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Geophysical Research Letters

RESEARCH LETTER
10.1029/2026GL124187

Tropical Eastern Pacific Decadal Variability Drives Atlantic Multidecadal Variability: Two Distinct Pathways

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Key Points:

- A pacemaker experiment reveals two distinct Atlantic Multidecadal Variability responses to tropical eastern Pacific decadal forcing
- The Atlantic responses include an immediate tripole via surface heat flux and a decadal-scale delayed monopole via overturning circulation
- The tropical eastern Pacific forcing accounts for a substantial fraction of the historical internal Atlantic Multidecadal Variability

Supporting Information:

Supporting Information may be found in the online version of this article.

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Citation:

Yang, J.-C., Lin, X., Li, J., Zhang, Y., Xu, W.-Y., & Lv, Z. (2026). Tropical eastern Pacific decadal variability drives Atlantic multidecadal variability: Two distinct pathways. *Geophysical Research Letters*, *53*, e2026GL124187. https://doi.org/10.1029/2026GL124187

Received 7 MAY 2026
Accepted 9 AUG 2026

Abstract

Using a tropical Pacific pacemaker experiment, we demonstrate that the tropical eastern Pacific (TEP) decadal sea surface temperature (SST) variability influences the Atlantic Multidecadal Variability (AMV) through two pathways. The TEP induces a North Atlantic SST tripole response that peaks 1 year later, primarily via modulation of net surface heat flux. The subpolar SST anomalies associated with this tripole response subsequently influence the strength of North Atlantic deep water formation and the Atlantic Meridional Overturning Circulation, ultimately leading to a monopole SST response in the North Atlantic about one decade later. The tripole and monopole responses account for 40% and 59% of the TEP-forced AMV signal, respectively. The TEP-forced AMV explains approximately 30% of the variance in the internal AMV, thus substantially modulating the phase transitions and intensity of the historical AMV. This identified TEP forcing should be carefully considered to improve simulations and predictions of the AMV.

Plain Language Summary

The Atlantic Multidecadal Variability (AMV) is a major climate mode that affects global climate and extreme weather. Understanding how the AMV forms is key to predicting it and its worldwide impacts. Using a climate model experiment, we show that the tropical eastern Pacific (TEP) decadal sea surface temperature variability is a crucial forcing of the AMV. We find a two-step AMV response to the TEP forcing involving rapid surface changes and slow ocean circulation adjustments. We highlight the importance of the TEP-forced AMV in the historical AMV evolution, as the TEP-forced AMV explains nearly one-third of the variance in the climate internally generated component of the AMV. Our study underscores the necessity to improve the simulation of the TEP forcing for better predicting the AMV.

1. Introduction

Atlantic sea surface temperature (SST) exhibits prominent multidecadal variability, termed the Atlantic Multidecadal Variability (AMV; Kerr, 2000; Sutton et al., 2018). The AMV shows broad climate impacts throughout the Atlantic Ocean and adjacent continents (Knight et al., 2006), modulating Atlantic hurricanes (Goldenberg et al., 2001; Klotzbach & Gray, 2008), Sahel rainfall (Folland et al., 1986; Martin & Thorncroft, 2014), and temperature and precipitation in Europe and North America (O'Reilly et al., 2017; Sutton & Hodson, 2005). The AMV also influences climate outside Atlantic-rim regions via atmospheric and oceanic teleconnections (Ruprich-Robert et al., 2017; Zhang et al., 2019). Due to the extensive impacts of the AMV, understanding its formation mechanisms is crucial for climate prediction and risk management. While Atlantic internal physics—including both externally forced and internal processes—contribute to the AMV (e.g., Booth et al., 2012; Clement et al., 2015; Delworth et al., 2017; Fang et al., 2021; He et al., 2023; Mann et al., 2021; Qin et al., 2020; Sun et al., 2015; Zhang et al., 2019), tropical eastern Pacific (TEP) SST variability also acts as an important modulator.

On interannual timescales, El Niño-Southern Oscillation (ENSO) could influence extratropical North Atlantic SST by affecting the North Atlantic Oscillation (NAO; Hurrell, 1995), through tropospheric and stratospheric teleconnections (Ineson & Scaife, 2009; Jiménez-Esteve & Domeisen, 2018; Scaife et al., 2024), albeit with non-stationarity (Rodríguez-Fonseca et al., 2016). ENSO also modulates tropical North Atlantic SST through generating tropical and extratropical atmospheric waves and circulation changes (Alexander et al., 2002; Cai et al., 2019; Klein et al., 1999; Lee et al., 2008). On decadal timescales, however, the focus of this study, observations and model simulations detect lead correlations between the Interdecadal Pacific Oscillation (IPO; Power et al., 1999) and the AMV (Deng et al., 2024; Meehl et al., 2021; Nigam et al., 2020; Xu et al., 2025), implying ENSO-like decadal variability in the TEP as a potential forcing of the AMV. Previous studies mainly investigated tropical mechanisms in the IPO-AMV linkage, including the adjustments of the Walker Circulation (Hong et al., 2022; Meehl et al., 2021) and water vapor-longwave radiation-SST positive feedback (Yao et al., 2022). Yet whether ENSO-like decadal variability (associated with the IPO) in the TEP influences the AMV via extratropical processes remains unclear.

Here, by using a tropical Pacific pacemaker experiment, we reveal that on decadal timescales, TEP SST is an important modulator of the historical AMV by influencing both tropical and previously ignored extratropical processes in the North Atlantic. Specifically, the extratropical processes on the decadal timescales share similar NAO-like teleconnections with those of interannual ENSO. Two distinct pathways, respectively associated with surface heat flux and Atlantic Meridional Overturning Circulation (AMOC), are discovered.

2. Data and Methods

2.1. Data

The observed TEP-forced AMV cannot be directly extracted from observations due to (a) limited observational records for decadal-to-multidecadal variability studies, and (b) AMV-TEP interactaions complicating correlation interpretations (Xu et al., 2025). To address this, we employ a 20-member tropical Pacific pacemaker ensemble based on the Community Earth System Model version 1 (CESM1 POGA; Deser, Guo, & Lehner, 2017) with experimental design following Kosaka and Xie (2013). Each member features identical radiative forcing (historical: 1920–2005; Representative Concentration Pathway 8.5: 2006–2013) and slightly different initial conditions. Critically, TEP SST anomalies (180°–American coast, 15°S–15°N; ±5° buffer) are nudged to the observations from the Extended Reconstructed Sea Surface Temperature version 3b (ERSSTv3b; Smith et al., 2008), while other regions remain fully coupled. We complement this with a 40-member fully coupled CESM1 Large Ensemble (LE; Kay et al., 2015) to isolate radiatively forced signals (1920–2013). Model validation confirms that: The LE ensemble well simulates the AMV despite multidecadal variance underestimation (Kim et al., 2018); the POGA ensemble reliably captures observed TEP-forced teleconnections (Deser et al., 2017a, 2017b; Meehl et al., 2021; Schneider & Deser, 2018; Yang et al., 2020; Zhang et al., 2018). The ensemble sizes of the two experiments are sufficient for robust AMV signal isolation (Text S1 in Supporting Information S1). Finally, observational SST data are obtained from the ERSSTv3b to assess the contribution of the model-estimated TEP-forced component to the observed AMV (Section 2.2).

2.2. Definitions and Methods

In this paper, anomalies are computed as annual-mean (January–December) data minus the long-term mean, and then decadal anomalies are derived via a 10-year Lanczos low-pass filter. We define the model-estimated TEP-forced component as the difference between the POGA ensemble mean and LE ensemble mean. This represents internal climate variability by removing radiatively forced signals (Deser, Guo, & Lehner, 2017; Yang et al., 2020).

We define several indices:

1. Tropical Pacific index (TPI): Decadal SST anomalies averaged in the TEP (180°W–90°W, 15°S–15°N), representing TEP decadal SST variability (Wang, Xie, et al., 2017).
2. AMV_{TEP}: Model-estimated TEP-forced decadal SST anomalies averaged in the North Atlantic (75°W–0°W, 0°N–70°N), tracking the TEP-forced AMV.
3. AMV_{OBS}: Observed AMV (North Atlantic average) from linearly detrended ERSSTv3b SST.
4. AMV_{LE}: Radiatively forced AMV computed as linearly detrended LE ensemble-mean AMV.

For the POGA ensemble, internal SST anomalies are computed as the difference between each POGA member and the LE ensemble mean. For observations, the North Atlantic internal decadal SST anomalies are derived by subtracting $k_0 \times AMV_{LE}$ from the detrended observed decadal SST anomalies, where the adjustment coefficient k_0 is the regression coefficient of the detrended observed decadal SST anomalies onto the AMV_{LE}. Significance of linear correlation and regression analyses is tested via non-parametric methods (Text S2 in Supporting Information S1).

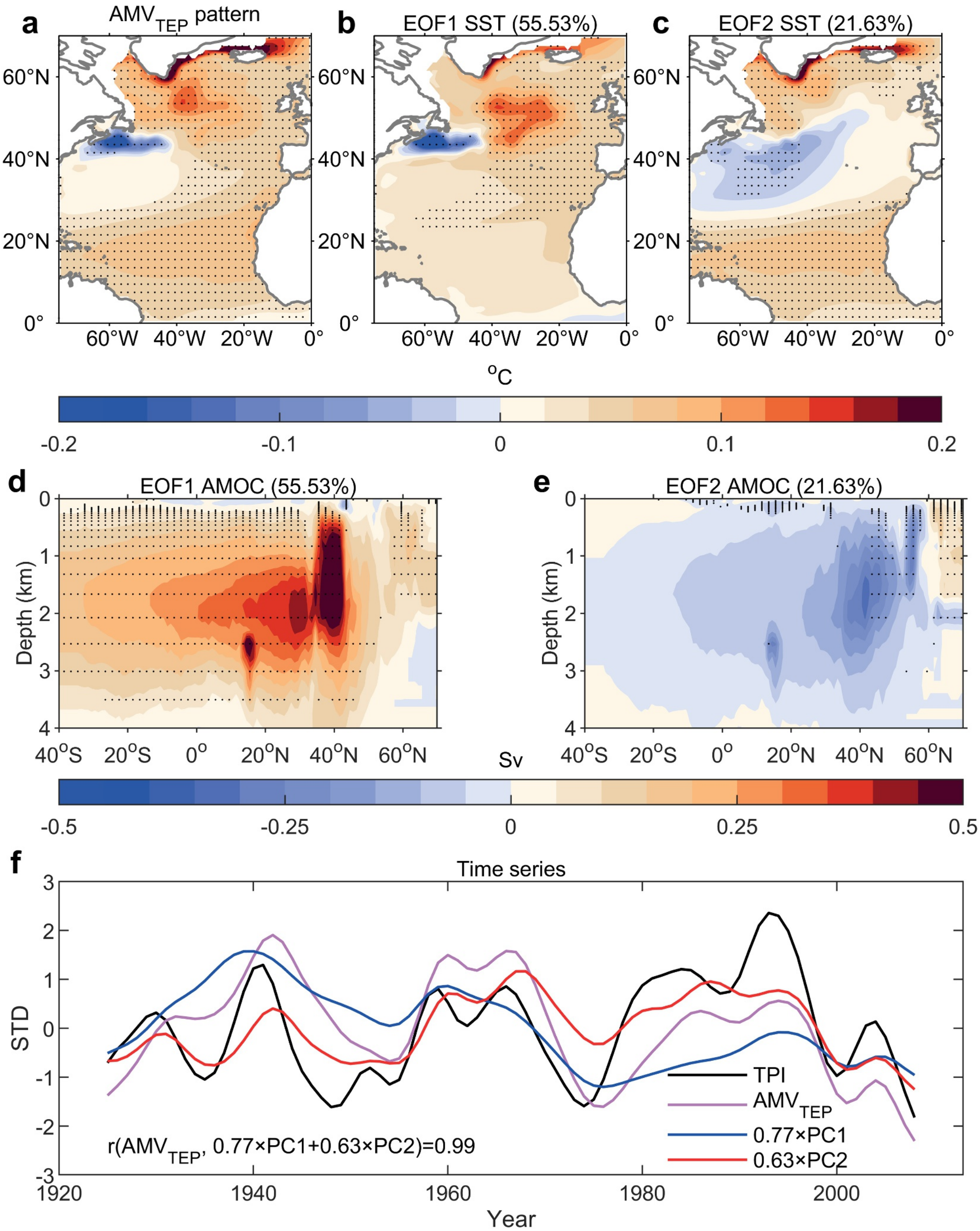

Figure 1. The two leading joint EOF modes and their PCs of North Atlantic SST and AMOC decadal variability. (a) Decadal SST anomalies regressed onto the AMV$_{TEP}$. Decadal SST (b) and AMOC (d) anomalies regressed onto the PC1. In panels (c, e) Same as (b, d) but for the PC2. Stipplings in (a–e) denote regression coefficients exceeding the 90% confidence level. (f) Time series of the TPI (black), AMV$_{TEP}$ (purple), and the normalized PC1 (blue) and PC2 (red) multiplied by their respective correlations with the AMV$_{TEP}$.

3. Results

3.1. Two Leading Modes of the TEP-Forced AMV

The AMV$_{TEP}$ generally exhibits a tripole pattern, with positive SST anomalies in the subpolar and tropical North Atlantic and negative SST anomalies near 65°W, 40°N (Figure 1a). To further decompose the AMV$_{TEP}$, we apply a joint Empirical Orthogonal Function (EOF) method to decadal anomalies of North Atlantic SST and Atlantic meridional overturning streamfunction in the model-estimated TEP-forced component. Both variables are normalized by their respective mean standard deviations before analysis. We include the overturning streamfunction to better isolate SST variations linked to AMOC changes, following Latif et al. (2022).

The EOF1 displays a monopole SST pattern with positive values across most of the North Atlantic, except near 65°W, 40°N (Figure 1b). Its signal remains largely within the extratropics. The EOF1 coincides with a basin-scale positive AMOC anomaly peaking near 40°N (Figure 1d). The EOF2 shows a tripole SST pattern with positive anomalies in the subpolar and tropical North Atlantic, and negative anomalies in midlatitudes (Figure 1c). The associated AMOC anomalies are negative, weaker than in the EOF1 (Figure 1e), and exhibit a poleward-shifted center with a shallow cell near 55°N. The AMV_{TEP} pattern (Figure 1a) resembles a combination of the two EOF SST patterns (Figures 1b and 1c).

The first principle component (PC1) exhibits strong multidecadal variability, while the PC2 varies on decadal to multidecadal timescales (Figure 1f). The AMV_{TEP} is largely explained by both the PCs (PC1: 59% variance; PC2: 40%). Notably, the extratropical AMV_{TEP} (averaging north of 25°N) is dominated by the PC1 ($r = 0.90$), while the tropical AMV_{TEP} (averaging south of 25°N) is dominated by the PC2 ($r = 0.86$). The 25°N boundary is chosen for illustration, but the result is insensitive to this choice. This spatial partitioning aligns with the structures of the two modes: the monopole response peaks in the extratropical latitudes (Figure 1b), whereas the tripole response shows comparable subpolar positive and midlatitude negative anomalies that cancel each other (Figure 1c). Consequently, regional contributions of the AMV_{TEP} primarily originate from the monopole in extratropics and the tripole in tropics. Thus, the two leading modes elucidate the mechanisms of the TEP-forced AMV. We will show that the TEP firstly forces the tripole SST pattern via atmospheric teleconnections and subsequently drives the monopole SST pattern through AMOC modulation.

3.2. Formation Mechanisms of the Tripole Response

The tripole SST pattern (EOF2) closely tracks TEP decadal SST variability, as evidenced by similar peaks/troughs in the PC2 and TPI (Figure 1f). However, the TPI leads the PC2 by 1 year, confirmed by lead-lag correlation (Figure 2a). To elucidate this relationship, we analyze TEP-forced atmospheric teleconnections and North Atlantic responses.

A positive TEP decadal SST anomaly excites tropospheric Rossby wave trains propagating from the tropical Pacific to North Atlantic (Figure 2b). The resultant strengthened Aleutian Low intensifies the polar vortex, inducing significant stratospheric geopotential height anomalies (Figure S1 in Supporting Information S1). These teleconnection patterns, resembling those raised by ENSO forcing (Ineson & Scaife, 2009; Jiménez-Esteve & Domeisen, 2018; Scaife et al., 2024), falicitate the formation of a sea level pressure anomaly resembling a negative NAO phase: low pressure over the midlatitude North Atlantic and high pressure to the northeast (Figure 2c). This NAO-like pattern decelerates surface westerlies and weakens upward turbulent heat flux (mostly sensible) in the subpolar North Atlantic (Figure 2c; Figure S2 in Supporting Information S1). In midlatitudes, TEP-forced shortwave and longwave radiation anomalies dominate the upper ocean cooling (Figure S2 in Supporting Information S1). However, because this NAO-like pattern shifts farther north than the original NAO, it cannot induce as strong a tropical North Atlantic response as the NAO (Sun et al., 2019). In addition to the extratropical NAO-like response, a positive TEP decadal SST anomaly also modulates the Walker Circulation and leads to descent anomalies in the equatorial Atlantic (contours in Figure 2d) that stabilize the boundary layer. Additionally, a warm TEP triggers tropospheric Kelvin waves propagating into the tropical North Atlantic, increasing tropospheric temperature (shadings in Figure 2d). Both processes can increase atmospheric water vapor content and warm the SST via enhanced downward longwave radiation (Figure 2e). Collectively, these processes force a tripole net surface heat flux anomaly, inducing a similar tripole SST response (Figure 2f), despite minor oceanic modulation. These responses in the POGA experiment align with those in idealized IPO experiments (Meehl et al., 2021; Yao et al., 2022).

The processes that lead to the delay of the tripole response (Figure 2a) need to be further clarified. In the extratropics, we attribute this lag partly to the so-called reemergence processes associated with mixed layer seasonal variations (Alexander & Deser, 1995; Deser et al., 2003; Watanabe & Kimoto, 2000). In the North Atlantic, mixed layer depth (MLD) peaks in boreal winter–spring (Figure S3 in Supporting Information S1), enabling SST signals to penetrate into deeper layers. The trapped signals reemerge during the subsequent winter–spring, reinforcing surface responses. Such seasonally varying MLD, along with seasonal atmospheric

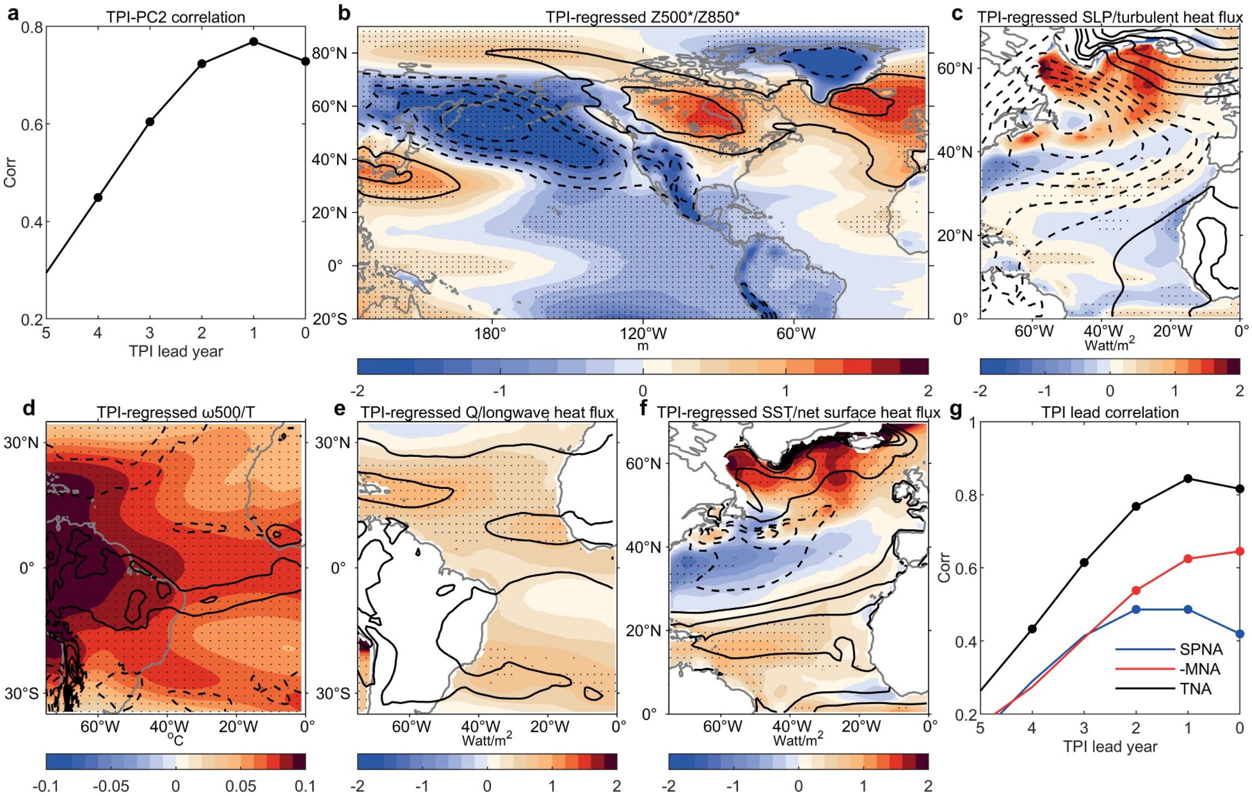

Figure 2. Mechanisms of the tripole response. (a) Lagged correlation between the TPI and PC2. Dots denote correlations exceeding the 90% confidence level. (b) TPI-regressed decadal 500-hPa (850-hPa) geopotential height anomalies with zonal mean removed are shown by contours (shadings). Contour interval is 1 m with solid for positive, dashed for negative, and zero line omitted. Stipplings denote regression coefficients shown by shadings exceeding the 90% confidence level. In panels (c) Same as (b) but for sea level pressure (contours; interval 0.02 hPa) and turbulent heat flux (shadings). In panels (d) Same as (b) but for vertical velocity in pressure coordinates at 500 hPa (contours; interval 5×10^{-4} Pa/s) and tropospheric mean temperature from 200 to 1000 hPa (shadings). (e) Same as (b) but for atmospheric water vapor content (contours; interval 0.1 kg/m^2) and longwave heat flux (shadings). In panels (f) Same as (b) but for SST (contours; interval 0.02°C) and net surface heat flux (shadings). (g) Lagged correlation between the TPI and decadal SST anomalies averaged over the subpolar North Atlantic (SPNA; 60°W–20°W, 55°N–65°N), midlatitude North Atlantic (MNA; 70°W–30°W, 35°N–45°N), and tropical North Atlantic (TNA; 75°W–10°W, 5°N–20°N). MNA correlations are multiplied by −1 for convenient comparison.

forcing, delays SST responses particularly in the subpolar North Atlantic (Figure 2g), with detailed modulation analyzed in Text S3, Figures S4a–S4c and S5 of Supporting Information S1. In the tropical North Atlantic, TEP-forced tropospheric Kelvin waves propagating into this region contribute to a seasonally delayed SST response (Text S3 and Figure S4d in Supporting Information S1).

3.3. Formation Mechanisms of the Monopole Response

The monopole response co-occurs with a basin-scale AMOC anomaly (Figures 1b and 1d). A positive AMOC anomaly enhances northward ocean heat transport, leading to upper ocean heat divergence in the Gulf Stream and convergence in the central subpolar North Atlantic (Zhang, 2008; Zhang & Zhang, 2015). This heat redistribution generates a positive monopole SST pattern (Figure 1b), consistent with Yan et al. (2018). We further identify the subpolar component of the tripole response as the key driver of AMOC changes.

The tripole SST response shallows the mixed layer—particularly east of the Labrador Sea deep water formation zone (Figure 3a). Decadal MLD anomalies in this region are primarily controlled by surface buoyancy fluxes (e.g., Danabasoglu et al., 2012; Delworth & Zeng, 2016). Here, surface warming (Figure 3b) outweighs freshening (Figure 3c) in driving negative surface buoyancy fluxes, thereby shoaling the mixed layer (Figure S6 in Supporting Information S1). This shoaling inhibits deep water formation, slowing the AMOC after a decadal-scale delay (e.g., Delworth & Zeng, 2016; Sun et al., 2015). To track the TEP-forced deep water formation and AMOC anomalies, we define a subpolar MLD index (40°W–55°W, 54°N–58°N, red box in Figure 3a) and an

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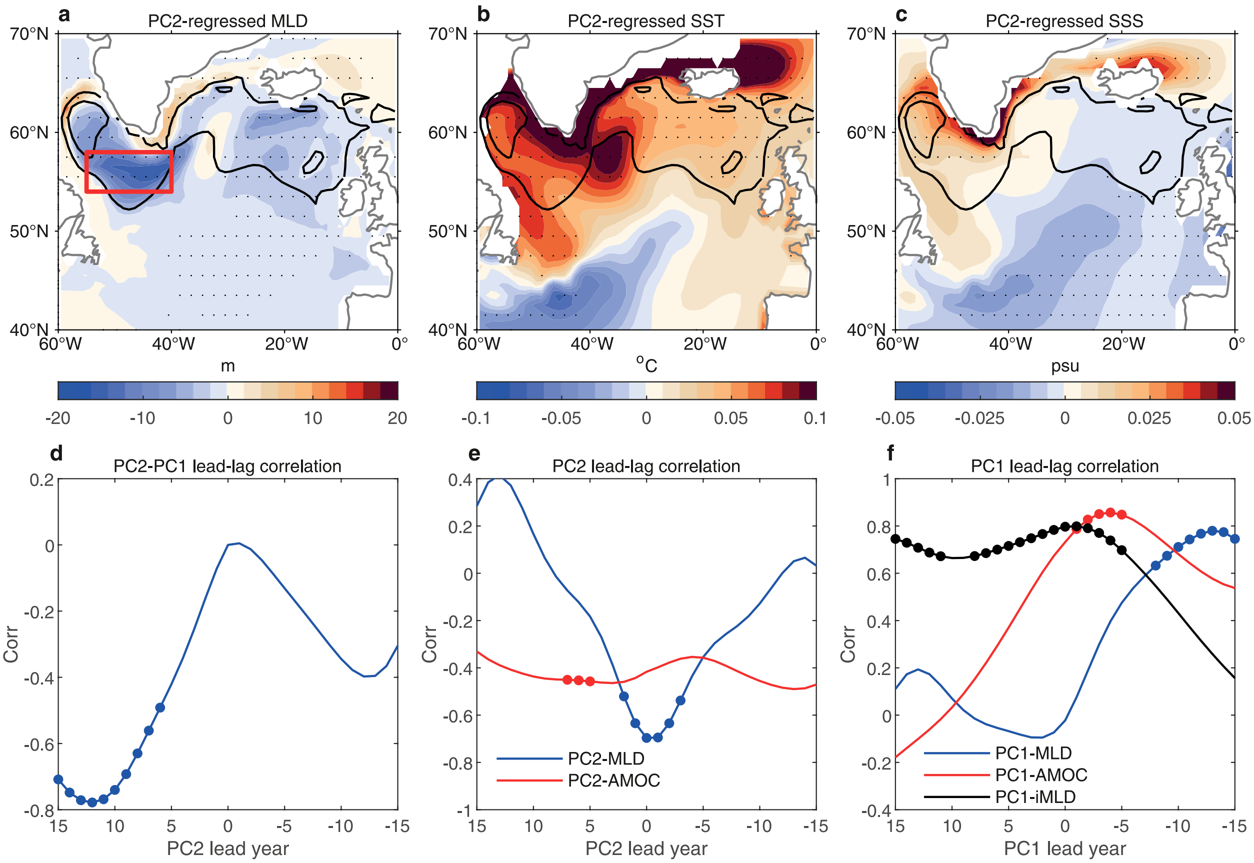

Figure 3. Mechanisms of the monopole response. The PC2-regressed anomalous MLD (a), SST (b), and sea surface salinity (c) shown by shadings. Contours denote climatological MLD at an interval of 200 m with zero line omitted. Stipplings denote regression coefficients exceeding the 90% confidence level. (d) Lead-lag correlation between the PC2 and PC1. Dots denote correlations exceeding the 90% confidence level. (e) Lead-lag correlation of the PC2 with the subpolar MLD anomaly (blue) averaged over 55°W–40°W, 54°N–58°N (red box in (a)) and AMOC anomaly (red). In panels (f) Same as (e) but for the PC1. Lead-lag correlation of the PC1 with the time-integrated subpolar MLD anomaly (iMLD) is also shown (black).

AMOC index (maximum streamfunction at 40°N). We calculate the lead-lag correlations between the two indices and the two leading PCs (representing the monopole and tripole responses).

The PC1 lags the PC2 by 12 years ($r = -0.78$; Figure 3d), meaning that the monopole response significantly lags the tripole response by about a decade. The subpolar MLD index correlates strongest with the PC2 contemporaneously ($r = -0.70$; blue line in Figure 3e) but with the PC1 at a 13-year lead ($r = 0.78$; blue line in Figure 3f), suggesting that tripole-induced deep water formation inhibition precedes the monopole response by about a decade. To further validate the dynamic effect of subpolar MLD anomalies on the PC1, we integrate the subpolar MLD index through time (iMLD), following the method of Sun et al. (2019), which focused on the integrated effect of the NAO on the AMV. There is a strong contemporaneous positive correlation between iMLD and the PC1 ($r = 0.80$; black line in Figure 3f), confirming the dynamic effect of subpolar deep water formation rates on the monopole SST response. AMOC-PC2 correlations lack clear extrema (red line in Figure 3e), likely because AMOC sensitively responds to multidecadal rather than decadal buoyancy forcing (Delworth & Zeng, 2016) while the PC2 dominates decadal time scales (Figure 1f). The PC1 significantly lags the AMOC by 4 years ($r = 0.86$; red line in Figure 3f), consistent with SST lagging AMOC-induced ocean heat transport (Tandon & Kushnir, 2015; Zhang & Wang, 2013). The full tripole-to-monopole transition is shown in Figures S7 and S8 of Supporting Information S1.

3.4. Contribution to the Historical AMV

We first quantify the contribution of TEP forcing to North Atlantic internal decadal SST variability under the POGA framework. Although the explained percentages in the subpolar region are moderate, there is a striking

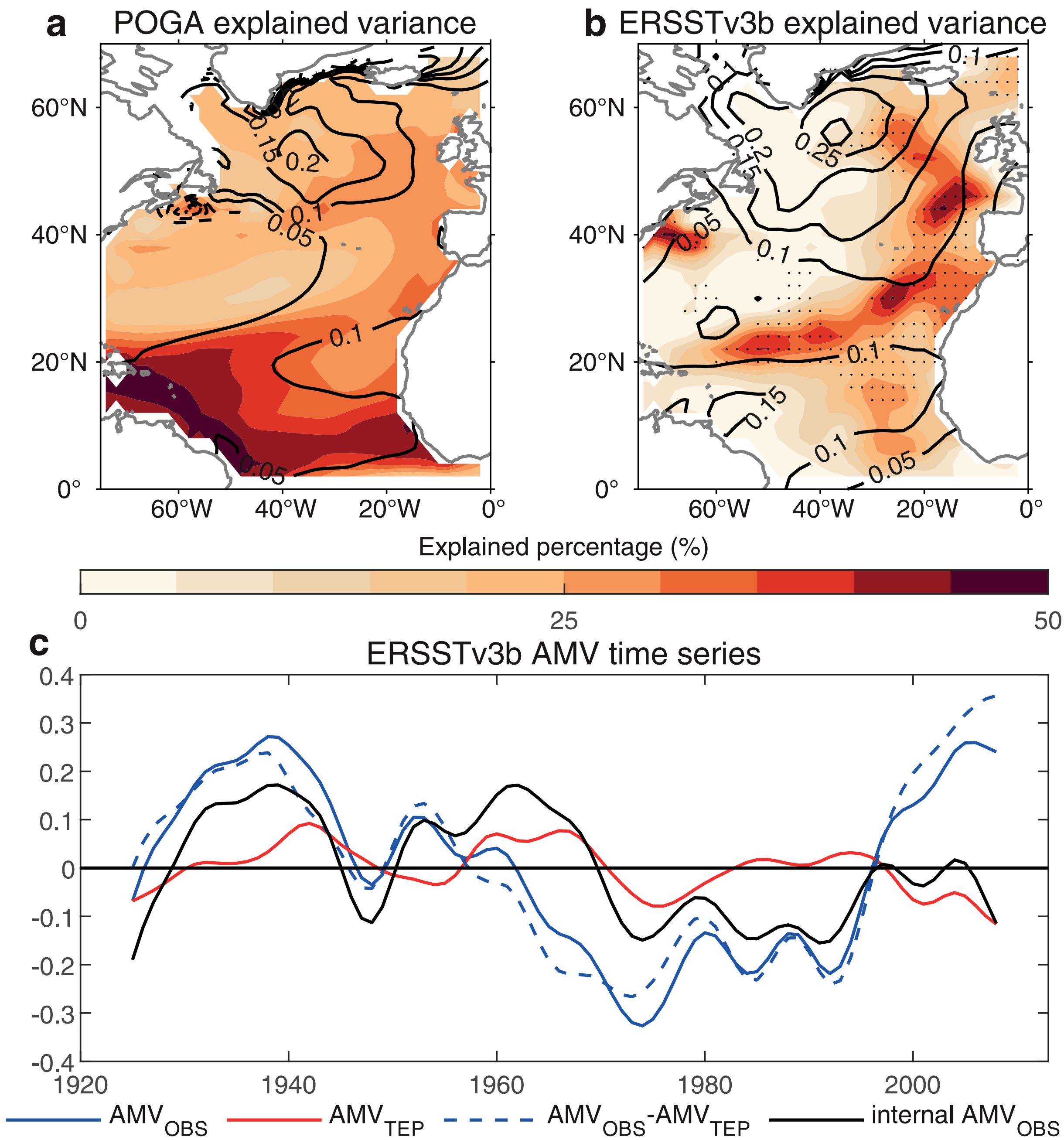

Figure 4. Contribution of the AMV_{TEP} to the historical AMV. The explained percentages (squared correlation coefficient; set to zero if correlation is negative) of TEP-forced SST variability on internal decadal SST variability averaged across the POGA ensemble members (a) and in the ERSSTv3b (b) are shown by shadings. Internal AMV-regressed internal decadal SST anomalies are shown by contours (°C). Dots in (b) denote that the observed explained percentages lie within the POGA ensemble range. (c) Time series in the ERSSTv3b: the AMV_{OBS} (blue solid), AMV_{TEP} (red), AMV_{OBS} minus AMV_{TEP} (blue dashed), and the internal AMV_{OBS} (black).

contribution along the horseshoe pathway from the midlatitude to the tropical North Atlantic in the eastern part of the ocean basin and in the region where the internal AMV loading is negative (Figure 4a).

Qualitatively similar results are found in the observations (Figure 4b). However, some differences remain. The most notable one is in the western tropical North Atlantic, where the model shows strong TEP-forced variance but the observations display weak TEP-forced variance. This discrepancy may arise because the model underestimates the total internal decadal SST varibility in this region (contours in Figures 4a and 4b), which may be generated by other sources. The second difference is a southwestward shift of the center of high explained percentages around 40°N near the North American coast from the model to the observations.

The AMV_{TEP} modulates the AMV_{OBS} through affecting its internal component. The AMV_{TEP} peaks/valleys substantially modulate the timing of phase changes and intensity of the AMV_{OBS} (Figure 4c). For example, an early-1960s AMV_{TEP} peak delays an AMV_{OBS} phase transition by 5 years; an early-2000s AMV_{TEP} valley weakens the positive phased AMV_{OBS} as its removal strengthens the AMV_{OBS} by 33% (1997–2008 mean).

4. Conclusions and Discussions

In this paper, we used a tropical Pacific pacemaker experiment to study the TEP-forced AMV (AMV_{TEP}). We decomposed the AMV_{TEP} into two distinct response modes with monopole and tripole SST pattern, explaining 59% and 40% variance in the AMV_{TEP}, respectively. The TEP forces the tripole SST pattern primarily through emission of atmospheric teleconnections that affect net surface heat flux. The subpolar portion of this tripole response induces anomalous surface buoyancy flux, thereby modulating deep water formation, the AMOC, and ultimately leading to the monopole response approximately one decade later. We further demonstrated that the AMV_{TEP} contributes to the historical AMV (AMV_{OBS}) evolution by substantially modulating both the intensity and the timing of phase changes. The contribution stems from the fact that the AMV_{TEP} explains a substantial fraction of the internal AMV_{OBS}, with the extratropical component more important than the tropical component.

This study suggests that the TEP decadal SST variability, which is linked to the IPO, acts as a decade-scale precursor for the AMV. This newly discovered factor should be considered in future efforts to improve decadal climate prediction of the AMV and its global impacts. Further studies are needed to assess whether climate models can adequately simulate the TEP forcing. Given that the TEP-forced sea level pressure pattern over the North Atlantic resembles the NAO, it is also important to investigate whether the TEP forcing contributes to common model biases in the NAO-AMOC-AMV relationship (Kim et al., 2018, 2023; Wang, Li, et al., 2017). However, it should be noted that the TEP-forced NAO-like pattern shifts northward relative to the NAO, thus its SST response shows some differences from NAO-forced AMV (Sun et al., 2019). The main difference is a weak modulation of the TEP-forced NAO-like pattern in the tropical North Atlantic (Figure 2c). TEP-forced tropical North Atlantic variability occurs mainly via tropospheric Kelvin waves and modulation of the Walker Circulation, not through the extratropical NAO-like pattern.

As previous studies also suggested AMV modulations on Pacific decadal variability (e.g., An et al., 2021; d'Orgeville & Peltier, 2007; Kucharski et al., 2016; Li et al., 2016; McGregor et al., 2014; Ruprich-Robert et al., 2017; Sun et al., 2021; Zhang & Delworth, 2007), our study improves the understanding of the potential interactions between the Atlantic and Pacific SST variability (e.g., Hong et al., 2022; Meehl et al., 2021; Hu et al., 2025; Xu et al., 2025 for decadal variability; Jiang & Li, 2021; Scaife et al., 2024 for interannual variability), which is crucial for global and regional decadal climate prediction (Chikamoto et al., 2015; Lv et al., 2022). Mechanisms such as the Walker Circulation, which play important roles in contributing to the bidirectional trans-basin effect between Atlantic and Pacific decadal variability (Hong et al., 2022; Sun et al., 2021), need more attention in the future to establish a potential two-way trans-basin mechanism. Climate models show strong systematic biases (Li et al., 2020; McGregor et al., 2018) and large inter-model spread (Ruprich-Robert et al., 2021; Yang et al., 2022) in simulating the AMV-forced IPO. A similar situation may exist for the IPO-forced AMV and warrants further investigation.

Some caveats remain due to the model design of our study. First, because of the common modeled SST bias in the TEP (Seager et al., 2019), the AMV_{LE} may be affected, which in turn influences the model-estimated AMV_{TEP}. Second, the observed SST anomalies used for nudging in the POGA experiment may include a radiatively forced component, therefore the AMV_{TEP} may contain some forced variability. Within the model framework, the modeled AMV_{TEP} can be more precisely calculated by nudging modeled TEP SST anomalies.

Conflict of Interest

The authors declare no conflicts of interest relevant to this study.

Availability Statement

The ERSSTv3b was obtained via Smith et al. (2008); the raw data of the CESM1 POGA and LE experiments were obtained from http://www.earthsystemgrid.org (requiring registration and log-in information), and we uploaded the processed CESM1 data via Yang (2026).

Acknowledgments

We thank Profs. Shang-Ping Xie and Jian Lu for constructive comments. J.-C.Y. and X.L. were supported by the National Key Research and Development Program of China (2023YFF0805200), the Shandong Provincial Natural Science Foundation (ZR2026MS0565), the National Natural Science Foundation of China (42105019), and the Fundamental and Interdisciplinary Disciplines Breakthrough Plan of the Ministry of Education of China (JYB2025XDXIM801); J.L. was supported by Laoshan Laboratory (LSKJ202202600); Y.Z. was supported by the National Natural Science Foundation of China (W2411039) and Laoshan Laboratory (LSKJ202202602).

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