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ARTICLE in JOURNAL OF CLIMATE · APRIL 2015
Impact Factor: 4.9 · DOI: 10.1175/JCLI-D-14-00390.1

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The Influence of the Indian Ocean Dipole on Antarctic Sea Ice*

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(Manuscript received 3 June 2014, in final form 7 November 2014)

ABSTRACT

This study explores the impact of the Indian Ocean dipole (IOD) on the Southern Hemisphere sea ice variability. Singular value decomposition (SVD) of September–November sea ice concentration and sea surface temperature (SST) anomalies reveals patterns of El Niño–Southern Oscillation (ENSO) in the Pacific and the IOD in the equatorial Indian Ocean. The relative importance of the IOD’s impact on sea ice in the Pacific sector of Antarctica is difficult to assess for two reasons: 1) ENSO generates larger anomalies in the Pacific and Weddell Sea and 2) many of the positive (negative) IODs co-occur with El Niño (La Niña). West of the Ross Sea, sea ice growth can be attributed to the negative heat fluxes associated with cold meridional flow between high and low pressure cells generated by the effects of the IOD. However, the locations of these positive and negative pressure anomaly centers tend to appear north of the sea ice zone during combined ENSO–IOD events, reducing the influence of the IOD on sea ice. The IOD influence is at a maximum in the region west of the Ross Sea. When ENSO is removed, sea ice in the Indian Ocean (near 60°E) increases because of cold outflows west of low pressure centers while sea ice near 90°E decreases because of the warm advection west of a high pressure center located south of Australia.

1. Introduction

The Indian Ocean dipole mode (IOD) discovered by Saji et al. (1999) and independently by Webster et al. (1999) is the presence of a zonal gradient in oceanic and atmospheric variables over the tropical Indian Ocean. The occurrence of this polarity over a latitudinal belt of 10°S–10°N is known to influence the weather and climate in many tropical and extratropical regions (Saji and Yamagata 2003). Australia experiences reduced rainfall because of anomalous subsidence over the eastern tropical and subtropical Indian Ocean during a positive IOD event (Ashok et al. 2003a). Influence of the IOD can be detected in Southern Hemispheric storm track activities in southern Australia and New Zealand (Ashok 2007a). Out of 21 significant bushfire seasons of Australia, 11 were preceded by positive IOD events (Cai et al. 2009a), one of which resulted in the notorious bush fire inferno called “Black Saturday.” The impact of the IOD can be traced to the extratropical and polar Southern Hemisphere. IOD events excite Rossby wave trains in the eastern tropical Indian Ocean, which are then trapped in the westerlies and propagate around Antarctica, modulating the Southern Hemisphere surface temperature anomalies (Saji et al. 2005; Cai et al. 2009b, 2011). This influence the weather in remote regions as far as South America, where the rainfall is modified by anomalous anticyclonic wind patterns induced by the Rossby wave trains associated with the IOD events (Chan et al. 2008). The impact of El Niño–Southern Oscillation (ENSO) over southern Australia and China occurs through ENSO’s coherence with IOD (Cai et al. 2011).

Sea ice, on the other hand, is one of the highly varying cryospheric parameters. The growth and decay of sea ice on different time scales, ranging from seasonal to decadal, is associated with numerous processes. The Antarctic

* National Centre for Antarctic and Ocean Research Contribution Number 06/2015 and Lamont-Doherty Earth Observatory Contribution Number 7874.

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DOI: 10.1175/JCLI-D-14-00390.1

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Oscillation (AAO) influences sea ice by means of anomalous mean surface heat flux and ice advection (Liu et al. 2004). The influence of ENSO generates a dipole pattern called the Antarctic dipole in western Antarctica (Yuan and Martinson 2000; Yuan 2004). Both AAO and the ENSO also synergistically influence the Antarctic sea ice (Pezza et al. 2012). The physical processes that influence the annual sea ice growth and decay are well documented; however, the manner in which they combine to promote sea ice variability on decadal time scales is not well studied (Zwally et al. 2002). The dynamic and thermodynamic processes that relate ENSO influence on Antarctic sea ice have been well documented only in the Western Hemisphere (Yuan 2004); what processes determine the sea ice variability in other Antarctic regions still largely remain unknown. Sea ice is a critical component of the climate system; hence, it is important to search for additional drivers of its variability. The IOD is another important tropical process that has far-reaching influence. Although the IOD is linked to extratropical weather, its relationship with the Southern Hemisphere sea ice has not been explored. The present study explores the impact of the IOD farther south to the seasonal sea ice zone.

2. Data and methods

Monthly sea ice concentration from the National Snow and Ice Data Center (NSIDC) (Comiso 2008) and Hadley Centre SST data (Rayner et al. 2003) for the time period 1979–2006 are used in this study. We also use geopotential heights from NCEP–NCAR reanalysis (Kistler et al. 2001). All anomaly series are generated relative to the climatology for the period of 1979–2006.

Singular value decomposition (SVD; Bretherton et al. 1992; Björnsson and Venegas 1997) is employed to investigate the covariability and possible links between the southern high-latitude sea ice and the tropical Indian Ocean SST. The analysis yields coupled spatial patterns for sea ice and SST anomalies, and their temporal variability in two time series [expansion coefficients (EC)]: one for the SST and another one for the sea ice for each mode. It also provides a squared covariance fraction (scf), which describes the amount of variability each mode represents in terms of a fraction of total coviability.

The season selected for the present study is September–November (SON), because the IOD peaks during this period. Also during SON (austral spring) the IOD can be identified as an intrinsic mode of variability (Hannachi and Dommenget 2009). In addition, our lead–lag correlations suggested that sea ice most responds to the IOD in this season despite the fact that quasi-stationary Rossby waves would be most favored in winter (Jin and Hoskins 1995). Moreover, Pacific sea ice and the Southern Oscillation index covary synchronously in SON but not in other seasons (Simmonds and Jacka 1995). Therefore, we focus our investigation over SON.

In addition, Rossby wave activity flux as described by Takaya and Nakamura (2001) is computed to investigate the Rossby wave propagation from the tropics. This is a diagnostic tool that illustrates the evolution and propagation of quasigeostrophic stationary or migratory wave disturbances on a zonally varying basic flow (Takaya and Nakamura 2001). For this study, 250-hPa geopotential height is regressed on to three indices: the leading mode SST time series, Niño-3 SST anomalies, and the Indian Ocean dipole mode index (DMI). These regressed values are used to compute the Rossby wave activity flux. To isolate the IOD’s impact on wave activity, we also remove the Niño-3 SST regressions from the geopotential height anomalies and then regress the residual on to the DMI. These four regressed fields are used to compute the Rossby wave activity flux. Since standing waves have a dominant impact on sea ice (Liu et al. 2002), the zonal phase speed in Eq. (38) of Takaya and Nakamura (2001) was set to zero. To understand the physical mechanisms that drive the sea ice variability, meridional heat flux associated with the standing wave pattern is computed as

\[ Qe = V^* T^*, \]

where \( V^* \) and \( T^* \) represent zonal anomalies of meridional winds and air temperature at the surface. We then regress these wind and air temperature anomalies on to various indices as described in the case of Rossby wave activity flux. Regression analyses were also conducted to ascertain the relationship of the dominant spatial patterns with the oceanic as well as atmospheric parameters, such as SST and geopotential height. Here various indices like the EC of the SVD modes and DMI ENSO are regressed onto the spatial fields of geopotential height, sea surface temperature (SST), sea level pressure (SLP), etc. To remove the effects of El Niño/La Niña, initially Niño-3 SST anomalies were regressed out of the spatial fields, followed by regressing its residual on to the DMI. To obtain the IOD-related \( Qe \), we first calculate partial regressions of \( V^* \) and \( T^* \), and then compute \( Qe \) following Eq. (1). Statistical significance tests were conducted based on a Student’s \( t \) test.

3. Results and discussion

The first three SVD modes between SST and sea ice are presented to illustrate the relationship between polar sea ice and extrapolar SST variability (Fig. 1). The scf (correlations) between two expansion coefficient series
for modes 1, 2, and 3 are 63.1 (0.71), 13.5 (0.75), and 5.9 (0.65) respectively. As expected, the SST and sea ice spatial patterns of the leading mode depict El Niño in the tropical Pacific and the IOD in the equatorial Indian Ocean (Fig. 1a). Correlation between the leading mode EC of SST and the SST anomalies also exhibited a similar pattern (Fig. 2a). The corresponding sea ice pattern (Fig. 1b) resembles the Antarctic dipole, a high-latitude mode triggered by ENSO (Yuan and Martinson 2000; Yuan 2004). Figure 1b represents sea ice anomalies in responding to an ENSO warm event. The leading mode EC of SST is well correlated with the Niño-3 and Niño-3.4 indices ($r \approx 0.9$) as well as with DMI ($r \approx 0.7$). Strong correlation of leading mode with the DMI raises the question as to whether the IOD always coexists with ENSO. This may not be the case all the time. Although positive (negative) IODs mostly co-occur with El Niño (La Niña), some IOD events occur in its absence (e.g., 1961, 1967, and 1994). During 2007, a positive IOD co-occurred with a La Niña. ENSO’s contribution to the variance of the IOD seems to be small even though the correlation between both the phenomena is high. This high correlation is due to many in-phase IOD and ENSO events (Ashok et al. 2003b). Mathematical analyses with filtered SST data in the Indian Ocean showed that the IOD is a physical mode of the Indian Ocean during autumn (Hannachi and Dommenguet 2009). Earlier coupled general circulation model studies also revealed this character. However, once generated, the IOD may interact with the ENSO (Behera et al. 2006). Paleoclimate records obtained from coral reefs suggest that only 56% of the moderate to strong IOD events co-occurred with ENSO (Abram et al. 2008). In the instrumental records, about 25 IOD events occurred when the Pacific was in neutral state (Meyers et al. 2007). These studies show that the IOD exists physically and sometimes independently from ENSO. It could generate its unique tropical forcing to extratropical regions.

The second mode displays a pattern characterized by warm (cold) SST anomalies in the western/central (eastern) tropical Pacific, cold anomalies in the eastern equatorial Indian Ocean, and warm anomalies in the subtropical gyre of the Indian Ocean (Figs. 1c and 2b). The mode-2 SST field has characteristics of two recently discovered phenomena, the subtropical Indian Ocean dipole (Behera and Yamagata 2001) and the El Niño Modoki (Ashok et al. 2007b) in the subtropical Indian Ocean and the tropical Pacific Ocean respectively. Indeed, SST EC of mode 2 is strongly correlated with the El Niño Modoki index ($r \approx 0.71$, 26 degrees of freedom). At the same time its correlation with the subtropical dipole is also statistically significant ($r \approx 0.41$, 26 degrees of freedom).
The time series, however, only weakly correlates with the Niño-3 index (-0.16). Hence the north–south pattern of SST anomalies in the central tropical Pacific or in the subtropical Indian Ocean is likely related to the sea ice pattern in Fig. 1d, particularly in the South Pacific. In addition, a wave-3 pattern was present in middle-to-high latitudes around Antarctica (Fig. 1c). The sea ice field of mode 2 also resembles the La Niña phase of Antarctic dipole anomalies in response to the cold anomaly in the eastern tropical Pacific (Fig. 1d). Contrasting with the leading mode with maximum sea ice responses in the Western Hemisphere, mode 2 influenced the southeastern and southwestern Pacific and western Indian Ocean (Figs. 1d, 2b). Mode 3 of SST exhibits a La Niña pattern in the tropics with stronger anomalies in the Pacific but weaker anomalies in the tropical Indian Ocean. (Figs. 1e, 2c) No large sea ice anomalies occur in the Antarctic dipole region. This tropical SST pattern is more associated with the sea ice anomalies in the Indian Ocean and Atlantic sectors of the Antarctic (Fig. 1f). In summary, the first mode depicts ENSO warm events with an associated Indian Ocean dipole pattern in the tropics and corresponding Antarctic dipole anomalies in the Antarctic sea ice field. The sea ice field of the second mode resembles the La Niña phase of the Antarctic dipole although the SST EC of this mode correlates strongly with El Niño Modoki. Thus SVD suggests that signatures of the IOD can be found mostly in the leading mode, and hence discussion is focused on this mode.

Regression of the leading mode SST series on 250-hPa geopotential anomalies is characterized by alternate regions of negative (low pressure) and positive (high pressure) anomalies, indicative of Rossby wave trains emanating from the tropics (Fig. 3a, shading). Similar patterns are found in regression of the geopotential height anomalies with the Niño-3 index and DMI (Figs. 3b,c, shading). Overlaid on each of these regression plots is the composite of Rossby wave activity flux (vectors) for three strong years of positive phase of leading mode (1982, 1987, 1997, 2002, and 2006) in (Fig. 3a), ENSO (1982, 1987, 1997, 2002, and 2006) in (Fig. 3b), and DMI (1994, 1997, 2002, and 2006) in (Fig. 3c). During ENSO and IOD years, Rossby wave activity fluxes emerge from both the tropical Pacific and tropical Indian Ocean. In the case of leading mode and Niño-3 (Figs. 3a,b), the flux emerging from the tropical Indian Ocean is weakened south of Australia, but restrengthened southeast of
it and merged with the flux from the tropical Pacific in the south-central Pacific. A part of this merged flux reaches the Weddell Sea and another part is reflected back into the tropics over South America and Africa. Thus the alternate regions of positive (high pressure) and negative (low pressure) regression coefficients represent Rossby wave trains emanating from the tropics. These are the Pacific–South American (PSA) pattern and the western and eastern Indian Ocean wave train arising from the eastern and western Indian Ocean SST variability (Cai et al. 2011). The IOD is also associated with Rossby wave trains in the Southern Hemisphere (Saji and Yamagata 2003; Saji et al. 2005). Figure 3c shows that the southeastern Indian Ocean emanates more wave energy to southern high latitudes during IOD events than during ENSO events. When ENSO is removed from the geopotential heights by linear regression, the flux from the Pacific is absent, while the flux from the eastern Indian Ocean is strengthened south of Australia and the southwestern Pacific sea ice region (Fig. 3d, vectors). One part of the wave energy is converged into the Ross Sea and another part is reflected north over South America. The response to these wave activity fluxes is manifested as positive and negative pressure centers as shown in Fig. 3d. This is the eastern Indian Ocean wave train described by Cai et al. (2011). Also there is an indication of an African wave train emanating from the tip of Africa. This could probably be a part of the equatorial African wave train suggested by Cai et al. (2011) or part of the wave train emanating from the eastern Indian Ocean and trapped in the westerly wind belt after one part of it exits over South Africa. Although they are weak compared to the ENSO-related Rossby waves, these wave trains also generate statistically significant signals in the height field (Fig. 3d, shading). Moreover the locations of the significant positive (south of Australia) and negative regression coefficients (near 60°E and in the Ross Sea) are closer to the Antarctic continent when the influence of ENSO is removed from both the fields (Fig. 3d) and hence more likely impact the ice field. Unlike the influence from ENSO, the height field responds to IOD with negative (positive) pressure anomalies in the Ross Sea region (south of Australia) (Fig. 3d). In the presence of ENSO (Figs. 3a–c), significant positive pressure anomalies are present near the Amundsen Sea as a result of the Pacific wave train. In fact, this anomalous pressure cell drives the Antarctic dipole (Yuan 2004). On the other hand, a positive IOD in the absence of ENSO generates a different wave train with a high pressure center south of Australia and low pressure system in the Ross Sea, without significant pressure anomalies in the Amundsen Sea. It seems that the IOD’s influence does not extend to the region east of the Ross Sea, where both southern annular mode (SAM) and ENSO exert significant influences on the sea ice extent synergistically (e.g., Pezza et al. 2012). However, the correlation between SAM and ENSO ($r \sim -0.24$) is not statistically significant at the 95% level during the time period of our analysis.

Regression analysis indicates that the negative pressure anomaly (low pressure region) during positive IOD and warm ENSO events in the Ross Sea (Figs. 3a–c) lies within the seasonal sea ice zone and can influence the sea ice variation. The leading mode spatial pattern of sea ice (Fig. 1b) displays an area of positive ice concentration...
from 180°E to 100°E followed by negative values near 90°E. The analysis described earlier suggests that the changes in sea ice in the western Ross Sea could be due to the negative pressure anomaly induced by both the IOD and ENSO, whereas south of Australia the changes are due to the positive pressure anomaly induced by the IOD alone. This can be explained by the flow structure around a low pressure anomaly. A low pressure anomaly will generate cold (warm) flow along its western (eastern) limb.

To further illustrate that the heat transport is a key factor linking the anomalous atmospheric circulation and sea ice anomalies, we computed the meridional heat fluxes associated with IOD excited wave trains. The regression of zonal SLP anomaly on DMI was also characterized by a low pressure cell in the Ross Sea. The southward (northward) flow along its eastern (western) limbs resulted in positive (negative) heat fluxes (Fig. 4). Thus the northward (cold) southward (warm) flow will increase (decrease) sea ice thermodynamically. Regression of sea ice anomalies with leading mode SST time series during SON also reveals a spatial pattern similar to the leading mode spatial pattern of sea ice (Figs. 5a and 1b). This spatial pattern is very similar to the regression of sea ice concentration anomaly on the Niño-3 index (Fig. 5b). The IOD is negatively (positively) correlated with sea ice in regions east (west) of the Ross Sea (Fig. 5c). When the ENSO effect is removed from the IOD, the Antarctic dipole is absent, and positive regressions persist and are strengthened west of the Ross Sea. Patches of negative sea ice anomalies can be seen near the prime meridian and near 90°E, whereas positive sea ice anomalies can be found near 60°E (Fig. 5d). But these are weaker than those observed just west of the Ross Sea. It is also noted that sea ice and ENSO exhibit a lead–lag relationship (Simmonds and Jacka 1995; Yuan and Martinson 2000). These lead–lag correlations reveal an eastward propagation of sea ice anomalies with a speed of approximately 45 km yr⁻¹ (White and Peterson 1996; Yuan and Martinson 2000). Since the strongest ENSO impact occurs in the Pacific, we expect minor impacts of ENSO to interfere with the interpretation of the correlations west of the Ross Sea, particularly when Antarctic dipole is absent in the Western Hemisphere. Thus both the regression and SVD analyses for SON point to the following results: 1) there is an identifiable covariability between the IOD and Antarctic sea ice; 2) both the IOD and ENSO have in-phase impact on sea ice east of the Ross Sea, while IOD could influence sea ice in the Indian Ocean sector (near 60°E and near 90°E) when ENSO is absent; and 3) the IOD’s impact on sea ice does not extend to the Western Hemisphere.

The covariability between the IOD and sea ice can be explained by the heat advection between anomalous low and high pressure centers. Why there is a weak response of sea ice to IOD during ENSO years and stronger IOD–sea ice correlations when ENSO was removed needs explanation. One reason for these phenomena is the different wave train paths during these two situations. The wave train generated by IOD alone curves in to the sea ice zone from the southern Indian Ocean to the southwestern Pacific (cf. the three significant correlation centers in the region in Fig. 3d with that in Figs. 3a–c). When ENSO events are present, the wave train originating in the tropical Indian Ocean shifts northward in the Indian Ocean and the negative pressure center in the southwestern Pacific shifts westward (cf. the negative pressure anomalies near the Ross Sea in Figs. 3a–c with Fig. 3d).

Regressions of sea ice and DMI reflect the impact of the westward and southward shifts in geopotential heights anomaly centers. The positive anomalies southeast of Australia are much stronger in sea ice–DMI regressions (Figs. 5c,d) than in the sea ice–ENSO regressions (Figs. 5a,b). Also regressions near 90°E in the Indian Ocean are weak in the presence of ENSO (Figs. 5a–c), owing to the fact that the responding pressure centers are located farther north away from the sea ice zone (Figs. 3a–c). In the absence of ENSO, the negative pressure center near 60°–70°E and positive pressure center south of Australia are much closer to the sea ice zone and capable of generating sea ice anomalies reflected in Fig. 5d through the thermodynamic process.

4. Conclusions

Singular value decomposition of Antarctic sea ice concentration and Southern Hemisphere SST reveals
that ENSO and the response of Antarctic sea ice to ENSO constitute the leading mode. However, contributions from the IOD cannot be ignored. Most of the significant differences between regressions of sea ice with IOD and ENSO lie within the Antarctic dipole region and south of Australia. Our results indicate that the influence of the IOD is strong in the central Pacific sector, west of the Ross Sea and Indian Ocean sector of the Antarctic. The IOD-generated wave train produces three pressure centers (low, high, low) from the Indian Ocean to the Ross Sea. All three centers are close to the sea ice zone. Northward (southward) flow associated with these anomalous pressure centers brings cold (warm) air to polar seas, resulting in weak positive (negative) ice anomalies near 60°E (90°E), while strong positive anomalies are present in the region west of the Ross Sea. When IOD and ENSO coexist, which happens frequently during IOD years, the IOD-generated wave

![FIG. 5. (a) Regressions of SST EC from the leading SVD mode on sea ice anomalies, (b) regressions of Niño-3 index on sea ice anomalies, (c) regressions of DMI on sea ice anomalies, and (d) partial regression of DMI on sea ice anomalies (see section 3). Regressions that are significant at 90% confidence level or greater are shown in color shades. Note that positive regressions west of the Ross Sea in (a) occupy a more westward position (near 120°E) than that in (d), reflecting that Rossby wave trains take different paths in these two situations.](image-url)
train tends to shift northward in the Indian Ocean and westward in the Ross Sea (see Fig. 6 for schematics). This shifted wave train path results in much less of an IOD impact on sea ice in the Indian Ocean sector and particularly less ENSO–IOD joint influence on sea ice west of the Ross Sea. The partial regression analysis suggests that the IOD’s influence on sea ice does not extend to east of the Amundsen Sea (e.g., the Bellingshausen Sea, Antarctic Peninsula region, and Weddell Sea), since the wave train generated by the IOD alone does not extend to the Amundsen Sea and beyond.

Acknowledgments. The authors thank the director of NCAOR for his keen interest and encouragement for this study. This work was carried out under a fellowship awarded by the Scientific Committee on Antarctic Research (SCAR). We thank the anonymous reviewers for their suggestions. Yuan is supported by U.S. NSF OPP Grant ANT10-43669. NCEP reanalysis data were provided by the NOAA/OAR/ESRL PSD, Boulder, Colorado, USA, from their web site http://www.esrl.noaa.gov/psd/. SSTs were obtained from Met Office Hadley Centre web site http://hadisst/data/download.html and sea ice data were downloaded from NSIDC.

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