Can we trust CMIP5/6 future projections of European winter precipitation?

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Abstract

IPCC models project a likely increase in winter precipitation over northern Europe under a high-emission scenario. These projections, however, typically rely on relatively coarse ~100 km resolution models that can misrepresent important processes driving precipitation, such as extratropical cyclone activity, and ocean eddies. Here, we show that a pioneering 50 km atmosphere–1/12° ocean global coupled model projects a substantially larger increase in winter precipitation over northwestern Europe by mid-century than lower-resolution configurations. For this increase, both the highest ocean and atmosphere model resolutions are essential: only the eddy-rich (1/12°) ocean projects a progressive northward shift of the Gulf Stream. This leads to a strong regional ocean surface warming that intensifies air–sea heat fluxes and baroclinicity. For this then to translate into a strengthening of North Atlantic extratropical cyclone activity, the 50 km atmosphere is essential, as it enables enhanced diabatic heating from water vapor condensation and an acceleration of the upper-level mean flow, which weaken vertical stability. Our results suggest that all recent IPCC climate projections using traditional ~100 km resolution models could be underestimating the precipitation increase over Europe in winter and, consequently, the related potential risks.

1. Introduction

Winter precipitation is projected to increase over northern and central Europe by the end of the century (likely under the Representative Concentration Pathway, RCP8.5), related to enhanced atmospheric moisture, moisture convergence, and extratropical cyclone activity [1]. Yet, the magnitude of the change in North Atlantic extratropical cyclone activity and its particular contribution to that precipitation increase remain uncertain [1–3]. This is due partly to the sensitivity to model resolution of key processes driving precipitation and atmospheric circulation changes at local and hemispheric scales (e.g. [1, 3–5]). Assessing such sensitivity in future projections is an important step, first, to reduce current uncertainty levels, second, to evaluate the impact of model biases on the climate change response to a greenhouse gas (GHG) increase, and, last, to provide more reliable projections to the development and implementation of more effective adaptation and mitigation policies (e.g. [6]).

Sensitivity studies have previously examined the influence of finer model resolutions in the atmosphere and the ocean on the representation of the mean state, variability, and future change of the mid-latitude large-scale atmospheric circulation and precipitation. Increased atmosphere resolution (usually from ~100 km to ~25–50 km) can strengthen extratropical cyclones and heavy precipitation in present-day climate simulations, likely because of a better resolved cyclone mesoscale structure [4, 7] and topography [8]. Similarly, increased atmosphere resolution can raise the sensitivity of storm tracks to a future increase in GHG forcing [4, 9–12] and of the transient eddy heat and moisture fluxes to surface ocean warming [13]. Increased ocean resolution
(from \(\sim 100 \text{ km} \) to \(\sim 10 \text{ km} \)) can also strengthen present-day extratropical cyclones and precipitation and shift the eddy-driven jet and storm tracks northward [5, 14–19]. In the light of these findings, more studies have lately advocated for increasing resolution in all model components beyond the traditional \(\sim 100 \text{ km} \) resolution used in most IPCC models for diverse applications, such as climate prediction and climate change studies (e.g. [4, 15, 20–23]). To date, most studies of future climate change at higher resolution (a few tens of kilometers) have relied on atmosphere-only global or regional models, due to the computational cost of performing long, fully coupled global climate projections at such resolutions; however, these models risk providing a partial view of the climate system (e.g. [4, 5, 7, 12–19]). Our study takes a step forward: the sensitivity to model resolution of precipitation changes over Europe in climate projections is studied in a hierarchy of general circulation models (GCMs) at a horizontal resolution that ranges from a conventional \(\sim 100 \text{ km} \) atmosphere–\(1^\circ \) ocean to a pioneering \(50 \text{ km} \) atmosphere–\(1/12^\circ \) ocean, considerably finer than in other IPCC projections.

2. Results

We use ensembles of simulations generated with the global coupled model HadGEM3-GC3.1 in the context of CMIP6 HighResMIP activities [24] and the H2020 PRIMAVERA project [25]. The simulations have five different resolutions: LL, MM, HM, MH, and HH (for low, L, medium, M, and high, H, resolution, with the first and second letters indicating the atmosphere and ocean resolutions respectively; see section 4 and [26] for more details). This experimental setup is conceived to disentangle the roles of the atmosphere and ocean resolutions in the representation of physical processes in climate models and their response to a GHG forcing increase. The simulations cover the period 1950–2050 under the shared socioeconomic pathway 5–8.5 (section 4). A previous model evaluation study found no major drifts in the surface climate in the control simulations at all resolutions, and reduced biases in the North Atlantic sea-surface temperatures (SSTs) and precipitation with increased resolution (section 4 and [26]). Here we compare changes in the winter climate (December–February; DJF) between the periods 2030–2050 and 1960–1980 (section 4).

2.1. Sensitivity of future precipitation changes to model resolution

Winter precipitation increases over large areas of the North Atlantic and northern and central Europe by 2050 at all resolutions (figure 1), consistent with the projected increase in previous-generation models in the IPCC 5th Assessment Report (AR5) [1]. This increase can to a large extent be attributed to rising GHG concentrations wherever anomalies exceed the 5th–95th range of internal variability in the control simulations (white masking in figure 1; section 4). The precipitation anomaly pattern, however, changes with resolution. Over the western North Atlantic, all resolutions project a north-northwestward shift in precipitation from the central subtropical North Atlantic to the Gulf Stream area. The shift becomes sharper and the anomalies deeper as the ocean resolution increases, from the non-eddy resolving (LL) to the eddy-present (MM and HM) and eddy-rich model resolutions (MH and HH). Increased precipitation over the Gulf Stream is the largest in MH and HH, and the smallest in LL, which instead shows a widespread increase from the Gulf Stream into Northern Europe.

Over the eastern North Atlantic and Europe, the largest increase in precipitation is projected in HH, with anomalies exceeding values at all the lower resolutions (stippling in figure 1(a)), as well as the projected change by the end of the century in both fully-coupled CMIP5 IPCC GCMs [1] and regional models [7, 10] and by mid-century in nearly all other CMIP6 and HighResMIP model projections (figure S1 (available online at stacks.iop.org/ERL/16/054063/mmedia)). Precipitation in HH substantially increases over the British Isles and surrounding seas, NW France, and the Alps. Changes in heavy precipitation (section 4) exhibit a similar pattern, and the largest increase in extreme events is also projected in HH (figure S2). Increased resolution in both the atmosphere and ocean is crucial for simulating the NW Europe precipitation increase. Over the Alps and British Isles, for example, the large increase in HH is not reproduced if a lower resolution in either the ocean (HM) or atmosphere (MH) is used (figure 1). These results suggest that models at a resolution lower than HH, including the traditional 100 km version, might considerably underestimate the projected winter precipitation increase over Europe.

2.2. Changes in the North Atlantic large-scale oceanic and atmospheric circulations

We investigate the physical processes leading to the exceptional precipitation increase in HH. We begin in the Gulf Stream region, where the strong low-level temperature contrast is key for the development of the extratropical cyclones (e.g. [27–29]) that later impact western Europe and deliver \(\sim 80\%–90\% \) of the net winter precipitation [30]. The Gulf Stream region next to Cape Hatteras undergoes an intense warming in MH and HH, both with an eddy-rich ocean model (figures 2 and 3). The warming outpaces the ones at lower ocean resolutions around the 2010s in HH and around the 2020s in MH, and it is on average \(\sim 2 ^\circ \text{C} \) (and locally up to \(\sim 7 ^\circ \text{C} \)) warmer by 2050. The magnitude of the Gulf Stream warming in MH and HH is consistent with that in observations in the Extended Reconstruction SST version 5 [31]
between the late 1990s and 2020, albeit it is somewhat larger in HH (figure 3). The observed warming is, by contrast, on the upper edge of that at lower resolutions in the early 21st century, which suggests that lower resolution models might be underestimating its magnitude.

Gulf Stream warming similar to that in HH and MH was shown in doubling-CO$_2$ simulations performed with the GFDL’s GCM at a HH-like resolution ($\sim$10/50 km in the ocean/atmosphere) but not at a lower, $\sim$100 km resolution [32]. The authors linked the warming to a northward shift in the Gulf Stream and slope waters in response to a weakening in the AMOC and North Atlantic deep western boundary current [32]. A similar Gulf Stream northward shift (of 2–3$^\circ$) is also projected both in MH and HH (figure 2) following an AMOC weakening by mid-century [33]. The Gulf Stream is, by contrast, positioned up to 3$^\circ$ farther north of Cape Hatteras at lower ocean resolutions (L and M) in the late-20th century, and its position stays unchanged over the 21st century (figures 2 and S3). As a result, LL, MM, and HM show weaker SST warming in the Gulf Stream compared to MH and HH. The magnitude of future Gulf Stream warming in the eddy-rich ocean model therefore crucially depends on the present-day representation of the Gulf Stream position and on its sensitivity to GHG forcing.

The warming in the Gulf Stream area and the associated strengthening of the low-level temperature gradient (not shown) are expected to enhance extratropical cyclone development by increasing baroclinic instability (e.g. [28, 29]) and, thereby, to increase precipitation farther downstream (e.g. [30]).

But, why do MH and HH show different precipitation responses for a similar Gulf Stream warming? To quantify changes in extratropical cyclone activity, we use the maximum Eady growth rate (EGR), which gives a measure of baroclinic instability, and storm tracks (section 4). Changes in EGR (figure 4(a)) suggest a northward shift in the western North Atlantic over the Labrador Peninsula, likely connected with the Gulf Stream warming at both resolutions. However, it is only in HH that positive EGR anomalies extend downstream from the Labrador Peninsula to the British Isles and the North Sea, the same regions exhibiting exceptional precipitation increase (figure 1(a)). None of the projected changes at lower
Figure 2. Projected Gulf Stream response and the associated SST warming. Upper panels: change in yearly SST (in °C) between 2030–2050 and 1960–1980. The stippling in (a) and the white shading are as in figure 1. Lower panels: change in the latitudinal position of the Gulf Stream, illustrated through the time evolution of the maximum gradient in the yearly sea-surface height, averaged between 70° W and 74° W (units in 10⁻⁶ m/m; area enclosed by a dashed black square in (e), upper panel). The dashed line indicates the mean position in the AVISO satellite observations for the period 1993–2018.

Resolution show comparable changes in EGR over the central and western North Atlantic and NW Europe (figures 4(a) and S4). The most consistent response at all resolutions is an EGR weakening around Iceland and the Nordic Seas, likely related to a northward retreat of the sea ice edge over the 21st century (e.g. [34, 35]).

The storm tracks strengthen over the subpolar North Atlantic at all resolutions (figures 4(b) and S4), in agreement with previous-generation and
Figure 3. Time evolution of the Gulf Stream SST in model simulations and observations. Yearly SST anomaly (in °C, compared to the 1960–1980 period) on average in the Gulf Stream region, between 60°W–80°W and 35°N–45°N. In LL, MM, and HM, thick lines are the ensemble means, while the shading represents the ensemble spread. The single MH and HH runs are bolder, and red and blue respectively. The black, dashed line shows the SST anomaly in ERSST version 5.

State-of-the-art CMIP models [1, 2, 36]. The largest increase is again projected in HH, especially at the genesis region over the Gulf Stream and over the eastern North Atlantic, where storms are more likely to impact Europe (figure 4(b)). Changes in both the North Atlantic EGR and storm tracks thus point to a larger increase in extratropical cyclone activity in HH compared to MH for a similar Gulf Stream warming. This warming directly drives the atmospheric response by increasing near-surface baroclinicity over the Gulf Stream area (figure S5). However, as the warming appears on its own unable to explain the differences in cyclone activity and precipitation, especially over the eastern North Atlantic and western Europe, we extend the comparison to two key variables that can contribute to enhancing extratropical cyclone development: the diabatic heating and the strength of the upper-level zonal mean flow [28, 29].

Atmospheric diabatic heating accounts for latent heat release during water phase changes (section 4). Positive diabatic heating anomalies are thus expected in regions of increased precipitation, and vice versa, due to increased condensation and, hence, latent heat release. In an extratropical cyclone, latent heat release from condensing water normally works to amplify and sustain both the lower-level cyclonic and the upper-level anticyclonic perturbations, which together fuel cyclone development especially over the eastern North Atlantic (e.g. [2, 7, 28, 29]). Increased diabatic heating can therefore be expected to enhance further extratropical cyclone development [7, 15, 16]. This is the case of HH, in which positive diabatic heating anomalies extend over the eastern North Atlantic and British Isles, where EGR increases as well (figures 4(a) and (c)). Compared to HH, lower resolutions show generally weaker or no change in diabatic heating in the eastern North Atlantic, consistent with the lack of significant changes in their eddy activity (figures 4(c) and S4). This suggests that diabatic heating increases with atmosphere resolution, likely because of better resolved mesoscale structures in extratropical cyclones [4, 7]. This increase can hence contribute to the strengthening of the extratropical cyclone activity in HH compared to MH.

Enhanced diabatic heating is tied to the Gulf Stream warming and the associated increase in turbulent air–sea heat flux (figure S6). The direct linear response to coastal warming is an increase in winter precipitation over the region (figure 1), related to strengthened atmospheric vertical motion (not shown) and convective precipitation (partially shown in figure 1) (e.g. [37, 38]). Further downstream, the Gulf Stream warming provides a heat source for additional latent heat release and, thereby, extratropical cyclone development. This process is, in fact, non-linear due to the Clausius–Clapeyron relationship, with a warmer atmosphere holding even more moisture (not shown) and thus having a larger heating potential from condensing water [4, 39, 40]. We note that the other projection with the finest atmosphere resolution, HM, shows no enhanced storm activity compared to MM (figure S4)—both with the same ocean resolution but with MM having a coarser atmosphere. We argue this is because HM lacks the extreme coastal warming found in HH. Thus, although the Gulf Stream warming alone is insufficient to explain the exceptional increase in cyclone activity and precipitation in HH, it is necessary for its development thanks to the associated strong surface heating (figure S6) and the related increase in baroclinicity over the region (figures 4, S4 and S5).
Figure 4. Changes in the large-scale atmospheric circulation in eddy-rich ocean model simulations. Change between 2030–2050 and 1960–1980 in the (a) DJF maximum EGR at 700 hPa (in day$^{-1}$), (b) DJF storm tracks (in Pa), (c) DJF atmospheric diabatic heating averaged between 850 hPa and 250 hPa (in K d$^{-1}$), and (d) zonal wind at 250 hPa (in m s$^{-1}$) in HH (left) and MH (right). Contours show the 1960–1980 climatological mean. The stippling in HH and the white shading are as in figure 1.

In tandem with enhanced diabatic heating, the different response of the eddy-driven jet can also contribute to explaining the larger increase in extratropical cyclone activity and hence precipitation over the North Atlantic and NW Europe in HH compared to MH for a similar Gulf Stream warming. An intensified upper-level mean flow can strengthen cyclogenesis through increased vertical wind shear and hence baroclinicity (e.g. [4, 7, 28, 29, 41–43]). The upper-level mean flow increases substantially more in HH than in MH (figure 4(d)) and lower resolutions (figure S4) over the North Atlantic and NW Europe. This results from strengthened eddy–mean-flow transfer of momentum at the core of the jet over the central North Atlantic by enhanced eddies in HH compared to lower resolutions (figures S7 and S8). Increases in cyclone activity and in the strength of the upper-level mean flow have been linked through a nonlinear feedback with atmosphere model resolution, where enhanced latent heat release in better-resolved storms strengthens EGR (figure 4) and increases the vertical wind shear that favors cyclone development [4]. The upper-level flow acceleration in HH is consistent with the increase in EGR, whose anomalies are co-located with those in the zonal wind at 250 hPa (figures 4(a) and (d)), as
ensured by thermal wind balance. The exceptional increase in extratropical cyclone activity in HH is therefore likely related to increased diabatic heating and an acceleration of the upper-level mean flow. Although our analysis assumes contemporaneous changes in atmospheric variables and therefore cannot identify whether they decisively contribute to the intensification in the storm activity in HH nor their relative roles, it finds support in previous literature linking such changes to an increase in storm activity over the North Atlantic [4, 7, 28, 29].

Put together, our results show that winter precipitation over NW Europe increases notably more in HH than at lower resolutions due to increased extratropical cyclone activity in the North Atlantic. As summarized in figure 5, we argue that this is in first instance related to a Gulf Stream SST warming and enhanced surface heating and then to a more unstable atmosphere further downstream, linked to intensified diabatic heating and a stronger upper-level mean flow.

These results contrast with the traditional ‘tug-of-war’ view to predict future changes in storm tracks in winter. It is frequently argued that future storm tracks will weaken or strengthen depending on whether changes in the meridional temperature gradients at lower- or upper-levels dominate (e.g. [3, 34]). In our simulations, changes in the meridional temperature gradients as well as in the zonally averaged tropospheric temperature are of similar magnitude across resolutions, including HH (figure S9, and table S1). This shows that changes in the meridional gradients play a subsidiary role in the exceptional strengthening of the North Atlantic storm tracks in HH compared to lower resolutions.

3. Discussion

Our analysis suggests that traditional coarse resolution GCMs as well as regional models might be missing key processes implicated in future climate change. As summarized in figure 5, the North Atlantic and NW Europe climate seems more sensitive to GHG forcing in the HadGEM3-GC3.1 GCM at a very high (50 km atmosphere, 1/12° ocean) resolution than it is at lower ones, with winter precipitation and storm activity increasing considerably more by mid-century under a high-emission scenario. These changes are ultimately driven by Gulf Stream coastal warming, which strongly increases air–sea heat fluxes and baroclinicity. Precipitation changes over NW Europe are the largest when both the ocean and atmosphere are at their highest resolution and are always smaller in the lower resolution models. Previous sensitivity studies have partially shown similar changes in the storm tracks and Gulf Stream at increased model resolution and thus lend support to our conclusions (e.g. [7, 32]). In summer, the strong surface warming increases precipitation over the Gulf Stream region as well, especially in HH but, in contrast to winter, it drives no major changes in precipitation further downstream—likely because of the weaker summer eddy activity over the North Atlantic (not shown).

We acknowledge two limitations of this study: the relatively small ensemble sizes, and the use of a single climate model. This is especially the case for the highest resolution, whose enormous computational costs currently limit the ensemble to one member. It is important to highlight, however, that these simulations are performed at a groundbreaking resolution that demands computational capabilities within reach of few modeling centers, for which producing multi-member ensembles is still currently unattainable. Having larger ensembles or more models would be ideal to test our conclusions and investigate the impact of each model’s characteristics and their intrinsic climate variability, on the future response of European winter precipitation to rising GHGs. We expect climate projections performed with other
climate models within HighResMIP [24] to shed light on the robustness of our findings.

Climate projections at such a groundbreaking resolution will become essential for identifying vulnerability hotspots, especially on a regional to local scale. Our findings provide palpable evidence that conventional lower-resolution IPCC models could actually be underestimating future risks. For example, a larger increase in precipitation such as the one identified at the highest model resolution would pose a higher flooding risk [44] than now projected in the IPCC AR5 over European land areas (e.g. [45–47]). Similarly, more energetic extratropical storms and wind extremes in the North Atlantic would have important consequences for several economic sectors, such as trade, navigation, and wind energy production, which would be impacted more heavily than currently estimated (e.g. [48–50]). A complete mapping of the range of potential future changes in the North Atlantic and European winter climate is crucial for designing better informed and more effective adaptation and mitigation strategies. Climate projections at an eddy-resolving resolution as used in this study can thus change the paradigm of climate change risk assessment.

## 4. Materials and methods

### 4.1. HadGEM3-GC3.1

We use simulations produced with the coupled climate model HadGEM3-GC3.1 [26, 51]. This model comprises a GA7.1/GL7.1 atmosphere/land configuration based on the MetUM and JULES [52], a GO6 ocean [53] based on NEMO [54] and GSI8 sea ice based on CICE [55]. We refer to these works for a more detailed description of the model characteristics. Simulations are produced at five different resolutions (table 1).

| Naming convention | CMIP6 nominal resolution | Ensemble size |
|-------------------|--------------------------|---------------|
|                   | Atmosphere | Ocean | hist-1950 | highres-future |
| LL                | 250 km      | 100 km | 8         | 1          |
| MM                | 100 km      | 25 km  | 3         | 3          |
| HM                | 50 km       | 25 km  | 3         | 3          |
| MH                | 100 km      | 8 km   | 1         | 1          |
| HH                | 50 km       | 8 km   | 1         | 1          |

4.2. Model simulations

Following the CMIP6 HighResMIP protocol [24], the simulations use observed historical forcings over the period 1950–2014 (hist-1950) and follow the shared socioeconomic pathway 5–8.5 (SSP5-8.5) between 2014 and 2050 (highres-future). The SSP5-8.5 compares best with the RCP8.5 used by projections in the IPCC AR5 and CMIP5 [58]. Given the computational cost of running simulations at such high resolutions, the simulated period is shorter than the standard CMIP1850–2100 period [59].

Ensemble sizes at each resolution are given in table 1. Because of the higher computational costs of those configurations, only one ensemble member is available at the MH and HH resolutions. In addition to these simulations, we use data from 100 year control simulations to estimate the range of natural variability at each resolution (below). Both the historical and control simulations are run after a multi-decadal spinup [26], whereas the scenario runs follow the historical ones. Despite the relatively short spinup, the control simulations show a relatively stable surface climate, with most of the climate drifts being of smaller magnitude than the climate biases (compared to 1950 climate conditions; [26]). This is why no drift is removed from the fields studied here. Increased ocean resolution leads to reduced common SST biases, such as the cold one along the Gulf Stream and at subpolar latitudes in the North Atlantic [26]. Similarly, a dry precipitation bias [26] and a too-strong storm-track bias present in the North Atlantic for the standard configuration (not shown) are both reduced as resolution increases.

### 4.3. Data and code availability

Model data used in the analysis are available from the CMIP6 Earth System Grid Federation and can be located using information in [60–66]. Data and scripts to reproduce this work are uploaded to Zenodo at https://doi.org/10.5281/zenodo.3952782.

### 4.4. Statistical significance of future changes

At each model resolution, statistical significance of the anomalies between the future (2030–2050) and present-day (1960–1980) climate is evaluated based on the likelihood of a random occurrence of a similar change in their respective control simulation. In these control simulations, we randomly sample
two non-overlapping periods, each 21 years long, and compute first their mean and then their anomaly, as done between the future and historical simulations. This calculation is repeated 500 times to obtain a distribution of anomaly maps. The 5th and 95th percentiles of this distribution are then used as confidence levels to test significance at the 10% level. Similar results are found if the 2.5th and 97.5th percentiles are used instead (5% level) or if statistical significance is tested through a Student’s t test (not shown).

4.5. Data analysis
We compare changes in winter (December–February or DJF) climate between the periods 2030–2050 and 1960–1980. Considering the relatively short length of the simulation, the 21 year 2030–2050 period is a reasonable compromise to capture long-term climatic impacts of a GHG increase. The period 1960–1980 leaves the first 10 years for further climate stabilization, although the control simulations do not exhibit major drifts [26]; also, this period does not extend too far into the present so that it may mask the impact of the climate change signal.

The maximum EGR is computed with ESMValTool [67] from monthly data as

\[ \sigma = 0.31 f \frac{\partial \nu}{\partial z}, \]

where \( \nu \) is the buoyancy frequency, \( f \) the Coriolis parameter, \( \nu \) the magnitude of the horizontal wind, and \( z \) the vertical distance.

The storm tracks are calculated as the standard deviation of the 2–6 day band-pass filtered daily sea-level pressure.

The diabatic heating \( (Q_{\text{diab}}) \) is calculated from the thermodynamic equation as

\[ \frac{Q_{\text{diab}}}{C_p} = \frac{\partial T}{\partial T} + \nabla \cdot (\nu T) + \omega \left( \frac{\partial T}{\partial P} - \frac{RT}{C_p P} \right), \]

where \( T \) is the temperature, \( P \) the pressure, \( \nu = (u, \nu) \) the zonal and meridional wind components, \( \omega \) the vertical velocity in a pressure coordinate, and \( C_p \) the specific heat capacity and specific gas contrast of dry air respectively. On the right-hand side, the first term represents the temperature tendency, the second term the horizontal advection of temperature, and the last term the vertical advection of temperature (all in K per unit time). The residual, \( Q_{\text{diab}} \), accounts for radiative heating/cooling and latent heat release due to water phase changes. \( Q_{\text{diab}} \) is calculated from daily data between November and March using a central finite-difference scheme as an approximation of all the derivatives. A temporal average from December to February (winter) and a vertical one between 850 hPa and 250 hPa are eventually calculated.

The zonal and meridional components of the Eliassen–Palm flux are computed respectively as \( \nabla^T T \) for the eddy heat flux at 850 hPa and \( \nu^T \nu^T \) for the eddy momentum flux at 250 hPa; the \( \cdot \) indicates a 2–6 day band-pass filter, and the overbar a time average.

Heavy precipitation is defined as the 95th percentile at each individual grid point from daily data for the periods 1960–1980 and 2030–2050 of the historical and scenario simulations. The percentiles are defined across all precipitation and non-precipitation days to avoid a change in the number of wet days affecting the percentile level [68].

Data availability statement
The data that support the findings of this study are openly available at the following URL/DOI: https://doi.org/10.5281/zenodo.3952782.

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Author contributions
E M C, L P C and P O conceived the original idea. M J R provided the model data. S L T assisted with the data analysis in ESMValTool. E M C analyzed the data and wrote the manuscript with input from all authors.

Conflict of interest
The authors have no competing interests.

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