Impacts of ENSO and Indian Ocean Dipole Events on the Southern Hemisphere Storm-Track Activity during Austral Winter

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ABSTRACT

Impacts of the ENSO and Indian Ocean dipole (IOD) phenomena on winter storm-track activity over the Southern Hemisphere are examined on the basis of the observed and reanalysis data for 1979–2003. The partial correlation technique is utilized to distinguish the impact of one phenomenon from that of the other. During an El Niño event, the subtropical jet stream tends to strengthen substantially, enhancing the jet bifurcation and thereby reducing storm-track activity over the midlatitude South Pacific and to the south of Australia. During a positive IOD event, the westerlies and storm-track activity also tend to weaken over southern Australia and portions of New Zealand. Thus both the positive IOD and, to a lesser extent, El Niño events act to reduce winter rainfall significantly over some portions of South Australia and New Zealand. Precipitation over the southeastern portion of the continent and over the northern portions of the two main islands of New Zealand is more sensitive to IOD. Significant reduction in precipitation associated with an El Niño event is seen over Tasmania. Over midlatitude South America, in contrast, the enhancement of the westerlies and storm-track activity tends to be more significant in a positive IOD event than in an El Niño event. It is demonstrated that despite the dominant influence of the Southern Hemispheric Annular Mode from a hemispheric viewpoint, the remote influence of ENSO and/or IOD on local storm-track activity can be detected in winter as a significant signal in particular midlatitude regions, including South Australia and New Zealand.

1. Introduction

Synoptic-scale baroclinic eddies that propagate along midlatitude storm tracks form an important constituent of the climate system, as they act to maintain the extratropical general circulation by meridionally transporting zonal angular momentum, sensible heat, and moisture. In this process, they also affect the local weather conditions in the extratropics through precipitation and temperature variations. Trenberth (1991) comprehensively studied the seasonal characteristics of the storm-track activity in the Southern Hemisphere (SH), based mainly on zonally averaged statistics of subweekly eddies and their relationship with the zonal mean circulation. Several other studies have examined the group velocity propagation of synoptic-scale eddies (Lee and Held 1993; Berbery and Vera 1996; Chang 1999; Rao et al. 2002). More synoptic and regional aspects of storm activity have been provided by Sinclair (1994, 1995, 1996) and Simmonds and Keay (2000), who tracked the centers of individual moving cyclones and anticyclones over the SH. More detailed synoptic aspects of the SH storm tracks have been shown through a sophisticated tracking analysis by Hoskins and Hodges (2005).

Although the seasonal dependency of the zonally averaged storm-track activity is weaker when compared to that of the zonal mean westerlies (Trenberth 1991), Nakamura and Shimpo (2004) found a marked seasonality in storm-track activity over the South Pacific depending on the intensity of the subtropical jet (STJ). In the absence of the developed STJ during austral summer, a single, well-defined circumpolar storm track
forms along a deep polar-front (or subpolar) jet stream (PFJ). During austral winter, in contrast, the main upper-level storm track over the South Pacific forms along the intense STJ, while a low-level storm track with vigorous baroclinic eddy growth forms along the surface baroclinic zone off the Antarctic coast. Trapping upper-level eddy activity into its core away from a surface baroclinic zone (Nakamura and Sampe 2002), the intense STJ over the South Pacific acts to suppress midlatitude storm-track activity (Nakamura and Shimpo 2004; Nakamura et al. 2004). Another important factor that influences the SH storm-track activity is the presence of a pronounced surface baroclinic zone over the south Indian Ocean, anchored by an intense oceanic frontal zone along the Antarctic Circumpolar Current (Nakamura and Shimpo 2004; Nakamura et al. 2004; Inatsu and Hoskins 2004). The core regions of the SH storm track and PFJ are anchored around the baroclinic zone throughout the year regardless of the intensity of the STJ (Nakamura and Shimpo 2004).

ENSO events can influence the SH storm-track activity by changing the strengths and positions of the STJ and/or the PFJ in response to anomalous convective activity in the Tropics (Trenberth 1998). The anomalous convection can influence the STJ via an anomalous divergent wind (Sardeshmukh and Hoskins 1985), whereas the PFJ tends to be influenced through stationary Rossby waves generated in response to anomalous tropical convection (Kidson et al. 2002). During austral summer, storm-track activity over the midlatitude South Pacific tends to be enhanced in an El Niño event and suppressed in a La Niña event in association with the intensification of the PFJ and STJ, respectively (Bhaskaran and Mullan 2003). Nakamura et al. (2004) found that the 1997 El Niño and 1998 La Niña events exerted opposing impacts on the winter storm-track activity over the South Pacific, associated with the marked intensification and weakening, respectively, of the STJ (cf. Bals-Elsenholtz et al. 2001) and PFJ. As ENSO is a very strong tropical phenomenon that exerts a vast impact on the climates around the world (e.g., Rasmusson and Wallace 1983; Ropelewski and Halpert 1987; Trenberth et al. 1998; Saji and Yamagata 2003), we further explore the impact of ENSO on the storm-track activity over the entire SH, including its influence on midlatitude precipitation.

The Indian Ocean dipole (IOD: Saji et al. 1999; Webster et al. 1999; see Yamagata et al. 2004 for further details and references) is another important tropical coupled phenomenon that influences the climates in many geographical regions around the world (Ashok et al. 2001, 2004b; Guan and Yamagata 2003; Saji and Yamagata, 2003; Behera et al. 1999; Rao et al. 2005; see Yamagata et al. 2004 for further details). In particular, Ashok et al. (2003a) showed that a positive IOD event tends to yield significant deficits in rainfall over the western and southern regions of Australia. Saji et al. (2005) have demonstrated a significant impact of IOD on surface air temperature in many regions over the SH, including Australia. Keeping in mind the significant impact IOD can exert on the Australian climate, we examine how significantly the IOD can influence the winter storm-track activity around the Australian continent.

In this study, we compare the impacts of ENSO and IOD on the SH storm-track activity and associated precipitation in austral winter (June–October). In section 2, we present details of the datasets used in this study, along with a description of our methodology. We discuss results of our analysis in section 3. In the final section, we present a brief summary of the work, along with conclusions and some discussion.

2. Data and methodology

In the present study, the National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) global reanalysis data with 6-hourly temporal resolution (Kalnay et al. 1996) are used for deriving systematic storm-track statistics. We limit our analysis to the period 1979–2003 for which the data quality is substantially higher, as compared to the period before 1979, due to the availability of satellite measurements. At each grid point, subweekly fluctuations associated with synoptic-scale transient eddies have been extracted from the 6-hourly data time series by means of a digital high-pass filter with a half-power cutoff period of 8 days. As argued by Nakamura and Shimpo (2004), eddy statistics based on the high-pass filtered quantities are unlikely affected by geographically fixed surface bogus data misplaced in the reanalysis.

Local instantaneous upper-level storm-track activity is measured by what may be called the “envelope function” \( Z_e \) (Nakamura and Wallace 1990; Nakamura and Shimpo 2004). The quantity is defined locally as

\[
Z_e = \left( \frac{1}{2Z'_{500}} \right) \left( \frac{\sin(45^\circ S)}{\sin(\text{lat})} \right),
\]

where \( Z' \) denotes the 8-day high-pass filtered 300-hPa height and the overbar the smoothing with an 8-day low-pass filter. The quantity \( Z_e \) thus represents the local instantaneous amplitude of 300-hPa height fluctuations with periods shorter than 8 days in terms of a geostrophic streamfunction.

As an index of low-level storm-track activity, a pole-
ward heat flux ($\mathbf{v}$) associated with subweekly disturbances was obtained as a product of the high-pass filtered time series of the meridional velocity and temperature at the 850-hPa level. This flux has been subject to the 8-day low-pass filtering to represent systematic transport of heat by the transient disturbances. In addition, barotropic feedback forcing by synoptic-scale eddies migrating along a storm track has been evaluated as a local tendency in 250-hPa height ($dZ_{250}$) that would be induced due solely to 250-hPa anomalous vorticity flux convergence associated with those eddies (Nakamura et al. 1997). The vorticity flux was calculated from the high-pass filtered velocity fields before smoothing with the low-pass filter. Compensated by the corresponding tendency due to eddy heat transport at lower levels, the barotropic feedback ($dZ_{250}$) should be regarded as an upper bound of the net feedback forcing by those eddies on quasi-stationary circulation (Lau and Nath 1991). In the following, we focus primarily on the eddy feedback forcing in terms of the equivalent westerly acceleration at the 250-hPa level ($dU_{250}$), as derived geostrophically from $dZ_{250}$. Since we are interested in the remote influence of the tropical interannual variability on the SH storm tracks, monthly mean fields of SH circulation including those for eddy statistics of $Z_e$, $\mathbf{v}$, and $dU_{250}$ are used. The anomaly fields are defined as monthly departures from their corresponding climatological-mean fields for a given calendar month for the period of 1979–2003.

Our indicators of ENSO and IOD phenomena are the Niño-3 sea surface temperature anomaly\(^1\) (SSTA) and the Indian Ocean dipole mode index\(^2\) (IODMI: Saji et al. 1999), respectively. These monthly indices have been derived from the Hadley Centre Global Sea Ice and Sea Surface Temperature (HadISST) analyses datasets (Rayner et al. 2003) for the period of 1979–2003. Our precipitation analysis has been carried out using a monthly gridded dataset ($2.5^\circ \times 2.5^\circ$) from the Climate Prediction Center (CPC) Merged Analysis of Precipitation (CMAP: Xie and Arkin 1996), which is based mainly on rain gauge measurements over land and satellite measurements over the ocean. A monthly index (downloaded from http://www.cpc.ncep.noaa.gov/products/precip/CWlink/all_index.html) for the SH Annular Mode (SAM: e.g., Thompson and Wallace 2000), also known as the Antarctic Oscillation, is also used. As well known, SAM is the dominant mode of variability in the extratropical SH on monthly and interannual time scales (Thompson and Wallace 2000) in which interactions between the anomalous midlatitude westerlies and transient eddies along the storm track are involved (Lorenz and Hartmann 2001).

Each of the monthly indices and anomaly fields of the atmospheric circulation and precipitation for a particular year has been averaged over a period from June to October, henceforth referred to as (austral) winter. This period includes early austral spring to capture robust influence of IOD events that tend to peak in November. Note that the climatological characteristics of storm tracks over the SH Indo–Pacific sector remain qualitatively the same throughout the period (Nakamura and Shimpo 2004). The partial correlation technique (e.g., Spiegel 1988) is employed to distinguish the impacts of the individual tropical phenomena (see the appendix for details).

3. Results

a. Climatological features and interannual variability

Figure 1 shows the climatological distributions of 300-hPa $Z_e$, 850-hPa poleward eddy heat flux ($\mathbf{v}$), and zonal wind speeds at the corresponding levels for austral winter. At the 300-hPa level, a STJ is pronounced over the eastern Indian Ocean and western Pacific (Fig. 1a), while a PFJ at this level is strongest over the Atlantic and Indian Oceans. The cores of the two jet streams are clearly separated at this level during this season. The overall distribution is qualitatively similar to those at the 200- and 250-hPa levels (figures not shown). The PFJ is very deep over the Atlantic and the Indian Oceans, and it can be recognized as the most intense westerly jet at the 850-hPa level (Fig. 1b). In contrast, the 850-hPa westerlies almost diminish below the STJ core, covered by the subtropical high pressure belt (Nakamura and Shimpo 2004). The mean 300-hPa $Z_e$ for winter exhibits its primary peak along the PFJ over the eastern South Atlantic and south Indian Oceans (Fig. 1c). Despite its prevailing strength, the STJ accompanies only modest eddy activity in the upper troposphere, and the activity is even weaker below the jet core in the lower troposphere (Nakamura and Shimpo 2004). By contrast, the lower-tropospheric eddy activity, as represented by 850-hPa $\mathbf{v}$, is strong along the PFJ, with its core in the western South Indian Ocean (Fig. 1d). Interestingly, the eddy activity is locally enhanced over southern Australia, forming a secondary core in the lower troposphere. As pointed out by Nakamura and Shimpo (2004), the primary core of the SH storm track is collocated with the core of the

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\(^1\) Defined as the area-averaged SSTA for ($5^\circ$S–$5^\circ$N, $150^\circ$E–$90^\circ$W).

\(^2\) Defined as the SSTA difference between the western ($50^\circ$–$70^\circ$E, $10^\circ$S–$10^\circ$N) and southeastern ($90^\circ$–$110^\circ$E, $10^\circ$S–$0^\circ$) portions of the tropical Indian Ocean.
deep PFJ, which is maintained by the downward transport of mean-flow westerly momentum via eddy heat fluxes. Over the South Atlantic and Indian Oceans, the association among the midlatitude storm track, PFJ, and oceanic subarctic frontal zone has been pointed out (Nakamura et al. 2004). Over the South Pacific, this association is less robust, especially in winter during which the upper-level eddy activity tends to be trapped in the core of the seasonally enhanced STJ.

Figure 2 shows the characteristics of interannual variability in the SH jet streams and storm tracks in winter. As shown in Fig. 2a, interannual variability in the upper-tropospheric westerlies, measured as the local standard deviation of 300-hPa zonal wind ($U_{300}$), is particu-
larly strong over the midlatitude South Pacific and Atlantic Oceans. The distribution is overall consistent with the interannual variability in 200-hPa winter zonal wind obtained by Hurrell et al. (1998), based on global analyses by the European Centre for the Medium-Range Weather Forecasts, although the variability over the south Indian Ocean is slightly weaker in our result than in theirs. Likewise, the interannual variability in the upper-level storm-track activity, measured as the local standard deviation of 300-hPa $Z_e$, is also strongest over the midlatitude South Pacific and Atlantic (Fig. 2b). In the regions of large interannual variability in $U_{300}$ and $Z_e$, the local correlation is generally positive between the two variables (Fig. 2c), exceeding the 90% confidence level (from a two-tailed Student’s $t$ test). The correlation is particularly high over the South Pacific and part of the Indian Ocean, and also over southern Australia. The positive correlations indicate that, as suggested by linear theory, baroclinic eddy activity tends to be enhanced (weakened) as the local westerly wind speed increases (decreases) in those midlatitude regions. It is hypothesized from Fig. 2 that the winter climate over the midlatitude SH is likely influenced by variations in the mean westerlies and the associated changes in the storm-track activity.

b. ENSO influence

Nakamura et al. (2004) have carried out a case study to examine the impacts of the 1997 El Niño and 1998 La Niña events on the SH storm-track activity in winter. They found that in the 1997 winter the meridional bifurcation of the storm tracks over the South Pacific was even more distinct than in the climatological mean, while in the 1998 winter the bifurcation was diminished and a single circumpolar storm track formed as typically observed in summer. They also found that the bifurcation between the STJ and PFJ was also enhanced in the 1997 winter, whereas the STJ was diminished in the following winter. They hypothesized that the enhanced (diminished) storm-track activity along the STJ in the 1997 (1998) winter was due to the stronger (weaker) trapping effect of upper-level eddy activity by the intense (diminished) STJ core. Since the correlation between the Southern Oscillation index and the degree of the jet bifurcation over the South Pacific

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**Fig. 2.** The 300-hPa distribution of winter standard deviations of (a) zonal wind anomalies (m s$^{-1}$) and (b) $Z_e$ anomalies (m), and (c) local correlation between the zonal wind and $Z_e$ anomalies. Shading in (c) denotes the correlations significant at the 90% or higher confidence level.
is not particularly high (Bals-Elsholz et al. 2001), it is of interest to examine whether the hypothesis is valid, in general, for ENSO events for the period 1979–2003.

The winter distribution of the partial correlation between 300-hPa $Z_e$ and the Niño-3 index is presented in Fig. 3a from which the influence of IOD, the other predictor, has been removed (see Guan et al. 2003; Ashok et al. 2003a). This procedure is necessary because of the seasonal phase locking in the occurrence of IOD and ENSO events in some years such as during 1997 (Yamagata et al. 2003; Saji and Yamagata 2003; Ashok et al. 2003b, 2004b). The correlation between the Niño-3 index and 300-hPa $Z_e$ is generally weak in the subtropics except over the eastern Pacific (Fig. 3a). The storm-track activity is more significantly correlated with the Niño-3 index over the midlatitude Indian and Pacific Oceans and in the tip of southeast Australia and Tasmania (Fig. 3b) where the significance of the negative correlation exceeds the 90% confidence level based on the two-tailed Student’s $t$ test. In these midlatitude regions, the corresponding partial correlation of 850-hPa $\bar{v}T^*$ with the Niño-3 index is also significantly negative (Fig. 3c). Compared to the corresponding upper-level statistic (Fig. 3a), the significance of the positive correlation between the low-level storm-track activity and the Niño-3 index increases slightly over the subtropical eastern Pacific. Figure 3 indicates a tendency that the occurrence of El Niño (La Niña) leads to the reduction (strengthening) of storm-track activity in a midlatitude sector spanning from the Indian Ocean to the Pacific, whereas the opposite tendency is observed in the subtropical eastern Pacific.

The aforementioned changes in the SH storm-track activity are associated with significant intensification of the STJ and weakening of the midlatitude PFJ in the upper troposphere, as indicated by the partial correlation between $U_{300}$ and the Niño-3 index (Fig. 4a). In an El Niño winter, the STJ tends to strengthen with enhanced jet bifurcation over the South Pacific and to the south of Australia, in agreement with the findings by Chen et al. (1996) and Kiladis and Mo (1998). Over the south Indian Ocean, the enhanced jet bifurcation is also contributed to by the weakening of PFJ, but the weakening is not significant in the South Pacific except in its central portion. The impact of ENSO thus tends to be significantly stronger on the STJ than on PFJ, in agreement with Bals-Elsholz et al. (2001). In the lower troposphere (Fig. 4b), ENSO exerts a similar but weaker impact on the midlatitude PFJ, as indicated by the corresponding correlation with 850-hPa zonal wind ($U_{850}$). Except over the eastern Pacific, no significant ENSO influence on $U_{850}$ is found below the STJ where the axis of a subtropical high pressure belt is located, and no significant baroclinic eddy growth tends to occur (Nakamura and Shimpo 2004). Our findings based on Figs. 3 and 4 agree overall with the hypothesis of Nakamura et al. (2004) about the ENSO impact on the winter storm-track activity over the South Pacific, while generalizing it for the Tasmanian region based on the multiple ENSO events. The tendency for eddy activity to be enhanced (suppressed) where the background westerlies are stronger (weaker) as a remote ENSO influence is in good agreement with a synoptic study by Sinclair et al. (1997). However, this tendency cannot apply to the entrance and core regions of the STJ over the Indian Ocean and southwestern Pacific, where no

![Fig. 3. Distribution of the winter partial correlation coefficients over the SH for (a) 300-hPa $Z_e$ anomalies with (b) an enlarged format around southeastern Australia, and (c) 850-hPa $\bar{v}T^*$ anomalies with Niño-3 SSTAs. Shading denotes correlations significant at the 90% or higher confidence level. The box in (b) is the domain for the ENSO-sensitive index, as a measure of the ENSO-related anomalous activity of the local storm track (see text for its definition).](image-url)
significant ENSO impact is found in the storm-track activity (Fig. 3).

To verify our finding that changes in storm-track activity during an ENSO event are due largely to the anomalous intensity of the westerlies, the meridional SST gradient for the winter of 1987 in an El Niño event and that for the winter of 1988 in a strong La Niña event have been compared (figures not shown). The comparison reveals that the intensity and location of the oceanic subarctic frontal zone [Antarctic Polar Frontal Zone (APFZ)], which acts to anchor the main atmospheric baroclinic zone (Nakamura and Shimpo 2004), exhibits no significant changes in the course of the 1987–88 ENSO cycle. This result is consistent with the findings by Nakamura et al. (2004) for the 1997–98 ENSO cycle. It is therefore unlikely that the profound changes in the South Pacific storm-track activity observed in these ENSO cycles are due mainly to changes in the midlatitude near-surface baroclinicity associated with the APFZ.

El Niño events tend to be associated with winter and spring droughts over the eastern third of the Australian continent (McBride and Nicholls 1983; Ropelewski and Halpert 1987; Nicholls 1989), and southern Australia and Tasmania (Ropelewski and Halpert 1987). From this perspective, the significant ENSO influence on storm-track activity observed over the eastern and southern portions of the Australian continent (Fig. 3b) bears certain implications. The southern and southeastern portions of the continent coincide with a local maximum in eddy activity (Fig. 1d), and the activity tends to be suppressed in an El Niño winter, as the subpolar storm track tends to stay to the south, away from the continent. In contrast, the local storm activity tends to be enhanced in a La Niña winter, as the storm track tends to strengthen and stay closer to the continent.

c. IOD influence

In the following, we examine the influence of IOD events on the SH winter storm-track activity. Figure 5a shows a map of the partial correlation between IODMI and 300-hPa $Z_{500}$, from which the ENSO influence has been removed. As also shown in the enlarged format in Fig. 5b, the correlation is significantly negative over a broad region spanning from the central to southeast Australian continent. The correlation is significantly positive over the central portion of the south Indian Ocean, a midlatitude Atlantic region off South America, and, importantly, another maritime region off southeastern Australia including Tasmania (Fig. 5a). The corresponding partial correlation with 850-hPa $u^\prime T^\prime$ is distributed in a similar manner (Fig. 5c), but the positive correlation is weaker and less significant to the south of Australia.

Figure 6 presents partial correlations between IODMI and zonal winds at the 300-hPa (Fig. 6a) and 850-hPa (Fig. 6b) levels. As indicated by Saji et al. (2005), the IOD-related circulation anomalies are significant in
several regions over the midlatitude SH, especially in the upper troposphere (Fig. 6a). Over the Eastern Hemisphere, the zonal extent of the IOD-related westerly wind anomalies is less than that of the ENSO-related anomalies. Over the Western Hemisphere, in contrast, the IOD-related westerly anomalies are more significant than the ENSO-related anomalies. In the IOD-related correlations shown in Figs. 5 and 6, the tendency for the enhanced (suppressed) storm-track activity to be collocated with the strengthened (weakened) westerlies is apparent, particularly around Australia and New Zealand and around South America.

To assess the potential impact of any changes in APFZ, the SH meridional SST gradient during 1994, a positive IOD year, and for 1992, a negative IOD year, are compared (figures not shown). The comparison reveals no apparent changes in the position and intensity of APFZ between the positive and negative IOD years, indicating that the changes in the storm-track activity between these years were likely due to the remote influence of the IOD events. Hence, we propose that the IOD-induced circulation anomalies can influence the austral winter storm-track activity over the SH. During a positive (negative) IOD event, the westerlies and

Fig. 5. As in Fig. 3 but with IODMI. (b) The box is the domain for the IOD-sensitive index, as a measure of the IOD-related anomalous activity of the local storm track (see text for its definition).

Fig. 6. As in Fig. 4 but with IODMI.
storm-track activity tend to weaken (enhance) over the Australian continent and northern New Zealand, whereas they tend to enhance (weaken) to the south of the continent and over southern New Zealand.

d. Eddy feedback forcing

It can be realized from Figs. 4a and 6a that both ENSO and IOD tend to generate stationary atmospheric responses over the SH in winter, including anomalous westerlies, leading to systematic changes in midlatitude storm-track activity (Figs. 3 and 5). The anomalous storm-track activity changes eddy westerly momentum transport, which in turn acts to maintain the anomalous westerlies. This positive feedback can be represented as the anomalous westerly acceleration at the 250-hPa level ($dU_{250}$) due solely to the anomalous convergence of eddy vorticity fluxes (e.g., Nakamura et al. 1997). To confirm whether the eddy feedback forcing is, indeed, operative in each of the ENSO and IOD responses over the SH, maps of the winter partial correlation of $dU_{250}$ with the Niño-3 SSTA and IODMI are shown in Figs. 7a and 7b, respectively. The Niño-3 correlation is significantly positive in the subtropical Indian Ocean off Australia, whereas it is significantly negative over a midlatitude zone extending zonally over the south Indian Ocean, New Zealand, and South America (Fig. 7a). The negative correlation is consistent with the reduced activity of storm tracks in those midlatitude regions (Fig. 3). The IODMI correlation shown in Fig. 7b is significantly positive in midlatitude regions to the south of Australia and over the south Indian Ocean and South America, while it is significantly negative over South Australia and the subpolar Atlantic and Indian Oceans. A fairly good spatial correspondence between the eddy feedback correlations in Fig. 7 with the two tropical indices and the corresponding correlations of the upper-level westerlies (Figs. 4a and 6a) indicates the positive feedback forcing from anomalous eddy activity on the anomalous westerlies. Our finding is consistent with previous studies that have highlighted the importance of the anomalous storm-track activity in shaping the extratropical stationary response to ENSO over the winter Northern Hemisphere (e.g., Hoerling and Ting 1994).

e. Relationship between precipitation and storm-track activity

Analysis of the CMAP rainfall data (1979–2003) shows that the mean annual rainfall is less than 40 cm over central and western Australia, except for the extreme southwestern portion of the country (Fig. 8a). More precipitation is observed annually in eastern Australia, Tasmania, and New Zealand. A substantial fraction of the annual rainfall is accumulated in winter [June–October (JJASO)]; 35%–60% in southern Australia, ∼50% in Tasmania, and 40%–45% in New Zealand.

Fig. 7. Distribution of the winter partial correlation coefficients for the eddy feedback forcing in terms of the anomalous westerly acceleration at 250 hPa ($dU_{250}$) with (a) Niño-3 SSTA and (b) IODMI. Shading denotes the correlation significant at the 90% or higher confidence level.
Zealand (Figs. 8b and 8c). Hence, any impact that the IOD or ENSO could exert on SH storm tracks in winter may have important implications for the interannual rainfall variability in these regions. In this subsection, we assess qualitatively how the winter modulations in storm-track activity as a remote response to the tropical interannual variability in the Indo-Pacific sector can influence precipitation in and around Australia and New Zealand.

An index of the local storm-track activity is defined as the 300-hPa $Z_e$ anomaly in winter averaged over the “ENSO sensitive” region [38°–44°S, 144°–163°E], in which the ENSO impact on the eddy activity is strongest (Fig. 3b). An “IOD sensitive” eddy activity index is also defined as the 300-hPa $Z_e$ anomaly averaged over [23°–32°S, 133°–144°E] (based on Fig. 5b). Figure 9a shows a map of the partial correlation of CMAP precipitation with this ENSO-sensitive index for eddy activity from which the signal correlated with the aforementioned IOD impact index for austral winter has been excluded. In that figure, the sign has been reversed for the actual correlation coefficient at each grid point to suit the negative correlation shown in Figs. 3b and 5b. As the negative correlation in Fig. 3b indicates the weakening tendency for the local eddy activity over Tasmania during an El Niño winter, the significant negative correlation in Fig. 9a suggests the potential for significant deficits (surplus) in cyclone-associated precipitation over Tasmania in an El Niño (a La Niña) winter. The marginally significant correlations in part of the coastal areas of southern New Zealand may suggest a possibility of drought during a strong El Niño winter. In general, the partial correlations are in agreement with similar partial correlations between CMAP rainfall anomalies and the Niño-3 index over an area spanning from southwest Australia to New Zealand except over the southeast coast of Australia.

Figure 9b shows a map of the partial correlation (with sign reversed just as in Fig. 9a for easier interpretation) between local CMAP precipitation anomaly and the IOD-sensitive index for eddy activity defined above, superimposed on the partial correlation between the local precipitation and IODMI. In a positive (negative) IOD winter, cyclone-related precipitation tends to be reduced (enhanced) significantly over the northernmost part of New Zealand and the southeastern portion of continental Australia, a maritime region off the east coast of Australia, consistent with reduced (enhanced) eddy activity in those regions (Fig. 5a). The negative correlations are also significant over the northern regions of the two major islands of New Zealand, indicating a tendency for deficit rainfall in a strong IOD winter. Over southeastern Australia and New Zealand, the correlations of the IOD-sensitive precipitation exhibit slightly higher significance than the ENSO-sensitive precipitation. In fact, the corresponding partial correlations of sea level pressure anomalies (SLPAs) with IODMI indicate a notable tendency for stronger anticyclonic circulation during positive IOD events (Fig. 10), in agreement with the rainfall response presented in Fig. 9b. In contrast, significant partial correlations between SLPAs and the Niño-3 index are confined to the northern half of Australia (figure not shown), consistent with the fact that ENSO events

![Figure 8](image-url)
mainly affect the STJ (Fig. 4) and our finding that the relatively weaker impact of ENSO on wintertime precipitation over continental southeast Australia (Fig. 9). The relatively weak influence of IOD activity over Tasmania, despite the apparently stronger influence of the IOD on upper-level storm-track activity over the maritime region of southern Australia (Fig. 5a), may be related to the weaker IOD signal in the local eddy activity in the lower troposphere than aloft, as shown in its weaker correlations with IODMI (Fig. 5c). However, it should be noted that the partial correlations showed in Fig. 9b are robust to the definition of the IOD-related storm-track activity, as verified by repeating the analysis shown in Fig. 9b with another index based on the positive correlations with IODMI off southeastern Australia presented in Fig. 5a (figure not
shown). The only difference was that the significant correlations are observed over a wider area around New Zealand.

**f. Sensitivity to extratropical influence**

Apart from the aforementioned tropical impacts, because of its location, the SH storm-track activity is prone to be modulated by extratropical climate variability, including SAM. On the basis of their ocean–atmosphere coupled model experiment, Lau and Nath (2004) argued that IOD events may be triggered by the positive phase of SAM, while another coupled model study by Hendon et al. (2006, manuscript submitted to *J. Climate*, hereafter HZA) suggested that they may be triggered by the negative phase of SAM. A higher-resolution coupled simulation by Behera et al. (2006) does not indicate any apparent association between IOD and SAM. Meanwhile, it is hinted in an observational analysis that the development of a positive IOD event in June–August may precede the development of the negative phase of SAM in September–November (HZA). In this background, it may be interesting to see whether the impacts of IOD and ENSO on the SH storm-track activity presented above are merely a manifestation of their possible association with SAM.

It is clear from Fig. 11a that SAM exerts significant modulations on storm-track activity in the extratropical maritime regions and their surroundings (e.g., Thompson and Wallace 2000), including southeastern Australia and New Zealand. Nevertheless, the impacts of IOD and ENSO events on the storm tracks, as shown in Figs. 11b and 11c versus Figs. 3a and 5a, respectively, are not modified in any substantial way in the presence of SAM. We must keep in mind, however, that there is a possibility of statistical over-fitting due to the relatively short time span as compared to the number of variables used to obtain partial correlations presented in Fig. 11. To verify the robustness of the ENSO and IOD impacts, the partial correlations of 300-hPa $Z_e$ with the Niño-3 index (IODMI), from which only the influence of the SAM index has been removed, are presented in Fig. 12a (Fig. 12b). Figure 12 indicates that overall features in the ENSO and IOD influence on storm-track
activity are unchanged with the removal of the SAM contributions only, but the wintertime ENSO correlation over continental Australia tends to be rather sensitive to SAM, and to IOD, for the 25 recent years. This may be due to the tendency for IOD events to peak by October, while ENSO events are likely to be still in the developing stage. This may also be due to the fact that climatic impacts of ENSO events over Australia have been under modulation in recent decades (Meyers et al. 2007). We have also verified that the ENSO and IOD impacts on the low-level eddy activity, as shown in Figs. 3c and 5c, are not significantly affected by the influence of SAM (figures not shown).

Figures 11b and 11c indicate that in particular mid-latitude regions over the SH, including southern Australia, the remote influence of IOD/ENSO on local storm-track activity can be detected as a significant signal in winter, despite the dominant influence of SAM from a hemispheric viewpoint. Even in those regions, however, the correlation of the ENSO/IOD signal is not high enough to claim that the tropical influence accounts for the major fraction of the interannual variance. The storm-track activity in those regions for a particular winter may also be under the influence of extratropical variability including SAM or the so-called wavenumber-three pattern. Interestingly, a recent study (Na et al. 2005) hints that IOD events may potentially influence the wavenumber-three pattern, though the opposite may not be true (L. Na 2006, personal communication). However, investigating this aspect is beyond the scope of the present study.

Influence of SAM on the IOD-associated precipitation anomalies has been assessed in the same manner. Partial correlations of CMAP precipitation with IODMI and those of precipitation with the index of IOD-associated storm-track activity from which only the SAM influence has been eliminated have been computed (figures not shown). The significant negative correlations presented in Fig. 9b over continental Australia and New Zealand are still present without the SAM signal, demonstrating a significant impact of IOD on winter rainfall variability in these regions. A closer comparison with Fig. 9b reveals that the signal of IOD-related rainfall reduction is slightly enhanced over southwest Australia, while it is marginally weakened around Tasmania and New Zealand. The corresponding partial correlations of the Niño-3 index with the rainfall indices of ENSO after removal of SAM also demonstrate the robust impact of ENSO on Tasmania (figure not shown).

4. Summary and concluding remarks
In this paper, the influence of ENSO and IOD phenomena on SH storm-track activity during austral winter (June–October) has been studied. We have found that ENSO influences the storm-track activity by modulating the STJ and PFJ, especially over the Indo-Pacific sector. In an El Niño winter the STJ tends to
strengthen, enhancing the equatorward dispersion of upper-level eddy activity and thus resulting in the weakening of the midlatitude storm-track activity over the central and eastern Pacific. The weakening is associated with a weakened midlatitude PFJ throughout the troposphere. The La Niña impact is the opposite. Our assessment of eddy feedback forcing via anomalous vorticity transport confirms that the anomalous storm-track activity acts to maintain the anomalous westerlies. From the teleconnection point of view, the above findings have implications for the Tasmanian winter climate. During an El Niño (a La Niña) winter, the midlatitude storm track tends to be located farther away from (closer to) Australia and New Zealand.

Fig. 13. (a) As in Fig. 3a but for 13 odd-numbered years (1979, 1981, . . . , 2003). (b) As in Fig. 5a but for the odd-numbered years. In (a) and (b), 0.38 (0.48) are the significant correlations at 80% (90%) confidence levels from 1000 randomized Monte Carlo simulations. (c) As in Fig. 3a but for 12 even-numbered years (1980, 1982, . . . , 2002). (d) As in Fig. 5a but for the even-numbered years. In (c) and (d), 0.4 (0.5) are the significant correlations at 80% (90%) confidence levels from 1000 randomized Monte Carlo simulations.
Anomalous westerlies as a remote response to IOD also influence the SH storm-track activity in winter. During a positive (negative) IOD event, the westerlies and storm-track activity tend to weaken (be enhanced) over the southeastern Australian continent. Rainfall anomaly signatures due to IOD- and ENSO-related anomalous storm-track activity are robust and largely are not part of the SAM influence. IOD impacts are stronger than ENSO in the study region. A concomitant occurrence of an El Niño event and a positive IOD event may have the potential to cause a significant deficit in winter precipitation over the Australian–New Zealand region, as actually observed during 1997 (figure not shown). In addition, IOD acts to modulate the storm-track activity over midlatitude South America. This may not necessarily be a statistical artifact, as Saji et al. (2005) showed through a wave activity flux diagnosis (Takaya and Nakamura 2001) applied to observed 200-hPa streamfunction anomalies that the IOD teleconnection extends almost over the entire extratropical SH in the form of a stationary Rossby wave train propagating through the midlatitude westerly jet. They substantiated this diagnostic result by carrying out idealized experiments with a mechanistic atmospheric model. As seen in Fig. 5, IOD events appear to have some association with the winter climate also over a few subpolar regions, in agreement with Na et al. (2005).

It should be stressed that the focus of the present study is placed on austral winter (June–October) when an IOD event usually develops and then peaks. Therefore, our statistics are likely to extract the maximum possible impact of IOD. Since an IOD event diminishes after November and an ENSO event does not necessarily peak in the June–October period, most of the tropical influence on the SH storm-track activity and associated precipitation in austral summer is likely due to ENSO. It should also be stressed that this study is subject to certain limitations, including the one arising from the quality of the reanalysis data, the relative shortness of our analysis period (1979–2003), and the possible decadal modulation of the signals. Therefore, the statistics concerning ENSO and IOD obtained in this study through linear analysis may not necessarily be quite robust. We have assessed the robustness of our ENSO- and IOD-related statistics shown in Figs. 3a and 5a, respectively, by repeating the same analysis separately for the even- and odd-numbered years (Fig. 13). Overall higher local correlations for the odd-numbered years mean a certain level of sampling fluctuations included in our statistics. Inclusion of the 1983 and 1997 warm events and the 1989 cold event contributes to the higher Niño-3 correlation for the odd-numbered years (Fig. 13a). The ENSO signal in storm-track activity appears to be robust over the south Indian Ocean, South Pacific, and Atlantic, while it appears to be less robust around Australia. For each of the two subsets (Figs. 13b and 13d), most of the remote IOD signals in storm-track variability shown in Fig. 5a are reproduced qualitatively, including the weakening of eddy activity over southern Australia and the enhancement off the southeastern coast of the continent and over the south Indian and Atlantic Oceans. Because of the sampling fluctuations that appear in Fig. 13, robustness of the statistics presented in this paper should be reassessed in the future with much longer data time series. Nevertheless, our findings based on currently available data for 25 recent years have some potentially important implications for seasonal climate studies and forecasts for SH winter. Hence, we propose to use station rainfall data in the Australia–New Zealand region in our future study for a more quantitative assessment of the IOD and ENSO impacts on modulations of storm-track activity. Furthermore, to understand the impacts of the Indo–Pacific sector coupled phenomena on the SH winter storm tracks in detail, high-resolution atmospheric as well as coupled general circulation models will also be used in future studies. Given the apparent decadal modulations in ENSO (e.g., Nitta and Yamada 1989; Luo et al. 2003) and IOD (Ashok et al. 2004a; Tozuka et al. 2007), it will be interesting to explore in the future how these decadal-scale modulations and variations in the tropical variability could modulate the SH storm-track activity and the associated precipitation during austral winter.

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APPENDIX

Partial Correlation Coefficients

The partial correlation coefficient \( r_{12,3} \) between two variables, \( A_1 \) and \( A_2 \) after removing the influence of the variable \( A_3 \), is given by
\[ r_{12,3} = \frac{r_{12} - r_{13} r_{23}}{\sqrt{(1 - r_{13}^2)(1 - r_{23}^2)}}. \]  

(A1)

In Eq. (A1) the term \( r_{12} \) represents the linear correlation coefficient between \( A_1 \) and \( A_2 \), and the same definition holds for other terms in Eq. (A1).

On similar lines, the partial coefficient \( r_{12,34} \) between two variables \( A_1 \) and \( A_2 \), after removing the influence of variables \( A_3 \) and \( A_4 \), is obtained by

\[ r_{12,34} = \frac{r_{12,4} - r_{13,4} r_{23,4}}{\sqrt{(1 - r_{13,4}^2)(1 - r_{23,4}^2)}} \]

\[ = \frac{r_{12,3} - r_{14,3} r_{24,3}}{\sqrt{(1 - r_{14,3}^2)(1 - r_{24,3}^2)}}. \]  

(A2)

The number of degrees of freedom for partial correlations was fixed in this study at \( N - 3 \) for the first order, and \( N - 4 \) for the second order, with \( N \) being the number of data values in the time series.

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