

The Impact of Wintertime NPO on ENSO Triggering

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1 **The Impact of Wintertime NPO on ENSO Triggering**

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Abstract

Recent studies highlight the key role of midlatitude climate variability in modulating tropical El Niño–Southern Oscillation (ENSO) activity. This study investigates the impact mechanisms of wintertime North Pacific Oscillation (NPO) in triggering ENSO events in the following year and its dependence on the structural characteristics of the NPO. When the NPO is stronger and located further south, anomalous southwesterly winds in the lower troposphere weaken the background trade winds and enhance local sea surface temperature (SST) anomalies. These SST anomalies are further amplified through the wind–evaporation–SST (WES) feedback, leading to the formation of the North Pacific Meridional Mode (NPMM). Through the seasonal footprinting mechanism (SFM), these SST and wind anomalies propagate toward the tropics, facilitating the westerly wind burst and ENSO onset. In contrast, weaker, more northerly and zonally elongated NPO anomalies confine SST anomalies to the subtropics, where they gradually dissipate. In the upper troposphere, a northeast–southwest tilted NPO-related geopotential height anomaly favors the southward propagation of Rossby wave, supporting the development of tropical anticyclonic circulation anomaly and associated vertical easterly wind shear, which also contribute to ENSO triggering. Therefore, the structure of NPO is crucial for its impact on ENSO. Further analysis reveals that the oceanic feedback can modulate the evolution of NPO structure. The NPO-related wind anomalies adjust air–sea heat exchange and SST patterns, which exert feedback onto the atmosphere by changing diabatic heating and transient eddy forcing. Diabatic heating dominantly contributes to maintaining the circulation anomaly of the NPO at lower layers, while transient eddy vorticity forcing sustains the equivalent barotropic structure of the NPO. These findings clarify the coupled ocean–atmosphere processes linking wintertime NPO to ENSO, providing insights into extratropical–tropical interactions and ENSO predictability.

Keywords: North Pacific Oscillation; El Niño–Southern Oscillation; Midlatitude ocean–atmosphere interaction; Transient eddy feedback

1 Introduction

El Niño-Southern Oscillation (ENSO) is one of the most significant natural climate variability signals in the tropical ocean-atmosphere system and acts as an important source of interannual climate variability. ENSO not only affects the climate in the tropics but also influences the global climate through atmospheric teleconnections (Lau 1997; Alexander et al. 2002; McPhaden et al. 2006). The development and evolution of ENSO are primarily determined by tropical ocean-atmosphere interaction dynamics (Bjerknes 1969; Eisenman et al. 2005; Cai et al. 2020). However, many studies have revealed that the extratropical climate variability can also influence ENSO activity on interannual or decadal time scales (Lau and Yang 1996; Vimont et al. 2003b; Chang et al. 2007; Yu and Kim 2010; Ding et al. 2017).

Extratropical atmospheric variability modulates the tropical Pacific mean state, thereby influencing the triggering and evolution of ENSO (Ding et al. 2015; Chakravorty et al. 2021; Capotondi et al. 2023). Specifically, as a significant atmospheric variability during the winter in the North Pacific, the North Pacific Oscillation (NPO) can impact ENSO activity in the following year on interannual scales. Statistically, the NPO is significantly correlated with ENSO activity when the ENSO signal lags the NPO by approximately 12 to 15 months, and this relationship remains consistent across different regions of SST anomalies (Fig. 1a). Previous studies have demonstrated that the cyclonic circulation anomaly associated with the positive phase of the NPO during winter generates anomalous southwesterly winds (Vimont et al. 2001; Anderson 2003). These anomalous southwesterly winds reduce the upward turbulent heat flux from ocean to atmosphere, resulting in localized SST warming. The SST footprint formed in winter leads to the development of the North Pacific Meridional Mode (NPMM) in the following spring (Vimont et al. 2003a, b; Vimont 2004). Through the so-called seasonal footprinting mechanism (SFM), the associated SST and wind anomalies subsequently propagate into the tropical central Pacific in late spring, favoring the development of westerly wind bursts over the western tropical Pacific in autumn, a key process leading to El Niño onset (McPhaden et al. 2006). In addition to low-level air-sea coupling processes, geopotential height anomalies associated with the NPO can also directly stimulate equatorward wave propagation through atmospheric teleconnections (Zhao et al. 2023a, b), leading to anomalous geopotential heights and

winds in the tropics. These anomalies can trigger anomalous convection and precipitation in the lower troposphere, which can further influence ENSO activity through the Gill-type circulation response (Gill 1980). However, the formation of the upper-level circulation anomaly associated with the NPO and its relationship with the southward propagation of wave energy underlying this mechanism remain unclear.

Previous studies have primarily focused on how the NPO initiates ENSO, particularly emphasizing weakened background trade winds and associated turbulent heat flux changes. Our preliminary analysis of selected NPO and ENSO events indicates that ENSO occurs following approximately 62.5% of NPO events (Table 1). Although the basic relationship between NPO events and subsequent ENSO development has been established, it remains unclear why certain NPO events fail to effectively initiate ENSO, suggesting that the spatial structure and intensity of the NPO may have potential roles in facilitating ENSO activity. Park et al. (2013) identified that the spatial structure of the NPO, along with changes in wind speed and net heat flux in the northeast Pacific, are key factors through which the NPO influences ENSO activity via the SFM. These factors determine the SST anomaly footprint in the northeast Pacific during late spring and summer. Thus, the definition of the NPO index influences the revelation of the relationship between NPO and ENSO (Chen and Wu, 2018). When the NPO index well captures the subtropical cyclonic anomalies that induce warm SST anomalies, the connection between the NPO and ENSO becomes stronger. Wang and Bao (2024) proposed that the southern lobe cyclone structure of the NPO can be divided into typical and atypical modes. Only the atypical mode, which overlaps with the Hawaiian region, can influence the NPM and further promote ENSO activity. Other studies found that stronger positive-phase NPO forcing enhances El Niño development, while a westward shift in the NPO's negative phase weakens its effect on subtropical northeast Pacific SST anomalies, leading to a weaker La Niña event (Chen and Yu 2020; Hu et al. 2024). These studies have deepened our understanding of the complex relationship between North Pacific SST variations and ENSO. However, the role of the structural characteristics of the NPO in impacting subsequent ENSO evolution has not been fully considered.

These findings suggest that the relationship between the NPO and the subsequent ENSO is largely dependent on the NPO structure, driven by internal climate variability

(Linkin and Nigam 2008; Tanaka et al. 2016; Sung et al. 2022). It is further necessary to explore the factors that cause these structural differences in the NPO. Moreover, the NPO, as an anomaly in mid-latitude atmospheric circulation, is influenced by both thermal and eddy-driven forcings (Lau and Nath 1991; Wang et al. 2017; Joh and Di Lorenzo 2019; Huang et al. 2024). Although the diabatic heating caused by SST anomalies is relatively weak and mainly confined to the lower troposphere, it can still have a significant impact on the maintenance of mid-latitude circulation anomalies in the lower troposphere. Additionally, the mid-latitude atmosphere exhibits strong baroclinicity, with active synoptic-scale transient eddy activities. The position of the NPO coincides with the mid-latitude oceanic frontal zone, which features storm tracks (Ren et al. 2010; Liu et al. 2014; Wang et al. 2017). Transient eddies can systematically transport heat and momentum through nonlinear wave-mean flow feedback, driving and maintaining the circulation anomalies, especially in the higher troposphere (Xiang and Yang 2012; Fang et al. 2022). Therefore, mid-latitude SST anomalies can influence the structure and evolution of the NPO by altering both diabatic heating and transient eddy forcing. Understanding the oceanic feedback to the atmosphere is important for elucidating how structural variations in the NPO influence ENSO initiation.

This study aims to investigate the specific mechanisms by which the structural characteristics of wintertime NPO anomalies affect ENSO triggering and to analyze the structural evolution of the NPO. The manuscript is structured as follows: Section 2 presents the data and research methods used, Section 3 categorizes the impact of the NPO on subsequent climate processes into two types and analyzes the evolutionary pathways of each event category separately, Section 4 examines the influence of oceanic feedback on the structure of the NPO and the last section is a summary and discussion for further study.

2 Data and methods

The analytical data used in the present study are derived from the European Centre for Medium-Range Weather Forecasts (ECMWF) ERA5 monthly reanalysis product (Hersbach et al. 2020), with a horizontal resolution of $1^\circ \times 1^\circ$. To further ensure the robustness of our findings, a cross-dataset consistency check is conducted using independent datasets from NCEP and ERSSTv5. The surface variables include SST, sea level pressure (SLP), latent heat flux (LHF, upward positive), sensible heat flux (SHF,

upward positive), and total precipitation. The multi-level variables include geopotential height (Z), horizontal wind fields, temperature (T), and specific humidity, with the stream function calculated from Z. Additionally, to analyze the impacts of transient eddy heat and vorticity forcing, daily ERA5 reanalysis products are used in Section 4, including the variables of SST, horizontal wind fields, vertical velocity (ω), and temperature. A 3-11 day Butterworth bandpass filter is applied to the daily data to extract the synoptic-scale eddies. The study period spans from 1950 to 2023. Prior to variable analysis, all data are detrended and processed to remove the seasonal cycles. Prior to the calculation of the NPO index and the composite analysis, the linear influence of concurrent ENSO, represented by the cold tongue index (defined as area-averaged SST anomalies over the region 180°–90°W, 6°S–6°N, see Vimont et al. 2003b), is removed to eliminate its effect on the evolution of the NPO and subsequent processes. A two-tailed Student's t-test is applied in this study, and a p-value less than 0.05 indicates the anomalies are significant. Because two types of event sequences exhibit temporal autocorrelation, the effective degrees of freedom (EDF) are calculated prior to composite significance testing (Zwiers and Von Storch 1995). To further assess the robustness of the composite patterns, a bootstrap resampling test with 10,000 iterations is conducted for key variables, and the resulting empirical confidence intervals are used to evaluate the stability of the anomalies.

The spatial pattern of the NPO is defined as the second leading mode derived from empirical orthogonal function (EOF) analysis of the standardized monthly SLP anomalies over the North Pacific region (100°E–80°W, 20°N–70°N) during the winter (DJF), accounting for 21.34% of the total variance. The corresponding annual NPO index is obtained by projecting the winter SLP anomalies onto this spatial mode. The Niño indices (Niño3, Niño4, Niño3.4) used in this study are calculated following the method of the National Center for Atmospheric Research (NCAR) Niño SST Indices. The maximum Eady Growth Rate (EGR) (Eady 1949; Lindzen and Farrell 1980) is used in this study to characterize atmospheric baroclinicity, which can be calculated using the formula $\sigma_{BI} = 0.3098 \cdot |f| \cdot \left| \frac{\partial u}{\partial z} \right| / N$, where u is the zonal wind, f is the Coriolis parameter, and N is the Brunt-Väisälä frequency. The Rossby wave activity flux (WAF) (Takaya and Nakamura 2001) is used to track the propagation path of Rossby wave energy, and it can be calculated using the following equation:

$$170 \quad \text{WAF} = \frac{P \cos \varphi}{2|\vec{V}|} \left(\begin{aligned} &\frac{u}{a^2 \cos^2 \varphi} \left[\left(\frac{\partial \psi'}{\partial \lambda} \right)^2 - \psi' \frac{\partial^2 \psi'}{\partial \lambda^2} \right] + \frac{v}{a^2 \cos \varphi} \left[\frac{\partial \psi'}{\partial \lambda} \frac{\partial \psi'}{\partial \varphi} - \psi' \frac{\partial^2 \psi'}{\partial \lambda \partial \varphi} \right] \\ &\frac{u}{a^2 \cos \varphi} \left[\frac{\partial \psi'}{\partial \lambda} \frac{\partial \psi'}{\partial \varphi} - \psi' \frac{\partial^2 \psi'}{\partial \lambda \partial \varphi} \right] + \frac{v}{a^2} \left[\left(\frac{\partial \psi'}{\partial \varphi} \right)^2 - \psi' \frac{\partial^2 \psi'}{\partial \varphi^2} \right] \end{aligned} \right) \quad (1)$$

171 where $\vec{V} = (u, v)$ is the horizontal wind field, P is the normalized pressure ($p/1000$
 172 hPa), a is the Earth's radius, λ and φ denote longitude and latitude, respectively,
 173 and $\psi = gz/f$ is the stream function, with g being the gravitational acceleration
 174 and the prime indicating anomalies relative to the monthly mean.

175 **3 The impact mechanism of the NPO on ENSO triggering**

176 In this section, we investigate how different structural characteristics of the NPO
 177 influence ENSO initiation by examining the interface air–sea interactions and upper-
 178 level energy propagation processes.

179 **3.1 Role of the ocean-atmosphere coupling processes**

180 The regression of SST and concurrent SLP anomalies onto the NPO index is first
 181 examined (Fig. 1b). During the boreal winter, the spatial pattern of the positive NPO
 182 phase is characterized by a meridional dipole mode, with an anomalous high pressure
 183 located north of 45 °N and an anomalous low pressure centered around 35°N over the
 184 mid-to-high latitudes of the North Pacific. Associated with the negative SLP anomaly,
 185 the horizontal wind anomalies weaken (strengthen) the background trade winds,
 186 inducing characteristic SST anomalies concurrently. These SST anomalies are
 187 characterized by cooling in the central and western North Pacific and warming
 188 extending southwestward from the western coast of California. Lag regression analysis
 189 further reveals significant NPO-related positive SST anomalies near the equatorial
 190 Pacific during the subsequent winter, displaying a typical El Niño pattern (Fig. 1c). This
 191 result indicates the potential of the NPO to trigger ENSO activity in the following year
 192 through oceanic and atmospheric processes from the midlatitudes to the tropics.

193 As discussed above, not all NPO events are followed by significant ENSO activity.
 194 This is confirmed by our preliminary analysis, which indicates that about 62.5% of
 195 NPO events are followed by ENSO (Table 1). Whether ENSO initiation occurs depends
 196 not only on the initial tropical ocean thermodynamic state but may also be associated

with variations in the spatial structure and evolution of NPO anomalies. To clearly investigate the key characteristics and processes that determine the impact of NPO on ENSO initiation, we categorize the observed NPO events into two groups based on their subsequent ENSO occurrence: events followed by significant ENSO activity (NPO-ENSO) and those not followed by ENSO events (NPO-NONE). Given that ENSO asymmetry (El Niño vs. La Niña) has a negligible influence on this study (Fan et al. 2023; Anderson et al. 2013a), the positive NPO phases followed by El Niño and the negative NPO phases followed by La Niña are combined into the NPO-ENSO category. Conversely, NPO phases without subsequent ENSO events are grouped into the NPO-NONE category. The detailed event classifications are summarized in Table 1.

By comparing the distribution characteristics of the low-level circulation fields and SST anomalies between the two types of events, it is found that the southern lobe anomalous cyclonic circulation of the NPO gradually strengthens and expands during winter, with corresponding southwesterly wind anomalies extending equatorward in NPO-ENSO events (Fig. 2a, c, e). Consequently, significant warm (cold) SST anomalies develop on the southeastern (southwestern) flank of the negative SLP anomaly due to reduced (increased) upward heat flux. The negative SLP anomaly persists until March, accompanied by westerly wind anomalies that propagate to the equatorial western Pacific via Wind–Evaporation–SST (WES) feedback (Xie and Philander 1994; Martinez-Villalobos and Vimont 2017). This process leads to the intensification of positive SST anomalies, resulting in a well-defined NPM during spring. By May of the following year, significant ENSO signals emerge in the equatorial central and eastern Pacific.

To better explore how wind and SST anomalies extend toward the tropics, we selected 15 points at equal latitude–longitude intervals along the propagation path (red points in Fig. 1b). The temporal evolution of these variables is analyzed by computing regional averages around each selected point to investigate differences in air–sea coupling processes between the two types of events. At each point, a 3-month running mean is applied, and regional means are calculated using a 9-grid-point area centered on each point. In NPO-ENSO events, notable southwesterly wind anomalies are already present in the regions from point 2 to point 6 in December (Fig. 3a). These anomalies induce positive SST anomaly during winter by reducing upward turbulent heat fluxes

(mainly LHF). Due to oceanic thermal inertia, the SST anomaly responds with a lag of approximately one month to changes in wind and LHF anomalies, reaching its peak in early spring, consistent with findings of Vimont (2003a). Subsequently, the SST anomaly induces an atmospheric response, causing southwesterly wind anomalies on its southwest side that further drive warm SST anomaly. Thus, via this WES feedback, the SST anomaly and associated southwesterly wind and LHF anomalies propagate toward the tropical western Pacific. These anomalies are observed to reach the tropics (points 12-15) around April to May of the following year, subsequently triggering a westerly wind bursts. This wind burst further interacts with the tropical ocean, leading to the generation of eastward-propagating Kelvin waves (Weisberg and Wang 1997), which ultimately induce warm SST anomalies in the equatorial central and eastern Pacific.

In contrast, during NPO–NONE events, although a strong NPO anomaly may also develop in early winter, the negative SLP anomaly begins to contract in January. By February, it becomes markedly elongated in the zonal direction and splits into two centers, with the main core retreating westward toward the date line and shifting northward (Fig. 2b, d, f). This structural change causes the southwesterly wind anomalies over the subtropical Pacific to shrink rapidly in spatial extent. Under these circumstances, even though notable SST anomalies persist over the subtropical ocean in February, the rapid decay of the negative SLP anomaly prevents the effective activation of the WES feedback, thereby inhibiting the equatorward propagation of the associated wind and SST anomalies toward the western equatorial Pacific. Consequently, the NPO-related wind and SST anomalies dissipate quickly during the following spring, failing to affect the oceanic thermal conditions in the tropics (Fig. 2h, j, l). It can also be seen that although horizontal wind anomalies initially develop near the southern flank of the low-pressure anomalies in NPO–NONE, the positive feedback process between these winds and SST anomalies is confined to the subtropics during winter and rapidly dissipates in spring as the cyclone anomaly weakens (Fig. 3b). Thus, no well-defined NPMM forms to effectively trigger subsequent ENSO events. These differences between the two types of events highlight that the intensity and spatial characteristics of the initial negative SLP anomaly significantly influence the efficiency of the WES feedback. Specifically, the anomalous southwesterly winds associated with the southern branch of the NPO play an important role in determining the strength and

extent of this feedback process.

3.2 Role of the NPO related upper-level circulation anomalies

Besides differences in SST and low-level atmospheric circulation, NPO-ENSO and NPO-NONE events also exhibit significant discrepancies in the evolution of upper-level circulation anomalies. In NPO-ENSO events, the geopotential low anomaly gradually acquires a northeast-southwest tilt during winter, and an anticyclonic anomaly emerges near 160°W in the tropics, accompanied by remarkable easterly anomalies on its southern flank in February (Fig. 4a, c, e). Notably, the anticyclonic anomaly persists and intensifies into spring, with associated easterly wind anomalies extending progressively equatorward and reaching the equatorial Pacific by May (Fig. 4g, i, k). By contrast, the NPO anomalies in NPO-NONE events gradually shift northward throughout winter, most clearly observed in the poleward migration of the northern anomalous high lobe. The low anomaly during winter is characterized by a pronounced zonal elongation, similar to features observed at lower levels (Fig. 4b, d, f). Moreover, the anticyclonic anomaly in the subtropics is weak and not significant during winter, failing to produce easterly anomalies and dissipating rapidly in spring (Fig. 4h, j, l). The following analysis further investigates how the structural differences in upper-level NPO anomalies influence the initiation of ENSO events.

The NPO exhibits a Western Pacific (WP) teleconnection pattern in the upper troposphere (Nakamura et al. 1987; Tanaka et al. 2016), which can influence ENSO activity through teleconnection processes. To further elucidate this upper-level pathway, we analyze the WAF (Eq .1) associated with two types of events. Composite results indicate that the northern anticyclonic anomalies of the NPO in both NPO-ENSO and NPO-NONE events stimulate downstream wave energy propagation (Fig. 5a, b), consistent with its known influence on North American weather and climate (Baxter and Nigam 2015; Sung et al. 2019). However, in the NPO-ENSO events, the southern lobe of the NPO exhibits prominent equatorward energy propagation during winter, and this energy propagation gradually intensifies throughout the season. This southward energy propagation sustains the tropical anticyclonic circulation anomaly throughout spring and helps expand the extent of easterly anomalies on the southern side of the anticyclone. According to Zhao et al. (2023a, b), the southward energy propagation is closely related to the structural characteristics of the southern lobe cyclone. Specifically,

a northeast-southwest tilted structure favors effective southward propagation of WAF, as clearly observed in NPO-ENSO events (Fig. 5e).

Further analysis reveals that the upper-level easterly anomalies associated with the subtropical anticyclone help establish a vertical easterly shear over the equator (Fig. 6 left panels). According to Wang and Xie (1996a, b), vertical easterly shear confines atmospheric disturbances to the lower troposphere, thereby intensifying Ekman pumping and boundary-layer frictional moisture convergence. This moisture-induced condensational heating ultimately facilitates the development of westerly wind anomalies over the equatorial central Pacific. In NPO-ENSO events, vertical easterly shear anomalies initially emerge near the equator and tropical regions during late winter (indicated by dashed lines, Fig. 6a). With the onset of spring, the extent of easterly shear between the upper and lower troposphere gradually expands (Fig. 6c, e, g), further intensifying the strength and range of low-level moisture transport. Generally, the enhancement of moisture convergence is accompanied by increased precipitation, and in NPO-ENSO events, the precipitation coverage in the tropical and equatorial Pacific expands during spring (Supplementary Fig. S1c, e, g). The increase in precipitation is accompanied by diabatic heating processes, which facilitate the formation of a Gill-type circulation response (Gill 1980) and subsequently promote the development of equatorial westerlies, thereby facilitating the onset of El Niño.

In contrast, NPO-NONE events exhibit primarily zonal elongation of the geopotential low, lacking the meridional tilt necessary for effective equatorward energy propagation. As a result, although slight equatorward propagation of WAF occurs briefly in December during NPO-NONE events, the main wave energy propagation remains zonally oriented (Fig. 5b, d, f). Due to this limited southward energy input, the associated tropical anticyclonic anomaly rapidly weakens and dissipates during spring, failing to sustain significant easterly anomalies. Thus, the vertical easterly shear is weak and gradually decays from winter to spring, inhibiting the moisture transport to the tropics (Fig. 6b, d, f, h). Correspondingly, precipitation differences in tropical regions between the two events are relatively minor in February (Supplementary Fig. S1a, b). However, as spring progresses and moisture convergence intensifies, NPO-ENSO events exhibit significantly greater precipitation coverage over the tropical and equatorial Pacific compared to NPO-NONE events (Fig. S1c-f). Notably, positive

precipitation anomalies are observed in the tropical region between 150°E and 170°E in NPO-ENSO events, whereas NPO-NONE events exhibit drought conditions in the same region. This distinct precipitation pattern explains the early emergence of westerly anomalies in the equatorial western Pacific during NPO-ENSO events (Fig. 2e).

The analyses presented in this section highlight that the upper-level atmospheric circulation associated with the NPO also significantly contributes to ENSO triggering through dynamic and thermodynamic processes related to southward wave-energy propagation. Collectively, these findings indicate that the role of the NPO in triggering ENSO is primarily achieved through the combined effects of lower-tropospheric air–sea coupling and upper-tropospheric wave-energy propagation. The structural differences in the southern branch of the NPO play an important role in modulating these two processes and their associated impacts.

To further quantitatively evaluate the key factors governing the ability of the NPO to trigger ENSO, we conduct a five-fold cross-validated logistic regression analysis using five standardized NPO characteristic indices, including the latitude of the southern low-pressure center, the central sea level pressure, the axial tilt of the southern low-pressure center, the intensity of the southward-propagating meridional WAF, and the zonal-to-meridional aspect ratio (Fig. 7). The regression results further support the above findings. Specifically, a southward-extended and stronger southern low-pressure center of NPO, combined with enhanced southward propagation of upper-tropospheric Rossby waves, greatly favors ENSO triggering. By contrast, a more zonally elongated and tilted southern low-pressure center suppresses ENSO onset. It is worth noting that, although the intensity of the southern low-pressure center contributes positively to ENSO triggering, its effect is relatively weak. That means the spatial structure of the NPO is more important than its intensity in determining its ability to trigger ENSO. Nevertheless, the processes governing the structural evolution of the NPO itself remain unclear. The following section explores this issue from the perspective of oceanic feedback.

4 Influence of oceanic feedback on the structural evolution of NPO

As a midlatitude atmospheric circulation pattern, the formation and evolution of the NPO are primarily driven by internal atmospheric processes (Lau 1988; Li and

Wettstein 2012; Tanaka et al. 2016; Sung et al. 2022). However, the oceanic feedback also exerts a substantial modulating effect on the spatial structure of the NPO during its evolution.

To further quantify the underlying dynamical processes, the time-mean quasi-geostrophic potential vorticity (QGPV) equation following Fang and Yang (2016) is employed to diagnose the contributions from different forcing terms,

$$\left(\frac{\partial}{\partial t} + \overline{\vec{V}_h} \cdot \nabla\right) \left[\frac{1}{f} \nabla^2 \bar{\phi} + f + \frac{\partial}{\partial p} \left(\frac{f}{\sigma_1} \frac{\partial \bar{\phi}}{\partial p} \right) \right] = \underbrace{-f \frac{\partial}{\partial p} \left(\frac{\alpha}{\sigma_1} \overline{Q_d} \right)}_{F_1} - \underbrace{f \frac{\partial}{\partial p} \left(\frac{\alpha}{\sigma_1} \overline{Q_{eddy}} \right)}_{F_2} + \underbrace{\overline{F_{eddy}}}_{F_3} \quad (2)$$

$$\overline{Q_{eddy}} = -\nabla \cdot \overline{\vec{V}'T'} - \frac{\partial \overline{\omega'T'}}{\partial p} + \frac{R}{c_p p} \overline{\omega'T'} \quad (3)$$

$$\overline{F_{eddy}} = -\nabla \cdot \overline{\vec{V}'\zeta'} \quad (4)$$

where the overbar denotes the monthly mean, ϕ is the geopotential, α is the reciprocal of atmospheric density, $\overline{\sigma_1} = \frac{\overline{\sigma_0 R}}{p}$ is the atmospheric static stability parameter, and $\overline{Q_d}$ is the monthly-mean diabatic heating that can be diagnosed as the residual of the thermodynamic equation (Yanai and Tomita 1998). $\overline{Q_{eddy}}$ is the monthly-mean transient eddy heating that is determined by the 3-dimensional convergence of heat flux transport by transient eddies (Eq. 3), and $\overline{F_{eddy}}$ is the transient eddy vorticity forcing induced by vorticity flux transport convergence by transient eddies (Eq. 4). The prime in Eq. 3 and Eq. 4 denotes the synoptic-scale perturbations. Eq. 2 suggests that the monthly-mean midlatitude atmospheric circulation is driven by three terms: diabatic heat forcing (F_1), transient eddy thermal forcing (F_2), and transient eddy vorticity forcing (F_3). Therefore, the midlatitude oceanic surface thermal condition can affect the atmosphere by changing diabatic heating and transient eddy activities. That is to say, the NPO-induced SST anomalies could exert feedback to the atmosphere through direct and indirect processes. Eq. 2 can be reformulated as the geopotential tendency equation, with the same three forcing terms in the right side. Then, we calculate the NPO-related anomalies of $\overline{Q_d}$, $\overline{Q_{eddy}}$ and $\overline{F_{eddy}}$, as well as the corresponding F_1 , F_2 , and F_3 anomalies for both types of events. Then, the successive over-relaxation (SOR) method is applied to obtain the initial geopotential tendency ($\frac{\partial \bar{\phi}}{\partial t}$) responses to each anomalous

forcing term.

In the NPO–ENSO events, during the early and middle winter, anomalous westerlies around the low-pressure center strengthen the background trade winds, enhancing the release of THF from the ocean to the atmosphere and inducing cold SST anomalies over the midlatitude North Pacific (20°–40°N), accompanied by positive $\overline{Q_d}$ anomalies in the lower troposphere (Fig. 2, Fig. 8a, b and Fig. 9a, b). Meanwhile, easterly anomalies south of the high-pressure center and southwesterly anomalies southeast of the low-pressure center weaken the background trade winds, reducing ocean-to-atmosphere heat release and thereby generating local warm SST anomalies, together with negative $\overline{Q_d}$ anomalies in the lower troposphere. The negative correlation between SST and THF anomalies (e.g. negative SST anomaly corresponds to upward THF anomaly) suggests that during December and January, these events are primarily characterized by atmospheric forcing of the ocean through near-surface dynamical processes. By February, however, the THF anomalies over the warm SST anomalies in the 40°–50°N reverse its sign (Fig. 8c). Through the decomposition of SHF and LHF, we demonstrate that this reversal of THF anomalies is mainly caused by the enhanced air–sea temperature contrast induced by warm SST anomalies (Supplementary Fig. S2). The enhanced upward heat release subsequently leads to significant positive $\overline{Q_d}$ anomalies in the lower troposphere over this region (Fig. 8c and 9c), suggesting that the SST anomalies can exert thermal feedback on the atmospheric circulation via modifying the surface heat fluxes in late winter. The diabatic heating anomalies also undergo a transition from atmosphere-driven in early winter to ocean-driven in late winter.

On the other hand, in February, the large-scale meridional SST gradient anomalies not only modify the intensity and position of the oceanic frontal zones, but also influence the distribution of low-level atmospheric baroclinicity through thermal effects, resulting in the co-location of the large atmospheric baroclinicity and the oceanic frontal zones. According to baroclinic instability theory, atmospheric transient eddies are generated in baroclinic regions (Charney 1947; Eady 1949; Simmons and Hoskins 1978; Hoskins and James 2014). Therefore, regions with strong baroclinicity are associated with vigorous transient eddy activities, which significantly influence time-mean atmospheric circulation by transporting moisture, heat, and vorticity (Kug et al. 2010;

Tao et al. 2020; Fang et al. 2022). In the NPO–ENSO events, cold SST anomalies in February are observed over the 20°–35°N band, while warm SST anomalies dominate north of 35°N, with the maximum warming located near 42°N (Fig. 10a). This SST anomaly pattern weakens the meridional SST gradient between 35°N and 42°N, which is accompanied by a reduction in lower-tropospheric baroclinicity. In contrast, north of 42°N, the meridional SST gradient is enhanced, inducing increased atmospheric baroclinicity. According to Eq. 3 and 4, $\overline{Q_{eddy}}$ and $\overline{F_{eddy}}$ represent the divergence of eddy-induced heat or vorticity fluxes, and thus their patterns are nearly perpendicular to the axis of strongest baroclinicity in the meridional direction. Notably, atmospheric baroclinicity near 40°N is substantially weakened in February (Fig. 10a), while pronounced negative anomalies of transient eddy forcing emerge to its north (Fig. 9f, i). These results suggest that the baroclinicity changes associated with oceanic feedback may further influence the atmospheric circulation by modulating transient eddy thermal and dynamical forcing.

For the atmospheric responses induced by the different forcing terms, the geopotential tendency response to F_1 exhibits a pronounced baroclinic structure in the vertical direction, with negative (positive) geopotential tendency anomalies collocated with the low-pressure (high-pressure) centers below 700 hPa (first column of Fig. 11). As a result, F_1 primarily acts to maintain NPO structure in the lower troposphere, while tending to weaken the NPO anomalies in the upper troposphere. The geopotential tendency response induced by F_2 is generally similar to that of F_1 , but with a weaker amplitude and a slightly northward-shift response center (second column of Fig. 11). In contrast, only the geopotential tendency induced by F_3 exhibits a clear equivalent-barotropic structure throughout the troposphere (third column of Fig. 11). Its spatial pattern closely resembles the observed NPO-related geopotential height anomalies (fourth column of Fig. 11), suggesting that F_3 is more effective in maintaining the NPO structure throughout the troposphere. This again confirms that the transient eddy vorticity forcing is the primary factor for the formation and maintenance of the equivalent barotropic structure of the midlatitude atmospheric circulation (Fang and Yang 2016; Sun et al. 2018; Tao et al. 2019, 2020; Tao et al. 2022).

In NPO-ENSO events, although all three forcing terms contribute positively to the development of the southern negative SLP anomaly of the NPO, the F_1 induced

geopotential tendency is strongest at lower atmospheric layers. Moreover, its horizontal distribution closely matches the NPO pattern (Fig. 12a-c), indicating that diabatic heating plays a dominant role in shaping the low-level NPO structure. By contrast, in the upper troposphere, the NPO structure aligns more closely with the geopotential tendency induced by F_3 (Fig. 12g-i), suggesting that F_3 plays a more prominent role aloft.

In contrast, oceanic feedback is substantially weaker in the NPO–NONE events. On the one hand, the relatively weak THF anomalies limit the ability of the ocean to exert a direct thermal influence on the midlatitude atmosphere through $\overline{Q_d}$ (Supplementary Fig. S3). In these events, the initial atmospheric forcing produces positive $\overline{Q_d}$ anomalies south of 40°N and negative anomalies to its north (Fig. 8d and Fig. 13a). However, the positive anomalies gradually weaken and contract after January. By February, only weak $\overline{Q_d}$ signals remain in the lower and middle troposphere south of 40°N, while no significant positive $\overline{Q_d}$ anomalies are found over the midlatitudes (Fig. 8f and 13c). These results suggest that the diabatic heating exerts comparatively limited contribution to the structural evolution of the NPO during NPO–NONE events, and such diabatic heating is barely modulated by the ocean. On the other hand, although a local minimum in the meridional SST gradient also develops over the midlatitudes during late winter, the anomalies of meridional SST gradients do not spatially coincide with the EGR anomalies in this region (Fig. 10b). Such spatial decoupling implies that the influence of oceanic anomalies on atmospheric baroclinicity and transient eddy activity is relatively weak. Consequently, the structural evolution of the NPO is predominantly controlled by internal atmospheric dynamics. From the vertical distribution of the transient eddy-related forcing terms, $\overline{Q_{eddy}}$ is characterized by negative anomalies at high latitudes, whereas positive anomalies are mainly confined to the region south of 40°N (Fig. 13d). Meanwhile, $\overline{F_{eddy}}$ exhibits positive anomalies primarily between 35°N and 45°N (Fig. 13g). Under the influence of internal atmospheric dynamics, the positive anomaly centers of both $\overline{Q_{eddy}}$ and $\overline{F_{eddy}}$ expand poleward (Fig. 13f, i), consistent with the northward displacement of the NPO-related circulation anomalies.

Correspondingly, because the $\overline{Q_d}$ anomalies in the NPO–NONE events are largely confined to the lower troposphere, the associated geopotential response is relatively

weak and displaced poleward (Fig. 14a, e, i). The geopotential tendency induced by F_2 also exhibits a baroclinic vertical structure. By February, the geopotential tendency responses associated with both F_2 and F_3 become substantially stronger. However, in the absence of effective local oceanic feedback, these responses are not anchored near the southern lobe of the NPO and instead tend to shift toward the mid–high latitudes (Fig. 14j, k). And the response induced by F_3 is primarily concentrated in the middle and upper troposphere, while its largest low-level anomalies are also located farther poleward (third column of Fig. 14), contributing to a northward displacement of the NPO circulation. At the same time, the southwesterly anomalies south of the NPO low-pressure center gradually become detached from the warm SST anomalies, making it difficult to establish effective WES feedback. In addition, under the influence of internal atmospheric processes, the geopotential tendency responses induced by F_1 and F_3 exhibit a double-center structure near the southern lobe of the NPO, causing the low-pressure anomalies to become zonally elongated and eventually split (Fig. 12f, l). Such a zonally elongated structure is unfavorable for the equatorward propagation of upper-level WAF, thereby weakening the influence of upper-level teleconnections on the triggering of ENSO.

The analyses presented above reveal oceanic feedback as a key modulator of NPO structure. By reshaping the diabatic heating and the atmospheric baroclinicity, the oceanic feedback further sustains NPO-related circulation anomalies, ultimately modulating its potential to trigger ENSO events in the subsequent year.

5 Summary and discussion

This study focuses on the structural characteristics of the winter NPO and their influence on the triggering of ENSO in the following year. By comparing NPO events followed by ENSO (NPO-ENSO) with those not followed by ENSO (NPO-NONE), it reveals how differences in the structure and intensity of the NPO affect air–sea coupling and upper-level atmospheric processes, thereby impacting the onset of ENSO. The study also identifies the role of oceanic feedback in the evolution of the NPO in the two types of events to explain the causes of the structural diversity of the NPO.

In the lower troposphere, anomalous southwesterly winds associated with the southern branch of the NPO weaken the background trade winds, inducing localized

SST warming. This SST anomaly is amplified and maintained through the WES feedback into spring, leading to the formation of the NPM. Subsequently, both the SST anomaly and associated southwesterly wind anomalies propagate toward the tropics through the SFM, triggering equatorial westerly wind bursts that promote ENSO development. This process is favored in NPO-ENSO events, of which the negative SLP anomaly strengthens during winter and shifts toward the equator, thereby enhancing SST anomalies and facilitating their tropical propagation. In contrast, during NPO-NONE events, the negative SLP anomaly weakens and becomes zonally elongated while shifting eastward in winter. These changes suppress the positive feedback between wind and SST anomalies, confining the SST anomalies to the subtropics where they gradually dissipate.

At the upper levels, the northeast–southwest tilted geopotential low anomaly during NPO-ENSO events facilitates the southward propagation of wave energy, maintaining the tropical anticyclone and its associated easterly anomalies to the south. This enhances easterly wind shear over the tropical regions, confines disturbances to the lower troposphere, and promotes Ekman pumping and moisture convergence. These processes favor convective precipitation and a Gill-type circulation response, which strengthens the equatorial westerlies and thus facilitates ENSO onset. In contrast, the anomalous geopotential low in NPO-NONE events displays a pronounced zonally extended structure, resulting in zonally wave energy propagation rather than southward propagation, which cannot affect tropical regions and further trigger ENSO.

The difference in ENSO triggering effect between the two types of events is closely related to differences in the NPO structure, which can be modulated by SST anomalies during the late winter. In the NPO–ENSO events, the oceanic contribution to THF rises continuously during winter. By February, it exceeds the contribution of the atmosphere, indicating that the ocean can exert robust feedback on the atmosphere through heat flux processes. The resulting meridional SST gradient anomalies weaken the Subarctic Frontal Zone (SAFZ), reduce the midlatitude atmospheric baroclinicity, and further modify the associated transient eddy activities. Under the combined effects of diabatic heating, transient eddy heat forcing, and transient eddy vorticity forcing, the southern low-pressure anomaly of the NPO persists into early spring, favoring the formation and continued development of the NPM and facilitating the tropical

propagation of upper-level teleconnection signals. Although all three forcing terms contribute positively to maintaining the NPO structure to some extent, the diabatic heating and the transient eddy thermal forcing mainly contribute in the lower troposphere, with diabatic heating playing the dominant role. Meanwhile, the transient eddy vorticity forcing is the primary driver in the upper troposphere and helps sustain the equivalent barotropic structure of the NPO. In contrast, in the NPO–NONE events, the SST anomalies are relatively weak and exert only limited feedback on the atmosphere by late winter. Under this condition, the internal atmospheric processes control the evolution of the midlatitude baroclinic zone and transient eddy activities, leading to a northward displacement of the NPO-related circulation anomalies. Meanwhile, as the diabatic heating weakens and the spatial distribution of transient eddy forcing changes, the cyclonic circulation anomaly gradually shifts westward in February and develops a pronounced zonally elongated structure, weakening the NPO pattern. Such evolution not only inhibits the propagation of anomalous southwesterlies toward the tropical western Pacific through the SFM, but is also unfavorable for the equatorward propagation of upper-level wave energy.

This study further analyzes the impact mechanisms of the NPO on ENSO triggering via upper-level and lower-level processes. Notably, Zhao et al. (2023b) have demonstrated that the southward propagation of upper-level WAF and the lower-level SFM are not independent. Regression analysis between the standardized WAF southward propagation index over the Hawaiian region (10°N–30°N, 170°W–130°W) and precipitation reveals that the southward wave energy propagation significantly affects precipitation in the subtropical Pacific (Supplementary Fig. S4). It is therefore inferred that precipitation changes induced by upper-level wave energy propagation may modulate the position of the Intertropical Convergence Zone (ITCZ), thereby enhancing the SST and wind anomalies triggered by the NPO (Amaya et al. 2019; Park et al. 2021). Although qualitative analyses have indicated that both upper- and lower-level processes contribute to ENSO triggering, their interaction and relative importance in the actual triggering process require further quantification. Additionally, the NPO-related anomalous southwesterly winds at the sea surface may influence ENSO activity through other pathways (Solomon et al. 2008; Alexander et al. 2010; Anderson et al. 2013b), and the specific impacts of the initial NPO structure and SST anomaly location on these pathways warrant further investigation.

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Statements and Declarations

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Competing Interests

The authors have no relevant financial or non-financial interests to disclose.

Author Contributions

All authors contributed to the study conception and design. The main idea of the study was put forward by Jiabei Fang. Material preparation, data collection and analysis were performed by Yuhui Zhao and Jiabei Fang. The manuscript was organized and written by Yuhui Zhao and Jiabei Fang. All authors read and approved the final manuscript.

Data Availability

The ECMWF ERA5 reanalysis datasets analyzed during the current study are available in the Climate Data Store website at <https://cds.climate.copernicus.eu/>. The calculation of the Niño index in this study follows the method provided by NCAR, available at <https://climatedataguide.ucar.edu/climate-data/nino-sst-indices-nino-12-3-34-4-oni-and-tni>.

803 **Tabels**

804 **Table 1.** NPO and ENSO Event Classification

Event classification	ENSO phase	Year of wintertime NPO
NPO-ENSO	+NPO-El Niño	1956/1957, 1967/1968, 1981/1982, 1990/1991, 2008/2009, 2013/2014, 2014/2015
	-NPO-La Niña	1954/1955, 1974/1975, 1987/1988, 1998/1999, 2006/2007, 2007/2008, 2019/2020, 2020/2021
NPO-NONE	+NPO-none	1968/1969, 1977/1978, 1980/1981
	-NPO-none	1959/1960, 1975/1976, 1989/1990, 2000/2001, 2011/2012, 2012/2013, 2015/2016

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Figure caption

Fig. 1 (a) Lead-lag correlations between the NPO index and different Niño indices. The x-axis represents the lead-lag in months, with positive (negative) values indicating that ENSO lags (leads) the NPO index. The solid line indicates that the correlation is significant at the 95% confidence level. (b) Regression of the winter (DJF) NPO index on concurrent SST (shaded, unit: K) and SLP (contours, interval: 0.4 hPa) anomalies. The red line represents the Hovmöller transect path shown in Fig. 3, which spans 15 evenly spaced points along the longitude-latitude axis, with the first point at 118°W, 22°N and the 15th point at 180°E, 5°N. (c) As (b) but for DJF SST of the subsequent year (shaded, unit: K). Stippling in (b) and (c) indicates statistical significance at the 95% confidence level

Fig. 2 Composites of SST (shaded, unit: K), surface wind (vectors, unit: $\text{m}\cdot\text{s}^{-1}$) and SLP (contours, interval: 0.8 hPa) anomalies for NPO-ENSO events (left panel) and NPO-NONE events (right panel), respectively. Stippling in shaded and purple color in line indicate statistical significance at the 95% confidence level, and horizontal wind anomalies with only confidence levels above 95% are shown

Fig. 3 Temporal evolution of composite SST (shaded, unit: K), latent heat flux (LHF, contours, interval: $2 \text{ W}\cdot\text{m}^{-2}$), and horizontal wind (vectors, unit: $\text{m}\cdot\text{s}^{-1}$) anomalies along the propagation path illustrated in Fig. 1b (from Nov (0) to Jun) for NPO-ENSO events (left panel) and NPO-NONE events (right panel), respectively. Values at each point are smoothed with a 3-month running average, and the regional average of 9 grid points centered on each point is applied. For LHF and horizontal wind anomalies, only values with confidence levels above 95% are shown, crosses indicate regions where SST anomalies are statistically significant at the 95% confidence level

Fig. 4 Composites of Z (shaded, unit: gpm) and horizontal wind (vectors, unit: $\text{m}\cdot\text{s}^{-1}$) anomalies at 300 hPa for NPO-ENSO events (left panel) and NPO-NONE events (right panel), respectively. Stippling indicates statistical significance at the 95% confidence level, and horizontal wind anomalies are shown with only confidence levels above 95%

Fig. 5 Composites of stream function (shaded, unit: $10^7 \text{ m}^2 \cdot \text{s}^{-1}$) and WAF (vectors, unit: $\text{m}^2 \cdot \text{s}^{-2}$) anomalies at 300 hPa for NPO-ENSO events (left panel) and NPO-NONE events (right panel), respectively. Stippling indicates statistical significance at the 95% confidence level and vectors with magnitude below $1 \text{ m}^2 \cdot \text{s}^{-2}$ are masked out

Fig. 6 Composites of vertical easterly shear [$U_{300\text{hPa}-850\text{hPa}} - U_{850\text{hPa}-1000\text{hPa}}$] (contours, interval: $2 \text{ m} \cdot \text{s}^{-1}$), with red contours indicating statistical significance at the 95% confidence level, moisture flux anomaly averaged from 1000-850 hPa (vector, unit: $\text{g} \cdot \text{s}^{-1} \cdot \text{cm}^{-1}$), only confidence levels above 95% are shown, and moisture flux divergence anomaly averaged from 1000-850 hPa (shaded, unit: $10^{-4} \text{ g} \cdot \text{s}^{-1} \cdot \text{cm}^{-2}$), with stippling indicating statistical significance at the 95% confidence level. The left panel represents NPO-ENSO events, and the right panel represents NPO-NONE events

Fig. 7 Regression coefficients and their confidence intervals from the logistic regression model. The dependent variable is ENSO occurrence (0/1). From top to bottom, the predictors on the y-axis are meridional WAF intensity, the latitude of the southern low-pressure center (defined as positive when displaced southward relative to the standardized NPO pattern), central sea level pressure, axis tilt, and the zonal-to-meridional aspect ratio. All predictors are standardized using the training dataset (standard deviation $\sigma = 1$). The x-axis represents the change in the log-odds of ENSO occurrence associated with a one-standard-deviation increase in each predictor. Dots denote the maximum likelihood estimates (MLEs). Thin gray error bars indicate ± 1.96 standard errors (approximately corresponding to the 95% two-sided confidence intervals), while thick black error bars indicate ± 1.645 standard errors (approximately corresponding to the 90% two-sided confidence intervals). The vertical dashed line denotes zero effect; positive (negative) coefficients indicate an increase (decrease) in the probability of ENSO = 1

Fig. 8 Composites of $\overline{Q_d}$ anomalies at 1000 hPa (contour, interval: $2 \times 10^{-6} \text{ K} \cdot \text{s}^{-1}$) and THF anomalies (shaded, unit: $\text{W} \cdot \text{m}^{-2}$) for NPO-ENSO events (left panel) and NPO-NONE events (right panel). Stippling marks regions significant at the 95% confidence level

Fig. 9 Latitude-altitude sections of the composite $\overline{Q_d}$ (left panel, unit: $10^{-6} \text{ K} \cdot \text{s}^{-1}$), $\overline{Q_{eddy}}$ (middle panel, unit: $10^{-6} \text{ K} \cdot \text{s}^{-1}$), and $\overline{F_{eddy}}$ (right panel, unit: 10^{-11} s^{-1}) anomalies for NPO-ENSO events from Dec (0) to Feb, averaged between 160°E - 240°E . Stippling indicates statistical significance at the 95% confidence level

Fig. 10 Composite meridional distributions of anomalous SST (red solid line, unit: $10^{-6} \cdot \text{s}^{-1}$), $-\partial\text{SST}/\partial y$ (purple dashed line, units: $10^{-6} \text{ K} \cdot \text{m}^{-1}$), 1000 hPa $\overline{Q_d}$ (green solid line, units: $10^{-6} \text{ K} \cdot \text{s}^{-1}$) and 850 hPa EGR (orange solid line, units: $10^{-6} \cdot \text{s}^{-1}$) in February, averaged over the 160°E - 140°W sector for the NPO-ENSO events. The two shaded orange bands indicate the climatological Subtropical Frontal Zone (STFZ) and Subarctic Frontal Zone (SAFZ), respectively. (b) Same as in (a), but for the NPO-NONE events

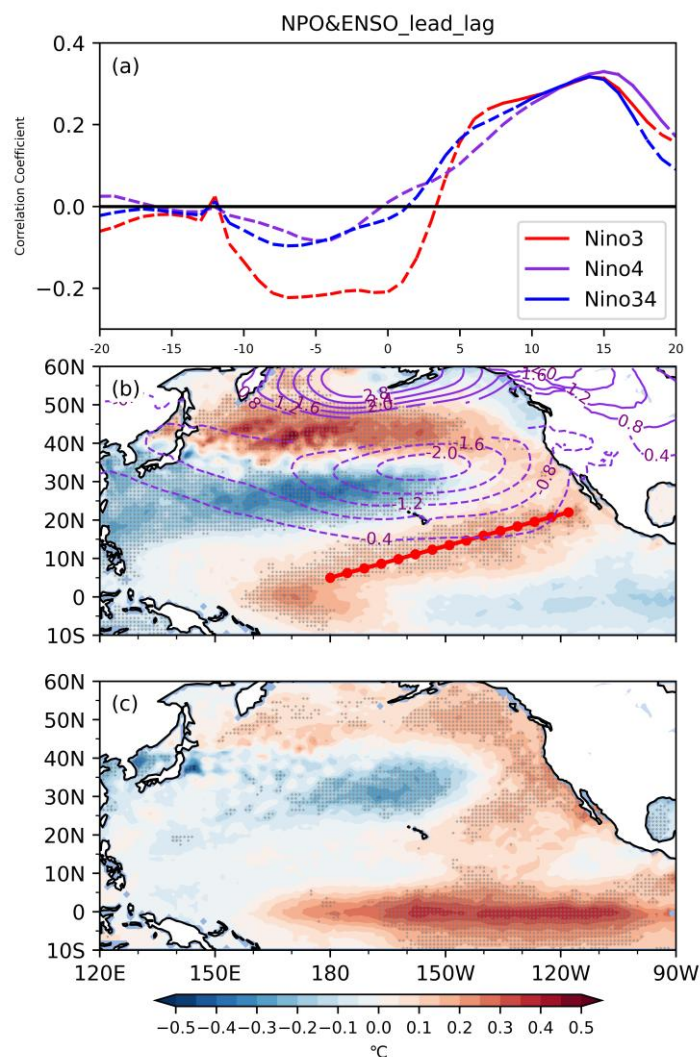
Fig. 11 Latitude-altitude sections of the composite geopotential tendencies (unit: $10^{-4} \text{ m}^2 \cdot \text{s}^{-3}$) induced by F_1 (first panel), F_2 (second panel) and F_3 (third panel), as well as Z anomaly (fourth panel, unit: gpm) for NPO-ENSO events from Dec (0) to Feb. All the variables are averaged between 160°E - 240°E and stippling indicates statistical significance at the 95% confidence level

Fig. 12 Composites of geopotential tendencies (shaded, unit: $10^{-3} \text{ m}^2 \cdot \text{s}^{-3}$) induced by F_1 at 1000 hPa (left panel) and F_3 at 300 hPa (right panel), with stippling indicating statistical significance at the 95% confidence level, and Z anomaly at 1000 hPa and 300 hPa (contours, interval: 8 gpm), with purple contours indicating statistical significance at the 95% confidence level. The first three rows correspond to NPO-ENSO events, while the last three rows correspond to NPO-NONE events

Fig. 13 Same as Fig. 9, but for NPO-NONE events

Fig. 14 Same as Fig. 11, but for NPO-NONE events

902 **Figures**



903
 904 **Fig. 1** (a) Lead-lag correlations between the NPO index and different Niño indices. The
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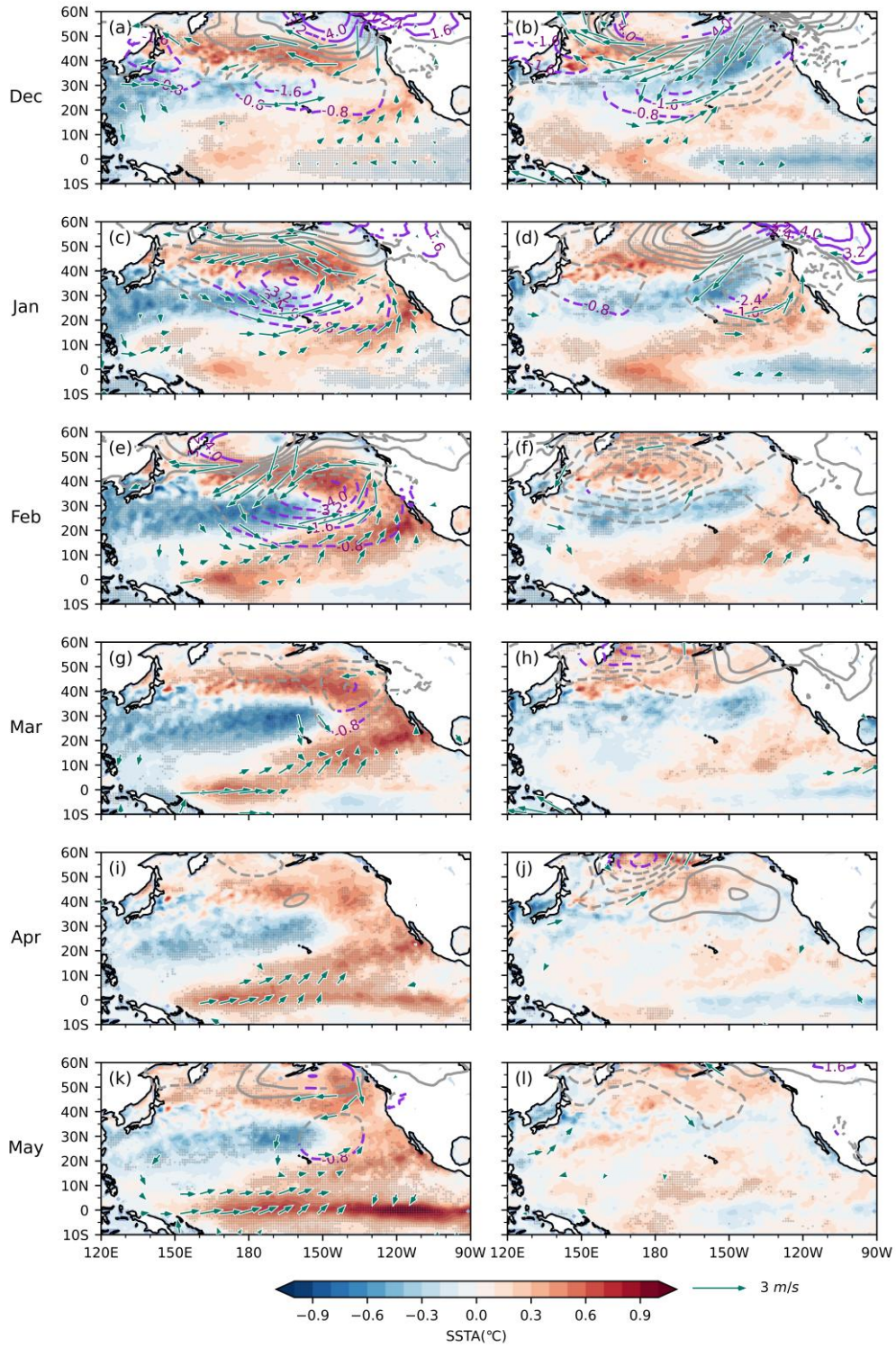


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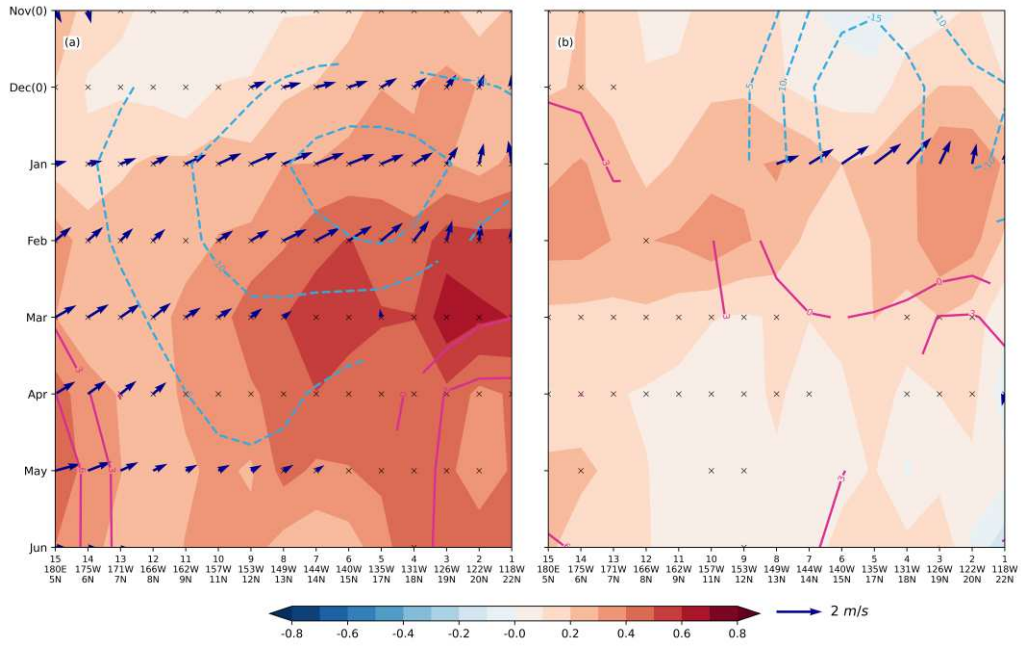


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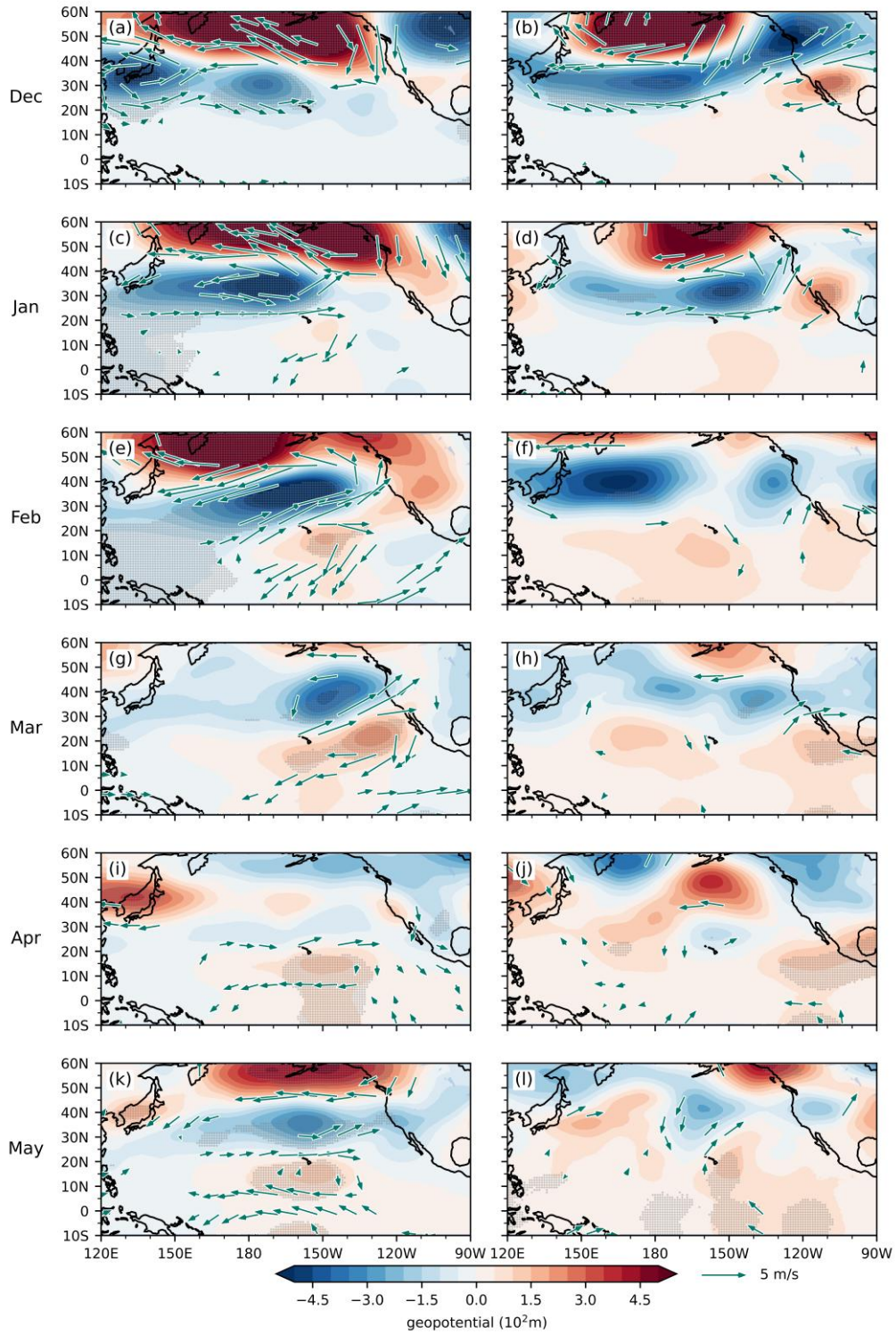


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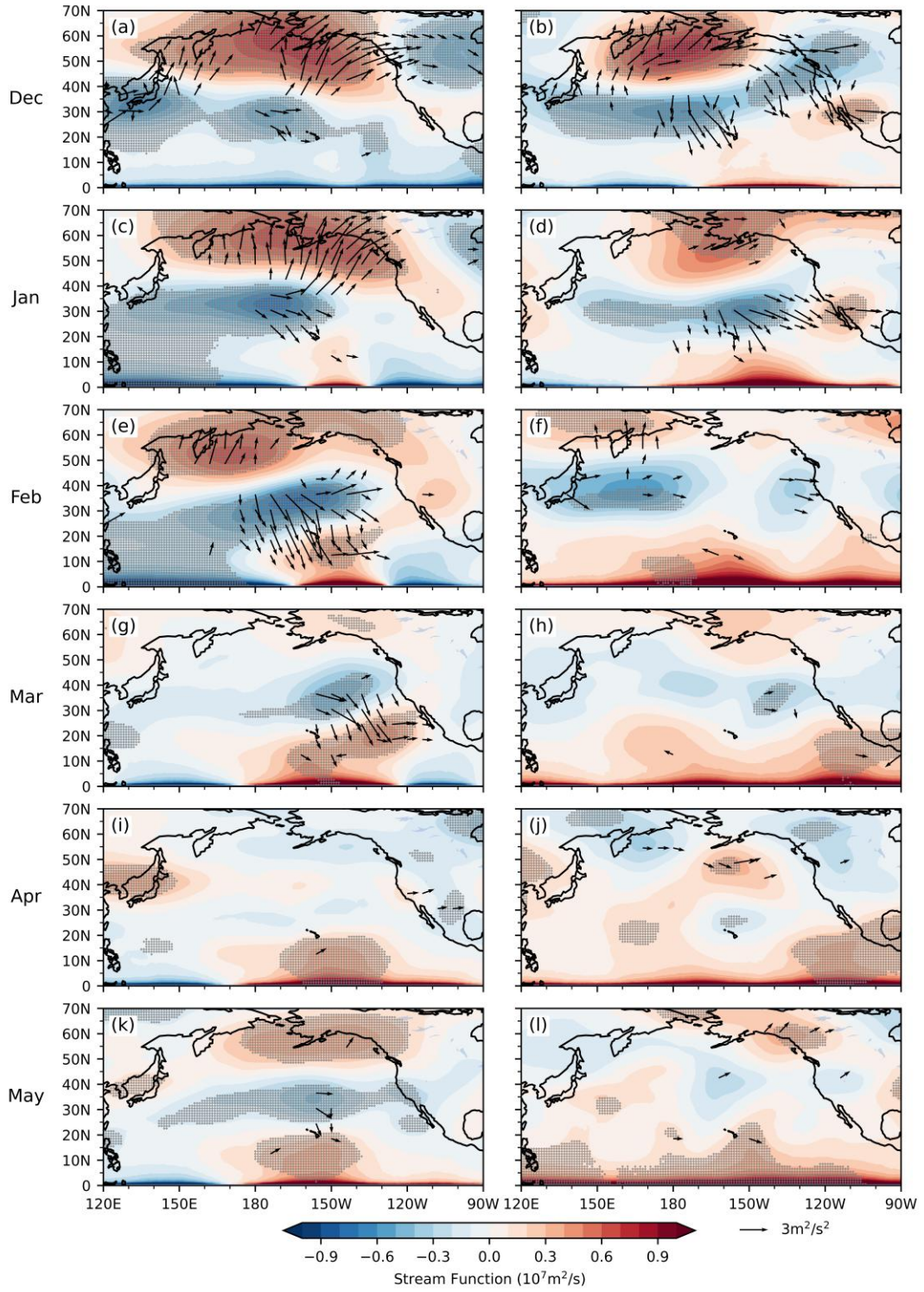


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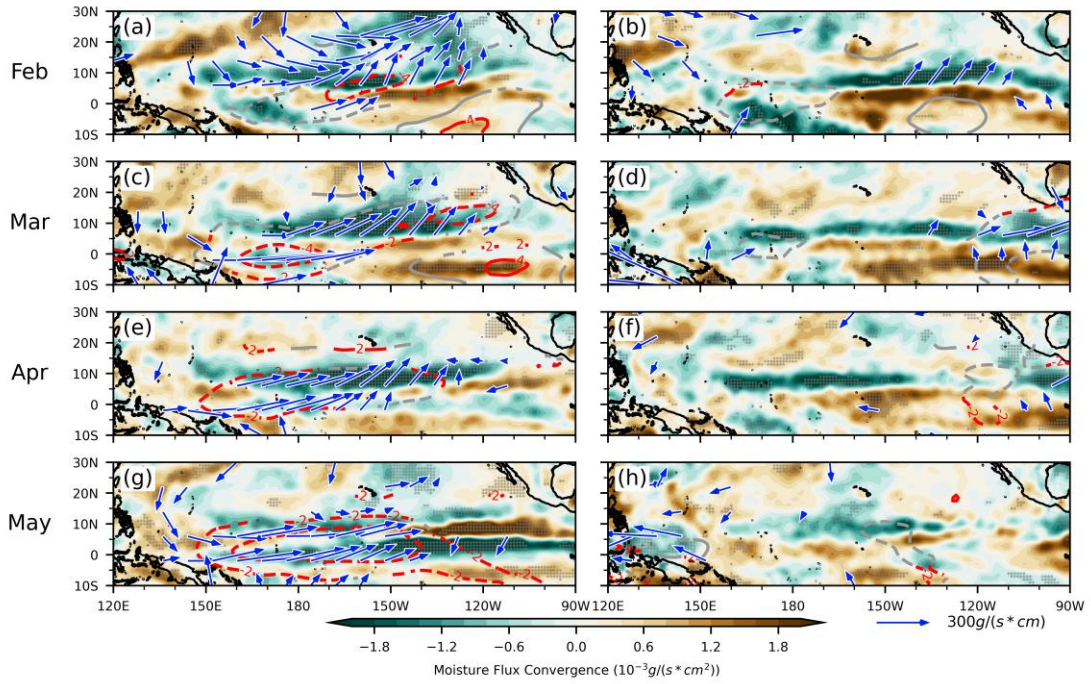


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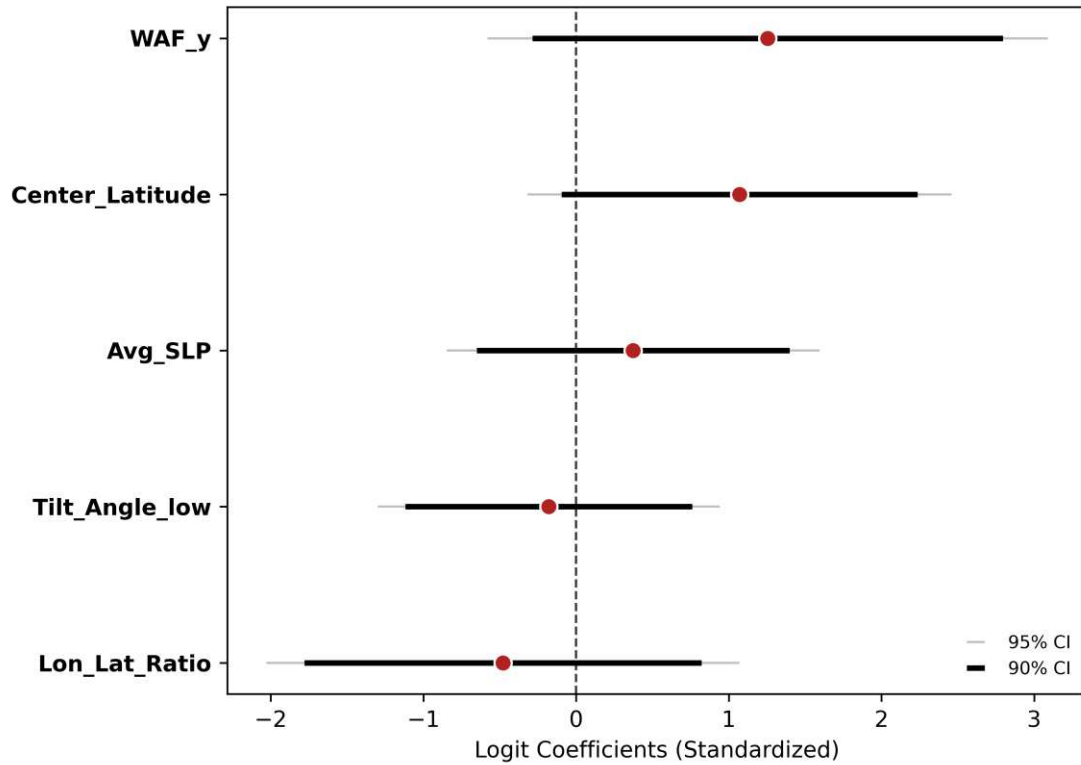


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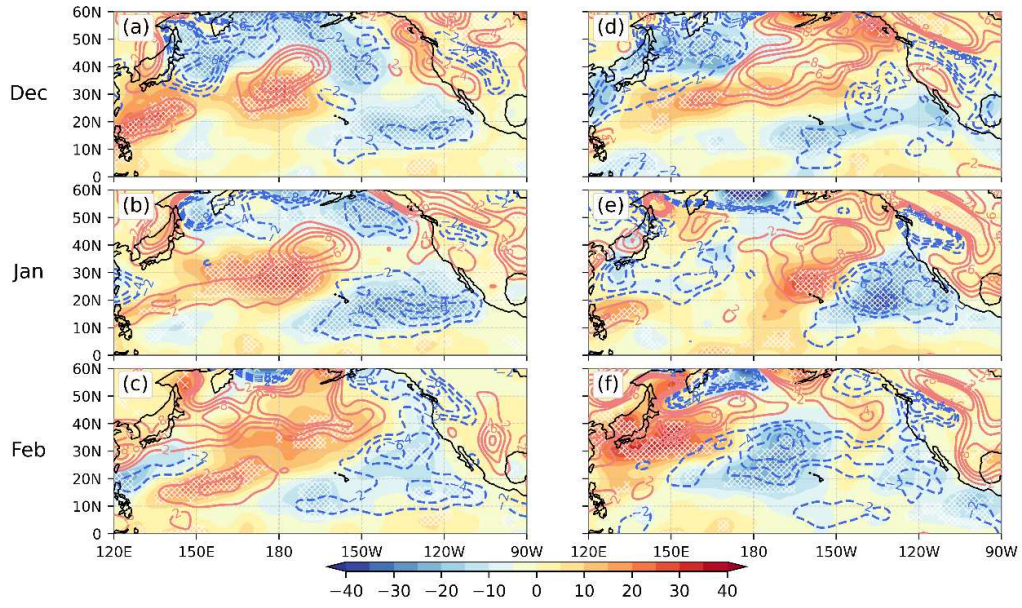


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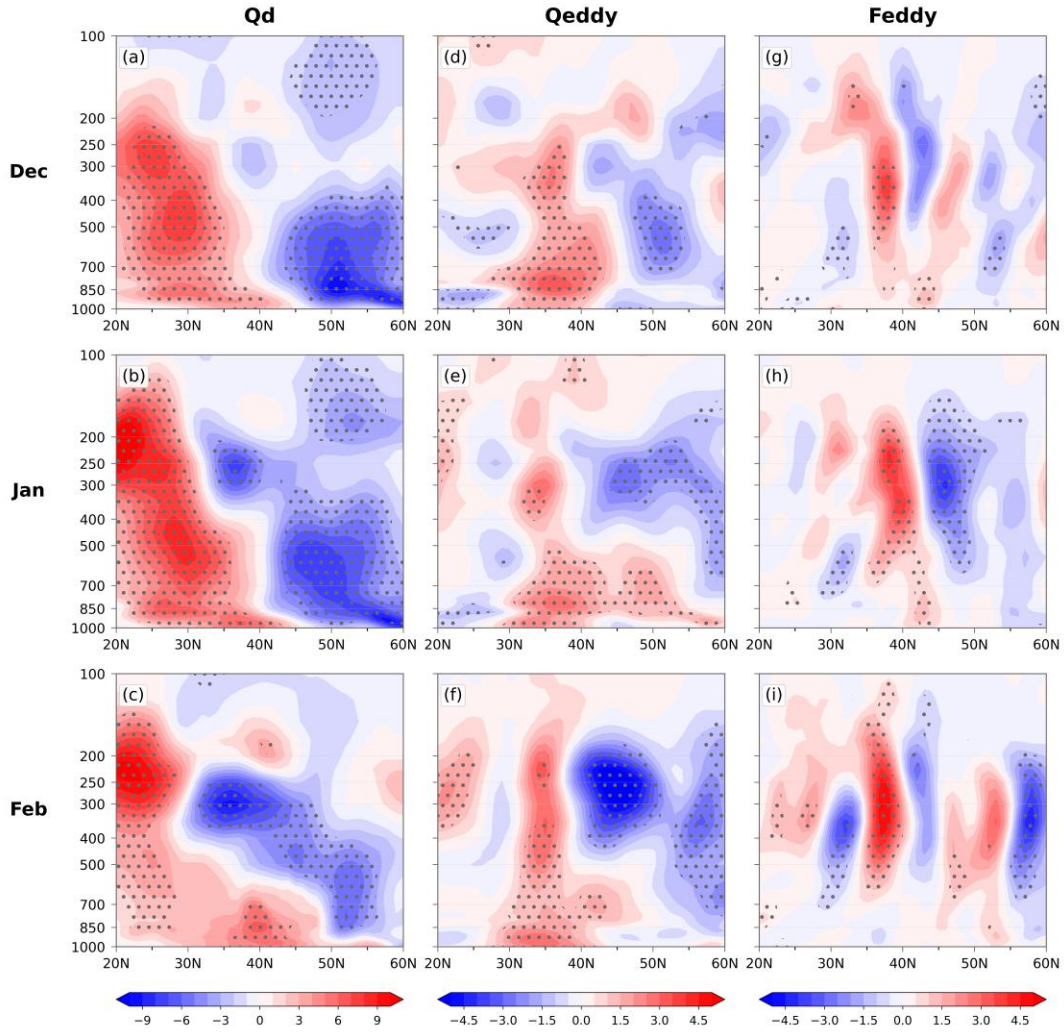


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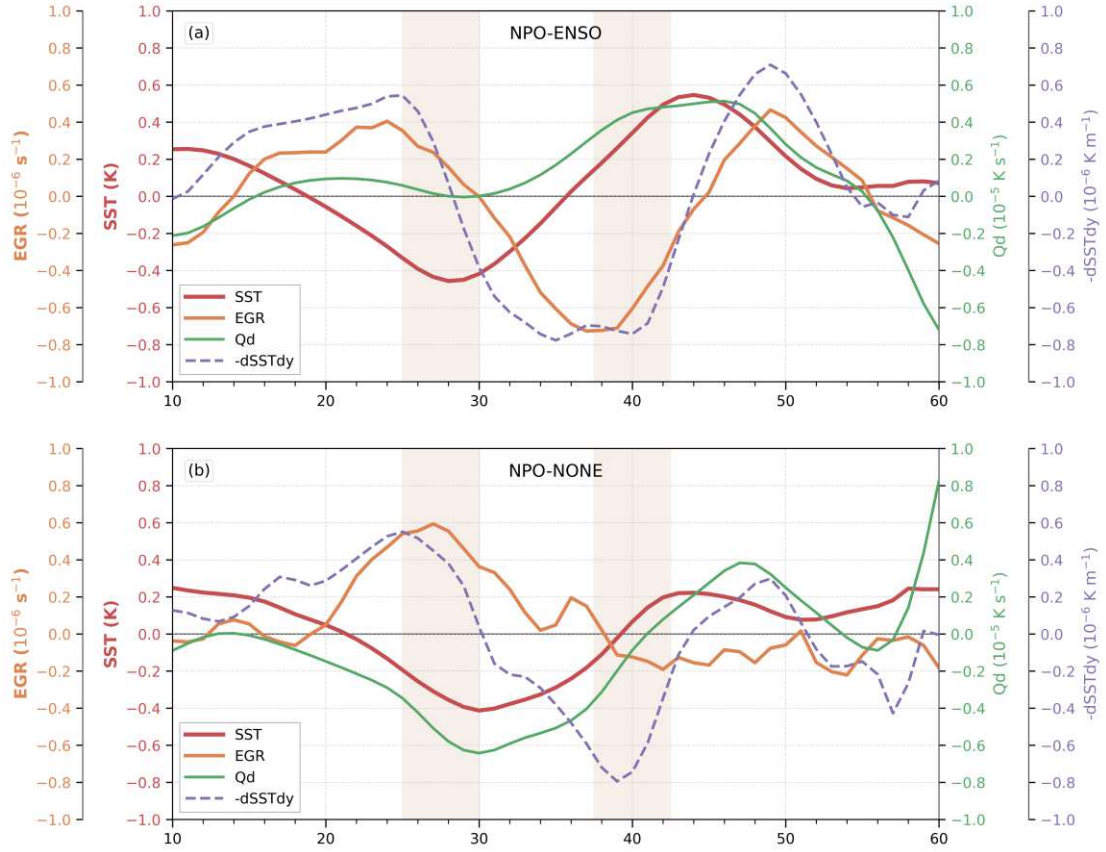


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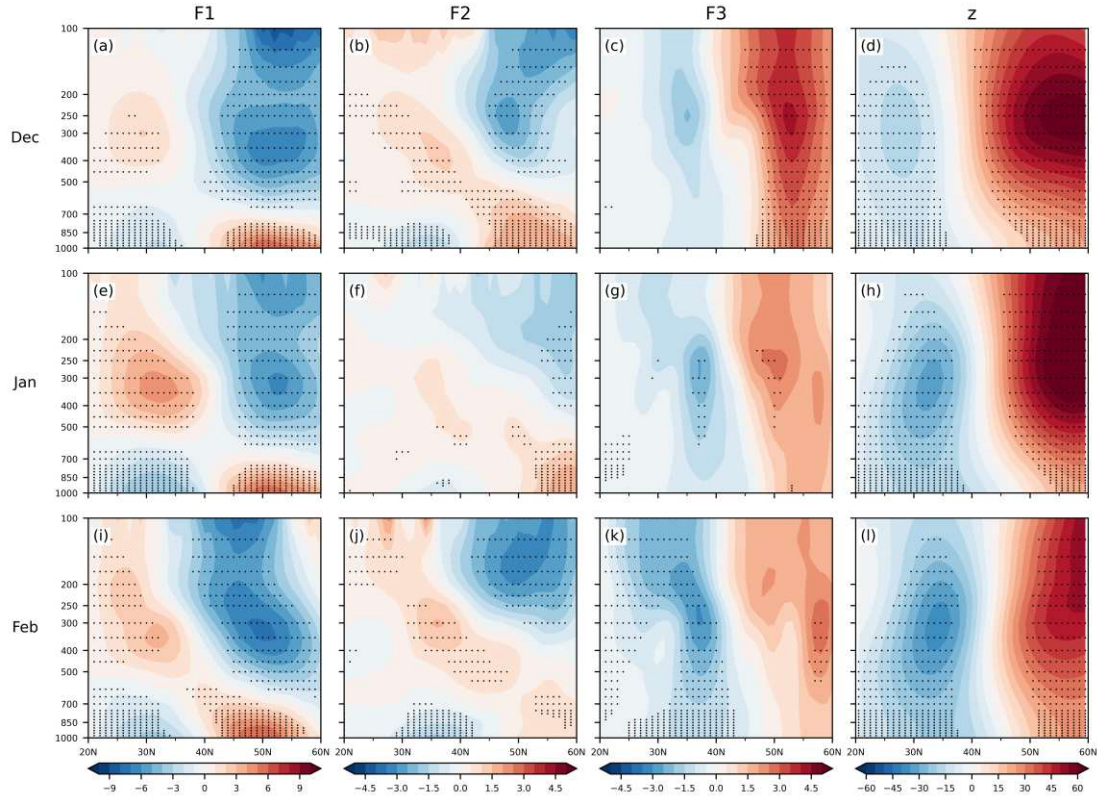


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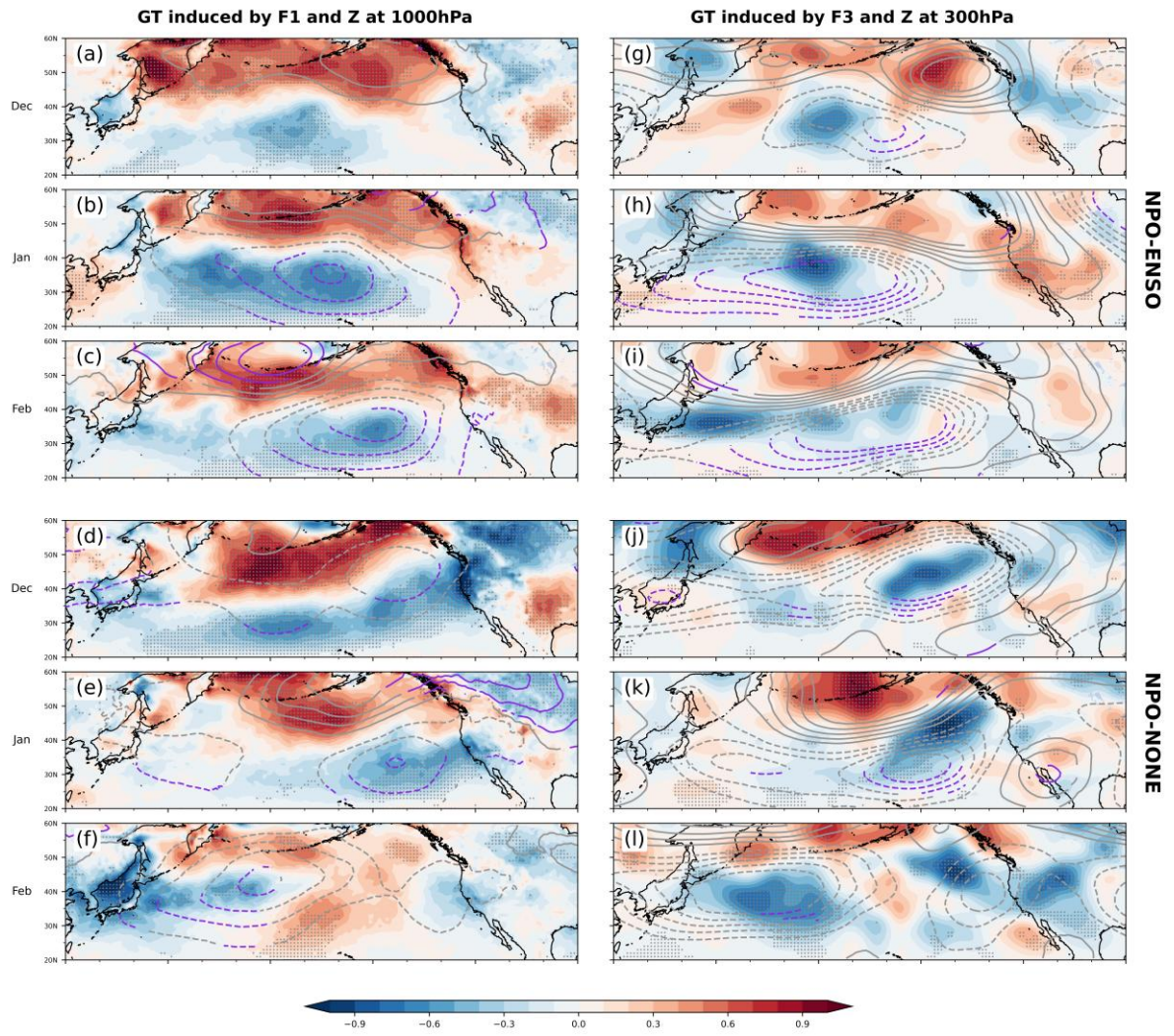


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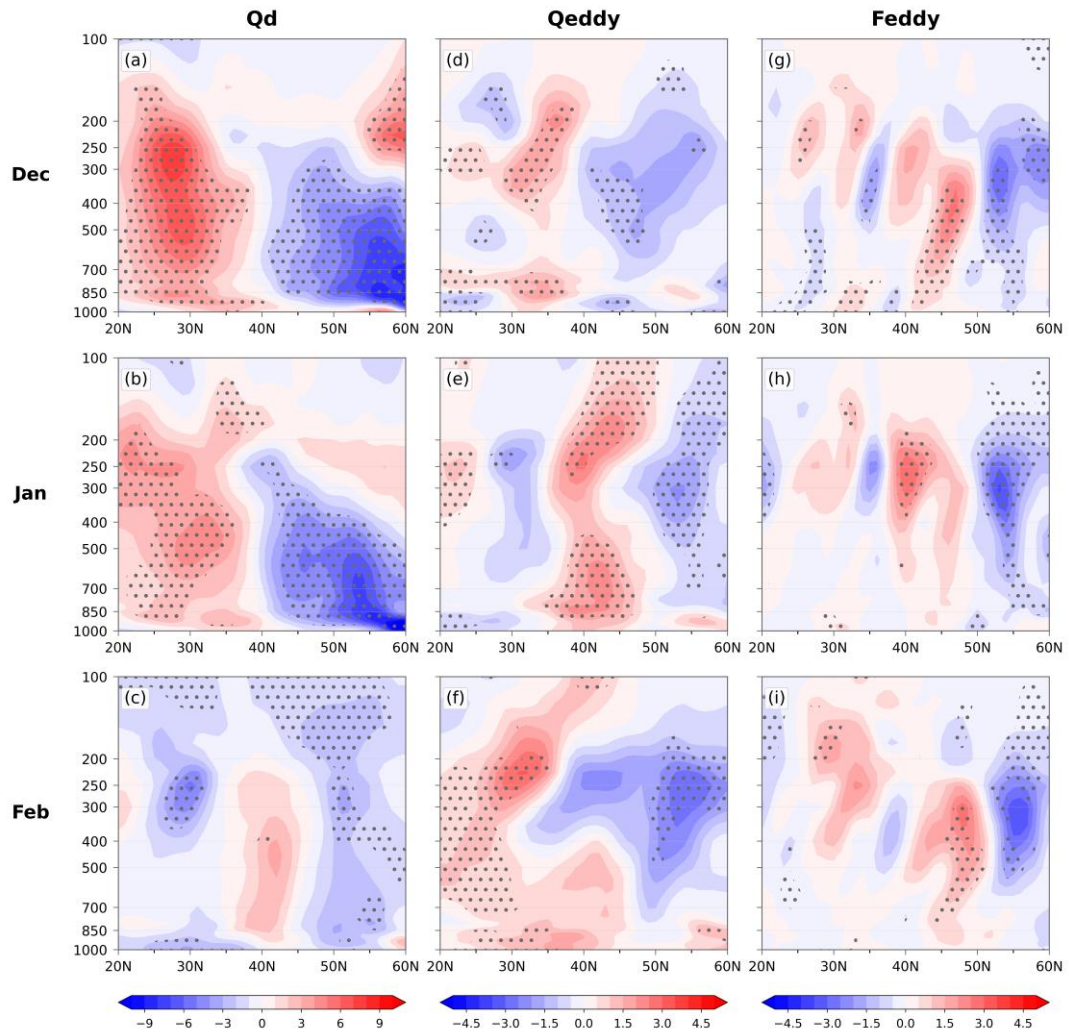


Fig. 13 Same as Fig. 9, but for NPO-NONE events

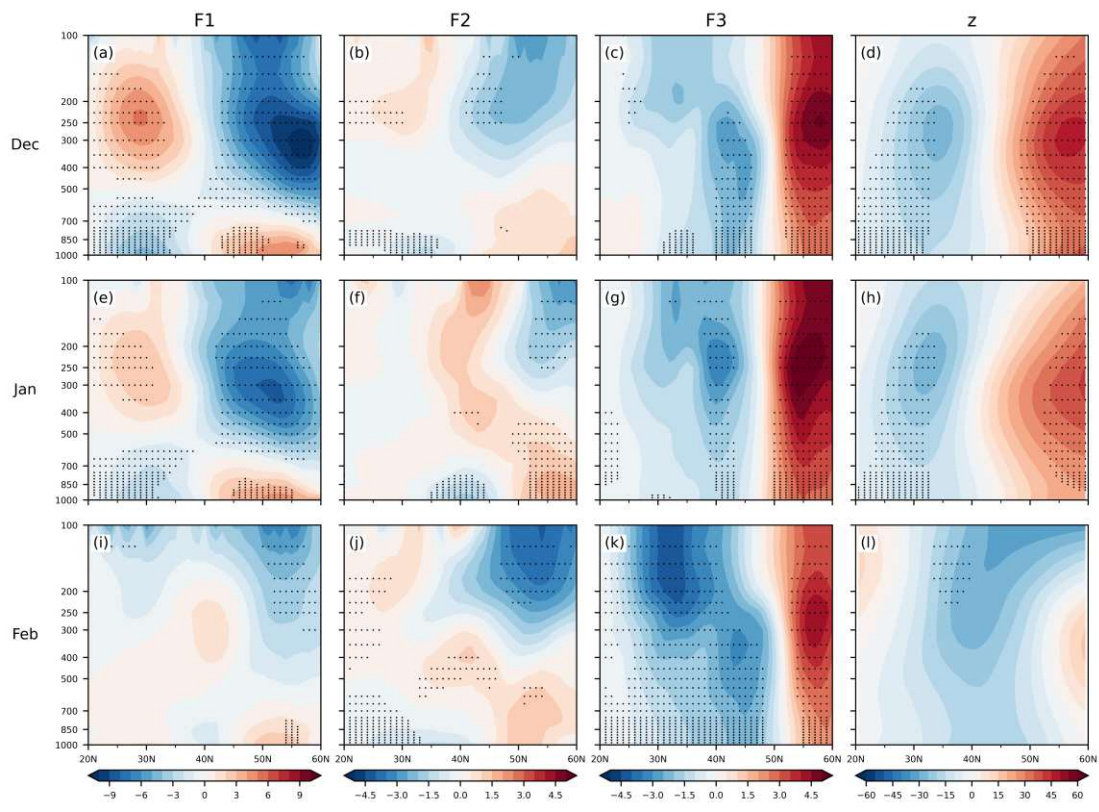


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