Observed Evidence of an Impact of the Antarctic Sea Ice Dipole on the Antarctic Oscillation

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ABSTRACT

A lagged maximum covariance analysis (MCA) is applied to investigate the linear covariability between monthly sea ice concentration (SIC) and 500-mb geopotential height (Z500) in the Southern Hemisphere (SH). The dominant signal is the atmospheric forcing of SIC anomalies throughout the year, but statistically significant covariances are also found between austral springtime Z500 and prior SIC anomalies up to four months earlier. The MCA pattern is characterized by an Antarctic dipole (ADP)-like pattern in SIC and a positively polarized Antarctic Oscillation (AAO) in Z500. Such long lead-time covariance suggests the forcing of the AAO by persistent ADP-like SIC anomalies. The leading time of SIC anomalies provides an implication for skillful predictability of springtime atmospheric variability.

1. Introduction

The interaction between the ocean and atmosphere plays an important role in shaping the climate and its variations. There has been considerable interest in and research on the predictability of atmospheric variability at seasonal to interannual time scales related to the interaction between the extratropical ocean and its overlying atmosphere [see the review by Kushnir et al. (2002)]. In the Southern Hemisphere (SH), recent numerical studies have shown that changes in winter sea surface temperature (SST) in the extratropics have a profound effect on the southern annular mode [SAM, also called Antarctic Oscillation (AAO); Thompson and Wallace 1998] (Watterson 2001; Marshall and Conolley 2006; Sen Gupta and England 2007).

Sea ice is also an active component in the climate system. Not only is it sensitive to dynamic and thermodynamic forcings from overlying atmosphere and underlying ocean, it also modulates atmospheric and oceanic circulations through altered radiative properties, atmosphere–ocean heat exchanges, momentum flux, and brine rejection/ freshwater release. Recent observational studies have shown that summer to autumn sea ice anomalies in the Arctic significantly impact the following wintertime atmospheric circulation in the Northern Hemisphere (Francis et al. 2009; Honda et al. 2009; Wu and Zhang 2010). The influence of wintertime arctic sea ice on the atmosphere has been investigated in many modeling studies (e.g., Deser et al. 2004, 2007). In the SH, modeling studies have also shown that changes in Antarctic sea ice can modify atmospheric conditions (e.g., Raphael 2003; Lachlan-Cope 2005). Stammerjohn and Smith (1997) find that Southern Ocean sea ice variability plays an important role in modulating regional heat budgets. They suggest that feedback processes between sea ice and the atmospheric circulation may act to enhance low-frequency variability in some high-latitude regions. However, few studies have provided observational evidence of the possible feedback of Antarctic sea ice anomalies on the large-scale atmospheric circulation.

In this study, we investigate the linear covariability between the Antarctic sea ice concentration (SIC) and SH atmosphere circulation in observations by using a lagged maximum covariance analysis [MCA; also known as singular value decomposition (SVD) analysis; e.g., Bretherton et al. (1992)]. The lagged MCA has been increasingly utilized to investigate the impact of midlatitud
SST on the atmosphere (e.g., Czaja and Frankignoul 2002; Frankignoul and Sennechael 2007; Wu 2008). As noted in Czaja and Frankignoul (2002), the lagged covariance is powerful in distinguishing between cause and effect. On monthly or longer time scales, the atmosphere primarily acts as a white noise forcing on the midlatitude ocean and sea ice. If the Southern Ocean sea ice only responds passively, there should be no covariance between sea ice and the SH atmospheric variable when the sea ice leads by more than the atmospheric persistence time. If Southern Ocean sea ice fluctuations have an impact on the SH atmosphere, there should exist significant cross covariance when the sea ice leads. Such signatures are searched for here as a function of time lag along the course of a year. The rest of this paper is arranged as follows: section 2 describes the data sources and analysis techniques, section 3 presents MCA results, and section 4 provides a summary.

2. Datasets and methodology

The atmospheric datasets used in this study are geopotential height, wind, temperature, sea level pressure (SLP), air temperature at 2 m (T2m), and surface energy heat fluxes in the National Centers for Environmental Prediction (NCEP)–National Center for Atmospheric Prediction (NCAR) reanalysis (Kalnay et al. 1996). The SIC data are the newly updated homogeneous monthly high-resolution data derived from passive microwave satellite data with the bootstrap algorithm from 1978 to 2008 (Comiso et al. 2008). We use the monthly SIC in units of percent (%) available on a 448 × 304 horizontal grid (25-km mesh) and restrict our analyses to the period from 1979 to 2007. We also use the global optimum interpolated SST (Smith et al. 1996). The climatology (29-yr mean) of each month is removed separately for atmosphere, SIC, and SST datasets. To reduce the influence of trends, linear trends were removed from the monthly anomalies by the least squares fit at each grid point for all fields. At each grid point for the SIC, atmosphere, and SST anomalies, we used the method by Wu (2008) to filter out the tropical Pacific influence using a regression against the Niño-3.4 (5°N–5°S, 120°–170°W) SST anomalies of the preceding months. The regression coefficient is selected as the maximum regression coefficient within the preceding six months.

To search an SIC anomaly influence on the atmosphere, the MCA is applied as a function of time lag and season to monthly SIC and 500-mb geopotential height (Z500) anomalies in the domain from 20° to 90°S. Sets of three successive months [months are denoted by the first letter of each month, e.g., January–March (JFM)] were considered with SIC leading or lagging Z500 anomalies. For each season and lag, the MCA is based on 28-yr (84 months) data. We evaluate the significance of statistics in the MCA, the squared covariance (SC), and the temporal correlation $r$ between the expansion coefficients of Z500 and SIC, by exactly the same Monte Carlo approach described in Czaja and Frankignoul (2002). One hundred ensembles of MCA between the scrambled Z500 and original SIC are performed. The predictability of the atmospheric signal is calculated by the cross validation, removing successive sets of 3 yr of the original anomalous fields before the MCA and then using the resulting MCA patterns to determine their amplitude in the middle year that was removed.

In both MCA and empirical orthogonal function (EOF) analysis, area weighting is accomplished by multiplying the SIC by the square root of the area at each grid point and Z500 by the square root of the cosine of latitude before computing the covariance matrix to ensure that equal areas are afforded equal weight in the analysis. The leading EOFs of JFM (and other seasons) monthly Z500 and SIC anomalies are readily identifiable as being the AAO and Antarctic dipole (ADP) patterns, respectively (Thompson and Wallace 1998; Yuan and Martinson 2001). The corresponding standardized principal component (PC) time series are defined as the AAO and ADP indices. At last, the leading EOFs of sets of three successive months [e.g., August–October (ASO)] of 28-yr SST in the extratropics of SH were calculated.

3. Results

a. MCA results

In Fig. 1, we present the SC and $r$ of the first MCA mode in the analysis with anomaly fields of SIC and Z500 between lag −5 and lag +4. Throughout most of the year, significant SCs are found when Z500 anomalies lead SIC from 0 to +2 months, indicating that the atmosphere–sea ice connection is dominantly characterized by the atmospheric circulation forcing on SIC. This is consistent with Hall and Visbeck (2002), Yuan and Li (2008), and many others. Significant SC and at lags +3 and +4 are only found for Z500 assigned at late winter to spring seasons. In addition, significant SCs and $r$ associated with the second leading MCA mode (not shown) are also found at lags 0 to +2 for most seasons, showing that the SIC also responds to other leading modes of Z500. On the other hand, significant SCs are only found associated with the first MCA mode when SIC leads ASO, September–November (SON), and October–December (OND) Z500 by 1–5 months, while there are no significant SCs identified during the rest of the year. This suggests a seasonality of SIC impact on atmospheric circulation.
To illustrate the association of the springtime atmospheric signal with the SIC from the preceding austral fall to winter, we show in Fig. 2 the MCA patterns of the first MCA mode for Z500 in OND and SIC at the time lags between $-2$ and $+1$ months when significant SC and $r$ are found at the 10% level. Each pair of patterns is formed from the heterogeneous covariance map for Z500 and the homogeneous covariance map for SIC, which are constructed by regressing the Z500 and SIC fields onto the normalized MCA–SIC time series at each lag. When the SIC leads, the MCA patterns of SIC (Figs. 2a–d) are characterized with the ADP-like pattern, exhibiting an out-of-phase relationship between anomalies in the central/eastern Pacific and anomalies in the Atlantic sectors of the Southern Ocean. Also, the coefficient time series of these successive SIC patterns in Figs. 2a–d are significantly correlated with each other (correlations greater than 0.85). This indicates that the progression of the SIC anomalies is coherent both spatially and temporally. The atmospheric signal in OND resembles the positive phase of the AAO, with a strong anomalous low reaching 25 m to west of Antarctic Peninsula. The MCA–Z500 time series are highly correlated with the AAO index in OND with correlations above 0.96, and the MCA–SIC time series correlated with the ADP indices with correlations about 0.80 for lag $-2$ and above 0.90 for other lags (Table 1). Cross validation (Table 2) indicates that correlations are robust at lags $-4$ to $-1$. Based on the square of $r^2$ between the MCA–Z500 and MCA–SIC time series (Table 2), about 21% of the variance of the dominant AAO-like Z500 pattern in OND can be predicted from the SIC anomalies two months earlier.

Similar MCA patterns are found at lags $-4$ to $-1$ months when Z500 is fixed in SON, and at lags $-2$ to $-1$ months when Z500 is fixed in ASO, with significance in SC, $r$, and $r^2$ at the 10% level. Table 1 indicates high correlations between the MCA–Z500 (MCA–SIC) time series and the corresponding AAO (ADP) indices at about 0.90 (0.90) for lags $-4$ to $-1$ months when Z500 is fixed at SON. The significant SC, $r$, and $r^2$ of SON and OND atmosphere to SIC at the lags $-4$ to $-1$ months, ASO atmosphere to SIC at lags $-2$ and $-1$ months appears to be relevant to the persistence of ADP-like SIC anomalies from the preceding fall to winter. The relationship between the ASO, SON, OND Z500, and SIC several months earlier are so strong that it is also found implicitly in the SIC field alone. The time evolution of the ADP-like MCA SIC patterns shown in Fig. 2 and the high correlations between the MCA–SIC time series and the ADP indices primarily reflect the persistence of the ADP-like SIC anomalies. The integration of the leading SIC and springtime Z500 anomaly information suggests that a persistent ADP-like SIC anomaly from late fall to winter would cause a polarization of AAO toward its positive phase several months later.

ADP is the largest interannual variability in the Antarctic sea ice field. It has been related to variability in the tropical Pacific SST anomalies associated with the El Niño–Southern Oscillation (ENSO) (Peterson and White 1998; Yuan and Martinson 2000) and the atmospheric forcing associated with the AAO, quasi-stationary wave-3 pattern, and Pacific South American pattern (Hall and Visbeck 2002; Liu et al. 2004; Holland et al. 2005; Holland and Raphael 2006; Yuan and Li 2008).
Since the forcing associated with the ENSO was removed prior to the analysis, Fig. 3 shows how the above SIC persistence is related to the circulation changes through the simultaneous regression patterns of SLP and T2m fields onto the normalized MCA–SIC time series associated with the SIC patterns from JJA to SON seasons in Fig. 2. The SLP is characterized with a large low pressure anomaly centered on the South Pole and a ring of high pressure anomalies at midlatitudes, reminding us the influence of the AAO on the ADP-like SIC in Liu et al. (2004) and Yuan and Li (2008). Anomalous strong cyclonic circulations in the southeast Pacific in Figs. 3a–d cause an anomalous equatorward (poleward) mean heat flux at the surface in the Ross–Amundsen Sea (the Bellingshausen–Weddell Sea) (Liu et al. 2004). The above SLP regression patterns are indicative of a northwesterly wind component over the regions of reduced SIC northwest of the Ross Sea and around the Antarctic Peninsula from the Bellingshausen Sea toward the Atlantic sector and an anomalous southeasterly wind component over the area of increased SIC stretching from the eastern part of the Bellingshausen Sea across the Amundsen Sea.

**Fig. 2.** (left) Heterogeneous Z500 and (right) homogeneous SIC covariance maps of OND Z500 and SIC in units of percent in the first MCA mode at lags from −4 to +1 months. Contour interval is 5 m for Z500. Negative contours are dashed and the zero line is omitted.
toward the northern part of the Ross Sea. Patterns of T2m show that large warming is coincident with the reduced SIC in the Ross Sea and the Weddell Sea and cooling is coincident with the increased SIC in the Bellingshausen–Amundsen Sea. The broad-scale pattern of T2m is consistent with advection by the AAO-like atmospheric circulation. In addition, the response of the sea ice to the AAO-like atmospheric circulations in Figs. 3a–d might also be due to ice advection, as suggested in Liu et al. (2004). Anomalous strong intensification of the surface westerlies induces an enhanced Ekman drift to the north. This drift transports cold water equatorward and leads to enhanced equatorward ice advection in the Ross–Amundsen sector and ice divergence away from both sides of the Antarctic Peninsula (particularly the Weddell Sea).

The AAO has a strong signature in SST that is closely related to the SIC response: SIC reductions are found in nearby regions of AAO-induced SST warming and vice versa (e.g., Karpechko et al. 2009). To determine how changes in the ADP-like SIC anomalies in Figs. 2a–d are related to that in SST, we calculate the correlations between the MCA–SIC time series associated with the homogeneous SIC patterns in Figs. 2a–d and the PC time series related to the leading EOFs modes of the SST for lag 0 or negative lags (SST leads SIC). The significant correlations are about 0.20–0.35 (statistically significant at the 95% confidence level), indicating that the change in SIC in Fig. 2 is not mainly forced by in SSTs. Therefore, the influence of SIC on the AAO-like atmospheric circulation shown in Fig. 2 may be independent of that of SST. However, the possibility exists that the SAM is forced by both wintertime SST and SIC anomalies in reality. Marshall and Connolley (2006) find that warmer (colder) SST anomalies in winter, immediately north of the sea ice zone around Antarctica, weaken (enhance) the SAM.

One of the most important features in Figs. 2 and 3 is the extensive anomalous low in the Amundsen Sea, which breaks the annular pattern of AAO. Usually, the annular pattern of AAO is weak in the region and tends to bulge out to the north. The extensive interruption of annular patterns in Z500 and SLP are unique for the response of the atmosphere and happen simultaneously while sea ice anomalies persist.

In contrast to results in Figs. 2a–d, which describe the SIC forcing on the atmosphere, the dominant relationship depicted in Figs. 2e,f is the AAO-like atmospheric forcings on SIC at lag 0 and 1. Figures 2e,f also exhibit ADP-like patterns with anomalies of one sign in the Pacific and an opposite sign in the Atlantic but the action center gradually retreated to smaller areas in the high latitude. Such an impact of springtime AAO-like atmospheric forcing on the ADP-like SIC variability should result from a combination of the anomalous mean surface heat flux and ice advection, as reported in previous studies (e.g., Liu et al. 2002, 2004; Yuan and Li 2008).

b. Impact of sea ice on atmospheric variables

Variations in sea ice can affect the atmospheric circulation through changes in surface albedo and surface fluxes of heat, moisture, and momentum. In SH winter seasons, the albedo effect is suppressed owing to the low insolation. To determine how the above ADP-like SIC anomalies in Figs. 2a–d are related to heat flux changes, we have shown in Fig. 4 the difference in heat flux anomaly in each season when the normalized MCA–SIC index is below and greater than 1 standard deviation. Positive values indicate above-normal heat transfer from the ocean to the atmosphere when the MCA–SIC index is below

### Table 1

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### Table 2

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normal. The largest heat flux changes occur near the ice edge, where they are much larger than those over the open ocean (70–80 W m\(^{-2}\) as opposed to 20 W m\(^{-2}\)). Positive upward heat flux anomalies are associated with reduced ice cover in the Weddell–Ross Seas followed by a negative response downstream. Negative heat flux anomalies are associated with enhanced ice concentration in the Bellingshausen–Amundsen Seas and South Indian Ocean, followed by a positive response downstream. For comparison, climatological upward heat flux for June–November (JJASON) is shown in Fig. 4e. Near the ice edge, the upward surface heat flux anomalies are of similar magnitude to that of climatological heat fluxes. For these areas, the sensible heat flux anomalies are much larger than the latent heat flux anomalies; and the latter are \(\sim 20\%–30\%\) of the former.

It has been found that strong transient eddy forcing is important for explaining the extratropical response to midlatitude SST, sea ice, and snow anomalies (e.g., Deser et al. 2007; Fletcher et al. 2009). Deser et al. (2007) show that the transition of the response from a thermally direct response to a reduction in sea ice cover in the Greenland Sea and an extension of the ice edge in the Labrador Sea in the onset phase to a North Atlantic Oscillation (NAO)-like response involves strong eddy–mean flow interactions. Here, the SIC-related storm track changes are explored by computing the bandpass-filtered 300-hPa transient eddy kinetic energy \(\frac{(u^2 + v^2)}{2}\), associated with 2–8-day-filtered fluctuations and 300-hPa eddy momentum flux \(u'v'\), associated with 2–8-day-filtered fluctuations), 700-hPa northward transient eddy heat flux \(u'T\), associated with 2–8-day-filtered fluctuations), and their simultaneous and lagged regression patterns onto the normalized MCA–SIC time series associated with the homogeneous sea ice patterns in Fig. 2. In contrast to the contemporaneous regression, the lagged regressions can be interpreted as the response of the eddies to anomalies in SIC.

The storm-track response for ASO (Fig. 5a) is indicative of a general increased storm activity in the high-latitude regions and decreased storm activity in the midlatitude oceans. Similar changes are very persistent and robust in the regression at lag of +1 and +2 months (storm tracks lagging) (Figs. 5b,c). The above shifts in storm activity are companioned with the poleward shift of the westerly jet (not shown). The response patterns of the high-pass-filtered 700-hPa northward transient eddy heat flux \(u'T\), shown in Figs. 5d and 5f, are such that it is negative over the area of a positive storm activity
anomaly and positive in the midlatitude and subtropical oceans, where storm-track activity has decreased. Thus, the responses in the transient meridional energy fluxes support the changes in storm-track activity, which were found in previous studies (e.g., Karoly 1990; Randel 1989).

The ADP-like SIC anomalies in Fig. 2c are followed by anomalous poleward (equatorward) eddy momentum fluxes in the high latitude (midlatitude) (Figs. 5g–i). The associated regions of anomalous eddy momentum flux divergence and convergence reflect the forcing of the extratropical zonal flow by the eddies (e.g., Limpasuvan and Hartmann 1999, 2000; Lorenz and Hartmann 2003). At a lag of 1–2 months, the upper troposphere in the high latitude (midlatitude) is marked by anomalous convergence (divergence) of the eddy momentum flux. Poleward eddy momentum fluxes interact with the zonal mean flow to sustain latitudinal displacements of the midlatitude westerlies associated with the AAO (Limpasuvan and Hartmann 2000; Rashid and Simmonds 2005). Similar features are found for other seasons (not shown).

To document the atmospheric circulation anomalies that follow ADP-like SIC, we have generated the regression maps of the OND anomaly fields of geopotential height at 850 and 250 hPa onto the MCA–SIC time series at lag $2_2$ when $Z500$ is fixed at OND. It is found that the regression patterns (Figs. 6a,b) are very similar to Fig. 2c, indicating that the atmospheric signal is primarily equivalent barotropic. It is consistent with the barotropic atmospheric responses to the reduced SIC in the Arctic seas in the observational studies by Wu and Zhang (2010) and in the model simulations by Deser et al. (2007).

We also attempted to determine which part of the ASO SIC anomaly had the strongest influence on the atmosphere by considering the averaged SIC anomaly in

![Fig. 4. (a)–(d) Composite response in JJA to SON surface energy heat flux of negative minus positive months of the MCA–SIC time series associated with homogeneous SIC pattern in Figs. 2a–d from JJA to SON seasons exceeding 1.0 standard deviation in absolute values. (e) Climatology JJASON mean surface energy flux. Contour interval is 20 W m$^{-2}$. Negative contours are dashed and the zero line is omitted.](image-url)
FIG. 5. Simultaneous and lagged regression maps of (a)–(c) bandpass-filtered 300-hPa transient eddy kinetic energy \(\left(\frac{u^2 + v^2}{2}\right)\), (d)–(f) 700-hPa transient eddy meridional heat flux \(\left(\bar{v}'T'\right)\), and (g)–(i) 300-hPa transient eddy momentum flux \(\left(\bar{u}'\bar{v}'\right)\) to the MCA–SIC time series at lag \(-2\) when Z500 is fixed at OND. Contour interval is 2.5 m\(^2\) s\(^{-2}\) for \(\left(\frac{u^2 + v^2}{2}\right)\), 0.25 K m s\(^{-1}\) for \(\bar{v}'T'\), and 2 m\(^2\) s\(^{-2}\) for \(\bar{u}'\bar{v}'\). Negative contours are dashed and the zero line is omitted. The thin lines indicate the estimated 5% significance level.
boxes centered on the two main centers of action of the SIC anomaly pattern—namely, 63°–57°S, 60°–20°W (South Pacific center) and 68°–62°S, 150°–90°W (South Atlantic center). Projecting the OND Z500 anomaly on the two SIC time series gave in each case a Z500 anomaly pattern (Figs. 7a,b), which resembled the AAO-like patterns. Hence, both centers of action seem to influence the atmosphere. The maximum atmospheric response is about 15 m for both the South Pacific center and the South Atlantic center. This is substantially smaller than in Fig. 2c, which is not surprising since the sea ice estimates are based on box-averaged values rather than the maximum and the MCA maximizes the covariance. Note that this calculation is independent of the MCA, except for the choice of season and box location, suggesting that the MCA results are robust.

4. Summary

We have found a significant covariance between SIC anomalies and the springtime atmospheric circulation at lags when SIC anomalies lead up to four months, which suggests an impact of the late fall–winter SIC anomalies on the springtime atmosphere. Such influences are not associated with trends and ENSO forcing. The link between the springtime atmosphere and sea ice up to four months earlier seems to stem from the remarkable persistence of ADP-like sea ice anomalies. Such persistence may be explained by dynamic and thermodynamic processes, but further study might be needed.

Our results suggest that ADP-like sea ice anomalies at late fall–winter are more efficient at exciting the atmospheric response than at other seasons. This may be due to the seasonality of ADP-like sea ice anomalies. EOF analysis was performed on SIC anomalies binned into groups of three months. The most robust feature found in the EOF analysis is the persistence of the austral winter ADP pattern discussed in Yuan and Martinson (2001) from May–June (MJJ) to OND, which is consistent with the MCA patterns of SIC in Fig. 2. The first leading EOF (EOF1) for each set of three successive months from austral late spring to fall [from November–January (NDJ) to March–May (MAM)] also exhibits ADP-like patterns with anomalies of one sign in the Pacific and opposite sign in the Atlantic, but the action centers retreat to much smaller area in the high-latitude. The influence of austral summer ADP-like sea ice variability on the atmosphere is not detected, which is due to the low signal-to-noise ratio at monthly to interannual time scales. This indicates that there is significant sensitivity of the atmospheric response in the SH to the calendar month.

The AAO-like atmospheric signal is hemispheric in extent and primarily equivalent barotropic through the troposphere. Therefore, large seasonal forecast skill should result in austral spring from the lagged relations established here. Our results support the modeling studies by Raphael (2003) and Lachlan-Cope (2005), which found the atmosphere is sensitive to Antarctic sea ice extent anomalies. The feedback of ADP-like SIC
anomalies on the atmospheric circulation involves the diabatic heating owing to the air–sea heat exchanges and transient eddy flux changes. It should also involve the nonlinear transient eddy vorticity fluxes found in Deser et al. (2007) for the atmospheric response to arctic sea ice but in a way that remains to be established. Modeling studies are needed to investigate the growth and maintenance of the AAO-like atmospheric response to late fall–winter ADP-like SIC anomalies found here.

Note that we ignore the possible uncertainties of the 500-mb geopotential height in the NCEP–NCAR reanalysis that used a frozen state-of-the-art global data assimilation system to generate the atmospheric datasets (Kalnay et al. 1996). We have repeated the same analyses in this study using the atmospheric data between 1979 and 2002 from the 40-yr European Centre for Medium-Range Weather Forecasts (ECMWF) Re-Analysis (ERA-40; Uppala et al. 2005) and found that the main conclusions are very robust to different reanalysis datasets.

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REFERENCES


