Influence of tropical Atlantic meridional dipole of sea surface temperature anomalies on Antarctic autumn sea ice

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Abstract

Antarctic sea ice plays an important role in polar ecosystems and global climate, while its variability is affected by many factors. Teleconnections between the tropical and high latitudes have profound impacts on Antarctic climate changes through the stationary Rossby wave mechanism. Recent studies have connected long-term Antarctic sea ice changes to multidecadal variabilities of the tropical ocean, including the Atlantic Multidecadal Oscillation and the Interdecadal Pacific Oscillation. On interannual timescales, whether an impact exists from teleconnection of the tropical Atlantic is not clear. Here we find an impact of sea surface temperature (SST) variability of the tropical Atlantic meridional dipole mode on Antarctic sea ice that is most prominent in austral autumn. The meridional dipole SST anomalies in the tropical Atlantic force deep convection anomalies locally and over the tropical Pacific, generating stationary Rossby wave trains propagating eastward and poleward, which induce atmospheric circulation anomalies affecting sea ice. Specifically, convective anomalies over the equatorial Atlantic and Pacific are opposite-signed, accompanied by anomalous wave sources over the subtropical Southern Hemisphere. The planetary-scale atmospheric response has significant impacts on sea ice concentration anomalies in the Ross Sea, near the Antarctic Peninsula, and east of the Weddell Sea.

1. Introduction

Antarctic sea ice has a profound influence on the polar and global climate system (Li et al. 2021). While Arctic sea ice is rapidly reducing in association with increasing surface air temperature (SAT) (Stroeve et al. 2007), observations clearly show a gradually but uneven increasing trend of Antarctic sea ice extent before 2014 (Holland 2014), followed by a precipitous decline (Turner et al. 2017, Parkinson 2019). Antarctic sea ice changes are affected by several factors, including but not limited to anthropogenic forcing (Thompson and Solomon 2002, Arblaster and Meehl 2006), atmospheric circulation and oceanic heat flux changes (Hobbs et al. 2016), and internal multidecadal variability of the Antarctic–Southern Ocean system (Lecomte et al. 2017, Meehl et al. 2019, Zhang et al. 2019). From the perspective of atmospheric circulation, previous studies have focused on the tropical-polar teleconnections driven by tropical ocean variability (Yuan et al. 2018, Li et al. 2021).

The stationary Rossby wave dynamics is the most widely recognized tropical-polar teleconnection mechanism. Since El Niño–Southern Oscillation (ENSO) events dominate tropical climate variability on interannual timescales, the teleconnection associated with ENSO is used as the foundation for the understanding of the tropical–Antarctic connections (Alexander et al. 2002). Specifically, the positive sea surface temperature (SST) anomalies (SSTAs) in the
tropical Pacific trigger anomalous deep convection, and excite an atmospheric Rossby wave train referred to as the Pacific-South American (PSA) pattern (Mo and Higgins 1998, Kidson 1999, Zhang et al 2016). Low-level atmospheric circulation anomalies associated with the PSA alter the Antarctic sea ice (Kohyama and Hartmann 2016), through both thermal advection and wind-driven drift (Lefebvre et al 2004).

Such mechanism also operates establishing the link between the interdecadal variation of the Pacific SST and Antarctic sea ice (Ding et al 2011, Meehl et al 2016, Ding et al 2020), i.e. the Interdecadal Pacific Oscillation (IPO). Similarly, SSTAs from other tropical oceans have impacts on the Antarctic sea ice variability as well. In the Indian Ocean, the Madden–Julian Oscillation and the Indian Ocean Dipole are known to drive Rossby wave trains that propagate to high latitudes, contributing to the Antarctic climate variability on weekly- to-subseasonal (Pohl et al 2010) and interannual timescales (Cai et al 2014, Wang et al 2019), respectively.

For the Atlantic, recent studies have connected long-term changes of the Antarctic sea ice to the warming trend and multidecadal variability of tropical SST (namely, the Atlantic Multidecadal Oscillation, AMO). Since the satellite era, observed warming trends, associated with positive phase of AMO, in the tropical Atlantic deepen the Amundsen Sea Low (ASL; Hosking et al 2013), contributing to the increase of sea ice extent in the Ross Sea and the reduction in the Amundsen–Bellinghausen–Weddell seas (Li et al 2014, Simpkins et al 2014, 2016). Moreover, the strong pan-tropical interactions may lead to a nonlinear response of Antarctic climate to tropical SST variability, e.g. the negative IPO and the positive AMO can reinforce each other on multidecadal timescales (Li et al 2016, Cai et al 2019), deepening the ASL and bringing a broad impact on the Antarctic climate (Li et al 2021).

Amid extensive studies, how interannual equatorial Atlantic SST variability affects Antarctic sea ice remains largely unexplored. Here we examine the relationship between the tropical Atlantic and Antarctic on interannual timescales, focusing on the associated mechanisms based on the stationary Rossby wave analysis. The results show that SSTAs of the tropical Atlantic dipole mode drive a series of anomalies in atmospheric circulation that modulate the Antarctic sea ice via thermal and mechanical forcing.

2. Data and methods

For monthly SST, sea ice concentration (SIC), and other variables, we use the fifth generation European Center for Medium Range Weather Forecasts reanalysis (ERA5) with a 0.25° horizontal resolution and 37 vertical levels in the atmosphere components (Hersbach et al 2020). Monthly means from January 1979 to December 2020 are used to study the relationship between tropical Atlantic SST (20.5°N–20.5°S, 84.5°W–16.5°E) and Antarctic SIC (60.5°S–89.5°S).

The seasonal cycle and linear trend of SST and SIC are removed beforehand. The influence of the ENSO is also removed through a multiple linear regression approach (An 2003). By using empirical orthogonal function (EOF) analysis on the tropical Pacific (5.5°N–5.5°S, 170.5°W–120.5°E) SST, we get the first two principal components (PCs) $Z_1$ and $Z_2$ representing the ENSO signal (figure S1), then the new filed $\xi$ is derived from this formula: $\xi = \frac{\text{cov}(Z_1)}{\text{var}(Z_1)} Z_1 - \frac{\text{cov}(Z_1, Z_2)}{\text{var}(Z_2)} Z_2$, where $\xi$ denotes the original filed. Then, anomalies in atmosphere variables associated with an anomaly of tropical Atlantic SST are calculated using linear regression, and the significance is obtained by the t-test. We also utilize SST from Hadley Centre Sea Ice and Sea Surface Temperature (HadISST; Rayner et al 2003) to explore the long-term variation of the tropical Atlantic.

We apply the lagged maximum covariance analysis (MCA) to get the spatial distribution and temporal evolution of tropical Atlantic SST and Antarctic SIC monthly anomalies. The leading mode of MCA maximizes the covariance between the two fields, with the significance of MCA measured by the squared covariance (SC) representing the overall correlation (CO) degree of the two fields, the squared covariance fraction (SCF) representing the contribution to the total SC, and the correlation of time coefficients. The specific mathematical procedure of SC and SCF is introduced in the supplemental material text S1. Based on the Monte Carlo test, the confidence level is given by computing the MCAs with the original Antarctic SIC anomalies and 100 randomly permuted tropical Atlantic SST time coefficients. The probability distribution function of the 100 SCs/SCFs/COs is then constructed to rank the confidence level for the actual statistic.

3. Results

3.1. Tropical Atlantic teleconnection on interannual timescales: seasonal dependence

To examine the influence of tropical Atlantic SST on Antarctic SIC throughout the year, we apply the MCA as a function of time lag in month for anomalies of SST and SIC. Figure 1(a) shows the SCs of the leading mode of MCA, and its ordinate is the SIC calendar month and abscissa is the time lag in month, with a negative value indicating that SST leads SIC. It displays a strong seasonal dependence, i.e. the SCs are large when SST leads May SIC by 0–2 months, and small but insignificant when SST leads SIC in January and September. This indicates a strong influence of austral autumn tropical Atlantic SST on Antarctic SIC. The SCFs and COs are consistent with SCs, as shown in figure S2.
Figure 1. (a) Squared covariance (contours) of the leading maximum covariance analysis (MCA) mode between tropical Atlantic SST (20.5° N–20.5° S, 84.5° W–16.5° E) and Antarctic SIC (60.5° S–89.5° S, 179.5° W–179.5° E). Ordinate is the SIC calendar month and abscissa is the time lag in month, with negative for SST leading SIC. Shaded areas exceed 80% (dark blue), 85% (celeste) and 95% (yellow) confidence level based on the Monte Carlo test. (b) Heterogeneous map for SIC (shading, %) in May and (c) homogeneous map for SST (shading, °C) in March corresponding to the leading MCA mode at lag –2 month. (d) Corresponding normalized time coefficients of SST and SIC, and normalized NEA-minus-SEA time coefficients. Areas of > 95% confidence level are marked by black dots in (b) and (c).

Figures 1(b) and (c) display spatial patterns of May SIC anomalies and March SSTAs associated with the leading MCA mode at lag –2 months (SCF, 56%), depicted by regressions of SIC and SST onto the normalized SST time coefficients. The SIC anomalies show a tri-pole structure (figure 1(b)), accompanied by the aggregated SIC increase near the Antarctic Peninsula, and decrease in the Ross Sea and east of the Weddell Sea. This structure resembles the leading mode of the EOF of Antarctic SIC in May, accounting for 20% of the total variance (figure S3(a)). The CO coefficient between the MCA1 SIC time coefficients at lag –2 months and the PC1 of May SIC EOF is 0.81, suggesting large impact of the austral autumn tropical Atlantic SST on SIC anomalies.

Figure 1(c) shows the March SSTAs regressed against the MCA1 SST index (over the Atlantic and Pacific). SSTAs in the tropical Atlantic show a meridional-dipole-like pattern (hereinafter SST-dipole), with positive values north of the equator and relatively small negative values south of the equator. The SSTAs obtained by our MCA highly resemble the Atlantic meridional mode (AMM, Amaya et al 2016, Chiang and Vimont 2004) both in spatial distribution.
(figure S4) and interannual variation ($r = 0.77$). The AMM is defined as the leading MCA mode of the SST and the zonal and meridional components of the 10 m winds over the tropical Atlantic ($32^\circ$ N–$22^\circ$ S, $74^\circ$ W to the African coast; Chiang and Vimont 2004). The dynamic mechanisms of the AMM have been extensively studied, including the trade wind variations (Nobre and Shukla 1996) and wind-evaporation-SST (WES) feedback (Chang et al. 1997). The AMM modulates the interannual variability of the coupled ocean-atmosphere system over the tropical Atlantic, and is often considered independent of the ENSO (Chiang and Vimont 2004). SSTAs in the tropical Pacific, on the other hand, are much weaker than in the Atlantic (recall figure 1(b)), highlighting the role of tropical Atlantic SSTAs in modulating the SIC. The CO coefficient between the normalized SST and SIC time coefficients (figure 1(d)) is 0.66, and the power spectrums (figure S5) of them both exhibit a spectral peak around five years. At lag $-1$ and 0 month, the SIC anomalies (figures S6(b) and (d)) show a similar distribution as in figure 1(b). The dipole-like structure become more pronounced as the negative anomalies in the tropical South Atlantic intensifies from March to May (recall figures 1(c), S6(a) and (c)).

We define an index as the difference between the area-averaged SSTA (with ENSO influence removed) in north equatorial Atlantic (NEA, $0.5^\circ$ N–$20.5^\circ$ N, $84.5^\circ$ W–$16.5^\circ$ E) and south equatorial Atlantic (SEA, $0.5^\circ$ S–$20.5^\circ$ S, $84.5^\circ$ W–$16.5^\circ$ E), which is widely adopted in previous studies (Servain 1991, Melice and Servain 2003). The NEA-minus-SEA index is highly correlated with the MCA1 SST time coefficients at lag $-2$ months at 0.95, well above the 99.9% confidence level, and the regression of SIC anomalies against the NEA-minus-SEA index shows a similar tri-polar SIC anomalies as the one obtained by the MCA in figure 1(b). This result is independent from the MCA, reassuring the robustness of the MCA analysis. The time coefficients defined by the pan-tropical Atlantic area-averaged SST (the combined NEA and SEA region) have been used to study the impacts of the AMO and warming trends in the tropical Atlantic on long-term changes of Antarctic sea ice, especially in austral winter and spring (Li et al. 2014, Simpkins et al. 2014). Note that the NEA-minus-SEA denotes the regional difference (i.e. the meridional dipole mode), whereas the pan-tropical Atlantic region denotes area-averaged SST (i.e. similar to the equatorial zonal mode). Different from the tri-pole SIC anomalies (recall figure 1(b)), regression of Antarctic SIC on the pan-tropical SST displays less significant positive SIC anomalies in the Ross Sea. Thus, what we report in this study is an additional component, i.e. the tropical Atlantic meridional dipole-Antarctica teleconnections that operates on interannual timescales.

It should be noted that lagged response of the SIC primarily reflects the persistent SSTAs rather than the occurrence of the SST forcing in advance, as pointed out by Czaja and Frankignoul (2002). To explore the persistence of ocean and atmosphere variables, we take convective precipitation as a representative of atmosphere variables, and calculate the autocorrelations of NEA-minus-SEA index and area-averaged convective precipitation over the equatorial Atlantic within $5.5^\circ$ N–$5.5^\circ$ S, $84.5^\circ$ W–$16.5^\circ$ E from March to August (figure S7). The SST-dipole can persist for more than four months, which explains why the influence of SSTAs on SIC in late austral autumn can be detected using SSTAs in March. Indeed, MCA results also show that SST-dipole in the tropical Atlantic in March (figure 1(c)) persists into May (figures S6(a) and (c)). For convective precipitation, however, the autocorrelation between March and May is much lower and insignificant, indicating that signals of tropical atmosphere anomalies cannot persist into the subsequent months.

### 3.2. Tropical convection over the Atlantic and Pacific

The teleconnection from the tropical Atlantic to the Amundsen–Bellingshausen seas low is built upon a stationary Rossby wave mechanism (Li et al. 2014, 2015a, 2015b). Here we aim to establish a dynamical link between season-dependent SST dipole and SIC anomalies on interannual timescales. We start by analyzing regressions of various tropical anomalies in May with the normalized MCA1 SST time coefficients at lag $-2$ months. For the regressions in the following paper, either SST or SIC index can be used, as these two time coefficients represent the synergistic anomalies of SST and SIC (Zhang et al. 2018, 2021).

In response to the SST-dipole, both the tropical Atlantic and Pacific undergo a series of variations. For the Atlantic (see red box in figure 2), convective precipitation response shows a meridional-dipole-like pattern, with decreased precipitation over the equatorial Atlantic (figure 2(a), brown shading), especially on the east coast of tropical South America, and increased precipitation over $5^\circ$ N–$10^\circ$ N (figure 2(a), green shading). This decreased precipitation is associated with the intensified low-level divergence (figure 2(a), vectors) and the corresponding downward vertical motion over the equatorial Atlantic (figure 2(b), red shading). Figure 2(c) displays the structure of upper troposphere divergence (shading) and 200 hPa divergent winds (vectors). The anomalous winds converge over the equatorial Atlantic, and diverge over South America at $\sim 30^\circ$ S, suggesting the altered Hadley cell.

The SST-dipole in the tropical Atlantic also modulates the convective activity over the tropical Pacific (see blue box in figure 2). There are significant positive convective precipitation anomalies over
the equatorial Pacific despite much weaker SSTAs compared to the Atlantic (recall figure 1(c)). It is known that the Walker cell acts as an atmospheric bridge between the tropical Atlantic and Pacific. Here, the anomaly Walker cell is displayed as descending motion over the equatorial Atlantic and ascending motion over the equatorial Pacific (figure 2(b), shading), which forms a clockwise circulation together with the zonal wind anomalies (figure 2(b), contours). The increased precipitation over the equatorial Pacific is due to the ascending motion associated with the anomalous Walker cell in response to the SSTAs in the tropical Atlantic. The resultant anomalous divergence in the upper troposphere over the equatorial Pacific then induces anomalous convergence over the South Pacific via anomalous Hadley cell.

We suggest that anomalous warming of NEA in March promotes the development of SST dipole mode and the northward movement of convergence zone, including decreased convection over the equatorial Atlantic, and these anomalies peak in May. Such decreased convection drives enhanced convection over the equatorial Pacific. Thus, the tropical Atlantic-induced teleconnection can be expressed as two pathways: the one through perturbations of the local Hadley cell over the Atlantic and South America; and the other one that arises from the Atlantic–Pacific interaction. Both driven by the SST-dipole in the tropical Atlantic, the atmospheric responses over subtropical South America and South Pacific are opposite, which is key to generate the collaborative Rossby waves into mid- to high-latitude Southern Hemisphere (SH) as discussed in the next subsection.

Figure 2. Regressions of (a) convective precipitation (shading, mm) and 950 hPa wind (vectors, m s\(^{-1}\)), (b) vertical velocity (shading, Pa s\(^{-1}\)) and zonal wind (contours, interval 0.5 m s\(^{-1}\)) averaged over 2\(^\circ\)–2\(^\circ\) S, (c) 200 hPa divergence (shading, s\(^{-1}\)) and divergent wind (vectors, m s\(^{-1}\)) in May onto the normalized SST time coefficients at lag \(-2\) months. The blue and red box represent the Pacific and Atlantic sectors, respectively. In (a) and (c), green dashed lines represent the position of the equator. In (b), red (blue) shading represents downwelling (upwelling) flow, and solid (dashed) contours denote positive/eastward (negative/westward) winds. Areas of >95% confidence level are marked by white slashes in (b) or black dots in (a) and (c).
Figure 3. Regressions of (a) 200 hPa geopotential height (contours, interval 4 geopotential meter) and RWS (shading, s$^{-2}$) (b) 500 hPa geopotential height (contours, interval 4 geopotential meter) and TN flux (vectors, m$^2$ s$^{-2}$) in May onto the normalized SST time coefficients at lag $-2$ months. (c) Climatological mean of 200 hPa Rossby wavenumber $K$ (shading, m$^{-1}$) and zonal wind (contours, interval 3 m s$^{-1}$) in May. In (a), the pink and green lines indicate the propagation of the wave trains, and areas of $>95\%$ confidence level are marked by black dots.

The atmospheric responses at lag $-1$ and 0 month are consistent with that at lag $-2$ months (not shown). Similar Walker cell response has also been found in preceding study of the long-term warming trends in the tropical Atlantic SST in austral spring (Simpkins et al. 2016).

3.3. Development of Rossby wave train and responses around Antarctic

To consider the responses in subtropic regions, we first estimate the activity of the Rossby wave source (RWS). According to Sardeshmukh and Hoskins (1988), the RWS is given by $S = -\zeta_x D + \mathbf{v}_x \cdot \nabla \zeta_x$. The first term on the right-hand side accounts for the vortex stretching, and the second term represents the absolute vorticity gradient advection by the divergent flow. The specific mathematical procedure of RWS is introduced in the supplemental material text S2. Figure 3(a) (shading) shows the structure of anomalous RWS in response to the SST-dipole in the tropical Atlantic. These positive (negative) anomalies represent anticyclonic (cyclonic) RWS anomalies (RWSs),
which drive anticyclonic (cycloic) eddy stream function anomalies in the SH (Wang et al. 2019). Further inspection shows that the total RWS is dominated by the vortex stretching (figure S8). Due to the dependence on absolute vorticity, the RWSs are small within the tropics, while RWS appears active in the midlatitude of the SH. As pointed by Sardeshmukh and Hoskins (1988), Rossby wave activity in the upper troposphere is generated by divergent flow in the presence of nonzero absolute vorticity. Specifically, positive RWSs located near east coast of Australia and subtropical South America (figure 3(a), R1 and R4) are attributed to the corresponding anomalous divergent out-flow (recall figure 2(c), D1 and D4). Similarly, the pronounced negative anomalous RWS region emerges (figure 3(a), R2 and R3) over the subtropical central and eastern South Pacific, corresponding to the upper-level anomalous convergence (recall figure 2(c), blue shading) and the anomalous divergent in-flow (recall figure 2(c), D2 and D3). The Rossby waves produced by the RWSs manifest as alternating negative and positive 200 hPa geopotential height anomalies (figure 3(a), contours).

Next, we consider the evolution and propagation of the Rossby waves. Following Takaya and Nakamura (1997, 2001), the Rossby wave activity flux (TN flux) is used to investigate the Rossby wave propagation (specific mathematical procedure in the supplemental material text S3). This diagnostic tool illustrates the evolution and propagation of stationary Rossby wave disturbances on a zonally varying basic flow. For slowly varying basic flows, the TN flux is parallel to the local group velocity. Figure 3(b) (vectors) gives the TN flux anomalies in the SH. The RWS anomaly R3 and R4 (figure 3(a), shading) are the source of the wave train that arcs poleward to the midlatitude, and then eastward across the Antarctic Peninsula along the waveguide of the subtropical and subpolar jets (pink line in figure 3(a)). Meanwhile, there is another Rossby wave train originating over the central-western Pacific (recall R1 and R2 in figure 3(a)), that propagates southward and eastward (green line in figure 3(a)). This suggests that these RWSs over subtropical South America and the central-western South Pacific are effective to trigger stationary Rossby waves. The height anomalies at 500 hPa (figure 3(b), contours) show an almost identical pattern to the one at 200 hPa (recall figure 3(a), contours), suggesting the barotropic nature of the Rossby wave trains. These two wave trains meet over the Antarctic Peninsula, and then propagate together over the Southern Ocean in the form of alternating negative and positive height anomalies. At lag −1 and 0 month, the planetary-scale atmospheric responses are similar to lag −2 months (not shown). We suggest that the superposition of teleconnection from the Atlantic and Pacific constitutes the total atmospheric responses in the SH high latitudes.

The subtropical and subpolar jets play a critical role in the teleconnection, creating a Rossby waveguide. The climatological mean of Rossby wavenumber $K$ (figure 3(c), shading) and meridional vorticity gradient $\beta^\ast$ in May (figure S9, details of $K$ and $\beta^\ast$ in the supplemental material Text S4) illustrates the underlying cause of the waveguide and reflecting surfaces. Regions where $K$ (figure 3(c)) tends toward zero (the boundaries of the white regions) mark reflecting surfaces, which exist because of the curvature of the flow along the polar flank of the subtropical jet and equatorward flank of the midlatitude background jet stream (figure 3(c), contours). The polar reflecting surface is a consequence of the decline in $\beta^\ast$ (figure S9, associated with the decline in $\beta$) as one approaches the pole. The subtropical reflecting surface, however, owes its existence to the curvature of the flow ($U_y$) along the polar flank of the subtropical jet and equatorward flank of the midlatitude jet stream, which is sometimes sufficiently strong to overcome the gradient in planetary vorticity $\beta$. Regions of negative $\beta^\ast$ are marked in dark blue, and are most prominent over the west Pacific. This reversal in the vorticity gradient forms a ‘barrier’ to Rossby wave propagation, keeping the wave activity trapped in the extratropic.

Figure 4(a) shows the anomalies of sea level pressure (SLP) and 10 m winds with the normalized SIC time coefficients at lag −2 months. Positive SLP anomalies are centered over the Antarctic Peninsula, accompanied with anticyclonic circulations. The anomalous southerlies induce cold air advection and offshore drifts of sea ice, contributing to the increase of SIC near the Antarctic Peninsula (figure 1(b)). In contrast, the negative SLP centers near 150° W and 10° E over 50° S–60° S induce warm air advection and onshore drifts to the Ross–Amundsen seas and Atlantic–Indian basins near Antarctica, respectively, and lead to the decrease of SIC therein (figure 1(b)).

We then verify SAT anomalies over the SH high latitudes. There are warming indications in the Ross–Amundsen seas and east of the Weddell Sea, while an apparent cooling signal around the Antarctic Peninsula, agrees with the regression results of SIC (figure 1(b)). It has been reported that SAT datasets in reanalysis may be biased over Antarctica to some degree (Bracegirdle and Marshall 2012), so we further compare the SAT data measured by 40 stations in the Antarctic. The COs between SAT observations from 40 stations and normalized SIC time coefficients at lag −2 months are in good agreement with the reanalysis, especially in the Antarctic Peninsula (figure 4(c)). Positive COs with a high confidence level can also be seen near the Ross Sea and east Antarctica between 0° and 30°E, further supporting our analysis.
4. Conclusions and discussion

Our study identifies an interannual teleconnection between the tropical Atlantic meridional dipole mode and Antarctic SIC anomalies in the austral autumn season, as summarized schematically in figure S10. The meridional-dipolar SSTAs with positive (negative) SSTA north (south) of the equator are accompanied by the tri-polar SIC anomalies in the Ross Sea, Antarctic Peninsula, and east of the Weddell Sea. The teleconnection driven by the meridional SST-dipole in the tropical Atlantic have two pathways: (a) anomalous deep convection over the equatorial Atlantic drives the anomalous local Hadley cell and induces anomalous upper-level divergent flows over subtropical South America that generate anomalous RWSs; (b) perturbations to the Walker cell, driven by the Atlantic–Pacific interaction, promote development of RWSs of opposite sign over the subtropical South Pacific. A subsequent analysis based on the TN fluxes and geopotential heights shows that the wave trains originating from the two regions meet in the Antarctic Peninsula, and propagate eastward and poleward together. We suggest that the superposition of teleconnections from the Atlantic and Pacific constitutes the total atmospheric responses in the SH high latitudes. The resultant low-level response such as SLP and wind anomalies affect Antarctic SIC through thermal and mechanical processes. SAT variations in the reanalysis and station observations are both highly consistent with SIC anomalies.

Considering the seasonal differences of oceanic mode and atmosphere background, here we briefly discuss the reasons for the seasonal dependence of the above interannual teleconnection. The standard deviation (figure S11(a) and power spectrum of
NEA-minus-SEA (figure S11(b)) suggest the interannual variability of the tropical Atlantic SSTA peaks in austral autumn (March–May), followed by austral summer (December to February). The climatological zonal wind at 200 hPa (figure S11(c)) is larger in austral autumn than in summer, providing stronger background vorticity and forming stronger RWS (Li et al 2015b). We have noticed that although the zonal wind in austral winter is the largest in all four seasons, the weak SSTA variability (figure S11(a)) limiting the influence of the ocean on the atmospheric circulation. In general, only during austral autumn, the conditions are most suitable for the tropical-polar teleconnection reported in our study, that the RWS is strong enough and the strength of subtropical jet is appropriate to serve as a waveguide (Li et al 2015b).

In addition, an open question arises as to which of NEA and SEA plays a significant role? Although the Atlantic dipole mode starts from NEA warming in March, it further develops in March–May based on WES feedback (Amaya et al 2016, Chiang and Vimont 2004) and the SEA SSTAs is comparable to the NEA SSTAs in May. It is difficult to separate their effects in a statistical assessment. To reveal the relative importance of SST variations in the Atlantic dipole mode, sensitivity experiments conducted by a fully coupled climate model and its stand-alone atmospheric component are needed in the future study, respectively.

Figure S12(a) shows the 29 year running standard deviation of NEA-minus-SEA using the HadISST dataset with ENSO removed. The standard deviation of meridional dipole mode is stronger in austral autumn than in any other seasons, with strong decad variability. The standard deviation between the 1940s and 1990s is much higher than in the recent 40 years. Before the 1940s, the long-term trends of the SST in the NEA and SEA are almost in phase (figures S12(b) and (c)). They both decreased before 1910s and then increased until the 1940s. The variability of the NEA-minus-SEA is weak during this time. Then the NEA and SEA show opposite trends during 1940–1984, with NEA cooling and SEA warming, favoring the enhancement of the meridional-dipole-like SSTA. In the following decades, the SST trend in the NEA is reversed, i.e. NEA warming, again in phase of the SST trend of the SEA, weakening the NEA-minus-SEA variability. In fact, mechanisms for decadal-multidecadal variability of the NEA and SEA SST are relatively independent (Mehta 1998, Hurrell et al 2006). We suggest that the decadal variability of NEA represents the AMO with a period around 60 years. This indicates that the impact of the Atlantic SSTA on Antarctic climate may be modulated by the low-frequency decadal- to-multidecadal variation of the background SST. The detailed mechanisms of the decadal modulation of the tropical-polar teleconnection remain unclear and need further examination.

Data availability statement

The data that support the findings of this study are openly available at the following URL/DOI: www.ecmwf.int/en/forecasts/datasets/reanalysis-datasets/era5.

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