

Anthropogenic aerosol forcing of European windstorms in CMIP6 climate models

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Abstract.

A recently developed set of historical storm reconstructions, ~~which~~ were extensively validated by insurance loss data, ~~and~~ revealed how European windstorm damages were three times higher in the 1980s and '90s compared to a few decades before
10 and since. A better understanding of these slower fluctuations could improve how this costly risk is managed. Here, we explore the impacts of anthropogenic aerosols (AA) on European property damage using results from DAMIP (Detection and Attribution Model Intercomparison Project) climate model experiments. Multimodel mean DAMIP results indicate AA boosted European wind losses by 45% in the late 20th century relative to preindustrial times, with the signal varying from zero to 100% between the six models. A ~~review~~validation of these modelling results using independent data from previous climate
15 studies suggested the signal is more likely to be at the higher end of this range, ~~though significant uncertainties remain. The results indicate. While this evidence suggests~~ AA forcing ~~could have been a major driver of~~contributed significantly to recent multidecadal changes in European windstorm losses, ~~there remains significant uncertainties such as the observational data used to assess model responses.~~ Further research ~~into observational and modelling uncertainties~~ would benefit those exposed to this ~~risk~~hazard.

20 1 Introduction

European windstorm activity has varied at multidecadal scales over the past few centuries (e.g. Dawson et al., 1997; WASA Group, 1998; Brázdil et al., 2004; Mellado-Cano et al., 2019; Hu et al., 2022; Brönnimann et al., 2025). The most recent and best-documented variations in storminess were particularly large, with average property damages (indexed to 2022) rising from €2.8 Bn in 1960-1979 to €6.7 Bn in 1980-1999, before falling to €2.5 Bn in 2000-2019, based on the storm loss
25 reconstructions in Cusack (2023).

~~Research into these long-~~The causes of longer timescale changes in European wind climate ~~has identified four main drivers, namely major tropical volcanic eruptions, internal climate variability, solar variations, and anthropogenic aerosol forcing. The evidence pointing to their impacts on multidecadal storminess is now summarized.~~

The first driver to be considered is are not fully understood. Zanchettin et al. (2013) presented evidence that major volcanic eruptions injecting huge amounts of sulphur into the tropical stratosphere. Empirical studies of a variety of climate reconstructions suggest this type of volcanic eruption strengthens the European winter wind climate at multiannual (e.g. Groisman, 1992; Robock and Mao, 1995; Shindell et al., 2004; Fischer et al., 2007; Ortega et al., 2015; Birkel et al., 2018; Brönnimann et al., 2025; Orme et al., 2025) and decadal timescales (Dawson et al., 1997; Zanchettin et al., 2013). Further, the quasi-periodicity of major tropical volcanic eruptions over the past 600 years (Ammann and Naveau, 2003) has coincided with the phase and timescale of multidecadal variations of climate contribute to some extent. They reported positive anomalies in the North Atlantic sector (e.g. Otterå et al., 2010; Mann et al., 2021; Dai et al., 2022). More detailed analysis has identified the mechanisms connecting eruptions to stormier conditions (e.g. Robock, 2000; Stenchikov et al., 2009; Mignot et al., 2011). Climate models consistently produce slower climate variations in the North Atlantic sector in the absence of all external forcings, which is referred to as internal variability (e.g. Zhang et al., 2019; Fang et al., 2021). This multidecadal driver can be broken down into two different components. The first concerns the Atlantic Meridional Overturning Circulation (AMOC) system of currents flowing through the full depth of the Atlantic Ocean. Variations in AMOC strength cause changes in northward heat transport (e.g. Yan et al., 2018; Robson et al., 2023), which alter meridional heat gradients in the northern Atlantic, which in turn modulate storminess in the overlying atmosphere (e.g. Brayshaw et al., 2009; Gastineau and Frankignoul, 2012; Peings and Magnusdottir, 2014; Omrani et al., 2014; Woollings et al., 2015). The mechanisms that cause decadal-scale variations in AMOC are not fully understood, though anomalous atmosphere forcing over the northern Atlantic is considered critical (e.g. Böning et al., 2006; Buckley & Marshall, 2016). The second component is atmosphere-based, and consists of shorter-timescale extremes of sufficient magnitude to alter multidecadal averages of storminess. This could be due to extremes in phenomena which drive shorter-timescale variations, such as the Quasi-Biennial Oscillation (QBO; Baldwin et al., 2001), El Niño-Southern Oscillation (ENSO; Brönnimann, 2007) and the NAO for up to approximately 15 years after major tropical sulphur-rich eruptions, with NAO anomalies in several consecutive five-year periods significantly different from zero at around the 5% level. The 11-year solar cycle (Gray et al., 2016), and their interactions (Labitzke and Kunze, 2009). In addition, the rare chance that weather conditions conducive to extreme events also hit large exposure centres can affect multidecadal mean losses.

There is a significant amount of evidence that the main 11-year solar cycle drives has been established as a driver of changes in European winter winds (e.g. Ineson et al., 2011; Scaife et al., 2013; Thiéblemont et al., 2015; Gray et al., 2016; Kuroda et al., 2022). The amplitudes of solar cycles vary at century timescales, and influence European winter climate (e.g. Otterå et al., 2010; Wang, 2017). This driver appears to be a minor contributor to storm changes over the past several decades, because the modulation of recent solar cycles corresponds to weak forcing (e.g. Figure 2.10 of Gulev et al., 2021), and was out of phase (2021). Multidecadal storm variations have also been reported to emerge from internal variability of the climate system. Climate model simulations with constant external forcings produce slower climate variations in the North Atlantic sector (e.g. Zhang et al., 2019; Fang et al., 2021), which can affect storminess (e.g. Brayshaw et al., 2009;

Gastineau and Frankignoul, 2012; Peings and Magnusdottir, 2014; Omrani et al., 2014; Woollings et al., 2015). ~~the multidecadal rise in storminess during the second half of the 20th century.~~

~~Anthropogenic~~Recent research indicates anthropogenic aerosols (AA) may also have exerted a large significant influence on temperatures in the North Atlantic (e.g. Booth et al., 2012) and Arctic (e.g. Aizawa et al., 2022) regions in the 20th century, and recent research indicates their influence extends to European wind climate. For example, Qin et al. (2020) found steeper meridional surface pressure gradients over Europe in the late 20th century in the multimodel means from both the fifth and sixth phases of the Coupled Model Intercomparison Project (CMIP5 and CMIP6, e.g. Eyring et al., 2016), while Hassan et al. (2021) found reported AA-forced declines in AA from 1990 to 2020 caused about one-third of the observed slackening of pressure gradients between the subpolar Atlantic and Mediterranean areas, in CMIP6 simulations. Needham and Randall (2023) and Needham et al. (2024) analysed a large ensemble of simulations from multiple forcing experiments, and found(2024) described how changes in AA emissions over North America and Europe caused a strengthening storm track in northern mid-latitudes over the second half of the 20th century, and subsequent weakening this century.~~Observed multidecadal changes in European storm losses align with the AA forced signals, in their climate model. Further, a reconstruction of windstorm losses (with socio-economic effects removed) in Cusack (2023) contains anomalies that co-vary with storm track changes over the past few decades. These recent results point to a material contribution from AA forcing to multidecadal loss variations over several decades.~~

At the present time, there are limits to our understanding of multidecadal variations of storminess. Reliable observational records of the European storm climate consist of one complete multidecadal cycle, which is insufficient to describe typical historical variations, and impossible to define the relative roles of various drivers and their interactions. Meanwhile, the information from modern climate models is compromised by documented problems simulating the processes generating low frequency storm variations. Specifically, volcanic and solar forcings both operate via a stratosphere pathway and models have known weaknesses in their simulation of processes in this layer of the atmosphere, and their effects on storminess (e.g. Hermanson et al., 2020; Bushell et al., 2020). Further, the Atlantic Meridional Overturning Circulation (AMOC) has been found to influence storminess at longer timescales, and models tend to underestimate the low frequency variability of AMOC (e.g. Yan et al., 2018) and ocean forcing of cyclones (e.g. Sheldon et al., 2017). The size of AA forcing of storminess is of particular interest to those exposed to wind risk, because future aerosol emissions from northern mid-latitudes are regarded as more predictable than those from natural forcings. Empirical analysis has a fundamental difficulty distinguishing AA forcing from multiple other drivers during the relatively short multidecadal period in which AA forcing was strong. Therefore, in this study, we use climate models to quantify the AA forcing of European storminess. The persistence of these modelling issues indicates full solutions are unlikely to be found soon.

While a complete understanding of multidecadal variations may require time to improve observational records and modelling, there remain opportunities to make progress in understanding multidecadal storm variations in the present-day. For instance, external forcing by AA has only recently been connected to storm tracks, and investigations into the modelled impacts on European windstorm risk have not been done. Such investigations would be worthwhile, since AA climate forcing originates

in the troposphere hence the fidelity of its simulations may differ from those forcings acting along stratosphere and ocean pathways, described above. Moreover, a focus on AA forcing of storminess enables more effort to be spent on measuring and validating model simulations, and learning more about this forcing. Finally, from a practical view, future aerosol emissions from northern mid-latitudes are more predictable than other forcings (e.g. major, tropical volcanic eruptions), hence more knowledge of AA-drive storm impacts has greater potential to improve management of future wind risk. These considerations suggest a study of the AA forcing of multidecadal storm loss anomalies could be of value to the insurance industry. Here, we analyse special climate model experiments to estimate the AA forcing of European storm losses. Near-surface winds from the DAMIP (Detection and Attribution Model Intercomparison Project) set of climate model experiments (Gillett et al., 2016) were processed into storm events, then an established model converted these event-winds into property losses-of-concern-to-the-insurance-industry-. The contribution of AA forcing was found by comparing DAMIP-AA forced results to control simulations with external forcings set to preindustrial values. Section 2 contains a description of the DAMIP climate model experiments, and the conversion of their near-surface winds to European storm losses. Results from the DAMIP-AA-forced model experiments are presented in Section 3, then explored-furtherassessed in Section 4, with-the-aid-of using additional information from previously-published research. Conclusions are given in Section 5.

2 Data and methods

2.1 DAMIP climate model data

CMIP6 includes a series of 23 sub-projects targeting specific questions about the climate. In particular, the DAMIP sub-project was designed to measure the relative roles of various climate forcings during the industrial period. Its Tier 1 experiments consist of climate model simulations covering the 1850-2014 period, with one type of forcing set to historical values and all others fixed at preindustrial values, and initial conditions from corresponding preindustrial control runs (from the main climate model experiments of CMIP6). We focus here on results from the DAMIP Tier 1 test of AA forcings. This study used climate models from six different modeling centres with the necessary diagnostics for at least five different ensemble members for the Tier 1 test, at the time of download in July and August 2024. Table 1 summarizes the models and simulations.

Table 1. Summary details of DAMIP-AA climate model simulations analysed in this study.

Model	No. of ensemble members	Control length (yr)	Atmosphere resolution Horizontal cells; vertical levels	Ocean resolution (km)
CMCC-CM2-SR5	8	500	288 x 192; 30 levs to ~ 2 hPa	100
CanESM5	9	600	128 x 64; 49 levs to 1 hPa	100
HadGEM3-GC31-LL	15	500	192 x 144; 85 levs to ~ 0.01 hPa	100

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<u>MIROC6</u>	<u>10</u>	<u>500</u>	<u>256 x 128; 81 levs to 0.004 hPa</u>	<u>100</u>
<u>MPI-ESM1-2-LR</u>	<u>15</u>	<u>600</u>	<u>192 x 96; 47 levs to 0.01 hPa</u>	<u>250</u>
<u>MRI-ESM2-0</u>	<u>5</u>	<u>250</u>	<u>320 x 160; 80 levs to 0.01 hPa</u>	<u>100</u>

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Gillett et al. (2016) ~~suggested~~recommended that starting conditions for different ensemble members be taken from well-separated states of the long Control integration. The historical AA forcings for the forced simulations are based on inventories in the Community Emissions Data System described in Hoesly et al. (2018), with associated radiative forcings described in several studies (e.g. Lund et al., 2018, 2019; Gulev et al., 2021; Bauer et al., 2022). In brief, sulphates have ~~produced~~been responsible for the largest multidecadal anomalies of shortwave radiative ~~forcings~~forcing in the industrial period, growing from a relatively small amount at the start of the 20th century to ~~peaks~~peak negative values in the 1980s and 1990s. The main source regions for 20th century sulphates were Europe, Russia, and to a lesser extent North America (e.g. Lund et al., 2019). There have been large cuts in sulphate emissions from these territories since the late 20th century leading to much smaller burdens this century. This is notable, ~~since~~because studies such as Diao et al. (2021) and Needham and Randall (2023) found the strength of the storm track in the northern hemisphere was more sensitive to emissions from mid-latitude regions.

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2.2 Converting Model Winds to Losses

Windstorm losses in climate simulations ~~were~~have been estimated using a method published by Klawns and Ulbrich (2003). ~~Results from the method were~~, which was validated ~~by a long~~using timeseries of insurance losses ~~which~~. Further insights

into this method were provided in a review of storm damage functions by Prah1 et al. (2015). Its established validity has led to its widespread use (e.g. Leckebusch et al., 2007; Pinto et al., 2007; Karremann et al., 2014; Priestley et al., 2024). A brief overview of how their method is applied is now given.

Daily maximum near-surface (10 m) winds in the October to April windstorm season were extracted from model simulations for every grid cell in a study domain consisting of 16 countries in northern and central Europe which experience the vast majority of all insured wind losses, shown in Figure 1. The first stage of processing defines the 98th percentile of the daily maximum wind was computed in each grid cell ($v_{i,98}$). This quantity is defined uniquely for each model, based on all windstorm seasons in its Control simulation, then (see Table 1). The second stage is to calculate a proxy of wind damage (D_d) in day d for the whole domain was, for each ensemble member of each model, defined as:

$$D_d = \sum_{i=1}^N \left[\max \left(\frac{v_{i,d}}{v_{i,98}} - 1, 0 \right) \right]^3 \quad (1)$$

where there are N grid cells in the domain, $v_{i,d}$ is the daily maximum wind for the ith grid cell on day d, and $v_{i,98}$ is the climatological 98th percentile of wind for the ith grid cell. Storm defined above. The third stage is to identify storm events (s) are then in each model integration: an event is defined as a series period of up to three days centred on the days with peak values of D_d , and though may be of shorter duration in practice when the maximum winds are below the 98th percentile throughout the domain. For each identified storm, the event maximum wind ($v_{i,s}$) is set to the maximum of the daily values comprising the storm, for each grid cell.

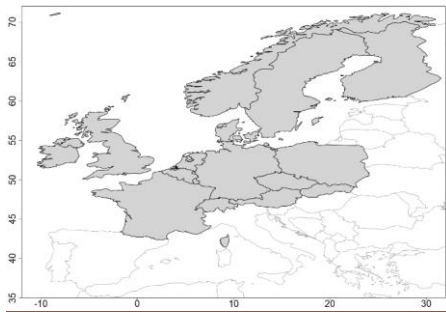
Finally, domain-wide event losses were estimated from the storm-maximum winds:

$$L_s = c \cdot \sum_{i=1}^N P_i \left[\max \left(\frac{v_{i,s}}{v_{i,98}} - 1, 0 \right) \right]^3 \quad (2)$$

where L_s is the loss for storm s, and P_i is the population count for the ith cell from Gridded Population of the World, version 4, at 2.5 minutes of arc resolution (CIESIN, 2018). The constant of proportionality (c) in the above loss equation was used by KU03 to re-scale their index values to actual loss data. This study analysed the relative change in losses, and setting this coefficient to unity does not alter results.



165 While populations evolve over time, we follow the common practice of using population data fixed to a common year when applying Eq. (2). This ensures all modelled loss anomalies are caused by changes in storm climate, in isolation from socio-economic factors that drive changes in nominal losses.



170 **Figure 1: A map of Europe with grey shading highlighting the region for which losses were computed.**

Losses from a model simulation with AA forcing were converted to anomalies using the appropriate control simulation, then grouped into ensemble means for each model, and the multimodel mean too, using linear averaging. Subsequently, the storm anomalies were filtered using a low-frequency second-order Butterworth filter with a 20-year cutoff, to highlight the multidecadal timescale of variation in [resultsplots](#).

Uncertainties2.3 Statistical testing

Two types of statistical tests have been used to measure the significance of modelled loss anomaly amplitudes.

First, uncertainties in multimodel mean quantities due to sampling errors have been estimated using the standard error of the mean, as follows. The this quantity. The first step defined the standard deviation of the annual storm anomaly for model m (σ_m) is computed using the full time series of unfiltered annual means from each model's corresponding Control run integration, then the standard error ($\epsilon_{m, \sigma, N}$) of its N -year model-mean for an AA-forced simulation is given as $\sigma_m / \sqrt{N}$, divided by the square root of N . The standard error of the N -year multimodel mean anomaly is computed similarly, with adjustments for the varying defined as the average standard error over all simulations, divided by the square root of the number of members per model (all simulations (62, see Table 1).

Second, uncertainties in ensemble mean signals from each model are assessed using t-tests of a null hypothesis that each model's mean anomaly over the 1960-2009 period was zero. This time period was chosen to maximise the signal-to-noise in model results, based on physical and statistical considerations. In more detail, Needham and Randall (2023) identified northern mid-latitude storm tracks to be more sensitive to AA emissions from the same zone, and their Figure 6 indicates the period 1960-2009 covers the period of maximum AA forcing (i.e. larger signal), and sufficiently long to reduce sampling error in results (i.e. lower levels of noise). The testing of each model consists of inputting unfiltered ensemble-mean loss anomalies for each year in the 1960-2009 period into a t-test, to assess the extent to which the mean value over this time period is consistent with no response, given the variability of unfiltered annual anomalies.

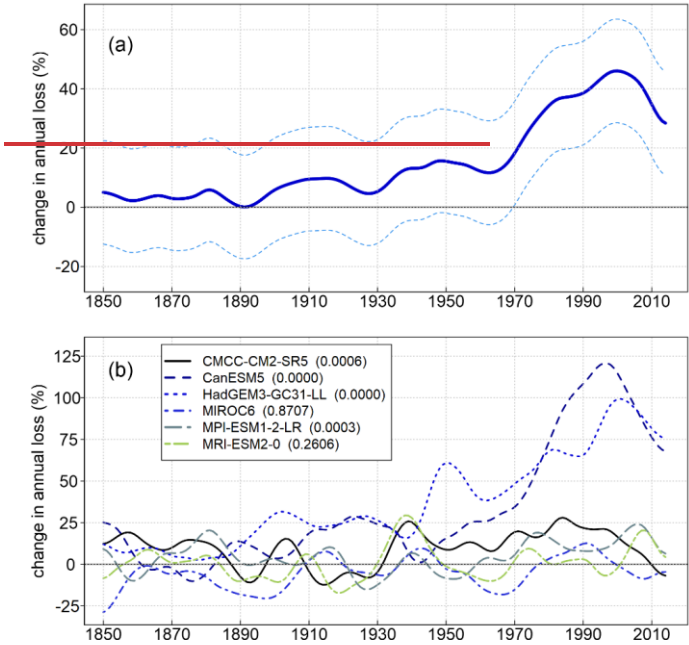
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3 Results

Figure 2a shows the timeseries of multimodel mean change in Europe-wide storm losses due to AA forcing, based on all 62 DAMIP-AA simulations. The main feature is a DAMIP-AA model peak at the end of the 20th century with 45% greater windstorm losses than its preindustrial control run; the model signal is at least four more than 10 standard errors above zero throughout the last two decades of the 20th century and indicates significant increases in European windstorm losses due to AA forcing. This conclusion is strengthened by the observation that the timing of the peak multi-model mean loss anomalies coincides with peak AA forcing from northern mid-latitudes (e.g. Lund et al., 2019). An observed loss reconstruction for the 1950-2023 period is also shown in Fig. 2a, and it can be seen that the AA-forced changes have smaller magnitude and slightly different phase from that experienced over the past seven decades. This is expected, since research indicates forcing from two major volcanic eruptions in the late 20th century and internal variability will exert influence on multidecadal variations in storminess.

AA-forced anomalies were computed for individual model ensemble-means, and shown in Figure Fig. 2b. A t-test was performed on the p-values of the null hypothesis that each model's ensemble mean anomaly was zero, using over the longer 1970-1960-2009 period for more robust results. In more detail, the t-test takes ensemble-mean loss anomalies for each year in the 1970-2009 period as input, then assesses whether the mean value over the whole period was zero is consistent with no response, given the variability of annual anomalies. The p-values of the null hypothesis for each model, given in parentheses in the legend of Figure Fig. 2b, These p-values indicate four models with have a signal above zero at the 99.9% confidence

level, including two models projecting a near-certain increase in storm loss due to AA. However, two models simulate no significant AA forcing on European storm losses. At this point in the analysis, there is no evidence to prefer one model over another, and the fact that only four of six have a significant signal suggests AA forcing of storminess is likely, but less certain than that portrayed by the multimodel mean [changesignal](#) in [FigureFig. 2a](#). It can be seen from [FigureFig. 2b](#) that the size of the multimodel mean signal is mainly due to strong AA forcing of storminess in two of the six models (CanESM5 and HadGEM3-GC31-LL). This wide range of storm responses to AA forcing indicates large uncertainty in the amplitude of the AA signal, and a key feature of model results. Additional information on the AA forcing and climate responses in the six models will be presented in Section 4, to reduce [this](#) uncertainty in [resultsstorm responses to this forcing](#).



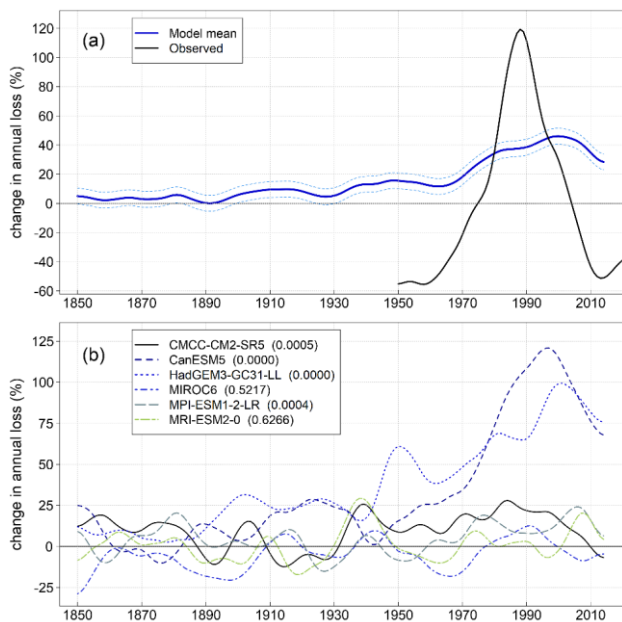
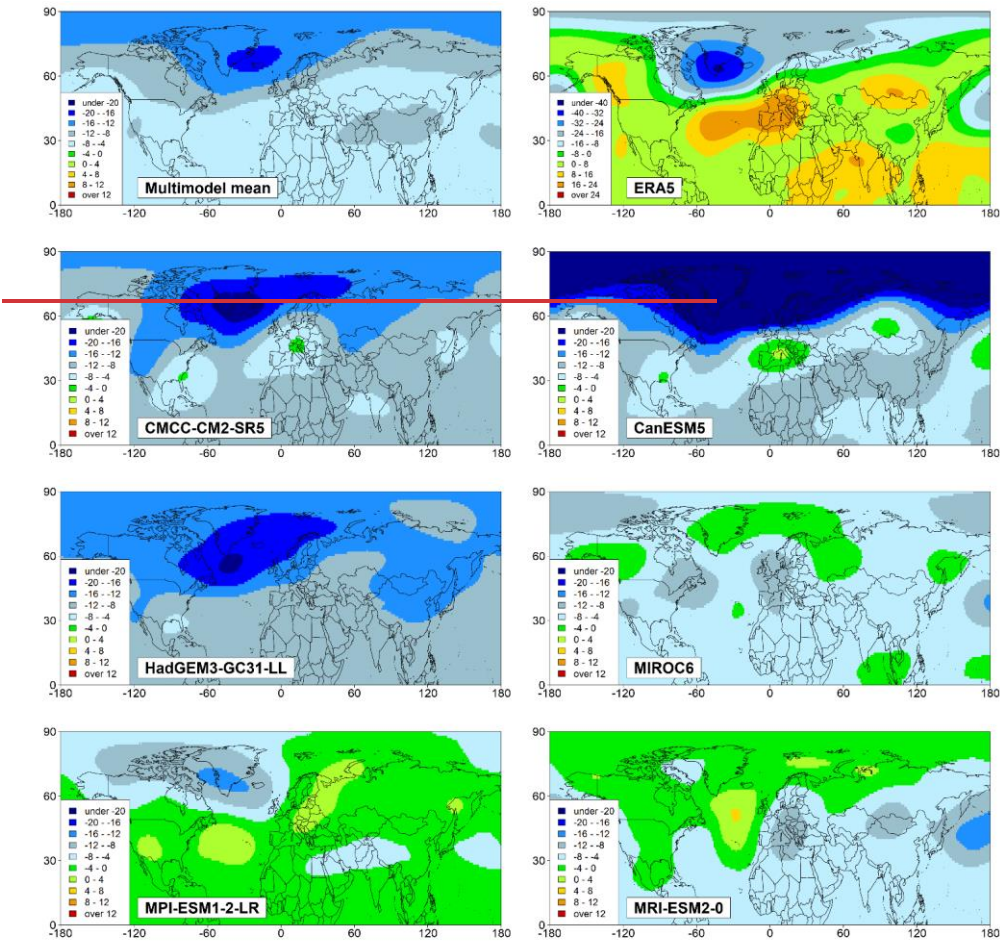


Figure 2: AA-forced change in modeled losses. (a) Timeseries of change (%) Change in DAMIP multimodel mean Europe-wide storm losses due to AA forcing, together with (solid blue line) and uncertainty (two standard errors of the mean, dashed lines), together with observed anomalies from the windstorm loss reconstruction in Cusack (2023). Note that observed anomalies are expressed relative to its long-term mean (1950-2023), whereas model anomalies are relative to long control integrations. (b) Ensemble-mean AA forcing of European windstorm losses in each of the six climate models. The p-value of a null hypothesis that the model has zero anomaly in the 1970-2009 period is given in the legend.

The evolution of circulation anomalies through the 20th century are summarised in Figure 3, showing Figure 3 presents maps of the anomalies in the 500 hPa geopotential heights (ϕ_{500}) for the 30-year period 1970-99 versus the earlier, less stormy 1940-69 period, for both the DAMIP-AA multimodel mean, ERA5, and ensemble means from each ERA5 reanalyses, to illustrate the evolution of the six DAMIP-AA models. The top-left panel circulation anomalies in the later part of the 20th century. Figure 43a indicates how the DAMIP-AA multimodel mean contains significant meridional gradients of ϕ_{500} anomalies over the northern Atlantic area, with an amplitude of about 12 m. This anomalous gradient is about one-third of the magnitude in the ERA5 reconstruction (top-right panel of Figure 4). This relativity between model and observed is similar for storm losses, where the multimodel anomaly (Figure 2) is roughly one-quarter of that from an extended set of storm loss reconstructions in Cusack (2023). Further, the individual model means show a similar consistency: their changes in losses in Figure 3 are generally aligned with the size of the model's meridional gradients in ϕ_{500} anomalies, with CanESM5 containing the steepest

gradient in ϕ_{500} anomalies, etc. In brief, storminess changes are generally consistent with circulation anomalies through the 20th century (Fig. 3b), and consistent with the model-to-observed relativity found for losses (Fig. 2a). Therefore, the changes in modelled storm losses are physically connected to larger-scale circulation anomalies.



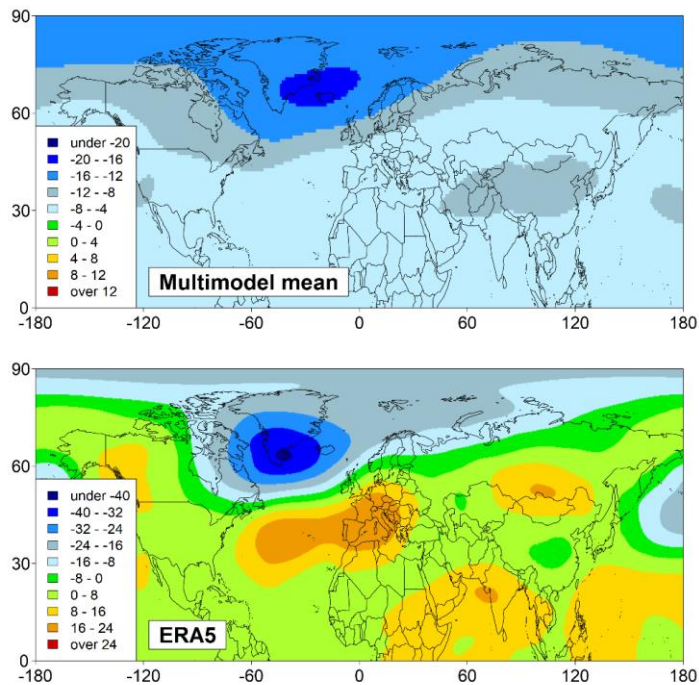


Figure 3: geopotential height anomalies at 500 hPa (m) for the 1970-1999 period relative to 1940-1969, for (a) DAMIP-AA multimodel-mean, and (b) ERA5.

The corresponding circulation anomalies for each model ensemble are shown in Figure 4. A similar consistency between circulation and loss anomalies is found, namely that the changes in losses in the late 20th century for each model (Fig. 2b) are generally aligned with the size of the model's meridional gradients in ϕ_{500} anomalies over Europe in Fig. 4. For example, CanESM5 contains the steepest gradient in ϕ_{500} anomalies over the Euro-Atlantic sector, and also produces the largest storm loss anomalies of all models.

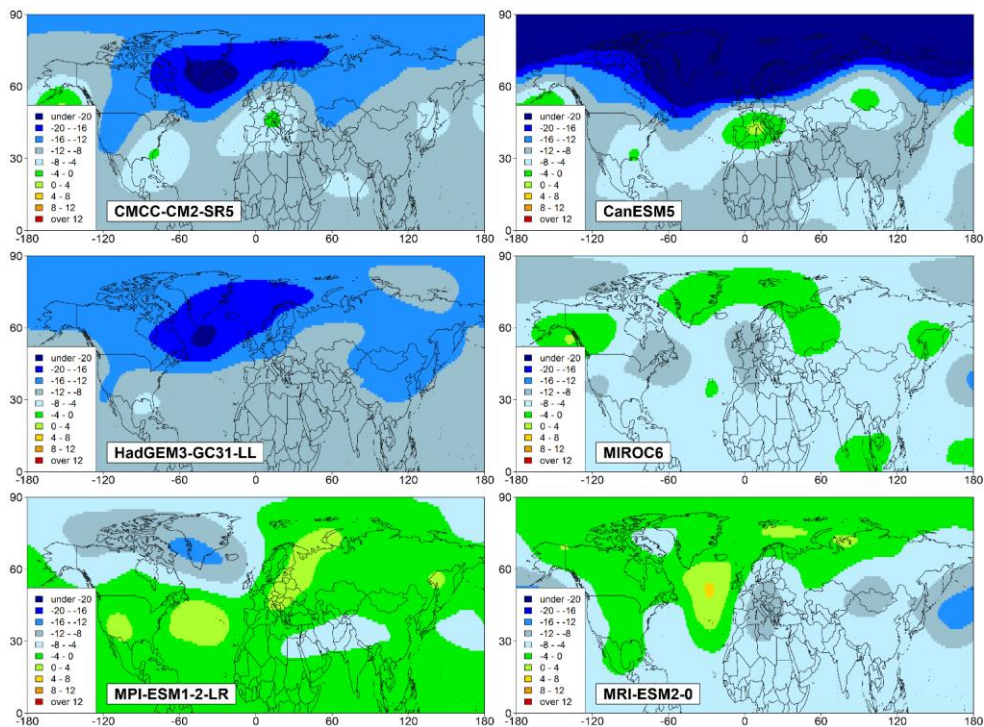


Figure 4: geopotential height anomalies at 500 hPa (m) for the 1970-1999 period relative to 1940-1969, for the DAMIP-AA multimodel-mean (top-left panel) and ERA5 (top-right panel), and the six individual DAMIP-AA model ensemble means.

4 Discussion

4 Assessment of modelled results

The spread in DAMIP-AA results suggests significant large uncertainty in the size of the AA forcing of European windstorm losses from models. For example, just only four of the six models produce a significant forcing of European windstorm losses, and further, the mean signal is largely determined by only mostly due to contributions from just two models. We review findings from climate research studies in In this section, with the intention of bringing new information and reducing our aim

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to reduce this uncertainty in-by conducting an assessment of the AA forcing of windstorm lossesmodelled storm responses using independent data from past studies.

The review begins by examining the strength of AA radiative forcing at the top of atmosphere in models and observed, using the global effective radiative forcing (ERF) diagnostic. The assessment methodology is founded on a relation between variables reported in past studies. In brief, AA radiative forcing drives changes in poleward energy transports (PET) that are manifested in the northern mid-latitudes as anomalies in both AMOC and storm tracks. Therefore, consideration of both modelled AA forcing, and associated AMOC responses to such forcing, can help understand the different storm track responses to AA forcing in models. Further, the validity of both the modelled AA forcing, and AMOC responses to it, can inform on the accuracy of modelled storm track responses. More details of this methodology are now given, before it is applied to modelling results. Needham and Randall (2023) described how the total PET is connected to the latitudinal gradient of net radiation at the top of atmosphere (TOA), therefore, the perturbed TOA radiation balance from AA forcing drives anomalies in total PET. In turn, Needham et al. (2024) showed how the total PET anomalies in northern mid-latitudes in a modern climate model (CESM2; Danabasoglu et al., 2020) are largely the sum of changes in two quantities, namely ocean heat transport via AMOC and atmosphere heat transport via eddies, or storm tracks. Therefore, the validity of the modelled change in storm tracks can be inferred from assessments of the accuracy of anomalies in both TOA radiation and AMOC in model simulations. We begin by assessing perturbations to the TOA radiation balance over northern mid-latitudes due to AA forcing, measured using effective radiative forcing (ERF). There are no published estimates of zonal-mean AA net radiation anomalies in northern mid-latitudes for CMIP6 models. Instead, we analyse global ERF, based on the assumption that it is reasonably aligned with corresponding values over the northern storm-tracks, since most of the sulphur emissions in the historical period have been from the northern mid-latitudes. Forster et al. (2021) provided a summary of the latest modelling and observational estimates of AA ERF at 2014, relative to the pre-industrial period, and includes four of the six models used in this study. Table 2 presents the data for these four models (extracted from table 7.6 of Forster et al., 2021). The IPCC authors also provide a best observational estimate of -1.3 Wm^{-2} . The mean of the four models and their average is -1.12 Wm^{-2} , which is 13% smaller magnitude than observed and suggests a best observational estimate of -1.3 Wm^{-2} from Forster et al. (2021), suggesting the DAMIP-AA ensemble used here may underestimate true AA forcing, on average. Further, the values in Table 2 show little alignment between Forster et al. (2021) also indicate it is very likely that the true global AA ERF values and storm loss changes for these four models. This is illustrated by both MIROC6 and MRI-ESM2-0 having the smallest storm changes due to AA forcing in 2014 lies in Figure 2b, yet their ERF values are the smallest and largest values of the four models, respectively. Moreover, the spread range from -0.6 to -2.0 Wm^{-2} , which in global turn suggests some possibility that the AA ERF between the four models is within 12% of the mean, and much smaller forcings in these models are stronger, than the spread in their loss impacts true value. In conclusion summary, the studied models tend to have slightly weaker global radiative forcing from AA than best observational estimates, and the inter-model differences in strength of global ERF do not explain differences in their AA forced storm loss anomalies though within the very likely range of true forcing.

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300 **Table 2.** DAMIP-AA climate model diagnostics. Global top-of-atmosphere effective radiative forcings are [taken](#) from Forster et al. (2021). The difference in the AMOC trends between 1950-1990 (increases) and 1990-2020 (reductions) are [taken](#) from [Figure S1 of Hassan et al. \(2021\)](#). The final column shows the AA-forced windstorm loss anomalies in the 1970-1999 period relative to corresponding control runs, from results presented in Sect. 3.

Model	Global AA forcing (ERF, Wm ⁻²)	Change in AMOC trends (%)	1970-1999 loss anomaly (%)
CMCC-CM2-SR5	---	---	23.0
CanESM5	-1.11	8	87.2
HadGEM3-GC31-LL	-1.17	13	69.3
MIROC6	-0.99	21	4.3
MPI-ESM1-2-LR	---	10	11.3
MRI-ESM2-0	-1.22	58	5.9

305 ~~The next step is to explore how AA radiative forcings affect the climate system. Needham and Randall (2023) framed this problem in terms of zonal-mean poleward energy transport (PET), and analysed how the radiative forcing perturbed the energy flows between zones at a planetary scale. While this study concerns impacts specific to the European sector, the zonal-mean scale of their~~
310 ~~The validity of CMIP6 model simulations of historical AMOC anomalies has been addressed by both Menary et al. (2020) and Hassan et al. (2021). Both studies conclude that the AMOC changes over the past several decades from an ensemble of CMIP6 models were largely driven by AA forcing, and the magnitude of the simulated AMOC anomalies are likely to be larger than best estimates of observed change. However, there are significant uncertainties in the estimates of AMOC anomalies and trends throughout the 20th century. Direct AMOC measurements began in 2004 with the RAPID array (Rayner et al., 2011), and estimates before this time span a wide range, extending to uncertainty in the sign of 20th century trends as well as its magnitude (e.g. Tett et al., 2014; Karspeck et al., 2015; McCarthy and Caesar, 2023). In this context, it is~~
315 ~~notable that the estimates of historical AMOC used by both Menary et al. (2020) and Hassan et al. (2021) were based on relations between sea surface temperature (SST) anomalies in the North Atlantic’s subpolar gyre and AMOC, from CMIP5 climate models (Rahmstorf et al., 2015). This SST proxy tends to estimate significantly weaker AMOC trends in the 20th century relative to other reconstruction methods, as can be found from comparing SST-based AMOC reconstructions in Rahmstorf et al. (2015) with those from ocean reanalyses (e.g. Danabasoglu et al., 2016) or ocean observations (e.g. Fraser and Cunningham, 2021). These estimates of weaker AMOC trends and may be due to relative SST anomalies in the subpolar gyre being driven by other processes, in addition to AMOC deviations (e.g. Keil et al., 2020; Little et al., 2020). Therefore,~~

325 while CMIP6 models tend to simulate AMOC changes in the 20th century that are more positive than those reconstructed from SSTs, it is unclear whether models are above the true AMOC trends during that period.

The above validation indicates CMIP6 models produce slightly low magnitudes of net TOA radiation anomalies, and slightly
330 too large AMOC anomalies, which in turn implies modelled storm track responses to AA forcing are generally too weak in CMIP6 models. However, uncertainties in the reference historical estimates of net TOA radiation and AMOC are large, and include the possibility that simulated storm loss changes are valid.

Other sources of uncertainty exist, in addition to those related to past anomalies of ERF and AMOC. For example, the zonal-mean framework of the analysis is consistent with the observed zonal scale of storm track anomalies in the late 20th century
335 (e.g. Chang and Fu, 2002; Woollings et al., 2014), hence zonal-scale findings are relevant to recent multidecadal variations in European windstorm.

Needham and Randall (2023) showed how ERA5 total PET values in northern mid-latitudes increased from 1959 but it is likely that processes local to the 1980s, then declined to 2014, and how PET anomalies from CESM2 historical simulations were consistent with ERA5 observational estimates. They proceeded to analyse various CESM2 external forcing experiments, and
340 found AA forcing to be Atlantic sector create a zonally asymmetric response, to some extent. Further, the main cause validation used global AA ERF as a proxy of the rise then fall strength of PET values in northern mid-latitudes from the mid-20th century to the present day. Needham et al. (2024) broke down the total PET into more detailed component parts, and their Figures 3e and d indicate how the northern mid-latitude heat transport anomalies in the 1960–1989 period were mainly caused by the North Atlantic Ocean heat transport (via the Atlantic Meridional Overturning Circulation, AMOC), together with a smaller contribution from atmosphere eddies. In brief, the observed increase of PET in northern mid-latitudes from the mid- to late 20th century was driven by AA forcing, and was manifested as increases in AA forcing, which is reasonable, but will introduce some error. On the other hand, extra data in Table 2 offer supporting evidence for a validation framework that equates AA forcing to the sum of PET anomalies from both AMOC and storm track intensity.

Menary et al. (2020) described how AMOC changes over the past several decades simulated by an ensemble of CMIP6 models
345 were likely to be larger than observed tracks. Hassan et al. (2021) provides quantitative information on the provided AMOC trends for a selection of CMIP6 models in two consecutive periods, one of which has with an upward trend (1950–1990) and the other with a declining trend (1990–2020), and they present data for five of the six DAMIP AA models studied here. Were reported how the change in the trend was largely driven by AA forcing. Here, we use the difference in their trends (%) from the earlier to later period (i.e. [1950–1990 trend] minus [1990–2020 trend]), in % as a measure of a model’s AMOC responsiveness to change in AA forcing, and results are listed in Table 2. These tabulated values data contain a pattern, whereby models with greater AMOC changes have smaller European windstorm loss changes in the late 20th century. For example, the two models with the smallest change in storminess (MRI-ESM2-0 and MIROC6) have the largest change in AMOC trend (58% and 21% respectively), while the two models with the largest change in storm losses (CanESM5 and HadGEM3-GC31-LL) have smaller changes in AMOC strength (8% and 13% respectively). Moreover, Hassan et al. (2021) identified CanESM5
350 as one of the two models in their ensemble of 24 CMIP6 models which were consistent with observed AMOC trends over the

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past 90 years, and this model simulates the largest change in storminess of the six DAMIP-AA models studied here. This suggests DAMIP-AA simulations with less responsive AMOC to AA forcing are more consistent with estimates of past AMOC changes, and Table 2 indicates this model subset produces larger storm loss changes on average, and vice-versa. Such compensation between AMOC and storm track anomalies supports the assessment method.

In conclusion, summary, findings from past studies were used to measure the review of climate research studies validity of modelled storm track responses to AA forcing. They suggest CMIP6 models may underestimate mid-latitude storm increases in the late 20th century, because (a) the magnitude of modelled AA forcing is slightly smaller than best estimates of observed, and (b) models tend to overestimate AMOC responses to recent AA forcing, implying an underestimate of which would tend to reduce their storm track response to the increases same forcing. However, confidence levels in these findings are lowered by the storm track severity large uncertainties in estimates of AA forcing and AMOC in the late 20th century. However, significant sources of uncertainty remain. First, Forster et al. (2021) indicate it is very likely that global ERF due to AA forcing in 2014 lies in a range from -0.6 to -2.0 Wm⁻², therefore, it is possible that the models studied here have too strong AA forcing (mean of -1.12 Wm⁻² from Table 2). Second, the total PET anomaly in response to Further research into northern mid-latitude AA forcing has been validated for the CESM2 model, however, actual PET radiative forcing, and AMOC anomalies for the six models studied here are needed to confirm the assumption that this aspect of CESM2 is representative of modern climate models. Third, past AMOC values and trends are ambiguous: direct AMOC measurements began in 2004 with the RAPID array (Rayner et al., 2011) and its value in earlier periods have large uncertainty, and it is possible that AMOC was more responsive to AA forcing than current best estimates. Finally, while the PET analysis of zonal mean changes is consistent with the zonal scale of storm track anomalies in the late through the 20th century, it is likely that simulations of local Atlantic storm track processes create extra uncertainty in model storm responses displayed in Figure 2. Future research into AA forcing of storminess, including the points mentioned above, could yield a more robust estimate would improve our knowledge of the contribution of AA forcing to the pronounced multidecadal peak of wind losses in the late 20th century, and benefit those exposed to European windstorm risk.

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5 Conclusions

We measured the AA-forced contribution to recent multidecadal variations in European windstorm losses using results from multiple simulations of six climate models participating in the DAMIP-AA experiment. The multimodel ensemble mean had 45% higher European windstorm losses in the late 20th century relative to preindustrial times, though anomalies range from zero to 100% between models and indicates signify large uncertainty in the strength of AA-forced impacts.

Past climate findings from earlier studies provide additional information to understand insights into the DAMIP-AA estimates of storm responses to AA forcing. First, the global ERF radiative forcing due to AA forcing was found to be 10 to 15% lower in models than the best estimate of observed, which would lead to a low bias in DAMIP-AA storm responses. Second,

modelled AMOC responses to AA forcing are larger than ~~best~~-estimates of historical AMOC behaviour, ~~which implies an~~. ~~These two results indicate the models~~ underestimate ~~of~~ the size of storm track anomalies in the late 20th century, ~~also suggesting~~ ~~the~~~~hence~~ DAMIP-AA ~~storm responses may be~~~~windstorm loss changes are~~ biased low.

Significant uncertainties remain. For example, past values of both global ~~ERF~~radiative forcing and AMOC ~~are unclear and~~ provide relatively weak constraints on the modelled storm responses to AA forcing, while zonal-mean diagnostics ~~cannot do~~ not resolve local processes in the Atlantic sector which may modulate European storminess. Future research into these uncertainties could lead to a better understanding of past multidecadal storminess in Europe and its future trajectory, and better management of this risk.

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Code and Data Availability

CMIP6-DAMIP climate model results were downloaded from the Earth System Grid Federation (ESGF): <https://aims2.llnl.gov/search/cmip6/> (accessed July-August 2024). Model data references are as follows: CMCC-CM2-SR5 (Lovato & Peano, 2020 and 2024); CanESM5 (Swart et al., 2019a, b); HadGEM3-GC31-LL (Ridley et al., 2018; Jones, 2019); MIROC6 (Tatebe & Watanabe, 2018; Shiogama, 2019); MPI-ESM1-2-LR (Wieners et al., 2019; Müller et al., 2019); MRI-ESM2-0 (Yukimoto et al., 2019a, b).

The dataset of Europe windstorm loss reconstructions is described in Cusack (2023), with licensing restrictions, and are not accessible to the public or research community. Contact Stormwise Ltd at information@stormwise.co.uk to obtain access.

All analyses were performed using R Statistical Software (v4.2.2; R Core Team, 2022), and the low-pass filtering used the dplR package (v1.7.4; Bunn et al., 2022).

Competing interests

The author declares no competing interest.

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