On the drivers of temperature extremes on the Antarctic Peninsula during austral summer

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Abstract
This study investigates the role of atmospheric circulation on intraseasonal and synoptic time scales in driving the temperature extremes over the Antarctic Peninsula (AP) during austral summer. It is found that the advection term induced by intraseasonal oscillations (ISOs) makes the largest contribution to the formation and development of temperature extremes. The synoptic variation-induced advection term affects the temperature anomalies around the peak time. The upstream ISO Rossby wave packet propagating along the jet stream south of Australia precedes the extreme temperature events. The ISO wave packet propagates eastward and eventually affects downstream circulation and relevant temperature anomalies over the AP. The perturbations triggered by the ISO wave packet can also gain energy from the mean flow through the potential energy conversion, contributing to the development and maintenance of ISO Rossby wave. The eastward propagation of the synoptic-scale Rossby wave is faster than that of the intraseasonal one. As such, the anomalous synoptic circulations rapidly change and facilitate the development of extreme temperature events. The synoptic Rossby wave activity before the warm events is much stronger than that before cold, which can be partly attributed to the background of more active baroclinicity before the warm events.

Keywords Temperature extremes · Antarctic Peninsula · Temperature advection · Rossby wave train

1 Introduction

Variations in surface air temperatures (SATs) have remarkable impacts on glaciological, oceanographic, chemical and biological processes in the Antarctic and surrounding ocean areas (Scambos et al. 2000; Cook et al. 2005; Meredith and King 2005; Turner et al. 2017). In recent decades, the Antarctic Peninsula (AP) has become one of the most rapidly warming regions in the world (King 1994; Turner et al. 2005; Turner et al. 2016; Marshall et al. 2006; Jones et al. 2019). Driving by the warming trend, the melting process of surface ice is the dominant cause of the disintegration and collapse of ice shelves over the AP (Scambos et al. 2000; Fahnestock et al. 2002; Cook et al. 2005). Furthermore, the warming trend and related sea ice loss along the western AP increased competition among krill-eating predators (Trivelpiece et al. 2011). In a warming climate, the cryosphere and ecosystems over the AP may become more susceptible to extreme temperature events. It is therefore important to investigate and understand the mechanism that drives the extreme temperature events over the AP. This can improve our understanding of the extreme SAT anomalies over the AP and provide important guidance for the subseasonal prediction of extreme temperature events, which is also one of the most important aims of the Polar Prediction Project (Jung et al., 2016).

Over the last few decades, many studies have investigated the mechanisms governing the trend and inter-annual
variability in the SATs over the AP. Several studies emphasized the important role of the tropical–polar teleconnections in influencing the SATs over the AP through Rossby wave dynamics (Ding and Steig 2013; Li et al., 2014, 2015). Specifically, the stationary Rossby wave trains generated by the tropical heating propagate poleward and eastward toward Antarctica, eventually affecting Antarctic climate. For example, the AP warming trend during the austral autumn, winter and spring can be partly attributed to the changes in the autumn sea surface temperatures (SSTs) in the tropical Pacific (Ding and Steig 2013). During austral winter, through deepening the ASL, the warming of SSTs associated with the Atlantic Multidecadal Oscillation also contributes to the warming trend over the AP (Li et al., 2014). The El Niño-Southern Oscillation (ENSO) and associated teleconnection between the tropical Pacific and high latitudes of southern hemisphere can affect the interannual variations of SATs over the AP during austral winter and spring (Marshall and King 1998; Clem et al. 2016, 2017). The above mechanism involves the formation of a Rossby waveguide across the Pacific, which depends critically on the intensity and extension of the subtropical jet (Li et al. 2015). In summer, the subtropical jet is too weak to trap the Rossby wave activity, leading to the breakdown of the tropical–polar teleconnection (Li et al. 2015). The polar front jet in the South Pacific acts as the waveguide (Hoskins and Ambrizzi, 1993; Ambrizzi et al. 1995), along which the stationary Rossby wave train propagates eastward and influences the SATs over the AP during summer (Wang et al. 2021), even in the absence of the tropical forcing. The variability in the phase of the Southern Annular Mode (SAM) can also exert influence on the trend and inter-annual variability in the SATs over the AP (Orr et al. 2004; Marshall et al. 2006; Marshall and Thompson 2016; Turner et al. 2020). During austral summer, a warming trend appeared over the northeastern AP (Turner et al. 2005; Orr et al. 2008), which can be partly attributed to the increasing intensity of the SAM (Orr et al. 2004; Marshall and Thompson 2006). In addition, the interannual variability in SAT across the AP is dominantly governed by the variability in the phase of the SAM (Marshall and Thompson 2016; Turner et al. 2020). The positive SAM leads to an increase in the SAT, whereas the negative SAM can decrease SAT over the AP (Marshall and Thompson 2016).

The aforementioned efforts help improve our understanding of the mechanisms of the interannual-multidecadal variations of SATs over the AP. However, the mechanisms of extreme temperature events over the AP during summer remain unclear. There have been few studies into extreme temperature extremes across Antarctica and the typical circulation anomalies for the extreme temperatures at some Antarctic stations. However, they did not investigate the dynamic mechanisms and relevant precursory signals of the extreme temperature events. Wang et al. (2021) proposed a mechanism that a Rossby wave along the polar front jet in the South Pacific can impact the inter-annual variability in the SATs over the AP during austral summer. This result inspires us to further explore whether this mechanism still works in driving the Peninsula-wide temperature extremes. Thus, this study attempts to reveal whether the Rossby wave along the polar jet can exert influence on the extreme temperature events over the AP. If so, the upstream Rossby wave packet can be considered as the potential precursory signal, which is essential for the prediction of the extreme temperature events.

The rest of the paper is organized as follows. The data and methods used in this study are introduced in Sect. 2. The results are presented in Sect. 3. Section 4 provides a summary and discussion.

## 2 Data and methods

This study used the daily air temperature, geopotential height, horizontal wind at multiple pressure levels, sea-level pressure (SLP), and SAT datasets, which were obtained from the European Centre for Medium-Range Weather Forecasts (ECMWF) ERA-Interim reanalysis (Dee et al. 2011). The atmospheric data from ERA-Interim reanalysis have a horizontal resolution of 2.5° × 2.5° and are available from 1979 to 2018. The ERA-Interim reanalysis has been widely applied for investigating the Antarctic climate variability (Turner et al. 2021; Wang et al. 2021) due to its good representation of recent Antarctic climate (Bracegirdle and Marshall 2012). This study also used observational SATs from eight weather stations at the AP, which can be downloaded from the Reference Antarctic Data for Environmental Research project (ftp://ftp.bas.ac.uk/scar/EGOMA/). Table 1 presents a brief description of these stations. The observational SATs were subject to quality control, which was detailly described in Turner et al.

| Station name | Latitude | Longitude | Elevation (m) | Country |
|--------------|----------|-----------|---------------|---------|
| Bellinghausen | −62.183333° | −58.833333° | 16 | Russia |
| Esperanza     | −63.4°    | −56.983333° | 13 | Argentina |
| Great_Wall    | −62.216667° | −58.966667° | 10 | China |
| Jubyany       | −62.233333° | −58.633333° | 4 | Argentina |
| Marambio      | −64.233333° | −56.716667° | 198 | Argentina |
| Rothera       | −67.57002°  | −68.12361° | 32 | UK |
| San_Martin    | −68.116667° | −67.133333° | 4 | Argentina |
| Vernadsky     | −65.25°    | −64.266667° | 11 | UK |
The daily SAT is obtained by averaging the four values observed at six-hour intervals (0000, 0600, 1200, and 1800 UTC).

The variability of the Peninsula-wide SAT can be measured by the AP SAT index, which is defined as the area-mean ERA-Interim SAT averaged over the AP region (72.5° S–62.5° S, 75° W–55° W). As shown in Fig. 1a, the daily AP SAT index can represent well the variability of Peninsula-wide SAT during austral summer (Fig. 1a). The daily AP SAT index is also significantly correlated with the observed SAT at each station (Fig. 1b). The correlation coefficients range from 0.58 at the Vernadsky Station to 0.74 at the Rothera Station. The explained variance of 34%–72% reflects that processes driving the Peninsula-wide SAT variations are not always the dominant contributors to the SAT variabilities at the stations of AP. In fact, the temperature variabilities at AP are also influenced by the orographic and local factors such as the Foehn effect (Turner et al. 2021). Extreme warm events are identified when the normalized AP SAT exceeds 1 standard deviation and maintains for more than 3 days. Similarly, the extreme cold events are determined when the normalized AP SAT exceeds –1 standard deviation. According to such definition, 44 warm events and 44 cold events were identified during the period 1979–2018. Note that the results do not change when selecting a slightly longer duration, such as more than 5 days. The day when the normalized AP SAT reaches the maximum is called day 0. The n\textsuperscript{th} day before the day (i.e., day 0) of the maximum AP SAT is referred to as day –n. Similarly, the n\textsuperscript{th} day after day 0 is called day + n. Composite analyses were performed to obtain the common features of the extreme events. Before composite analyses, the daily climatological-mean annual cycle was removed from the raw data. The two-tailed Student’s t-test was used to evaluate the statistical significance of composite analyses.

Several previous studies have reported that temperature anomalies during extreme warm/cold events are mainly contributed by horizontal advection (Song and Wu 2017; Song et al. 2018; Yang and Li 2016). The horizontal temperature advection term at the surface was calculated using the method of Lee et al. (2011). To evaluate the contribution of atmospheric flow on different time scales in driving the temperature anomalies, the advection term is decomposed using the method of Song et al. (2018). Variable $A$ can be decomposed into three parts as follows:

![Fig. 1](image-url) Spatial distribution of the correlation of daily SAT anomalies in (a) ERA-interim reanalysis and (b) station on AP with the daily AP SAT based on ERA-interim reanalysis during austral summer. The rectangle in (a) indicates the areas used to calculate the AP SAT. The station names used in the text are also shown in (b).
\[ A = A_h + A_b + A_t, \]  
(1)

where the subscripts \( h \), \( b \), and \( t \) denote the synoptic, intraseasonal, and low-frequency components, respectively. In this study, we extracted synoptic, intraseasonal and low-frequency time-scale components from the original fields by using the 2–8-day bandpass-filter, 8–90-day bandpass-filter and 90-day low-pass filter. Here, the weights of the filters are 17 for synoptic and 181 for intraseasonal and low-frequency time scales. Then, the advection terms can be decomposed into three terms:

\[
\vec{V} \cdot \nabla T = (\vec{V} + \vec{V}_b + \vec{V}_h) \cdot \nabla (T + T_b + T_h)
\]

\[
= u_b \frac{\partial T_b}{\partial x} + u_b \frac{\partial T_i}{\partial x} + u_i \frac{\partial T_b}{\partial y} + v_i \frac{\partial T_b}{\partial y} + \frac{\partial T_i}{\partial y} + v_i \frac{\partial T_i}{\partial y}
\]

\[
+ u_b \frac{\partial T_h}{\partial x} + u_b \frac{\partial T_i}{\partial x} + u_i \frac{\partial T_h}{\partial y} + v_i \frac{\partial T_h}{\partial y} + \frac{\partial T_i}{\partial y} + v_i \frac{\partial T_i}{\partial y}
\]

\[
+ u_b \frac{\partial T_h}{\partial y} + u_i \frac{\partial T_h}{\partial y} + \frac{\partial T_b}{\partial y} + v_i \frac{\partial T_b}{\partial y}
\]

\[ \text{synoptic mixture} \]

(2)

The above three terms represent the advection terms associated with the intraseasonal oscillations (ISOs), synoptic variations, and the mixture contribution of the ISOs and synoptic variations, respectively.

To investigate the interactions between the mean flow and the ISO perturbation, the barotropic energy conversion (\( CK \)) and baroclinic energy conversion (\( CP \)) with the climatological-mean flow (Cai et al. 2007) are applied:

\[
CK = -u' \frac{\partial \overline{T}}{\partial x} - v' \frac{\partial \overline{T}}{\partial y} - u' v' \frac{\partial \overline{T}}{\partial y} - u' v' \frac{\partial \overline{T}}{\partial x}
\]

(3)

\[
CP = -1 \gamma^{-1} \left( \frac{T'}{\partial x} + v' \frac{T}{\partial y} \right)
\]

(4)

where the single prime denotes the intraseasonal perturbation, and the overbar denotes the summer time mean. \( \gamma \) is the stability factor: \( \gamma = \frac{p}{R} \left( \frac{C_p}{\gamma} \right) \). \( R \) is the gas constant of dry air, \( C_p \) is the specific heat of air at constant pressure, and the hat operator denotes horizontal averaging over the Southern Hemisphere. Following Nakamura and Wallace (1990) and Nakamura et al. (1997), the storm track activity, which is defined as root mean square of the 2–8 day bandpass filtered 300-hPa geopotential height, is calculated to represent the amplitude of the high-frequency transient eddies. As a diagnostic tool, the maximum Eady growth rate (Hoskins and Valdes 1990) is applied to investigate the change in the synoptic-scale disturbance:

\[
\sigma = 0.31 \frac{f}{N} \frac{| \partial V |}{| \partial z |},
\]

(5)

where \( f \) is the Coriolis parameter, \( N \) is the static stability, \( z \) is the vertical coordinate, and \( V \) is the horizontal wind vector. The Eady growth rate was calculated using daily data over the 850–300-hPa layer to represent the tropospheric baroclinicity. The method of wave activity fluxes (Takaya and Nakamura 2001) was also used to illustrate the propagation of atmospheric Rossby waves.

3 Results

3.1 The evolution of SAT anomalies

We examine the evolution of the AP SAT anomalies of compositied warm and cold events during the period from day −10 to day +10. For both warm and cold events, the magnitudes of the raw SAT anomalies are relatively small at the beginning (from day −10 to day −6) and ending (day +6 to day +10) periods of the temperature events (black line in Fig. 2). For the warm events, the SAT anomalies increase rapidly after day −6 and reach the peak on day 0, and then decrease rapidly from day 0 to day +6 (Fig. 2a). For cold events, the time series of the AP SAT anomalies shows a similar evolution but with opposite sign (Fig. 2b). It can also be detected in Fig. 2 that different timescale variations modulate the AP SAT anomalies during different periods of the evolution of the warm and cold events. The low-frequency temperature anomaly during the whole lifetime of the extreme temperature events maintains a relatively stable value, which is the dominant contributor to the raw AP temperature anomaly from day −10 to day −4 and from day +4 to day +10 (green line in Fig. 2). During the period from day −3 to day +3, the intraseasonal temperature anomaly accounts for the majority of the raw AP temperature anomaly (blue line in Fig. 2), indicating the importance of the ISOs to the occurrence and strength of extreme temperature events. On day 0, the synoptic temperature anomaly also partly contributes to the raw AP temperature anomaly (red line in Fig. 2). This indicates the important role of the synoptic temperature anomaly in affecting the peak time of the temperature anomaly.

The spatiotemporal evolution of SAT anomalies for the warm and cold events is presented in Fig. 3. For warm events, a weak but significant warm anomaly can be observed around the Drake Passage on day −4 (Fig. 3a). After that, the warming anomalies clearly expand southward and become more significant from day −4 to day 0 (Fig. 3b–c). On day 0, the warm anomalies reach the peak, with the maximum over the AP region (Fig. 3c). On day 2, the warm anomalies gradually weaken (Fig. 3d). For cold
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events, the spatiotemporal evolution of SAT anomalies shows a similar but opposite-sign feature (Fig. 3e–h). The cold anomalies develop and expand southward to the AP region from day –4 to day –2 and reach a minimum of SAT anomalies over the AP region on day 0, and then begin to recover on day 2 (Fig. 3e–h).

Figure 4 displays the spatiotemporal evolution of SAT anomalies of warm events contributed by intraseasonal, synoptic, and low-frequency variations. The spatial pattern of the ISO-related temperature anomalies (Fig. 4a–d) is very similar to the original one (Fig. 3a–d) from day –4 to day +2, manifesting the most important contribution to the evolution of the original SAT anomalies around the AP. For synoptic variations, the SAT around the AP shows significant cold anomalies on day –2 (Fig. 4f), but is rapidly replaced with significant warm anomalies on day 0 (Fig. 4g). This indicates that the contribution of synoptic variations quickly fluctuates and partly leads to the maximum of warm events. The low-frequency variations can maintain a stable warm anomaly around the Drake Passage and AP regions (Fig. 4i–l). By comparing these different timescale variations, it is found that the ISOs play the most important role in governing the spatiotemporal evolution of the AP SAT anomalies during the period from day –4 to day +2. For cold events, the spatial pattern of the ISO-induced temperature anomalies (Fig. 5a–d) is also similar to that of the original one during the lifetime of the cold events (Fig. 3e–h). This result implies that the ISOs also dominantly regulate the spatiotemporal evolution of the AP SAT anomalies of cold events.

3.2 Diagnostic analysis of SAT anomalies

In this subsection, we assess the contributions of the horizontal advection term using the ERA-Interim data, which helps to explore the influence of the anomalous atmospheric flow on the SAT anomalies over the AP. Figure 6a, c present the temporal evolution of the composite anomaly of temperature tendency (red curve) and horizontal advection (black curve) averaged over the AP region (72.5°S–62.5°S, 75°W–55°W), respectively, for the warm and cold events. Overall, for the warm events (Fig. 6a), the tendency of the observed air temperature anomaly is positive from lag –6 to lag 0 days, which corresponds well to the increase in AP SAT. The tendency anomaly increases since lag –6 day and reaches the largest positive value on lag –1 day (Fig. 6a). After day 0, the regionally averaged temperature tendency becomes negative (Fig. 6a), which corresponds to the decay of the warm anomalies over the AP region. For the cold events, the composite anomaly of temperature tendency shows a similar evolution but with opposite sign (Fig. 6c). The observed tendency is generally negative before lag 0 day (Fig. 6c), corresponding to the enhancement of the cold anomalies over the AP region. After the peak day (day 0), the observed tendency of temperature turns positive (Fig. 6c), indicating the weakening of the cold anomalies. For both warm and cold events, the horizontal advection term facilitates the evolution of the temperature anomalies during the period from day –6 to day +2 (Fig. 6a, c). At the developing stage of the temperature events (lag –6 to lag 0 days), the magnitudes of the horizontal advection are overall larger than those of the observed tendencies (Fig. 6a, c), implying the important contribution of horizontal advection term to the development and maintenance of temperature anomalies. As shown in Fig. 6b, the advection term associated with the ISOs (blue curve) shows a process of development, maturation, and decay, higher than those associated with synoptic variations (red curve) and the mixture effect of the ISOs and synoptic variation (orange curve) during the period day –6 to day +4. The higher ISO-related advection term
suggests the importance of the ISOs in sustaining the duration of the warm events. Around day 0, the advection term associated with synoptic variations reaches a secondary high level (Fig. 6b), also making a non-negligible contribution to the occurrence of the maximum of warm events. Similarly, the ISOs also play an important role in maintaining the duration of cold events (Fig. 6d). The advection term associated with synoptic variations can contribute to and determine, to some extent, the peak time of cold events (Fig. 6d). During the lifetimes of both warm and cold events, the temperature tendencies induced by the mixed terms are small (Fig. 6b, d), indicating their weaker contributions to the evolution of the AP SAT anomalies.

Figure 7 further compares the contributions of different parts of the advection terms associated with the ISOs and synoptic variations to warm and cold events on day −1. The term \(v_b \frac{\partial T}{\partial y}\) is the primary contributor to the positive anomalies of temperature tendency of the warm events (Fig. 7a). This reveals that the development of warm anomalies over the AP region is mainly induced by the advection of intra-seasonal meridional wind of low-frequency temperature gradients. The term \(v_h \frac{\partial T}{\partial y}\) is the second important contributor (Fig. 7b), implying that the synoptic meridional wind anomalies play a secondary role in increasing the temperature anomalies around the peak time. For the cold events, the
term \( \frac{\partial T}{\partial y} \) acts as the primary contributor to the maintenance of cold anomalies (Fig. 7c). Also, the terms \( \frac{\partial T}{\partial y} \) makes a non-negligible contribution to the cooling tendency (Fig. 7d). Generally, intraseasonal and synoptic wind anomalies are vital to the formation of temperature anomalies around the peak time of warm and cold events.

### 3.3 Intraseasonal atmospheric circulation anomalies

The above analyses indicate the important role of intraseasonal wind anomalies in inducing temperature tendency anomalies. We further assess the intraseasonal evolution of atmospheric circulation anomalies in detail. Figure 8 shows the spatiotemporal evolution of intraseasonal components of SLP and surface wind anomalies for warm and cold events, respectively. From day \(-4\) to day \(0\) of the warm events, a dipole pattern with the anomalous cyclone over the Amundsen/Bellingshausen Sea (ABS) and the anomalous anticyclone to the east of the Drake Passage can be observed over the Drake Passage (Fig. 8a–c). The dipole pattern is favourable for the development of the warm anomalies around the AP. From day \(-4\) to day \(0\), the dipole pattern over the Drake Passage develops in its magnitude (Fig. 8a–c), corresponding well to the enhancement of the warm anomalies over the AP region. Subsequently, the dipole pattern over the Drake Passage weakens in its magnitude (Fig. 8d). In general, the development of the cold events is also accompanied with a dipole pattern with the opposite sign (Fig. 8e–h). Under the influence of the dipole pattern, anomalous southeasterlies prevail around the Drake Passage, which reduces the SATs over the AP through inducing cold advection. The pattern of the atmospheric circulation anomalies can sustain from...
day − 4 to day 0 (Fig. 8e–g), which is conducive to the maintenance of cooling over the AP region.

Figures 9 and 10 display the spatiotemporal evolution of intraseasonal components of 300-hPa geopotential height anomalies (shading) and the attendant wave activity flux (vector) for warm and cold events, respectively. Here the 300-hPa summer mean zonal wind (green contours) represents the jet stream. Generally, the precursors of the extreme temperature events can be traced back to an upstream wave packet propagating along the jet stream to the south of Australia (Figs. 9, 10). On day − 8 of warm events, significant negative height anomalies can be observed at the exit of the jet to the south of Australia (Fig. 9a). Due to the continuous eastward propagation of wave activity fluxes, positive height anomalies are stimulated over the southeast of New Zealand on day − 6 (Fig. 9b). From day − 4 to day 0, the cyclonic anomalies over the ABS and anticyclonic anomalies over the east of the Drake Passage forms and intensifies due to the continuous eastward propagation of wave activity fluxes (Fig. 9c–e). Subsequently, due to the disappearance of the upstream wave packet, the dipole pattern over the Drake Passage weakens in its magnitude (Fig. 9f). Also, the intraseasonal Rossby wave train along the polar jet over the Southern Ocean precedes the occurrence of cold events (Fig. 10). On day − 8 of the cold events, significant increase in the geopotential height occurs at the exit of the jet to the south of Australia, accompanied with upstream wave packet (Fig. 10a). Meanwhile, the downstream height anomalies centered over the southeast of New Zealand, ABS, and east of the Drake Passage start to develop with continuous eastward propagation of wave activity fluxes (Fig. 10a). On day − 4, the dipole pattern with opposite-sign height anomalies forms over the Drake Passage (Fig. 10c). Under the influence of the dipole pattern, anomalous southeasterlies prevail around the Drake Passage, which reduces the SATs over the AP through inducing cold advection. The dipole pattern over the Drake

Fig. 5  The same as Fig. 4 but for the cold events
Passage can sustain from day − 4 to day 0 (Fig. 10c–e), which is conducive to the maintenance of cooling over the AP region.

The above analysis indicated that the upstream wave packet along the jet over the Southern Ocean plays an important role in triggering ISO disturbances over the
Fig. 8  Composite anomalies of intraseasonal component of SLP (shading; hPa) and surface winds (vectors; ms$^{-1}$) on days a + 4, b + 2, c 0 and d + 2 of the warm events. e–h are the same as a–d but for cold events. The dots indicate anomalies significant at the 95% confidence level. Only the surface winds anomalies that exceed 95% significance level are drawn.

Fig. 9  Composite anomalies of intraseasonal component of 300-hPa geopotential height (Z300; shading; m) and the attendant wave activity flux (vector; m$^2$s$^{-2}$) on days a + 8, b + 6, c + 4, d + 2, e 0 and f + 2 of the warm events. The climatological summer mean zonal wind at 300 hPa is indicated with green contours with interval 8 (16, 24, …; ms$^{-1}$).
Drake Passage. Once triggered, these ISO disturbances can be maintained and developed through the wave–flow interaction processes. To illustrate the mechanisms for the maintenance of the ISO disturbances, we further examine the spatiotemporal evolution of the barotropic and baroclinic energy conversions with the climatological mean state. For warm events, the ISO disturbances gain net baroclinic energy from the local mean flow over the Drake Passage from day $-2$ to day 0 (Fig. 11e–g). This process is conducive to the amplification of the height anomalies around the Drake Passage. In contrast, a significant increase in the barotropic energy conversion only occurs over the AP region on day 0 (Fig. 11c). Furthermore, the net conversion contributed by the barotropic process is trivial compared to the CP conversion (Fig. 11). From day $-2$ to day 0 of the cold events, the ISO disturbance over the Drake Passage can also extract energies from the mean flow through the baroclinic process (Fig. 12e–h). Furthermore, significant positive CP anomalies can also be observed between 150°W to 120°W from day $-4$ to day $-2$ of the cold events (Fig. 12e, f), which is conducive to the development and maintenance of the height anomalies over the southeast of New Zealand and ABS.

### 3.4 Synoptic atmospheric circulation anomalies

The spatiotemporal evolutions of synoptic atmospheric circulation anomalies for warm and cold events are shown in Figs. 13 and 14, respectively. A synoptic Rossby wave train can be identified from day $-2$ to day 0 (Figs. 13 and 14). However, unlike intraseasonal Rossby wave, the synoptic wave train has a smaller spatial scale and shows a baroclinic structure, with the phase tilting slightly westward with the increase in altitude (Figs. 13 and 14). Furthermore, synoptic atmospheric circulation anomalies systematically propagate eastward during the period from day $-2$ to day 0 (Figs. 13 and 14). The synoptic variable quickly fluctuates between positive and negative values when the synoptic Rossby wave passes by. For example, the synoptic height anomalies to the east of AP switch quickly from negative values on day $-2$ to positive values on day 0 of the warm events (Fig. 13a, c). The fast propagation of the synoptic atmospheric circulation anomalies also leads to the rapid switch of synoptic SAT anomalies over the AP (Figs. 4e–h and 5e–h).

It should be noted that the magnitude of the synoptic circulation anomalies before the cold events is not as strong as that before warm events (Fig. 13 vs Fig. 14). The instantaneous amplitude of synoptic Rossby wave at a given location is usually depicted by the storm track activity (Nakamura and Wallace 1990; Nakamura et al. 1997). The discrepancy between the synoptic eddy activities before warm events and those before cold events can also be reflected by the storm track activity. As shown Fig. 15a, significant enhancement of the storm track activity anomalies can be observed over the west of Chile and the region from AP to Weddell Sea before warm events (Fig. 15a). The enhanced storm track activity corresponds to the development of a synoptic wave train. For cold events, the storm track activities are suppressed over the Drake Passage (Fig. 15b), which is consistent with the loosely-organized and weakened synoptic Rossby wave train before the cold events. By comparing the composite storm track activities for the warm events and that for the cold events (green contours in Fig. 15a, b), we find that more high-frequency transient eddies pass through the Antarctic Peninsula before the warm events than those
before cold events. The source of energy for the storm track activity is baroclinic generation (Cai and Mak 1990; Hoskins and Valdes 1990; Shaw et al. 2016; Chang et al. 2002). Therefore, the composite anomalies of background baroclinicity for the warm and cold events, which is measured by the maximum Eady growth rate, are examined (Fig. 15c, d). Figure 15c shows a clear increase in the tropospheric baroclinicity over the west of Chile and the region from AP to Weddell Sea before warm events, which is broadly consistent with the anomalous pattern of the storm track activities. Corresponding to the weakening of the storm track activities before cold events, negative baroclinicity anomalies appear over the Drake Passage region (Fig. 15d). The results imply that the anomalous synoptic Rossby wave activities before the temperature events are closely related to the tropospheric baroclinicity. The discrepancy between the synoptic eddy activities before warm events and those before cold events may be related to the occurrence of the opposite-sign dipole pattern before the warm and cold events.

4 Summary and discussion

Using the ERA-interim SATs, this study identified 44 austral summer warm events and 44 cold events over the AP during the period from 1979/80 to 2018/19. During the life cycle of warm and cold events, the development of the AP SAT anomalies can be maintained by temperature advection. Further analyses reveal that the advection terms induced by ISOs make the largest contribution to the maintenance of temperature extremes. The advection terms associated with synoptic variations also make non-negligible contributions to the formation and development of temperature anomalies around the peak time of temperature events. The advection term associated with ISOs on day –1 is mainly induced by the intraseasonal winds. Similarly, the advection term associated with synoptic variations is mainly induced by the synoptic winds.

The extreme temperature events over the AP region are preceded by the intraseasonal Rossby wave packet along the polar jet over the Southern Ocean. The intraseasonal Rossby
wave packet before the warm events leads to the formation and development of cyclonic anomalies over the ABS and anticyclonic anomalies to the east of the Drake Passage. The associated intraseasonal northerly wind anomalies over the Drake Passage can induce persisting warm anomalies over AP region. The occurrence of the cold events is preceded by the ISO Rossby wave train with the opposite sign. Influenced by the intraseasonal Rossby wave, the dipole pattern with anticyclonic anomalies over the ABS and cyclonic anomalies to the east of the Drake Passage maintains and develops, which is vital to the formation of persisting cold anomalies over AP region. The wave–flow interaction also plays an important role in maintaining circulation anomalies around the AP. Specifically, the anomalous efficient conversion from available potential energy of the mean flow contributes to the energy supply of the ISO disturbances triggered by continuous Rossby wave energy dispersion.

Concomitant with the rapid eastward propagation of Rossby wave, the anomalous synoptic circulations rapidly change and facilitate the development of the extreme temperature events. However, the synoptic Rossby wave activity before cold events is much weaker than that before warm events. This difference can be partly attributed to the anomalous tropospheric baroclinicity. Before the warm events, the baroclinicity increases to the west of Chile and extends from the AP to the Weddell Sea, contributing to the enhancement of the synoptic Rossby wave activity. However, the baroclinicity decreases over the Drake Passage before cold events, which suppresses the development of the synoptic Rossby wave activity.

The atmospheric flows on intraseasonal and synoptic time scales may influence each other. On one hand, the amplitude of the synoptic eddies is modulated by the flow with longer timescales (Wallace et al. 1988; Nakamura and Wallace 1990, 1993; Nakamura et al. 1997; Chang and Yu 1999; Wang et al. 2020). The intraseasonal atmospheric flow may affect the tropospheric baroclinicity and consequently modulate the intensity of the synoptic wave activity. On the other hand, the self-interaction among transient eddies also contributes to the development and maintenance of the extreme temperatures.
Fig. 13 Composite anomalies of synoptic component of geopotential height at 300 hPa (shading; m) and the attendant wave activity flux (vector; m² s⁻²) on days a – 2, b – 1 and c 0 of the warm events. d–f are the same as a–d but for SLP (shading; hPa) and surface winds (vectors; ms⁻¹). In d–f, only the surface winds anomalies that exceed 95% significance level are drawn. The dots indicate anomalies significant at the 95% confidence level.

Fig. 14 As in Fig. 13, but for the cold events.
of intraseasonal Rossby wave trains (Feldstein 2002; Wang et al. 2020). It should be noted that the proposed mechanisms may only work in the present climate background. According to some model studies, the Southern Hemisphere jet streams and storm tracks shift poleward and strengthen in response to global warming (Yin 2005; Chang et al. 2012; Barnes and Polvani 2013; Simpson et al. 2014). These changes may affect the propagation and intensity of the Rossby wave on intraseasonal and synoptic time scales and alter the wave–flow interaction and the self-interaction among transient eddies. As a result, the mechanisms of the Peninsula-wide temperature extremes may change in the future. It is worthy of further study how the mechanisms change in a warmer climate background.

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Fig. 15 Composite anomalies of 8-day low-pass-filtered storm track activity anomalies (shading; m) averaged from days −2 to 0 of the a warm events and b cold events. c–f are the same as a–c, but for anomalies of Eady growth rate (shading; day⁻¹). The green contours in a and b indicate the composite storm track activity (m) averaged from days −2 to 0 of the warm events and cold events, respectively. The dots indicate anomalies significant at the 95% confidence level.

Declarations

Conflict of interest The authors declare no potential conflicts of interest.

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