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To cite this article: Florian Börgel *et al* 2022 *Environ. Res. Lett.* **17** 044040

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ENVIRONMENTAL RESEARCH
LETTERS

LETTER

OPEN ACCESS

RECEIVED

14 September 2021

REVISED

7 March 2022

ACCEPTED FOR PUBLICATION

10 March 2022

PUBLISHED

24 March 2022

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Atlantic multidecadal variability and the implications for North
European precipitationFlorian Börgel^{1,*} , H E Markus Meier^{1,2}, Matthias Gröger¹, Monika Rhein³, Cyril Dutheil¹
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E-mail: florian.boergel@io-warnemuende.de**Keywords:** climate variability, AMV, AMOCSupplementary material for this article is available [online](#)

Abstract

The North Atlantic exhibits temperature variations on multidecadal time scales, summarized as the Atlantic multidecadal variability (AMV). The AMV plays an essential role for regional climate and is a key driver of the low-frequency variability in Northern Europe. This study analyzed the interaction between the atmosphere and the ocean using Coupled Model Intercomparison Project 6 (CMIP6) control runs. The results showed that the physical mechanisms underlying decadal or longer time scales differ among CMIP6 models, which allowed them to be sorted into two clusters. For the first cluster, a significant coherence between the North Atlantic Oscillation (NAO) and the AMV was found. Further, it showed a strong negative NAO response and decreasing precipitation over Northern Europe. In contrast, the second cluster showed no significant coherence between NAO and AMV. This non-coherent cluster developed a low-pressure anomaly in the subpolar gyre and showed increasing precipitation over Europe. Differences in the northward extension of the Atlantic meridional overturning circulation (AMOC) between the two clusters were identified and linked to the different atmospheric responses. Our findings have important implications for European climate, since predictions of an increase or decrease in precipitation over Northern Europe will be model-dependent.

1. Introduction

The climate variability in the North Atlantic Ocean plays an important role in the global climate system and has strong regional impacts. The climate of the North Atlantic region is frequently assessed as the Atlantic multidecadal variability (AMV), which is related to large multi-decadal fluctuations in the Atlantic's sea surface temperature. The AMV influences summer climate, cold winter episodes, precipitation patterns in Europe, and weather patterns in Africa and North America (Enfield *et al* 2001, Knight *et al* 2006, Casanueva *et al* 2014, Peings and Magnusdottir 2014, Sutton and Hodson 2005, Ruprich-Robert *et al* 2017, Börgel *et al* 2018, Simpson *et al* 2019).

The origin of the AMV is heavily debated. There is strong observational and modeling evidence that the

multidecadal variability of the Atlantic meridional overturning circulation (AMOC) is a crucial driver of the AMV (Zhang *et al* 2019). Wills *et al* (2019) argued that the interaction between the North Atlantic oscillation (NAO) and the AMOC leads to a low-frequency response of the ocean, a conclusion also supported by many other studies (Sun *et al* 2015, Delworth and Zeng 2016, Delworth *et al* 2017). According to Wills *et al* (2019), AMV warm events are triggered by strong zonal winds associated with positive NAO (NAO+) conditions, which in turn cause an anomalous heat loss in the Labrador Sea, leading to stronger deep water formation and ultimately to a strengthening of the AMOC circulation. During the peak of an AMV warm event, the higher temperatures drive the formation of a basin-wide low-pressure anomaly, followed by negative NAO (NAO-) conditions that terminate the AMV warm event. By analyzing

observational data, Li *et al* (2013) found that the NAO precedes the AMV signal by approximately 15–20 years, a finding confirmed by other studies (Gastineau and Frankignoul 2015, Sun *et al* 2015, Delworth and Zeng 2016, Delworth *et al* 2017).

In contrast to this dynamic interpretation, which focuses on the unforced component of the AMV, some modeling studies have found that the frequency of the AMV is dominated by external forcing, such as greenhouse gases, volcanic eruptions, and stratospheric ozone (Booth *et al* 2012, Zanchettin *et al* 2013, Tandon and Kushner 2015, Bellucci *et al* 2017, Singh *et al* 2018, Bellomo *et al* 2018). A recent study by Mann *et al* (2021) further suggested that the observed periodicity in the AMV during the past millennium was driven exclusively by volcanic radiative forcing.

The NAO itself is the dominant and most recurrent mode of climate variability in the northern hemisphere during winter. For example, between 1970 and 1998, the NAO accounted for >40% of the annual variability in sea level pressure (SLP) across the northern hemisphere (Kauker and Meier 2003). The NAO can be defined as the difference in the SLP between the subpolar low-pressure system near Iceland (Icelandic Low) and the subtropical anticyclone in the Atlantic near the Azores (Azores High) (Hurrell 1995). In addition, it can be defined using empirical orthogonal functions (EOFs) (see Methods). Like the AMV, the NAO has been linked to climate variations over North America and Europe, such as storm tracks and variability in temperature and precipitation (Zorita and Laine 2000, Scaife *et al* 2008, Delworth and Zeng 2016).

A frequent shortcoming in evaluations of the AMV is that the amplitude of the simulated unforced AMV is substantially underestimated compared to observations, such that the relative importance of the forced component of the AMV is accordingly stronger (Zhang *et al* 2019). Therefore, it remains challenging to assess the importance of the unforced part of the AMV. It has also been shown that model simulations with prescribed changes in external radiative forcing, but without a proper initialization of ocean states, cannot predict the observed AMV (Zhang *et al* 2019). Therefore, the atmosphere-ocean interaction appears to be a key component of the unforced AMV. To reliably analyze the impact of the AMV on climate variability, these interactions must be captured by model simulations.

In this study, Coupled Model Intercomparison Project 6 (CMIP6) pre-industrial control simulations (piControl) were used to evaluate the impact of the AMV on Europe. CMIP6 piControl simulations are forced with pre-industrial conditions and are used as a benchmark to evaluate the natural variability of CMIP6 models. Hence, the effect of external forcing such as greenhouse gases, aerosols, ozone, and solar variability can be neglected. The piControl simulations start after an initialization

period and cover a minimum period of 500 years. Consequently, the following analysis focuses on the unforced component of the AMV and the different model responses.

As global climate models give projections on how Earth's climate may change, the impacts of changing climate and the corresponding adaptation strategies will occur on national levels. As the results of regional climate modeling impact these adaptation strategies, it is crucial to be aware of possible significant multi-decadal discrepancies for Europe.

These discrepancies become particularly important when looking at the Baltic Sea, which is in the center of Northern Europe and to which many marine protection measures in Scandinavian, Baltic, Central European countries, and Russia refer. The ecosystem of the Baltic Sea is heavily influenced by river runoff, which is accumulated over the catchment area. The catchment area covers large parts of Northern Europe. Consequently, differences in precipitation of global circulation models (GCMs) will affect regional downscaling and the results of climate projections performed for the semi-enclosed Baltic Sea (Meier *et al* 2022).

The analysis concentrated on the interaction between the atmosphere, i.e. the NAO, and the ocean, as a key physical mechanism driving the variability and the impact of the AMV. Our results showed that the models could be sorted into two clusters, depending on the coherence between the NAO and AMV. The differences in the behaviors of the two clusters were significant. They included winter precipitation responses opposite in their signs over large parts of Europe, especially Northern Europe and the Baltic Sea.

2. Materials and methods

2.1. Modes of variability

In this study, the AMV was defined as the area-weighted average sea surface temperature (SST) across the North Atlantic domain (0°–60° N, 0°–80° W). As proposed by Trenberth and Shea (2006), the global signal is removed by subtracting the global mean SST, thus allowing extraction of only the Atlantic variability.

To analyze the low-frequency component of the AMV, a ten year running mean was applied. The NAO was defined as the first EOF (Hannachi *et al* 2007), calculated from the monthly sea level pressure (SLP) anomalies of the winter season (December, January, February) (20° N–70° N; 90° W–40° E).

2.2. AMOC

The AMOC stream function Ψ was calculated using the three-dimensional meridional velocity $v(t, z, y, x)$ consisting of t (time), z (depth), y (latitude) and x (longitude)

Table 1. CMIP6 preindustrial control simulations used in this study and the number of model years. Models with meridional velocity fields available are denoted with an asterisk.

Model	Years	Model	Years
ACCESS-CM2*	500	GISS-E2-1-G*	851
ACCESS-ESM1-5*	900	GISS-E2-1-H	401
BCC-CSM2-MR*	600	HadGEM3-GC31-LL*	500
BCC-ESM1	451	HadGEM3-GC31-MM	500
CESM2-FV2*	500	INM-CM4-8	531
CESM2-WACCM-FV2*	500	INM-CM5-0	1200
CanESM5*	1000	IPSL-CM6A-LR	1200
CESM2-WACCM*	500	MCM-UA-1-0*	500
CESM2*	1200	MIROC6*	800
CIESM	500	MPI-ESM-1-2-HAM*	780
E3SM-1-0	500	MPI-ESM1-2-HR*	500
EC-Earth3-Veg-LR*	500	MPI-ESM1-2-LR*	1000
EC-Earth3*	500	MRI-ESM2-0	701
EC-Earth3-Veg*	500	NESM3*	500
FGOALS-g3	700	NorCPM1	500
FIO-ESM-2-0	575	NorESM2-MM	500
GFDL-CM4	500	SAM0-UNICON	700
GFDL-ESM4	500		

$$\Psi(t, y, z) = \int_{-HX_{\text{west}}}^0 \int_{X_{\text{east}}} v(t, z, y, x) dx dz,$$

with H as the sea bottom and X_{east} and X_{west} as the eastern and western ocean boundary of the Atlantic basin.

2.3. Wavelet analysis

The continuous wavelet transform (CWT) expands a time series into a time-frequency space, thus revealing time-localized oscillations and areas with high power. Therefore, the CWT is appropriate for studying periodic phenomena that change over time. The most common wavelet function $\psi(t)$ is the Morlet wavelet (Torrence and Compo 1998)

$$\psi(t) = \pi^{-1/4} e^{i\omega t} e^{-t^2/2},$$

with ω as non-dimensional frequency, here taken to be six. The CWT is defined as the convolution between a set of Morlet wavelet functions and discrete-time time series $x(t)$

$$\text{CWT}(\tau, s) = \sum_t x(t) \frac{1}{\sqrt{s}} \psi\left(\frac{t-\tau}{s}\right).$$

The position of a wavelet function is defined by the time parameter τ that is shifted by the time increment dt . The scaling factor s defines the temporal width of the wavelet, which determines the frequency of oscillation due to its fixed shape. Like the power spectrum, the wavelet power spectrum is defined as

$$\text{Power}(\tau, s) = \frac{1}{s} |\text{CWT}(\tau, s)|^2.$$

In summary, the CWT allows analyses of the frequency spectrum and its energy while retaining the time information.

2.4. Datasets and model simulations

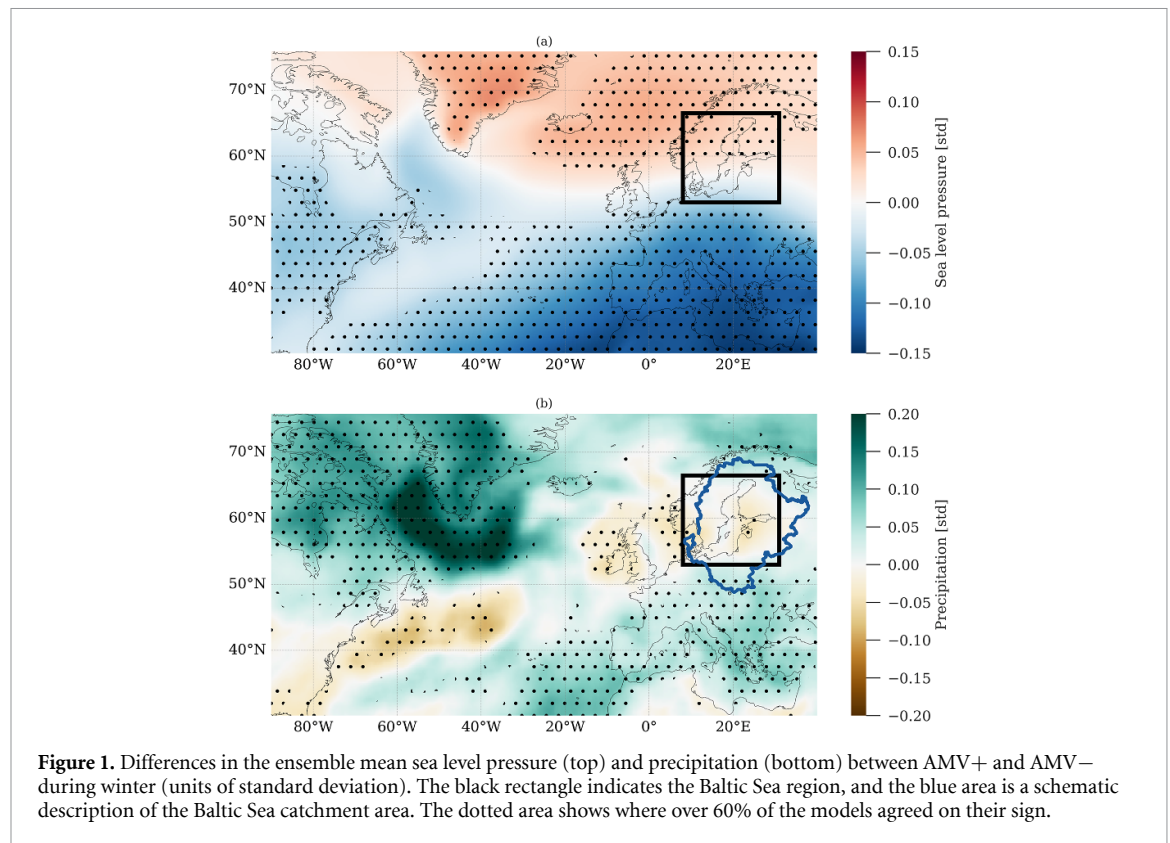
In this study, we used 33 piControl climate simulations from phase 6 of the CMIP (Eyring *et al* 2016). We included all available data to analyze variability in the North Atlantic; however, we set a threshold to at least 450 years needed for the analysis. We used sea surface temperature, sea level pressure, and precipitation for our study. Further, we used the meridional ocean velocity to compute the AMOC circulation. Unfortunately, the velocity data was only available for 21 models, denoted by an asterisk in table 1. The data were accessed using Google Cloud Storage as described by the Pangeo/ESGF Cloud Data Working Group.

The extended reconstructed sea surface temperature (ERSST) dataset used in the appendix is a global monthly analysis of SST data derived from the International Comprehensive Ocean–Atmosphere Dataset. It has a horizontal resolution of $2^\circ \times 2^\circ$ and covers the period from 1854 to 2018. As data sources it contains COADS 3.0, which combines SST from Argo floats (above 5 m), Hadley Centre Ice-SST version 2 ice concentration (1854–2015), and National Centers for Environmental Prediction (NCEP) ice concentration (2016–2018). The ERSST dataset illustrates the behavior of the forced component of the AMV.

3. Results

3.1. Response to the AMV

Individual models and their representation of the AMV differ. We find differences in amplitude, persistence, and spatial pattern associated



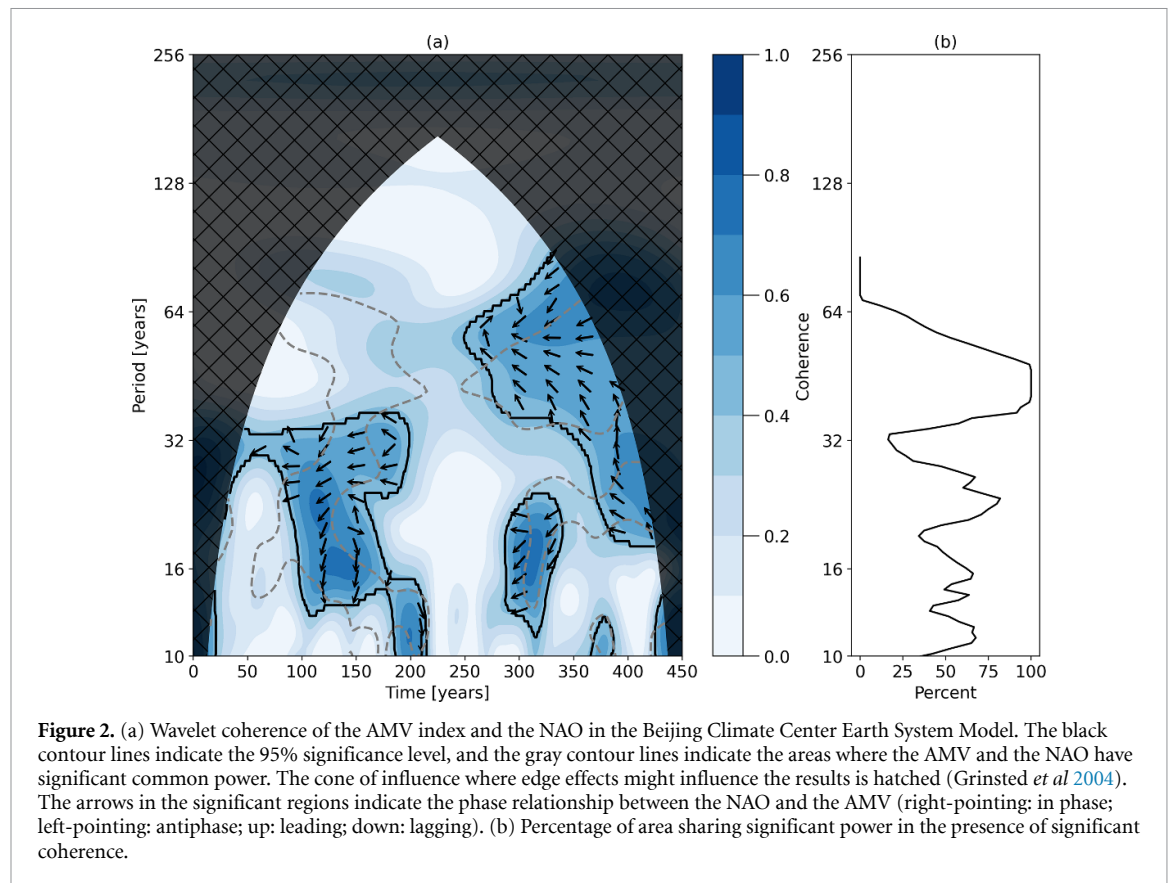
with the AMV (appendix S1 available online at stacks.iop.org/ERL/17/044040/mmedia). Unforced variability is subject to considerable uncertainty, but instead of focusing on the individual models, we analyze a multimodel composite that summarizes the driving mechanisms of the CMIP6 ensemble. The difference between AMV+ and AMV− states for SLP and precipitation during winter is computed to analyze the atmospheric response associated with the AMV. Hence, our analysis is mainly based on the difference between mean atmospheric states during AMV+ and AMV− phases. Also, the regression analysis in section 3.2 is based on linear regression. While this approach is common to previous studies, assessing nonlinear relationships could substantially improve the results, as discussed in Ruggieri *et al* (2021).

Figure 1 shows the mean response of the CMIP6 ensemble during winter associated with AMV variability in units of standard deviation. The upper panel shows the difference in SLP between AMV+ and AMV− phases, yielding a positive anomaly over Iceland and a negative anomaly over the Azores. This north-south dipole of SLP differences corresponds to an NAO− pattern. The lower panel shows the difference between AMV+ and AMV− states for precipitation. The strongest response is found in the subpolar gyre with stronger precipitation during AMV+ phases. South of the subpolar gyre, decreasing precipitation is observed. The precipitation response in the subpolar gyre and further south is likely linked

to air-sea heat fluxes associated with AMV variability, which will be discussed later. Further, lower precipitation is found for Great Britain and Northern Europe; higher precipitation for Central and South Europe. These results are in good agreement with the precipitation pattern associated with an NAO− state (Hurrell 1995). An analysis of the ensemble standard deviation showed that the AMV related SLP and precipitation responses differ most in the subpolar gyre (not shown).

The SLP difference between AMV+ and AMV− states of the piControl simulations is consistent with an NAO− pattern. To assess the robustness of the ensemble mean response, we hatched regions where more than 60% of the models agree on their sign. The SLP shows a good inter-model agreement. However, no agreement is found within and near the subpolar gyre matching the high variability in the mean ensemble standard deviation in this region. The model agreement for the AMV related precipitation response is generally smaller and shows no agreement over Northern Europe.

To further analyze a potential relationship between the AMV and the NAO, the coherence between the AMV and the NAO for every model is analyzed using wavelet analysis. It should be noted that the AMV and the NAO are not low pass filtered before the wavelet analysis and have an annual resolution. As an example, figure 2 shows the wavelet coherence between the AMV and the NAO for the Beijing Climate Center Earth System Model

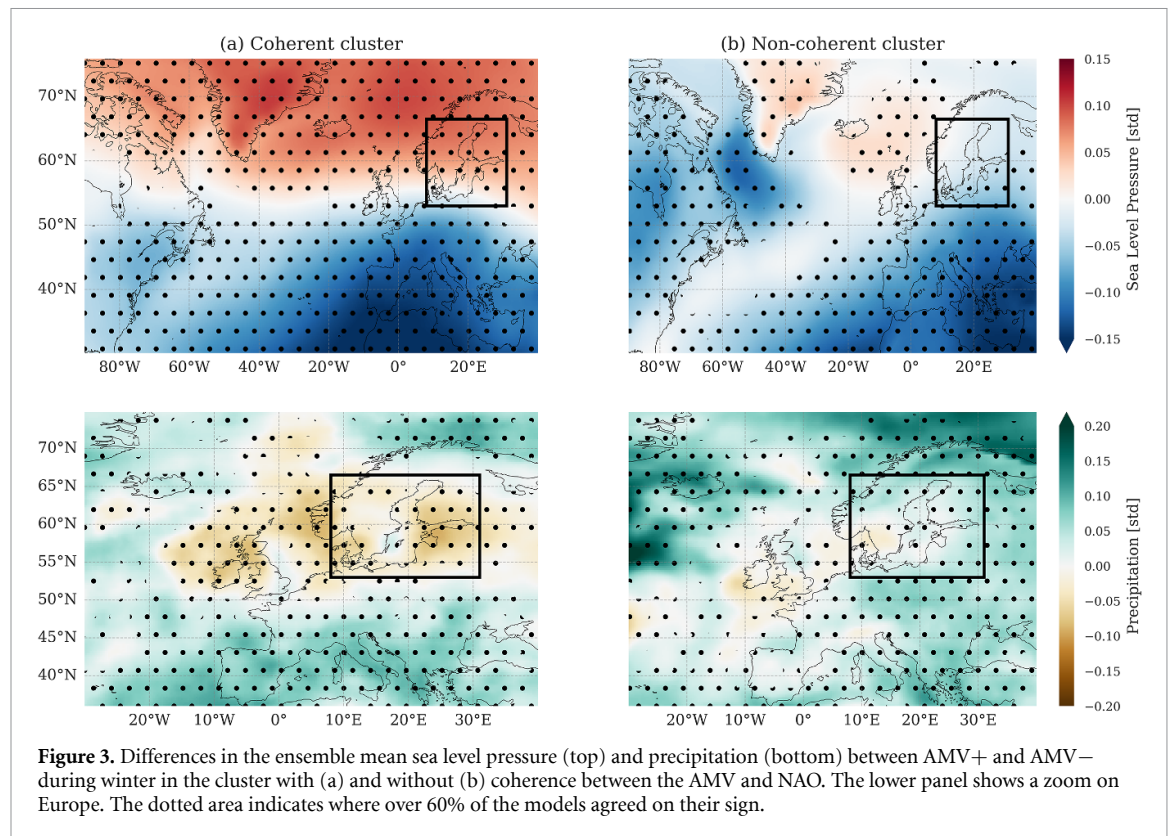


(BCC). The AMV and the NAO are characterized by significant coherence at low frequencies (periods > 10 years) throughout the model simulation. Still, no frequency band shows persistent coherence across the entire simulation. The dashed gray contours in figure 2 indicate areas where the AMV and the NAO share a high common power. For the BCC model, these areas coincide with areas where the AMV and the NAO have a significant coherence, suggesting a physical relationship. These areas are also displayed in figure 2(b), which shows the percentage of area sharing considerable power in the presence of significant coherence between the AMV and the NAO.

This analysis was repeated for every model and resulted in two different clusters. The first ($n = 18$) is characterized by a significant coherence between the AMV and the NAO, and the second cluster ($n = 15$) is not. In the determination of the former, only those models with a significant coherence between the NAO and the AMV over large parts of the model simulation at periods greater than ten years are included. A further requirement is that the percentage of area sharing significant power in the presence of significant coherence must be >50%. If these conditions are not met, the models are added to the cluster of non-coherent models. Figure S1 shows which of the two clusters each model belongs to. In the following, the coherent cluster, is referenced as (CC), and the non-coherent cluster is referred to as (NCC).

The differences between AMV+ and AMV− phases for SLP and precipitation are calculated for both clusters (figure 3). Models characterized by a significant coherence between the AMV and the NAO (figure 3(a)) have a strong meridional sea level pressure gradient resembling an NAO− pattern. The precipitation response (figure 3(a)) shows strong similarities with the ensemble mean precipitation pattern associated with AMV variability (figure 1(b)). A comparison of the CC (figure 3(a)) with the mean response of the ensemble (see figure 1) shows a near doubling of the amplitude for both SLP and precipitation. Further, the ensemble standard deviation of the CC is consistently smaller than the ensemble mean standard deviation, with the strongest difference at the subpolar gyre (not shown). This indicates a more robust response in the CC compared to the ensemble mean.

In contrast, the SLP response of the NCC (figure 3(b)) is not consistent with an NAO− pattern. Further, compared to the CC (figure 3(a)), the response of the NCC is relatively weak (figure 3(b)) and shows an increase in the ensemble standard deviation compared to the ensemble mean standard deviation (not shown). The corresponding precipitation pattern for the NCC shows an increase in precipitation over northern Europe, parts of Great Britain and small parts of western Europe. In contrast, there is a decrease in precipitation in the CC across northern Europe and large parts of Great Britain. The different



responses between the two clusters also affect southern Europe (south of 45° N) as the CC shows a much stronger increase in precipitation.

Lastly, we found that intra-model consensus is high, as for most areas the sign of the inherent models agrees. Comparing the inter-model consensus of the clusters to the mean ensemble inter-model consensus, we find that clustering increases the robustness of the results.

3.2. Relationship between AMV and AMOC

The two clusters in figure 3 show different atmospheric mean states associated with AMV variability. In the following,

we regress SST anomalies and air-sea heat flux anomalies onto the AMV index to identify the underlying processes (figure 4; surface). We consider lags -12 to -2 years, -2 – 0 years, and 0 – 12 years before and after an AMV+ event. For clarity, note that the following passages describe an AMV+ event, but the regression analysis also applies for an AMV– phase, but with opposite sign. The discussed heat fluxes consist of sensible and latent heat fluxes.

The left column shows the mean response of the SST (surface contours), heat fluxes (surface shading), and AMOC (front) for the CC. For lags -12 to -2 years, we find a positive SST response throughout the North Atlantic, being the strongest in the subpolar region (figure 4(a); left column). The corresponding air-sea heat fluxes depict the subsequent effect on ocean-atmosphere coupling (surface shading). Substantial oceanic heat loss is found north of

50° N in the western Atlantic and the Labrador Sea. Both spatial patterns—regressed SSTs and air-sea heat fluxes—look similar. The persistent ocean’s heat loss to the atmosphere is only possible if anomalous ocean heat fluxes convergence sustains the warm temperatures, thus, more poleward heat transport (Wills *et al* 2019). This is confirmed as the regression of the AMV onto the AMOC also shows a strengthening of the AMOC across the analyzed latitude range, reinforcing the mean AMOC circulation, with a maximum of about 0.9 Sv at 45° N (figure 4(a); left column).

Comparing lags -12 to -2 years and -2 – 0 years of the mean response (figures 4(a) and (b); left column) reveals similar spatial patterns. However, we find a stronger AMOC response for lags -12 to -2 years during the growth phase of an AMV+ event, which indicates the strengthening of the AMOC before an AMV+ event. In accordance, the heat fluxes and the SST pattern show a stronger response at lags -12 to -2 . After an AMV+ event (figure 4(c); left column), the imprint of the AMV on the AMOC fades. Consequently, the associated SST and heat flux pattern also disappear. The results of the mean response suggest that air-sea interactions over multi-decadal time scales are driven by an oceanic response as proposed by e.g. Gulev *et al* (2013).

The right column of figure 4 shows the difference between the mean response of the CC and the NCC. Again, lags at -12 to -2 years and -2 – 0 years (figures 4(a) and (b); right column) show a similar spatial pattern for the SST and heat fluxes. North of 50° N, the response of the NCC is stronger, resulting

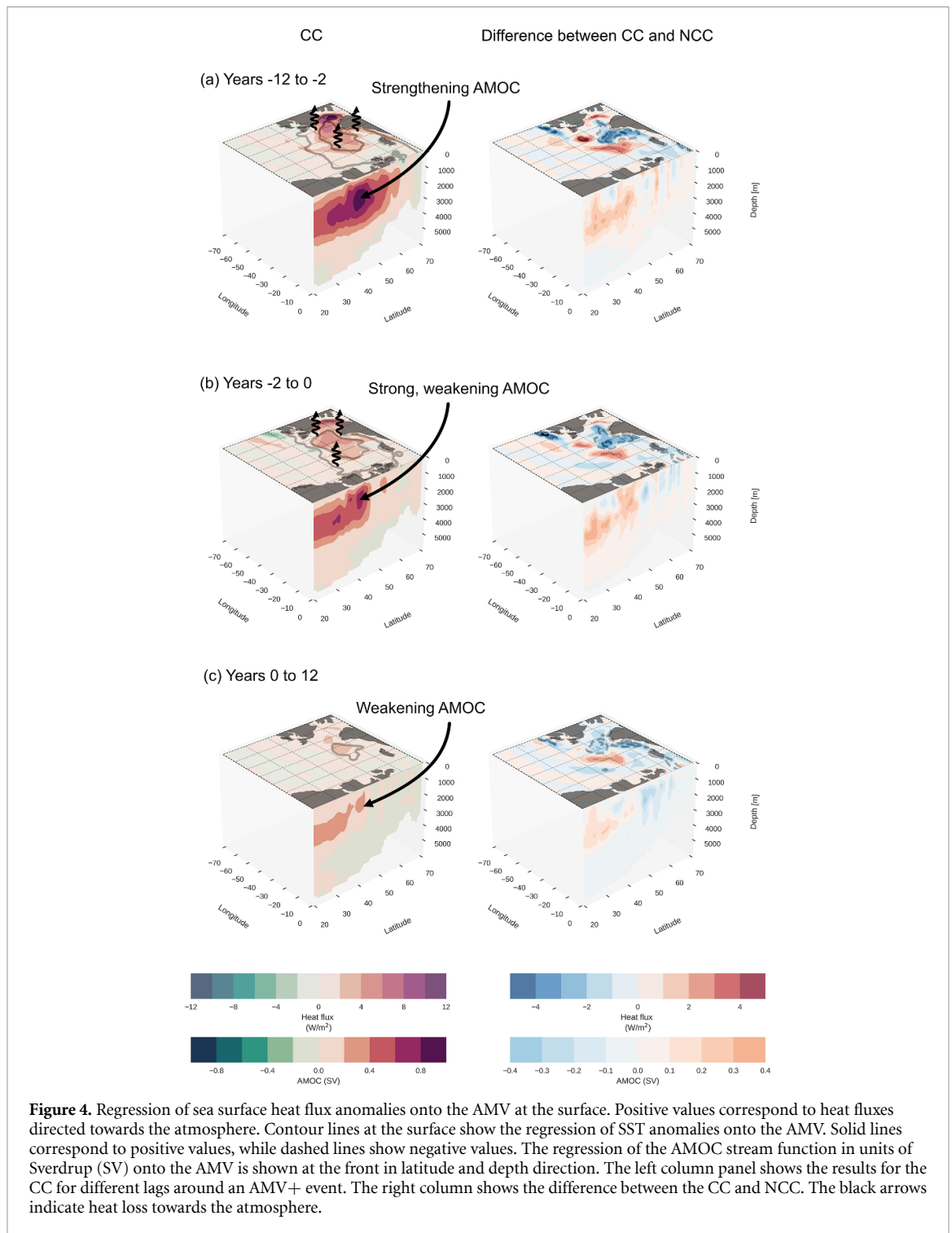


Figure 4. Regression of sea surface heat flux anomalies onto the AMV at the surface. Positive values correspond to heat fluxes directed towards the atmosphere. Contour lines at the surface show the regression of SST anomalies onto the AMV. Solid lines correspond to positive values, while dashed lines show negative values. The regression of the AMOC stream function in units of Sverdrup (SV) onto the AMV is shown at the front in latitude and depth. The left column shows the results for the CC for different lags around an AMV+ event. The right column shows the difference between the CC and NCC. The black arrows indicate heat loss towards the atmosphere.

in larger heat fluxes towards the atmosphere than in the CC. In contrast, south of 50° N, the SST and heat flux response is stronger in the CC.

The differences in the AMOC response at lags -12 to -2 years and -2 – 0 years show a stronger and deeper response across the latitude range up to 45° N for the CC. North of 45° N in the subpolar region, the AMOC response is of opposite sign, which corresponds to a stronger response of the AMOC in the NCC. This indicates weaker northward heat transport and also weaker transfer to the atmosphere north of

45° N in the CC, confirmed by the observed differences in air-sea heat fluxes. This suggests a stronger role of the AMOC related to AMV variability for the CC in the subpolar region. Lastly, it should be noted that by comparing the AMOC response at lags -12 to -2 years and -2 – 0 years (figures 4(a) and (b); right column), we find that the difference north of 45° N increases at lags -2 – 0 years. This is likely related to a complex two-dimensional response as indicated by the differences in heat fluxes at lags -12 to -2 years that the AMOC definition cannot capture.

Compared to prior lags, for lags 0–12 years (figure 4(c); right column) we find a weaker SST and heat flux response which is likely related to the declining impact of the AMV as shown in the left column of figure 4(c). The declining impact of the AMV is also found in the difference of the AMOC response, as the magnitude decreases compared to prior lags. However, the spatial pattern still looks similar, with a stronger response south of 45° N and a weaker response north of 45° N.

For most regions of the Atlantic north of 50° N, the stronger imprint of the AMV on heat fluxes could indicate a higher amplitude in oceanic vertical mixing and thus upward heat transport in the NCC. In the Labrador Sea, the differences between the two clusters are likely related to differences in the deep oceanic convection during winter and the subsequent impact on sea ice cover, suppressing heat transfer to the atmosphere.

4. Discussion

This study showed that the multidecadal Atlantic SST variability in unforced coupled climate models projects onto a reduced meridional SLP gradient, thus resembling a negative NAO pattern. In that regard, the behavior of the CMIP6 models resembled that of the CMIP5 models (Wills *et al* 2019). However, we found that CMIP models can be divided into two clusters: one characterized by coherence, i.e. a strong linkage between the AMV and the NAO, as well as a negative NAO response, and a second that lacked significant coherence between the AMV and the NAO and in which the SLP response did not match an NAO— pattern, instead of forming a weak negative SLP anomaly located in the subpolar gyre. These findings suggested that the physical mechanisms important for decadal or longer time scales differed between the two clusters.

Our results showed that the two clusters have a different SST response associated with AMV variability. This leads to different air-sea heat fluxes that need to be maintained by an anomalous vertical ocean heat transport north of 50° N and is finally reflected in an anomalous AMOC. While the NCC shows a weaker coupling between AMOC and AMV variability up to 40° N, it has a stronger response in the subpolar region. A stronger AMV-driven AMOC response in the subpolar region (NCC) leads to stronger air-sea heat fluxes resulting in a negative SLP anomaly located further north. It does not project strongly onto the NAO. Weaker heat fluxes in the subpolar region and a stronger AMOC response in lower latitudes, as found for the CC, are more likely to project onto a negative NAO pattern. Hence, models in the CC show a significant coherence between the AMV and the NAO, as found in the wavelet analysis (figure 2).

Wills *et al* (2019) argue that a plausible reason for different warming patterns between models is that they differ in their representation of the shape of the subpolar gyre and the geographic location of deep-water formation, which our findings support. However, it should be noted that the AMOC and the AMV varied considerably among the different models, which is attributable to model biases (Yan *et al* 2018). The CMIP6 models show a relatively large spread in their AMOC strength, ranging from 8.2 Sv to 42.1 Sv, with a mean of 19.4 Sv. Further, different lengths of model simulations are also likely to influence our results, as the IPSL-CM6A-LR model features a prominent and quasi-regular AMOC oscillation with a 200 year period (Jiang *et al* 2021). Such low-frequency oscillations cannot be analyzed when only 500 years of data are available.

Nevertheless, our results stress the importance of the oceanic circulation and its interaction with the atmosphere for North European precipitation. Our analysis links different SLP and precipitation responses to slow oceanic processes. In that sense, our results complement the work of Ruggieri *et al* (2021) since they use a constant prescribed, idealized AMV forcing, focusing on shifts in the jet stream that cause different atmospheric responses as well. Following our results, future work could focus on the jet stream's position in response to different ocean AMV responses.

Lastly, in examining the impact of the unforced AMV in the CMIP6 models, our study showed that most models had significant low-frequency power. However, no distinct frequency was dominant across all models (figures S2 and S3). The wavelet analysis of the AMV revealed time-localized periods with significant power rather than a persistent AMV periodicity. Moreover, the autocorrelation analysis showed that, on average, the unforced component of the AMV had lower predictability when compared to the ERSST dataset. This finding was in good agreement with several studies attributing a significant role to external forcing in the observed variability of the AMV (Bellomo *et al* 2018). Recently, Mann *et al* (2021) proposed that the observed multi-decadal frequency of the AMV (50–70 years) could be attributed solely to volcanic forcing. Their analysis of CMIP5 piControl runs revealed no distinct dominant frequency across models, suggesting no apparent predictability of the AMV in piControl runs. In contrast to Mann *et al* (2021), we argue that the unforced AMV signal can be disentangled by clustering the models. The AMV-related response in precipitation and sea level pressure of the models can be predicted by analyzing the relationship between the AMV, AMOC, and the NAO.

The importance of the difference between the two clusters was demonstrated by analyzing the mean

precipitation pattern for Europe that is associated with AMV variability. There was a decrease in precipitation over Northern Europe for the CC, which is in accordance with a lower heat and moisture transport as indicated by the lower air-sea coupling, less northward extension of AMOC, and the dominating NAO— conditions. In contrast, there was an increase in precipitation over this area in the NCC. This leaves us with the conclusion that the precipitation sign for Northern Europe is dependent on the model's cluster.

During late winter, precipitation observations showed more precipitation over Portugal and less precipitation over Great Britain on multidecadal time scales during AMV+ phases (Simpson *et al* 2019). As the observed decrease in precipitation is found only in the CC but not in the NCC, this might indicate that the CC is more likely to represent a realistic multidecadal precipitation response. In addition, the CC can reproduce the linkage between AMOC, AMV, and NAO as proposed by Wills *et al* (2019); Delworth and Zeng (2016); Delworth *et al* (2017). In addition, the response of the CC is more robust as indicated by the smaller standard deviation compared to the ensemble mean. However, as we focus on unforced climate variability, it is also possible that important mechanisms driven by external forcing are neglected.

In summary, our results demonstrate the potential bias of any regional downscaling approach, depending on whether the deployed GCM is a member of the coherent or NCC. Consequently, the results of studies analyzing multidecadal variability, such as Börgel *et al* (2018), in which multi-decadal precipitation variability associated with AMV variability was shown to lead to multidecadal salinity changes of 10% of the mean salinity of the Baltic Sea, will be model-dependent.

In conclusion, our study revealed a characteristic ocean-atmosphere interaction preceding an AMV warm event. This is characterized by an anomalous ocean heat transfer to the atmosphere that can only be maintained by an anomalous AMOC response (figure 4; left column). Our results revealed two clusters of models within CMIP6 piControl runs that were different in their atmospheric response with respect to AMV variability. These clusters showed different AMOC responses as well, which likely drive the different atmospheric responses. These differences have important implications for regional climate modeling, whose findings are often the basis for decision-making in coastal management. If the low-frequency response of the underlying general circulation models fails to reproduce the interaction between atmosphere and ocean, future projections will also be biased. Hence, for coastal seas that suffer from eutrophication, oxygen depletion, and other environmental threats, management decisions should be based on not one but several models, thus supporting the use of regional ensembles.

Data availability statement

The data that support the findings of this study are available upon reasonable request from the authors.

Acknowledgments

The research presented in this study is part of the Baltic Earth program (Earth System Science for the Baltic Sea region, see www.baltic.earth).

We would like to thank Dr. Leo Borchert and an anonymous reviewer whose constructive comments greatly improved earlier drafts of the manuscript.

Most of the analysis in this article was conducted using xarray (Hoyer and Hamman 2017). Hence, we would like to thank the xarray community for their amazing work. Further, we'd like to thank all institutes that have contributed to CMIP6 and make their data publicly available.

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