Seasonality and Dynamics of Atmospheric Teleconnection from the Tropical Indian Ocean and the Western Pacific to West Antarctica

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Abstract: The global impact of the tropical Indian Ocean and the Western Pacific (IOWP) is expected to increase in the future because this area has been continuously warming due to global warming; however, the impact of the IOWP forcing on West Antarctica has not been clearly revealed. Recently, ice loss in West Antarctica has been accelerated due to the basal melting of ice shelves. This study examines the characteristics and formation mechanisms of the teleconnection between the IOWP and West Antarctica for each season using the Rossby wave theory. To explicitly understand the role of the background flow in the teleconnection process, we conduct linear baroclinic model (LBM) simulations in which the background flow is initialized differently depending on the season. During JJA/SON, the barotropic Rossby wave generated by the IOWP forcing propagates into the Southern Hemisphere through the climatological northerly wind and arrives in West Antarctica; meanwhile, during DJF/MAM, the wave can hardly penetrate the tropical region. This indicates that during the Austral winter and spring, the IOWP forcing and IOWP-region variabilities such as the Indian Ocean Dipole (IOD) and Indian Ocean Basin (IOB) modes should be paid more attention to in order to investigate the ice change in West Antarctica.

Keywords: tropical-Antarctic teleconnection; barotropic Rossby wave theory; Indian Ocean and Western Pacific forcing; climate of West Antarctica

1. Introduction

West Antarctica is experiencing severe ice loss due to global warming. The ice sheet is losing mass faster than in early 2020 due to an increase of ice discharge [1,2]. The enhanced ice flow from the glaciers of West Antarctica is mainly attributed to the basal melting of ice shelves caused by the inflow of the Circumpolar Deep Water (CDW) from the nearby ocean. Since the ice shelf supports ice flows on the ground, a thinning of the ice shelves by basal melting produces huge amounts of meltwater from glaciers and consequently causes sea level rise [3]. These changes in Antarctic ice are closely related to atmospheric disturbance. The zonal wind at high latitudes derives Ekman transport offshore and the upwelling of the CDW along the coast, generating flow toward the underside of ice shelves, and leads to the basal melting of ice shelves (e.g., the Thwaites and Pine Island glaciers) [4–6]. The consequent retreat of the grounding line modifies ice-sheet instability, which induces massive ice discharge into the ocean. The meridional wind is also important for the poleward advection of warm moist air, which is responsible for sea ice melting, the accumulation of snowfall, and ice movement due to wind shear [7,8].

These seemingly regional circulation changes are mainly associated with global-scale atmospheric variability [9]. Many studies have found that high-latitude atmospheric variability such as the Southern Annular Mode (SAM), the Pacific-South America (PSA) pattern, and the Zonal Wavenumber 3 (ZW3) pattern are responsible for changes in sea ice, ice discharge, and the accumulation of snowfall over West Antarctica [10–14]. Other
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studies have suggested that tropical forcing can be a trigger for the high-latitude circulation anomaly [15,16]. For example, the Eastern Pacific variability such as the El Niño-Southern Oscillation (ENSO) and the Interdecadal Pacific Oscillation (IPO) modulates the anomalous SAM and ZW3 pattern by forming teleconnection [17–19] and in particular causes the Amundsen Sea Low (ASL) anomaly [20]. Furthermore, related to the recent dramatic ice discharge in West Antarctica, Nicolas et al. [21] and Holland et al. [22] found that the Eastern Pacific variability can accelerate the effect of global warming on ice loss in West Antarctica. Additionally, the Atlantic Multidecadal Oscillation (AMO) contributes a dipole-like sea ice concentration anomaly in West Antarctica during the austral winter by deepening the surface pressure in the Amundsen Sea as a part of globe-crossing teleconnection and has been revealed to account for the recent sea ice trend, which cannot be explained by the ENSO [23]. Since tropical forcing has a large influence in the high latitudes, there have been many attempts to examine the connection between the tropics and ice changes in West Antarctica.

However, the impact of the tropical Indian Ocean and the Western Pacific (IOWP) has not yet been fully understood because it is located far upstream. By conducting a coupled climate model simulation and regression analysis, Purich and England [24] and Wang et al. [25] documented that, unlike the teleconnection from the Pacific Ocean that may be reinforcing the ASL deepening, the teleconnection from the Indian Ocean tends to contribute to the formation of the ZW3 pattern. Several studies reported that during the austral winter, the influence of the IOWP can extend to West Antarctica [26,27], whereas Nunci and Yuan [28] argued that the wave train induced by the Indian Ocean Dipole (IOD) mode does not reach the Amundsen Sea and beyond during the austral spring. In the IOWP, the sea surface temperature has shown progressive warming due to global warming, and this region is thus expected to be increasingly influential for global circulation as a potential source of convection in the future. Therefore, it is essential to understand the dynamics of teleconnection from the IOWP to West Antarctica. Since the teleconnection shows different features depending on the season, each season should be examined separately.

Based on the Rossby wave theory, the formation of teleconnection from tropical forcing can be explained by two consecutive processes: the generation of the Rossby wave source (RWS) and the propagation of the barotropic Rossby wave. Diabatic forcing induces a divergent wind anomaly in the upper atmosphere and leads to the generation of the RWS by interacting with the climatological absolute vorticity [29–31]. The RWS is a dynamic source of propagating barotropic Rossby waves and acts as a catalyst in energy conversion from potential energy to rotational kinetic energy [32,33]. To adapt to the growth of the RWS, the barotropic Rossby wave consecutively arises and spreads, releasing kinetic energy. Generally, when the barotropic Rossby waves propagate, they are directed toward the pole at which the potential vorticity is large and the teleconnection is well-constructed in the winter hemisphere. There are characteristics of the Rossby wave propagation for identifying the preferred propagation region. For the stationary Rossby wave, (1) the wave is trapped inside the jet stream and moves eastward up to the exit of the jet, (2) the wave is reflected at a reflecting latitude where the meridional wavenumber of the wave is near zero, and (3) the wave dissipates its energy at a critical latitude where the background zonal wind is near zero and the meridional wavenumber is infinite [34–36]. Thus, the wave propagation depends on the background wind flow.

In this study, we theoretically explore the characteristics and dynamics of the IOWP-West Antarctica teleconnection using a primitive model simulation and Rossby wave theory. Its main goal is to determine the seasonal differences in the teleconnection for the four seasons.

2. Materials and Methods

2.1. Data Used

Monthly sea surface temperature (SST) data for the period 1870–2019 are obtained from the Met Office Hadley Center SST dataset (HadISST) version 1.1. For atmospheric
variables, daily ERA5 reanalysis data for 250 hPa over the period 1979–2019 are obtained from the European Centre for Medium-range Weather Forecasts (ECMWF). The anomaly for each variable is calculated by subtracting the mean value of each month or day to remove the annual cycle. December–February (DJF), March–May (MAM), June–August (JJA), and September–November (SON) are considered as the Austral summer, autumn, winter, and spring, respectively.

2.2. Model and Experimental Design

To investigate the circulation response generated by tropical forcing, a numerical experiment is performed using an atmospheric linear baroclinic model (LBM) that is available from: http://ccsr.aori.u-tokyo.ac.jp/~lbm/sub/lbm.html, accessed on 10 June 2021. The model has a horizontal resolution of T42 with 20 sigma levels. The horizontal diffusion has an e-holding decay time of 6 h and the vertical diffusion has a damping timescale of 1000 days. The timescale of linear drag is set to 0.5 days for the lowest three levels, 1 day for the topmost two levels, and 20 days for the middle levels to mimic Rayleigh friction and Newtonian damping. Each climatological seasonal-mean field of the three-dimensional wind, temperature, geopotential height, specific humidity, and mean sea level pressure is initialized as a background field in the model. To imitate the diabatic heating, circle-shaped heating rate with the 20° longitudinal and latitudinal scales is added to the temperature tendency equation at longitudes ranging from 40° E to 180° E at intervals of 20° along the equator. The vertical heating profile has a peak of 2.25 K day$^{-1}$ around the sigma level of 0.454. The wide IOWP forcing region is arbitrarily partitioned into the several independent forcing regions, but it helps to better understand the explicit teleconnection process. To concentrate the extratropical response and suppress the continuous tropical response, the forcing is turned on until day 15 (i.e., no forcing after day 15), so the modeled intensity of the response is slightly smaller than the observed intensity; however, this difference does not change our conclusion. The simulated circulation anomaly is plotted as an average over 20–25 days as the steady response.

The simulation performed here is designed to examine the steady response to the short-lived tropical forcing, but the result will be nearly similar to the response to the steady (or long-lived) forcing if the spatial features of the basic state and the forcing are maintained. Because the response to the steady forcing is structurally identical to the sum of the response to a sequence of pulses of the forcing [37]. Thus, the model has been used in teleconnection analysis on the inter-annual timescale as well.

2.3. Teleconnectivity

Teleconnectivity is calculated with the daily 250-hPa zonal-eddy stream function anomaly on a 2.5 × 2.5° grid. To reveal the strength of teleconnection, one-point correlation analysis is conducted for every grid point, then the strongest negative correlation in their respective correlation maps is plotted at the location of each grid point in the same way as in Wallace and Gutzler [38], i.e.,

\[ T_i = \left| \left( R_{ij} \text{ minimum for all } j \right) \right|, \]

where \( i \) and \( j \) are the grid points and \( R_{ij} \) is the correlation between \( i \) and \( j \). The absolute value of the teleconnectivity is taken and is multiplied by 100 for the sake of convenience. In the teleconnectivity map, the arrow connects the grid points for which the strongest negative correlation is exhibited in the respective one-point correlation maps.

2.4. Stationary Total Wavenumber

From the modified dispersion relationship [39,40], the stationary total wavenumber is calculated as follows:

\[ K_s^2 = \frac{\sigma_y k - \sigma_x l}{\sigma_M k + \sigma_M l - \omega'}, \]
where $\omega$ is the angular frequency; $k$ and $l$ are the zonal and meridional wavenumbers, respectively; and $\overline{\Pi}_M$ and $\overline{\Pi}_M$ are the basic zonal and meridional flow, respectively, on a Mercator projection. The case of $\omega = 0$ is determined to describe the stationary Rossby wave. $\overline{q}_x$ and $\overline{q}_y$ represent the zonal and meridional gradients of the absolute vorticity, respectively, and are expressed as follows:

$$
\overline{q}_x = \frac{1}{a^2 \cos \varphi} \left( \frac{\partial^2 \Pi}{\partial \lambda^2} - \frac{\partial^2 \Pi}{\partial \lambda \partial \varphi} \cos \varphi + \frac{\partial \Pi}{\partial \lambda} \sin \varphi \right)
$$

and

$$
\overline{q}_y = \overline{\beta}_M + \frac{\partial^2 \Pi}{\partial \lambda \partial \varphi} + \tan \varphi \frac{\partial \Pi}{\partial \lambda},
$$

where $\overline{\beta}_M = \frac{\partial f}{\partial \varphi} - \frac{\partial^2 \Pi_M}{\partial \varphi^2}$, where $f$ is the planetary vorticity. Here, $k$ is normally given as an integer and $l$ should be calculated by a cubic polynomial equation written as follows:

$$
f(l) = \Pi_M l^3 + k(\Pi_M - c)l^2 + \left(k^2 \Pi_M + \overline{q}_x\right) l + \left(k^2(\Pi_M - c) - \overline{q}_y\right) k,
$$

which is driven by the dispersion relationship. The parameter $l$ has at least one and at most three solutions. Positive and negative values of $l$ are associated with waves propagating northward and southward, respectively. If there is a local maximum $K_s$ ranging from $K_1$ to $K_2$, a wave with a zonal wavenumber between $K_1$ and $K_2$ is trapped in the localized $K_s$ region. In general, the stationary total wavenumber is plotted to identify the preferred region for the barotropic Rossby wave propagation since the wave is refracted toward the region with higher values of $K_s$ [35].

3. Results

3.1. Basic Flow and Teleconnectivity in Different Seasons

Before investigating tropical-extratropical teleconnection, the basic flow in different seasons should be examined. In Figure 1, the climatological zonal wind shows strong values in westerly regions near 30°S and 50°S—that is, the locations of the subtropical and polar jets, respectively—regardless of the season. The subtropical jet is strong in JJA and SON with a maximum near 180°E (Figure 1c,d) and disappears in DJF (Figure 1a). The polar jet has a similar intensity in all seasons but the longitudinal location of its center differs depending on the season; in DJF, the polar jet has two centers around 30–60°E and 100°W (Figure 1a), while the latter center does not exist in MAM/JJA/SON (Figure 1b–d). Thus, the polar jet in MAM/JJA/SON becomes zonally localized. An important feature of the austral winter is the presence of a double jet which is identified as the westerly regions with values greater than 35 m/s centered near Australia and 50°S, 40°E (Figure 1c). In the tropics, there is a region where the zonal wind is near to or less than zero, which extends from the tropical West Atlantic Ocean to the date line and South America in DJF/MAM (Figure 1a,b). According to the classic Rossby wave theory, the barotropic Rossby wave cannot penetrate this easterly region, so the wave only crosses the equator through the westerly region over the Central and Eastern Pacific; this tropical westerly region is referred to as “westerly ducts” [41]. During JJA/SON, since the easterly region appears over the tropics (Figure 1c,d), the cross-equatorial wave propagation cannot be easily explained [36]. However, in the modified Rossby wave theory considering an effect of the meridional wind, the background meridional flow makes it possible for the Rossby wave to pass through the easterly region [39,40]. Particularly, the dominant northerly over the eastern Indian Ocean and the Western Pacific in JJA/SON allows inter-hemispheric propagation, and these “northerly ducts” can be a waveguide for the southward-propagating Rossby wave (Figure 1g,h) [27,42].
The seasonal mean sea level pressure (MSLP) for the four seasons is shown in Figure 1 to demonstrate the annual cycle of the ASL, which is closely associated with the climate of West Antarctica and the surrounding ocean. The cyclonic circulation over the Bellingshausen and Amundsen seas shows a seasonal difference in terms of its location and strength; the longitudinal position is confined to the Amundsen sea in MAM/JJA/SON (Figure 1j–l) and the deepest low pressure appears in SON (Figure 1i). By developing an ASL index, Hosking et al. [43] calculated the relative ASL pressure with regard to the background pressure by subtracting the area-averaged MSLP over the Bellingshausen and Amundsen seas (60°–70° S, 170–290° E) to isolate the regional ASL variability and found that the strength of the relative ASL pressure is greatest in JJA (not shown). This seasonal ALS climatology is generated by the interaction between the topography of West Antarctica and the westerly wind jet [44]; therefore, it is speculated that the annual zonal-wind cycle plays an important role in the seasonal modification of the ASL. If atmospheric disturbance occurs by teleconnection, the ASL becomes deeper or weaker over time.

To identify teleconnection structures in the mid-to-high latitudes of the Southern Hemisphere, teleconnectivity maps are presented in Figure 2. In DJF, the regions of strongest teleconnection are found at latitudes lower than 60° S and are aligned in the zonal direction (Figure 2a). In MAM, the strongest teleconnection occurs farther south (Figure 2b), and in JJA/SON, a new region of maximum teleconnection emerges near the north of the Amundsen Sea and the Antarctic Peninsula (Figure 2c,d). This indicates that teleconnection patterns tend to occur more in the higher latitudes over the Western Hemisphere during the austral winter than the austral summer. Most of the strong teleconnectivity exists near jet streams. In JJA when the double jet occurs, the region with the strongest teleconnection is located between Australia and southern South America along the subtropical jet and it is located between the South Atlantic Ocean and the south Indian Ocean near 45° S along the polar jet. It is suggested that the jet acts as a waveguide in the teleconnection process. A similar structure is observed in standard deviation maps of the eddy stream function anomaly (Figure 3). In DJF, the region of largest standard deviation is confined to...
the lower latitudes (Figure 3a), but in JJA, it gradually migrates to the high latitudes and almost reaches West Antarctica (Figure 3b,c). This corresponds well to the annual cycle of the ASL, which in the austral winter is located farther southwest and has the lowest depth, while in the austral summer is located farther to the northeast and has the highest depth [43,45,46]. In MAM, a relatively small standard deviation is observed over Australia, and this area of low standard deviation moves southeastward beyond New Zealand during the next two seasons (Figure 3c,d). The existence of the area of low standard deviation indirectly represents that waves detour around this area. This behavior is also observed in the teleconnectivity map, which shows that strong teleconnection appears at 30–40° S and 50–60° S, but not at 40–50° S near New Zealand (Figure 2c); the former two latitude ranges are associated with the double jets. Thus, it can be inferred that the Rossby wave propagation along the polar jet is very likely to affect the atmospheric disturbance over West Antarctica during the austral winter.

![Figure 2](image-url)

**Figure 2.** The teleconnectivity of the daily 250-hPa eddy stream function anomaly for the four seasons. The values are multiplied by −100 and values smaller than 40 are not expressed. Contours greater than 40 m s⁻¹ are denoted with the interval of 5 m s⁻¹. Arrows indicate locations with maximum teleconnectivity. (a) DJF; (b) MAM; (c) JJA; (d) SON.

![Figure 3](image-url)

**Figure 3.** The standard deviation of the daily 250-hPa eddy stream function anomaly for the four seasons. The contour interval is 2.0 × 10⁶ m² s⁻¹ and the regions with values greater than 10 × 10⁶ m² s⁻¹ are shaded. (a) DJF; (b) MAM; (c) JJA; (d) SON.

3.2. Simulated Circulation in the LBM

To understand the formation process of the teleconnection developed by the tropical forcing over the IOWP, model simulations are performed. External diabatic heating is forced at varying longitudes at 20° intervals from 40 to 100° E for the Indian Ocean forcing and from 120 to 180° E for the Western Pacific forcing. For the sake of convenience, the simulation results are presented with the IOWP forcing divided into the Indian Ocean forcing and the Western Pacific forcing, because the wave emitted from the former forcing penetrates the tropics through the northerly duct over the eastern Indian Ocean, while the wave from the latter forcing passes through the northerly duct over the Western Pacific,
which has been documented in previous studies [27,42]. The simulations are re-performed with different backgrounds for each season with fixed forcing to isolate the effect of the seasonal cycle in the background state. Figure 4 presents the circulation response to diabatic forcing at 80° E and 160° E as representative experiments for the Indian Ocean and the Western Pacific forcing, respectively. For the response to the 80° E forcing, the simulation result shows a seasonal difference in the wave propagation pattern. There is a weak wave train for the SON/DJF/MAM background states (Figure 4a,b,d), while a strong wave train appears for the JJA background state (Figure 4c). In JJA, the wave starting near East Africa reaches the Ross Sea after propagating southeastward and returns to the lower latitudes near South America forming an arc-like route. It develops an anticyclonic circulation centered in the Ross Sea and a cyclonic circulation in the ocean north of the Amundsen Sea. In response to the 160° E forcing, similar circulation anomalies are observed in the Eastern Hemisphere in all seasons, although the anomaly is strongest in JJA (Figure 4e–h). The wave arises near the Indian Ocean and moves to the Ross Sea, generating a cyclonic circulation at the date line at 55° S, and subsequently propagates until the Weddell Sea with a weak anticyclonic circulation there in JJA. This result suggests that the teleconnection from the tropical forcing arrives in West Antarctica in the austral winter, whereas it does not in other seasons. During winter, the teleconnection is likely to contribute to heavy snowfall in West Antarctica, where the climatological snowfall is highest from May to September (not shown). An analogous result is obtained in previous studies which revealed that ENSO teleconnection to the Amundsen Sea is not strong in the austral autumn and summer [19,46].

Figure 4. The simulated zonal-eddy stream function anomalies at σ = 0.229 averaged over 20–25 days of simulation integration in response to the tropical forcing at 80° E (left panels) and 160° E (right panels). The initiated forcing is represented as blue dot. Each experiment is initialized with the climatological background states of (a,e) DJF, (b,f) MAM, (c,g) JJA, and (d,h) SON. The contour interval is $2.0 \times 10^6$ m$^2$ s$^{-1}$.

Figure 5 shows standard deviation maps of the circulation responses to the various forcings located between 40–100° E and 120–180° E. During JJA, the forcing over the Indian
Ocean induces strong atmospheric variation over the Ross and Bellingshausen seas, and the Western Pacific forcing leads to variation over the Ross and Weddell seas (Figure 5c,g). Although in the figure, a subtropical response seems to initiate in the west or east sides of the Indian Ocean in response to the Indian Ocean forcing or the Western Pacific forcing, the individual subtropical response to each forcing arises near the location of the forcing. Since the divergent wind anomaly emitted from the forcing produces the RWS, the longitude of the RWS is nearly in the vicinity of the tropical forcing even though the RWS can be amplified by the background flow [29,31]. Therefore, the effect of the season tends to be more critical for wave propagation than for RWS generation.

Figure 5 shows standard deviation maps of the circulation responses to the various forcings. The contour interval is 2.0 × 10^4 m^2 s^-1. The simulated zonal-eddy stream function anomalies are the same as in Figure 4 except for the location of the tropical forcing. The left panels (a–d) are for the simulations forced at various longitudes ranging from 40° to 100° E and the right panels (e–h) are for the simulations forced at various longitudes ranging from 120° to 180° E at 20° intervals. The contour interval is 2.0 × 10^6 m^2 s^-1.

3.3. Stationary Total Wavenumber

To clarify the wave propagation, the stationary total wavenumber (Ks) is plotted in Figure 6 for the negative meridional wavenumber (l < 0) that is associated with the wave heading toward the south. By examining Ks, we can diagnose where the wave propagates because the wave refracts toward the region with higher Ks values than the wavenumber of the wave. The cubic polynomial equation solution of Equation (5) for l is solved by priority, so the Ks value is calculated where there is a solution for l; wave propagation is not theoretically achieved except for where the solution exists. In the figure, the Ks value exists in the climatological northerly region in the subtropics (see Figure 1a–h). The wave can propagate into the Southern Hemisphere through the northerly near the IOWP during JJA and near the western Pacific during SON, while the wave can hardly penetrate the entire tropics except for the Eastern Pacific during DJF/MAM due to the existence of the southerly in this region (Figure 6a,b). This is the reason why the simulation result shows a weak wave train response when the climatological DJF/MAM background state is initialized (Figures 4 and 5).
Figure 6. (a) DJF; (b) MAM; (c) JJA; (d) SON. The 250-hPa stationary total wavenumber (Ks) (units of \(a^{-1}\), where \(a\) is the radius of the earth) for a zonal wavenumber of 3 and for negative meridional wavenumbers. White areas represent areas where there is no solution for \(l\). Red contours represent zero climatological zonal wind and black contours denote climatological zonal wind of 25 and 35 m s\(^{-1}\).

Meanwhile, it is suggested that the wave propagation in the mid-latitudes is more affected by the climatological zonal wind because the localized stationary total wavenumber is placed in the jet stream. During JJA and SON, the subtropical jet waveguide is placed along the strong westerly region near 25° S where the localized Ks appears (Figure 6c,d). The wave guide traps waves with shorter wavelengths in the jet and makes them propagate eastward to the exit of the jet. The trapped waves are well represented in the response to the Indian Ocean forcing, and have less impact on West Antarctica (Figure 4c,d). During JJA, to the south of the subtropical jet around 40° S, there is a region where the stationary total wavenumber is smaller than 2; this region is referred to as the reflecting latitude (Figure 6c). At this latitude, the meridional direction of wave propagation changes. For example, when the wave propagating southward reaches the reflecting latitude, the wave is reflected and moves towards the equator. In the same manner, if the wave propagates from high latitudes and is close to the reflecting latitude, the wave turns poleward; this forces wave trapping in the high latitudes [26,36]. In general, waves with longer wavelength propagate farther poleward than those with shorter wavelengths [34]. Thus, waves with longer wavelengths, which reach the high latitudes as they move southward from the Indian Ocean, are blocked by the reflecting latitude and cannot return to the tropics. Consequently, the waves zonally propagate and arrive in West Antarctica (Figure 4c,g).

4. Summary and Discussion

This study investigates how the teleconnection between the IOWP and West Antarctica differs seasonally and what mechanisms affect this difference. We used the modified Rossby wave theory in which the effect of the background meridional wind is considered to examine the practical wave propagation in the teleconnection process. The simulation results showed that, during JJA/SON, the wave response to the tropical IOWP forcing propagates into the Southern Hemisphere through the climatological northerly wind and moves southeastward along the jet waveguide. Waves with long wavelengths reach an area near to West Antarctica by trapping in the high latitudes due to the reflecting latitude.
Meanwhile, during DJF/MAM, since the climatological southerly wind is located in the IOWP region, the wave is more likely to propagate toward the Northern Hemisphere and not reach West Antarctica even if it moves southward. This result demonstrates the importance of the background flow to the teleconnection between the IOWP and West Antarctica and the significant effect of the IOWP on West Antarctica during the austral winter and spring.

The RWS is important in the teleconnection process since it provides information about where the wave practically starts. It is known that the location and the amplitude of the RWS depend on the season. Generally, the RWS is induced near the subtropical jet where the climatological absolute vorticity gradient is strong, and it acts as a medium of energy conversion from divergent flow to rotational flow, indicating that it is indeed a remote source of the barotropic Rossby wave [32,33]. Here, the RWS generated by the 80°E or 160°E forcing is mainly located near the subtropical jet close to the forcing, regardless of the season, and the amplitude is stronger in JJA and SON (not shown). It may induce the enhanced Rossby wave response in those seasons, but it does not mean that the stronger RWS helps the wave reach West Antarctica. Rather, the effect of seasonal modulation of the wave propagation by the basic state is more important to the teleconnection toward the high latitudes.

In an additional simulation in which the forcing near the Eastern Pacific is initiated for JJA, the cyclonic circulation over the Bellingshausen and Amundsen seas is roughly twice as strong as that of the IOWP forcing (not shown). It indicates that the strength of the response of the IOWP forcing is less than that of the Eastern Pacific forcing. Nevertheless, our result suggests that the tropical IOWP forcing during the austral winter and spring should be more investigated in order to better understand the ice change in West Antarctica. Contrary to the ENSO, which shows a peak in the austral summer, the Indian Ocean Basin (IOB) mode following an El Nino event has a peak in the early austral autumn and persists through winter, and theIOD known as an intrinsic mode peaks in the austral winter. The teleconnection generated by these IOWP-related variabilities modulates the ASL, which can influence the ice mass balance and total melting rate during winter, when snowfall peaks in West Antarctica. Recently, accelerated SST warming over the Indo-Pacific warm pool has been observed with the steepest trend slope than other oceans, so the global influence of the warming trend has been revealed as external convective forcing strengthens under global warming [47–50]. Thus, the impact of the IOWP forcing on the climate of West Antarctica is expected to grow in the future.

In this study, we analyze wave propagation operating in the climatological state. However, if there is a change in the background state, wave propagation can be expected to exhibit different behavior. The latest Coupled Model Inter-comparison Project phase 6 (CMIP6) model simulations reported poleward movement of the subtropical jet in response to CO₂ forcing (although its poleward shift is smaller than in the Coupled Model Inter-comparison Project phase 5 (CMIP5)), and the poleward jet shift is unanimously accepted [51,52]. Not only the tropical SST trend, but also the long-term climate variability, modifies the strength and location of the jet [15,53]. The changes in the jet and the strengthening of the forcing over the IOWP region have potential implications for the teleconnection between the IOWP and West Antarctica; however, the impact and detailed processes need to be further investigated.

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