, these analyses provide confidence that the reported frequencies primarily reflect methodological differences rather than detection errors.

However, the technique used to identify COLs is purely statistical and relies on geometrical criteria. Depending on the synoptic situation, the distinction between a polar depression and a COL can sometimes be ambiguous. Consequently, in a few isolated cases, other types of extratropical cyclones may be identified as COLs. In our validation dataset, among the 50 randomly selected cases, we identified one case (see Fig. SS47 in COL_algorithm_validation_part1.xlsx) that we believe is more appropriately classified as a polar cyclone than as a COL.

The association of future precipitation events with detected COLs may also be influenced by the applied bias-adjustment procedure. In this study, precipitation is corrected using a classical empirical quantile mapping approach. Although conceptually simple, univariate quantile mapping has been shown to perform competitively with more sophisticated multivariate bias-adjustment techniques (Van de Velde et al., 2022). Nevertheless, for events located in the extreme tails of the distribution, the correction factors derived from the historical period may not fully represent the biases affecting the most exceptional future events (Pierce et al., 2015). Despite this limitation, the systematic underestimation of COL counts during the historical period suggests that the low number of projected COL-associated events is likely to be influenced, at least in part, by an underrepresentation of COL occurrence in the driving ESMs rather than solely by the bias-adjustment procedure.

Beyond these methodological considerations, it is also useful to compare the proposed framework with existing COL climatologies. Among similar previous studies, Mishra et al. (2025) reported approximately 50 COLs yr^{-1} over a broadly similar European domain for the 1979–2004 period using ERA5 reanalysis data, while Muñoz et al. (2020) obtained comparable frequencies using NCEP/NCAR reanalyses over 1960–2014. These differences are likely explained by methodological choices. In particular, the detection algorithm of Muñoz et al. (2020) requires the presence of an eastward-propagating thermal front associated with the COL, whereas the methodology proposed here identifies upper-level cut-off circulations solely from the z_{500} field, independently of the position of the associated frontal system. This distinction is important because the precipitation associated with impactful COLs is not systematically confined to the eastern flank of the circulation, as illustrated by the July 2021 (Bernd) and September 2024 (Boris) events (Figs. C.3a and b). Likewise, Mishra et al. (2025) identify COLs using potential vorticity anomalies rather than closed geopotential structures. Consequently, the proposed methodology detects a broader range of mature upper-level cut-off circulations, including systems in a decaying stage that still exhibit well-defined closed z_{500} contours.

In their study, Mishra et al. (2025) also suggest that the frequency of particularly long-lasting systems might increase in the future. Their conclusions are derived from projected COL frequency anomalies computed for the 2071–2100 period. Similarly, Muñoz et al. (2020) reported slight long-term trends in COL frequency over Europe for the period 1960–2017. Using the detection algorithm developed here, we find no evidence for a systematic long-term increase in the number of detected COLs on a yearly basis. Instead, the historical record is dominated by pronounced inter-decadal variability, with periods such as the 1970s exhibiting enhanced COL activity and persistence.

Nevertheless, we do find slight but statistically significant positive trends in the JJA and SON seasons (Fig. C.4e,f), suggesting that seasonal dynamical changes may influence COL frequency. The significance of these trends is however highly sensitive to the inclusion of the 1940–1950 period in the analysis. When restricting the analysis to the same period as Muñoz et al. (2020) (1960–2017), the positive trends disappear and become non-significant (see Fig. D.4 for details). Furthermore, when restricting the analysis to the post-satellite era (1980–2024), no statistically significant yearly trend in COL occurrence is detected (Fig. D.5). Depending on the season, the estimated trends can even become negative, as observed during MAM. We believe that this sensitivity to the selected analysis period further highlights the dominant role of inter-decadal variability in the historical evolution of COL occurrence.

The seasonal distribution of COLs found in this paper is consistent with previous studies (Nieto et al., 2005), with a lower frequency of COLs during winter. This behaviour has been attributed to weaker jet streams during summer, which allow greater meridional flow and increase the probability of COL formation (Nieto et al., 2005). The enhanced COL activity during the 1970s is mainly observed during MAM, JJA and SON, suggesting that the circulation changes associated with the 1970s climate regime shift primarily affected the seasons in which the background flow is already more favourable for COL formation. The spatial pattern associated with COL frequency (Fig. C.4b) is also very similar to (Nieto et al., 2005) with three main hot-spots identified west of the Iberian Peninsula, over Italy and over the Balkans/Black Sea.

To conclude, our algorithm also offers several advantages. First, it does not require the synoptic feature to exhibit a consistent spatial pattern that can be tracked over time, as is the case for analogue-based approaches. This makes the method adaptable to a wide range of COL structures, without imposing explicit spatial constraints. Second, the algorithm relies on input variables that are widely available, including in ESM outputs, which facilitates its application across different datasets and modelling frameworks. Finally, the sensitivity of the results to key algorithm parameters can be easily assessed, as demonstrated in this study. This allows for transparent robustness checks and strengthens confidence in the main conclusions.

C.4.2 Toward a Better Understanding of Physical Drivers Behind COL Impacts

While this study focuses on the detection of COLs and the attribution of extreme precipitation events to these systems, a broader understanding of the physical mechanisms controlling their impacts remains essential. In this section, we discuss several dynamical and thermodynamic drivers that may help interpret the observed results and guide future research.

A first important result is the 1970s peak which is observed across MAM, JJA and SON. This period has also been identified in previous studies as a phase of major circulation changes over the North Atlantic–European sector associated with a climate regime shift (e.g. (Dai et al., 2018; Sarkar and Maity, 2021)). Although the exact causes of this shift remain uncertain, it has been linked to changes in the large-scale background circulation. Dai et al. (2018) suggest that it likely resulted from a complex interplay between oceanic and atmospheric circulation variability. In addition, Sarkar and Maity (2021) hypothesize that Arctic amplification may have favoured a more meridional atmospheric circulation, potentially enhancing Rossby wave breaking and the persistence of cut-off lows, thereby contributing to an increase in extreme precipitation events over Europe.

Our findings are also consistent with recent studies that emphasize the dominant role of thermodynamic processes in shaping future changes in extreme precipitation over Europe. For instance, Pfahl et al. (2017), based on CMIP6 simulations, concluded that projected increases in European precipitation extremes are primarily driven by thermodynamic factors rather than by changes in large-scale circulation patterns. At the Northern Hemisphere scale, Crawford et al. (2025) also showed that projected increases in cyclone-associated precipitation are primarily driven by thermodynamic changes, whereas projected changes in cyclone frequency and dynamics contribute comparatively

little. Although their analysis considers all extratropical cyclones rather than specifically COLs, these findings are consistent with our results, which show increasing precipitation extremes despite the absence of a systematic long-term increase in COL frequency.

Because COLs are frequently associated with the most intense precipitation events in Europe, this internal variability likely contributes to the temporal evolution of extreme precipitation occurrence. Interestingly, the same threshold dependence is also obtained when considering all extreme precipitation events irrespective of their association with a COL (Appendix Fig. D.7). The past periods of enhanced COL activity are reflected in extreme precipitation days counts, but their irregular timing points to a strong contribution from internal climate variability. However, extreme precipitation over Europe is not exclusively driven by COLs. Other processes, such as atmospheric rivers (Eiras-Barca et al., 2021) and localized convective events (Lombardo and Bitting, 2024), also make an important contribution to the occurrence of extreme rainfall. The present study does not investigate the evolution of these processes. Instead, it quantifies how the occurrence and natural variability of COLs modulate extreme precipitation within the broader context of the projected intensification of precipitation extremes across Europe (Rajczak and Schär, 2017; Coppola et al., 2021).

Topography and regional differences in climate change responses may also further modulate the observed trends. Moustakis et al. (2020) showed that changes in extreme precipitation intensity across Europe are strongly conditioned by terrain characteristics and local-scale processes. Although our analysis considers Europe as a single domain, marked regional disparities can exist. For instance, extreme precipitation intensity is projected to decrease in parts of the Mediterranean basin, while increasing trends are reported over northern Europe and Scandinavia (Rajczak and Schär, 2017; Coppola et al., 2021). In this context, our results highlight that topography plays a key role in amplifying precipitation during COL events, both at continental and local scales.

Nevertheless, recent high-impact flood events illustrate that even without systematic changes in COL frequency, additional physical processes can substantially amplify the precipitation associated with individual COLs. Despite the absence of a systematic increase in COL frequency, several exceptionally damaging flood events have occurred during the past five years—notably in Belgium, eastern Spain (Valencia), and Central Europe—associated with record-breaking rainfall totals. Many of these events were linked to COLs, suggesting that additional physical mechanisms may have amplified their impacts beyond what would be expected from circulation changes alone. These mechanisms remain insufficiently understood and warrant further investigation. Recent work by Insua-Costa et al. (2022) demonstrated that approximately 50% of the moisture contributing to the July 2021 extreme precipitation event (daily precipitation on the 13th of July) in eastern Belgium and western Germany originated from local recycling, through evaporation from Central and Northern European forests. Such recycling processes are expected to become more effective under future warming conditions (Fernández-Alvarez et al., 2023), as several regional moisture sources are projected to release larger amounts of water vapour. This highlights the potential importance of land–atmosphere interactions in modulating precipitation intensity during extreme events.

To provide additional context, we briefly examine the evolution of total evapotranspiration across several European regions using ERA5 reanalysis data (Fig. C.9). Over recent decades, multiple regions exhibit statistically significant trends in total evapotranspiration during specific months, suggesting that land-surface processes may increasingly influ-

ence the atmospheric moisture available for high-impact precipitation events. In Central Europe, for example, total monthly evapotranspiration shows significant long-term increases in April, August, September and October over the 1940–2024 period. Other regions across Europe, such as the Baltic Sea and the North Sea, also exhibit significant positive evapotranspiration trends, particularly during April, May, and June (Fig. C.9). These trends suggest an increasing availability of moisture in some regions, which may contribute to a more humid lower troposphere during the most active COL seasons.

When combined with enhanced atmospheric instability under warming conditions, such as increased convective available potential energy (CAPE), these factors could favour more intense convective responses when large-scale dynamical triggers are present (Calvo-Sancho et al., 2026). In this context, Calvo-Sancho et al. (2026) showed that the Dana COL, which struck Valencia in October 2024 (see Fig. C.3b and d), produced approximately 20% more heavy precipitation than it would have under pre-industrial conditions, corresponding to an increase of about 55% in the spatial extent of the precipitation event. Their analysis indicates that this intensification is linked to enhanced moisture fluxes from the Mediterranean Sea (about +9%) as well as increased updraft velocities within the associated convective systems.

Taken together, COLs may act primarily as the synoptic-scale trigger, while the ultimate severity of extreme precipitation events may depend on a complex interplay between atmospheric circulation, land–atmosphere feedbacks, and thermodynamic conditions. These interactions remain insufficiently constrained and represent an important avenue for future research.

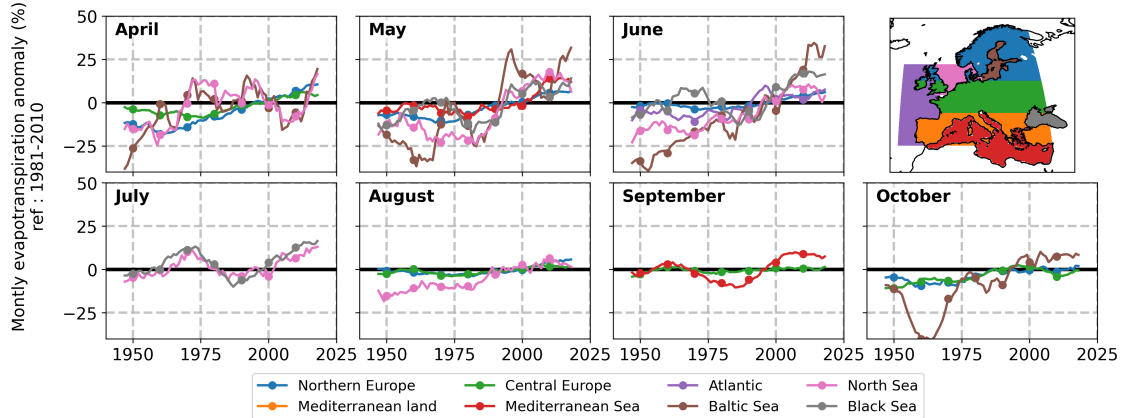


Figure C.9: Monthly anomalies of total evapotranspiration (%) over selected European regions based on ERA5 data, expressed relative to the 1981–2010 climatological mean. Linear trends are shown only where they are positive and statistically significant (p -value < 0.05), as determined using a Mann-Kendall test. A 15-year running mean is applied.

C.4.3 Implications of the results

We find that a large fraction of extreme precipitation days over Europe are associated with COLs (Fig. C.5a). A similarly dominant role of persistent synoptic-scale systems has also been reported over Australia by Barnes et al. (2023), although direct quantitative comparisons remain difficult because different weather systems and precipitation thresholds were considered. During summer, the proportion of COL-associated extreme precipitation

days can reach up to 68% (Fig. C.5i). Although the numbers of COL-associated days may appear relatively high, they should be interpreted in the context of the large European domain considered in this study and the relatively permissive precipitation threshold used to associate the event with a COL. These findings indicate that accurately representing COL occurrence is essential for future assessments of hydroclimatic risk over Europe.

Our results also highlight the dominant role of natural variability in modulating the annual frequency of COLs over the 1940–2024 period. This strong inter-decadal variability, also reported in recent studies (Brajkovic et al., 2025; Hundhausen et al., 2024; Kendon et al., 2023; Jain et al., 2023), implies that future climate conditions may still be characterized by extended periods of both enhanced and reduced COL activity. Because COLs represent one of the primary synoptic drivers of extreme precipitation in Europe, this variability is expected to translate directly into pronounced fluctuations in the occurrence of extreme precipitation events. Although natural variability dominates the timing of COL occurrence, the intensity of COL-associated precipitation is nevertheless expected to increase under global warming. This increased sensitivity of COL-associated precipitation to warming has already been demonstrated using pseudo-global-warming experiments (Thompson et al., 2025; Ludwig et al., 2023) or using model simulations (Ferreira, 2021).

All considered ESMs underestimate the occurrence of COLs over Europe, with particularly large biases over Southern Europe. Similar biases in the simulated spatial distribution of European COLs have also been reported by Mishra et al. (2025). More generally, comparable biases have been identified in CMIP6 simulations of storm-track densities (Priestley et al., 2023), and have been linked to deficiencies in the representation of atmospheric blocking over Europe (Zappa et al., 2014). In addition, diabatic processes play a key role in the life cycle of COLs (Portmann et al., 2018). Priestley et al. (2023) showed that coupled air–ocean interactions enhance cyclone moisture uptake and latent heat release, whereas latent heating associated with condensation is still believed to be underestimated in many CMIP6 ESMs. Model horizontal resolution may also influence the representation of COLs by improving the simulation of horizontal temperature gradients and synoptic-scale circulation (Bracegirdle et al., 2021; Baker et al., 2019; Priestley et al., 2020; Zappa et al., 2013). However, our results indicate that horizontal resolution alone cannot explain the differences in model performance. The absence of a simple relationship between horizontal resolution and model performance suggests that deficiencies in model physics and dynamics also contribute substantially to the simulated biases.

The underrepresentation of COLs is also reflected in projections of COL-associated extreme precipitation. Among the selected models, EC-Earth3-Veg exhibits trends most consistent with those derived from ERA5. Nevertheless, no ESM is able to reproduce the recent observed levels of COL-associated extreme precipitation. In most cases, present-day ERA5 values are only reached—or even exceeded—by the end of the 21st century in EC-Earth3-Veg under the SSP5-8.5 scenario. This finding motivated our focus on this high-emissions scenario, as even under such forcing, simulated COL-related extremes remain lower than those currently observed. Moreover, the SSP5-8.5 scenario is particularly useful because it maximizes the climate change signal and therefore facilitates the comparison between the selected ESMs. Nevertheless, the conclusions of this study are not specific to this scenario. Simulations performed under the SSP1-2.6 scenario exhibit a similar behaviour, characterized by strong inter-decadal variability in COL-associated precipitation and by EC-Earth3-Veg providing the closest agreement with the

ERA5-based attribution. These additional results are presented in Appendix Fig. D.8. This underrepresentation implies that current climate projections may underestimate the contribution of COLs to future European precipitation extremes.

Topography emerges as a key modulating factor of precipitation intensity during COL events. Mountainous regions such as the southern Dinaric Alps consistently appear as hotspots of COL-related precipitation impacts. However, COLs are also capable of triggering intense frontal precipitation that affect flatter regions, including the Netherlands and Belgium. This indicates that, although some regions are more frequently exposed because of their geographical setting, severe COL-associated precipitation can also occur over lowland regions across Europe.

Finally, these results emphasize the need for further research into the drivers of COL variability and the mechanisms leading to periods of clustered extreme events. In particular, future work should investigate potential low-frequency modes of variability in COL occurrence, as well as the combined influence of land–atmosphere feedbacks, moisture availability, and changes in atmospheric stability under climate change. Improving our understanding of these interacting processes will be essential for better anticipating the timing, frequency, and severity of future large-scale precipitation extremes across Europe.

C.5 Conclusions

In this study, we developed an objective algorithm based solely on 500 hPa geopotential height fields to detect COLs and to systematically attribute extreme precipitation events to their potential synoptic influence over Europe. Applied to the ERA5 reanalysis over the 1940–2024 period, this approach allowed us to construct a long-term climatology of COL occurrence and to quantify their contribution to extreme precipitation events using a consistent and transparent attribution framework.

Our results show that the total number of COLs over Europe exhibits strong inter-decadal variability but no statistically significant long-term trend. Nevertheless, COLs are found to be associated with a substantial fraction of European precipitation extremes, accounting for up to approximately half of the most extreme daily events, depending on the precipitation threshold and season considered. Importantly, while moderate COL-related precipitation events show no significant temporal trend, the most intense COL-associated extremes display statistically significant increases, particularly in summer. This threshold dependence highlights that changes in extreme precipitation cannot be inferred from COL frequency alone and points toward a dominant thermodynamic contribution to the observed intensification of extreme events.

When applied to six CMIP6 ESMs, the detection algorithm reveals a systematic underestimation of COL occurrence, especially over the Mediterranean basin, and a northward displacement of simulated COL activity compared to ERA5. Although some models reproduce the sign of observed trends in COL-associated extreme precipitation, none are able to reproduce present-day reanalysis levels, even under the high-emissions SSP5-8.5 scenario. These results suggest that current-generation ESMs struggle to accurately represent both the occurrence of COLs and their associated precipitation impacts, which has important implications for the interpretation of future projections of extreme precipitation over Europe.

Overall, this study suggests that the observed increase in European COL-associated extreme precipitation is not necessarily driven by an increase in COL frequency itself, but is instead consistent with a growing influence of thermodynamic factors. This interpretation is supported by the absence of a systematic increase in observed COL occurrence together with the documented increase in extreme precipitation intensity reported in previous studies. At the same time, the strong role of natural variability implies that periods of clustered high-impact COL events may continue to occur, even in the absence of clear long-term dynamical trends. Nevertheless, because current-generation ESMs systematically underestimate the occurrence of COLs over Europe during the historical period, confidence in projections of future COL-associated extreme precipitation remains limited. Improving the representation of COL dynamics in climate models therefore constitutes an important priority for future research.

The proposed detection and attribution framework provides a flexible and transparent tool that can be readily applied to other regions, reanalysis products, or climate model outputs. Future work should focus on improving the representation of COL dynamics in climate models and on better understanding the interactions between synoptic forcing, land–atmosphere feedbacks, and atmospheric instability that ultimately control the severity of COL-related extreme precipitation events.

Code availability

The cut-off low (COL) detection algorithm developed in this study is publicly available at <https://doi.org/10.5281/zenodo.19604862>.

The archived version provides the full Python implementation of the method, together with a documented example script and sample datasets enabling reproducibility of the detection workflow.

Data availability

ERA5 reanalysis data are available from the Copernicus Climate Data Store (CDS). CMIP6 data are available via the Earth System Grid Federation (ESGF).

Author contributions

Josip Brajkovic, Sébastien Doutreloup and Nicolas Ghilain conceived the study. Josip Brajkovic developed the algorithm, performed the analysis and led the writing of the manuscript. Nicolas Ghilain and Sébastien Doutreloup contributed to the manuscript structure and writing. Pierre Archambeau provided input on both the methodology and the manuscript.

APPENDIX **D**

Second paper: supplementary material

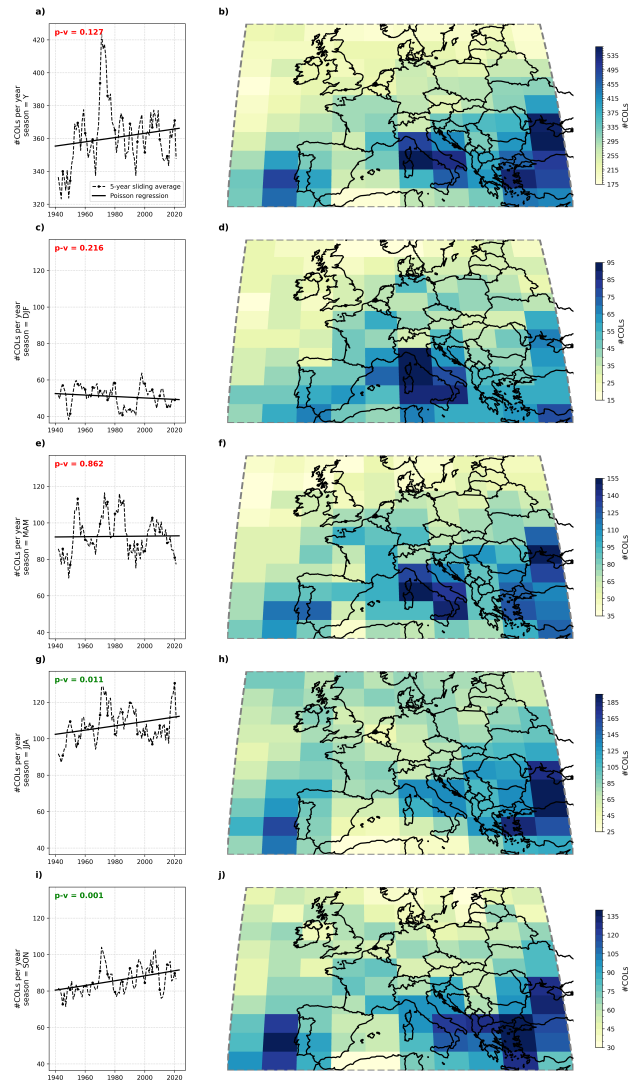


Figure D.1: This figure is associated with Figure C.4 of the original paper. (a) Annual count of COLs per year and (b) spatially cumulated count over the same period. (c–j) Same as (a–b) but for the other seasons.

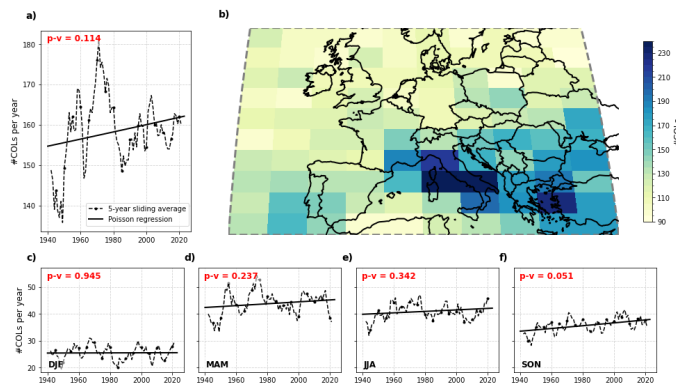


Figure D.2: Same as Fig. C.4, but after applying a spatio-temporal declustering procedure. Two detections are considered to belong to the same COL if their centres are located within 500 km of each other and occur within a 7-day period. Each cluster is therefore counted only once.

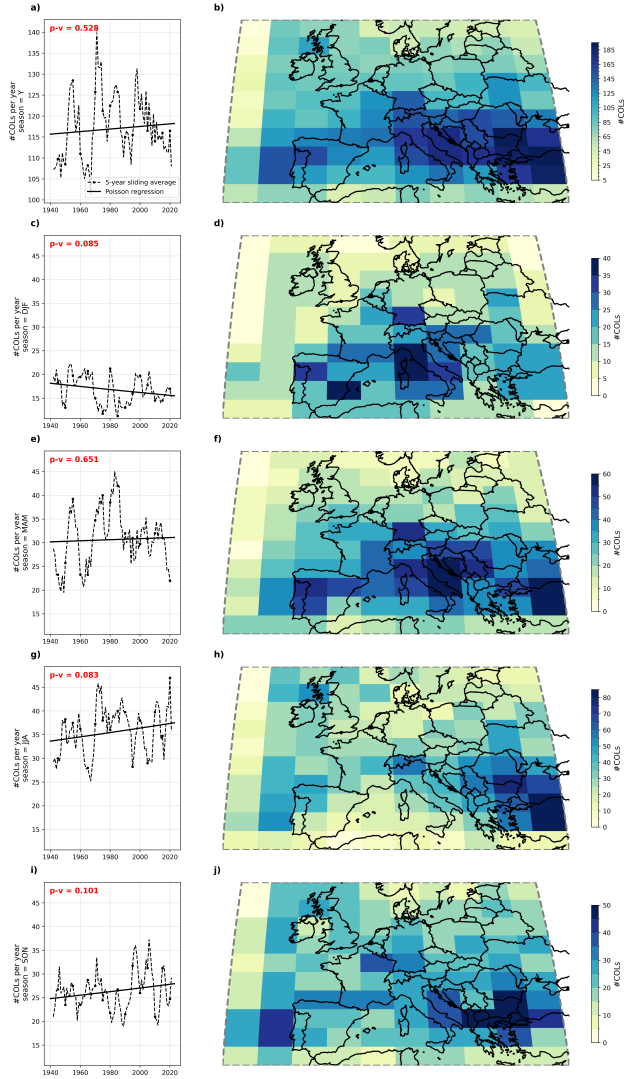


Figure D.3: Same as Fig. D.1, but with $A_{\text{crit}} = 10^6 \text{ km}^2$.

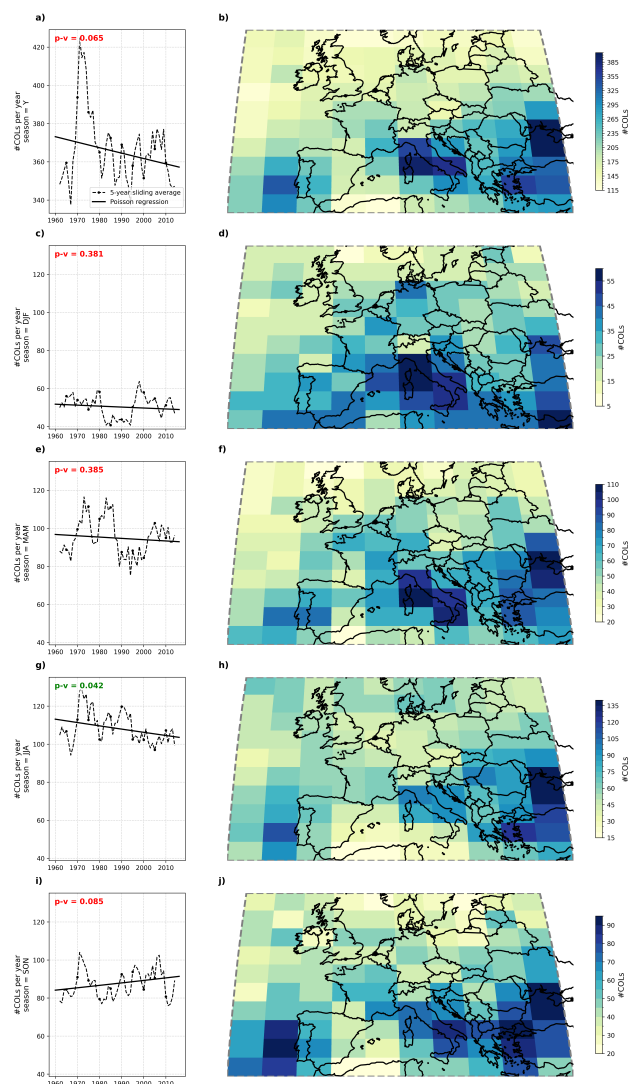


Figure D.4: Same as Fig. C.4, but over 1960–2017.

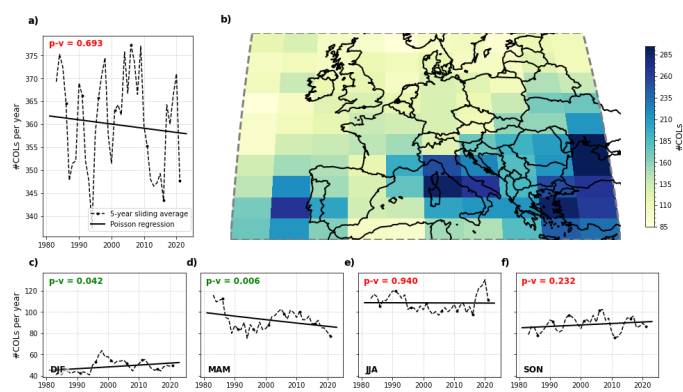


Figure D.5: Same as Fig. C.4, but over 1980–2024.

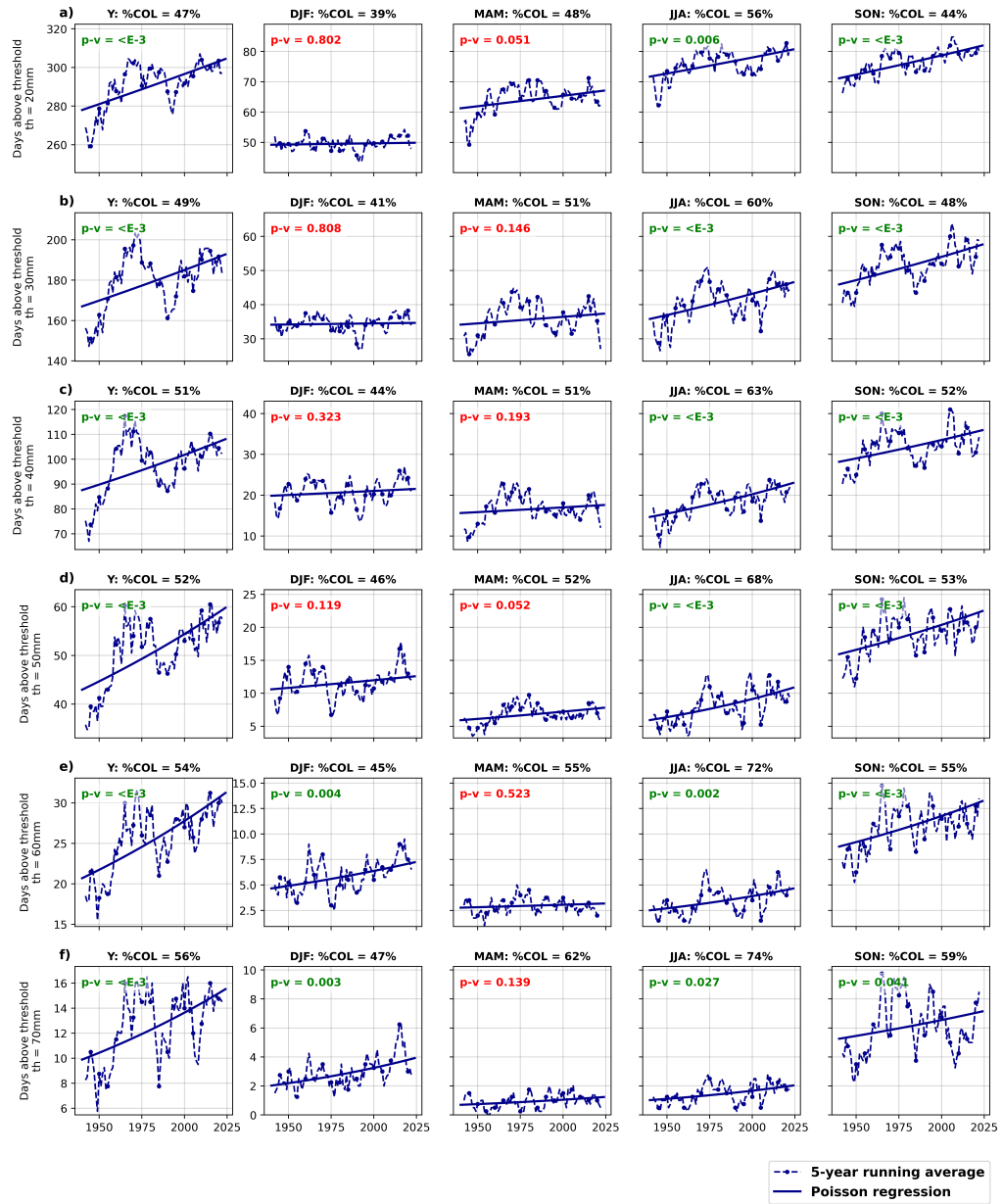


Figure D.6: Same as Figure C.4 of the original paper, but with different precipitation thresholds (a–f).

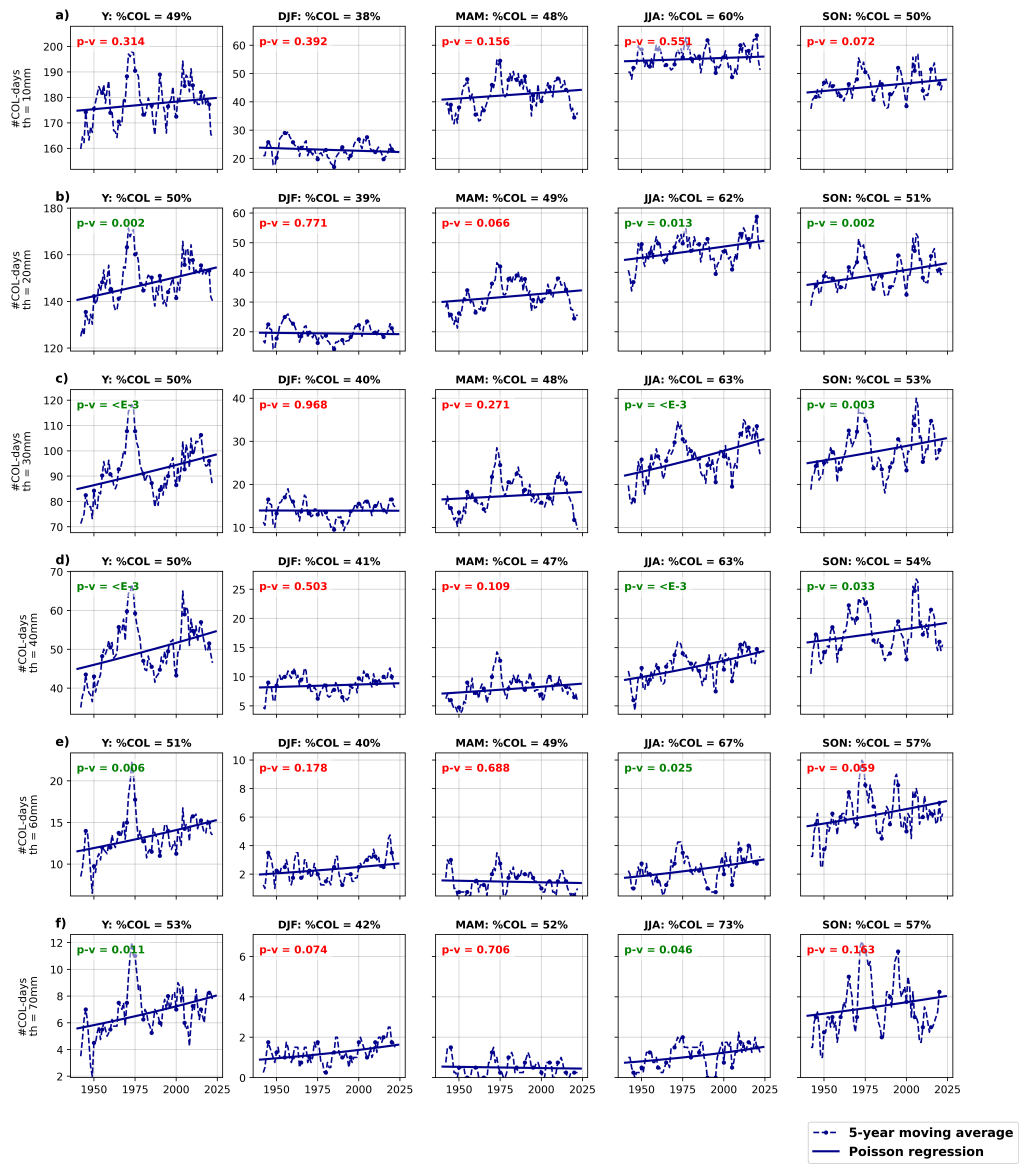


Figure D.7: Same as Figure C.5 of the original paper, but with different precipitation thresholds (a–f) and for all the days regardless of their association with COIs.

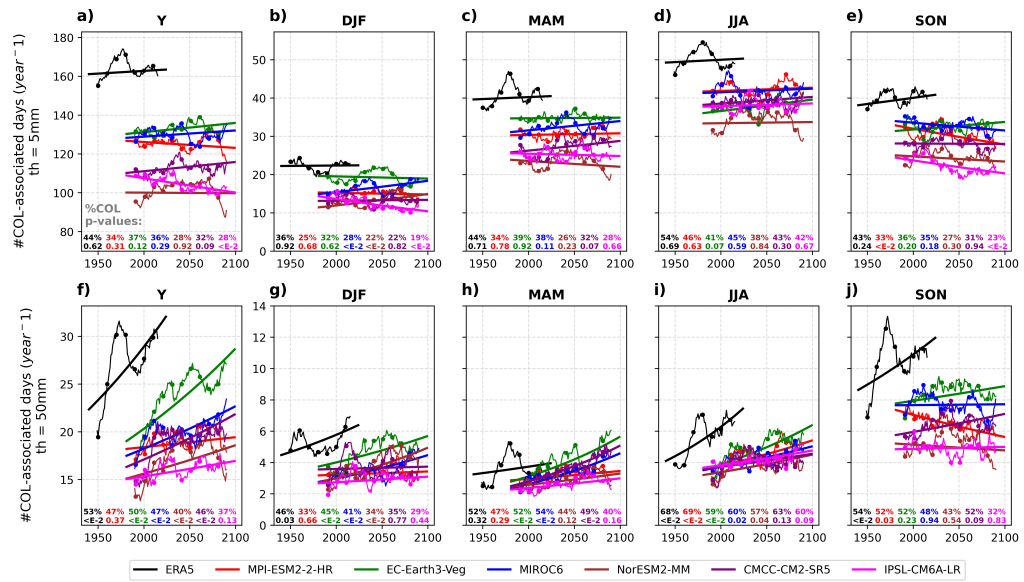


Figure D.8: Same as Figure C.8 of the original paper, but with the scenario SSP1-2.6.

APPENDIX E

Chapter 4: Additional elements

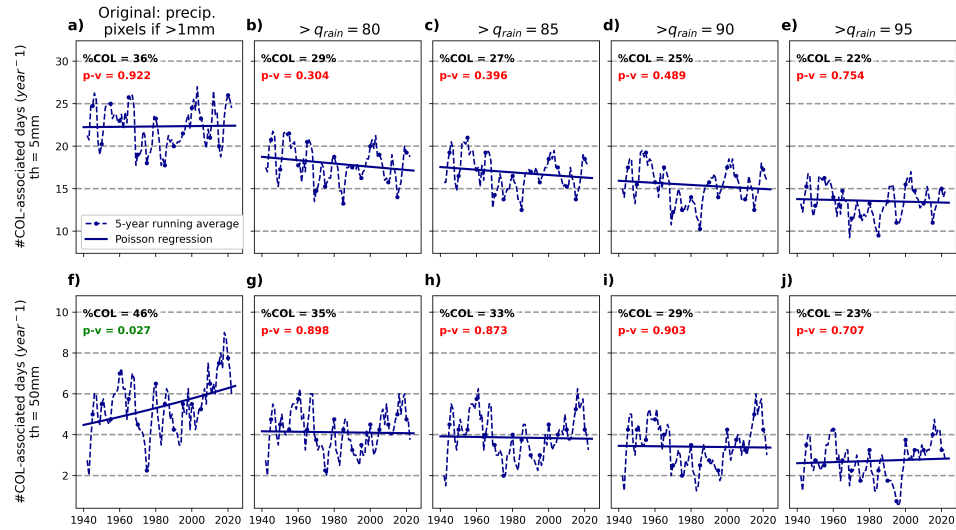


Figure E.1: Same as Fig. 4.7, but for DJF.

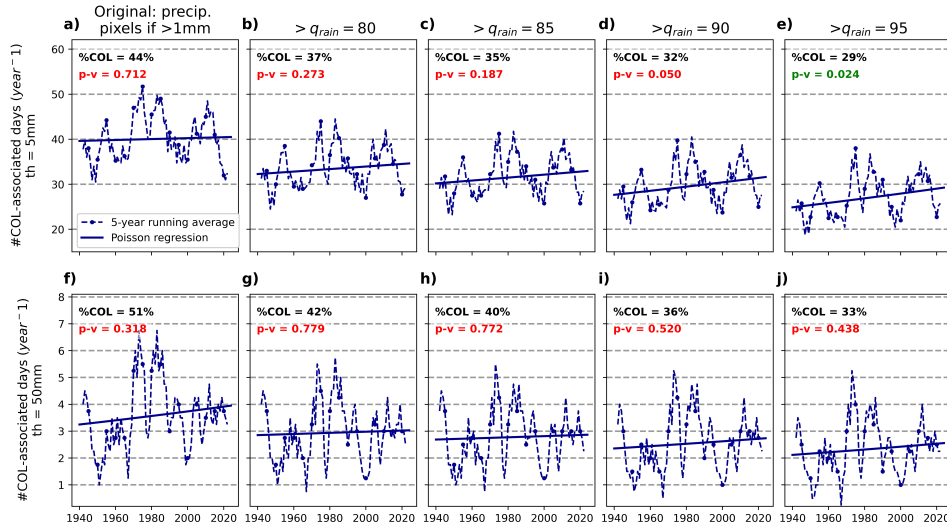


Figure E.2: Same as Fig. 4.7, but for MAM.

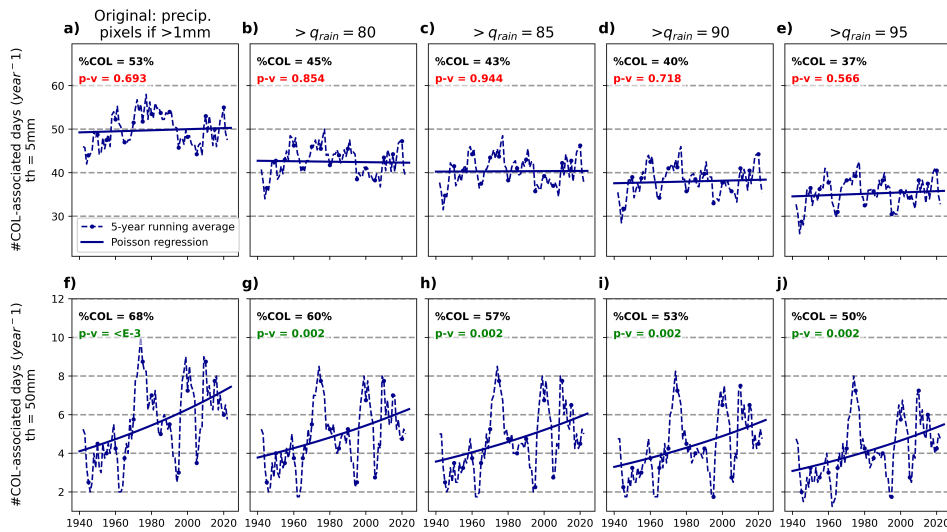


Figure E.3: Same as Fig. 4.7, but for JJA.

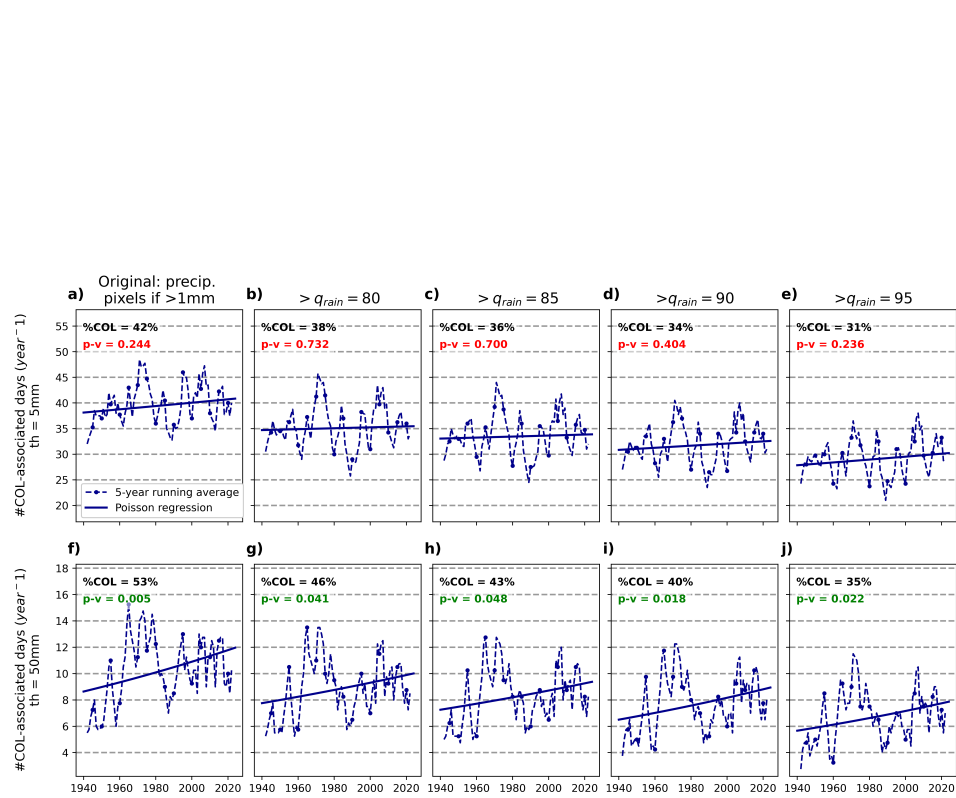


Figure E.4: Same as Fig. 4.7, but for SON.

APPENDIX F

Chapter 5: Additional elements

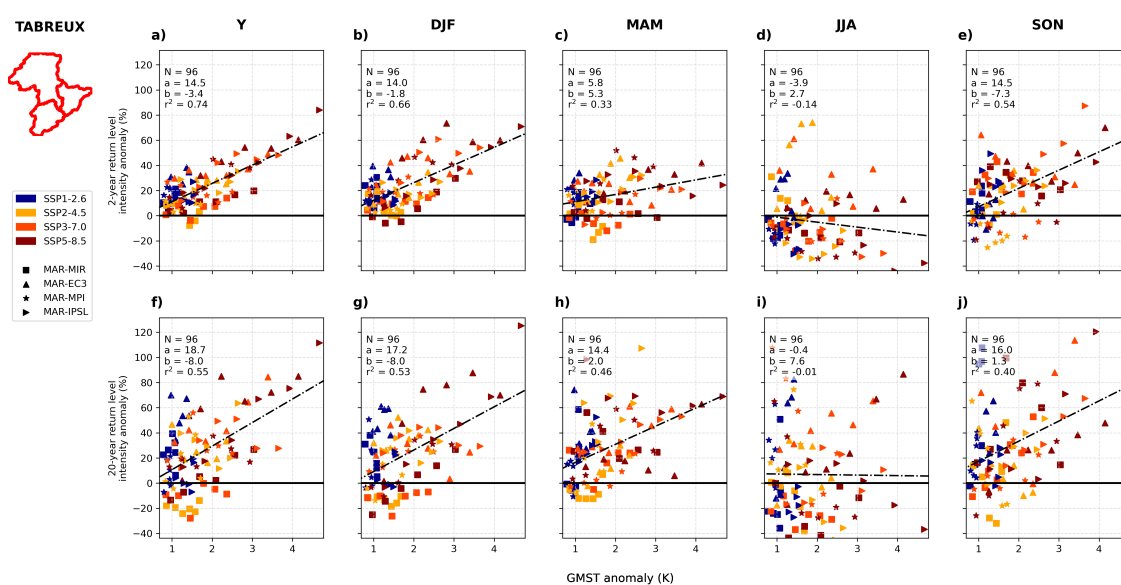


Figure F.1: Same as Fig. 5.13, but for all seasons.

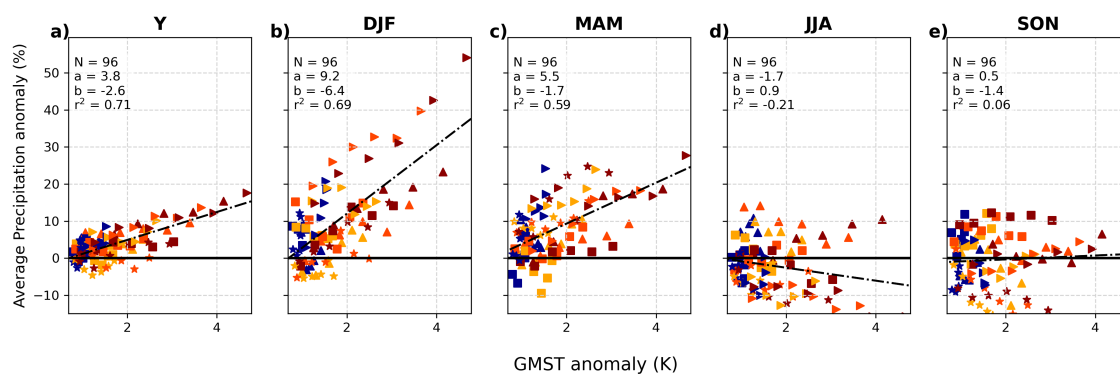


Figure F.2: Anomalies of the Average precipitation over the Ourthe catchment at Tabreux as a function of the Global Mean Surface Temperature Anomaly (GMST). Results are shown for the whole year (Y; a), DJF (b), MAM (c), JJA (d) and SON (e). Linear regressions of slope (a) and intercept (b) are fitted to the (N)-point data clouds, and the corresponding coefficient of determination (r^2) is reported

List of publications

Publications in peer reviewed journals

Brajkovic, J., Delhasse, A., & Fettweis, X. (2023). Impact du contenu en eau sur la capacité de rétention simulée du manteau neigeux de la calotte du Groenland. *Bulletin de la Société Géographique de Liège*, 80(2023/1), 5–17. <https://doi.org/10.25518/0770-7576.7061>

Brajkovic, J., Fettweis, X., Noël, B., Van De Vyver, H., Ghilain, N., Archambeau, P., Piroton, M., & Doutreloup, S. (2025). Increased intensity and frequency of extreme precipitation events in Belgium as simulated by the regional climate model MAR. *Journal of Hydrology: Regional Studies*, 59, 102399. <https://doi.org/10.1016/j.ejrh.2025.102399>

Brajkovic, J., Archambeau, P., Ghilain, N., and Doutreloup, S.: Past and future evolution of cut-off-low-associated extreme precipitation in Europe, *EGUsphere* [preprint], <https://doi.org/10.5194/egusphere-2026-2315>, 2026.