Effect of the Late-1990s Change in Tropical Forcing on Teleconnections to the Amundsen–Bellingshausen Seas Region during Austral Autumn

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ABSTRACT: Tropical sea surface temperature (SST) and associated precipitation, acting as diabatic heat forcing, has far-reaching climatic impacts across the globe through exciting poleward-propagating Rossby waves. It is found that the leading mode of tropical Pacific forcing in austral autumn experiences a significant interdecadal shift from an eastern Pacific (EP) to a central Pacific (CP) type around the late 1990s. More specifically, the EP-type precipitation anomaly mode before 1998 drives a quadrupole-like teleconnection pathway emanating from the tropical Pacific to the Ross Sea and Amundsen–Bellingshausen Seas (ABS) region, whereas the CP-type mode after 1999 excites a Pacific–South American (PSA)-like teleconnection orienting along a great circle. Divergent flows induced by different precipitation anomaly modes primarily determine the generation of Rossby waves by means of the vortex stretching and vorticity advection processes. Furthermore, the synoptic high-frequency transient eddy activity along with its dynamic forcing effect differs greatly before and after 1998/99, contributing to different locations of the teleconnection lobes at mid- to high latitudes. In contrast, the subseasonal low-frequency transient eddy activity exerts a limited influence. Our findings also indicate that the CP-type (EP-type) tropical forcing mode could significantly modulate the zonal displacement (strength) of the Amundsen Sea low, which could lead to distinct climate responses of West Antarctica and the Antarctic Peninsula in austral autumn.

KEYWORDS: Antarctica; ENSO; Teleconnections; Interdecadal variability; Tropical variability

1. Introduction

Climate variability over the Southern Hemisphere (SH), especially Antarctica, is beginning to be better understood in recent decades owing to the advent of the satellite era. As a result of the rapid temperature rising in West Antarctica and the Antarctic Peninsula (Bromwich et al. 2012; Steig et al. 2009; Vaughan et al. 2003), more attention has been devoted to variations of the SH atmospheric circulation. Recent literature found that some changes in West Antarctica and the Antarctic Peninsula cannot be fully explained by anthropogenic forcing (Bromwich et al. 2012; Li et al. 2014; Shepherd et al. 2018; Steig et al. 2009). Atmospheric teleconnections, which are significantly associated with the tropical variability, to the Amundsen–Bellingshausen Seas (ABS) region (Fig. 1a) have been proved important in the Antarctic climate modulation (Ciasto et al. 2015; Clem and Fogt 2013; Clem et al. 2017; Ding et al. 2011; Li et al. 2021; Purich et al. 2016; Steig and Ding 2013; Turner et al. 2016; Yiu and Maycock 2019). Consequently, improving the predictability of the SH polar climate calls for a better understanding of tropical–polar interactions, in which the SH atmospheric teleconnection represents a key role.

Tropical–polar teleconnections have been reviewed by Stan et al. (2017), Yuan et al. (2018), and Li et al. (2021). They provide a comprehensive overview of the characteristics of the SH atmospheric teleconnections and their importance in linking the tropical and polar climate. The Pacific–South American (PSA) pattern, one of the SH teleconnection patterns detected by previous researches (Mo and Ghil 1987; Mo and Higgins 1998; Szeredi and Karoly 1987), emanates from the tropical central Pacific and propagates to Argentina with large magnitudes in the PSA sector, and it is also recognized as an essential teleconnection to bridge the tropics and Antarctica. The teleconnection lobe near the ABS region—part of such a pattern—can modulate the strength and zonal oscillation of the Amundsen Sea low (ASL), which is characterized by a climatological low pressure located over the ABS region at the latitude band of 60°–75°S (shading in Fig. 1b). Faced with a deepening ASL in recent decades and its tremendous influence on the Antarctic climate (Raphael et al. 2016), it is important to access a great insight into the ASL: the variability, its linkage with tropical forcing, and its effect on the Antarctic climate.

Tropical forcing, El Niño–Southern Oscillation (ENSO) in particular, influences teleconnections to the ABS region via arousing a poleward-propagating Rossby wave train in the SH (Ciasto et al. 2015; Ding et al. 2011; Fogt and Bromwich 2006; Hoskins and Karoly 1981; Mo 2000; Trenberth et al. 1998; Wallace and Gutzler 1981). It has been previously noted that ENSO events could primarily drive the PSA pattern and then regulate the ASL strength (Fogt and Wovrosh 2015; Mo 2000; Stammerjohn et al. 2008; Turner 2004; Yuan 2004). Moreover, it has been demonstrated that the ENSO diversity, specifically...
eastern Pacific (EP) and central Pacific (CP) El Niño events (Ashok et al. 2007; Kug et al. 2009), could exert variable effects on the Antarctic atmospheric circulation and climate (Ciasto et al. 2015; Lee et al. 2010; Song et al. 2011; Zhang et al. 2021). With the late-1990s change in ENSO properties on the interdecadal time scale being revealed (Hu et al. 2020), studies intended to address the changed ENSO–SH climate connection have been increasingly undertaken in recent decades (Clem and Fogt 2015; Fogt and Bromwich 2006; Meehl et al. 2016; Yu et al. 2015). Most of these studies focus on the austral spring season, claiming that the link between ENSO and the SH atmospheric circulation is the closest then compared with that in other seasons (Jin and Kirtman 2009; Schneider et al. 2012; Yu et al. 2015; Zhang et al. 2021). However, the uniqueness of the ENSO-related tropical precipitation fluctuation in austral autumn [March–May (MAM)] has been exposed by plentiful evidence (Cai and Cowan 2009; Feng and Li 2011; Guo et al. 2016; Taschetto and England 2009). More specifically, one of the overarching features of interdecadal oscillation of MAM-mean precipitation anomaly over the tropical Pacific was highlighted in our previous work (Guo et al. 2016)—namely, the leading empirical orthogonal function (EOF) mode of MAM-mean precipitation anomalies, shown as Fig. 2 [modified from Figs. 2 and 3 in Guo et al. (2016)], exhibits a significant shift around 1998. The first EOF mode (EOF1) in 1979–98 (the pre-1998 period) is characterized by a zonal dipole pattern, showing positive precipitation anomalies over the equatorial EP and negative ones over the equatorial western Pacific corresponding to a positive principal component (PC1), referred to here as the EP-type precipitation mode (Fig. 2a). The counterpart in 1999–2021 (the post-1999 period) shifts to a zonal tripole-like pattern with a positive center over the equatorial CP and two negative centers over the western North Pacific and the equatorial southeastern Pacific, respectively, referred to as the CP-type precipitation mode (Fig. 2b). A similar interdecadal shift has also been recognized in the SST anomalies in austral autumn (Figs. 2c and 2d), which may result from a more frequent occurrence of the CP ENSO events after the 1990s (Hu et al. 2020; Yeh et al. 2009).

It has been gradually recognized that it is in austral winter, spring, and autumn that Rossby wave trains can propagate toward the South Pole owing to a sufficiently strong westerly jet, as reviewed by Li et al. (2021). During the El Niño episode, the subtropical westerly jet strengthens and the storm track in the South Pacific sector tends to shift equatorward (Yuan 2004). Bonding to the change in the storm track, behaviors of the transient eddy activity could significantly modulate the mean flow, which is considered as another essential process in the change of extratropical atmospheric teleconnections through the transportation of both eddy heat and momentum fluxes (Lau 1988; Lorenz and Hartmann 2001; Solomon 1997). Hence, investigating the role of the transient eddy activity on changing the SH teleconnections to the ABS region from an interdecadal perspective is an interesting but also uncharted issue.

Motivated by these unsolved questions, this study aims to address whether and how the atmospheric teleconnections to the ABS region would change as a result of an interdecadal shift of tropical Pacific precipitation anomaly pattern in austral

![Fig. 1](image-url)
autumn around the late 1990s. We also attempt to examine the underlying physical processes through the lens of Rossby waves and transient eddy feedbacks. In addition, whether such an interdecadal shift of tropical forcing affects the West Antarctic climate will be briefly discussed.

The remaining paper is laid out as follows. Data and methodology are given in section 2. The linkage between the tropical Pacific precipitation and the SH atmospheric teleconnections from an interdecadal perspective is described in section 3, and mechanisms of the tropically induced teleconnections are presented in section 4. Section 5 gives a discussion about the role of the changed teleconnections in modulating the climate of the Antarctic and its adjacent ocean. Section 6 summarizes the major findings.

2. Data and methodology

a. Data

Monthly variables, including geopotential height, horizontal wind, near-surface (2-m) air temperature, 10-m wind, and mean sea level pressure (MSLP), were obtained from the fifth generation of the global European Centre for Medium-Range Weather Forecasts (ECMWF) reanalysis data (ERA5) with a horizontal resolution of 2.5° x 2.5° (Hersbach et al. 2020). Daily fields of geopotential height, horizontal wind, and air temperature, derived from 12-hourly variables from the ERA5 dataset, were used to measure the transient eddy activities based on the Lanczos bandpass filter (Duchon 1979). In addition, monthly precipitation was obtained from the CPC Merged Analysis of Precipitation (CMAP) and Global Precipitation Climatology Project version 2.3 (GPCPv2.3) (Adler et al. 2003; Xie and Arkin 1997). Owing to the highly similar results obtained from CMAP and GPCPv2.3, only the results of data manipulation utilizing the GPCPv2.3 are shown in this study. The sea ice concentration (SIC) and SST were derived from the Met Office Hadley Centre’s SIC/SST datasets (Rayner et al. 2003).

The southern annular mode (SAM) index, defined by the zonal pressure difference between the latitudes of 40° and 65°S, is available at [https://legacy.bas.ac.uk/met/gjma/sam.html](https://legacy.bas.ac.uk/met/gjma/sam.html) (Marshall 2003). A positive SAM index corresponds to enhanced westerlies over the latitude belt of 50°–70°S and weakened westerlies of 30°–50°S (Fogt and Marshall 2020). The studying period spans from 1979 to 2021, with 43 years involved in total. We focus on the austral autumn season.
[March–May (MAM)], during which the interdecadal change in tropical precipitation anomalies is pronounced (Guo et al. 2016) and the SH westerly jet is sufficiently strong to allow the propagation of Rossby waves. Anomalies were calculated by subtracting the climatological mean of 1979–2021.

b. Rossby wave source

Following previous studies (Sardeshmukh and Hoskins 1988; Shimizu and de Albuquerque Cavalcanti 2011), the quasigeostrophic vorticity equation mainly concerning linear dynamics with tilting and nonlinear effects neglected can be written as

\[
\frac{\partial \zeta}{\partial t} = -\frac{1}{\rho} \nabla \cdot \mathbf{V}_\phi - \zeta \nabla \cdot \mathbf{V}_{\chi} - \mathbf{V}_{\chi} \cdot \nabla \zeta + \text{residual,} \tag{1}
\]

where \(\zeta\) is absolute vorticity, and \(\mathbf{V}_\phi (\mathbf{V}_{\chi})\) is rotational (divergent) wind. On the right-hand side of Eq. (1), the first term (referred to as \(P\)) represents the propagating term and the following two terms (identical to \(S\)) represent the Rossby wave source (RWS). The term \(S\) can be further rewritten in the form of the anomaly:

\[
S' = -\frac{1}{\rho} \nabla \cdot \mathbf{V}_\phi' - \zeta \nabla \cdot \mathbf{V}_{\chi}' - \mathbf{V}_{\chi}' \cdot \nabla \zeta' - \mathbf{V}_{\chi}' \cdot \nabla \zeta. \tag{2}
\]

In Eq. (2), the overbar is the climatology of autumn, and the prime denotes departures from the climatological mean of 1979–2021. Note that \(S\) consists of four physical processes: \(S_1\) is the vorticity source forced by the interaction between the anomalous divergence and mean vorticity, \(S_2\) reveals interaction between the mean divergence and anomalous vorticity, \(S_3\) denotes the mean absolute vorticity advection by anomalous divergence flow, and \(S_4\) represents the anomalous absolute vorticity advection by mean divergence flow. Usually, the sum of \(S_1\) and \(S_2\) is considered as the vortex stretching term, the sum of \(S_3\) and \(S_4\) is the vorticity advection term by divergent flows. The vorticity equation encompassing the RWS is a helpful diagnostic tool to examine the Rossby wave generations that are excited by tropical diabatic heating (Johnson and Kosaka 2016; Yiu and Maycock 2020; Zhu et al. 2020).

c. Eliassen–Palm flux

The horizontal component of three-dimensional Eliassen–Palm (E-P) flux proposed by Plumb (1985) was used to diagnose the propagation of Rossby waves, which can be calculated by the following equation:

\[
F_{EP} = p \cos \varphi \times \left[ \frac{\nu^2}{2 \Omega R \sin 2 \varphi} \frac{\partial (\nu \varphi)}{\partial \lambda} - \frac{g}{2 \Omega R \sin 2 \varphi} \frac{\partial (u \varphi)}{\partial \lambda} \right]. \tag{3}
\]

In Eq. (3), \(\nu\) and \(v\) are zonal and meridional geostrophic wind anomalies, respectively, \(\phi\) is geopotential height anomaly, \(\Omega\) is Earth’s rotation rate, \(R\) is the radius of Earth, and \(p\) = pressure/1000. Moreover, \(\varphi\) and \(\lambda\) represent the latitude and longitude, respectively. The E-P flux is widely used to measure the energy propagation of Rossby waves, as its direction is parallel to the group velocity of Rossby waves.

3. Connection between the tropical precipitation anomaly pattern and teleconnections to the ABS region

To capture the dominant coupled modes between the tropical Pacific forcing and the SH teleconnection patterns over the South Pacific sector in different epochs, the maximum covariance analysis (MCA) (Wallace et al. 1992) was conducted onto the MAM-mean precipitation anomalies and the SH geopotential height anomalies at 300 hPa before and after 1998/1999 (Fig. 3).

The first MCA mode during the pre-1998 period, explaining 62.6% of the total covariability, shows negative and positive centers of 300-hPa geopotential height anomalies over the tropical western and central-eastern Pacific near 10°S, respectively (Fig. 3c), locating south to the EP-type precipitation anomaly mode shown in Fig. 3a. By probing into the extratropical SH atmospheric circulation, a quadrupole-like teleconnection pathway is observed over the South Pacific (Fig. 3c), with four action centers near 35°S, 170°E; 60°S, 180°; 60°S, 80°W; and 30°S, 95°W, respectively.

For the post-1999 period, the first MCA mode during the post-1999 period explains 51.7% of the total covariance. Specifically, the CP-type precipitation anomaly mode (Fig. 3b) is correlated with a wavelike pattern along a great circle route, which shows an anomalous positive center near 10°S, 170°E, and other three alternate teleconnection actions appearing in sequence: east to New Zealand, the Amundsen Sea, and the Weddell Sea (Fig. 3d). This teleconnection pattern bridging the tropical Pacific and South Pacific resembles the PSA teleconnection first illustrated by Mo and Ghil (1987).

Similar MCA modes could be obtained by analyzing the MAM-mean SST anomalies and SH teleconnections (contours in Figs. 3a–d), as the tropical precipitation change is closely related to the SST in the tropics. It is well known that the tropical SST could change the convective activities and then the atmospheric heating by releasing plenty of latent heat. Redistributed atmospheric heating can excite poleward-propagating Rossby waves and modulate the extratropical atmospheric circulation. Some studies indicated that the SST–convection relation varies with regions and seasons (Lau et al. 1997; Sabin et al. 2013). Sometimes slight differences of the SST anomalies can lead to very different responses of tropical precipitation (Guo et al. 2017; Johnson and Kosaka 2016). Therefore, owing to a more direct excitation of teleconnections by tropical precipitation compared with that by SST, our emphasis is on the effects of tropical precipitation instead of SST on the SH teleconnections hereinafter.

Figure 4 portrays the vertical structures of the SH teleconnections associated with different tropical precipitation anomaly modes in the two epochs. It is obtained from the regressed atmospheric circulation anomalies at 850 and 300 hPa against the principal components of the EP-type and CP-type precipitation modes (Figs. 2a and 2b). During the pre-1998 period, an equivalent barotropic structure of teleconnection patterns is observed in the extratropical South Pacific with two action centers solidly lying close to 60°S, 160°W (60°S, 170°W) and
60°S, 75°W (60°S, 75°W) at 850 hPa (300 hPa), respectively (Figs. 4a and 4b). During the post-1999 period, the associated atmospheric circulation anomalies in both the lower and upper troposphere share a PSA-like teleconnection propagating along a great circle route (Figs. 4c and 4d), which bears a strong resemblance to the Rossby wave trajectory over the North Pacific sector elaborated by Hoskins and Karoly (1981). Moreover, these centers of action south of 50°S—that is, an anomalous anticyclone at 60°S, 100°W (60°S, 110°W) and an anomalous cyclone at 60°S, 15°W (60°S, 15°W) at 850 hPa (300 hPa)—also act together to vertically form an equivalent barotropic structure (Figs. 4c and 4d).

Zhang et al. (2021) examined the spring SH teleconnections responding to two types of warm ENSO events, EP and CP El Niño. They indicated that EP El Niño is accompanied by two branches of Rossby wave trains (their Fig. 4a), resulting in an anomalous anticyclone at 60°S, 100°W (60°S, 110°W) and an anomalous cyclone at 60°S, 15°W (60°S, 15°W) at 850 hPa (300 hPa)—also act together to vertically form an equivalent barotropic structure (Figs. 4c and 4d).

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It is also worth mentioning that SAM, acting as the dominant pattern of the SH atmospheric variability (Fogt and Marshall 2020; Marshall 2003; Thompson and Wallace 2000), may significantly modulate the SH teleconnections, paving the way to influence the ASL’s strength and location (Clem and Fogt 2013; Clem et al. 2017; Fogt et al. 2012). Importantly, we found that the MAM-mean precipitation anomaly modes (Figs. 2a,b) exhibit extremely weak connections with SAM, with small correlation coefficients of 0.09 (EP type) and −0.11 (CP type), respectively. It is reasonable to infer that these tropical precipitation anomaly modes in autumn could be treated as a climate forcing independent of SAM. Some papers considered SAM and ENSO as two large-scale internal climate variabilities manipulating the ASL variability (Clem and Fogt 2013; Clem et al. 2017). At the same time, others suggested that SAM and ENSO themselves are coupled together to some extent since ENSO events could
contribute to SAM by planetary waves in the stratosphere (Seager et al. 2003; Zubiaurre and Calvo 2012). Stemming from our results, SAM is practically independent of the above-mentioned tropical Pacific precipitation anomaly modes in austral autumn.

As introduced in section 2b, the quasigeostrophic vorticity equation [Eq. (1)] delineates several physical processes associated with the Rossby wave generation and propagation, which can be further estimated by different terms calculated by divergent and rotational horizontal motion [Eq. (2)]. It is interesting to know whether and how the associated divergent and rotational flows would differ in the two epochs. To address this question, Fig. 5 provides the regressed velocity potential and divergent wind anomalies, as well as the regressed streamfunction and rotational wind anomalies at 300 hPa against the corresponding PC1 of the dominant mode of the tropical Pacific precipitation anomaly during two subperiods.

For the pre-1998 period, there is a large-scale zonally oriented dipole pattern in tropics with anomalous upper-level (300 hPa) convergence over the western Pacific and divergence over the central-eastern Pacific (Fig. 5a), which is speculated to be induced by the tropical diabatic heating that is tightly associated with the EP-type anomalous precipitation mode. Upon examination of the rotational components, action centers of teleconnection, particularly those south to 30°S observed in Fig. 4b, are found to be mainly contributed by related rotational flows (Fig. 5c). Note that the signals of divergent flow almost vanish south to 30°S, contrary to significant centers of rotational flows in the extratropics, which is consistent with common knowledge that the tropical large-scale atmospheric motions are generally governed by divergent flows, and those in the mid- to high latitudes are mainly determined by rotational flows.

For the post-1999 period, a zonal tripole-like pattern of anomalous divergent flow exists in the tropics (Fig. 5b), characterized by divergence anomalies expanding over the tropical central-eastern Pacific and two anomalous convergence centers near the Philippines and tropical South America. This tripole-like divergent flow in the tropics is tightly coupled with the CP-type precipitation anomaly mode (Fig. 2b). By contrast, the associated rotational wind and streamfunction anomalies are characterized by a PSA-like wave train extending from the subtropical CP to the ABS region and then the Weddell Sea (Fig. 5d), which is significantly distinguishable from the rotational component of the teleconnection bonding to the EP-type precipitation mode (Fig. 5c). Discrepancies of the SH atmospheric circulation before and after 1998/99 make it reasonable for us to hypothesize that the interdecadal shift of the tropical precipitation anomaly mode might exert distinguishable effects on the SH teleconnections during different subperiods. The underlying mechanisms will be discussed in the following section.

4. Physical processes of distinct teleconnection associated with different precipitation anomaly modes

a. From the perspective of the Rossby wave source

Rossby wave propagation is emphasized by many previous studies for the rationale behind the stimulation of atmospheric teleconnections forced by tropical diabatic heating (Hoskins and Karoly 1981; Hoskins and Ambrizzi 1993; Trenberth et al. 1998; Wallace and Gutzler 1981; Webster and Chang 1998). Generations of the heat-induced Rossby waves would be diagnosed in detail according to the quasigeostrophic vorticity balance in this subsection.

1) PRE-1998 PERIOD

Figure 6 gives regressions of the anomalous RWS superimposed by the associated E-P flux, the vortex stretching term
[i.e., the sum of S1 and S2 in Eq. (2)], and the vorticity advection term [the sum of S3 and S4 in Eq. (2)] against the pre-1998 and post-1999 PC1s. Corresponding to a positive phase of EP-type precipitation anomaly mode during the pre-1998 period, an east–west dipole of RWS anomalies, with significant negative (positive) values over the subtropical western (central-eastern) Pacific south to the equator} highlighted as D1 (D2) in Fig. 6a—was found between 30° and 10°S. A striking southeastward propagation of E-P flux is observed south to D1, and then Rossby wave energy propagates eastward to the ABS region between 50° and 75°S. The flux subsequently changes its direction near the coast of South America and propagates northward to Chile. There is a slightly poleward propagation of Rossby wave energy south to D2 between 30° and 40°S, which may be important to forming the teleconnection centers of action over there. Linked to the generation of the vorticity source over the subtropical South Pacific and the propagation route of Rossby waves, this dipole pattern of RWS anomalies, especially the D1 center, provides a possibility for the efficient generation and poleward propagation of Rossby wave trains as shown in Figs. 4a and 4b.

Concerning the associated vortex stretching term (S1 + S2) and vorticity advection term (S3 + S4), in the first epoch, both of them feature a zonal dipole pattern in D1 and D2 (Figs. 6c and 6e), having many similarities to the RWS regressions in Fig. 6a. The relative contribution rate of the vorticity advection term to the total RWS within D1 is approximately 67%, whereas that of the vortex stretching term is 33% (Fig. 7a), implying a dominant role of the vorticity advection process in D1. Within D2, the vorticity advection term approximately contributes to less than one-third of the total RWS (28%), while the contribution rate of the vortex stretching term is 72%. It is interesting to note that the dominant physical process controlling the RWS anomalies in D1 and D2 differs substantially. Concerning a more important role of RWS anomalies within D1 in generating poleward-propagating Rossby waves than D2 (Fig. 6a), the vorticity advection process acts as a dominant contributor in inducing the local RWS and exciting the teleconnection patterns.

We also examined the propagating term [P in Eq. (1)] consisting of the rotational components associated with different precipitation anomaly modes in two subperiods (not shown). It is found that the propagating term plays a relatively minor role in contributing to the vorticity anomaly source, hence, the physical processes linked to RWS anomalies mentioned above are mainly discussed in this study.

The RWS change could be further decomposed into four interaction processes between the heat-associated perturbations ($\zeta'$ and $V'_x$) and mean state ($\overline{\zeta}$ and $\overline{V_x}$), termed as S1–S4 in Eq. (2). Their relative contribution rates are illustrated in Fig. 7b. In D1, S3 [$-V'_x \cdot \nabla f + \overline{\zeta}$] essentially hinges on the coupling of the anomalous divergent wind and the gradient of mean absolute vorticity, dominating approximately 53% of the local RWS anomalies. The spatial distribution of S3 shows negative anomalies in D1, where the isolines of mean absolute vorticity are dense (figure not shown). The second important factor is S1, expressed by $-(f + \overline{\zeta})\nabla \cdot V'_x$, which contributes
up to one-third of the total RWS (36%; Fig. 7b). Significant anomalies of S1 appear between only 30° and 10°S, where the anomalous convergence is large over the western Pacific, also accompanied by sufficiently strong mean absolute vorticity (figure not shown). These results highlight the considerable importance of the interaction between the heat-induced flow and mean states—particularly the coupling between subtropical divergent winds and the gradient of mean absolute vorticity (i.e., S3)—in generating Rossby waves.

2) POST-1999 PERIOD

It sequentially needs to be settled how the RWS anomaly associated with the CP-type precipitation anomaly mode behaves in the recent epoch. As shown in Fig. 6b, the CP-type precipitation anomaly mode links to significant positive RWS anomalies in the subtropical CP south to the equator (indicated as D3 in Fig. 6b, the region of 30°–10°S, 150°–110°W). Evident southeastward-propagation of Rossby wave energy along a great circle route appears over the South Pacific sector, extending eastward to the Weddell Sea (vectors in Fig. 6b). This confirms that the PSA-like wave train might be excited by the CP-type precipitation anomaly mode.

Upon examination of the regressed vortex stretching and vorticity advection processes after 1999, it is found that the vortex stretching term (Fig. 6d), compared with the vorticity advection term (Fig. 6f), shares a stronger resemblance to the total RWS anomalies (Fig. 6b) concerning both the spatial distribution and amplitude. Not surprisingly, the contribution rate of the vortex stretching term to the total RWS within D3 reaches 75%, much higher than that of the vorticity advection term (25%; Fig. 7a).

By analyzing behaviors of the S1–S4 anomalies associated with the CP-type precipitation anomaly mode, it turns out that S1 practically dominates the total RWS anomalies averaged within D3 with a contribution rate of 82% (Fig. 7b). More specifically,
the CP-type precipitation anomaly mode links to two significant anomalou
s divergence centers, situated over the CP and D3, respectively (contours in 
Fig. 5b). The divergence anomaly center in D3, together with large climatological absolute 
vorticity between 30° and 10°S (not shown), favors positive anomalies of S1 [−(f + \nabla \cdot \mathbf{V})] there, facilitating the in situ efficient generation of Rossby waves (Fig. 6b).

Based on the diagnoses by the quasigeostrophic vorticity balance and E-P flux in the foregoing investigation, disparities of the teleconnections to the ABS region can be well explained by the heat-induced Rossby waves, which propagate toward the South Pole. However, there is a debate as to whether Rossby waves can be regarded as a rationale for the ENSO-induced PSA teleconnection (Lou et al. 2021). Some studies raised a lack of direct dynamic evidence that Rossby waves can bridge the tropical heating and the PSA mode since the synoptic-scale Rossby waves will be trapped locally within the SH subtropical and polar jets (Karoly et al. 1989; Li et al. 2015; O’Kane et al. 2015). Apart from the Rossby wave excitation, another plausible mechanism involving the indirect effects of tropical heating on the SH subtropical jet by modulating the thermal winds is proposed to explain the ENSO–PSA connection (O’Kane et al. 2017). Our results, obtained from both the regression analyses of RWS and E-P flux anomalies, further consolidate the rationale that the poleward-propagating Rossby waves can link the tropical heating and the PSA teleconnections (Cai et al. 2011; Karoly 1989; Mo and Paegle 2001).

b. Role of transient eddy activities

Previous studies found that characteristics of the synoptic high-frequency (HF) and subseasonal low-frequency (LF) eddy activities exhibit various preferred sites and orientations in both hemispheres (Blackmon et al. 1977; Cai et al. 2007; Solomon 1997). The interaction between different time-scale eddies and mean flow serves as one of the underlying physical mechanisms accounting for the variability of the extratropical atmospheric circulation via transportation by the momentum and heat fluxes (Lau and Holopainen 1984). Notwithstanding many previous studies emphasizing the eddy–mean flow interaction in the winter hemisphere when the eddy activity is peaking, one may wonder whether the autumn eddy activity is sufficiently strong to modify the mean flow as well as the time-mean circulation. Hence, we compared the spatial distribution and magnitude of transient eddy activities in austral winter and autumn (Fig. 8).

The transient eddy activity is estimated by the root-mean-square of the 2–8-day (HF) or 10–90-day (LF) bandpass filtered geopotential height at 300 hPa. The maximum of HF (LF) transient eddy activity over the South Pacific sector reaches 975 (1240) gpm in austral winter, whereas the maximum is 852 (1215) gpm in austral winter. This result supports comparable variations of transient eddy activities over the South Pacific in the latitude band of 70°–40°S in between autumn and winter. Sequentially, the role of transient eddy activities in autumn will be addressed in this subsection.

Figure 9 presents the regressed maps of the MAM-mean HF and LF transient eddy activities against the PC1s of the EP- and CP-type precipitation anomaly modes in two subperiods. When PC1 anchoring in the positive phase before 1998, the HF transient eddies (shading in Fig. 9a) are reinforced along the midlatitude South Pacific between 60° and 30°S (with two anomalous centers locating near New Zealand and 50°S, 145°W) and suppressed near Argentina. Along 80°–60°S, the suppressed signals are confined north to the Ross Sea, while the intensified signals extend from the Amundsen Sea to the Antarctic Peninsula. In terms of the related LF transient eddies (contours in Fig. 9a), they are characterized by a similar distribution to the counterpart of the HF transient eddies (shading in Fig. 9a), but with faint and less significant signals near Argentina. This result implies a northward displacement of storm track over the South Pacific sector when the positive phase of EP-type precipitation anomaly mode is dominant in the first epoch, given that the climatological storm track center locates over the latitude band of 65°–50°S (Fig. 8a).
quantitatively measured by a quasigeostrophic potential vorticity tendency is associated with the convergence of transient eddy transports of heat and vorticity. Thus, the simplified formula of Eq. (4) can be rewritten as follows:

$$\frac{\partial \phi}{\partial t} = \nabla^{-2} \left[ -f \cdot \nabla \cdot \left( \nabla^c \phi' \right) \right] + \nabla^{-2} \left[ \frac{f^2}{g} \frac{\partial}{\partial p} \left( \nabla \cdot \left( \nabla^c \phi' \right) \right) \right] + \text{residual},$$

(5)

where $\phi$ and $\theta$ indicate geopotential height and potential temperature, respectively. Additionally, $g$ is the acceleration of gravity, and $\sigma$ is the static stability parameter. The overbar denotes the seasonal mean, and the prime represents components of filtered variables on the HF or LF time scale. Parallel lines ($\parallel$) indicate the SH average. According to Eq. (4), the quasigeostrophic potential vorticity tendency is associated with the convergence of transient eddy transports of heat and vorticity. Thus, the simplified formula of Eq. (4) can be rewritten as follows:

$$\frac{\partial \phi}{\partial t} = \nabla^{-2} \left[ -f \cdot \nabla \cdot \left( \nabla^c \phi' \right) \right] + \nabla^{-2} \left[ \frac{f^2}{g} \frac{\partial}{\partial p} \left( \nabla \cdot \left( \nabla^c \phi' \right) \right) \right] + \text{residual},$$

(5)

The abbreviations F1 and F2 represent dynamic and thermal forcing by the transient eddy activities, respectively. In sum, the geopotential height tendency is proportional to the anti-Laplacian values of convergence by the transient eddy vorticity fluxes (F1) and by the transient eddy heat fluxes (F2). Regressions of F1 on both HF and LF time scales against the PC1s of different precipitation anomaly modes in two sub-periods are given in Fig. 10.

During the pre-1998 period, the geopotential height tendency by the divergence of HF eddy vorticity fluxes (shading in Fig. 10a) is characterized by a zonal dipole pattern with significant negative and positive centers lying near 60°S, 165°W and 60°S, 85°W, which well matches the teleconnection lobes south to 50°S with the same signs (contours in Fig. 10a, also identical to the shading in Fig. 4b). This result exemplifies the assumption that the HF transient eddy activity may expect to favor the maintenance of the mid- to high-latitude action centers comprising the teleconnection pattern induced by the EP-type precipitation anomaly mode. On the other hand, the dynamic forcing of LF transient eddy activity has feeble effects, apart from facilitating the northward extension of the cyclonic anomaly south to New Zealand (Fig. 10c).
A more striking contribution of HF transient eddy activities in the second epoch, compared with that of LF transient eddy activity, is illustrated by Figs. 10b and 10d. Specifically, the geopotential height tendency induced by the divergence of HF eddy vorticity fluxes shows a relatively weak negative tendency near 45°S, 130°W, a positive tendency over the Amundsen Sea, and a negative tendency over the Weddell Sea (shading in Fig. 10b), which is in phase with the PSA-like teleconnection associated with the CP-type precipitation anomaly mode (contours in Fig. 10b). Note that the LF eddy-related signals are extremely weak along 60°S (shading in Fig. 10d). Therefore, the vorticity forcing by the HF transient eddy activities plays a prominent role to sustain the mid- to high-latitudinal action centers belonging to the PSA-like teleconnection during the post-1999 period.

In Eq. (5), the geopotential height tendency is proportional to the anti-Laplacian values of the vertical gradient of transient eddy heat fluxes, referred to as F2 (thermal forcing of transient eddy activities). The distribution of F2 is not displayed because the thermal forcing of both HF and LF transient eddy activities does not match the teleconnection patterns bonding to either EP-type or CP-type precipitation anomaly modes (figure not shown), suggesting that the transient eddy activities could not modulate the heat-induced teleconnections via their thermal effect, namely via the redistribution of eddy heat fluxes. It agrees with the understanding that the transient eddy activity helps to maintain the barotropic atmospheric teleconnection mainly via its dynamic forcing, instead of its thermal forcing (Lau and Holopainen 1984).

5. Does the shift of tropical precipitation anomaly mode matter on the West Antarctic climate?

The ASL is characterized by a climatological low pressure covering the Ross Sea and ABS region between 80° and 60°S, with a minimum near 75°S, 145°W in austral autumn (shading in Fig. 1b), whereas its standard deviation shows the most significant interannual variability over the ABS (contours in Fig. 1b). Furthermore, it has been perceived as one of the most important atmospheric internal variabilities that change the Antarctic climate (Clem et al. 2017; Hosking et al. 2013; Raphael et al. 2018; Raphael et al. 2016; Turner et al. 2013; Turner et al. 2016), which calls for a need to establish a link between the heat-induced teleconnections and the ASL variability in autumn.

To connect different tropical driven modes with the ASL variability as well as the Antarctic climate, regression maps of the MAM-mean MSLP, 10-m wind, and 2-m air temperature anomalies against the PC1s were given in Fig. 11. During the pre-1998 period, the EP-type precipitation anomaly mode is associated with a zonal dipole of MSLP anomalies with out-of-phase signals over the eastern Ross Sea and the Bellingshausen Sea (shading in Fig. 11a), suggesting that the transient eddy activities could not modulate the heat-induced teleconnections through their thermal effect, namely via the redistribution of eddy heat fluxes. It agrees with the understanding that the transient eddy activity helps to maintain the barotropic atmospheric teleconnection mainly via its dynamic forcing, instead of its thermal forcing (Lau and Holopainen 1984).
Amundsen Sea and Marie Byrd Land via the warm-air advection from mid- to high latitudes. Correspondingly, a decrease of associated SIC anomalies is observed in the Amundsen Sea, but is insignificant at the 90% confidence level (not shown).

The zonal dipole of MAM-mean MSLP identified by Fig. 11a reveals, to some extent, the longitudinal oscillation of the ASL, which bears a strong resemblance with the second EOF mode of MAM-mean ASL (not shown), accounting for 31% of the total variance. Hosking et al. (2013) emphasized that the ASL’s longitudinal position has an important influence on the surface climate of West Antarctic and its adjacent ocean. Our work suggests that in austal autumn, these teleconnections induced by prescribed tropical Pacific forcing could affect changes in the low-level circulation and surface temperature over the Southern Ocean and West Antarctic by modulating the ASL’s longitudinal position during the pre-1998 period.

In contrast, Fig. 11b shows the Antarctic atmospheric circulations with respect to the positive phase of the CP-type precipitation anomaly mode. An anomalous high pressure center appears over the ABS region (shading in Fig. 11b); to its east, an anomalous low pressure center locates over the Weddell Sea. This seesaw pattern of MSLP anomalies between the ABS and the Weddell Sea leads to anomalous southwesterly between 80° and 30°W and anomalous northwesterly between 150° and 110°W, which could bring cold air to the Antarctic Peninsula and the Weddell Sea (green contours in Fig. 11b) and warm air to the ABS region (red contours in Fig. 11b). This distinct contrast of surface air temperature anomalies over the Weddell Sea and the ABS region is similar to the so-called Antarctic dipole pattern (ADP), proposed by Yuan and Martinson (2000, 2001). Owing to the existence of ADP, the associated SIC anomalies significantly increase in the Weddell Sea and weakly decrease in the Amundsen Sea (figure not shown).

Figure 11 was also re-examined by removing the large-scale SAM-related signals through the method suggested by Hosking et al. (2013). The result is similar to Fig. 11, indicating that SAM has limited effects on changing the linkage between the tropical Pacific precipitation and the West Antarctic climate in austral autumn. On the other hand, the regressed SIC anomaly pattern in autumn shares a strong consistency with Figs. 2c and 2g in Zhang et al. (2021), who investigated the responses of the Antarctic sea ice to the CP and EP El Niño events. However, their work emphasized that this connection is strongest in austral spring, that is, September–November (SON), rather than the austral autumn we concentrate on. Note that our studies attempt to understand the plausible impact of different tropical Pacific precipitation anomaly modes in austral autumn on the teleconnection to the ABS region from the perspective of interdecadal time scale. Our findings suggest that the more frequent occurrence of the CP-type precipitation anomaly mode in the second epoch might contribute to significant changes in the ASL strength, and hence the sea ice variability over the Weddell Sea since the late 1990s, which has not been explicitly documented until now.

Concerning the effect of the tropical Pacific precipitation anomaly modes on the Antarctic climate, one may wonder whether or not the Antarctic climate experiences a significant interdecadal shift before and after 1998/99. Figure 12 gives the variability of the MAM-mean Antarctic climate, including the year-to-year variation of the ASL’s longitudinal position and the SIC in the Weddell Sea. The ASL’s longitudinal position is measured by the longitude of the minimum MAM-mean MSLP over the ABS region (60°–75° S, 170°E–70°W, highlighted as the black box in Fig. 1b). Furthermore, the sea ice change in the Weddell Sea is defined by the area-average MAM-mean SIC over 60°–75°S, 60°–10°W (outlined by the red box in Fig. 1b). Observational evidence suggests that the interannual variance of the ASL’s longitudinal oscillation in austral autumn tends to be suppressed after the late 1990s, with its standard deviation decreasing from 1.18 to 0.82 (significant at $p = 0.1$). It is consistent with the interdecadal shift of tropical Pacific precipitation anomaly modes mentioned above, as the EP-type precipitation anomaly mode, which is closely related to the longitudinal position of the ASL (Fig. 11a), is not dominant in the second subperiod anymore (Fig. 2).

Furthermore, the interannual variability of MAM-mean SIC anomalies in the Weddell Sea becomes amplified in the second subperiod, with a significant interdecadal increase of the standard deviation from 0.79 to 1.17 ($p = 0.09$). Such an interdecadal change might be partly explained by the perennial CP-type precipitation anomaly mode in the second epoch, which significantly modulates the variation of ADP and SIC anomalies in the Weddell Sea (Fig. 11b). Turner et al. (2020) recently noticed a dramatic decrease in summer SIC over the Weddell Sea in 2015/16. From their Fig. 1, the summer SIC in the Weddell Sea indeed exhibits a stronger interdecadal variability since the late 1990s, although their focus is on the change in 2015/16. Our result gives a hint that the autumn SIC in the Weddell Sea shows a similar feature from the interdecadal perspective, that is, a significant amplification of its interannual variance. It remains unclear whether and how the SIC change in the Weddell Sea is influenced by the tropical forcing from the interdecadal perspective. This work provides a clue of this issue, entailing further investigation in the future.

![Image](65x574 to 275x615)

![Image](65x646 to 275x687)
6. Summary and discussion

The linkage between the tropical forcing and the SH atmospheric circulation in austral autumn has not been widely discussed so far. As the most significant shift of tropical Pacific precipitation anomaly mode occurred in austral autumn around the late 1990s (Guo et al., 2016), the present work focuses on

...does not widely discuss the interaction between the EP and equatorial western Pacific, named the EP-type precipitation anomaly mode for short. It usually coincides with an EP El Niño–like SST distribution. The prescribed tropical forcing closely related to the precipitation distribution inclines to drive a quadrupole-like teleconnection pathway over the SH. This teleconnection emanates from the tropical western Pacific and orients southeastward to the Ross Sea and then the ABS region, accompanied by a clear poleward-propagating Rossby wave energy. The RWS-based diagnosis reveals that the vorticity advection process over the subtropical Pacific south to the equator dominantly contributes to the generation of Rossby waves.

In consideration of the mid- to high-latitude teleconnection lobes (south to 50°S) associated with the EP-type tropical forcing mode, there is a dipole pattern of geopotential height anomalies performing in an equivalent barotropic structure, with the two centers of action fixed over the Ross Sea and the ABS region, respectively. We interpret the formation and maintenance of such a dipole pattern from the view of not only the heat-induced teleconnection but also the dynamically driven amplification by the synoptic HF transient eddy activities near the storm track. It was also found that the subseasonal LF transient eddy activities have a weak influence on the action centers of the heat-induced teleconnection at mid-to-high latitudes. As a result of this dipole between the Ross Sea and the ABS region, the ASL is supposed to shift westward when a positive phase of the EP-type precipitation anomaly mode occurs. Both the Amundsen Sea and Marie Byrd Land tend to experience a warmer condition owing to the warm-air advection by the local anomalous northerly, which might fuel the sea ice melting in the Amundsen Sea.

During the post-1999 period, the prominent EOF mode of the autumn tropical precipitation anomaly shifts to a CP-type mode, driven by an interdecadal shift of the SST anomalies toward a CP El Niño–like pattern. The CP-type precipitation mode features an above-normal precipitation center over the CP with two significant below-normal ones to its western and eastern side. Further analysis suggests that a PSA-like teleconnection pattern along a great circle route usually coincides with the CP-type precipitation pattern in the second epoch. The PSA-like pattern may result from the heat-induced divergent flow over the subtropical CP south to the equator, which dominantly contributes to the generation of relevant Rossby waves via the vortex stretching process. In addition, the structure of synoptic HF transient eddy activities, distinct from the counterpart during the pre-1998 period, also contributes to the PAS-like teleconnection lobes south to 50°S via the dynamical forcing. Both the heat-induced and eddy-forced atmospheric circulation anomalies favor the variability of ASL’s strength and atmospheric circulations near the Weddell Sea in austral autumn, which could lead to the change in the surface temperature in the Antarctic Peninsula and sea ice in the Weddell Sea.

It is worth mentioning that such interdecadal changes in the first coupling MCA mode can only account for approximately half of the total covariability between the tropical Pacific forcing and upper-level geopotential height over the SH. Whether the interdecadal differences of these total fields are significant still needs to be further discussed. After analyzing the differences of the mean precipitation and 300-hPa geopotential height during the two subperiods (figure not shown), a dipole distribution has been observed with suppressed precipitation over the tropical central-eastern Pacific and enhanced precipitation to its northwest. Meanwhile, a PSA-like pattern is manifested in the interdecadal difference of 300-hPa geopotential height. Moreover, the high values of the interannual standard deviation of the tropical precipitation shift more westward to the CP in the second subperiod than those in the first subperiod. Accordingly, the standard deviation of 300-hPa geopotential height also gets larger over the ABS region and the Weddell Sea during the second subperiod. These disparities derived from both the climatological mean and their standard deviation resemble the first MAC mode after 1999 to some extent (Fig. 3). In other words, the interdecadal shift of tropical Pacific forcing noted in our study very likely dominates the interdecadal change in both the tropical Pacific precipitation and the SH atmospheric circulation.

Our attention was devoted to the tropically forced SH teleconnection in austral autumn. It has to be recognized that their interdecadal variability (e.g., the PSA) varies with seasons. Concerning that the CP ENSO events occur more frequently

FIG. 12. Normalized time series of (a) the ASL’s longitudinal position, defined by the longitude where the minimum MAM-mean MSLP over the ABS region (60°–75°S, 170°E–70°W; outlined by the black box in Fig. 1b) locates, and (b) the area-averaged SIC anomalies over the Weddell Sea (60°–70°S, 60°–10°W; outlined by the red box in Fig. 1b). The light gray bar represents the years when the interdecadal shift occurs. Numbers on both sides of this gray bar mean the standard deviation of each index before and after 1998/99. Numbers in parentheses represent the p value corresponding to the F test, which is used to estimate the significance of the interannual variance contrast in two epochs.
since the 1990s (Hu et al. 2020; Yeh et al. 2009), the ENSO-related tropical precipitation correspondingly would change not only in autumn but also in its mature phase, that is, the austral summer season (Jo et al. 2015). The leading mode of tropical Pacific precipitation anomalies exhibits a slight westward shift after 1999, and its associated teleconnections to the ABS regions differ as well in different epochs (not shown). Whether and how the interdecadal change in tropical forcing modulates the SH circulation and Antarctic climate in summer needs to be further analyzed.

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Data availability statement. The fifth generation of the global European Centre for Medium-Range Weather Forecasts (ECMWF) reanalysis data (ERA-5) are openly available at https://cds.climate.copernicus.eu/cdsapp#!/dataset/reanalysis-era5-pressure-levels-monthly-means?tab=overview and https://cds.climate.copernicus.eu/cdsapp#!/dataset/reanalysis-era5-pressure-levels?tab=overview. Precipitation data, including CPC Merged Analysis of Precipitation (CMAP) and Global Precipitation Climatology Project version 2.2 (GPCPv2.2), are provided by NOAA/OAR/ESRL PSL, Boulder, Colorado, USA, from their web site at https://psl.noaa.gov/data/index.html. Both the sea ice concentration (SIC) and sea surface temperature (SST) data analyzed in this study are derived from the Met Office Hadley Centre ice dataset at https://www.metoffice.gov.uk/hadobs/hadisst/data/download.html as cited in Rayner et al. (2003).

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