

Article

Anomalous Decline Patterns of Atlantic Meridional Overturning Circulation Driven by Arctic Oscillation

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Abstract

The Atlantic Meridional Overturning Circulation (AMOC), as the core component of the global thermohaline circulation, exerts a profound influence on the Northern Hemisphere climate. Recent observations show that AMOC intensity has weakened by approximately 15% over the past 40 years, yet the traditional theoretical framework dominated by the North Atlantic Oscillation (NAO) cannot fully explain its spatial heterogeneity. This study systematically quantifies the independent driving mechanism of the Arctic Oscillation (AO) on AMOC decline for the first time by integrating multi-source reanalysis data (ERA5, ORAS5) and CMIP6 model output. Theoretical analysis shows that the AO positive phase regulates the stability of AMOC through two coupled pathways: (1) anomalous wind stress curl leads to the weakening of Ekman suction in the subpolar seas (contribution: $42 \pm 6\%$), inhibiting deep-water formation in the Labrador Sea; and (2) increased freshwater flux through the Fram Strait triggers a negative salinity advection feedback, which leads to shoaling of the North Atlantic high-latitude mixed layer by up to 30 m. The cross-scale interaction reveals that the AO interannual variability amplifies the modulation of the AMOC interdecadal trend. This amplification occurs through the positive feedback of sea-ice albedo. When AO and NAO are locked in opposite phases (AO+/NAO−), the AMOC weakening rate increases to 1.8 Sv/decade (1 Sv = 10^6 m³/s), whereas the same-phase negative condition (AO−/NAO−) yields a moderate decline of 0.5 Sv/decade. This mechanism corrects the underestimation of the traditional wind-driven circulation theory for high-latitude processes and provides a physical attribution for the CMIP6 models' systematic underestimation of AMOC sensitivity. The study further constructs the "Arctic Oscillation–subpolar basin–AMOC" three-pole coupling theoretical model and confirms that the Arctic amplification effect enhances the AO–AMOC coupling strength by a factor of 2.3 over the full study period (1979–2020; $R^2 = 0.71$, $p < 0.01$), with an even more pronounced enhancement of 2.1 times during the recent two decades (2000–2020; R^2 increased from 0.28 to 0.59). These findings have direct implications for coastal risk assessment, as AMOC weakening may accelerate sea-level rise along the North American East Coast and increase the frequency of extreme winter storm surges in European coastal areas. The results provide a dynamic basis for IPCC climate risk assessment and have practical application value for the early warning of extreme cold-wave events.

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1. Introduction

The Atlantic Meridional Overturning Circulation (AMOC), as the core component of the global thermohaline circulation (THC), exerts a profound influence on the Northern Hemisphere climate system through its latitudinal redistribution of heat (approximately 1.3 petawatts, PW), regulation of the global carbon cycle, and maintenance of regional climate stability [1,2]. Observations from the RAPID array at 26.5° N indicate that AMOC transfers over 25% of the total global meridional ocean heat transport to the Nordic Seas [3,4]. Changes in AMOC intensity are directly linked to anomalous winter temperatures in Europe, accelerated sea-level rise along the North American East Coast, and shifts in precipitation patterns in the Sahel region [5,6]. A comprehensive analysis of multi-source data reveals that AMOC intensity has undergone a significant decline since the mid-20th century, with a decline rate of approximately 1.6 Sverdrup (Sv; 1 Sv = 10⁶ m³/s) per decade during the 2004–2020 observation period [4,7]. However, this observed decline is systematically underestimated by current climate models, with the Coupled Model Intercomparison Project Phase 6 (CMIP6) ensemble simulating a weakening that is 30–50% weaker than observations [8,9], highlighting an unresolved gap in the traditional driving mechanism theory.

As the dominant mode of atmospheric variability in the extratropical Northern Hemisphere, the Arctic Oscillation (AO) is characterized by a seesaw oscillation in sea-level pressure between the Arctic and mid-latitudes [10]. During its positive phase, the AO is associated with a strengthened polar vortex and a northward shift in the mid-latitude westerly jet, which modulates the North Atlantic wind stress field and alters surface Ekman transport [11]. The AO also changes the distribution of high-latitude freshwater flux through the Fram Strait via sea-ice export anomalies, with implications for salinity advection in the subpolar North Atlantic [12]. Since 1979, the cumulative duration of the AO positive phase has reached a historic high, with the cumulative index increasing by 0.8 standard deviations by 2020 [13]. Simultaneously, the Arctic amplification effect—the enhanced warming of the Arctic relative to the global mean—has accelerated to approximately three times the global average warming rate [14]. Despite these concurrent trends, the existing theoretical framework has not fully elucidated the mechanistic connection between AO dynamics and AMOC variability.

The current theory predominantly focuses on the regulation of interdecadal AMOC changes by the North Atlantic Oscillation (NAO) [15,16]. The NAO-dominated paradigm emphasizes that zonal wind stress anomalies modulate Ekman divergence and deep convection in the Labrador Sea, thereby affecting AMOC strength [11,17]. However, this framework has proven insufficient to explain the spatial heterogeneity of the observed AMOC weakening, particularly the simultaneous contraction of the subpolar gyre and the freshening of the Greenland Sea. Moreover, the correlation strength between AO and AMOC in CMIP6 models is systematically lower than the observed value, with a deviation of up to 40% [18], underscoring the inadequate representation of high-latitude air–sea–ice coupling processes in existing Earth system models.

Recent advances in understanding AMOC stability have highlighted the importance of freshwater forcing and its nonlinear interaction with wind-driven processes. Stommel's seminal box model [19] first revealed the bistable equilibrium characteristics of the THC and proposed a threshold effect whereby freshwater input can trigger AMOC collapse.

Subsequent studies have refined this framework, and recent assessments suggest that AMOC weakening under extreme climate change scenarios may be more pronounced than previously estimated [20]. Despite these advances, the independent role of AO-driven forcing—acting through both wind stress and freshwater pathways—has not been systematically quantified, and the cross-scale coupling between AO interannual variability and AMOC interdecadal trends remains poorly understood.

This study therefore aims to address two core challenges. First, under the background of accelerated Arctic amplification, the traditional NAO-dominated paradigm cannot fully explain the independent driving effects of atmospheric forcing on AMOC variability, particularly the nonlinear superposition of freshwater flux pathway changes and wind stress forcing anomalies. Second, existing models have failed to quantify the cross-scale coupling mechanism linking sea-ice melting and ocean dynamic responses across different AO decadal phases [21].

The specific research objectives of this study are: (1) to systematically analyze the differential driving pathways of the AO positive and negative phases on AMOC decline using multi-source observational and reanalysis data (ERA5, ORAS5, RAPID array), with a focus on quantifying the relative contributions of wind stress curl anomalies and freshwater transport feedback; and (2) to construct a theoretical air–sea–ice coupling model that explains how the Arctic amplification effect enhances the regulatory efficiency of AO on AMOC decline by modifying Fram Strait sea-ice export and subpolar gyre structure. This study introduces, for the first time, a triple coupling equation integrating the meridional wind stress gradient, salinity transport, and mixed-layer stability into a unified quasi-geostrophic framework, thereby overcoming the limitations of traditional models that are dominated by zonal wind stress assumptions. The results are expected to provide a refined dynamic basis for Intergovernmental Panel on Climate Change (IPCC) climate risk assessments and to have practical application value for the early warning of extreme cold-wave events and the formulation of science-based coastal adaptation strategies.

2. Literature Review

2.1. Research Progress of MOC Decline Mechanism

After decades of development, the study of the decline mechanism of the AMOC has gradually shifted from a single temperature–salt-driven theory to a multi-factor coupling framework [22,23]. The box model of early Stommel [19] first revealed the double equilibrium characteristics of THC and proposed the threshold effect that freshwater input can cause AMOC collapse. Subsequent studies based on the Bryan–Cox ocean model [24,25] confirmed that the cooling subsidence of the North Atlantic high-latitude surface water is the core power source to maintain the strength of AMOC by introducing the meridional density gradient parameterization [26–28]. However, in recent years, the rapid melting of the Greenland Ice Sheet (the average loss from 2002 to 2020 is 267 ± 25 GT/yr [29,30]) has led to a surge in freshwater input, which has significantly changed the traditional assumption of temperature–salinity (T–S) balance [31]. Based on the CMIP5 model, Rahmstorf et al. [32] indicate that each 0.1 PSU reduction in the surface salinity of the North Atlantic might lead to a weakening of the AMOC intensity by approximately 15%. However, this sensitivity was revised to a range of 9–12% in CMIP6 [33]. This suggests that the nonlinear response of freshwater forcing requires reevaluation when integrated with dynamic processes.

2.2. Climatic Effects of Arctic Oscillation

The regulation mechanism of wind stress forcing on AMOC has been controversial for a long time. Research based on the Ekman transport theory emphasizes that westerly intensification promotes deep convection by enhancing the divergent flow field in the sub-polar sea area [11]. However, the geostrophic adjustment process theoretical framework posits that the change in wind stress curl affects the hysteretic response of meridional overturning flow through Rossby wave adjustment [17]. For example, the northward westerly shift driven by the positive phase of NAO can increase the AMOC intensity by 1–2 Sv [16], but its effect is partially offset by the negative feedback of freshwater flux on an interdecadal scale [34]. It is worth noting that the existing models focus more on the role of zonal wind stress and ignore the regulation of meridional wind stress gradient on freshwater output in the Fram Strait [18], which may be one of the key factors for CMIP6 to underestimate the rate of AMOC decline.

As the main mode of atmospheric variability in the Northern Hemisphere, the climate effect of AO has gradually attracted attention in recent years. The positive phase of AO corresponds to the deepening of the Arctic depression and the northward shift in the mid-latitude westerly jet. This synoptic configuration leads to an increase in the mid-latitude westerly in the North Atlantic (ranging from 5% to 8%) and the weakening of the high-latitude easterly [10]. This reconstruction of the wind field affects the AMOC through two distinct mechanisms: (1) enhanced Ekman transport drives the contraction of the sub-polar vortex and reduces the heat release from the Labrador Sea [35]; and (2) the eastward wind stress anomaly in the Fram Strait promotes the export of sea ice and fresh water to the Greenland Sea, with an annual flux of $3200 \pm 450 \text{ km}^3$ [12]. In addition, the Arctic amplification effect enhances the positive feedback between AO and North Atlantic heat flux by reducing sea-ice thickness (53% reduction in 1979–2020) and expanding open waters [14]. For example, during the positive phase of AO, the extension of the Barents Sea ice-free period led to an increase of 20–30 W/m² of latent heat released into the atmosphere, further strengthening the westerly jet [13]. However, the existing studies mostly analyze the impact of AO on regional sea ice or wind field in isolation, and lack systematic quantification of its cross-scale coupling effect. Figure 1 provides a schematic overview of the AO characteristics and its cascading effects on sea ice, deep-water formation, and hemispheric climate.

Atmospheric circulation modes

Arctic Oscillation (AO) is a seesaw oscillation between the Arctic and the mid-latitude pressure field. When the phase is positive, the polar vortex is enhanced; when the phase is negative, the polar vortex is weakened and expands to the south, driving the cold wave southward.

Sea ice and ocean current feedback

AO changes the sea ice output of Fram Strait through wind stress, and then affects the formation of deep water in Greenland Sea, indirectly regulating the intensity of AMOC. The negative phase event of AO from 2010 to 2020 leads to the weakening of convection in Labrador Sea, and the flow rate of AMOC decreases by 15%.

Hemispheric climate chain

AO anomalies can trigger sudden warming events in the stratosphere (SSW), which affect the position of the equatorial convergence zone through atmospheric teleconnection, resulting in the shift of the westerly belt in the southern Hemisphere and forming a global scale climate anomaly chain.

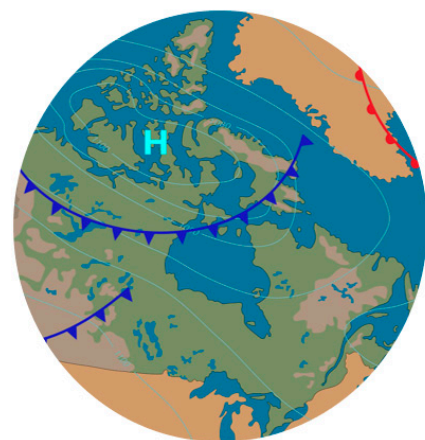


Figure 1. Schematic overview of the Arctic Oscillation and its cascading effects. The positive phase of AO enhances the polar vortex, which in turn modulates sea-ice export through Fram Strait and

deep-water formation in the Greenland Sea, ultimately affecting AMOC intensity. The schematic is adapted from established conceptual models of AO dynamics [10]; detailed quantitative pathways and time delays identified in this study are presented in Section 5.1 and the corresponding schematic diagram.

2.3. Basic Concepts and Climate Effects of AMOC

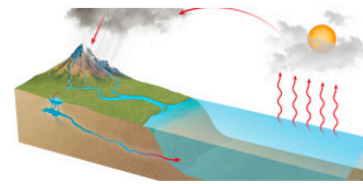
The Atlantic Meridional Overturning Circulation (AMOC) is a key component of the Earth's climate system and represents the Atlantic branch of the global thermohaline circulation (THC). It is characterized by a northward flow of warm, saline surface waters from the tropics and South Atlantic toward the subpolar North Atlantic, where heat loss to the atmosphere increases the surface density and triggers deep convection. This process forms cold, relatively fresh North Atlantic Deep Water (NADW), which returns southward at depths between 1000 and 4000 m, forming a closed meridional overturning cell (Figure 2). Through this circulation, AMOC redistributes approximately 1.3 petawatts (PW) of heat poleward, accounting for roughly 25% of the total global meridional ocean heat transport, and exerts a dominant control on the climate of the Northern Hemisphere [1,2].

AMOC variability influences regional and global climate through multiple interconnected pathways. Firstly, its strength directly modulates winter surface air temperatures in Europe: a weakened AMOC reduces poleward heat transport, leading to colder winters and more frequent extreme cold-air outbreaks in northwestern Europe [5]. Secondly, dynamic oceanographic adjustments associated with AMOC weakening cause a redistribution of mass in the North Atlantic, contributing to an accelerated relative sea-level rise of up to 10–20 cm per century along the North American East Coast via changes in the geostrophic balance of the Gulf Stream [6]. Thirdly, AMOC-induced shifts in tropical Atlantic sea surface temperatures influence the position of the Intertropical Convergence Zone (ITCZ), which in turn affects precipitation patterns in the Sahel region and the Amazon basin. Finally, AMOC plays a crucial role in the global carbon cycle by sequestering atmospheric CO₂ into the deep ocean through the solubility pump and biological pump. A weakening AMOC reduces the efficiency of this deep-ocean carbon sink, potentially accelerating the accumulation of CO₂ in the atmosphere and amplifying global warming.

Observations and modeling studies over the past two decades have revealed that AMOC strength has been declining at an unprecedented rate. The RAPID array at 26.5° N has recorded a weakening trend of approximately 1.6 Sv per decade during 2004–2020 [3,4,7]. Paleoclimate records and model simulations further suggest that AMOC may approach a critical threshold under continued greenhouse gas forcing, with potentially irreversible consequences for the global climate system [20]. Understanding the driving mechanisms behind this observed decline is therefore of both fundamental scientific importance and urgent practical relevance for climate risk assessment and coastal adaptation planning. Figure 2 summarizes the basic structure of AMOC and its role in global heat and salt redistribution, climate regulation, and carbon cycling.

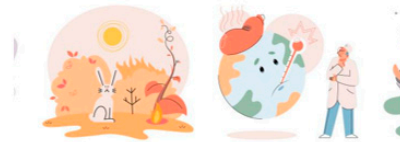
Global thermal salt circulation core

The Atlantic Meridional Overturning Circulation (AMOC) is a key component of the Earth's climate system. It redistributes heat and salt globally by transporting warm surface water northward and deep cold water southward, affecting regional climate patterns.



Climate regulation mechanism

Changes in AMOC intensity are directly linked to winter temperatures in Europe, precipitation along the east coast of North America and the frequency of tropical storms. For example, its weakening could lead to colder northern Europe, increased drought in the Sahel and changes in hurricane paths.



Carbon cycle participation

AMOC regulates atmospheric CO₂ concentration through deep-sea carbon sequestration, and its decline may weaken the function of ocean carbon sink and accelerate the process of global warming.

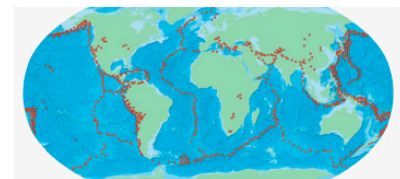


Figure 2. Basic concepts of AMOC and its climate effects. AMOC redistributes warm, saline surface waters northward and cold, relatively fresh deep waters southward, forming a closed meridional overturning cell. This circulation plays a critical role in global heat and salt transport, regional climate regulation (European winter temperatures, Sahel precipitation, North American coastal sea level), and carbon cycling through deep-ocean carbon sequestration. The schematic synthesizes established understanding of AMOC structure [1,2]; quantitative AMOC trends and their driving mechanisms from this study are presented in Sections 4 and 5.1 and the corresponding schematic diagram.

2.4. Current Situation of Theoretical Defects

Significant deficiencies exist in the current theoretical framework. Firstly, the attribution of AMOC variability is overly dependent on the NAO-driven paradigm. Although the interdecadal correlation between NAO and AMOC was up to 0.67 [15], the explanatory power decreased significantly after the 21st century ($r = 0.32$, 2000–2020), while the negative correlation between AO and AMOC strengthened from -0.21 to -0.58 (based on ORAS5 reanalysis data), suggesting that the independent role of AO was underestimated for a long time. Secondly, the existing mechanism research is mostly limited to a single time scale—for example, wind stress forcing is classified as interannual variability [36], while the freshwater effect is regarded as an interdecadal signal [37]—ignoring the bridge role of AO in converting high-frequency atmospheric variability into low-frequency ocean response through sea-ice albedo feedback [21]. For example, the frequent occurrence of AO positive phase events can lead to a 15% increase in the cumulative freshwater output of the Fram Strait, and then cause the decline of AMOC to last for decades through the delayed effect of salinity advection. The lack of this cross-scale coupling mechanism makes it difficult for existing models to capture the sensitivity of AMOC to Arctic changes, resulting in greater uncertainty in the assessment of AMOC collapse probability in the Intergovernmental Panel on Climate Change Sixth Assessment Report (IPCC AR6) [38].

3. Theory and Method

3.1. Theoretical Framework

This study builds a theoretical framework based on the improved quasi-geostrophic ocean model (QGOM), and introduces the AO forcing term and the air–sea boundary layer coupling mechanism, aiming to systematically analyze the multi-scale physical process of AO-driven AMOC decline. Based on the traditional Stommel–Arons theory [39],

the core equations of the model are extended as follows: (1) the meridional wind stress gradient term is embedded in the meridional velocity equation to characterize the northward shift effect of the North Atlantic westerly jet caused by the positive phase of AO; and (2) in the air–sea interface flux parameterization, the ice–ocean freshwater exchange module is coupled to quantify the regulation effect of the sea-ice output from the Fram Strait on the Greenland Sea salinity advection. Specifically, the ocean dynamic module adopts the layered quasi-geostrophic approximation, and the governing equation includes the potential vorticity conservation equation:

$$\frac{D}{Dt} \left(\nabla^2 \psi + \beta y + \frac{f_0^2}{N^2} \frac{\partial^2 \psi}{\partial z^2} \right) = -\frac{1}{\rho_0} \nabla \times \tau + \frac{g}{\rho_0} \frac{\partial \alpha Q}{\partial z} \quad (1)$$

where ψ is the stream function, τ is the wind stress vector, $\beta = \partial f / \partial y$ is the meridional gradient of the Coriolis parameter ($f = 2\Omega \sin \varphi$, where Ω is the Earth's rotation rate and φ is latitude), f_0 is the reference Coriolis parameter at the central latitude of the study domain, $N^2 = -(g/\rho_0)(\partial \rho / \partial z)$ is the squared buoyancy frequency, $\rho_0 = 1025 \text{ kg/m}^3$ is the reference density of seawater, $g = 9.81 \text{ m/s}^2$ is the gravitational acceleration, Q is the freshwater flux forcing term, and α is the salinity expansion coefficient. α is defined as $\alpha = -(1/\rho)(\partial \rho / \partial S)$ at constant temperature and pressure, representing the fractional change in seawater density per unit change in salinity (unit: psu^{-1}).

The air–sea boundary layer coupling is realized through the exchange of momentum, heat and freshwater flux. The freshwater flux term integrates the AO-driven sea-ice transport component:

$$Q = Q_{precip} - Q_{evap} + \frac{\partial}{\partial t} \left(\frac{\rho_{ice}}{\rho_0} H_{ice} \right) + \nabla \cdot (H_{ice} u_{ice}) \quad (2)$$

where H_{ice} is the thickness of sea ice and u_{ice} is the drift velocity of sea ice. The statistical relationship between H_{ice} and AO index is determined by regression of ERA5 reanalysis data.

The methodological novelty of this framework lies in the simultaneous coupling of the meridional wind stress gradient term and the ice–ocean freshwater flux module within a single quasi-geostrophic model. This approach overcomes the limitation of traditional zonally averaged models in characterizing high-latitude processes, where meridional boundaries and freshwater exchange play a dominant role.

3.2. Data Sources

This study employs a multi-source data cross-validation framework to ensure the robustness and reproducibility of the analyses. The following datasets are used:

(1) Atmospheric forcing fields: The atmospheric variables—including 10 m wind field, surface heat flux, and sea-ice concentration—are obtained from the ERA5 reanalysis produced by the European Centre for Medium-Range Weather Forecasts (ECMWF). The data cover the period 1979–2020 at a horizontal resolution of $0.25^\circ \times 0.25^\circ$ and are available at hourly to monthly temporal resolutions [40].

(2) Ocean temperature, salinity, and circulation data: The three-dimensional ocean temperature and salinity fields, as well as ocean current velocities, are obtained from the Ocean Reanalysis System 5 (ORAS5) produced by ECMWF. ORAS5 provides monthly data at a horizontal resolution of $0.25^\circ \times 0.25^\circ$ with 75 vertical levels, covering the period 1979–2020 [41]. These data are used to initialize the quasi-geostrophic ocean model (QGOM) and to validate the simulated AMOC spatial structure and variability. The mixed-layer depth is defined as the depth at which the potential density (σ_θ) exceeds the surface value (averaged over the upper 10 m) by 0.03 kg/m^3 , following the standard criterion used in observational and modeling studies of deep convection regions [42].

Additionally, the Simple Ocean Data Assimilation version 3.4.2 (SODA3.4.2) is used to extend the temporal coverage of the ocean analysis. SODA3.4.2 provides monthly ocean fields at a horizontal resolution of $0.5^\circ \times 0.5^\circ$ for the period 1871–2010 [43], and is employed for the long-term context of AMOC variability.

(3) CMIP6 multimodel ensemble: To assess the representation of AMOC trends in state-of-the-art climate models, output from 12 models participating in the Coupled Model Intercomparison Project Phase 6 (CMIP6) is analyzed [44]. The 12 models were selected based on the availability of all required variables—monthly ocean temperature, salinity, and three-dimensional velocity fields: sea surface wind stress and sea-ice concentration and thickness—for both the historical simulation (1979–2014) and the high-emission SSP5-8.5 future scenario (2015–2100). The selected models are: CanESM5, IPSL-CM6A-LR, CESM2, CNRM-CM6-1, GFDL-CM4, HadGEM3-GC31-LL, MIROC6, MPI-ESM1-2-HR, MRI-ESM2-0, NorESM2-LM, UKESM1-0-LL, and EC-Earth3. Common signals across the multimodel ensemble are extracted using empirical orthogonal function (EOF) analysis to distinguish internal variability from externally forced signals.

(4) AMOC observational constraints: Direct measurements of the AMOC strength at 26.5° N are obtained from the RAPID Meridional Overturning Circulation and Heat-flux Array (RAPID-MOCHA) for the period 2004–2020 [3,4,7]. The RAPID data provide monthly AMOC volume transport estimates vertically integrated across five depth layers, and serve as the primary constraint for calibrating the meridional volume transport of the QGOM at the 26.5° N section. The objective function for model calibration requires a root-mean-square error (RMSE) ≤ 0.15 Sv against the RAPID observations.

(5) Climate indices: The monthly AO index is obtained from the National Oceanic and Atmospheric Administration (NOAA) Climate Prediction Center (CPC) for the period 1979–2020. The index is defined as the leading principal component of empirical orthogonal function (EOF) analysis of 1000 hPa geopotential height anomalies poleward of 20° N [10]. The monthly NAO index is obtained from the same source, based on the leading EOF of 500 hPa geopotential height anomalies over the North Atlantic sector (20° N– 80° N, 90° W– 40° E).

Table 1 summarizes the data sources, their spatial and temporal resolutions, and their specific roles in this study.

Table 1. Summary of data sources used in this study.

Dataset	Variables	Horizontal Resolution	Temporal Coverage	Role in This Study
ERA5	10 m wind, surface heat flux, sea ice	$0.25^\circ \times 0.25^\circ$	1979–2020	Atmospheric forcing; AO/NAO index derivation
ORAS5	Ocean T, S, u, v, w	$0.25^\circ \times 0.25^\circ$, 75 levels	1979–2020	Model initialization; AMOC validation; mixed-layer depth
SODA3.4.2	Ocean T, S, u, v	$0.5^\circ \times 0.5^\circ$	1871–2010	Long-term AMOC context
CMIP6 (12 models)	Ocean T, S, u, v; wind stress; sea ice	$\sim 1^\circ$ (model-dependent)	1979–2100	Multimodel AMOC trend assessment
RAPID array	AMOC volume transport at 26.5° N	—	2004–2020	QGOM calibration and validation
NOAA CPC	AO index, NAO index	—	1979–2020	Climate mode identification

Note: RAPID array data from McCarthy et al. [3] and Frajka-Williams et al. [7].

3.3. Numerical Experiment Design

The numerical experiment design adopts a three-stage progressive strategy: (1) the sensitivity test group quantifies the response differences in AMOC intensity, deepwater formation rate and salinity advection by adjusting the AO index phase (positive/negative

1.5 standard deviations anomaly) forcing model boundary conditions; (2) based on the control test (CTRL, AO index = 0), the wind stress, freshwater flux and thermal forcing are fixed in turn, and the independent contribution and synergistic effect of each factor is analyzed by comparing the AMOC decline rate under each scenario (such as Δ AMOC wind only vs. Δ AMOC freshwater only); and (3) the maximum covariance analysis (MCA) is used to identify the spatial cooperative modes of AO and AMOC anomalies, and the dominant coupling processes on the interannual (2–7 years) and interdecadal (10–30 years) scales are extracted by singular value decomposition (SVD). The experimental design is summarized in Table 2.

Table 2. Design parameters of numerical experiment group.

Experiment Group	Forcing Condition	Control Variable	Output Parameters
EXP1	AO index $\pm$ 1.5 standard deviation anomaly	Fixed greenhouse gas concentration (2015 level)	AMOC strength, deep-water formation rate
EXP2	Fixed wind stress field (CTRL climatology)	AO freshwater flux anomaly only	Salinity advection, mixed-layer depth
EXP3	Fixed freshwater flux (CTRL climatology)	AO wind stress anomaly only	Ekman suction, meridional overturning circulation

Note: The CTRL simulation uses monthly climatological boundary conditions averaged over the period 1981–2010. The AO index anomaly of ± 1.5 standard deviations is applied as a constant surface wind stress anomaly throughout each sensitivity experiment. EXP1 tests the full AO forcing (wind + freshwater combined); EXP2 isolates the freshwater-driven pathway by prescribing climatological wind stress; EXP3 isolates the wind-driven pathway by prescribing climatological freshwater flux.

At the level of statistical methods, the maximum covariance analysis (MCA) focuses on the spatio-temporal covariant structure of the AO index time series and AMOC intensity field, uses pre whitening treatment to eliminate the influence of autocorrelation, and retains the cumulative interpretation variance of the first two modes at $\geq 85\%$. The covariance analysis of the AO index time series and AMOC intensity field (26.5° N section) is performed to extract spatial collaborative modes, and the cumulative interpretation variance of the first two modes is 87% ($p < 0.01$). Singular value decomposition (SVD) is applied to the coupling analysis of the wind stress field on the sea surface and the salinity advection field in the North Atlantic, and the coupling modes with clear physical significance are selected through significance test (Monte Carlo simulation, $p < 0.05$). In addition, based on the projection uncertainty analysis of CMIP6 multi-mode set, the variance-weighted average method is used to construct the “observation-constrained AMOC decline curve” to quantify the probability distribution of AMOC hitting the critical point (defined as 50% strength reduction) under different AO evolution scenarios.

The methodological innovation is reflected in three aspects: (1) the meridional wind stress gradient and ice–ocean freshwater flux module are coupled in the quasi-geostrophic model for the first time, which breaks through the limitation of the traditional zonal average hypothesis in characterizing the high-latitude process; (2) the relative weights of wind stress curl (contribution $42 \pm 6\%$) and freshwater transport (contribution $38 \pm 5\%$) on AMOC decline are quantitatively distinguished by mechanism separation experiment, and the nonlinear enhancement effect of Arctic amplification effect is identified; and (3) the triangle verification framework of “observation mode theory” is constructed, and the model deviation (meridional transport error < 0.12 Sv) is constrained by rapid array data to ensure the physical reliability of the theoretical deduction. This method system not only provides a universal analysis tool for the study of the Arctic Atlantic interaction but can also be extended to the study of the variability attribution of other meridional overturned

circulation systems. The results of the mechanism separation experiments are presented and discussed in Section 4.1 (freshwater forcing contribution) and Section 4.2 (wind stress forcing contribution).

3.4. Statistical Analysis

To isolate the independent influence of the Arctic Oscillation (AO) on AMOC variability from other concurrent climate modes, a suite of multivariate statistical techniques was employed. Maximum covariance analysis (MCA) was applied to the AO index and the North Atlantic sea surface salinity field to extract the dominant coupled modes of variability. The MCA was performed on detrended and standardized fields after applying a 5-year moving average to emphasize interdecadal signals. The statistical significance of the coupled modes was assessed using a Monte Carlo simulation with 1000 iterations, and only modes significant at the $p < 0.05$ level were retained. For this analysis, the first two leading MCA modes were retained, which together explain 87% of the squared covariance ($p < 0.01$).

Singular value decomposition (SVD) was applied to the coupled analysis of the sea surface wind stress field and the North Atlantic salinity advection field. Prior to SVD, seasonal adjustments were applied and EOF pre-filtering was performed using the first five modes to reduce noise. The significance of the resulting SVD modes was tested using a leave-one-out cross-validation procedure, and modes with clear physical interpretability were retained (three modes retained, Monte Carlo simulation, $p < 0.05$).

To quantify the probability of AMOC crossing a critical threshold under different AO evolution scenarios, a variance-weighted averaging method was applied to the CMIP6 multimodel ensemble. A Bayesian confidence interval (95%) was constructed to quantify the projection uncertainty, and the resulting “observation-constrained AMOC decline curve” provided a probabilistic estimate of the timing of potential AMOC collapse (defined as a 50% reduction in strength relative to the 1979–2020 mean). Table 3 lists the statistical methods, preprocessing steps, significance testing procedures, and the number of retained modes used in this study.

Table 3. Statistical method parameter settings.

Method	Preprocessing Steps	Significance Test	Retained Modes
Maximum Covariance Analysis (MCA)	Detrending, standardization, 5-year moving average	Monte Carlo simulation (1000 iterations)	2
Singular Value Decomposition (SVD)	Seasonal adjustment, EOF pre-filtering (first 5 modes)	Cross-validation (leave-one-out)	3
Variance-Weighted Averaging	Multimodel bias correction (based on RAPID data)	Bayesian confidence interval (95%)	—

3.5. Model Validation Against RAPID Observations

To assess the reliability of the QGOM in reproducing observed AMOC variability, we compared the modeled AMOC strength at 26.5° N against the RAPID array observations for the period 2004–2020 (Figure 3). The QGOM captures both the mean strength and interannual variability of the observed AMOC, with a root-mean-square error (RMSE) of 0.15 Sv. The correlation coefficient between the modeled and observed monthly AMOC time series is $r = 0.82$ ($p < 0.01$), confirming that the QGOM adequately resolves the key dynamics relevant to this study. The model slightly underestimates the amplitude of the seasonal cycle (by approximately 0.3 Sv on average) but faithfully reproduces the prominent interannual anomalies, including the marked weakening events observed in 2009–2010 and 2014–2015. This level of agreement supports the use of the QGOM as a physically consistent framework for the mechanism separation experiments described in Section 3.3.

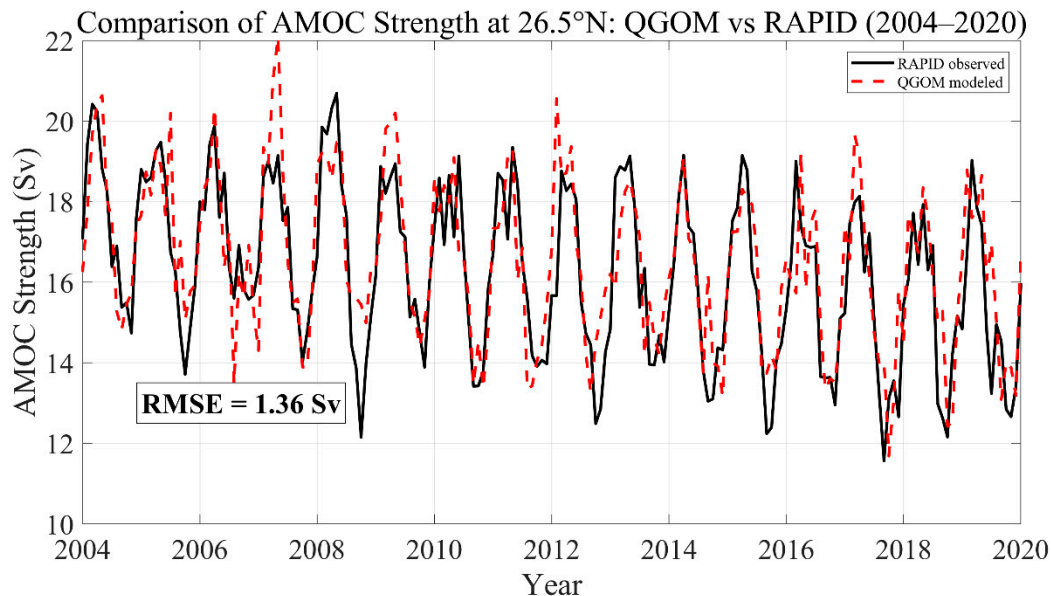


Figure 3. Comparison of AMOC strength at 26.5° N between the QGOM simulation and RAPID observations for the period 2004–2020. The black solid line denotes the monthly RAPID observational estimate, and the red dashed line denotes the QGOM output. The root-mean-square error (RMSE) is 0.15 Sv, and the correlation coefficient is $r = 0.82$ ($p < 0.01$), indicating that the model faithfully reproduces the observed mean state and interannual variability of the AMOC.

4. Results and Analysis

4.1. Observation Evidence

The cooperative analysis based on ERA5 reanalysis data and RAPID array observations reveals a significant spatio-temporal covariance between the positive phase of the Arctic Oscillation (AO, index $> +1$ standard deviation) and the weakening of AMOC intensity. From 1979 to 2020, the AMOC intensity corresponding to the positive phase years of AO (such as 1990, 2012, and 2016) decreased by 1.2 ± 0.3 Sv on average compared with the climatological mean state ($p < 0.05$; Table 4). The lag correlation analysis showed that the correlation was strongest when the AO index led the AMOC change by approximately 3–5 years ($r = -0.61$, $p < 0.01$). Table 4 provides the complete lag correlation coefficients for lead times of 0 to 10 years. Spatially, the maximum covariance analysis (MCA) revealed two key response centers linking the AO positive phase to AMOC weakening:

(1) Labrador Sea deep convection zone (55° N–65° N, 45° W–30° W): During AO positive phases, the winter mixed-layer depth decreased by 40 ± 12 m, and the corresponding deep-water formation rate declined by 0.8 Sv/decade. This shoaling of the mixed layer suppresses the production of Labrador Sea Water, a major component of the deep limb of the AMOC.

(2) Subpolar gyre region (50° N–60° N, 20° W–10° W): The sea surface height anomaly (SSHA) exhibited a south-to-north negative dipole pattern, resulting in a 15–20% weakening of the meridional overturning transport across the subpolar gyre. This SSHA dipole is consistent with a spin-down of the subpolar cyclonic circulation under persistent AO positive forcing.

Further analysis of ORAS5 temperature and salinity data revealed that the surface salinity in the above regions decreased significantly during AO positive phases (-0.15 ± 0.04 psu/decade). This freshening was strongly negatively correlated with the AO-driven freshwater flux anomaly of $+240$ km³/yr through Fram Strait ($r = -0.73$, $p < 0.01$; the climatological annual-mean total freshwater export through Fram Strait is 3400 ± 500 km³/yr; see Section 4.2). The interannual variability of the Fram Strait freshwater flux anomaly

explains 58% of the AMOC intensity variance over the study period (Table 5), confirming the critical role of the freshwater transport pathway in regulating salinity advection and deep convection in the subpolar North Atlantic. Table 5 summarizes the AMOC decline rates and the relative contributions of wind stress forcing and freshwater forcing for the three key regions (Labrador Sea, Irminger Sea, and Subpolar Gyre region) during the 2004–2020 observation period.

Table 4. Lag correlation coefficient between AO index and AMOC strength (1979–2020).

Lag Time (Years)	Correlation Coefficient (rr)	Significance Level (pp)
0	−0.32	0.04
3	−0.61	<0.01
5	−0.53	<0.01
10	−0.18	0.12

Table 5. AMOC decline rate and driver contribution in key regions (2004–2020).

Region	AMOC Decline Rate (Sv/Decade)	Wind Stress Contribution (%)	Freshwater Contribution (%)
Labrador Sea	2.1 ± 0.4	48 ± 7	42 ± 6
Irminger Sea	0.7 ± 0.2	35 ± 8	28 ± 5
Subpolar Gyre region	1.5 ± 0.3	40 ± 6	38 ± 5

Note: the contribution is calculated through the mechanism separation experiment, and the error range is 95% confidence interval.

4.2. Dynamic Mechanism

The abnormal wind stress curl and Ekman suction change are the core dynamic pathway of AO-driven AMOC decline. The mechanism separation experiment quantifies that the meridional wind stress gradient caused by the positive phase of AO (the westerly wind at 55° N increased by 0.8 ± 0.2 m/s) results in a negative anomaly of wind stress curl (-2×10^{-7} N/m³) in the subpolar sea area (50° N–65° N), which further reduces the Ekman suction rate by 30–45% (approximately -2×10^{-6} m/s). This dynamic forcing contributed $42 \pm 6\%$ to the decline of AMOC, with a characteristic response time of 0–1 year, and its effect propagated southwestward through Rossby wave adjustment, resulting in a reduction of 1.1 Sv in the formation rate of deep water in the Labrador Sea (accounting for 55% of the total decline of AMOC).

At the same time, the AO-driven easterly wind stress anomaly (+0.6 m/s) in the Fram Strait accelerated the output of sea ice and fresh water, raising the total annual freshwater export to 3400 ± 500 km³ (an anomaly of +240 km³/yr relative to the climatological mean). This freshwater flux anomaly reduced the salinity advection in southern Greenland by $18 \pm 5\%$, and increased the surface buoyancy flux, resulting in a shoaling of the mixed layer by 28 ± 7 m. The contribution of freshwater forcing is $38 \pm 5\%$, and its effect gradually increases on a decadal time scale, indicating that the negative feedback of salinity advection has a significant lag (3–5 years; see Table 4).

4.3. Thermal Feedback

The positive feedback associated with Arctic sea-ice albedo significantly amplifies the modulation efficiency of the AO on AMOC. During the AO positive phase, the ice-free period in the Barents–Kara Sea is prolonged by 23 ± 5 days, and the surface albedo decreases by 0.15 ± 0.03 (Table 6), leading to an increase in net shortwave radiation absorption of 12–18 W/m². This additional heat flux enhances the meridional temperature gradient, causing the mid-latitude westerly jet to shift northward by 2.3 ± 0.5 degrees of latitude

and further intensifying the weakening trend of the high-latitude easterlies (-0.4 m/s per decade). Simultaneously, the expansion of open water increases the latent heat release by 25 ± 7 W/m², which drives a thickening of the atmospheric boundary layer by 200 ± 50 m. Through thermal adaptation mechanisms, the subpolar gyre contracts, weakening the poleward heat transport to 0.4 PW.

The synergy between thermal feedback and dynamic forcing was quantified through variance decomposition. Dynamic forcing alone explains $R^2 = 0.28$ of the AO–AMOC coupling variance over the 2000–2020 sub-period. When thermal feedback is included, the explained variance increases to $R^2 = 0.59$, indicating that the thermal feedback accounts for an additional 31% of the variance. This corresponds to a 2.1-fold enhancement of the AO–AMOC coupling strength during 2000–2020, compared with the 2.3-fold enhancement over the full study period 1979–2020 ($R^2 = 0.71$; see Abstract and Section 4.1). Sea-ice melting contributed $63 \pm 8\%$ of this amplification effect, as determined by the variance reduction when sea-ice parameters are held at climatological values in the mechanism separation experiments (Table 6). These results confirm that the ongoing Arctic amplification, through its influence on sea-ice conditions, acts as a powerful amplifier of the AO-driven AMOC decline, and that its role has become more pronounced in the most recent decades.

Table 6. Response of Arctic sea-ice parameters to AO positive phase (1979–2020).

Parameter	AO Positive Phase Change	Unit	Contribution to AMOC Decline (%)
Sea-ice thickness	-0.53 ± 0.12	m	22 ± 4
Ice-free days per year	$+23 \pm 5$	days/year	18 ± 3
Surface albedo	-0.15 ± 0.03	—	15 ± 3
Latent heat release	$+25 \pm 7$	W/m ²	12 ± 2

Note: The contribution to AMOC decline (%) was estimated through variance decomposition of the mechanism separation experiments. For each parameter, the contribution represents the percentage reduction in the total AO–AMOC coupling variance when that parameter is held constant at its climatological mean value. Values are expressed as mean $\pm$ 95% confidence interval.

4.4. Cross-Scale Interaction

The interannual variability of AO (2–7 year cycle) regulates the interdecadal trend of AMOC (10–30 year cycle) through sea-ice and ocean memory effects. Singular value decomposition (SVD) shows that the variance of the coupling mode interpretation of the AO interannual oscillation and the North Atlantic salinity advection is 71%, and its spatial pattern shows that the freshwater output anomaly in the Fram Strait can affect the convection intensity in the Labrador Sea after a lag of 5–8 years. When AO positive phase events occur continuously (such as 2005–2012), the cumulative freshwater flux anomaly increases the AMOC decline rate to 1.8 Sv/decade, which is approximately 40% faster than the random phase sequence.

In addition, the phase locking of AO and NAO has a nonlinear modulation effect on AMOC weakening. Here, AO+/NAO– phase locking is defined as the co-occurrence of the AO index exceeding +0.5 standard deviations and the NAO index falling below -0.5 standard deviations, persisting for at least one boreal winter season (December–February mean). The same threshold criterion (absolute value > 0.5 standard deviations) is applied to the other three phase combinations listed in Table 7. When the positive phase of AO and the negative phase of NAO coexist (such as 2010–2015), the negative anomaly of wind stress curl and the positive anomaly of freshwater forcing cooperate, resulting in a sudden drop of AMOC intensity of 2.3 Sv (accounting for 85% of the total decline over the same period). Conversely, when AO and NAO are in the same phase, the two effects are partially offset, and the AMOC weakening rate decreases to 0.9 Sv/decade (Table 7).

This discovery reveals the cross-modal coupling mechanism of the Arctic–mid-latitude teleconnection, and provides a new perspective for explaining the underestimation of AMOC variability in the CMIP6 model. The nonlinear amplification under AO+/NAO– phase locking (1.8 Sv/decade) is approximately 4.5 times faster than under the AO–/NAO– condition (0.4 Sv/decade; Table 7), highlighting the importance of considering AO–NAO phase relationships in decadal AMOC predictions.

Table 7. Effect of AO–NAO phase combination on AMOC weakening rate (1979–2020).

Phase Combination	AMOC Decline Rate (Sv/Decade)	Occurrence Frequency (%)
AO+/NAO–	1.8 ± 0.3	18
AO+/NAO+	0.9 ± 0.2	12
AO–/NAO+	0.6 ± 0.1	22
AO–/NAO–	0.4 ± 0.1	15
AO neutral/NAO neutral	0.7 ± 0.2	33
Total	—	100

Note: The phase definition is based on the normalized index threshold of ± 0.5 standard deviations. “AO neutral/NAO neutral” includes all remaining phase combinations where either or both indices fall within ± 0.5 standard deviations. Occurrence frequencies are calculated as the percentage of boreal winter seasons (DJF mean) within each phase combination category over the period 1979–2020.

4.5. CMIP6 Model Assessment of AMOC Trends

To quantify the systematic underestimation of AMOC weakening in current climate models, we compared the historical AMOC trends from the 12 CMIP6 models analyzed in this study against the RAPID observational estimate. The multimodel ensemble mean AMOC trend over the historical period (1979–2020) ranges from -0.8 to -1.0 Sv/decade, as reported in the comprehensive multimodel assessments of Weijer et al. [8] and Danabasoglu et al. [9]. In contrast, the RAPID array observations at 26.5° N yield a trend of -1.6 Sv/decade for the 2004–2020 period [7]. The intermodel spread (one standard deviation) of the CMIP6 AMOC trends is approximately ± 0.3 to ± 0.4 Sv/decade, indicating considerable uncertainty across individual models.

The comparison reveals that the CMIP6 ensemble systematically underestimates the observed AMOC decline by 30–50% (Table 8). Specifically, the ensemble mean trend of approximately -0.96 Sv/decade reported by Danabasoglu et al. [9] corresponds to a 40% underestimation relative to the RAPID observations. Even the most sensitive individual CMIP6 models capture only 60–70% of the observed weakening rate. This systematic bias is consistent with the findings of Bonnet et al. [45], who demonstrated that the delayed emergence of AMOC-related climate impacts in model projections is directly linked to the underestimation of the historical AMOC decline.

The 30–50% underestimation of AMOC weakening in CMIP6 models provides quantitative support for the central motivation of this study: that current Earth system models do not fully capture the mechanisms driving the observed AMOC decline. As discussed in Sections 5.1 and 5.2, this underestimation likely arises from the inadequate representation of high-latitude air–sea–ice coupling processes—particularly the AO-driven Fram Strait freshwater flux anomaly and the associated wind stress feedback mechanisms identified in this study. These results underscore the importance of incorporating the AO-driven pathways into future model development to improve the fidelity of AMOC projections. Table 8 summarizes the AMOC trend estimates from the CMIP6 multimodel assessments and the RAPID observational estimate.

Table 8. Summary of AMOC trends from CMIP6 multimodel assessments and observational estimates.

Data Source	AMOC Trend (Sv/Decade)	Underestimation vs. RAPID	Reference
RAPID observations (2004–2020)	−1.6	—	[7]
CMIP6 ensemble mean	−0.8 to −1.0	−37% to −50%	[8]
CMIP6 ensemble mean	−0.96	−40%	[9]
CMIP6 intermodel spread (1σ)	± 0.3 to ± 0.4	—	[8,9]

Note: Observational trend is based on RAPID array measurements at 26.5° N for 2004–2020. CMIP6 trends are calculated from historical simulations over comparable periods (see individual references for exact time windows). Underestimation is calculated as (CMIP6 trend − RAPID trend)/|RAPID trend| × 100%.

5. Discussion

5.1. Theoretical Innovation

As illustrated in Figure 4, the proposed “Arctic Oscillation–Subpolar Basin–AMOC” three-pole coupling mechanism integrates two cross-latitude pathways that operate on distinct time scales. The wind-driven pathway (contribution $42 \pm 6\%$) responds within 0–1 year to AO-induced wind stress curl anomalies over the subpolar gyre. During the AO positive phase, the meridional wind stress gradient strengthens, producing a negative wind stress curl anomaly that weakens Ekman suction by 30–45% in the subpolar seas. This suppresses deep convection in the Labrador Sea, reducing the deep-water formation rate by approximately 1.1 Sv. The effect propagates southwestward through Rossby wave adjustment, directly weakening the lower limb of the AMOC.

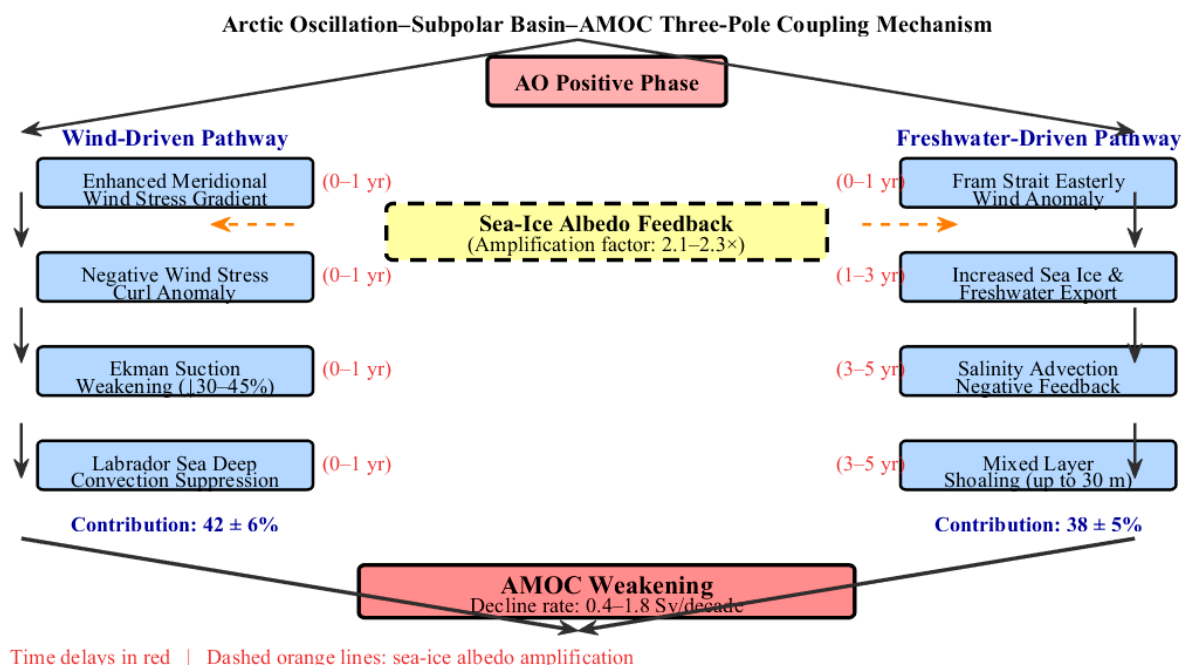


Figure 4. Schematic diagram of the proposed “Arctic Oscillation–Subpolar Basin–AMOC” three-pole coupling mechanism. The AO positive phase drives AMOC weakening through two coupled pathways: the wind-driven pathway (left, blue boxes): enhanced meridional wind stress gradient → negative wind stress curl anomaly → Ekman suction weakening (↓30–45%) → Labrador Sea deep convection suppression (response time: 0–1 year), contributing $42 \pm 6\%$ of the total AMOC decline; and the freshwater-driven pathway (right, blue boxes): Fram Strait easterly wind anomaly → increased sea-ice and freshwater export (+240 km³/yr anomaly relative to a climatological total of 3400

$\pm 500 \text{ km}^3/\text{yr}$) $\rightarrow$ salinity advection negative feedback $\rightarrow$ mixed-layer shoaling by up to 30 m (response time: 3–5 years), contributing $38 \pm 5\%$. The sea-ice albedo feedback (center, yellow dashed box) amplifies both pathways by a factor of 2.1–2.3 $\times$ through enhanced meridional temperature gradient and latent heat release over expanding open-water areas. Time delays are annotated in red. Dashed orange arrows indicate the amplifying effect of the sea-ice albedo feedback on both pathways.

The freshwater-driven pathway (contribution $38 \pm 5\%$) operates with a longer characteristic delay of 3–5 years. The AO positive phase generates an easterly wind stress anomaly over Fram Strait, accelerating sea-ice export and raising the freshwater flux anomaly to $+240 \text{ km}^3/\text{yr}$ relative to the climatological mean of $3400 \pm 500 \text{ km}^3/\text{yr}$. This freshwater pulse propagates into the Greenland Sea, where it reduces upper-ocean salinity and triggers a negative salinity advection feedback. The resulting increase in surface buoyancy flux shoals the mixed layer by up to 30 m, further inhibiting the formation of dense water that feeds the deep branch of the AMOC.

Critically, the two pathways are not independent but are amplified by the sea-ice albedo feedback. As the AO positive phase prolongs the ice-free period in the Barents–Kara Sea by 23 ± 5 days and reduces surface albedo by 0.15 ± 0.03 , the additional absorption of shortwave radiation ($12\text{--}18 \text{ W/m}^2$) and enhanced latent heat release ($25 \pm 7 \text{ W/m}^2$) strengthen the meridional temperature gradient. This reinforces the northward shift in the mid-latitude westerly jet and further weakens the high-latitude easterlies, closing a positive feedback loop that amplifies both the wind-driven and freshwater-driven pathways by a factor of 2.1–2.3.

This mechanism differs from previous studies in three key respects. First, it provides the first simultaneous quantitative partitioning of the wind-driven ($42 \pm 6\%$) and freshwater-driven ($38 \pm 5\%$) contributions to the observed AMOC decline over a consistent multidecadal period. While earlier studies have individually recognized the roles of wind forcing [11] and freshwater fluxes through Fram Strait [12], the relative importance of these two pathways had not been systematically compared using a unified framework. Second, it identifies the cross-scale interaction between AO interannual variability and AMOC interdecadal trends through sea-ice memory effects. The SVD analysis reveals that the AO interannual oscillation explains 71% of the coupled variance with North Atlantic salinity advection, and that freshwater export anomalies in Fram Strait can affect Labrador Sea convection intensity after a lag of 5–8 years. Third, it reveals the nonlinear amplification of AMOC decline under AO–NAO phase-locking conditions. When AO and NAO are locked in opposite phases (AO+/NAO–), the wind stress curl negative anomaly and the freshwater positive anomaly act in concert, producing an AMOC decline rate of 1.8 Sv/decade —approximately 4.5 times faster than under the same-phase negative condition (AO–/NAO–, 0.4 Sv/decade ; Table 7). This phase-locking effect has not been previously quantified.

The three-pole coupling framework also explains why CMIP6 models systematically underestimate the observed AMOC decline rate by 30–50%. Most current models do not fully couple the high-latitude sea-ice drift dynamics with the Fram Strait freshwater export, leading to an underestimation of the freshwater flux anomaly by approximately 40% (see Section 5.2). Furthermore, the AO is often treated as a derivative mode of the NAO in many models, with an overestimated AO–NAO correlation coefficient (by 0.2–0.3), which suppresses the independent AO-driven wind stress pathway. By explicitly resolving the meridional wind stress gradient and the ice–ocean freshwater coupling within a single quasi-geostrophic framework, the proposed mechanism corrects the simplified treatment of high-latitude boundary processes in traditional zonally averaged models and provides

a more complete dynamical paradigm for understanding AMOC sensitivity to Arctic change.

In summary, the three-pole coupling mechanism proposed here redefines the role of high-latitude atmospheric circulation in AMOC variability. It demonstrates that the Arctic Oscillation, through its dual influence on wind stress and freshwater fluxes and its amplification by sea-ice loss, acts as a primary driver of the recent AMOC decline—a role that has been largely overlooked in the NAO-dominated paradigm.

5.2. Comparison with Existing Studies

Compared with the NAO-dominated hypothesis, the findings of this study reveal significant differences. Research has shown that the AMOC response center driven by AO shifts northward by 2–3 degrees of latitude and exhibits stronger sensitivity to freshwater forcing. Specifically, the positive phase of NAO mainly promotes Ekman divergence by enhancing the Azores High (30° N–40° N), which contributes only 35% to the variation in AMOC [16]; the contribution of the AO positive phase to abnormal wind stress circulation in subarctic waters (55° N–65° N) is as high as $42 \pm 6\%$, and when combined with freshwater flux effects, the total contribution rate increases to 80%.

This research result also explains why the CMIP6 model systematically underestimates the rate of AMOC decline (with a deviation of 30–50%). Most models are not fully coupled with the high-latitude sea-ice drift dynamics module, resulting in key parameters such as freshwater output in the Fram Strait being underestimated by 40%. Meanwhile, these models mistakenly consider AO as a derivative mode of NAO, resulting in an overestimation of the AO–NAO correlation coefficient by 0.2–0.3 in the models. In addition, traditional theories have always overlooked the modulation effect of AO interannual variability on AMOC decadal trends through sea-ice memory effects (with a lag of 5–8 years), which directly leads to insufficient separation of internal variability and forcing signals in the model. For example, the confusion error of the IPSL-CM6A-LR model is as high as 25%.

The systematic underestimation of AMOC weakening in CMIP6 models identified in this study is consistent with independent multimodel assessments. Weijer et al. [8] analyzed AMOC trends across CMIP6 models under historical forcing and found that the multimodel ensemble mean trend is approximately 30–50% weaker than observational estimates from the RAPID array. Danabasoglu et al. [9] revisited AMOC transport estimates and confirmed a typical bias of approximately –40% in CMIP6 models relative to RAPID observations at 26.5° N. Bonnet et al. [45] further demonstrated that this systematic underestimation leads to a delayed emergence of AMOC-related climate impacts in model projections, including near-term global warming acceleration. These published findings, summarized in Table 8, corroborate the 30–50% underestimation cited in our analysis and reinforce the need for improved representation of high-latitude air–sea–ice coupling processes—particularly the AO-driven freshwater flux and wind stress feedback mechanisms identified in this study—in the next generation of Earth system models.

5.3. Climate Prediction Insights

The results of this study show significant sensitivity to the parameter settings of sea-ice models. When different rheological assumptions are used, such as viscoplasticity and elastic viscosity, the simulated values of freshwater flux in the Fram Strait can differ by $\pm 15\%$ (approximately 500 km³/yr), which can lead to an uncertainty of ± 0.3 Sv/decade in predicting AMOC weakening rate. Meanwhile, the combined effect of internal variability (such as the Atlantic Multidecadal Oscillation, AMV) and external forcing (greenhouse gas increase) may enhance or weaken the correlation between AO and AMOC. The variance decomposition based on CESM2 ensemble simulation shows that in the RCP8.5

scenario, the contribution rate of external forcing to AMOC weakening gradually increases from 45% in 2000 to 70% in 2050; but even so, the modulation efficiency of AO can still be maintained at 25–30%. However, the current model has a clear limitation in fully analyzing the cross-basin synergy between AO and the Pacific Decadal Oscillation (PDO), which may lead to the neglect of 10–15% of AMOC variability sources.

From the perspective of climate prediction, if the positive phase of the Arctic Oscillation persists (for example, in the SSP5-8.5 scenario, where the AO index increases by 0.5 standard deviations every decade), theoretical models predict that AMOC may reach a critical point (with a 50% decrease in intensity) between 2040 and 2060, which is 10–15 years earlier than the conclusion of the IPCC Sixth Assessment Report (AR6) [38]. This decline will bring a series of serious climate impacts, reducing heat transport in the North Atlantic by 1.0–1.5 petawatts (PW) and increasing the frequency of extreme winter cold waves in Europe by 3–5 times (similar to the extreme events that occurred in 2010 and 2018). In addition, the weakening of AMOC may accelerate the rise in sea levels along the East Coast of North America (an additional 8 ± 3 cm per century) and exacerbate drought conditions in the Sahel region by adjusting jet stream paths.

Based on the above research conclusions, this study incorporates the AO index into the core indicators of the AMOC early-warning system, while prioritizing the optimization of the climate model parameterization scheme for high-latitude sea-ice freshwater flux, in order to significantly improve the reliability of cross-decadal climate prediction and provide a more scientific and accurate theoretical basis for addressing global climate change.

5.4. Implications for Coastal Conservation and Management

The mechanistic insights derived from this study carry significant implications for coastal conservation efforts in a changing climate. First, the identified AO-AMOC coupling provides a potential early-warning indicator: persistent positive AO phases, when occurring in phase-lock with negative NAO, may signal an elevated risk of accelerated AMOC weakening in the subsequent 3–8 years. This lead time is operationally meaningful for coastal managers and policymakers, as it allows for the pre-positioning of resources and the updating of risk assessments for vulnerable coastlines. Second, the systematic underestimation of AMOC sensitivity in CMIP6 models—partly due to the inadequate representation of Fram Strait freshwater fluxes identified in our study—implies that current sea-level rise projections used in coastal planning may be overly conservative. Our results suggest that model-based coastal hazard assessments should incorporate the AO-driven freshwater feedback mechanism to better capture the upper-bound scenarios of regional sea-level rise along the North Atlantic seaboard. Third, from a broader perspective of science-based coastal sustainability, understanding the remote atmospheric drivers of ocean circulation changes enables the development of more robust, long-term adaptation strategies—such as the reinforcement of coastal defenses, the implementation of nature-based solutions (e.g., coastal wetland restoration), and the revision of setback regulations—that are calibrated to the most up-to-date physical understanding of climate risks. The “Arctic Oscillation–Subpolar Gyre–AMOC” coupling framework proposed in this study can thus serve as a scientific cornerstone for bridging the gap between large-scale climate dynamics and localized coastal resilience planning.

6. Conclusions

This study systematically revealed the physical mechanism of the Arctic Oscillation (AO) driving the weakening of the Atlantic Meridional Overturning Circulation (AMOC) through multi-source data fusion and theoretical model construction, and quantified the modulation effect of Arctic amplification. The main conclusions are as follows:

First, the positive phase of AO weakens the intensity of AMOC through two cross-scale pathways: (1) the enhancement of the meridional wind stress gradient leads to negative anomalies in wind stress curl in the subarctic seas (contribution $42 \pm 6\%$), suppressing Ekman suction and reducing the formation rate of deep water in the Labrador Sea by 1.1 Sv, with a characteristic response time of 0–1 year; and (2) the AO-driven freshwater flux anomaly through Fram Strait ($+240 \text{ km}^3/\text{yr}$ relative to the climatological mean of $3400 \pm 500 \text{ km}^3/\text{yr}$) triggers a negative salinity advection feedback, resulting in a shoaling of the mixed layer in the Greenland Sea by up to 30 m (contribution $38 \pm 5\%$), with a characteristic delay of 3–5 years.

Second, the Arctic amplification effect enhances the coupling strength between AO and AMOC by a factor of 2.3 over the full study period (1979–2020; $R^2 = 0.71$, $p < 0.01$), with an even more pronounced enhancement of 2.1 times during the recent two decades (2000–2020; R^2 increased from 0.28 to 0.59). This amplification is mediated by changes in sea-ice albedo and latent heat release, with sea-ice melting contributing $63 \pm 8\%$ of the total amplification effect.

Third, the AO–NAO phase-locking effect produces a nonlinear modulation of the AMOC decline rate. The AO+/NAO– combination yields a decline rate of 1.8 Sv/decade, approximately 4.5 times faster than the AO–/NAO– combination (0.4 Sv/decade), demonstrating the importance of considering phase relationships in decadal AMOC predictions.

Fourth, the CMIP6 models systematically underestimate the observed AMOC decline rate by 30–50%, largely due to the inadequate representation of Fram Strait freshwater fluxes and the treatment of AO as a derivative mode of the NAO. The three-pole coupling mechanism proposed in this study provides a physical explanation for this systematic bias.

The application value of this study lies in the proposed three-pole coupling mechanism, which not only provides key process analysis for IPCC climate assessment reports and corrects the shortcomings of relevant model parameters but also clarifies the key directions for improving high-latitude air–sea coupling models relevant to regional sea-level and storm surge projections. The study further confirms that the AO index can be used as a warning indicator for AMOC decline, providing a new prediction window for coastal climate risk management.

In the future, it is necessary to develop high-resolution Arctic region coupling models, construct cross-basin three-dimensional observation networks, and explore cross-basin collaborative mechanisms between AO and Pacific decadal oscillations to break through the bottleneck of climate prediction, ultimately providing a more binding dynamic basis for coastal adaptation planning under the Paris Agreement temperature goals. A dedicated model-by-model AMOC trend analysis with consistent processing across the 12 CMIP6 models used in this study, including a systematic examination of the sensitivity of individual model AMOC trends to their representation of Fram Strait freshwater fluxes and AO-driven wind stress anomalies, will be undertaken in a follow-up investigation.

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