Multidecadal variability from ocean to atmosphere in the North Atlantic: Perturbation potential energy as the bridge

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ABSTRACT

The North Atlantic Ocean forcings are considered an important origin of the North Atlantic atmospheric multidecadal variability. Here we reveal the energetics mechanisms of the phenomenon using the perturbation potential energy (PPE) theory. Supporting the previous model studies, a cyclic pattern involving the Atlantic multidecadal oscillation (AMO) and North Atlantic tripole (NAT) is observed: positive AMO phase (AMO⁺, similarly hereafter) → NAT⁻→ AMO⁻→ NAT⁺, with a phase lag of approximately 15~20 years. An atmospheric mode characterized by basin-scale sea level pressure anomaly in the North Atlantic is associated with the AMO, which is termed the North Atlantic uniformity (NAU). The AMO⁺ induces positive uniform PPE anomalies over the ocean through precipitation heating, leading to decreased energy conversion to perturbation kinetic energy (PKE) and a large-scale anomalous cyclone. For the NAT⁺, tripolar precipitation anomalies result in tripolar PPE anomalies. Anomalous energy conversions occur where the PPE anomaly gradient is large, explained by an energy balance derived from thermal wind relationship. The PKE around 15°N and 50°N (25°N and 75°N) increases (decreases), forming the anomalous anticyclone and cyclone at subtropical and subpolar region, respectively, known as the North Atlantic Oscillation (NAO). The reverse holds for the NAT⁻ and AMO⁻. As the phases of the ocean modes alternate, the energetics induce the NAU⁻, NAO⁻, NAU⁺, and NAO⁺ sequentially. In the multidecadal cycle, the accumulated energetics process is related to delayed effect, and the difference in variance explanation between the NAU and NAO is attributed to the feedback mechanisms.

SIGNIFICANCE STATEMENT

The North Atlantic Ocean’s multidecadal changes affect the atmosphere above it. Our study explores the energy processes behind this phenomenon. The North Atlantic Ocean’s temperature distribution goes through a shift every 15~20 years, persistently affecting the air’s potential energy through the heat release related to vapor condensation. The changed potential energy converts into kinetic energy, causing the atmospheric circulation to alternate between different states. Our study provides a comprehensive explanation of how the ocean affects the region's climate. This insight may contribute to making more accurate models and predictions of climate changes in the North Atlantic.

1. Introduction
The atmosphere circulation over the North Atlantic is of great concern because of its extensive climate impacts across the region and globally. The North Atlantic Oscillation (NAO), characterized by the opposite sea level pressure (SLP) variations in the Azores high and Iceland low (Wallace and Gutzler, 1981), is the dominant mode of the atmosphere variability in the North Atlantic. Throughout history, the NAO has undergone significant multidecadal variability with widespread climate impacts (Hurrell 1995; Cohen and Barlow 2005; Deser and Teng 2008; Dieppois et al. 2013). The late 20th century witnessed a significant upward trend in the NAO (Thompson and Wallace 2000), followed by a pronounced downward trend post-1990. The multidecadal variability in the NAO has been identified as a key factor contributing to the anomalous temperatures over Eurasia and North America (Hurrell 1996) and the Northern Hemisphere (Cohen et al. 2012; Wang et al. 2010), along with its association with the multidecadal variability in the Asian winter monsoon (Gong et al. 2001; Li et al. 2019, 2022). At the multidecadal scale, the NAO emerges as an effective climate predictor (Li et al. 2013; Sun et al. 2015; Xie et al. 2021; Li et al. 2022). A deeper understanding of the atmospheric multidecadal variability in the North Atlantic is crucial for comprehending and predicting the multidecadal variability in the Earth's climate system.

The atmosphere over the North Atlantic exhibits relatively high multidecadal predictability (Latif et al. 2006), with the North Atlantic Ocean as a potential influential factor (Czaja and Marshall, 2000; Wu and Gorden 2002; Gulev et al. 2013; Gastineau and Frankignoul 2015). Observational and model studies have proved the coupled relationship and positive feedback between the North Atlantic tripole (NAT) and NAO (Czaja and Frankignoul 2002; Kushnir et al. 2002; Wu and Rodwell 2004; Wu et al. 2009; Gastineau et al. 2013). Additionally, the Atlantic multidecadal oscillation (AMO), characterized by basin-wide North Atlantic sea surface temperature (SST) anomalies (SSTA), is also linked to the atmospheric variability in the North Atlantic. The increase of the North Atlantic SSTA can induce low pressure anomalies (Sun et al. 2015) and ascending motion (Sun et al. 2017) over the North Atlantic. The anomalous cross-equatorial winds and the meridional movement of intertropical convergence zone are coupled with the SSTA variability in the northern tropical Atlantic (Zhang and Delworth 2006), which arise from the wind-evaporation-SST feedback (Xie 1999; Chiang and Vimont 2004). The anomalous precipitation (Frierson et al. 2013) and upper-level divergence induced by the AMO are thought to be the origin of the Africa-Asia multidecadal teleconnection (Sun et al. 2017).
The mechanism of the multidecadal variability of the North Atlantic SST is still under debate. Low frequency variability of ocean circulation (Medhaug and Furevik 2011; Danabasoglu et al. 2016; Buckley and Marshall 2016) and atmospheric forcing (Clement et al. 2015; O’Reilly et al. 2016; Booth et al. 2012) have been both proposed as contributors to the AMO. The NAO can drive the Atlantic meridional overturning circulation (AMOC) through a delayed effect (Álvarez-García et al. 2008; Delworth and Greatbatch 2000), thus affecting the poleward heat transport to induce the delayed AMO mode (Delworth et al. 2017). It has also been suggested that air-sea interaction may be the key process in the multidecadal variability of the air-sea system in the North Atlantic (Wills et al. 2019). In the model simulation of Sun et al. (2015), it is found that the AMO can further translate to the NAT through ocean dynamics, which further provides feedback on the NAO due to the coupling relationship between the NAT and NAO. In a complete multidecadal cycle, the North Atlantic is dominated by the positive NAT phase (NAT+, similarly hereafter), AMO+, NAT− and AMO− in sequence, with a phase lag of about 15~20 years. As an external forcing, the oscillation of the ocean modes provides a possibility for the formation of multidecadal oscillation of the atmosphere modes.

Energetics is an effective tool for explaining how the ocean affects atmosphere. Though the atmosphere possesses a large amount of total potential energy, only a small fraction can be converted into kinetic energy to influence circulation, known as available potential energy (APE) (Margules 1910). Lorenz (1955) derived a mathematical expression for the APE, explaining its physical meaning as the difference between the total potential energy of the actual atmosphere and that of the reference state atmosphere. The reference state is the state of the minimum total potential energy that the system can reach through adiabatic adjustment. The concept of APE has been widely used to analyze the energetics of the ocean and atmosphere (Oort et al. 1989; Peixoto and Oort 1974). However, APE is the concept for the global atmospheric system, which cannot explain the local energetics, and the definition of Lorenz’s reference state does not considers the effect of topographic heterogeneity (Gao et al. 2006). To address these limitations, Gao et al. (2006) proposed the concept of conditional minimum total potential energy as the reference state. This led to the concept of perturbation potential energy (PPE) (Li and Gao 2006), which exhibits a greater magnitude than APE at the local scale. Therefore, PPE dominates the local energy availability, which has better practicability in local atmospheric circulation energetics.
Subsequent developments extended the concept of PPE to layer perturbation potential energy (LPPE) (Wang et al. 2012) and applied it to the analysis and diagnosis of climate dynamics. PPE has a close physical relationship with the annular mode (Wang et al. 2014; Wang et al. 2015). The anomalous conversion between PPE and perturbation kinetic energy (PKE) is an important mechanism of monsoon variability (Huyan et al. 2017; Zhang et al. 2019). PPE induces equatorial zonal winds in the negative feedback process of the phase transition of El Nino-Southern Oscillation (Dong et al. 2017). PPE also plays a bridging role in the process where the east pole anomaly induces the west pole anomaly in the Indian Ocean dipole events (Wang et al. 2019), which can be triggered by the PPE anomaly of the South China Sea summer monsoon (Zhang et al. 2019; Zhang et al. 2023).

Although some studies suggest that the atmospheric multidecadal variability in the North Atlantic may be attributed to the forcings from the North Atlantic Ocean, few studies focused on the underlying energetics mechanisms, which may be clarified by the PPE theory. This study aims to use the PPE framework to analyze the diabatic heating from the North Atlantic Ocean, its effect on PPE and the impact of PPE anomaly on atmospheric circulation. The remainder of this paper is organized as follows. Section 2 describes the datasets and methods used. In Section 3, we detected a multidecadal transition between the NAO and a uniform atmospheric pattern, which is associated with the North Atlantic SST modes. Then we analyze the energetics of the two atmospheric modes. Section 4 is about how the ocean affects PPE. Section 5 is about how PPE affects atmospheric circulation. Section 6 analyzes the accumulated process and feedback involved in the energetics mechanisms. A summary and discussion of the results are provided in Section 7.

2. Data and methodology

a. Data and indices

The Twentieth Century Reanalysis Version 3 dataset (Compo et al. 2011; Slivinski et al. 2019) is employed to analyze the three-dimensional field variability of the atmosphere. The dataset provides atmospheric reanalysis variables dating back to 1836, with a horizontal resolution of 1°×1°. Various variables including air temperature, wind, vertical velocity, specific humidity, precipitation rate, sensible heat net flux, latent heat net flux, longwave radiation flux, and shortwave radiation flux are used. The SLP data are from the HadSLP2 dataset (Allan and Ansell 2006) with a resolution of 5° × 5°, available from 1850 onward. The SST data is from the Extended Reconstructed SST Version 5 dataset (Huang et al. 2017).
with a horizontal resolution of $2^\circ \times 2^\circ$, starting from 1854. All analyses in this study are conducted on annual mean values for the period 1854–2015, which is the intersection of the temporal range of the above datasets.

To quantify key climate variabilities, the following definitions are employed. The AMO index (AMOI) is defined as the area-weighted average of the detrended SST in the North Atlantic domain ($75^\circ-7.5^\circ W$, $0^\circ-60^\circ N$). The NAO index (NAOI) is defined as the difference between the normalized SLP zonally averaged from $80^\circ W$ to $30^\circ E$ at $35^\circ N$ and $65^\circ N$, following Li and Wang (2003). The NAT index (NATI) is defined as follows:

$$NATI = \frac{1}{2} \left( SST_{60^\circ-30^\circ W,25^\circ-50^\circ N} - \frac{1}{4} \left( SST_{40^\circ-15^\circ W,55^\circ-70^\circ N} + SST_{75^\circ-45^\circ W,0^\circ-20^\circ N} \right) \right),$$

where the subscripts denote the ranges of the area-weighted average (the ranges are based on the Figs. A1a,e).

b. PPE and its governing equation

According to the PPE theory (Li and Gao 2006), PPE is the local difference between the total potential energy of the actual atmosphere and that of the conditional minimum reference state:

$$PPE = TPE^\sim = TPE - \overline{TPE} = \frac{1}{(1 + \kappa)\gamma_d p_{00}} \int_{\theta_S}^{\theta_T} \left( p^{1+\kappa} - \overline{p}^{1+\kappa} \right) d\theta,$$

where $\overline{()}$ denotes the global mean value on isobaric surface, $(())^\sim$ denotes the difference between the local value and the global mean value, and $TPE$ is the total potential energy of the air column. By expanding Eq. (1) in series, the mathematical formula of PPE in isobaric coordinates is expressed as (detailed derivations can be found in the Supplemental Material):

$$PPE = \sum_{i=1}^{\infty} PPE_i = \sum_{i=1}^{\infty} \frac{p_{00}^{(i-1)\kappa}}{i!} \left[ \prod_{j=0}^{i-1} (1 + \kappa - j) \right] \int_0^{p_s} \left( \frac{T^i}{p^{(i-1)(1+\kappa)}} \right) \left( -\frac{\partial \theta}{\partial p} \right)^{-i+1} dp,$$

where $i (i=1, 2, \ldots)$ is the order of moment term of $PPE$; $p_{00}$ is the reference pressure (usually taken as 1000 hPa); $\kappa = \frac{R}{c_p}$, where $R$ is the gas constant of dry air and $c_p$ is the specific heat at constant pressure; $\gamma_d = g/c_p$ is the dry adiabatic lapse rate, where $g$ is the gravitational acceleration; $\theta$ is potential temperature. The higher-order moment terms are of much smaller magnitude than the lower-order moment terms, and can be omitted in local analyses (Li and Gao 2006). The expressions for the first and second-order moment terms of PPE are as follows:
\[ PPE_1 = \frac{1}{\gamma_d} \int_0^{P_s} T^{-\gamma} \, dp , \]  
\[ PPE_2 = \frac{\kappa \rho_0^\kappa}{2 \gamma_d} \int_0^{P_s} T^{-2} \left( \frac{-\partial \hat{\theta}}{-\partial p} \right)^{-1} \, dp , \]  
(3) (4)

Note that the first-order moment term can be positive or negative, and the second-order moment term is always positive. By averaging PPE globally, the average of \( PPE_1 \) is zero, and the average of \( PPE_2 \) corresponds to the Lorenz APE. Thus, the \( PPE_1 \) dominates the local energy availability, while the APE is the energy availability of the global atmospheric system.

The governing equations of \( PPE_1 \) and \( PKE \) (\( PKE = \frac{1}{g} \int_0^{P_s} K^- \, dp \), where \( K = \frac{1}{2} (u^2 + v^2) \)) are as follows:

\[
\frac{\partial PPE_1}{\partial t} = \frac{1}{g} \int_0^{P_s} (\omega \alpha)^- \, dp + \frac{1}{g} \int_0^{P_s} Q^- \, dp + \frac{c_p}{g} \int_0^{P_s} -\nabla_h \cdot (V_h T) \, dp ,
\]
\[
\frac{\partial PKE}{\partial t} = -\frac{1}{g} \int_0^{P_s} (\omega \alpha)^- \, dp + \frac{1}{g} \int_0^{P_s} (V_h \cdot F_h)^- \, dp + \frac{1}{g} \int_0^{P_s} (-\nabla_h \cdot (V_h K) - \nabla_h \cdot (V_h \Phi)) \, dp ,
\]
(5) (6)

where \( PC_k \) is the perturbation energy conversion between \( PPE_1 \) and \( PKE \), \( \omega \) is the pressure vertical velocity and \( \alpha \) is the specific volume. \( PC_k \) is related to the variability of the local atmospheric circulation. When \( PC_k < 0 \), \( PPE_1 \) is converted to \( PKE \), and when \( PC_k > 0 \), \( PKE \) is converted to \( PPE_1 \). \( G \) is the generation term, dependent on sensible heat, latent heat release and radiative forcing. \( HBF_p \) is the horizontal boundary flux term, where \( \nabla_h \) horizontal gradient operator and \( V_h \) is the horizontal wind vector. \( HBF_K \) is the horizontal boundary flux of \( PKE \), where \( \Phi \) is the geopotential. \( D \) is the frictional dissipation term, where \( F_h \) is the horizontal friction force. Refer to the Supplemental Material for detailed derivations of these governing equations.

c. Methods

To suppress high frequency variabilities in the data, which are considered as noise for this study, a Gaussian low-pass filter was applied before analysis. The filter window of the filter is set to 11 years. Additionally, linear trends are removed to eliminate potential anthropogenic influence on long-term climate change. The statistical significance of the
linear correlation coefficient between two filtered time series is assessed using a two-tailed student t-test with the effective number of degrees of freedom (Pyper and Peterman 1998; Li et al. 2013). The effective number of degrees of freedom \((N_{\text{eff}})\) is approximated as follows:

\[
\frac{1}{N_{\text{eff}}} \approx \frac{1}{N} + \frac{2}{N} \sum_{j=1}^{N-2} \frac{N-j}{N} \rho_{XX}(j)\rho_{YY}(j),
\]

where \(N\) is the sample size, and \(\rho_{XX}(j)\) and \(\rho_{YY}(j)\) are the autocorrelations of two time series \(X\) and \(Y\) at lag \(j\), respectively.

To investigate the properties of the forcing terms in the governing equations on a multidecadal scale, the forcing terms were integrated over time to represent the accumulated effect over a period (Hasselmann 1976). The integral form for a time series \(x(t)\) is given by:

\[
X(t) = \int_{t_0}^{t} x(u) du,
\]

where \(X(t)\) is a function of time \(t\) (the upper limit of integration), which reflects the accumulated effect of \(x\) in the previous period. For the governing equations mentioned above, the integral form of the equations can be obtained by conducting the integration. All the time integration are integrated from the beginning year (1854) of the study period. Since the anomalies are zero-mean, the accumulated quantities are stationary, and the periodic fluctuations in the forcing terms will be neutralized in the accumulation process. Therefore, the accumulation here is actually a net accumulation of the previous fluctuant forcing.

To further examine atmospheric diabatic heating and the contribution of latent heat of condensation, the apparent heat source \((Q_1)\) and apparent moisture sink \((Q_2)\) (Yanai and Tomita 1998; Hsu and Li 2011) were also analyzed:

\[
Q_1 = c_p \frac{\partial T}{\partial t} - c_p (\omega \sigma - \mathbf{V}_h \cdot \nabla_h T),
\]

\[
Q_2 = -L \left( \frac{\partial q}{\partial t} + \mathbf{V}_h \cdot \nabla_h q + \omega \frac{\partial q}{\partial p} \right),
\]

where \(\sigma = \frac{RT}{c_p p} - \frac{\partial T}{\partial p}\) is the static stability, \(\mathbf{V}_h\) is the horizontal velocity, \(\nabla_h\) is the isobaric gradient operator, \(L\) is the latent heat of condensation, and \(q\) is the specific humidity. In this study, the \(Q_1\) and \(Q_2\) are vertically integrated.
In order to test the degree to which an index explains the variability of the region, we calculate the regional average variance explanation for the index, performed as follows: 1) 11-year Gaussian low-pass filter is performed on each point. 2) The percentage of variance explanation for each grid point is calculated using the index. 3) Calculate the regional average of the gridded variance explanation.

3. Multidecadal transition between the NAO and the North Atlantic uniformity (NAU)

Model simulation has revealed a multidecadal transition between the AMO and NAT (Sun et al. 2015). The lead-lag correlation analysis between zonal mean SST\(A\) and the AMO index, as illustrated in Figs. 1a, b, confirms this oscillation. The AMO, characterized by uniform SST\(A\) in the North Atlantic basin, exhibits a close relationship with the NAT pattern, with a lag time of approximately 15~20 years. When the AMO lags by about 15~20 years, the North Atlantic has positive (negative) SST\(A\) at around 40°N (15°N and 60°N), vice versa when the AMO leads by about 15~20 years. Upon removing the AMO signal from SST data, the tripole mode becomes more pronounced (Fig. 1b). This enhanced visibility results from the fact that the explained variance of SST\(A\) by the AMO, approximately 70% for the average in the North Atlantic, is much greater than that of the NAT, which is around 11%. Based on the lead-lag correlation coefficient field (Figs. A1a, e), the NAT index is defined (described in section 2.1). Lead-lag correlation (Fig. A1f) demonstrates that the AMO can be transformed into the NAT with a lag time of about 15~20 years, confirming the cycle of SST\(A\) modes on multidecadal time scales: NAT\(+\) → AMO\(+\) → NAT\(−\) → AMO\(−\).
Fig. 1. (a) Lead-lag correlation coefficients between the AMOI and regionally (80°−5°W) zonal averaged SST. (b) As in (a), but for the AMOI and SST*. The SST* denotes that the AMO signal is removed from SST at lag 0 year by a linear regression. (c) and (d) as in (a) and (b), respectively, but for the SLP and SLP*. The SLP* denotes that the NAO signal is removed from SLP at lag 0 year by a linear regression. The black dotted area denotes significant values at the 90% confidence level using the $N_{eff}$. Negative (positive) values of the x-axis denote that the AMOI lags (leads) the corresponding variables.

Figs. 1c, d show the lead-lag correlation between the AMO and zonal mean SLP in the North Atlantic. When the AMO leads (lags) about 15~20 years, the North Atlantic is dominated by the negative (positive) phase of the NAO, which aligns with previous studies (Li et al. 2013; Sun et al. 2015; Delworth et al. 2017). The NAO drives the AMO through an accumulated effect of heat transport, following the stochastic climate model (Hasselmann, 1976). Subsequently, the AMO undergoes adjustment to the NAT through ocean dynamics, and the NAO develops through positive feedback loop with the NAT (Sun et al. 2015). Then the oscillation proceeds in the opposite sense. After removing the NAO signal from SLP, the AMO showed uniform negative SLP correlations simultaneously, indicating that the AMO is associated with a large-scale low-pressure anomaly. Notably, as the NAO explains the
majority of the atmospheric variance in the North Atlantic, the uniform feature becomes more pronounced after removing the NAO signal (Fig. 1d). Therefore, the two atmospheric modes coupled to the ocean modes can be extracted: the NAT is associated with the NAO, while the AMO is associated with basin-scale uniform SLP anomalies. Here, the large scale SLP anomaly pattern in the North Atlantic is referred to as the NAU, with its positive (negative) phase characterized by a basin-scale positive (negative) anomaly of SLP in the North Atlantic. The index for the NAU (NAUI) is defined as:

\[
    NAUI = SLP_{70^\circ-10^\circ W,0^\circ-80^\circ N}.
\]  

Contrary to the NAO, the NAU represents the alike SLP variations in the Azores high and Iceland low. The analysis of SLP variance in the North Atlantic shows that the NAO explains about 33% of variance for multidecadal variability, while the NAU explains about 21%, slightly less than the NAO mode.

Fig. 2 shows the atmospheric circulation and SST associated with the NAO and the NAU. The NAO exhibits an anticyclonic circulation near 35°N and a cyclonic circulation near 65°N at the lower level, and three circulation centers near 20°N, 35°N, and 65°N at the upper level (Figs. 2a, c). The SST associated with the NAO exhibit tripole characteristics (Fig. 2e). Above the warm (cold) SST, there is an anticyclonic (cyclonic) circulation, and the vorticity at high and low levels are in the same phase, which has equivalent barotropic structure. In contrast, the NAU shows a large-scale anticyclonic circulation and divergence at low level over the North Atlantic basin, and cyclonic circulation and convergence at high level (Figs. 2b, d). The SST associated with the NAU is characterized by the AMO (Fig. 2f), consistent with the lead-lag correlation in Fig. 3b.
Fig. 2. Correlation coefficients of the NAOI with 200-hPa wind field (a), 850-hPa wind field (c) and SST (e). (b), (d), and (f) as in (a), (c), and (e), respectively, but for the NAUI. In (a)-(d), the blue arrows denote the significant correlation coefficient vectors at the 90% confidence level using the $N_{eff}$, and the unit vector is in the bottom left corner. The black dotted areas in (e) and (f) represent significant values at the 90% confidence level using the $N_{eff}$. 

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Fig. 3. (a) Annual time series of the AMOI, NAUI, NATI and NAOI for the period 1854–2015. All the indices are smoothed using a 11-year Gaussian low-pass filter. (b) The lead-lag correlation coefficients of the NAUI with the AMOI (blue solid line), NAOI (purple solid line) and NATI (yellow solid line). The orange solid line is the autocorrelation of the NAUI. The correlations are based on the smoothed indices. Positive (Negative) values of the x-axis denote that the NAUI lags (leads) the corresponding indices. The dashed lines indicate the 90% confidence level using the $N_{eff}$. (c) Schematic diagram of the multidecadal cycle of atmospheric patterns in the North Atlantic.

Therefore, we identified two modes of the atmosphere over the North Atlantic coupled to the ocean at a multidecadal scale: the NAU and NAO, characterized by uniform and dipole SLP anomalies, respectively. Lead-lag correlation analysis showed that the NAU and AMO are negatively correlated simultaneously, while the NAU leads the NAO/NAT about 15–20 years with positive correlation and lags the NAO/NAT about 15–20 years with negative correlation (Fig. 3b). This suggests that the multidecadal SST mode cycle may influence the atmospheric multidecadal variability. Therefore, similar to the transition between the AMO
and NAT in the ocean, there is an alternating cycle of NAO\textsuperscript{+}, NAU\textsuperscript{−}, NAO\textsuperscript{−} and NAU\textsuperscript{+} in the atmosphere, as shown in Fig. 3c. Next, we will further investigate the energetics of these two atmospheric modes.

4. Diabatic heating from the multidecadal ocean modes

According to the theory of atmospheric PPE (Li and Gao 2006), the local external diabatic forcing cannot directly impact the local atmospheric circulation. Instead, it can only change PKE through PPE. Figs. 4a,b show the correlation between PPE and the AMO/NAT in the North Atlantic. Similar to SST\textsubscript{A}, the PPE anomalies have two modes: a uniform pattern (Fig. 4a) and a tripole pattern (Fig. 4b). The spatial distribution of PPE anomalies closely mirrors that of SST\textsubscript{A} in the North Atlantic. During the AMO\textsuperscript{+}, the North Atlantic shows uniform positive PPE anomalies, while during the NAT\textsuperscript{+}, the North Atlantic shows positive PPE anomalies around 40°N and negative PPE anomalies around 20°N and 60°N. The PPE tripole pattern is most pronounced when leading or lagging the PPE uniform pattern by about 15~20 years (Figs. 4c,d), resembling the lag time of tripole and uniform SST modes. This suggests that SST\textsubscript{A} acts as an external forcing of PPE anomalies on the multidecadal time scale.
Fig. 4. (a) Correlation coefficient map between the AMOI and PPE. (b) As in (a), but for the NATI and PPE. (c) As in Fig. 1a, but for the PPE. (d) As in (c), but for the uniform PPE index (defined as the areal weighted mean PPE in 0°-60°N, 70°-10°W) and PPE*. The PPE* denotes that the uniform PPE signal is removed at lag 0 year by a linear regression. Positive (Negative) values of the x-axis denote that the AMOI or uniform PPE index leads (lags) the PPE or PPE*, respectively. The black dotted areas denote significant values at the 90% confidence level using the $N_{eff}$.

To investigate the diabatic heating source of PPE anomalies, we calculated four atmospheric diabatic heating terms associated with two SST modes: sensible heat, latent heat, longwave radiation and shortwave radiation. Considering that only the condensation of water vapor can heat the atmosphere, we use the precipitation rate to represent the latent heat release. In addition, to assess the accumulated effect of these terms, as described in Section 2, we integrated the governing equation of $PPE_1$ from $t_0$ to $t$ to obtain the equation in the form of integral:

$$PPE_1(t) = PPE_1(t_0) + \int_{t_0}^{t} PC_k(u)du + \int_{t_0}^{t} G(u)du + \int_{t_0}^{t} HBF_p(u)du.$$  \hspace{1cm} \text{(12)}

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Therefore, the \( PPE_1 \) at time \( t \) equals the initial \( PPE_1 \) at time \( t_0 \) (a constant determined by the selection of \( t_0 \), set as the start time of the analysis period in this study) plus the integral of governing terms from \( t_0 \) to \( t \). This formulation captures the source of PPE anomalies considering the accumulated effect in the preceding period. In this paper, the terms on the right-hand side of Eq. (5) and Eq. (6) are referred to as instantaneous quantities, for they govern the instantaneous rate of change of \( PPE_1 \) and \( PKE \). The time integrations of these terms are referred to as accumulated quantities. The changes in PPE will be analyzed using the terms of the governing equation in both instantaneous and accumulated forms.

The diabatic heating associated with the AMO is shown in Fig. 5. Among the four terms, only the instantaneous latent heat shows significant positive correlation (\( \alpha = 0.1 \)). The correlation between the AMO and precipitation also exhibit uniform pattern (Fig. 5b). Moreover, both the instantaneous \( Q_1 \) and \( Q_2 \) also have a high positive correlation. The partial correlation analysis, after removing the signal of instantaneous latent heat, reveals a decrease in the correlation between the AMO and PPE uniform pattern (defined as the areal weighted mean PPE in 0°-60°N, 70°-10°W) from 0.83 to 0.59, rendering it statistically insignificant using the \( N_{eff} \). This suggests that the AMO influences the PPE through the latent heat of condensation released by precipitation. This is consistent with previous studies (Sun et al. 2015; Frierson et al. 2013), which identify the AMO's capacity to induce anomalous vertical motion and precipitation. Specifically, the warming (cooling) of SST in the North Atlantic basin leads to anomalous ascending (descending) motion, conducive to increased (decreased) water vapor condensation. This, in turn, results in the diabatic heating (cooling) of latent heat and the increase (decrease) of PPE.
Fig. 5. (a) Correlation coefficients of the AMOI with net sensible heat flux, released latent heat, net longwave radiation flux, net shortwave radiation flux, $Q_1$ and $Q_2$ averaged in the North Atlantic ($70^\circ W-0^\circ$, $0^\circ-65^\circ N$) for both instantaneous quantity (blue bars) and accumulated quantity (orange bars). The bars with black outline are significant at the 90% confidence level using the $N_{eff}$. (b) Correlation coefficients between the AMOI and instantaneous latent heat release. The black dotted areas represent significant values at the 90% confidence level using the $N_{eff}$.

The same method is applied to analyze the diabatic heating of the NAT mode. Fig. A2 indicates that the sensible heat, longwave radiation and shortwave radiation cannot explain the characteristics of the PPE tripole pattern. Notably, the characteristics of accumulated latent heat appear to account for the tripole pattern of PPE anomalies, while the instantaneous latent heat does not (Figs. 6a,b). Resembling the PPE and accumulated latent heat, both accumulated $Q_1$ and $Q_2$ exhibit tripole anomalies (Figs. 6c,d), providing further evidence of the role of accumulated latent heat release. Therefore, the diabatic heating process associated

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with the NAT is as follows: the warm (cold) SSTA area promotes anomalous increases (decreases) in precipitation. These precipitation anomalies induce the tripole PPE anomalies by releasing latent heat through accumulated effect.

![Graphs showing correlation coefficients between the NAT and instantaneous latent heat release.](image)

**Fig. 6.** (a) Correlation coefficients between the NATI and instantaneous latent heat release. (b) As in (a), but for the NATI and the accumulated latent heat release. Correlation coefficients of the NAOI with the accumulated $Q_1$ (c) and $Q_2$ (d). The black dotted areas represent significant values at the 90% confidence level using the $N_{eff}$.

When the ocean alternates between the AMO and the NAT, the atmospheric PPE over the ocean alternates between uniform and tripole patterns. In both cases, the diabatic heating processes can be attributed to the latent heat of condensation released by precipitation. An interesting observation is that the uniform PPE pattern is influenced by instantaneous latent heat, whereas the tripole PPE depends on accumulated latent heat. This discrepancy may be attributed to the amplitude of the SSTA modes. The AMO, with a larger explained variance of North Atlantic SSTA (70%) and a greater SSTA amplitude (about 0.4 K), may manifests

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its influence simultaneously. In contrast, the influence of the NAT, with smaller explained variance (11%) and SST amplitude (about 0.2 K), may be realized through accumulated effects. PPE acts as a bridge between external heating and atmospheric circulation changes. Therefore, the next section analyzes how PPE anomalies affect the multidecadal variability of atmospheric circulation over the North Atlantic through energy conversion.

5. Impact of energy conversion on the PKE

According to the governing equations of $PPE_1$ and $PKE$, the conversion term $PC_k$ serves as the link between their variabilities. In this section, we will analyze how PPE anomalies can be converted into PKE anomalies through vertical motion. Similar to the operation to $PPE_1$ governing equation, the governing equation of $PKE$ is integrated over time, as in Eq. (13). The $PC_k$ is examined in both instantaneous and accumulated forms.

$$PKE(t) = PKE(t_0) - \int_{t_0}^{t} PC_k du + \int_{t_0}^{t} D du + \int_{t_0}^{t} HB_{K}F_{K} du.$$ (13)

The correlation between the AMO and PKE is illustrated in Figs. 7a,b. Large-scale negative PKE anomalies are observed in the North Atlantic during the AMO*, particularly at the lower level. The corresponding lower atmosphere response manifests as a cyclonic circulation (the opposite phase of Fig. 2d), on the contrary to the climatological North Atlantic subtropical high and westerlies, resulting in a weakening of the lower-level circulation. However, at the middle and upper levels, PKE anomalies do not exhibit uniformity and may also be influenced by other factors, such as $HB_{K}F_{K}$, which is not discussed in this study.
Fig. 7. (a) Correlation coefficients between 900-hPa PKE and the AMOI. (b) Correlation coefficients between regionally (70°–5°W) zonal mean PKE and the AMOI. (c) As in (a), but for 300-hPa PKE and the NAOI. (d) As in (b), but for the NAOI. The black dotted area denotes significant values at the 90% confidence level using the $N_{eff}$.

Fig. 8 displays the energy conversion associated with the AMO. The instantaneous $PC_k$ anomalies (Fig. 8a) associated with the AMO cannot explain the negative PKE anomalies, because the instantaneous vertical motion is upward, which represents that PPE is converted to PKE anomalously. However, there is a positive correlation pattern for accumulated $PC_k$ at lower levels over the North Atlantic basin (Fig. 8b). The AMO is associated with accumulated descending motion (Fig. 8c), which decreases the conversion from PPE to PKE. Therefore, the weakening (strengthening) of lower-level atmospheric circulation over the North Atlantic during the AMO+ (AMO−) can be attributed to the accumulated energy conversion.
Fig. 8. (a) Correlation coefficients between the AMOI and instantaneous $PC_k$ at 1000-850 hPa. (b) As in (a), but for the accumulated $PC_k$. The black dotted areas in (a) and (b) denote significant values at the 90% confidence level using the $N_{eff}$. (c) Correlation coefficients between regionally (60°–10°W) zonal mean accumulated $PC_k$ and the AMOI (shading), correlation coefficient vectors between the AMOI and accumulated regionally (60°–10°W) zonal mean meridional circulation (arrows). The unit vector is in the upper left corner. The black arrows denote the significant correlation coefficient vectors at the 90% confidence level using the $N_{eff}$. The areas surrounded by black lines denote significant values at the 90% confidence level using the $N_{eff}$.

The PKE anomaly of the NAO is also characterized by a meridional quadrupole pattern (Fig. 7c), and strengthens with height (Fig. 7d). At the upper level of the atmosphere over the North Atlantic, climatologically dominated by westerly winds, variations in PKE primarily reflect the variations in westerly intensity. In the latitudes around 10°N and 50°N (30°N and 75°N) with positive (negative) PKE anomalies, there are westerly (easterly) anomalies in the same (opposite) direction as the climatological wind field (Fig. 2a). The positive (negative) PPE anomaly area corresponds to anticyclonic (cyclonic) circulation (Fig. 4b and Figs. 2a,c). Notably, different from the PPE anomalies and SST anomalies, PKE anomalies exhibit quadrupole characteristics rather than a tripolar pattern. The zonal average results show that the latitudes

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of the strongest PKE anomalies are situated between the positive and negative PPE anomalies, where the PPE gradient is the largest (Fig. 9).

![Zonal mean SST, PPE, PKE, PCₖ and u-wind of NAO/NAT mode](image)

**Fig. 9.** Correlation coefficients between the NATI and regionally (65°W–0°) zonal mean SSTA (blue line). Correlation coefficients of the NAOI with regionally (65°W–0°) zonal mean PPE (orange line), PKE (yellow line) and instantaneous $PC_k$ (purple line). The part of lines with black dots denotes the significant values at the 90% confidence level using the $N_{eff}$.

The spatial relationship between the PPE gradient and the wind anomaly is consistent with the thermal wind dynamics. In the region between anomalous heating and anomalous cooling, the anomalous wind reaches its peak at the upper level (Fig. 7d). Considering that the thermal wind relationship does not include energy in explicit form, we further derived a local equilibrium relationship between $PKE$ and $LPPE_1$ (integrated between isobaric surfaces of $p_1$ and $p_2$) to explain this phenomenon (refer to Appendix A for a detailed derivation). We refer to this relationship as shear energy balance, for it describes the relationship between the vertical shear of $PKE$ and the horizontal gradient of $LPPE_1$. According to the relationship,

$$\frac{1}{2} \left( u_g^2 + v_g^2 \right) \bigg|_{p_1}^{p_2} \propto \frac{\langle V_\theta \rangle}{f} \times \nabla_\theta LPPE_1 \cdot k,$$

where $\langle \cdot \rangle = \frac{1}{p_2-p_1} \int_{p_1}^{p_2} dp$ denotes the vertical average between pressure levels $p_1$ and $p_2$ ($p_2 > p_1$), $u_g$ and $v_g$ are the zonal and meridional components of the geostrophic wind, and $k$ is the vertical unit vector. Approximating the extratropical mid-upper-level winds as having

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only a zonal component, when \( \text{LPPE}_1 \) decreases with latitude, the \( \text{PKE} \) of the west wind increases with height, and the \( \text{PKE} \) of the east wind decreases with height, and vice versa. Over the North Atlantic, \( \langle u_g \rangle \) and \( f \) are positive. Where there is an anomalous positive (negative) meridional gradient in \( \text{LPPE}_1 \), there will be an anomalous decrease (increase) with height of anomalous \( \text{PKE} \), consistent with Fig. 7d.

From the perspective of energetics, the energy conversion between PPE and PKE is strongest in the region with the largest PPE gradient, corresponding to the maximum PKE anomalies. The instantaneous \( \text{PC}_k \) can indicate this physical process. From Fig. 10b, the accumulated \( \text{PC}_k \) associated with the NAO does not conform to the spatial distribution of PKE anomalies. Whereas the instantaneous \( \text{PC}_k \) shows meridional characteristics, which can explain the quadrupole PKE anomaly (Fig. 10a). The negative (positive) anomaly region of instantaneous \( \text{PC}_k \) corresponds to the positive (negative) anomaly region of PKE. Around 15°N and 50°N (25°N and 75°N), there are instantaneous ascending (descending) motion anomalies (Fig. 10c), and the conversion from PPE to PKE increases (decreases), leading to the strengthening (weakening) of westerly winds.

![Images of plots](image1.png)

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Fig. 10. (a) Correlation coefficients between the NAOI and instantaneous $PC_k$ at 1000-150 hPa. (b) As in (a), but for the accumulated $PC_k$. The black dotted areas in (a) and (b) denote significant values at the 90% confidence level using the $N_{eff}$. (c) Correlation coefficients between the NAOI and regionally (70°–5°W) zonal mean instantaneous $PC_k$ (shading). Correlation coefficient vectors between the NAOI and regionally (70°–5°W) zonal mean instantaneous meridional circulation (arrows). The unit vector is in the upper left corner. The black arrows denote the significant correlation coefficient vectors at the 90% confidence level using the $N_{eff}$. The areas surrounded by black lines denote significant values at the 90% confidence level using the $N_{eff}$.

The vertical distribution of energy can help us further understand the circulation structure of NAU and NAO. Notably, the NAO is a deep system while the NAU is a shallow system (Fig. 2). The difference is related to the characteristics of the vertical distribution of energy conversion. The uniformity of $PC_k$ gradually disappears with height (Fig. 8c), therefore the anomalies of PKE vary greatly with height (Fig. 7b). This leads to the difference between the circulation at high and low levels. For the NAO mode, the meridional anomaly of $PC_k$ exists at both high and low levels (Fig. 10c), therefore the pattern of PKE anomaly has little difference between high and low levels (Fig. 7d).

The difference in the energy distributions between the two modes can be attributed to the spatial nature of oceanic forcing. The PPE gradient is larger for the meridional modes, and the anomaly of PKE increases with height, due to the energy balance relationship in Eq. (14). Conversely, for the uniform modes, the PPE gradient is small, therefore kinetic energy anomalies are not deep structures. Therefore, as the state of the ocean changes, the two structures of the atmosphere alternate with each other.

In summary, similar to the characteristics of SSTA and PPE anomalies, the energy conversion and PKE anomalies exhibit two distinct patterns: a uniform pattern and a meridional pattern. During the NAU+ (NAU–), accumulated ascending (descending) motion anomalies result in negative (positive) $PC_k$ anomalies, converting PPE (PKE) to PKE (PPE) anomalously. This leads to an increase (decrease) in atmospheric PKE over the North Atlantic, forming a basin-scale anomalous anticyclonic (cyclonic) circulation that enhances (weakens) the atmospheric circulation at lower levels over the North Atlantic. During the NAO+, the instantaneous ascending (descending) motion around 15°N and 50°N (25°N and 75°N) leads to the anomalous energy conversion that strengthens (weakens) the westerly wind intensity. The reverse holds for the NAO–.
6. Delayed and feedback effect in the multidecadal cycle

Different from previous studies on PPE (Dong et al. 2017; Zhang et al. 2019; Zhang et al. 2023), this study distinguishes between instantaneous and accumulated (or time-integrated) physical processes. Actually, they reflect two kinds of relationship between forcing and response. The instantaneous quantity corresponds to an immediate response, for the forcing and response are almost synchronous. The accumulated quantity corresponds to a delayed effect, for the response lags the forcing. In Fig. 8, the positive AMO is associated with negative instantaneous $PC_k$ (instantaneous ascending motion) but positive accumulated $PC_k$ (accumulated descending motion). Fig. 11a gives the lead-lag relationship between the AMO and the uniform $PC_k$ index (in instantaneous form). The AMO is negatively correlated with uniform $PC_k$ simultaneously, and positively lagging the uniform $PC_k$ by 20~30 years. Notably, it is in negative AMO phase when the AMO$^-$ lags 20-30 years. Therefore, the accumulated descending motion associated with the AMO$^-$ actually comes from the influence of the previous negative phase, which has a delayed effect. In addition, the opposite effect of accumulated $PC_k$ and instantaneous $PC_k$ constitutes negative feedback, which is analyzed in detail later. Similarly, the accumulated tripole precipitation associated with the NAT also comes from the tripole precipitation (in instantaneous form) leading by 6~7 years (Fig. 11b), when the NAT just starts developing.
Fig. 11. (a) Lead-lag correlation between the AMOI and the uniform PC<sub>k</sub> index (defined as the areal weighted mean PC<sub>k</sub> in 0°-60°N, 70°-10°W) (red solid line). The autocorrelation of the AMOI (black solid line). (b) As in (a), but for the NATI and tripole precipitation index. Due to the complexity of the pattern, the tripole precipitation index is directly defined as the projection of the precipitation field on the field of Fig. 6b. The correlations are based on the smoothed indices. The dashed lines indicate the 90% confidence level using the N<sub>eff</sub>.

We observe that when instantaneous (accumulated) processes take effect, they correspond to modes exhibiting larger (smaller) variances (Fig. 12). Specifically, the explained variance of PPE anomalies by the uniform PPE pattern (about 34%) surpasses that by the meridional PPE pattern (about 16%), with diabatic heating driven by instantaneous precipitation forcing. Similarly, the PKE anomaly variance of the NAO mode (about 32%) is larger than that of the NAU mode (about 10%), and the energy conversion process of the NAO is predominantly influenced by instantaneous vertical motion. Conversely, the accumulated process corresponds to the NAT and the NAU, characterized by smaller
variances. This suggests that whether the relationship between forcing and response is synchronous or delayed may depend on the magnitude of the anomaly.

Fig. 12. Diagram of the energetics mechanisms of the atmospheric multidecadal variability forced by the North Atlantic SSTA modes. The percentage denotes the average variance explanation of the index on the corresponding variable in the North Atlantic region.

However, why does a high-variance AMO and PPE uniformity lead to a low-variance NAU, while a low-variance NAT and PPE tripole lead to a high-variance NAO? We found that this may be related to the positive (negative) feedback process that exists within the NAO (NAU).

Riviere and Orlanski (2007) discovered a quadrupole low-level humidity pattern associated with the NAO, emphasizing that these humidity anomalies facilitate the low-level destabilization anomalies. The pattern is spatially similar to the $PC_k$ quadrupole in Fig. 10a. Thus, there is a positive feedback process associated with the NAO: $PC_k$ quadrupole→NAO→humidity quadrupole at lower level→destabilization at lower level→vertical motion quadrupole→$PC_k$ quadrupole. This feedback mechanism helps us to further understand the formation of the NAO and the interaction between the NAO and NAT. For the NAU mode, as mentioned above, the effects of accumulated $PC_k$ and instantaneous $PC_k$ are opposite. This forms negative feedback in the NAU: positive anomaly of accumulated $PC_k$→NAU→anomalous cyclone and ascending motion→negative anomaly of instantaneous $PC_k$. These

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two feedback mechanisms explain the inconsistencies in the variance explanation of the oceanic and atmospheric modes (Fig. 12). The positive feedback loop can cause strong development of the NAO circulation, although the amplitude of oceanic forcing is relatively small. On the contrary, the negative feedback prevents the NAU circulation from developing strongly.

7. Summary and discussion

Supporting the previous model results (Sun et al. 2015), a cyclic ocean pattern involving the AMO and NAT is detected in observational data: \( \text{AMO}^+ \rightarrow \text{NAT}^- \rightarrow \text{AMO}^- \rightarrow \text{NAT}^+ \rightarrow \text{AMO}^+ \cdots \), with a phase lag of approximately 15~20 years. An atmospheric mode characterized by basin-scale uniform SLP anomalies, termed as the NAU, is found to be associated with the AMO. The NAU+ (NAU−) exhibits a large-scale anomalous anticyclone (cyclone) circulation at lower-level over the North Atlantic basin, which is distinct from the NAO. The lead-lag correlation reveals a cyclic sequence of the atmospheric modes: \( \text{NAU}^- \rightarrow \text{NAO}^- \rightarrow \text{NAU}^+ \rightarrow \text{NAO}^+ \rightarrow \text{NAU}^- \cdots \), which is associated with the cycle of the ocean modes.

We explored the energetics driving the atmospheric cyclic pattern in the North Atlantic, focusing on the North Atlantic Ocean diabatic heating and atmospheric energy conversion processes. Summarizing the energetics processes reveals the life cycle of the atmospheric circulation oscillation (Fig. 13). When the North Atlantic Ocean is in the NAT+, the SSTA induce accumulated tripolar precipitation anomalies, leading to tripolar PPE anomalies through latent heat release. The tripole PPE anomaly leads to quadrupole PKE anomalies through quadrupole instantaneous vertical motion. The latitudes where the strongest energy conversion occurs lie between the positive and negative PPE anomalies. Consequently, the PKE anomalies are situated in the region with the greatest PPE gradient, and they are most pronounced at the upper levels. This aligns with the energy balance relationship derived from the thermal wind relationship. The positive (negative) PKE anomaly corresponds to the westerly (easterly) anomaly and the strengthening (weakening) of background circulation, which forms the anomalous anticyclones over the warm SSTA (around 40°N) and the anomalous cyclones over the cold SSTA (around 20°N and 60°N). Thus, the atmospheric circulation of the NAO+ is formed.

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Fig. 13. Schematic diagram of the energetics of the multidecadal cycle involving the NAO and NAU in the North Atlantic. Solid (dashed) black arrows denote positive (negative) PKE anomalies and strengthened (weakened) atmospheric horizontal circulation. The prefix “A-“ (“I-“) denotes accumulated (instantaneous) process. The horizontal (vertical) shading denotes SST (PPE) anomalies. The red (blue) shading denotes the positive (negative) anomalies of SST and PPE. The green (brown) horizontal shading and cloud denote the positive (negative) anomaly of precipitation. The red (blue) arrows denote the anomalous descending (ascending) vertical motion and positive (negative) $PC_k$ anomalies.

Transitioning from the NAT$^+$ to the AMO$^+$, the SSTA induce simultaneous increase of basin-scale precipitation, transforming PPE into uniform positive anomalies through the instantaneous release of latent heat. Accumulated descending motion leads to positive accumulated $PC_k$ anomalies, reducing energy conversion from PPE to PKE. The atmospheric circulation at the lower levels over the North Atlantic weakens, large-scale anomalous cyclonic circulation and low-pressure occurs, and the NAU$^-$ is formed. Subsequently, the AMO$^+$ transforms into the NAT$^-$, and the oscillation continues, but reverses to the opposite
phase, leading to the NAO+. Over a quasi-70-year cycle, the two modes have about a quarter phase difference, and their energetics physical processes alternate, resulting in the periodic occurrence of the NAU and NAO circulations.

The instantaneous and accumulated (or time-integrated) physical processes are distinguished in the multidecadal cycle. For the instantaneous process, the forcing and response are almost synchronous. For the accumulated process, the response lags the forcing. When instantaneous (accumulated) processes take effect, they correspond to modes exhibiting larger (smaller) variances, namely the AMO and NAO (NAT and NAU). This indicates that whether the relationship between forcing and response is synchronous or delayed may depends on the magnitude of the anomaly. Additionally, the inconsistencies in the variance explanation of the oceanic and atmospheric modes are related to feedback mechanisms in the atmospheric energy conversion. The positive feedback loop can cause strong development of the NAO circulation, although the amplitude of oceanic forcing is relatively small. On the contrary, the negative feedback prevents the NAU circulation from developing strongly.

The SST influences PPE through the diabatic heating from precipitation, and PPE anomalies, in turn, influence atmospheric circulation through energy conversion induced by vertical motion. The partial correlations between SST modes and PPE patterns (PPE patterns and PKE patterns) by removing the PKE (SST) signals are still significant (Table 1). However, the partial correlations between the AMO and uniform PKE anomalies, as well as between the NAT and the quadrupole PKE anomalies become insignificant by removing the PPE signal. This demonstrate that diabatic forcing cannot directly influence atmospheric circulation variation without influencing PPE, which further supports that PPE serves as a crucial bridge between the two. The energetics mechanisms provide evidence supporting the attribution of atmospheric multidecadal variations in the North Atlantic to the North Atlantic oceanic forcings, which holds potential for enhancing the skills of decadal modeling and prediction for the atmosphere in the region. In addition, it is essential to acknowledge that besides North Atlantic Ocean forcing, other factors, such as greenhouse gas forcing (Shindell et al. 1999; Fyfe et al. 1999) or remote forcing from tropical Indian and Pacific Oceans, may also contribute (Hoerling et al. 2001; Bader and Latif 2003).

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Table 1. Correlation coefficients (second column) and partial correlation coefficients (third column) between the SST modes, PPE patterns, and PKE patterns. The index of uniform PPE anomaly is as in the caption of Fig. 4. The tripolar PPE anomaly index is defined as $\frac{1}{2}PPE_{80^\circ-10^\circ W,35^\circ-45^\circ N} - \frac{1}{4}(PPE_{80^\circ-10^\circ W,10^\circ-25^\circ N} + PPE_{70^\circ-10^\circ W,50^\circ-75^\circ N})$, where the PPE is integrated at 1000-300 hPa. The index of uniform PKE anomaly is defined as average PKE anomaly at $0^\circ-70^\circ N, 80^\circ W-0^\circ$ integrated at 1000-850 hPa. The quadrupolar PKE anomaly index is defined as $\frac{1}{4}(PKE_{70^\circ-5^\circ W,40^\circ-60^\circ N} + PKE_{70^\circ-5^\circ W,5^\circ-20^\circ N}) - \frac{1}{4}(PKE_{70^\circ-5^\circ W,60^\circ-80^\circ N} + PKE_{70^\circ-5^\circ W,20^\circ-40^\circ N})$, where the PKE is integrated at 1000-100 hPa. All the correlations are based on indices smoothed by a 11-year low-pass Gaussian filter. The postfix stars denote the significant value at the 90% confidence level using the $N_{eff}$.

The analysis in this paper is based on reanalysis data, which has some uncertainties due to the sparsity of the early observation. It is necessary to conduct sensitivity experiments to confirm the mechanisms involved, including the atmospheric response to the oceanic multidecadal variability and the energetics mechanisms. Moreover, the interaction between the atmosphere and ocean in the North Atlantic is bidirectional. The atmosphere over the North Atlantic can affect the North Atlantic Ocean through surface heat flux or wind stress (Halliwell 1997; Marshall et al. 2001). Traditionally, the NAT has been considered an instantaneous response to the NAO (Wu et al. 2009; Deser et al. 2010; Álvarez-García et al. 2008), while the oscillation of AMOC and the AMO are viewed as delayed responses to the NAO (Delworth and Greatbatch 2000; Eden and Jung 2001; Álvarez-García et al. 2008; Li et al. 2013; Sun et al. 2015; Delworth et al. 2017). While our study explores unidirectional mechanisms from ocean to atmosphere, future research should investigate bidirectional interaction energetics. This requires applying ocean PPE theory to analyze how the atmosphere influences North Atlantic Ocean circulation, thereby driving the evolution of multidecadal ocean modes.

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Data Availability Statement.

All the datasets used in this paper are available publicly. The atmospheric data are from Twentieth Century Reanalysis Version 3 dataset (https://psl.noaa.gov/data/gridded/data.20thC_ReanV3.html). The SST data are from the Extended Reconstructed SST Version 5 dataset (https://psl.noaa.gov/data/gridded/data.noaa.ersst.v5.html). The SLP data are from the Hadley Centre Sea Level Pressure dataset (https://www.metoffice.gov.uk/hadobs/hadslp2/).

APPENDIX A

Derivation of the shear energy balance

First, we agree on the following formulas in an isobaric coordinate system: the global mean of a variable $A$ on isobaric surface is $ar{A}$, and the perturbation $A^\sim = A - \bar{A}$; the vertical mean of $A$ between isobaric surface $p_1$ and $p_2$ is $\langle A \rangle = \frac{1}{p_2-p_1} \int_{p_1}^{p_2} Adp$, and the vertical perturbation $A^\wedge = A - \langle A \rangle$; the geostrophic wind $V_h = (u_g, v_g)$; the horizontal gradient operator is $\nabla_h$.

The thermal wind relationship is

$$\frac{\partial V_h}{\partial p} = \frac{R}{f p} \nabla_h T \times \mathbf{k}. \quad (A1)$$

The dot product of Eq. (A1) and $V_h$ is

$$V_h \cdot \frac{\partial V_h}{\partial p} = \frac{R}{f p} V_h \cdot (\nabla_h T \times \mathbf{k})$$

$$\frac{1}{2} \frac{\partial (u_g^2 + v_g^2)}{\partial p} = \frac{R}{f p} (V_h \times \nabla_h T) \cdot \mathbf{k}. \quad (A2)$$

The global mean of Eq. (A2) is

$$\frac{1}{2} \frac{\partial (u_g^2 + v_g^2)}{\partial p} = \frac{R}{p} \left( \frac{1}{f} V_h \times \nabla_h T \right) \cdot \mathbf{k}. \quad (A3)$$

Eq.(A2) minus Eq. (A3) equals the local perturbation:

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\[ \frac{1}{2} \frac{\partial (u_g^2 + v_g^2)}{\partial p} = \frac{R}{p} \left( \frac{1}{f} \mathbf{V}_h \times \nabla_h T - \frac{1}{f} \mathbf{V}_h \times \nabla_h T \right) \cdot \mathbf{k}. \quad (A4) \]

The vertical integration of Eq. (A4) is

\[ \frac{1}{2} \left( u_g^2 + v_g^2 \right) \bigg|^{p_2}_{p_1} = R \int_{p_1}^{p_2} \left[ \left( \frac{1}{f} \mathbf{V}_h \times \nabla_h T \right) - \frac{1}{f} \mathbf{V}_h \times \nabla_h T \right] dlnp \cdot \mathbf{k} \]

\[ = R \ln \frac{p_2}{p_1} \left[ \frac{1}{f} \mathbf{V}_h \times \nabla_h T \right] \cdot \mathbf{k}. \]

Using the formula \( \langle AB \rangle = \langle A \rangle \langle B \rangle + \langle A^* B^* \rangle \), it could be further written as

\[ \frac{1}{2} \left( u_g^2 + v_g^2 \right) \bigg|^{p_2}_{p_1} = R \ln \frac{p_2}{p_1} \left[ \frac{1}{f} \langle \mathbf{V}_h \rangle \times \nabla_h \langle T \rangle + \frac{1}{f} \langle \mathbf{V}_h \rangle \times \nabla_h \langle T^\sim \rangle \right] - \frac{1}{f} \mathbf{V}_h \times \nabla_h \langle T^\sim \rangle \cdot \mathbf{k}. \]

Obviously, \( \nabla_h \langle \mathbf{T} \rangle = 0 \), therefore \( \nabla_h \langle T \rangle = \nabla_h \langle T^\sim \rangle \). Then \( \langle LPPE_1 \rangle = \frac{1}{\gamma_d} \int_{p_1}^{p_2} T^\sim dp = \frac{p_2 - p_1}{\gamma_d} \langle T^\sim \rangle \implies \langle T^\sim \rangle = \frac{\gamma_d}{p_2 - p_1} LPPE_1. \)

\[ \frac{1}{2} \left( u_g^2 + v_g^2 \right) \bigg|^{p_2}_{p_1} = R \ln \frac{p_2}{p_1} \left[ \frac{\gamma_d}{p_2 - p_1} \left( \frac{1}{f} \langle \mathbf{V}_h \rangle \times \nabla_h \langle LPPE_1 \rangle \right) + \frac{1}{f} \langle \mathbf{V}_h \rangle \times \nabla_h \langle T^\sim \rangle \right] \]

\[ = \left( \frac{\gamma_d}{p_2 - p_1} \frac{1}{f} \langle \mathbf{V}_h \rangle \times \nabla_h \langle LPPE_1 \rangle \right) \cdot \mathbf{k}. \]

It leads to the following proportional relationship

\[ \frac{1}{2} \left( u_g^2 + v_g^2 \right) \bigg|^{p_2}_{p_1} \propto \frac{1}{f} \langle \mathbf{V}_h \rangle \times \nabla_h \langle LPPE \rangle \cdot \mathbf{k}. \quad (A5) \]

Eq. A5 is different from the thermal wind relationship (Eq. A1) in expression form. Eq. A1 describes the relationship between velocity and temperature, in this formula, velocity becomes kinetic energy and temperature becomes potential energy. It helps us understand the distribution and balance of energy. Second, Eq. A5 shows that the vertical change in kinetic energy is not only related to the gradient of the LPPE, but also to the magnitude of the velocity itself.

**APPENDIX B**

**Lead-lag relationship between the AMO and NAT in observational data**

Fig. A1 displays the temporal evolution of the two multidecadal SST modes. The positive pole of the NAT\(^+\) gradually expands over time, changing the NAT\(^+\) into the AMO\(^+\). Subsequently, negative SST modes emerge at the mid-latitudes in the North Atlantic Ocean,
changing the AMO into the NAT. The NATI in this study is defined using the boxes in Figs. A1a, e. Lead-lag correlation shows that NATI leads (lags) AMOI about 12 (21) years with significant positive (negative) correlation (Fig. A1f).

Fig. A1. (a)-(e) Lead-lag correlation coefficient maps of the AMO and detrended low-passed SSTAs. The black dotted areas denote significant values at the 90% confidence level using the $N_{\text{eff}}$. The boxes in (a) and (e) is used to define the NATI. (f) Lead-lag correlation coefficients between the AMOI and NATI (black solid line). The dashed lines indicate the 90% confidence level using the $N_{\text{eff}}$. In (a)-(f), the positive (negative) lags denote that the AMOI leads (lags) the corresponding variables.

APPENDIX C

Longwave radiation, shortwave radiation and sensible heat terms of the NAO mode

Fig. A2 displays the sensible heat flux and radiation flux at surface associated with the NAO. The tripole PPE anomalies cannot be explained by these terms. Note that the instantaneous shortwave radiation flux exhibits a tripole pattern, but the latitudes of the anomalies do not appear to align with those of tripole PPE anomalies in Fig. 4b.
Fig. A2 (a) Correlation coefficient map between the NAOI and instantaneous sensible heat flux. (b)-(c) as in (a), but for instantaneous net longwave radiation and instantaneous net longwave radiation. (d)-(f) as in (a)-(c), but for accumulated sensible heat, accumulated net longwave radiation and accumulated net shortwave radiation. The black dotted areas denote significant values at the 90% confidence level using the $N_{eff}$.

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