

Limited impact of Greenland meltwater on abruptness and reversibility of future Atlantic overturning changes

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Abstract

All climate models project that the Atlantic Meridional Overturning Circulation (AMOC) will weaken in the 21st century, but most models neglect increasing runoff from the Greenland ice sheet. Greenland meltwater is expected to exacerbate AMOC weakening, and omitting it increases the uncertainty in assessing the possibility of a future abrupt collapse or tipping of the AMOC. Here, we test the abruptness and reversibility of AMOC changes under strong future global warming in a state-of-the-art climate model with and without physically plausible Greenland meltwater forcing. While Greenland meltwater significantly exacerbates future AMOC weakening, modeled long-term AMOC changes are neither abrupt nor irreversible. While accounting for Greenland meltwater will increase the accuracy of climate projections, our results suggest that the importance of Greenland meltwater for assessing the risk of future AMOC tipping may be smaller than previously thought.

1 Introduction

The Atlantic Meridional Overturning Circulation (AMOC) has a prominent role in shaping the Earth’s mean climate and climate change [1–4]. For instance, recent studies have demonstrated the influence of an AMOC weakening on the position of the inter-tropical convergence zone [5], European temperature extremes [6–8], Arctic amplification [9], Pacific trade winds [10], and global monsoon systems [11].

Because of these near-global impacts, the possibility that global warming could induce an abrupt collapse or *tipping* of the AMOC, defined here as “a critical threshold beyond which a system reorganizes, often abruptly and/or irreversibly” [12], has been debated for several decades [13–15]. Early warning indicators suggest that the AMOC is currently undergoing destabilization compatible with approaching a tipping point [16–18], although recent work has also identified new mechanisms that are expected to stabilize the AMOC [19, 20]. In addition, the ability of the AMOC to weaken abruptly and irreversibly with respect to (idealized) fresh-water forcing has now been demonstrated across the hierarchy of ocean and climate models [18, 21–23].

Nevertheless, while there is consensus among climate models that the AMOC will weaken until 2100 [24], the IPCC Sixth Assessment Report (AR6) found that none of the CMIP6 models show an abrupt AMOC collapse during the 21st century [25]. In contrast to earlier assessments, however, only *medium confidence* was assigned to this finding, implying that a possible collapse of the AMOC is still under debate [25]. Besides model biases that could influence AMOC stability [26, 27], one main reason why current-generation models may not be able to simulate an AMOC collapse is that they neglect meltwater influx from the Greenland Ice Sheet [25], which is expected to accelerate with global warming [28].

Previous studies have shown that including realistic Greenland ice sheet meltwater in future emission scenarios can exacerbate future AMOC weakening [29–36], but the impact of Greenland meltwater on abruptness and irreversibility has not yet been investigated, even though meltwater is often hypothesized as a mechanism that would induce or accelerate a future AMOC collapse [25, 34, 37, 38]. Here, we assess the long-term AMOC response until 2300 under strong greenhouse gas emissions and Greenland meltwater forcing in model experiments carried out with current-generation (CMIP6) climate model EC-Earth3 [39]. This model features

one of the higher grid resolutions in CMIP6 and is one of the few CMIP6 models which simulates both a realistic present-day AMOC strength and a negative value of freshwater import into the South Atlantic (F_{ovS} ; Methods), which is consistent with observations and implies that the model is not expected to be biased towards a too stable AMOC [27]. Despite a significant impact of Greenland meltwater on AMOC weakening especially after 2100, we do not find an abrupt or irreversible AMOC collapse, suggesting that the omission of Greenland mass loss in climate models might impact the assessment of a potential future AMOC tipping less than previously thought.

2 Results

2.1 Meltwater-induced AMOC weakening

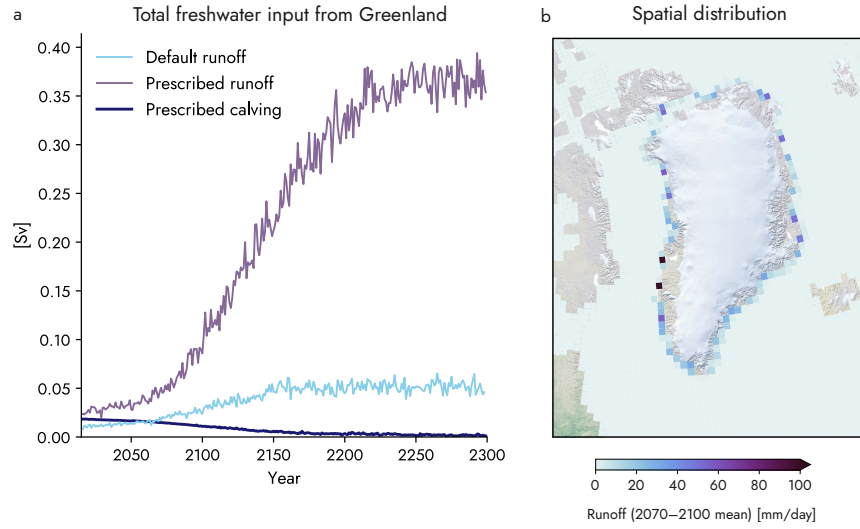


Fig. 1 Greenland meltwater forcing. (a) Annual mean runoff and calving used as input for the “Meltwater” simulations and runoff in the “Reference” simulations for comparison, (b) Example spatial distribution of runoff (averaged over 2070–2100) with non-uniform input at coastal grid points.

To reliably isolate and quantify the meltwater signal, we perform two four-member initial condition ensembles with and without the Greenland meltwater forcing (“Meltwater” and

“Reference”, respectively). We prescribe Greenland runoff and calving from a state-of-the-art (CMIP6) coupled climate–ice sheet model simulation [40] under a high-end greenhouse gas emission scenario (SSP5-8.5) until 2300 (Methods). Meltwater derived from the coupled climate–ice sheet model reaches 0.09 Sv by 2100 and more than 0.3 Sv by 2300 (Fig. 1a), larger values than those considered in previous studies with parametrized meltwater (see Methods). In contrast to most previous studies, the meltwater is also routed in a physically consistent way, non-uniformly in space and time, to coastal grid points around Greenland (Fig. 1b).

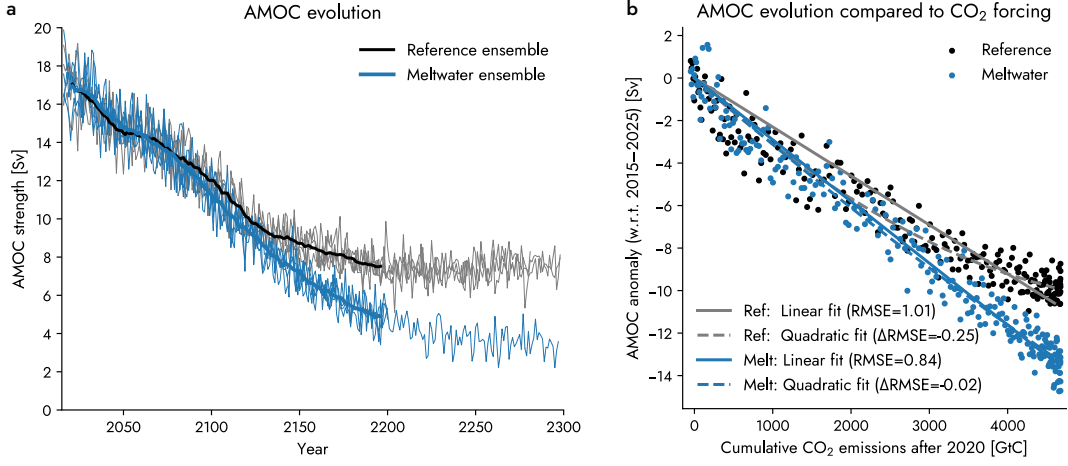


Fig. 2 AMOC projections until 2300 with and without Greenland meltwater. (a) Annual mean AMOC time series for individual ensemble members (thin lines) and 10-year running mean of ensemble means until 2200 (thick lines; as the full 4×2 member ensemble covers the period 2016–2200), (b) AMOC anomalies in one pair of ensemble members against cumulative CO₂ emissions since 2015.

We find that Greenland meltwater consistently exacerbates the weakening of the AMOC at 26.5°N in response to global warming (Fig. 2a). However, until 2100, the size of the effect is projected to be relatively small. For the reference ensemble without Greenland meltwater, the AMOC weakening under SSP5-8.5 is approximately linear during the 21st century and ranges between 3.8 and 7.3 Sv per century for different ensemble members, compared to 4.5 to 8.4 Sv per century in the meltwater ensemble. This enhanced weakening of, on average, 0.9 Sv per century is consistent for all pairs of ensemble members (Supplementary Fig. S1a)

and statistically significant ($p = 0.02$ using a paired Student’s t -test), although on the same order as the standard deviation of trends due to low-frequency variability (1.5 Sv per century in the reference ensemble).

We quantify when this significantly stronger AMOC weakening trend emerges in the meltwater ensemble compared to the reference ensemble using bootstrap resampling of the interannual to decadal AMOC variability (Methods). For the linear AMOC trend, a statistically significant difference emerges in 2092 (2086 to 2098, 66% confidence interval [CI]; Fig. 2b). Similarly, an emergence in 2100 (2091 to 2112, 66% CI) is found when the 15-year running mean of the AMOC index is used instead of the AMOC trend. This is at most a few decades after first differences between forcing scenarios emerge, given that the AMOC in CMIP6 models weakens almost independently of the scenario “until about 2060” [25].

Beyond the 21st century, the meltwater-induced AMOC weakening becomes stronger, in line with the strong increase in Greenland melting after 2100 (Fig. 1) due to the sustained high CO₂ emissions in the SSP5-8.5 scenario. By the end of the 22nd century (2180-2200), an additional AMOC weakening of 2.5 ± 0.2 Sv is attributable to Greenland meltwater. We extend one member per ensemble until the end of the 23rd century, by which the AMOC stabilizes at 7.5 Sv in the reference simulation but weakens further to about 3.5 Sv with added Greenland meltwater. In summary, Greenland meltwater induces an additional AMOC weakening on the order of 10–20% at the end of the 21st century and up to 40% at the end of the 23rd century, compared to the AMOC weakening induced by CO₂ and consequent atmosphere–ocean feedbacks which are accounted for in current climate models.

Even after experiencing strong meltwater input, a basin-wide Atlantic overturning cell remains in all simulations, although it is shallower and weaker in the meltwater simulation (Fig. S2). This differs from entirely collapsed or reversed Atlantic overturning states previously identified in some climate models [41–43]. Nevertheless, the AMOC in the meltwater simulations is on the threshold of being classified as “collapsed” using typical threshold definitions (e.g., 80% weakening compared to the pre-industrial AMOC strength [44]). AMOC characteristics such as northward heat transport due to overturning into the South Atlantic (Supplementary Fig. S3a) or shared outcropping isopycnals between the Northern hemisphere and the Southern Ocean [45] (Supplementary Fig. S3b,c) remain even at the end of the 23rd century, but they are strongly reduced with meltwater input. It should also be noted that

the Atlantic ocean heat transport due to overturning is indeed close to zero at some tropical latitudes in the meltwater simulation (not shown).

More importantly, we do not find signs of an *abrupt* AMOC change following the classical definition of a nonlinear response exceeding the rate of external forcing [46, 47]. As shown in Fig. 2b, the AMOC strength in the meltwater ensemble scales remarkably linearly with cumulative CO₂ emissions until emissions reach zero in 2250. This scaling is sub-linear in the reference ensemble due to the AMOC stabilization in the 23rd century, in agreement with previous modeling studies which did not account for GrIS melt [48]. To summarize, while it is ambiguous whether the AMOC can be classified as “collapsed”, the weakening is not abrupt.

2.2 Shift of source regions shapes response to meltwater

The significant additional AMOC weakening after around 2100 calls for a better understanding of associated mechanisms. To this end, we decompose the Atlantic overturning at 45°N and its meltwater-induced weakening into the contributions of different source regions north of 45°N (defined in Fig. 3d) based on the density-space overturning at the northern and southern gateways bounding these regions (Methods).

Both ensembles agree that the main source regions of the present-day AMOC in EC-Earth3, the Irminger and Nordic Seas, will decrease in importance through the 21st and 22nd centuries. In turn, the AMOC is characterized by a strong northward shift in source regions in response to global warming. The Arctic Ocean, whose present-day direct contribution to the AMOC at 45°N is negligible, becomes the most important AMOC source region by 2100 (Fig. 3a–b). This increased role of the Arctic Ocean is partly because the overturning strength across the Arctic gateways intensifies, but also because the overturning peak in density space moves more rapidly to lighter densities in the Arctic Ocean than at subpolar latitudes (Fig. S4), potentially due to Arctic Ocean amplification [49]. Consequently, the density of maximum overturning at the Arctic gateways coincides with that at 45°N σ_{MOC} during much of the 22nd century, whereas it is located at the larger densities of the overflow waters during the present-day, contributing to the overturning further south only indirectly (e.g., via mixing at the Greenland–Scotland Ridge).

In line with the northward shift of AMOC source regions, the Arctic Ocean contributes most to the meltwater-induced additional AMOC weakening in the 22nd century (Fig. 3c). This

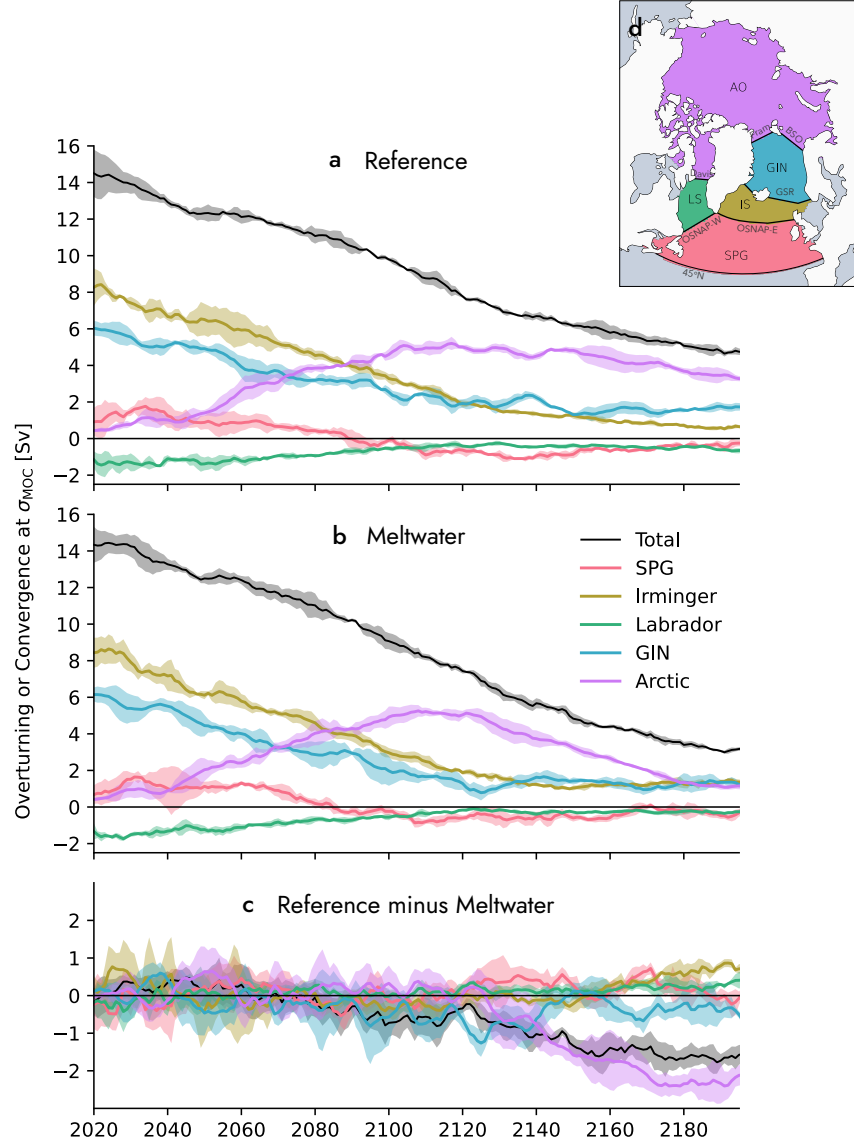


Fig. 3 Change in AMOC source regions. Maximum overturning at 45°N (black line) and convergence at σ_{MOC} by region: (a) Reference ensemble, (b) Meltwater ensemble, (c) Difference between the two ensembles. All lines are for 2015–2200 and are smoothed with a ten-year running mean; shadings indicate the ensemble standard deviation. The inset (d) shows the definition of the source regions and the gateways bounding them.

meltwater-driven weakening can be understood in terms of a volume and buoyancy budget for the Arctic Ocean (Fig. 4a-c; Methods). The overturning across the Arctic gateways weakens mostly due to a decrease in surface-forced water mass transformation (SFWMT), while the magnitude of interior mixing also decreases with meltwater input. Decreased SFWMT is linked to a decrease in both surface density (Fig. 4d) and surface buoyancy flux. While the former is a direct consequence of the meltwater input freshening the surface layers, the latter (as well as the smaller mixing component) is linked to a decrease in mixed layer depth (Fig. 4e). This is again a consequence of the lower surface density, which increases stratification. To summarize, Greenland meltwater influences the northward-shifted AMOC source regions both directly (via decreased surface density) and indirectly (via increased stratification that leads to less mixing and less heat loss).

To gauge the robustness of meltwater impacts, we calculate the Arctic Ocean contribution to the overturning at σ_{MOC} in eight other CMIP6 models which provide SSP5-8.5 simulations until 2300 (Fig. S5). In all models, the Arctic contribution to Atlantic overturning increases during the 21st century, in agreement with EC-Earth3. However, the timing of the maximum contribution is model-dependent. In most models, the Arctic contribution decreases to below 2 Sv by the mid-22nd century, while EC-Earth3 is the only model from this ensemble with a sustained Arctic contribution until 2300. This implies that EC-Earth3 probably provides an upper-end estimate of the meltwater-induced AMOC weakening via the Arctic Ocean after 2150, although other regions might play a significant role in modulating meltwater-induced AMOC weakening in other models.

2.3 Reversibility of meltwater-induced AMOC changes

To test the impact of the added Greenland meltwater on the reversibility of AMOC changes, we branch off idealized reversibility experiments in which the meltwater and/or greenhouse gas forcings are reversed. We conduct two sets of experiments starting from the year 2250, in which emissions reach zero in the extended SSP5-8.5 scenario. In the first set of experiments, the meltwater forcing is switched off in 2250 and the CO₂ concentration is kept constant after 2300. This enables assessing AMOC reversibility with respect to meltwater under sustained strong global warming. In the second set of experiments, CO₂ concentrations are ramped down at 1% per year for 170 years until they reach 2015 levels and held constant thereafter. This

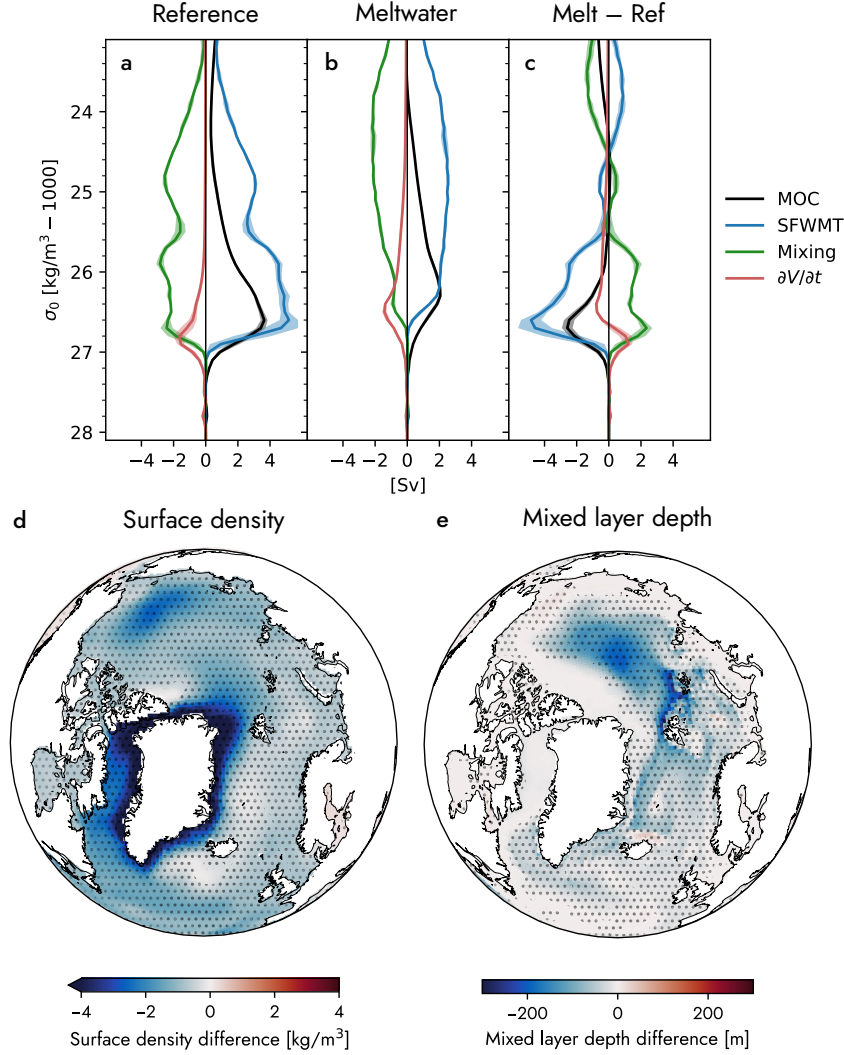


Fig. 4 Effect of Greenland meltwater on an ice-free Arctic Ocean. a–c: Volume and buoyancy budget (Methods; [50]) for the Arctic Ocean at the end of the 22nd century for (a) the reference simulation, (b) with Greenland meltwater, (c) difference between b and a. The overturning across the Arctic gateways (MOC) is decomposed into surface-forced water mass transformation (SFWMT), volume changes ($\partial V/\partial t$) and mixing, which is calculated as a residual. d–e: Ensemble mean differences (meltwater minus reference) at the end of the 22nd century: (d) annual mean surface density, (e) winter (March) mixed layer depth. Areas with significant differences ($p < 0.05$) using a t -test are stippled.

ramp-down experiment is initialized from both the meltwater and reference experiments to isolate differences in the reversibility behavior due to the previous meltwater forcing. Because no CO₂ ramp-down is available with the ice sheet model, we also switch off the additional meltwater forcing after 2250 for these CO₂ reversal experiments.

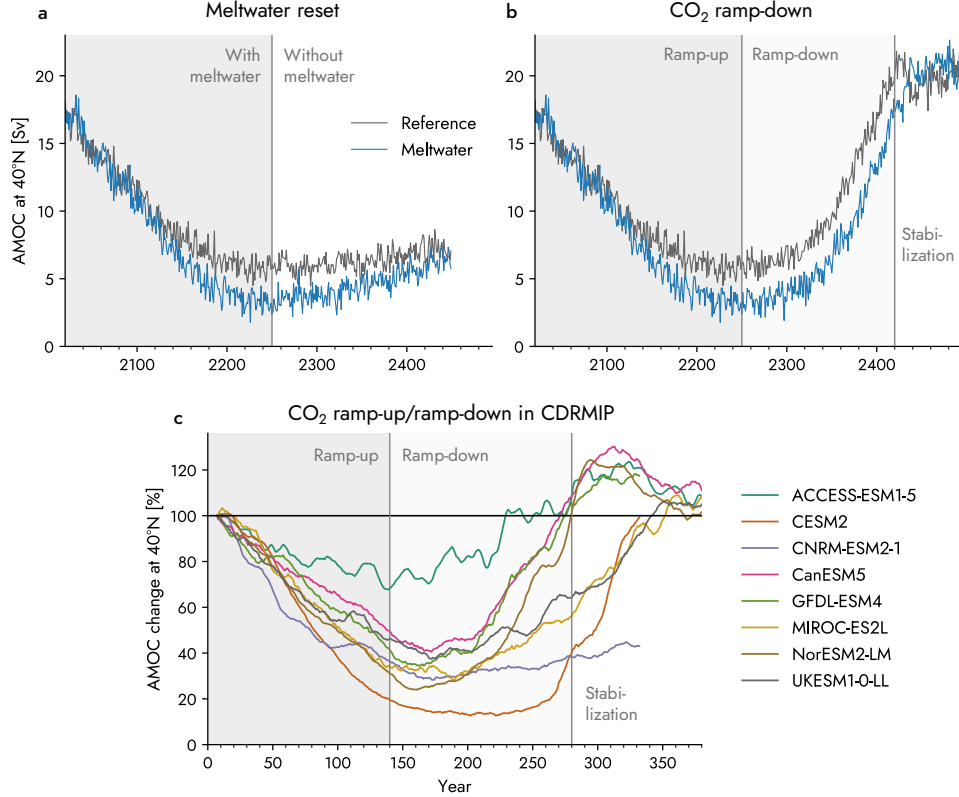


Fig. 5 AMOC reversibility in EC-Earth3 and CMIP6 models. Annual mean AMOC maximum at 40°N for (a) the “meltwater reset” experiments, (b) the “CO₂ ramp-down” experiments (see the text for experiment setups). The dark grey-shaded time spans in both subplots are identical to the standard “Reference” and “Meltwater” simulations, with the meltwater forcing stopped at 2250. (c) AMOC change in different CMIP6 models in the CDRMIP reversibility experiments [51]. Starting from pre-industrial conditions, these experiment consist of a CO₂ ramp-up at 1% per year for 140 years, followed by a ramp-down at 1% per year for another 140 years and a stabilization period at pre-industrial values for at least 50 years. AMOC changes are indicated as a percentage of the initial (year 0–15) AMOC strength and smoothed using a 15-year running mean.

Resetting the meltwater forcing under late-23rd century conditions leads to a gradual recovery of the AMOC (Fig. 5a). We continue the simulation for 200 years, after which the AMOC anomaly at 40°N has decreased from -2.7 Sv to -0.7 Sv compared to the reference simulation, which is extended using the same constant CO₂ concentrations after 2300. The larger recovery rate in the meltwater simulation suggests that both simulations would eventually converge to the same equilibrium. Therefore, while the timescale of recovery is (multi-)centennial, it appears that Greenland meltwater forcing, despite its large magnitude, does not induce an irreversible shift to a different equilibrium state of the AMOC under very high CO₂ concentrations.

Similarly, the CO₂ ramp-down experiments show reversibility of both the greenhouse gas- and meltwater-induced AMOC changes (Fig. 5b). During the ramp-down, the AMOC is about 2–3 Sv stronger without prior meltwater forcing, with the main regional driver of differences shifting from the Arctic to the Nordic Seas later during the ramp-down (Supplementary Fig. S6). This corresponds to a difference in recovery time to the same AMOC strength of, on average, 25 years. Both simulations eventually approach a similar stabilized AMOC strength of around 20 Sv after CO₂ concentrations have reached 2015 levels. This is stronger than the initial AMOC strength in 2015, in agreement with previous CO₂ ramp-down experiments without meltwater [52]. In EC-Earth3, this could be linked to a permanently winter ice-free Labrador Sea, whereas the Labrador Sea is intermittently ice-covered in the pre-industrial [53] and historical [39] simulations. This points toward a potential irreversibility of winter sea ice changes [54], which will be investigated elsewhere.

The reversibility to a ramp-down of CO₂ concentrations is not unique to EC-Earth3, but can be found across the ensemble of CMIP6 models in the *1pctCO2-cdr* simulations of the CDRMIP project [51]. These simulations followed a similar CO₂ ramp-down protocol at 1% per year from four times pre-industrial (Methods), albeit without explicitly considering Greenland meltwater. In the CDRMIP simulations, the AMOC fully recovers in 7 out of 8 CMIP6 models [55] (Fig. 5c). Our simulations suggest that this reversibility might not be sensitive to prior Greenland meltwater input, even when its magnitude is very large.

3 Discussion

In this study, we used an ensemble of future projections with a state-of-the-art climate model to probe the response of the future AMOC to high-end but physically plausible Greenland meltwater input. In contrast to a non-significant effect during the historical period [56], we found a small but significant additional AMOC weakening associated with Greenland meltwater of about 1 Sv until 2100 (about 10% of the CO₂-induced weakening) and up to 4 Sv until 2300 (nearly 40% of the CO₂-induced weakening) under very strong forcing. These results are in good agreement with previous studies using parametrized Greenland meltwater input [33, 34]. However, we found that the combined CO₂- and meltwater-induced AMOC weakening in our model was neither abrupt (with respect to the external forcing) nor irreversible, characteristics often associated with climate tipping points [12, 15]. Instead, with Greenland meltwater, the AMOC weakening scaled remarkably linearly with cumulative CO₂ emissions. This linearity, if corroborated by other climate models, might also yield additional insights into the physics of future AMOC weakening, such as why the AMOC strength diverges relatively late in different emission scenarios [24].

As climate projections beyond the 21st century are becoming more widely used, the increasing importance of Greenland meltwater for AMOC weakening after 2100 emphasizes the need to incorporate realistic meltwater estimates into these projections. This would also enable probing the robustness of the role of the Arctic Ocean, which played a key role in driving long-term AMOC changes under strong global warming, as well as shaping the AMOC response to Greenland meltwater, in EC-Earth3. This role of the Arctic Ocean is consistent with previous work demonstrating a northward shift of AMOC source regions [57] and areas of deep convection [58, 59] as the winter sea ice edge retreats northward under global warming, although only a relatively small number of CMIP6 models (including variants of EC-Earth) shows sustained deep convection in the Arctic Ocean at the end of the 21st century [60]. According to our source region analysis, models robustly project a significant Arctic contribution to the AMOC under strong CO₂ forcing, but it does not persist until 2300 in most models. This means that EC-Earth3 probably provides an upper bound on role of the Arctic Ocean to meltwater-induced future AMOC weakening.

The absence of abruptness and irreversibility even under strong melt rates of more than 0.3 Sv may seem at odds with the robustness of a freshwater-induced AMOC collapse across the model hierarchy [22, 43, 61]. There are several possible explanations. First, the results could be specific to EC-Earth3, for example due to model biases. However, the realistic AMOC strength and negative F_{ovS} as in observations increase our confidence that EC-Earth3 is not biased too stable. We also demonstrated that, as in EC-Earth3, the AMOC is reversible under idealized CO₂ reversal in most CMIP6 models. Second, even if the AMOC is in a bistable regime under pre-industrial conditions, it is possible that the AMOC shifts outside the bistable regime at very high CO₂ concentrations such as those prescribed by the high-end SSP5-8.5 scenario. This possibility calls for a better (conceptual) understanding of the interplay of CO₂ and freshwater forcing on the AMOC [14, 62], especially under transient forcing.

Finally, it has previously been shown that the same amount of North Atlantic freshwater input has a reduced effect on the AMOC under global warming due to changes in ocean stratification and in the North Atlantic gyre structure [63]. We will explore this aspect in more detail in a follow-up study. A corollary is that strong meltwater forcing under a moderate CO₂ increase might provide a higher risk for AMOC tipping. However, Greenland melt rates also generally scale with atmospheric temperatures and therefore CO₂ concentrations [33]. This creates a tug-of-war between an increasing amount of meltwater discharge and the weakening ocean sensitivity to meltwater input under global warming, although this relationship could break down on (multi-)millennial time scales if the Greenland ice sheet itself crosses a tipping point [64, 65].

In addition to comparing the response in different climate models, future studies should assess whether the overturning changes shown here are robust in high-resolution ocean models which resolve mesoscale eddies (e.g., ref. [66]). Nevertheless, ref. [67] demonstrated that the magnitude of AMOC weakening due to Greenland meltwater does not depend strongly on the ocean resolution despite different meltwater propagation pathways, increasing our confidence in estimates of meltwater-induced AMOC weakening from CMIP6-class models. Finally, since the uncertainties regarding the impact of Antarctic meltwater on the AMOC are considerable [66, 68, 69], simulations assessing the concurrent impacts of Greenland and Antarctic meltwater on the AMOC in long-term scenarios are needed in the future.

4 Methods

4.1 Model and AMOC characteristics

We perform model experiments with a state-of-the-art coupled climate model, EC-Earth3 [39], which participated in CMIP6 [70]. EC-Earth3 consists of the atmospheric model IFS cy36r4, the land-surface scheme H-TESSEL [71], the ocean model NEMO 3.6 [72], the sea ice model LIM3 [73], and the OASIS3-MCT coupler [74]. EC-Earth3 is run at its standard resolution for CMIP6, i.e., a horizontal resolution of about 80 km (TL255) for the atmosphere and 1° (about 100 km) for the ocean, with a grid refinement to $1/3^\circ$ in the tropical ocean. In the vertical, 91 levels are used for the atmosphere and 75 levels for the ocean, where layer depths range from 1 m near the surface to 200 m in the deep ocean.

EC-Earth3 successfully reproduces the most important basin-scale AMOC metrics at the beginning of the scenario runs (2016–2022). At 26°N , the AMOC strength is $17.1 \pm 0.9 \text{ Sv}$, in agreement with the observational value of $16.9 \pm 1.2 \text{ Sv}$ from the RAPID array [75]. In the sub-polar North Atlantic, the simulated overturning across the OSNAP-East line is $15.2 \pm 0.8 \text{ Sv}$, again in good agreement with observations ($16.3 \pm 0.6 \text{ Sv}$ [76]). While EC-Earth3 does not capture the overturning across the OSNAP-West line, this contribution is small (around 3 Sv) in observations, such that the total subpolar overturning of $12.5 \pm 0.9 \text{ Sv}$ in EC-Earth3 is only slightly weaker than observed ($16.7 \pm 0.6 \text{ Sv}$ [76]). An important metric for AMOC stability, the freshwater transport due to overturning at 34°S (F_{ovS}), is $-0.014 \pm 0.010 \text{ Sv}$ in EC-Earth3. This is slightly larger than in observations ($-0.16 \pm 0.09 \text{ Sv}$ [77]) but has the correct sign (negative as observed) often associated with the AMOC being in a bistable regime [14, 78, 79]. Hence, EC-Earth3 is one the very few CMIP6 models which both simulate a realistic AMOC strength and the correct sign of F_{ovS} [27].

4.2 Experiment setup

We conduct two ensembles (“reference” and “meltwater”) of future projections under the SSP5-8.5 scenario [80], the strongest global warming scenario considered by the IPCC Sixth Assessment Report with unabated greenhouse gas emissions beyond the 21st century [81]. This scenario can be regarded as a “worst-case, no-policy” scenario suitable to study extreme climate outcomes [82]. In SSP5-8.5, CO_2 concentrations reach more than 1000 ppm (about

4x pre-industrial) by the end of the 21st century and more than 2000 ppm (about 8x pre-industrial) by 2200, stabilizing at similarly high levels afterwards [81]. In EC-Earth3, this leads to a GMST increase of 5.5 K, 10.1 K and 10.9 K compared to pre-industrial at the end of the 21st, 22nd and 23rd centuries, respectively.

The “reference” experiments use the standard CMIP6 version of EC-Earth3, except for fixing the surface albedo at the location of present-day ice sheets to 0.8 (the default ice sheet albedo in EC-Earth). This change was implemented because IFS “parametrizes” ice sheets as a 10-meter snow pack and exposes low-albedo bedrock after it melts, inducing an unrealistically strong temperature feedback. In comparison, a fixed ice sheet albedo is more in line with the moderate albedo changes even under prolonged strong CO₂ forcing [83] and enables studying the effect of Greenland meltwater alone, leaving the opportunity for future model experiments to study the effect of changes in the ice cover separately. The “meltwater” experiments use the same EC-Earth3 configuration as the “reference” runs except for prescribing runoff and calving from the Greenland ice sheet, which is described in the following section.

As is common practice in climate projections [84], we use an ensemble of simulations initialized from different initial conditions to separate forced changes from internal variability. The initial conditions are sourced from a subset of the 50-member SMHI Large Ensemble with EC-Earth3 [85]. Because low-frequency variability – and therefore ensemble spread – in EC-Earth3 is dominated by the AMOC [39], we select four ensemble members that span the range of simulated present-day AMOC strength and projected 21st-century AMOC weakening under SSP5-8.5 (Fig. S7a). This includes the two members with the strongest and least pronounced weakening as well as two members with average AMOC characteristics, yielding an overall representative sample (Fig. S7b). Both our 4-member ensembles (“reference” and “meltwater”) are initialized from the same set of initial conditions at the start of 2016. Note that, due to the slight modification to the model, our reference simulations sample a different realization of internal variability compared to the SMHI Large Ensemble members, so that the rate of weakening is not expected to be identical. However, the strong correlation between the initial AMOC state and 21st century AMOC weakening (Fig. S7a) means that we still expect to sample a wide range of rates of AMOC weakening.

4.3 Meltwater forcing

In the “Meltwater” ensemble, we prescribe runoff and calving from a fully coupled climate–ice sheet model, the Community Earth System Model version 2 (CESM2) coupled to the Community Ice Sheet Model version 2 (CISM2) for the Greenland ice sheet [83]. CESM2 [86] has a nominal resolution of 1° for atmosphere and ocean and CISM2 [87] was run at 4 km resolution. Ice-sheet runoff is routed following topography gradients to the nearest ocean gridpoint, while calving is spread diffusively over a radius of up to 300 km from the coast to mimic icebergs [40]. Climate projections under the SSP5-8.5 scenario until 2100 were presented in ref. [40].

We prescribe CESM fields for ocean surface fluxes from runoff (monthly) and calving (annually) from 2016 until 2300, including the diffusive spreading of calving. We set grid cells with negative river runoff values, which are artifacts from water conservation in CESM’s land model, to zero, such that the total runoff into the ocean matches the runoff from the ice-sheet model well. These fluxes were remapped conservatively onto the native NEMO grid before being passed to EC-Earth3 following the implementation of ref. [88]. At coupling time, we set internally calculated runoff and calving over Greenland to zero to avoid double-counting. As in the standard EC-Earth3 configuration, runoff is inserted into the ocean at sea surface temperature, zero salinity, and through a depth of up to 150 m to prevent physical and numerical issues from injecting large amounts of runoff into the topmost ocean cell [72]. In the meltwater ensemble, the runoff depth mask is updated around Greenland to match the runoff climatology by the end of the 23rd century. For calving, we account for the latent heat flux of melting.

The area-integrated time series of runoff over Greenland is shown in Fig. 1a. During the historical period, both the fully coupled ice sheet model in CESM2–CISM2 (19.4 mSv, all values averaged 1981–2010) and the simple mass balance approach in EC-Earth (8.6 mSv) produce a realistic magnitude of runoff compared to observations (13.2 mSv; [89]). In the future projections, GrIS runoff strongly increases in CESM2–CISM2, reaching 0.09 Sv by the end of the 21st century and more than 0.3 Sv by the end of the 23rd century (Fig. 1a). In the EC-Earth reference simulations, Greenland runoff levels off at about 0.07 Sv after 2150, a similar value as in many other CMIP6 models (not shown). The meltwater forcing prescribed in the model intercomparison of Bakker et al. [34] was also about 0.07 Sv in 2300 [90], such that no significant

effect on the AMOC would be expected with their meltwater parametrization. Calving in CISM2 decreases during the 21st century and beyond ([40], Fig. 1a), but the uncertainty of this projection can be considered large because CISM2 does not account for ocean forcing at marine-terminating Greenland margins (cf. [91]). In any case, the contribution of calving in SSP5-8.5 is expected to be small compared to the strongly increasing runoff.

We note that there are several limitations to our approach. First, we prescribe monthly fields from a single model, such that the meltwater trajectory inherits the boundary conditions from the CESM2 simulation. A recent study comparing long-term projections from a standalone GrIS model forced with boundary conditions from different GCMs showed that CESM forcing produced the largest GrIS mass loss [92], implying that the meltwater forcing derived from CESM2-CISM2 is likely at the upper end of the CMIP6 range. Second, our implementation does not account for the differences in surface mass balance between CESM2 and EC-Earth3, such that the sea level rise in the EC-Earth3 “meltwater” runs is different from the integrated CISM2 mass loss. However, we do not analyze sea level metrics here, making this an acceptable trade-off over ad-hoc corrections. We preferred the forcing approach from a fully coupled model over parametrizations [e.g., 33, 34, 93] for several reasons: the extended SSP5-8.5 forcing extends far beyond their calibration range; parametrizations typically give regional averages while CESM2 runoff is resolved at 1° resolution; they do not eliminate the role of model biases, as the (regional or global) climate model forcing used for calibration can substantially affect the runoff sensitivity to warming [94]; and they also do not take into account differences in surface mass balance. As a final caveat, fixing the ice sheet geometry and albedo to present-day values allows to cleanly separate the effect of meltwater, but these changes are expected to further weaken the AMOC [e.g., 95, 96]. However, the effect is expected to be of second order compared to the effects of CO_2 and meltwater.

4.4 Overturning and source region diagnostics

First, we analyze the AMOC in depth-space at the conventional latitude of 26.5°N by taking the depth-maximum of the annual mean overturning streamfunction [1]. At subpolar latitudes, density (σ) coordinates are preferred over depth (z) coordinates because isopycnals tend to be strongly sloped in east-west direction [97]. Following ref. [98], we therefore compute the overturning north of 40°N in density coordinates referenced to the surface (σ_0). The streamfunction

is then defined as [50]

$$\Psi(\sigma) = - \int_{x_1}^{x_2} \int_{\sigma_{\text{bot}}}^{\sigma} v(x, \sigma', t) d\sigma' dx, \quad (1)$$

where v is the northward velocity, x_1 and x_2 are the western and eastern boundaries of the basin, and σ_{bot} is the bottom (maximum) density. Using this definition, the overturning is zero at the bottom and equal to the net volume transport for $\sigma \rightarrow 0$. Following previous modeling studies [99, 100], we do not apply any compensation term between neighboring straits (except for the comparison against observed values above). We compute $\Psi(\sigma)$ at 45°N, two sections east and west of Greenland that approximately follow the OSNAP array [97], the Greenland-Scotland Ridge (GSR), Davis Strait, Fram Strait and the Barents Sea Opening (Fig. 3d). Streamfunctions are computed from monthly data using an adapted version of the “line method” from the StraitFlux package [101].

Following ref. [100], we define the convergence of overturning M in a region as the difference between $\Psi(\sigma)$ at its southern and northern boundaries. This way, the overturning across 45°N is decomposed into the sum of convergences of the five regions north of 45°N (Fig. 3d; SPG: subpolar gyre, IS: Irminger Sea, LS: Labrador Sea, GIN: Nordic Seas, AO: Arctic Ocean and Baffin Bay):

$$\Psi_{45N}(\sigma) = M_{SPG}(\sigma) + M_{IS}(\sigma) + M_{LS}(\sigma) + M_{GIN}(\sigma) + M_{AO}(\sigma). \quad (2)$$

Here, we assume that the overturning across Bering Strait is zero and the contributions from other marginal seas (Baltic Sea, Hudson Bay etc.) are negligible, which is confirmed by the very good agreement between the left-hand and right-hand side of Eq. 2 (not shown).

Similarly to ref. [50], in our analysis we focus on the (time-dependent) density σ_{MOC} , which is defined as the density at which the overturning at 45°N is at its maximum. This can be interpreted as the density which bounds the AMOC lower limb. Evaluating Eq. 2 at σ_{MOC} therefore quantifies the net contribution of each region to the AMOC lower limb, regardless of which processes drive diapycnal transformations.

4.5 Volume and buoyancy budget

If a region is characterized by inflow of lighter and outflow of denser water masses, the overturning can be related to the transformation of water masses at the surface in the so-called

Walsh framework [102]. The volume and buoyancy budget in a region can be computed as [50]

$$F(\sigma) + G(\sigma) = \frac{\partial V(\sigma)}{\partial t} + \Psi_S(\sigma) - \Psi_N(\sigma), \quad (3)$$

where $F(\sigma)$ is the surface-forced water mass transformation defined below, $V(\sigma)$ is the ocean volume with a potential density larger than σ , $\Psi_S(\sigma)$ and $\Psi_N(\sigma)$ are the overturning across the southern and northern gateways of the region, respectively, calculated using Eq. 1. For the Arctic Ocean, $\Psi_N(\sigma) = 0$ because there is no overturning across Bering Strait. $G(\sigma)$ quantifies interior mixing which cannot be diagnosed directly from climate model output and is therefore calculated as a residual [50].

Surface-forced water mass transformation $F(\sigma)$ is given by

$$F(\sigma) = \frac{1}{\Delta\sigma} \iint_{\mathcal{A}} \left[-\frac{\alpha}{C_p} Q + \beta \frac{S}{1-S} \Phi_{FW} \right] \Pi(x, y; \sigma) \, dx \, dy, \quad (4)$$

where α and β are the thermal expansion and haline contraction coefficients, respectively, C_p is the specific heat capacity of seawater, Q is the air-sea heat flux, S is the sea surface salinity, Φ_{FW} is the surface freshwater flux and $\Pi(x, y; \sigma)$ selects the outcrop region of a density range $\sigma \pm \frac{\Delta\sigma}{2}$. It is defined as

$$\Pi(x, y; \sigma) = \begin{cases} 1 & \text{if } |\tilde{\sigma}(x, y) - \sigma| \leq \frac{\Delta\sigma}{2} \\ 0 & \text{elsewhere,} \end{cases} \quad (5)$$

where $\tilde{\sigma}(x, y)$ is the potential density at the location (x, y) . F is analogous to a streamfunction and computed from monthly fields for surface density, heat and freshwater fluxes (evaporation minus precipitation minus runoff minus sea-ice melt) and subsequently averaged into an annual mean climatology. A spacing of $\Delta\sigma = 0.1 \, \text{kg m}^{-3}$ is used for all budget terms in Eq. 3.

4.6 Time of emergence

For any given year, the difference in ensemble means between the reference and meltwater ensembles is compared using a one-sided, paired Student's t -test. Here, two simulations initialized from the same initial condition but subject to different meltwater trajectories are treated as paired observations due to the long memory in the AMOC time series arising from

low-frequency (centennial) variability [53, 103]. Since the AMOC weakening in all ensemble members is approximately linear in until at least 2120, the t -test is first applied to the linear AMOC trend (2016 to the specified year), and a sensitivity test is performed using 15-year running means instead of the trend. The year of emergence is then defined as the first year from which on the difference is always statically significant ($p < 0.05$).

For an uncertainty estimate of emergence times, we repeat this procedure on an ensemble of 1000 surrogate time series for each ensemble member. To construct the surrogates, the interannual-to-decadal variability (“residuals”) is obtained by subtracting a fifth-order polynomial fit from the original AMOC time series. Then, the residuals are resampled using the “random phasing” method of Ebisuzaki [104], which preserves the weak autocorrelation of the residuals better than conventional bootstrapping, and added back to the polynomial fit.

4.7 CMIP6 model data

To put the results from EC-Earth3 in context, we use output from the CMIP6 multi-model ensemble [70] for two experiments: extended SSP5-8.5 scenario simulations until 2300 from ScenarioMIP [81] and reversibility experiments (*1pctCO2* and *1pctCO2-cdr*) from the Carbon Dioxide Removal Model Intercomparison Project (CDRMIP) [51].

The SSP5-8.5 scenario runs use the same external forcing as the EC-Earth simulations (except for not prescribing meltwater) and we select all 8 models (cf. Fig. S5) which provide sufficient ocean output until 2300 to calculate the overturning in density space.

The CDRMIP experiments use an idealized CO₂ ramp-up/ramp-down protocol starting at pre-industrial conditions. First, CO₂ concentrations are increased by 1% per year (“ramp-up”) until they reach four times pre-industrial values (around 1140 ppm) after 140 years. Subsequently, CO₂ concentrations are decreased at 1% per year (“ramp-down”) for another 140 years until they return to pre-industrial values. The simulations are then extended for a minimum of 50 years under fixed pre-industrial conditions (“stabilization”). A total of 8 CMIP6 models listed in Fig. 5 provided AMOC output for this experiment.

Supplementary Figures

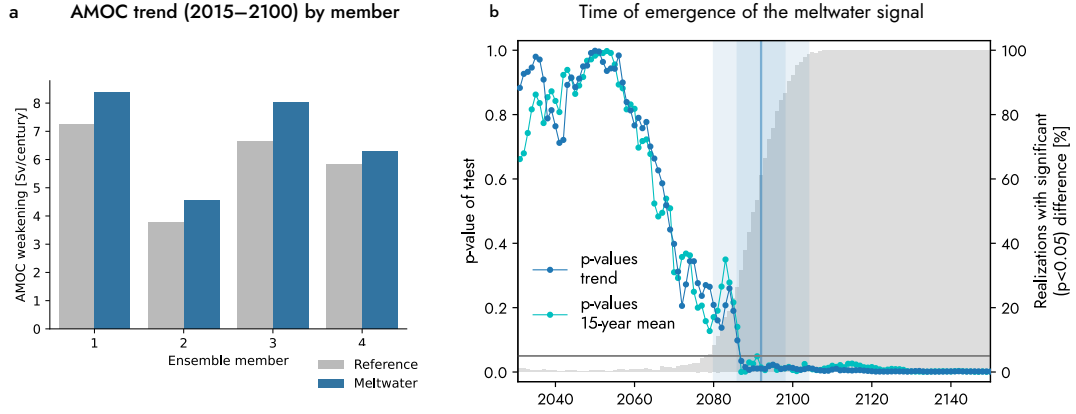


Fig. S1 21st-century AMOC weakening and emergence of the meltwater signal. (a) Linear trends of AMOC weakening (in Sv/century) during 2015–2100 for each ensemble member. Paired bars are initialized from the same initial conditions. (b) Left axis: p -values for the one-sided t -test whether trends (blue line) or 15-year running means (cyan line) differ significantly between the reference and meltwater ensembles. The grey line shows the 0.05 threshold. Right axis: Percentage of realizations using a bootstrap test (Methods) that show a significant difference in trends between the reference and meltwater ensembles. The blue line and shadings indicate the median, 66% and 95% confidence intervals of the time of emergence (Methods).

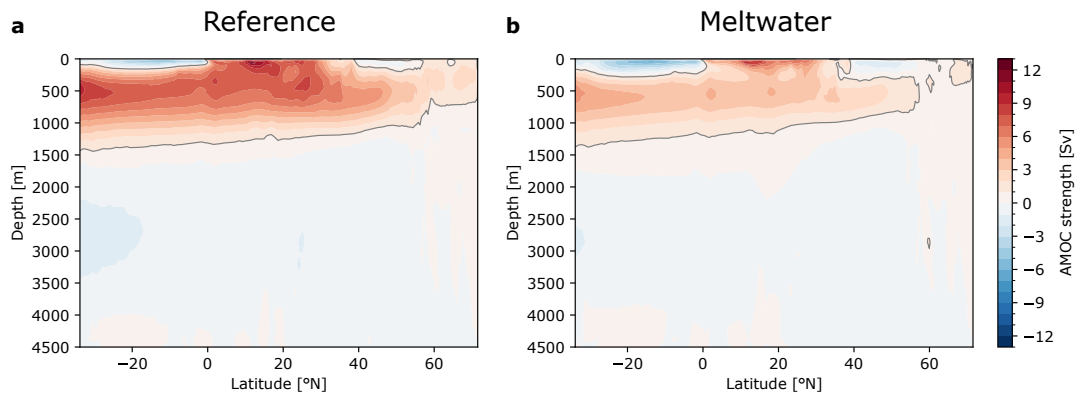


Fig. S2 Final AMOC streamfunctions. AMOC streamfunctions in depth space for the end of the simulation 2270–2300: (a) Reference (2 members), (b) Meltwater (1 member). The grey contour indicates the 1 Sv isoline.

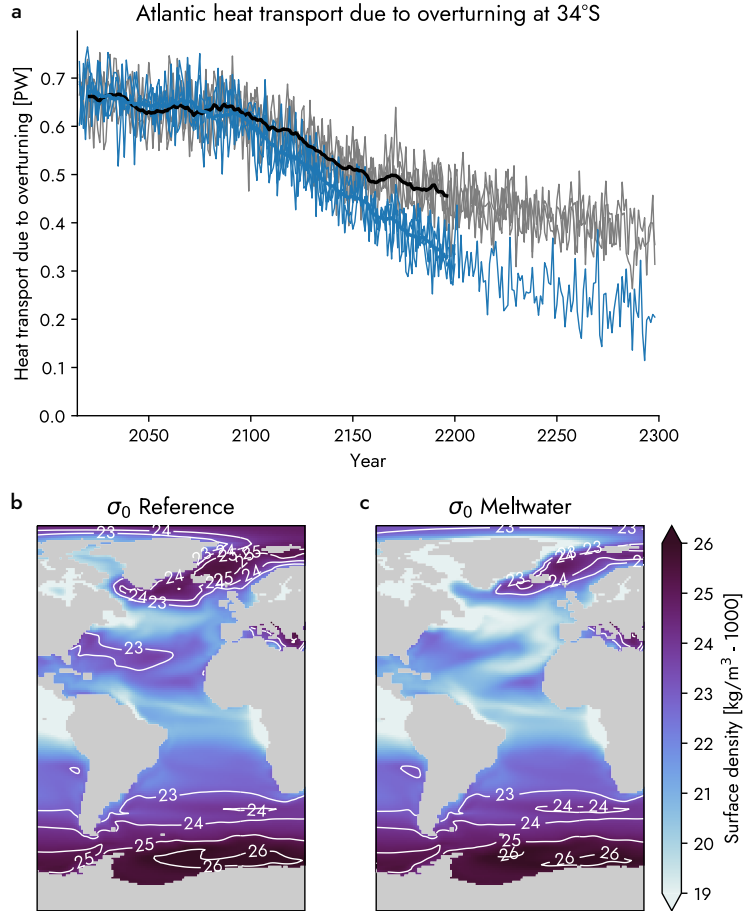


Fig. S3 Characteristics of a sustained AMOC. (a) Time series of Atlantic heat transport due to overturning (Methods) for the reference and meltwater ensembles, (b, c) Annual mean surface density (σ_0) at the end of the 23rd century for the reference and meltwater simulations. Shared isopycnals between the North Atlantic and Southern Ocean are shown as white contours.

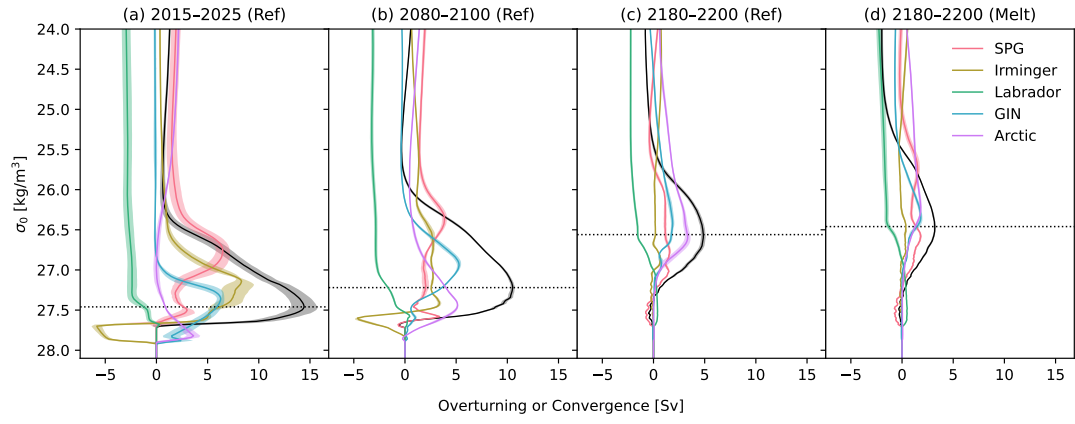


Fig. S4 Changing overturning convergence by region. Climatologies of σ -space overturning at 45°N (black lines) and convergence by region. Shadings indicate the ensemble standard deviation. The dotted horizontal line indicates σ_{MOC} , the density at which the overturning streamfunction at 45°N is at its maximum.

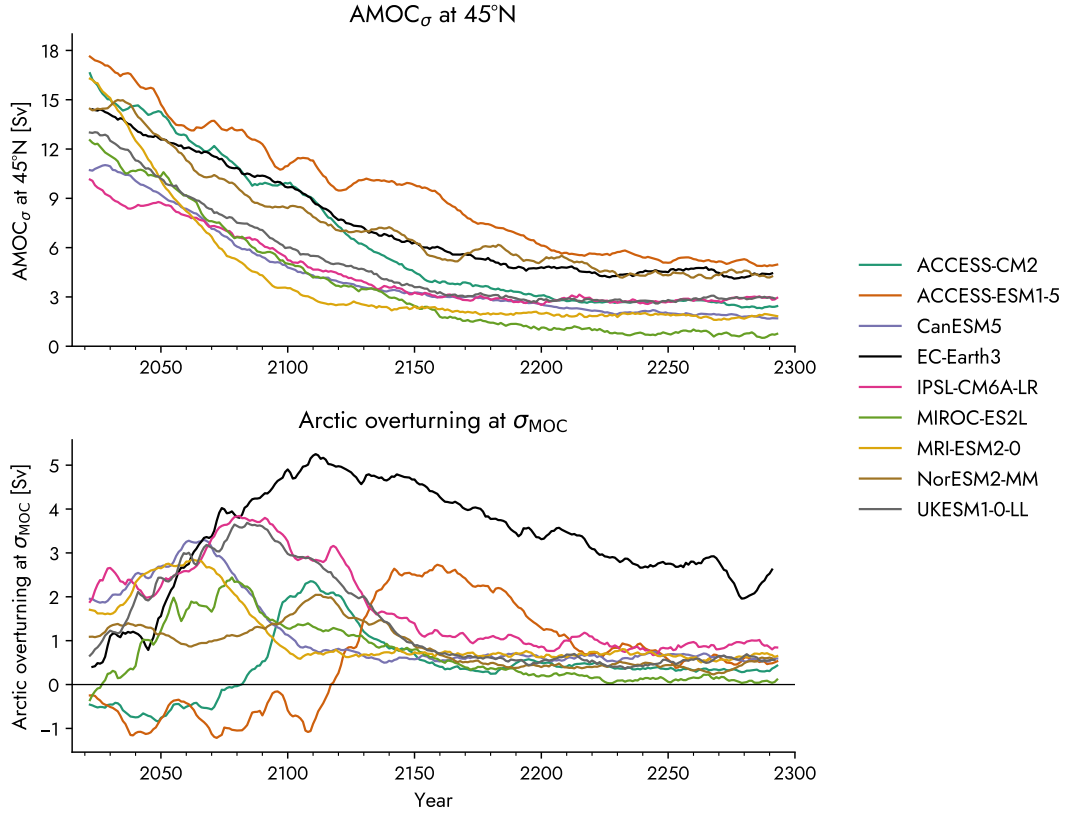


Fig. S5 Mid-latitude and Arctic overturning in CMIP6 models. (a) Maximum Atlantic overturning in density-space at 45°N, (b) overturning across the Arctic gateways at σ_{MOC} in the SSP5-8.5 scenario for all CMIP6 models providing suitable output until 2300 and the first ensemble member of the EC-Earth3 “Reference” ensemble. A 15-year running mean is applied to all time series for clarity.

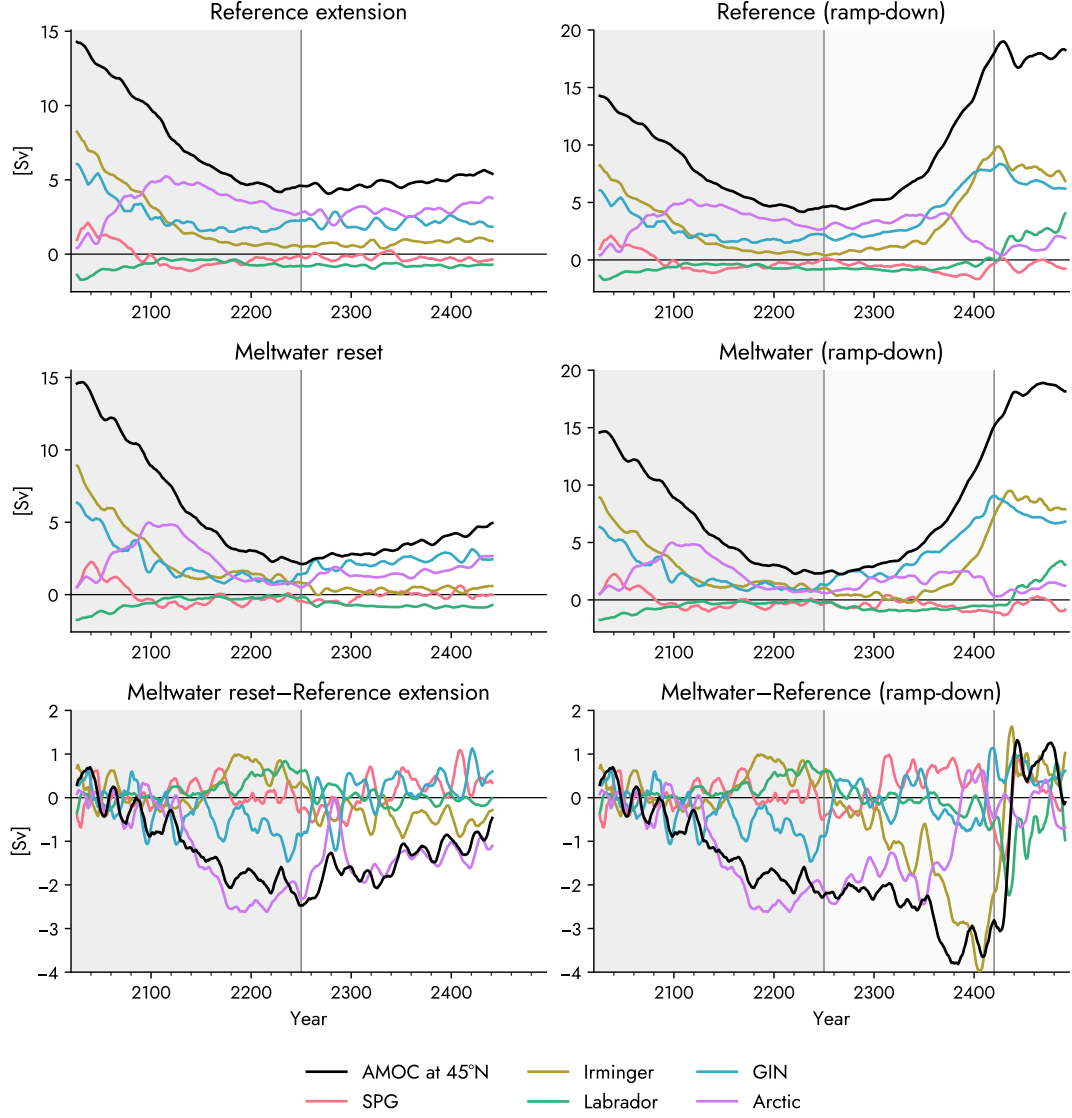


Fig. S6 Convergence at σ_{MOC} in EC-Earth3 reversibility experiments. Same as Fig. 3, but for the reversibility experiments shown in Fig. 5.

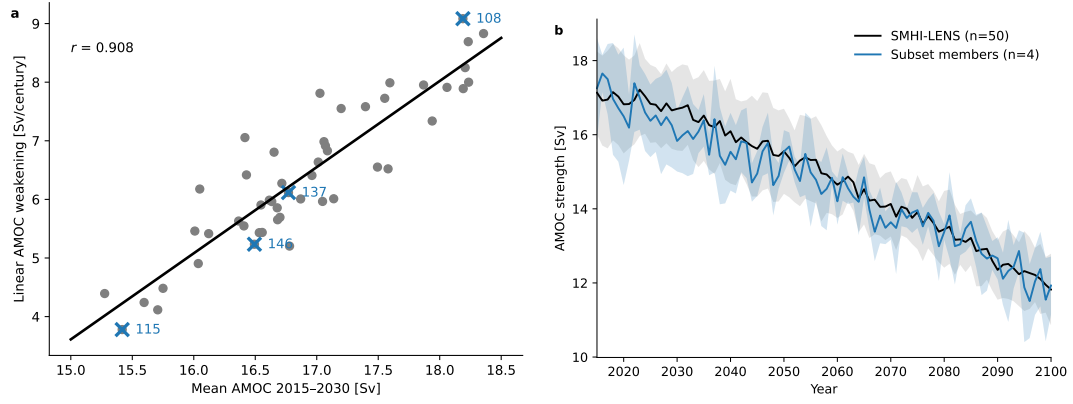


Fig. S7 Ensemble member selection strategy based on the SMHI Large Ensemble. (a) Relation between initial (2015–2030) mean AMOC strength at 26°N and linear AMOC trend during 2015–2100 in the 50-member EC-Earth3 large ensemble [85] for the SSP5-8.5 scenario. The selected ensemble members and their original variant labels (rXXXi1p1f1) are highlighted with blue crosses. (b) AMOC time series (ensemble mean plus and minus one standard deviation) under SSP5-8.5 forcing for the full 50-member ensemble (black) and the 4-member subset (blue). While the subset mean is slightly lower than the ensemble mean between the 2020s and 2040s, overall the subset captures the ensemble mean and spread well, especially in the second half of the 21st century.

References

- [1] Buckley, M. W. & Marshall, J. Observations, inferences, and mechanisms of the Atlantic Meridional Overturning Circulation: A review. *Rev. Geophys.* **54**, 5–63 (2016). <https://doi.org/10.1002/2015RG000493>.
- [2] Liu, W., Fedorov, A. V., Xie, S.-P. & Hu, S. Climate impacts of a weakened Atlantic Meridional Overturning Circulation in a warming climate. *Sci. Adv.* **6**, eaaz4876 (2020). <https://doi.org/10.1126/sciadv.aaz4876>.
- [3] Bellomo, K., Angeloni, M., Corti, S. & von Hardenberg, J. Future climate change shaped by inter-model differences in Atlantic meridional overturning circulation response. *Nat. Commun.* **12**, 3659 (2021). <https://doi.org/10.1038/s41467-021-24015-w>.
- [4] Hu, A. *et al.* Complexity of the Global Climate Dynamics: Interactions among ENSO, AMOC, and tropical basins. *Ocean-Land-Atmosphere Res.* **4**, 0096 (2025). <https://doi.org/10.34133/olar.0096>.
- [5] Cerato, G., Bellomo, K., D’Agostino, R. & von Hardenberg, J. Multi-model evidence of future tropical Atlantic precipitation change modulated by AMOC decline. *J. Clim.* **38**, 3093–3107 (2025). <https://doi.org/10.1175/JCLI-D-24-0333.1>.
- [6] Meccia, V. L., Simolo, C., Bellomo, K. & Corti, S. Extreme cold events in Europe under a reduced AMOC. *Environ. Res. Lett.* **19**, 014054 (2024). <https://doi.org/10.1088/1748-9326/ad14b0>.
- [7] Meccia, V. L., Simolo, C., Bellomo, K. & Corti, S. The impact of a weakened AMOC on European heatwaves. *Environ. Res. Lett.* **20**, 024005 (2025). <https://doi.org/10.1088/1748-9326/ada3e7>.
- [8] van Westen, R. M. & Baatsen, M. L. J. European Temperature Extremes Under Different AMOC Scenarios in the Community Earth System Model. *Geophys. Res. Lett.* **52**, e2025GL114611 (2025). <https://doi.org/10.1029/2025GL114611>.
- [9] Lee, Y.-C., Liu, W., Fedorov, A. V., Feldl, N. & Taylor, P. C. Impacts of Atlantic meridional overturning circulation weakening on Arctic amplification. *Proc. Natl. Acad. Sci.* **121**, e2402322121 (2024). <https://doi.org/10.1073/pnas.2402322121>.
- [10] Orihuela-Pinto, B., Santos, A., England, M. H. & Taschetto, A. S. Reduced ENSO Variability due to a Collapsed Atlantic Meridional Overturning Circulation. *J. Clim.* **35**, 5307–5320 (2022). <https://doi.org/10.1175/JCLI-D-21-0293.1>.
- [11] Ben-Yami, M. *et al.* Impacts of AMOC Collapse on Monsoon Rainfall: A Multi-Model Comparison. *Earth’s Future* **12**, e2023EF003959 (2024). <https://doi.org/10.1029/2023EF003959>.
- [12] Chen, D. *et al.* in *Framing, Context, and Methods* (eds Masson-Delmotte, V. *et al.*) *Climate Change 2021: The Physical Science Basis. Contribution of Working Group I to*

- the Sixth Assessment Report of the Intergovernmental Panel on Climate Change* 147–286 (Cambridge University Press, Cambridge, United Kingdom and New York, NY, USA, 2021).
- [13] Broecker, W. S., Peteet, D. M. & Rind, D. Does the ocean–atmosphere system have more than one stable mode of operation? *Nature* **315**, 21–26 (1985). <https://doi.org/10.1038/315021a0>.
 - [14] Weijer, W. *et al.* Stability of the Atlantic Meridional Overturning Circulation: A Review and Synthesis. *J. Geophys. Res. Oceans* **124**, 5336–5375 (2019). <https://doi.org/10.1029/2019JC015083>.
 - [15] Armstrong McKay, D. I. *et al.* Exceeding 1.5°C global warming could trigger multiple climate tipping points. *Science* **377**, eabn7950 (2022). <https://doi.org/10.1126/science.abn7950>.
 - [16] Boers, N. Observation-based early-warning signals for a collapse of the Atlantic Meridional Overturning Circulation. *Nat. Clim. Chang.* **11**, 680–688 (2021). <https://doi.org/10.1038/s41558-021-01097-4>.
 - [17] Ditlevsen, P. & Ditlevsen, S. Warning of a forthcoming collapse of the Atlantic meridional overturning circulation. *Nat. Commun.* **14**, 4254 (2023). <https://doi.org/10.1038/s41467-023-39810-w>.
 - [18] van Westen, R. M., Kliphuis, M. & Dijkstra, H. A. Physics-based early warning signal shows that AMOC is on tipping course. *Sci. Adv.* **10**, eadk1189 (2024). <https://doi.org/10.1126/sciadv.adk1189>.
 - [19] Baker, J. A. *et al.* Continued Atlantic overturning circulation even under climate extremes. *Nature* **638**, 987–994 (2025). <https://doi.org/10.1038/s41586-024-08544-0>.
 - [20] Bonan, D. B. *et al.* Observational constraints imply limited future Atlantic meridional overturning circulation weakening. *Nat. Geosci.* 1–9 (2025). <https://doi.org/10.1038/s41561-025-01709-0>.
 - [21] Stommel, H. Thermohaline Convection with Two Stable Regimes of Flow. *Tellus* **13**, 224–230 (1961). <https://doi.org/10.1111/j.2153-3490.1961.tb00079.x>.
 - [22] Rahmstorf, S. *et al.* Thermohaline circulation hysteresis: A model intercomparison. *Geophys. Res. Lett.* **32**, L23605 (2005). <https://doi.org/10.1029/2005GL023655>.
 - [23] van Westen, R. M., Kliphuis, M. & Dijkstra, H. A. Collapse of the Atlantic Meridional Overturning Circulation in a Strongly Eddying Ocean-Only Model. *Geophys. Res. Lett.* **52**, e2024GL114532 (2025). <https://doi.org/10.1029/2024GL114532>.
 - [24] Weijer, W., Cheng, W., Garuba, O. A., Hu, A. & Nadiga, B. T. CMIP6 Models Predict Significant 21st Century Decline of the Atlantic Meridional Overturning Circulation. *Geophys. Res. Lett.* **47**, e2019GL086075 (2020). <https://doi.org/10.1029/2019GL086075>.

- [25] Fox-Kemper, B. *et al.* in *Ocean, Cryosphere and Sea Level Change* (eds Masson-Delmotte, V. *et al.*) *Climate Change 2021: The Physical Science Basis. Contribution of Working Group I to the Sixth Assessment Report of the Intergovernmental Panel on Climate Change* 1211–1362 (Cambridge University Press, Cambridge, United Kingdom and New York, NY, USA, 2021).
- [26] Liu, W., Xie, S.-P., Liu, Z. & Zhu, J. Overlooked possibility of a collapsed Atlantic Meridional Overturning Circulation in warming climate. *Sci. Adv.* **3**, e1601666 (2017). <https://doi.org/10.1126/sciadv.1601666>.
- [27] van Westen, R. M. & Dijkstra, H. A. Persistent climate model biases in the Atlantic Ocean’s freshwater transport. *Ocean Sci.* **20**, 549–567 (2024). <https://doi.org/10.5194/os-20-549-2024>.
- [28] Trusel, L. D. *et al.* Nonlinear rise in Greenland runoff in response to post-industrial Arctic warming. *Nature* **564**, 104–108 (2018). <https://doi.org/10.1038/s41586-018-0752-4>.
- [29] Fichet, T. *et al.* Implications of changes in freshwater flux from the Greenland ice sheet for the climate of the 21st century. *Geophys. Res. Lett.* **30** (2003). <https://doi.org/10.1029/2003GL017826>.
- [30] Ridley, J. K., Huybrechts, P., Gregory, J. M. & Lowe, J. A. Elimination of the Greenland Ice Sheet in a High CO₂ Climate. *J. Clim.* **18**, 3409–3427 (2005). <https://doi.org/10.1175/JCLI3482.1>.
- [31] Mikolajewicz, U., Vizcaíno, M., Jungclaus, J. & Schurgers, G. Effect of ice sheet interactions in anthropogenic climate change simulations. *Geophys. Res. Lett.* **34** (2007). <https://doi.org/10.1029/2007GL031173>.
- [32] Gierz, P., Lohmann, G. & Wei, W. Response of Atlantic overturning to future warming in a coupled atmosphere-ocean-ice sheet model. *Geophys. Res. Lett.* **42**, 6811–6818 (2015). <https://doi.org/10.1002/2015GL065276>.
- [33] Lenaerts, J. T. M. *et al.* Representing Greenland ice sheet freshwater fluxes in climate models. *Geophys. Res. Lett.* **42**, 6373–6381 (2015). <https://doi.org/10.1002/2015GL064738>.
- [34] Bakker, P. *et al.* Fate of the Atlantic Meridional Overturning Circulation: Strong decline under continued warming and Greenland melting. *Geophys. Res. Lett.* **43**, 12,252–12,260 (2016). <https://doi.org/10.1002/2016GL070457>.
- [35] Golledge, N. R. *et al.* Global environmental consequences of twenty-first-century ice-sheet melt. *Nature* **566**, 65–72 (2019). <https://doi.org/10.1038/s41586-019-0889-9>.
- [36] Ackermann, L., Danek, C., Gierz, P. & Lohmann, G. AMOC Recovery in a Multicentennial Scenario Using a Coupled Atmosphere-Ocean-Ice Sheet Model. *Geophys. Res. Lett.* **47** (2020). <https://doi.org/10.1029/2019GL086810>.

- [37] Caesar, L., Rahmstorf, S., Robinson, A., Feulner, G. & Saba, V. Observed fingerprint of a weakening Atlantic Ocean overturning circulation. *Nature* **556**, 191–196 (2018). <https://doi.org/10.1038/s41586-018-0006-5>.
- [38] Drijfhout, S., Angevaere, J., Mecking, J. V., van Westen, R. & Rahmstorf, S. Shutdown of northern Atlantic overturning after 2100 following deep mixing collapse in CMIP6 projections. *Environ. Res. Lett.* (2025). <https://doi.org/10.1088/1748-9326/adfa3b>.
- [39] Döscher, R. *et al.* The EC-Earth3 Earth system model for the Coupled Model Intercomparison Project 6. *Geosci. Model Dev.* **15**, 2973–3020 (2022). <https://doi.org/10.5194/gmd-15-2973-2022>.
- [40] Muntjewerf, L. *et al.* Greenland Ice Sheet Contribution to 21st Century Sea Level Rise as Simulated by the Coupled CESM2.1-CISM2.1. *Geophys. Res. Lett.* **47**, e2019GL086836 (2020). <https://doi.org/10.1029/2019GL086836>.
- [41] Manabe, S. & Stouffer, R. J. Are two modes of thermohaline circulation stable? *Tellus Dyn. Meteorol. Oceanogr.* **51**, 400–411 (1999). <https://doi.org/10.3402/tellusa.v51i3.13461>.
- [42] Gregory, J. M., Saenko, O. A. & Weaver, A. J. The role of the Atlantic freshwater balance in the hysteresis of the meridional overturning circulation. *Climate Dynamics* **21**, 707–717 (2003). <https://doi.org/10.1007/s00382-003-0359-8>.
- [43] van Westen, R. M. & Dijkstra, H. A. Asymmetry of AMOC Hysteresis in a State-Of-The-Art Global Climate Model. *Geophys. Res. Lett.* **50**, e2023GL106088 (2023). <https://doi.org/10.1029/2023GL106088>.
- [44] Collins, M. *et al.* in *Extremes, Abrupt Changes and Managing Risk* (eds Pörtner, H.-O. *et al.*) *IPCC Special Report on the Ocean and Cryosphere in a Changing Climate* Ch. 6 (2019).
- [45] Nikurashin, M. & Vallis, G. A Theory of the Interhemispheric Meridional Overturning Circulation and Associated Stratification. *J. Phys. Oceanogr.* **42**, 1652–1667 (2012). <https://doi.org/10.1175/JPO-D-11-0189.1>.
- [46] Rahmstorf, S. in *Abrupt Climate Change* (eds Steele, J. H., Thorpe, S. A. & Turekian, K. K.) *Encyclopedia of Ocean Sciences* 1–6 (Academic Press, London, 2001).
- [47] Alley, R. B. *et al.* Abrupt Climate Change. *Science* **299**, 2005–2010 (2003). <https://doi.org/10.1126/science.1081056>.
- [48] Steinacher, M. & Joos, F. Transient Earth system responses to cumulative carbon dioxide emissions: Linearities, uncertainties, and probabilities in an observation-constrained model ensemble. *Biogeosciences* **13**, 1071–1103 (2016). <https://doi.org/10.5194/bg-13-1071-2016>.

- [49] Shu, Q. *et al.* Arctic Ocean Amplification in a warming climate in CMIP6 models. *Sci. Adv.* **8**, eabn9755 (2022). <https://doi.org/10.1126/sciadv.abn9755>.
- [50] Buckley, M. W., Lozier, M. S., Desbruyères, D. & Evans, D. G. Buoyancy forcing and the subpolar Atlantic meridional overturning circulation. *Phil. Trans. R. Soc. A* **381**, 20220181 (2023). <https://doi.org/10.1098/rsta.2022.0181>.
- [51] Keller, D. P. *et al.* The Carbon Dioxide Removal Model Intercomparison Project (CDRMIP): Rationale and experimental protocol for CMIP6. *Geosci. Model Dev.* **11**, 1133–1160 (2018). <https://doi.org/10.5194/gmd-11-1133-2018>.
- [52] Jackson, L. C., Schaller, N., Smith, R. S., Palmer, M. D. & Vellinga, M. Response of the Atlantic meridional overturning circulation to a reversal of greenhouse gas increases. *Clim Dyn* **42**, 3323–3336 (2014). <https://doi.org/10.1007/s00382-013-1842-5>.
- [53] Mehling, O., Bellomo, K. & von Hardenberg, J. Centennial-Scale Variability of the Atlantic Meridional Overturning Circulation in CMIP6 Models Shaped by Arctic–North Atlantic Interactions and Sea Ice Biases. *Geophys. Res. Lett.* **51**, e2024GL110791 (2024). <https://doi.org/10.1029/2024GL110791>.
- [54] Yu, H. *et al.* Incomplete Arctic Sea-Ice Recovery Under CO₂ Removal and Its Effects on the Winter Atmospheric Circulation. *Geophys. Res. Lett.* **52**, e2024GL113541 (2025). <https://doi.org/10.1029/2024GL113541>.
- [55] Schwinger, J., Asaadi, A., Steinert, N. J. & Lee, H. Emit now, mitigate later? Earth system reversibility under overshoots of different magnitudes and durations. *Earth Syst. Dyn.* **13**, 1641–1665 (2022). <https://doi.org/10.5194/esd-13-1641-2022>.
- [56] Devilliers, M., Yang, S., Drews, A., Schmith, T. & Olsen, S. M. Ocean response to a century of observation-based freshwater forcing around Greenland in EC-Earth3. *Clim. Dyn.* **62**, 4905–4923 (2024). <https://doi.org/10.1007/s00382-024-07142-0>.
- [57] Lique, C. & Thomas, M. D. Latitudinal shift of the Atlantic Meridional Overturning Circulation source regions under a warming climate. *Nat. Clim. Change* **8**, 1013–1020 (2018). <https://doi.org/10.1038/s41558-018-0316-5>.
- [58] Brodeau, L. & Koenigk, T. Extinction of the northern oceanic deep convection in an ensemble of climate model simulations of the 20th and 21st centuries. *Clim. Dyn.* **46**, 2863–2882 (2016). <https://doi.org/10.1007/s00382-015-2736-5>.
- [59] Lique, C., Johnson, H. L. & Plancherel, Y. Emergence of deep convection in the Arctic Ocean under a warming climate. *Clim. Dyn.* **50**, 3833–3847 (2018). <https://doi.org/10.1007/s00382-017-3849-9>.
- [60] Heuzé, C. & Liu, H. No Emergence of Deep Convection in the Arctic Ocean Across CMIP6 Models. *Geophys. Res. Lett.* **51**, e2023GL106499 (2024). <https://doi.org/10.1029/2023GL106499>.

- [61] Dijkstra, H. A. The role of conceptual models in climate research. *Physica D* **457**, 133984 (2024). <https://doi.org/10.1016/j.physd.2023.133984>.
- [62] Willeit, M. & Ganopolski, A. Generalized stability landscape of the Atlantic meridional overturning circulation. *Earth Syst. Dyn.* **15**, 1417–1434 (2024). <https://doi.org/10.5194/esd-15-1417-2024>.
- [63] Swingedouw, D. *et al.* On the reduced sensitivity of the Atlantic overturning to Greenland ice sheet melting in projections: A multi-model assessment. *Clim. Dyn.* **44**, 3261–3279 (2015). <https://doi.org/10.1007/s00382-014-2270-x>.
- [64] Bochow, N. *et al.* Overshooting the critical threshold for the Greenland ice sheet. *Nature* **622**, 528–536 (2023). <https://doi.org/10.1038/s41586-023-06503-9>.
- [65] Sinet, S., von der Heydt, A. S. & Dijkstra, H. A. West Antarctic Meltwater can Prevent an AMOC Collapse (2025). <https://doi.org/10.48550/arXiv.2502.17104>.
- [66] Li, Q., England, M. H., Hogg, A. M., Rintoul, S. R. & Morrison, A. K. Abyssal ocean overturning slowdown and warming driven by Antarctic meltwater. *Nature* **615**, 841–847 (2023). <https://doi.org/10.1038/s41586-023-05762-w>.
- [67] Martin, T. & Biastoch, A. On the ocean’s response to enhanced Greenland runoff in model experiments: Relevance of mesoscale dynamics and atmospheric coupling. *Ocean Sci.* **19**, 141–167 (2023). <https://doi.org/10.5194/os-19-141-2023>.
- [68] Shin, Y. *et al.* Fast and Slow Responses of Atlantic Meridional Overturning Circulation to Antarctic Meltwater Forcing. *Geophys. Res. Lett.* **51**, e2024GL108272 (2024). <https://doi.org/10.1029/2024GL108272>.
- [69] Wunderling, N. *et al.* Climate tipping point interactions and cascades: A review. *Earth Syst. Dyn.* **15**, 41–74 (2024). <https://doi.org/10.5194/esd-15-41-2024>.
- [70] Eyring, V. *et al.* Overview of the Coupled Model Intercomparison Project Phase 6 (CMIP6) experimental design and organization. *Geosci. Model Dev.* **9**, 1937–1958 (2016). <https://doi.org/10.5194/gmd-9-1937-2016>.
- [71] Balsamo, G. *et al.* A Revised Hydrology for the ECMWF Model: Verification from Field Site to Terrestrial Water Storage and Impact in the Integrated Forecast System. *J. Hydrometeorol.* **10**, 623–643 (2009). <https://doi.org/10.1175/2008JHM1068.1>.
- [72] Madec, G. & NEMO team. NEMO ocean engine. Notes Du Pôle de Modélisation de l’Institut Pierre-Simon Laplace (IPSL) (2016).
- [73] Rousset, C. *et al.* The Louvain-La-Neuve sea ice model LIM3.6: Global and regional capabilities. *Geosci. Model Dev.* **8**, 2991–3005 (2015). <https://doi.org/10.5194/gmd-8-2991-2015>.

- [74] Craig, A., Valcke, S. & Coquart, L. Development and performance of a new version of the OASIS coupler, OASIS3-MCT_3.0. *Geosci. Model Dev.* **10**, 3297–3308 (2017). <https://doi.org/10.5194/gmd-10-3297-2017>.
- [75] Johns, W. E. *et al.* Towards two decades of Atlantic Ocean mass and heat transports at 26.5° N. *Phil. Trans. R. Soc. A* **381**, 20220188 (2023). <https://doi.org/10.1098/rsta.2022.0188>.
- [76] Fu, Y. *et al.* Seasonality of the Meridional Overturning Circulation in the subpolar North Atlantic. *Commun. Earth Environ.* **4**, 181 (2023). <https://doi.org/10.1038/s43247-023-00848-9>.
- [77] Arumí-Planas, C. *et al.* A Multi-Data Set Analysis of the Freshwater Transport by the Atlantic Meridional Overturning Circulation at Nominally 34.5°S. *J. Geophys. Res. Oceans* **129**, e2023JC020558 (2024). <https://doi.org/10.1029/2023JC020558>.
- [78] de Vries, P. & Weber, S. L. The Atlantic freshwater budget as a diagnostic for the existence of a stable shut down of the meridional overturning circulation. *Geophys. Res. Lett.* **32** (2005). <https://doi.org/10.1029/2004GL021450>.
- [79] Huisman, S. E., den Toom, M., Dijkstra, H. A. & Drijfhout, S. An Indicator of the Multiple Equilibria Regime of the Atlantic Meridional Overturning Circulation. *J. Phys. Oceanogr.* **40**, 551–567 (2010). <https://doi.org/10.1175/2009JPO4215.1>.
- [80] O’Neill, B. C. *et al.* The Scenario Model Intercomparison Project (ScenarioMIP) for CMIP6. *Geosci. Model Dev.* **9**, 3461–3482 (2016). <https://doi.org/10.5194/gmd-9-3461-2016>.
- [81] Meinshausen, M. *et al.* The shared socio-economic pathway (SSP) greenhouse gas concentrations and their extensions to 2500. *Geosci. Model Dev.* **13**, 3571–3605 (2020). <https://doi.org/10.5194/gmd-13-3571-2020>.
- [82] Hausfather, Z. & Peters, G. P. Emissions – the ‘business as usual’ story is misleading. *Nature* **577**, 618–620 (2020). <https://doi.org/10.1038/d41586-020-00177-3>.
- [83] Muntjewerf, L. *et al.* Description and Demonstration of the Coupled Community Earth System Model v2 – Community Ice Sheet Model v2 (CESM2-CISM2). *J. Adv. Model. Earth Syst.* **13**, e2020MS002356 (2021). <https://doi.org/10.1029/2020MS002356>.
- [84] Deser, C. *et al.* Insights from Earth system model initial-condition large ensembles and future prospects. *Nat. Clim. Chang.* **10**, 277–286 (2020). <https://doi.org/10.1038/s41558-020-0731-2>.
- [85] Wyser, K. *et al.* The SMHI Large Ensemble (SMHI-LENS) with EC-Earth3.3.1. *Geosci. Model Dev.* **14**, 4781–4796 (2021). <https://doi.org/10.5194/gmd-14-4781-2021>.
- [86] Danabasoglu, G. *et al.* The Community Earth System Model Version 2 (CESM2). *J. Adv. Model. Earth Syst.* **12**, e2019MS001916 (2020). <https://doi.org/10.1029/2019MS001916>.

- [87] Lipscomb, W. H. *et al.* Description and evaluation of the Community Ice Sheet Model (CISM) v2.1. *Geosci. Model Dev.* **12**, 387–424 (2019). <https://doi.org/10.5194/gmd-12-387-2019>.
- [88] Devilliers, M. Experimental protocol and source code for realistic freshwater experiments with EC-Earth3 and IPSL-CM6-LR (2023). <https://doi.org/10.5281/zenodo.10214936>.
- [89] Bamber, J. L. *et al.* Land Ice Freshwater Budget of the Arctic and North Atlantic Oceans: 1. Data, Methods, and Results. *J. Geophys. Res. Oceans* **123**, 1827–1837 (2018). <https://doi.org/10.1002/2017JC013605>.
- [90] Schmittner, A. *et al.* Greenland ice sheet melting influence on the North Atlantic. *US Clivar Var.* **14**, 32–37 (2016).
- [91] Slater, D. A. *et al.* Estimating Greenland tidewater glacier retreat driven by submarine melting. *The Cryosphere* **13**, 2489–2509 (2019). <https://doi.org/10.5194/tc-13-2489-2019>.
- [92] Greve, R. & Chambers, C. Mass loss of the Greenland ice sheet until the year 3000 under a sustained late-21st-century climate. *J. Glaciol.* **68**, 618–624 (2022). <https://doi.org/10.1017/jog.2022.9>.
- [93] Fettweis, X. *et al.* Estimating the Greenland ice sheet surface mass balance contribution to future sea level rise using the regional atmospheric climate model MAR. *The Cryosphere* **7**, 469–489 (2013). <https://doi.org/10.5194/tc-7-469-2013>.
- [94] Little, C. M., Piecuch, C. G. & Chaudhuri, A. H. Quantifying Greenland freshwater flux underestimates in climate models. *Geophys. Res. Lett.* **43**, 5370–5377 (2016). <https://doi.org/10.1002/2016GL068878>.
- [95] Davini, P., von Hardenberg, J., Filippi, L. & Provenzale, A. Impact of Greenland orography on the Atlantic Meridional Overturning Circulation. *Geophys. Res. Lett.* **42**, 871–879 (2015). <https://doi.org/10.1002/2014GL062668>.
- [96] Andernach, M., Kapsch, M.-L. & Mikolajewicz, U. Impact of Greenland Ice Sheet disintegration on atmosphere and ocean disentangled. *Earth Syst. Dyn.* **16**, 451–474 (2025). <https://doi.org/10.5194/esd-16-451-2025>.
- [97] Lozier, M. S. *et al.* A sea change in our view of overturning in the subpolar North Atlantic. *Science* **363**, 516–521 (2019). <https://doi.org/10.1126/science.aau6592>.
- [98] Zhang, R. Latitudinal dependence of Atlantic meridional overturning circulation (AMOC) variations. *Geophys. Res. Lett.* **37** (2010). <https://doi.org/10.1029/2010GL044474>.
- [99] Menary, M. B., Jackson, L. C. & Lozier, M. S. Reconciling the Relationship Between the AMOC and Labrador Sea in OSNAP Observations and Climate Models. *Geophys. Res. Lett.* **47**, e2020GL089793 (2020). <https://doi.org/10.1029/2020GL089793>.

- [100] Jackson, L. C. & Petit, T. North Atlantic overturning and water mass transformation in CMIP6 models. *Clim. Dyn.* **60**, 2871–2891 (2023). <https://doi.org/10.1007/s00382-022-06448-1>.
- [101] Winkelbauer, S., Mayer, M. & Haimberger, L. StraitFlux – precise computations of water strait fluxes on various modeling grids. *Geosci. Model Dev.* **17**, 4603–4620 (2024). <https://doi.org/10.5194/gmd-17-4603-2024>.
- [102] Walin, G. On the relation between sea-surface heat flow and thermal circulation in the ocean. *Tellus* **34** (1982). <https://doi.org/10.3402/tellusa.v34i2.10801>.
- [103] Meccia, V. L. *et al.* Internal multi-centennial variability of the Atlantic Meridional Overturning Circulation simulated by EC-Earth3. *Clim. Dyn.* **60**, 3695–3712 (2023). <https://doi.org/10.1007/s00382-022-06534-4>.
- [104] Ebisuzaki, W. A Method to Estimate the Statistical Significance of a Correlation When the Data Are Serially Correlated. *J. Clim.* **10**, 2147–2153 (1997). [https://doi.org/10.1175/1520-0442\(1997\)010<2147:AMTETS>2.0.CO;2](https://doi.org/10.1175/1520-0442(1997)010<2147:AMTETS>2.0.CO;2).