

European climate extremes intensify unevenly across warming levels with weak contemporaneous AMOC associations

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European climate extremes intensify unevenly across warming levels with weak contemporaneous AMOC associations

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Abstract:

European climate extremes intensify with global warming, but whether their interannual variability covaries with the Atlantic Meridional Overturning Circulation (AMOC) within comparable warming states remains uncertain. Here, we combine bias-corrected, statistically downscaled CMIP6 projections with corresponding AMOC simulations to assess six temperature and precipitation extreme indices across Europe at global warming levels of 1.5 °C, 2.0 °C and 3.0 °C under SSP5–8.5. An 18-model ensemble projects progressively stronger but spatially differentiated responses: heavy precipitation intensifies most strongly in northern Europe, warm-spell duration increases most markedly around the Mediterranean, and annual minimum daily maximum temperatures show larger increases than annual maximum daily maximum temperatures. Among 11 models with matched AMOC and climate-extreme realizations, the ensemble-mean AMOC anomaly at 26.5° N, relative to 1995–2014, becomes more negative from –2.81 Sv at GWL1.5 to –5.10 Sv at GWL3.0. We then test contemporaneous associations between annual AMOC anomalies and climate extremes after removing fitted linear trends within model-specific 20-year GWL windows. Area-weighted, autocorrelation-adjusted R^2 estimates range from approximately –0.5% to 2.0%, and 17 of the 18 index–GWL combinations have no European land grid cells

surviving false-discovery-rate correction. Detrended interannual AMOC variability therefore shows little robust contemporaneous association with European climate extremes within the analysed windows. This result does not exclude lagged, low-frequency or nonlinear AMOC influences, which cannot be resolved by the present experimental design.

Keywords: European climate extremes; global warming levels; Atlantic Meridional Overturning Circulation; CMIP6; temperature extremes; precipitation extremes.

1. Introduction

Anthropogenic warming is intensifying climate extremes, with growing consequences for ecosystems, human health, infrastructure and socioeconomic systems^{1,2}. Global mean near-surface temperature in 2024 was 1.55 ± 0.13 °C above the 1850–1900 average, although this single-year exceedance does not indicate that the long-term 1.5 °C threshold has been crossed². Europe is warming approximately twice as rapidly as the global average and is experiencing increasingly severe heatwaves, heavy rainfall and compound hazards^{1,3–5}. Quantifying how these extremes evolve with additional warming is therefore critical for regional risk assessment and adaptation.

Climate extreme indices provide standardized measures of changes in the intensity and persistence of temperature and precipitation extremes^{6–10}. Evaluating these indices at prescribed global warming levels (GWLs), rather than fixed future periods, facilitates comparison among models with different warming rates and directly relates regional impacts to the Paris Agreement thresholds^{5,11,12}. Climate models consistently project stronger warm extremes and heavy precipitation over Europe, but their rates of amplification and regional maxima vary because of atmospheric circulation, land–atmosphere coupling and model sensitivity^{4,11,13–19}. Using bias-adjusted, high-resolution EURO-CORDEX simulations, Dosio and Fischer demonstrated robust changes in European temperature and precipitation indices between warming levels of 1.5 °C, 2 °C and 3 °C, including stronger changes in cold-side temperature indices than in maximum-temperature indices²⁰. Building on this established evidence, the present study uses an 18-model ensemble of bias-

corrected, statistically downscaled CMIP6 projections to compare six annual extreme indices within a consistent framework. We focus on their spatially differentiated regional hazard patterns, inter-model distributions and amplification beyond GWL1.5, while separately examining projected AMOC behaviour and its contemporaneous statistical associations with the extreme indices in a matched model subset.

North Atlantic ocean circulation may influence European climate variability. The Atlantic Meridional Overturning Circulation (AMOC) transports heat northwards and affects North Atlantic temperatures, atmospheric circulation and European hydroclimate²¹⁻²³. For example, Jackson et al.²⁴ found localised increases in winter precipitation associated with more winter storms and a strengthened storm track, and Meccia et al.²⁵ found fewer winter cold spells over Europe with a reduced AMOC. CMIP6 models consistently project AMOC weakening during the twenty-first century, although the magnitude and timing of this weakening vary substantially among models and forcing scenarios^{21,26,27}. AMOC variability has also been associated with European temperature, precipitation and weather extremes^{23,24,28}. However, the presence and strength of these relationships depend on the timescale, region and climate variable considered.

The concurrent intensification of European climate extremes and weakening of the AMOC does not, by itself, demonstrate that changes in the AMOC modify those extremes, because both evolve as the climate warms. A narrower unresolved question is whether annual AMOC anomalies and European climate extremes exhibit robust contemporaneous covariability within comparable warming states. Testing this relationship after removing fitted linear trends can identify residual statistical associations within individual GWL windows, but it cannot formally separate externally forced and internally generated variability or establish causal AMOC influence.

Here, we combine bias-corrected, statistically downscaled CMIP6 projections with corresponding AMOC simulations to examine six temperature and precipitation extreme

indices across Europe at global warming levels of 1.5 °C, 2.0 °C and 3.0 °C under SSP5–8.5. Using 18 models, we quantify continental changes, regional patterns and amplification beyond GWL1.5. We then characterize projected AMOC weakening in 11 models with matched AMOC and climate-extreme realizations and test contemporaneous associations between their detrended annual anomalies within model-specific 20-year GWL windows. We find spatially differentiated intensification of European climate extremes and progressively more negative AMOC anomalies across GWLs, but little robust contemporaneous association between their detrended interannual variability. These findings characterize two concurrent responses under SSP5–8.5 without attributing the projected changes in European extremes to AMOC weakening.

2. Results

2.1 Continental-scale intensification of European climate extremes across 1.5–3.0 °C global warming levels

2.1.1 Heavy precipitation intensifies unevenly across Europe

Heavy-precipitation extremes increase across most of Europe at all three warming levels (Fig. 1a,b). Both annual maximum 1-day precipitation (Rx1day) and annual maximum consecutive 5-day precipitation (Rx5day) show their strongest and most spatially coherent increases over northern Europe, with additional increases across parts of western-central and eastern Europe. Mediterranean responses are more heterogeneous, with localized decreases persisting alongside areas of increasing extreme precipitation.

The Europe-wide mean Rx1day change increases from approximately 3.3 mm at GWL1.5 to 3.7 mm at GWL2.0 and 4.4 mm at GWL3.0 (Fig. 1g). Corresponding Rx5day changes are approximately 5.6, 6.2 and 7.3 mm, respectively (Fig. 1h). Between GWL1.5 and GWL3.0, this represents additional increases of approximately 1.1 mm for Rx1day and 1.7 mm for Rx5day. Despite their different absolute magnitudes, both indices are approximately 30%

larger at GWL3.0 than at GWL1.5, indicating comparable relative amplification but pronounced regional differences in the precipitation response.

2.1.2 Coldest daytime temperatures warm faster than hottest daytime temperatures

Both the annual maximum of daily maximum (TXx) and annual minimum of daily maximum (TXn) increase across Europe at each warming level (Fig. 1c,d). TXx warming is strongest over western-central and southern Europe, reaching approximately 6 °C locally at GWL3.0. TXn shows a distinct spatial pattern, with changes approaching 8 °C over northern and eastern Europe. These contrasting patterns persist as their magnitude and spatial coherence increase from GWL1.5 to GWL3.0.

The Europe-wide mean TXx change increases from approximately 2.6 °C at GWL1.5 to 3.2 °C at GWL2.0 and 4.6 °C at GWL3.0 (Fig. 1i). The corresponding TXn changes are approximately 3.4, 4.5 and 6.0 °C, respectively (Fig. 1j). TXn warming therefore exceeds TXx warming by 0.8–1.4 °C across the three GWLs, demonstrating that the coldest daytime temperatures warm more strongly than the hottest daytime temperatures. This asymmetry distinguishes southern and western-central European amplification of hot extremes from the stronger northern and eastern European warming of cold-side daytime temperatures.

2.1.3 Warm-spell expansion dominates changes in temperature persistence

Warm- and cold-spell duration respond asymmetrically to increasing global warming (Fig. 1e,f). Warm-spell duration index (WSDI) increases throughout Europe, with the strongest changes over southern Europe and the Mediterranean and progressively larger increases extending into western-central and eastern Europe. At GWL3.0, local increases exceed approximately 180 days over parts of southeastern Europe and the eastern Mediterranean. By contrast, cold spell duration index (CSDI) decreases across almost the entire domain, with the largest reductions of approximately 8 days over northern and eastern Europe.

142 The Europe-wide mean WSDI change increases from approximately 27 days at GWL1.5 to
143 40 days at GWL2.0 and 68 days at GWL3.0 (Fig. 1k). Corresponding cold spell duration index
144 (CSDI) changes are approximately -5.3, -6.0 and -6.6 days, respectively (Fig. 1l). Thus,
145 warming from GWL1.5 to GWL3.0 adds approximately 41 warm-spell days but only 1.3 days
146 to the magnitude of cold-spell contraction. Warm-spell duration therefore continues to
147 expand strongly beyond GWL1.5, whereas cold-spell reductions are already widespread at
148 GWL1.5 and intensify comparatively modestly thereafter.

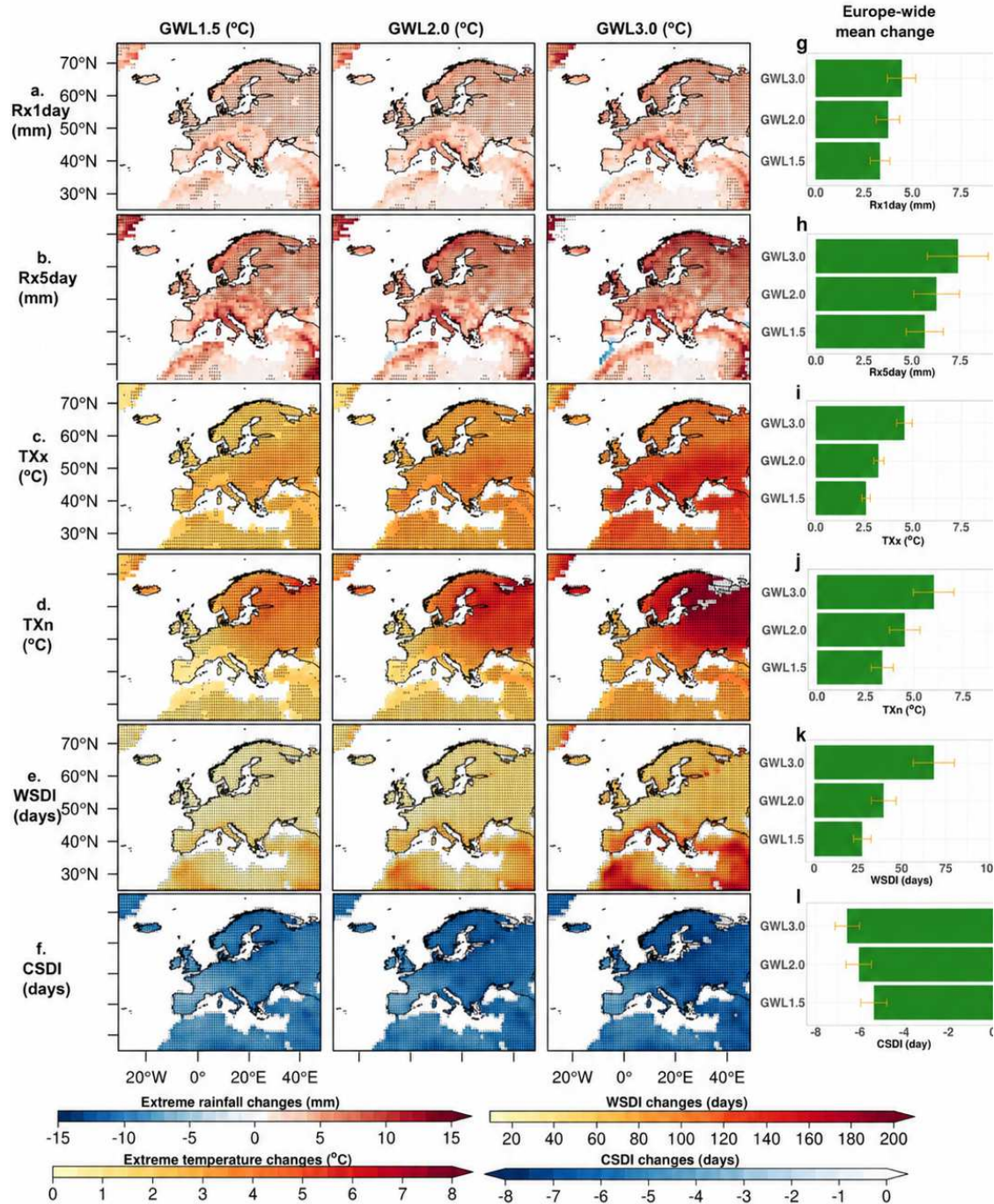


Figure 1 | Projected changes in European climate extremes across global warming levels. Multi-model ensemble-mean changes at GWL1.5, GWL2.0 and GWL3.0 relative to the 1995–2014 baseline. Maps show changes in **a**, annual maximum 1-day precipitation (Rx1day); **b**, annual maximum consecutive 5-day precipitation (Rx5day); **c**, annual maximum of daily maximum temperature (TXx); **d**, annual minimum of daily minimum temperature (TXn); **e**, Warm Spell Duration Index (WSDI); and **f**, Cold Spell Duration Index (CSDI). Europe-wide mean changes are shown for **g**, Rx1day; **h**, Rx5day; **i**, TXx; **j**, TXn; **k**, WSDI; and **l**, CSDI. Orange

error bars denote the uncertainty around the mean. Stippling indicates grid cells with changes statistically significant at the 95% confidence level.

2.2 Distinct regional patterns emerge with increasing warming

Europe-wide means mask pronounced differences in the regional expression of climate extremes (Fig. 2). Northern Europe shows the strongest amplification of heavy precipitation, the Mediterranean the largest expansion of warm-spell duration, and northern and eastern Europe the strongest TXn warming. These distinct regional maxima generally become more pronounced at higher GWLs, showing that the dominant climate extremes associated with a given level of global warming differs across Europe.

Figure 2 compares standardized regional responses with their corresponding inter-model distributions. The ensemble standardized anomalies characterize model agreement on the magnitude of the projected changes, while the violin distributions quantify uncertainty across models. Together, they show whether the regional changes remain distinguishable despite inter-model spread.

2.2.1 Regional responses diverge across Europe

Northern Europe shows the strongest regional precipitation response (Fig. 2a–b). Between GWL1.5 and GWL3.0, its ensemble-mean Rx1day change increases from approximately 3.7 to 5.0 mm and Rx5day from 6.2 to 8.8 mm. Mediterranean responses remain substantially weaker: Rx1day changes only slightly from approximately 3.0 to 3.1 mm, while Rx5day decreases from 4.5 to 4.1 mm. At GWL3.0, northern European changes exceed Mediterranean changes by approximately 1.9 mm for Rx1day and 4.7 mm for Rx5day.

Temperature-intensity indices exhibit a different regional pattern (Fig. 2e–h). TXx warming is strongest over western-central Europe, increasing from approximately 2.8 °C at GWL1.5 to 5.0 °C at GWL3.0. TXn warming is greatest over eastern Europe, increasing from approximately 3.7 to 6.6 °C, followed by northern Europe, where it increases from 3.6 to 6.1

°C. Mediterranean TXn changes are substantially smaller, increasing from approximately 1.8 to 3.2 °C.

Regional differentiation is strongest for warm-spell duration (Fig. 2i–l). Mediterranean WSDI increases from approximately 36 days at GWL1.5 to 98 days at GWL3.0, exceeding the GWL3.0 Europe-wide mean by approximately 29 days and the western-central European response by 38 days. In contrast, regional CSDI reductions at GWL3.0 span a comparatively narrow range, from approximately 5.2 days in the Mediterranean to 6.8 days in northern Europe. Thus, northern Europe is distinguished by heavy-precipitation intensification, the Mediterranean by warm-spell expansion, and northern and eastern Europe by strong warming of the coldest daytime temperatures.

2.2.2 Regional signals persist amid inter-model spread

Inter-model distributions shift towards heavier precipitation, warmer temperature extremes, longer warm spells and fewer cold-spell days as warming increases (Fig. 2b,d,f,h,j,l). The distributions also generally broaden towards GWL3.0, indicating increasing uncertainty in the magnitude of regional change. Absolute spread is largest for WSDI, particularly over the Mediterranean, while substantial overlap among regional distributions is evident for the precipitation indices and CSDI.

Despite this spread, the ensemble centres retain the principal regional fingerprints: northern Europe shows the largest mean precipitation response, the Mediterranean the largest WSDI increase and eastern Europe the strongest TXn warming. However, overlap among the distributions indicates that these regional rankings are not reproduced uniformly by every model. The ensemble therefore supports distinct regional hazard patterns, while the precise magnitude and ordering of regional responses remain model dependent.

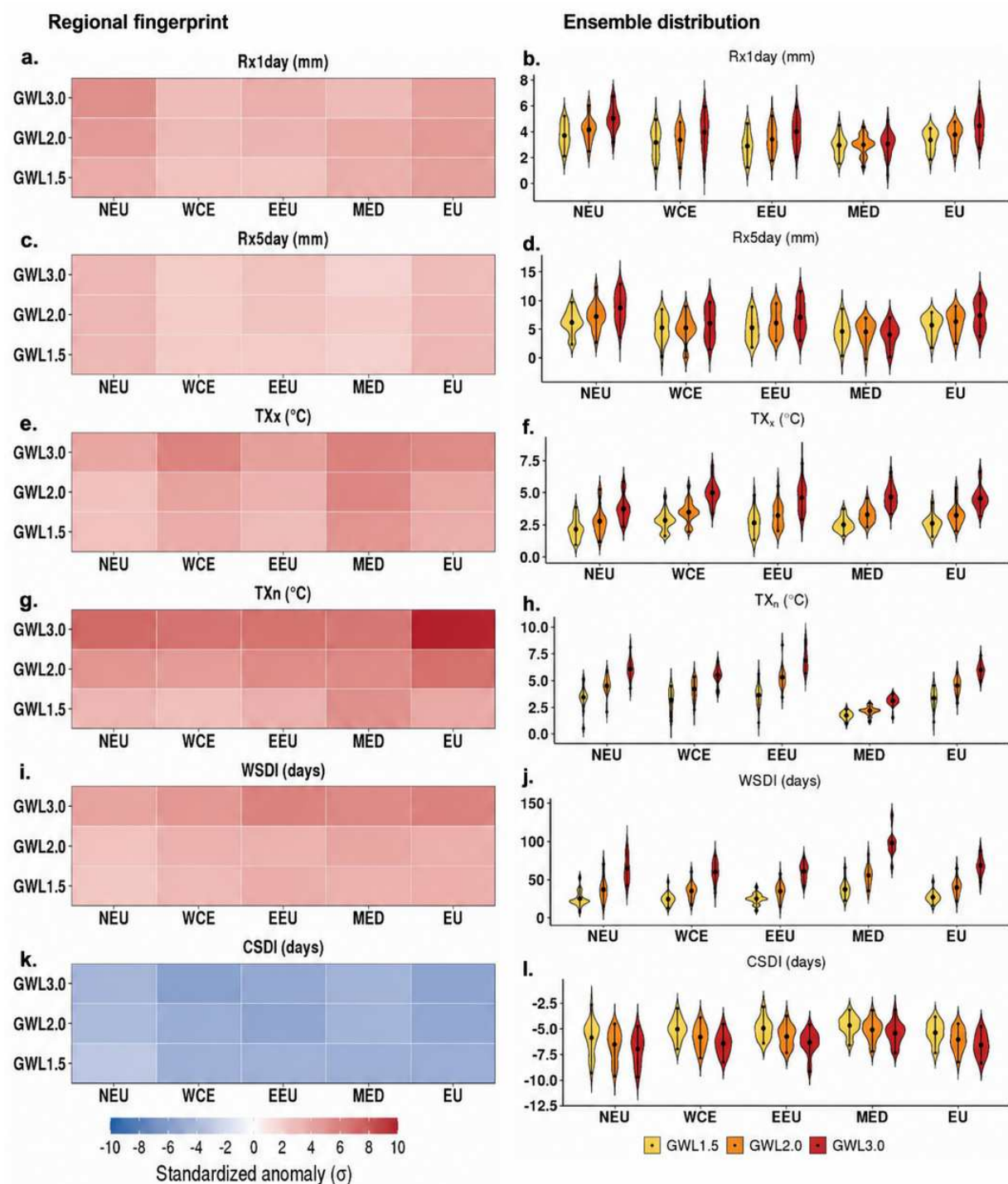


Figure 2 | Regional fingerprints and inter-model distributions of European climate extremes across global warming levels. Multi-model ensemble-mean standardized regional anomalies (left) and corresponding inter-model distributions in their original physical units (right) for **a,b**, Rx1day; **c,d**, Rx5day; **e,f**, TX_x; **g,h**, TX_n; **i,j**, WSDI; and **k,l**, CSDI at GWL1.5, GWL2.0 and GWL3.0 relative to the 1995–2014 baseline. Results are shown for northern Europe (NEU), western and central Europe (WCE), eastern Europe (EEU), the Mediterranean (MED)

and Europe overall (EU). In the violin plots, black points denote multi-model ensemble means, and violin widths represent model density. Yellow, orange and red denote GWL1.5, GWL2.0 and GWL3.0, respectively.

2.2.3 Warm-spell duration shows the strongest amplification beyond 1.5 °C

To compare indices expressed in different physical units, we calculated response-amplification factors at GWL2.0 and GWL3.0 relative to GWL1.5 (Fig. 3):

WSDI exhibits the strongest amplification across all European regions, with response factors increasing from approximately 1.4–1.5 at GWL2.0 and 2.4–2.6 at GWL3.0 relative to GWL1.5. Temperature-intensity responses show more moderate amplification, with TXx and TXn increasing by factors of approximately 1.2–1.4 at GWL2.0 and 1.7–1.9 at GWL3.0. At GWL3.0, TXx amplification is greatest over the Mediterranean, whereas TXn amplification is largest over Eastern Europe.

In contrast, precipitation extremes exhibit weaker amplification, with Rx1day and Rx5day generally remaining between approximately 1.0 and 1.4 at GWL3.0. The largest additional amplification occurs over Northern Europe, whereas Mediterranean Rx5day exhibits an amplification below 1, indicating a weaker regional mean response than at GWL1.5. CSDI exhibits modest amplification approximately 1.2 and 1.3 across all regions. At the Europe-wide scale, amplification reaches approximately 2.5 for WSDI, 1.8 for TXx and TXn, 1.3 for Rx1day and Rx5day, and 1.2 for CSDI, establishing a clear hierarchy in response to additional warming. Collectively, these results demonstrate that warming beyond 1.5 °C preferentially amplifies the persistence of heat extremes, while changes in temperature intensity are more moderate and precipitation extremes and cold-spell contraction exhibit comparatively limited additional scaling.

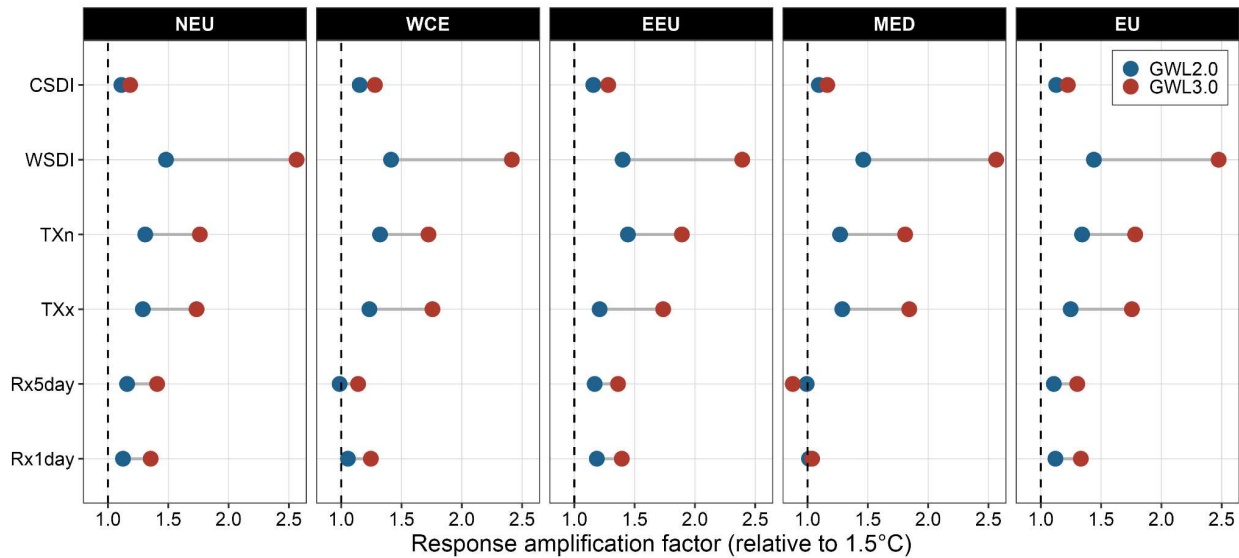


Figure 3 | Relative amplification of European climate-extreme responses beyond 1.5 °C global warming. Regional response-amplification factors at GWL2.0 and GWL3.0 relative to GWL1.5 for Rx1day, Rx5day, TXx, TXn, WSDI and CSDI over northern Europe (NEU), western and central Europe (WCE), eastern Europe (EEU), the Mediterranean (MED) and Europe overall (EU). Amplification factors are calculated as $A_{GWL} = |\Delta X_{GWL}| / |\Delta X_{GWL1.5}|$, where ΔX is the projected change relative to 1995–2014. Values greater than 1 indicate amplification relative to GWL1.5, values below 1 indicate a weaker response and the dashed line marks no change in magnitude. Blue and red points denote GWL2.0 and GWL3.0, respectively; connecting lines show the additional amplification between these warming levels.

2.3 Progressive AMOC weakening under SSP5–8.5 across global warming levels

Within the SSP5–8.5 simulations, AMOC states become progressively weaker at successive global warming levels. Relative to 1995–2014, the matched 11-model ensemble-mean AMOC anomaly is -2.81 Sv at GWL1.5 (95% bootstrap confidence interval, -4.65 to -1.30 Sv), -3.63 Sv at GWL2.0 (-5.64 to -1.99 Sv) and -5.10 Sv at GWL3.0 (-7.27 to -3.36 Sv; Fig. 4). Ten models project a weaker AMOC at GWL1.5 and GWL2.0, and all 11 do so at GWL3.0.

Paired model responses reinforce this progression (Fig. 4b and Table 1). Compared with GWL1.5, the ensemble-mean AMOC weakens by an additional 0.83 Sv at GWL2.0 and 2.29 Sv at GWL3.0. A further weakening of 1.47 Sv occurs between GWL2.0 and GWL3.0. All

11 models exhibit additional weakening in each successive comparison, although the magnitude of the response differs substantially among models.

These results characterize AMOC states along the SSP5–8.5 pathway. Because AMOC adjustment may depend on the rate and pathway of forcing, they should not be interpreted as scenario-independent responses at a given GWL.

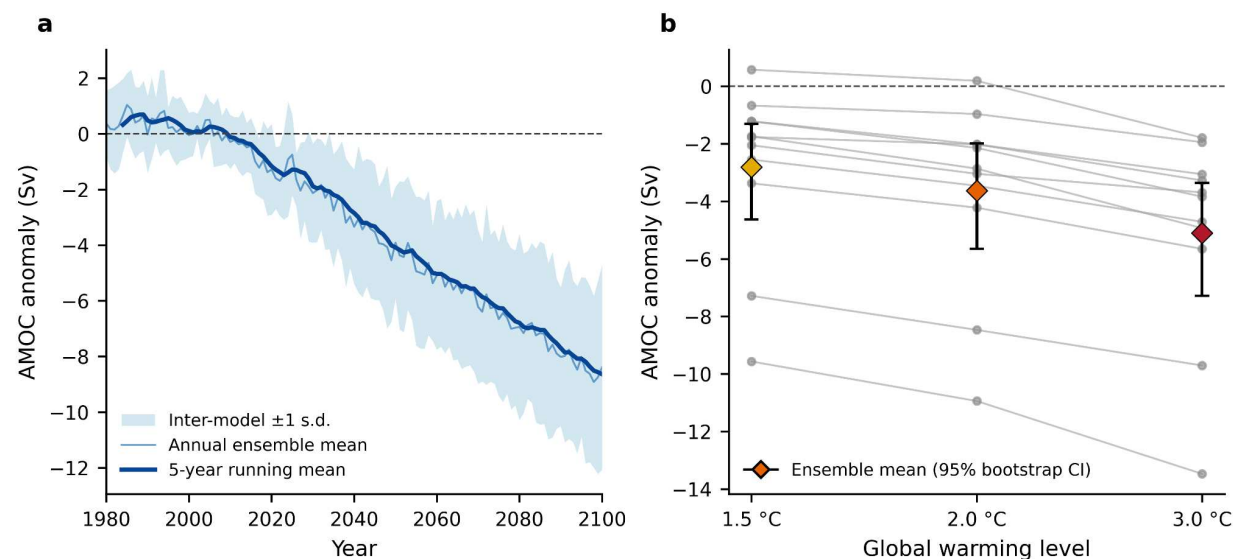


Figure 4 | Progressive AMOC weakening under SSP5–8.5 across global warming levels. a, Annual ensemble-mean AMOC anomaly at 26.5° N from 1980 to 2100, relative to 1995–2014. The light-blue and dark-blue curves show the annual and five-year trailing ensemble means, respectively; shading denotes the inter-model ± 1 s.d. b, AMOC anomalies averaged over the model-specific 20-year windows centred asymmetrically on each crossing year corresponding to GWL1.5, GWL2.0 and GWL3.0. Grey points and connecting lines show the same models as in a; coloured diamonds and error bars show the multi-model means and 95% bootstrap confidence intervals. Because the windows are model-specific, the values in b are not derived from a single common calendar interval in a.

Differences among the warming levels are statistically robust (Friedman $\chi^2(2) = 22.0$, $P = 1.67 \times 10^{-5}$; Kendall's $W = 1.0$), with all paired comparisons remaining significant after Holm correction ($P = 0.0029$; Table 1). Inter-model standard deviation increases from 3.00 Sv at GWL1.5 to 3.52 Sv at GWL3.0, indicating growing uncertainty in the magnitude of AMOC weakening. Nevertheless, the direction of change across successive GWLs is

consistent across all models. Progressively higher GWLs are therefore associated with systematically weaker AMOC states, providing the basis for assessing their statistical relationship with European climate extremes.

Table 1 | Projected AMOC anomalies and paired changes across global warming levels

GWL or comparison	Mean anomaly or difference (Sv)	95% bootstrap CI (Sv)	Model agreement
GWL1.5	-2.81	-4.65 to -1.30	10/11 weaker
GWL2.0	-3.63	-5.64 to -1.99	10/11 weaker
GWL3.0	-5.10	-7.27 to -3.36	11/11 weaker
GWL2.0 – GWL1.5	-0.83	-1.03 to -0.61	11/11 further weaken
GWL3.0 – GWL1.5	-2.29	-2.74 to -1.89	11/11 further weaken
GWL3.0 – GWL2.0	-1.47	-1.79 to -1.17	11/11 further weaken

Note. AMOC anomalies are calculated relative to each model’s 1995–2014 mean and averaged over model-specific 20-year GWL windows. Confidence intervals were obtained from 50,000 bootstrap resamples of the 11 models. Negative paired differences indicate additional weakening. Exact two-sided Wilcoxon signed-rank tests were significant for all paired comparisons after Holm correction ($P = 0.0029$). Overall differences were assessed using a Friedman test ($\chi^2(2) = 22.0$, $P = 1.67 \times 10^{-5}$; Kendall’s $W = 1.0$). AMOC, Atlantic Meridional Overturning Circulation; CI, confidence interval; GWL, global warming level; Sv, sverdrup.

2.4 Detrended AMOC variability shows limited contemporaneous associations with European climate extremes

We next tested whether annual AMOC anomalies and European climate extremes exhibit contemporaneous interannual covariability within comparable warming states. For each of the 11 models with matched AMOC and climate-extreme realizations, annual AMOC anomalies and six climate-extreme indices were linearly detrended within their model-specific 20-year GWL windows. This procedure removes the fitted linear trend from each series but does not formally separate externally forced and internally generated variability. The resulting regressions therefore quantify contemporaneous statistical associations and should not be interpreted as evidence that AMOC variability causes or modifies European climate extremes.

The fitted regressions accounted for little of the interannual variance in the climate-extreme indices (Fig. 5a,b). Mean area-weighted in-sample R^2 ranged from 4.9% to 7.3%, but declined to between -0.5% and 2.0% after adjustment for finite sample size and lag-1 autocorrelation (Table 2). Spatial median R^2 values were likewise generally low across indices and warming levels (Supplementary Fig. 9). The largest autocorrelation-adjusted values occurred for TXn at GWL1.5 (2.0%), WSDI at GWL3.0 (1.7%) and CSDI at GWL2.0 (0.9%); all remaining values were 0.5% or lower. Effective sample sizes remained close to the nominal 20 years, indicating that the reduction arose primarily from finite-sample adjustment rather than strong temporal persistence (Supplementary Table 4).

The direction and spatial expression of the associations were also weakly constrained. Area-weighted Fisher-mean correlations ranged from -0.029 to 0.046, and mean grid-cell agreement among models on the sign of the regression coefficient ranged from 61.0% to 70.8% (Fig. 5c, Supplementary Fig. 10 and Supplementary Table 4). No grid cells survived false-discovery-rate correction for 17 of the 18 index-GWL combinations (Fig. 5d). The only exception was Rx5day at GWL2.0, for which significance covered 0.15% of the analysed European land domain and was not reproduced at GWL1.5 or GWL3.0. The analysis therefore provides little evidence of a robust contemporaneous relationship between detrended interannual AMOC variability and European climate extremes within the analysed GWL windows. It does not test or exclude AMOC influences operating through lagged, low-frequency, nonlinear or circulation-mediated processes.

An exploratory inter-model diagnostic identified correlations between GWL-mean AMOC anomalies and some projected climate-extreme changes (Supplementary Fig. 11 and Supplementary Table 5). These descriptive associations may reflect differences among models in climate sensitivity, ocean heat uptake, circulation response or shared model genealogy. Given the sample of 11 structurally different models and the absence of controls for these potential confounders, the correlations cannot identify an AMOC-mediated response. They are reported only as a supplementary diagnostic and are not used to infer physical coupling, interpret the within-GWL results or support the principal conclusions.

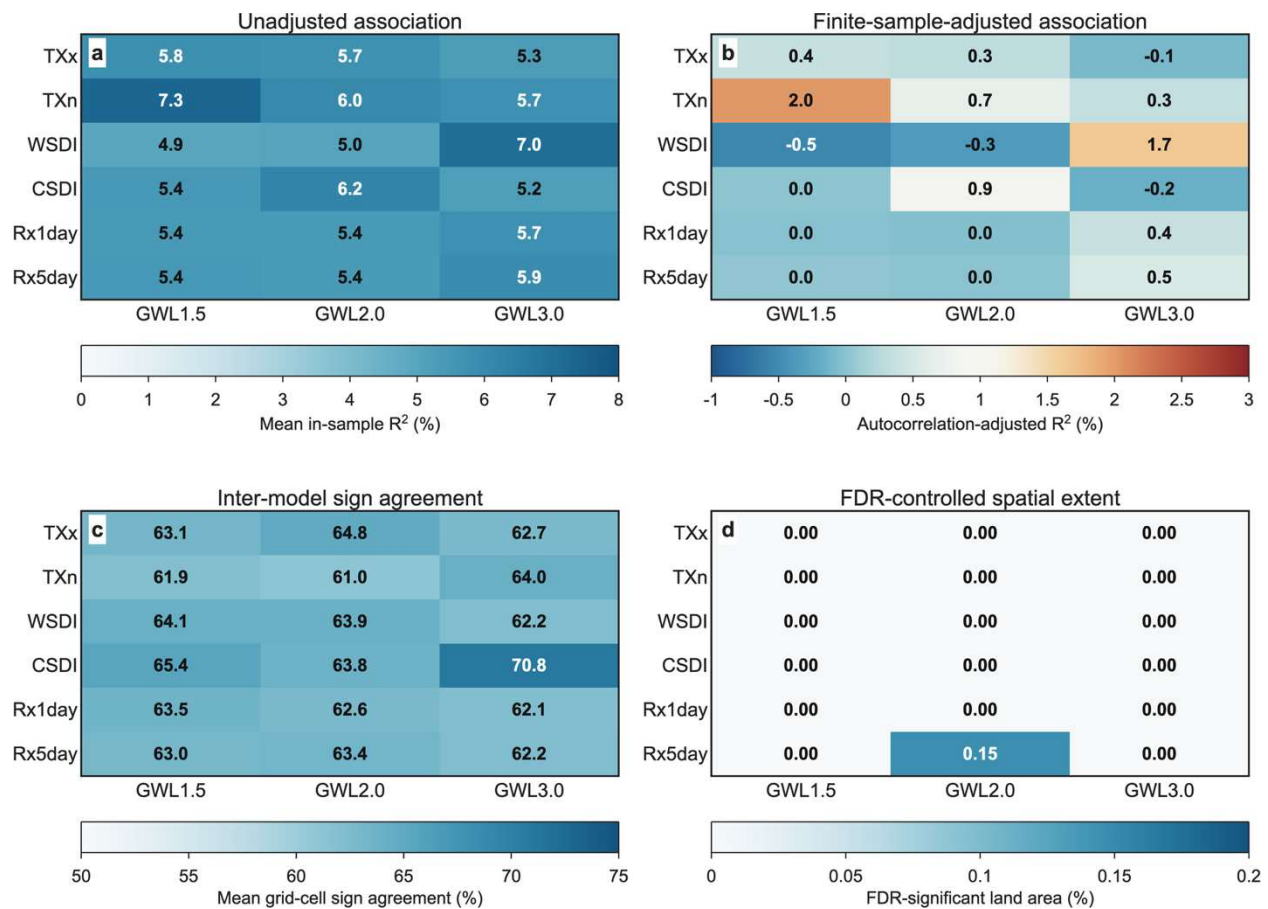


Figure 5 | Contemporaneous associations between detrended AMOC variability and European climate extremes across global warming levels. Area-weighted results across 11 matched CMIP6 models at GWL1.5, GWL2.0 and GWL3.0. **a**, Mean in-sample (R^2). **b**, Mean (R^2) adjusted for finite sample size and lag-1 autocorrelation. **c**, Mean grid-cell agreement among models on the majority sign of the regression coefficient. **d**, Percentage of the analysed land domain where the across-model association survives Benjamini–Hochberg false-discovery-rate correction at ($q=0.05$). Annual AMOC anomalies and climate-extreme indices were linearly detrended within each model-specific 20-year GWL window. The results represent contemporaneous statistical associations and do not establish causality.

Table 2 | Contemporaneous associations between detrended AMOC variability and European climate extremes

Extreme index	GWL1.5		GWL2.0		GWL3.0	
	Adjusted R ² (%)	FDR area (%)	Adjusted R ² (%)	FDR area (%)	Adjusted R ² (%)	FDR area (%)
TXx	0.4	0	0.3	0	−0.1	0
TXn	2.0	0	0.7	0	0.3	0
WSDI	−0.5	0	−0.3	0	1.7	0
CSDI	0.0	0	0.9	0	−0.2	0
Rx1day	0.0	0	0.0	0	0.4	0
Rx5day	0.0	0	0.0	0.15	0.5	0

Note. Adjusted R² accounts for finite sample size and lag-1 autocorrelation. Negative values indicate no improvement over an intercept-only model after adjustment. FDR area is the percentage of the analysed land domain where the across-model association remains significant after Benjamini–Hochberg correction at q = 0.05. Results are based on detrended annual data from 11 matched models within model-specific 20-year GWL windows.

3. Discussion

Our results show spatially differentiated intensification of European climate extremes between 1.5 °C and 3.0 °C global warming under SSP5–8.5. Heavy precipitation intensifies most strongly over northern Europe, hot extremes and warm-spell duration are amplified over western-central and southern Europe, and the largest TXn increases occur over northern and eastern Europe. In the matched 11-model subset, AMOC states at 26.5° N also become progressively weaker at successive GWLs. However, detrended annual AMOC variability shows little robust contemporaneous association with year-to-year variations in European extremes within the model-specific 20-year windows. These are related but statistically distinct findings: the analysis characterizes projected changes across GWLs and separately tests contemporaneous within-window associations, but it does not determine whether long-term AMOC weakening modifies the magnitude or spatial pattern of the projected extremes.

The contrasting responses of temperature intensity and persistence reflect the interaction of large-scale thermodynamic forcing with regional surface processes. Increasing

greenhouse-gas concentrations shift the temperature distribution towards warmer conditions, intensifying TXx and reducing the severity and persistence of cold conditions^{1,8, 10}. TXn shows larger increases than TXx at the European scale, consistent with the sensitivity of cold extremes to snow-cover loss, surface-albedo feedbacks, boundary-layer processes and enhanced high-latitude warming^{1,4}. The different spatial fingerprints are important: TXx amplification is concentrated over western–central Europe and the Mediterranean, whereas TXn warming is strongest over northern and eastern Europe. Hot- and cold-extreme responses should therefore not be interpreted as spatially uniform consequences of the same mean warming.

The pronounced increase in WSDI further shows that warming affects the persistence as well as the intensity of heat. Warm-spell duration exhibits the strongest relative amplification beyond GWL1.5, particularly over the Mediterranean, where soil-moisture depletion, increased sensible heat fluxes and persistent atmospheric circulation anomalies favour prolonged heat events^{15–18}. By contrast, CSDI declines across Europe but changes less strongly in relative magnitude. This asymmetry indicates a transition towards a climate in which persistent heat expands much faster than cold spells contract. Such changes are consequential because the impacts of prolonged heat can accumulate across health, agriculture, ecosystems, water resources and energy systems even where changes in annual temperature maxima are comparatively moderate.

The precipitation response is also spatially heterogeneous. Rx1day and Rx5day increase with warming across most of Europe, with larger absolute changes in Rx5day and the strongest regional amplification over northern Europe. This is in line with Bellomo et al.²⁸, who find that a decline in the AMOC leads to an increase in the frequency of the positive phase of the North Atlantic Oscillation (NAO+), which is associated with an increase in the occurrence of wet days over northern Europe. The broad response is also consistent with the increasing moisture-holding capacity of a warmer atmosphere and greater moisture availability for heavy precipitation¹⁴. Regional departures from this thermodynamic expectation reflect changes in circulation, moisture transport, storm characteristics and

land–atmosphere interactions. The weaker and less spatially uniform Mediterranean response illustrates that atmospheric moistening does not translate directly into geographically uniform precipitation change. The distinction between single-day and multi-day extremes is also societally relevant because sustained rainfall accumulations can generate different flood, landslide and infrastructure risks from short-duration downpours^{19, 30}.

Together, the regional fingerprints demonstrate that a common increment in global mean temperature produces different dominant hazards across Europe. Northern Europe is distinguished by heavy-precipitation amplification, the Mediterranean by rapidly expanding warm-spell duration, and northern and eastern Europe by strong warming of cold extremes. These findings complement earlier GWL-based assessments, including Dosio and Fischer²⁰, by evaluating six consistently processed annual indices in a bias-corrected and statistically downscaled CMIP6 ensemble and comparing their regional patterns, inter-model distributions and amplification beyond GWL1.5. The results caution against representing European climate risk using a single continental-scale metric because the dominant response depends on region, index and whether the hazard is characterized by intensity or persistence.

The AMOC analysis identifies a coherent progression within the SSP5–8.5 simulations. Relative to 1995–2014, the matched 11-model ensemble projects mean anomalies of –2.81 Sv at GWL1.5, –3.63 Sv at GWL2.0 and –5.10 Sv at GWL3.0. All 11 models exhibit additional weakening between successive warming levels, and differences among the GWL states are supported by the Friedman and paired-model tests. This behaviour is consistent with CMIP6 projections in which ocean warming and freshwater-related buoyancy changes reduce North Atlantic deep-water formation and weaken the overturning circulation^{1, 21–23}. The GWL framework shows that progressively warmer states reached under SSP5–8.5 coincide with progressively weaker AMOC states, although substantial uncertainty remains in the magnitude of the model responses.

These AMOC results are conditional on SSP5–8.5. The circulation may respond differently to pathways that reach the same GWL at different rates or through different combinations of greenhouse-gas, aerosol and other forcings. Consequently, the estimated AMOC anomalies should not be interpreted as universal or scenario-independent responses to a given warming level.

The concurrent intensification of European extremes and weakening of the AMOC does not, however, automatically confirm that the latter causes or modifies the former, although some of the results found here (CSDI and precipitation) are consistent with other AMOC focused studies. After removal of fitted linear trends within the model-specific 20-year windows, autocorrelation-adjusted (R^2) estimates range from –0.5% to 2.0%, and the residual correlations remain close to zero. No grid cells survive false-discovery-rate correction in 17 of the 18 index–GWL combinations. The only exception, Rx5day at GWL2.0, covers 0.15% of the analysed European land domain and is not reproduced at the other warming levels. The results therefore provide little evidence of a robust contemporaneous relationship between detrended interannual AMOC variability and European climate extremes within the analysed windows.

An exploratory inter-model diagnostic identifies stronger correlations for some GWL-mean quantities, particularly TXx and WSDI. However, differences among models in climate sensitivity, ocean heat uptake, sea-ice response, circulation adjustment and shared model genealogy can affect both AMOC and European climate projections. These descriptive correlations therefore cannot identify an AMOC-mediated response and do not underpin our interpretation of the within-GWL results. This conclusion does not contradict the established role of the AMOC in regulating North Atlantic and European climate. Sustained or abrupt AMOC changes can influence ocean heat transport, sea-surface temperature gradients, storm tracks and European hydroclimate^{26, 29, 30}. Our analysis addresses only contemporaneous annual covariability within relatively short GWL windows, revealing a coherent but spatially differentiated intensification of European climate extremes between

1.5 °C and 3.0 °C of global warming, accompanied by progressive weakening of the AMOC, and cannot resolve lagged, low-frequency, nonlinear or circulation-mediated influences.

Broader implications

These results emphasize the benefits of avoiding each additional increment of global warming for European climate extremes. Substantial changes are already present at GWL1.5, but their magnitudes and regional differentiation increase further at GWL2.0 and GWL3.0. Within the SSP5–8.5 simulations, lower GWLs also coincide with less-negative AMOC anomalies, although the magnitude of this difference may depend on the forcing pathway. Adaptation will therefore remain necessary even under strong mitigation. The regional fingerprints indicate that priorities should vary geographically, with greater emphasis on persistent heat across southern Europe, changing cold extremes over northern and eastern Europe, and increasing heavy-precipitation and flood hazards across northern Europe.

Strengths, limitations and conclusions

This study integrates six internationally recognized climate-extreme indices and AMOC projections within a common GWL framework. It combines continental-scale changes, regional model distributions, response-amplification factors, paired-model AMOC inference and matched model–realization association analyses. By evaluating projected changes across GWLs separately from detrended within-window associations, the design avoids treating coincident long-term behaviour as evidence of direct AMOC–extreme coupling. It does not, however, formally partition externally forced and internally generated variability.

Several limitations remain. The climate-extreme projections are derived from CMIP6 simulations under SSP5–8.5, and the matched AMOC analysis is limited to 11 models. Twenty-year GWL windows provide policy-relevant climate states but restrict the sample available for detecting low-frequency or lagged relationships. The AMOC is represented by a single overturning metric at approximately 26.5° N, whereas its influence may depend on

latitude, depth, timescale and circulation component. Multi-model means can also suppress internal variability and obscure model-specific mechanisms, although the regional distributions and matched-model analyses partly address this limitation. Finally, the cross-model GWL-mean associations remain susceptible to common-model controls and cannot provide causal attribution.

Future work using single-model initial-condition large ensembles, longer transient records, lagged analyses and targeted circulation-perturbation experiments could more effectively distinguish externally forced changes, AMOC-mediated responses and internal variability. Within the present framework, additional global warming intensifies European climate extremes and strengthens their regional differentiation, while progressively warmer states under SSP5–8.5 coincide with progressively weaker AMOC states. After removal of fitted within-window linear trends, however, detrended interannual AMOC variability shows little robust contemporaneous association with year-to-year variations in European extremes. These findings characterize the concurrent behaviour of the atmosphere and ocean across GWLs without attributing the projected extreme-event changes to AMOC weakening. They do not exclude AMOC influences operating through lagged, low-frequency, nonlinear or circulation-mediated processes that cannot be resolved by the present experimental design.

4. Data and Methods

4.1 Study area

Europe encompasses climatic regimes ranging from Arctic and maritime environments to continental and Mediterranean climates. Strong latitudinal, longitudinal and topographic gradients generate substantial spatial differences in temperature, precipitation and atmospheric circulation. The analysis covers European land areas within approximately 25° W–45° E and 30°–72° N.

To characterize regional heterogeneity, the domain was divided into four IPCC climate-reference regions: Northern Europe (NEU), Western and Central Europe (WCE), Eastern Europe (EEU) and the Mediterranean (MED) as outlined in Supplementary Fig. 1³¹. Regional analyses were conducted alongside Europe-wide assessments to determine how the magnitude, spatial distribution and inter-model spread of temperature and precipitation extremes vary across GWLs.

4.2 Climate Datasets

We combined statistically downscaled CMIP6 climate projections, native-resolution CMIP6 ocean-circulation simulations and gridded observations to assess European climate extremes and AMOC behaviour across GWLs.

Daily climate projections were obtained from the Global Downscaled Projections for Climate Impacts Research (GDPCIR) dataset³³. GDPCIR provides bias-corrected and statistically downscaled CMIP6 simulations of daily minimum temperature (tasmin), daily maximum temperature (tasmax), and total precipitation (pr) at 0.25° spatial resolution for the historical (1950-2014) and the future emission pathways (2015-2100).

SSP2–4.5 and SSP5–8.5 simulations were initially obtained for all available models (Supplementary Table 1). The GWL analysis was subsequently restricted to SSP5–8.5 because comparatively few SSP2–4.5 simulations reached GWL3.0 by 2100, which would have produced inconsistent model samples across warming levels. Eighteen models providing all three daily variables were retained for the climate-extreme analysis. These data were used to calculate six Expert Team on Climate Change Detection and Indices (ETCCDI) indices: annual maximum of daily maximum temperature (TXx), Annual minimum of daily maximum temperature (TXn), Annual count of days in periods of ≥ 6 consecutive days when daily maximum temperature exceeds the calendar-day 90th percentile (WSDI), Annual count of days in periods of ≥ 6 consecutive days when daily minimum temperature falls below the calendar-day 10th percentile (CSDI), Annual maximum 1-day precipitation

amount (Rx1day) and Annual maximum consecutive 5-day precipitation amount (Rx5day) as defined in Supplementary Table 2.

Annual global mean near-surface air temperature was obtained from the corresponding CMIP6 historical and SSP5–8.5 simulations for 1850–2100. Model-specific crossings of 1.5 °C, 2.0 °C and 3.0 °C above the 1850–1900 mean were identified using the procedure described in Section 4.5. The resulting crossing years and associated 20-year windows were used to extract the climate-extreme and AMOC data for each GWL (Supplementary Table 3). Multi-model means were calculated only after the model-specific GWL windows had been identified.

The annual AMOC index was defined as the maximum Atlantic overturning streamfunction (msftmz) at approximately 26.5° N, close to the latitude monitored by the RAPID–MOCHA–WBTS array ³² and expressed in sverdrups (Sv). Because continuous msftmz output is unavailable for all models included in the GDPCIR dataset, AMOC analyses were performed using a subset of 11 CMIP6 models with complete historical and future simulations (Supplementary Table 3).

To evaluate model performance in reproducing the European climate extremes, we obtained the ensemble E-OBS v28.0e gridded observational dataset, which provides quality-controlled daily minimum temperature, maximum temperature, and precipitation from 1950 to present at 0.25° spatial resolution across Europe ³³. The common 1950–2014 period was used for comparison with the historical GDPCIR simulations.

Although the GDPCIR dataset is bias-corrected using the ERA5 reanalysis and its performance has been evaluated against ERA5 in the original dataset description ³⁴, we additionally evaluated the selected CMIP6 models against E-OBS, an observation-based dataset widely used for European climate analyses. This provides an independent regional assessment of model performance for the variables and climate-extreme indices considered in this study and enables comparison with previous European climate-impact assessments.

4.3 Climate extreme indices

Six climate-extreme indices recommended by the Expert Team on Climate Change Detection and Indices (ETCCDI) were calculated using the Climpack2 software package, following the standard ETCCDI definitions and procedures⁶⁻¹⁰. These indices were selected because they capture key aspects of changes in temperature and precipitation extremes that are relevant to climate impacts in Europe and are widely used for evaluating observed and projected climate variability.

The selected indices comprise four temperature extremes: annual maximum of daily maximum temperature (TXx), annual minimum of daily maximum temperature (TXn), Warm Spell Duration Index (WSDI), and Cold Spell Duration Index (CSDI); and two precipitation extremes: annual maximum 1-day precipitation amount (Rx1day) and annual maximum consecutive 5-day precipitation amount (Rx5day). They characterize changes in the intensity and persistence of daytime hot and cold extremes, as well as the magnitude of heavy precipitation events. A summary description of the indices is provided in the supplementary Table 2.

All indices were computed on an annual basis from the daily GDPCIR bias-corrected and statistically downscaled CMIP6 projections (tasmax, tasmin and pr) and the E-OBS observational datasets using identical processing procedures to ensure methodological consistency. For the percentile-based temperature indices (e.g. WSDI and CSDI), climatological thresholds were calculated using a common 1979–2014 base period and subsequently applied to both the historical and future simulations, ensuring a consistent baseline for quantifying changes in warm and cold spell characteristics across all warming levels. The resulting annual time series formed the basis for model evaluation, assessment of projected changes across GWL1.5, GWL2.0, and GWL3.0, and analyses of their relationships with contemporaneous AMOC variability.

4.4. Model Evaluation

The ability of selected CMIP6 models to reproduce historical European climate extremes was evaluated using the E-OBS observational dataset. Model performance was assessed for each climate extreme index using Taylor diagrams ³⁵, which summarise the spatial correlation coefficient, centred root mean square error (RMSE), and spatial standard deviation relative to the observations.

Performance was evaluated for both the individual models and the multi-model ensemble mean over the common historical reference period (1950-2014). The multi-model ensemble was constructed by averaging the annual ETCCDI indices across all 18 models. This evaluation was undertaken to quantify the ability of the selected ensemble to reproduce the observed spatial characteristics of European temperature and precipitation extremes prior to analysing future projections. The performance of the individual models and the spatial comparison of the multi-model ensemble are presented in Supplementary Figures 2- 5.

4.5 Definition of Global Warming Levels

Global warming levels were used to compare climate responses at equivalent levels of global mean warming rather than at fixed calendar periods^{11,12,36_38}. Annual global mean near-surface air temperature (tas) was obtained from the CMIP6 historical simulations for 1850–2014 and SSP5–8.5 simulations for 2015–2100. Although SSP2–4.5 data were initially considered, the final GWL analysis was restricted to SSP5–8.5 because too few models reached GWL3.0 under SSP2–4.5 by 2100 to maintain a consistent model sample across warming levels.

For each model, annual global mean temperature anomalies were calculated relative to the 1850–1900 mean. A centred 30-year running mean was applied to reduce the influence of interannual variability, and the first year in which the smoothed anomaly reached 1.5 °C, 2.0 °C or 3.0 °C was identified as the corresponding GWL crossing year. A 20-year window centred on the model-specific crossing years (Supplementary Table 3) was then defined in equation 1.

$$W_{m,g} = [Y_{m,g} - 10, Y_{m,g} + 9] \dots\dots\dots(1)$$

where $Y_{m,g}$ is the crossing year for model m and warming level g . Each window therefore contained 20 annual values.

Projected changes in the six climate-extreme indices were calculated for each model by averaging over its corresponding 20-year GWL window and subtracting from the mean of the historical reference period 1995–2014. Multi-model ensemble means were subsequently calculated from these model-level changes. This warming-level alignment prevents differences in model climate sensitivity and warming rate from being conflated with differences in regional climate response.

Sections 2.1 and 2.2 used 18 CMIP6 models providing the required daily temperature and precipitation variables. Because corresponding AMOC data were available for 11 of these models, Sections 2.3 and 2.4 used a matched subset of 11 models with AMOC simulations, climate-extreme indices and model-specific GWL windows. The same 20-year window was applied to the AMOC and climate-extreme data for each matched model and GWL. For Section 2.3, each model's AMOC anomalies were calculated relative to its 1995–2014 mean and averaged within its model-specific GWL windows; ensemble statistics and paired differences were subsequently calculated across the 11 model-level values. For Section 2.4, annual AMOC anomalies and climate-extreme indices were extracted from the same windows and separately linearly detrended before the contemporaneous within-GWL association analysis. Detrending removed the fitted linear component from each series but did not formally partition externally forced and internally generated variability. A separate exploratory inter-model diagnostic used the corresponding model-level GWL-mean AMOC anomalies and climate-extreme changes.

4.6 Atlantic Meridional Overturning Circulation

AMOC strength was derived from the CMIP6 meridional overturning mass streamfunction (msftmz) for historical and SSP5–8.5 simulations. Annual AMOC strength was defined as the maximum Atlantic overturning streamfunction at approximately 26.5° N, close to the latitude monitored by the RAPID–MOCHA–WBTS array, and expressed in sverdrups (Sv).^{27,32}

Continuous annual AMOC simulations for 1850–2100 were available for 11 models. The AMOC analyses therefore used 11 models with corresponding climate-extreme simulations and model-specific GWL windows. For each model, annual AMOC anomalies were calculated relative to its 1995–2014 mean as defined in equation 2

$$\Delta M_m(t) = M_m(t) - \overline{M_m}^{1995-2014} \dots\dots\dots(2)$$

where $M_m(t)$ is AMOC strength for model m in year t , and the overbar denotes the corresponding mean value of reference period (1995–2014). Model-level GWL anomalies were calculated by averaging the annual anomalies over the model-specific 20-year windows defined in Section 4.5.

For each GWL, the ensemble mean, median and inter-model standard deviation were calculated across the 11 model-level anomalies. Uncertainty in the ensemble mean was estimated from 50,000 bootstrap resamples of the models. Because the same models contributed to all three GWLs, changes between warming levels were calculated as paired model differences.

A five-year trailing mean of the annual ensemble-mean anomaly was used only to display the long-term evolution of the AMOC in Fig. 4a. All GWL means, uncertainty estimates, paired comparisons and detrended within-GWL AMOC–extreme analyses were calculated from the unsmoothed annual or model-level data.

For the within-GWL analysis in Section 2.4, annual AMOC anomalies were extracted from each model-specific GWL window and linearly detrended. The corresponding annual climate-extreme indices were independently detrended for the same model, realization and years. Grid-cell regressions were then fitted between the two detrended series. The resulting statistics quantify contemporaneous interannual associations after removal of the fitted within-window linear trends. They do not formally separate forced and internal variability, establish a causal direction or determine whether long-term AMOC weakening modifies the projected climate extremes.

4.7 Statistical analysis

Projected changes in the six climate-extreme indices were calculated for each model relative to its 1995–2014 climatology and summarized across the 18-model ensemble at each global warming level (GWL). Continental-wide and regional standardized anomalies were also calculated using equation 3. The standardized anomalies express the ensemble-mean projected change relative to its inter-model spread, thereby providing a dimensionless measure of the strength and consistency of the simulated response.

$$Z = \frac{\bar{A}}{\sigma_A} = \frac{\frac{1}{N} \sum_{i=1}^N A_i}{\sqrt{\frac{1}{N-1} \sum_{i=1}^N \left(A_i - \frac{1}{N} \sum_{i=1}^N A_i \right)^2}}, \dots\dots\dots(3)$$

where Z is the standardized anomalies (or signal-to-noise ratio); A_i the anomaly for model i , $\bar{A}$ is the multi-model mean of the anomalies, σ_A is the multi-model standard deviation of the anomalies.

We further used inter-model distributions based on violin plots to characterize model spread and uncertainty in the magnitude of the projected changes.

To compare responses, the amplification factor at GWL2.0 and GWL3.0 was calculated relative to GWL1.5 as in equation 4

$$A_{GWL} = \frac{|\Delta X_{GWL}|}{|\Delta X_{GWL1.5}|} \dots\dots\dots(4)$$

where ΔX at the specified GWL denotes the projected change in a climate-extreme index relative to 1995–2014. Values greater than one indicate amplification beyond the response at GWL1.5.

AMOC differences among GWLs were evaluated using the 11 models with matched AMOC simulations, climate-extreme indices and model-specific GWL windows. For each model, the annual AMOC anomaly was calculated relative to its 1995–2014 mean and averaged over each 20-year GWL window. Ensemble means and 95% confidence intervals were

estimated by resampling the 11 models with replacement 50,000 times. Inter-model spread was expressed as one standard deviation.

Because the same models contributed to each GWL, differences in AMOC strength were analysed as paired responses. An omnibus Friedman test assessed differences among GWL1.5, GWL2.0 and GWL3.0, with Kendall's W used to quantify the associated effect size. Pairwise differences were evaluated using exact two-sided Wilcoxon signed-rank tests, and the resulting P values were adjusted for the three comparisons using the Holm procedure. Paired effect sizes were additionally expressed as the mean model difference divided by its inter-model standard deviation. The five-year trailing ensemble mean shown in Fig. 4a was used only for visualization and not in the statistical tests.

Contemporaneous within-GWL associations between AMOC variability and European climate extremes were analysed separately for each model, extreme index and GWL. Annual AMOC and climate-extreme series were linearly detrended within each model-specific 20-year window to remove their local linear trends. At every European land grid cell, ordinary least-squares regression was used to estimate the change in the climate-extreme index per 1 Sv increase in AMOC strength, and Pearson's correlation coefficient (r) quantified the contemporaneous linear association. The corresponding in-sample explained variance was calculated as $R^2 = r^2$.

Finite-sample effects were evaluated using the adjusted coefficient of determination defined as in equation 5.

$$R_{adj}^2 = 1 - (1 - R^2) \frac{n-1}{n-2} \dots\dots\dots(5)$$

where n = 20 is the nominal number of annual observations within each GWL window. The effective sample size was estimated from the lag-1 autocorrelations of the detrended AMOC and climate-extreme series following Bretherton et al. ³⁹ defined as in equation 6

$$n_{eff} = n \frac{1-r_{1,AMOC}r_{1,X}}{1+r_{1,AMOC}r_{1,X}} \dots\dots\dots(6)$$

The effective sample size was constrained to $3 \leq n_{eff} \leq n$. An autocorrelation-adjusted diagnostic was then calculated by replacing n with n_{eff} in the adjusted- R^2 equation defined as in equation 7.

$$R_{AC-adj}^2 = 1 - (1 - R^2) \frac{n_{eff}-1}{n_{eff}-2} \dots\dots\dots(7)$$

Negative adjusted values indicate that the fitted relationship performs no better than an intercept-only model after accounting for model complexity and sample size.

Grid-cell correlations were transformed using Fisher's z transformation before aggregation across the 11 models. The ensemble association was obtained by averaging the transformed correlations and back-transforming the result. Statistical evidence against a zero ensemble-mean association was evaluated using two-sided one-sample tests applied to the Fisher-transformed model correlations, treating models as the independent sampling units. Multiple comparisons were controlled separately for each index–GWL map using the Benjamini–Hochberg false-discovery-rate procedure at $q = 0.05$ ⁴⁰. Model agreement was calculated as the proportion of valid models sharing the modal sign of the regression coefficient.

Spatial summaries were restricted to European land within 25° W–45° E and 30–72° N and weighted by the cosine of latitude. An area-weighted statistic was first calculated separately for each model and subsequently averaged across models, thereby assigning equal weight

to each model. Confidence intervals for mean area-weighted R^2 were obtained from 50,000 bootstrap resamples of the 11 models.

As an exploratory inter-model diagnostic, relationships between GWL-mean AMOC anomalies and climate-extreme changes were examined across the 11 model means. For each GWL and grid cell, ordinary least-squares regression and Pearson correlation were calculated across models, and the resulting (P) values were controlled using the same false-discovery-rate procedure. This diagnostic did not control for differences among models in climate sensitivity, ocean heat uptake, circulation response, sea-ice feedbacks or model genealogy. The resulting correlations were therefore treated as descriptive inter-model covariation and were not used to infer an AMOC-mediated response, establish physical coupling or support the principal within-GWL conclusions. All statistical tests were two-sided, with statistical significance assessed at ($P < 0.05$), unless otherwise stated.

Data availability

The observational datasets analysed in this study are publicly available. The E-OBS daily gridded temperature and precipitation dataset (version v28.0e) is available from the European Climate Assessment & Dataset (ECA&D) (<https://www.ecad.eu>). CIL Global Downscaled Projections for Climate Impacts Research (GDPCIR cc-by-4.0) datasets are publicly available on Microsoft AI for Earth's Planetary Computer (<https://planetarycomputer.microsoft.com/dataset/group/cil-gdpcir/>). CMIP6 historical and SSP5-8.5 simulations, including Atlantic Meridional Overturning Circulation (AMOC) overturning streamfunction (*msftmz*) data, were obtained from the Earth System Grid Federation (ESGF) data portal (<https://esgf-node.llnl.gov/search/cmip6/>). Global Warming Level (GWL) periods were derived from the CMIP6 simulations as described in the Methods. The processed datasets used to generate the figures and support the findings of this study will be deposited in a public repository and made available upon publication.

Code availability

All custom scripts used to identify model-specific global warming levels, process the ETCCDI climate-extreme indices, standardize and analyse the AMOC simulations, perform the statistical analyses, and generate the figures and tables will be deposited in a public repository before publication. The repository will include the Bash and MATLAB scripts, model-realization mappings, GWL crossing years and associated 20-year windows, and the Climact2 processing settings. The GDPCIR projections were supplied as bias-corrected data, and no additional bias correction was performed. The code is available from the corresponding author upon reasonable request.

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Author contributions

E.O.E. and B.O.A. conceived and designed the study. E.O.E., K.A.I., and M.T.D. performed the formal analyses, data visualization, and figure preparation. E.O.E., K.A.I., and H.C.N. curated the datasets. E.O.E. and M.T.D. prepared the original manuscript draft. H.C.N,

K.A.I., and S.H. critically reviewed and revised the manuscript. All authors contributed to the interpretation of the results, reviewed the manuscript, and approved the final version.

Competing interests

The authors declare no competing interests.

Additional information

Supplementary information is available for this paper.

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