INTERNATIONAL JOURNAL OF CLIMATOLOGY
Int. J. Climatol. **24**: 1–31 (2004)
Published online in Wiley InterScience (www.interscience.wiley.com). DOI: 10.1002/joc.965

REVIEW

THE EL NIÑO–SOUTHERN OSCILLATION AND ANTARCTICA

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Received 18 April 2002
Revised 20 August 2003
Accepted 20 August 2003

ABSTRACT

This paper reviews our understanding of how the effects of the El Niño–southern oscillation (ENSO) might be transmitted from the tropical Pacific Ocean to the Antarctic, and examines the evidence for such signals in the Antarctic meteorological, sea ice, ice core and biological records. Many scientific disciples concerned with the Antarctic require an understanding of how the climatic conditions in the tropical and mid-latitude regions affect the Antarctic, and it is hoped that this review will aid their work.

The most pronounced signals of ENSO are found over the southeast Pacific as a result of a climatological Rossby wave train that gives positive (negative) height anomalies over the Amundsen–Bellingshausen Sea during El Niño (La Niña) events. However, the extra-tropical signature can sometimes show a high degree of variability between events in this area. In West Antarctica, links between ENSO and precipitation have shown variability on the decadal time scale. Across the continent itself, it is even more difficult to relate meteorological conditions to ENSO, yet analyses of the long meteorological records from the stations do indicate a distinct switch in sign of the pressure anomalies from positive to negative across the minimum in the southern oscillation index.

The oceanic signals of ENSO around the Antarctic are less clear, but it has been suggested that the Antarctic circumpolar wave could be forced by the phenomenon.

Ice-core data offer the potential to help in understanding the long-term relationship between ENSO and the climate of the Antarctic, but there are difficulties because of the need to smooth the ice-core data to overcome the mixing of snow on the surface. Nevertheless, analysis of methylsulphonic acid in a South Pole core has shown high variability on ENSO time scales.

It is clear that some evidence of ENSO can be found in the Antarctic meteorological and ice-core records; however, many of the relationships tend not to be stable with time, and we currently have a poor understanding of the transfer functions by which such signals arrive at the Antarctic from the tropical Pacific. Copyright © 2004 Royal Meteorological Society.

KEY WORDS: ENSO; Antarctica; climate variability

1. INTRODUCTION

The El Niño–southern oscillation (ENSO) is the largest climatic cycle on decadal and sub-decadal time scales (Diaz and Markgraf, 1992) and it has a profound effect on not only the weather and oceanic conditions across the tropical Pacific, where the ENSO has its origins, but also in regions far removed from the Pacific basin (Kiladis and Diaz, 1989). ENSO has a very direct influence on weather conditions in some mid-latitude areas, and its effects have been linked to changes in storm tracks across North America, producing significant seasonal anomalies in temperature and precipitation (Ropelewski and Halpert, 1987). It is also thought to be a factor behind major drought events in the African Sahel region (Ward, 1998). Not surprisingly, most research into the effects of ENSO has been concerned with tropical and mid-latitude areas, where there are

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long *in situ* meteorological records and high-quality atmospheric analyses extending back into the early part of the 20th century or earlier. These long records increase the possibility of finding *statistically significant* relationships between ENSO and climatic parameters in remote locations — what are now frequently referred to as teleconnections (Mo and White, 1985). However, finding such links is much more difficult in the Antarctic because of the relatively short length of the *in situ* meteorological records (about 50 years at best) and the poor quality of the analyses over the ocean in the pre-satellite era. The larger inter-annual variability of the climate of the Antarctic compared with lower latitude areas (King and Turner, 1997) also adds to the problem of finding signals of ENSO on the continent. Yet it is very important to understand if and how ENSO is affecting the present-day Antarctic climate so that we can interpret correctly the temperature and snow accumulation records derived from Antarctic ice cores and to determine whether rapid climatic changes observed in recent decades (e.g. the warming on the western side of the Antarctic Peninsula (King, 1994)) are affected by the changing nature of ENSO.

This paper reviews our current understanding of how signals of ENSO might be transmitted to the Antarctic and considers the degree to which we have been able to relate changes in ENSO to the various forms of data available from the continent. It is hoped that it will provide an up-to-date review of the possible effects of ENSO on the Antarctic environment for workers in the many scientific disciplines that have an interest in ENSO in the Antarctic, such as biologists, ice-core chemists, glaciologists and oceanographers, as well as meteorologists/climatologists. There is now quite a large body of literature on ENSO and the Antarctic and it is not possible to refer to all the papers on this subject, but I have tried to reference all the major papers that have been published and to give a flavour of the range of disciplines that ENSO affects.

A very brief outline of ENSO and the measures of the ENSO cycle are presented in the Section 2 before a discussion of the propagation of signals from the tropical to the high-latitude areas of the South Pacific. This is followed by assessments of the degree to which researchers have been able to relate ENSO variability to the meteorological data, Antarctic sea ice observations, the oceanic records, ice cores and the various biological records. In the final section I attempt to bring together the results of the various investigations into a picture of our current understanding of ENSO teleconnections in the Antarctic and consider the requirements for future work.

2. THE ENSO PHENOMENON OF THE TROPICAL PACIFIC

ENSO has received a great deal of attention in the literature over the last few years, and there are many papers discussing the nature of the phenomenon in the tropical Pacific (e.g. Philander, 1983, 1989; Rasmussen and Wallace, 1983; Philander and Rasmussen, 1985; Bigg, 1990; Trenberth, 1997). In this section we provide only a few brief comments on the various measures of ENSO activity.

The evolution of the ENSO cycle can be measured by a number of different indices. Here, we use the southern oscillation index (SOI), which is the twice-normalized difference in surface pressure between Tahiti and Darwin, Australia (Parker, 1983). The values of the SOI since 1950 are shown in Figure 1.

Sea-surface temperatures (SSTs) in various parts of the Pacific have been proposed as useful indicators of ENSO, with values in the Niño 3.4 region (150–90 °W and 5 °N–5 °S) being regarded as an important indicator of the phase of the cycle (Trenberth and Hoar, 1996). Here, SSTs can rise above 27 °C during ENSO 'warm' events (Trenberth, 1997).

Over the years there has been extensive discussion over what constitutes an El Niño event, and ideas have changed with time. Trenberth (1997) suggested that an El Niño event has occurred if the 5 month running mean anomaly of SST in the Niño 3.4 region exceeds 0.4 °C for 6 months or longer. If this definition is adopted then it suggests that El Niños occur on 31% of occasions, with La Niñas occurring during a further 23% of the time. During the remaining 56% of the time the circulation is in neutral conditions.

3. THE TRANSMISSION OF THE ENSO SIGNAL TO HIGHER LATITUDES

Using station data and series of operational numerical analyses, several workers since the 1970s have shown that statistically significant teleconnections exist between atmospheric conditions in the tropical Pacific and

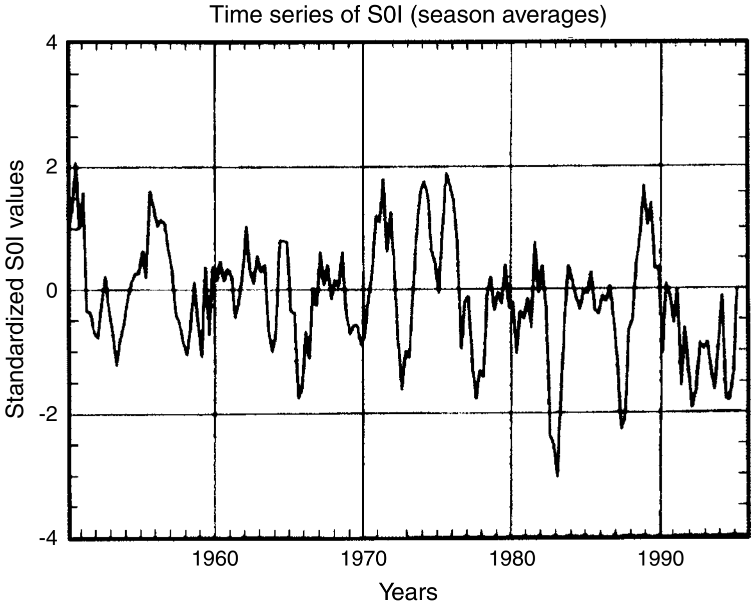

Figure 1. The SOI since 1950. The units are standard deviations from the climatological monthly mean sea-level pressure (MSLP) differences between Tahiti and Darwin. After Bigg (1990)

those at extra-tropical latitude areas of the Southern Hemisphere. Trenberth (1975a,b) found highly significant relationships between the southern oscillation and indices of the circulation in the middle- and high-latitude areas of the Australian sector for the period 1959–72. That work was then extended to cover more easterly locations (Trenberth, 1976) and suggested east–west circulations coupling the South Pacific high and the South Pacific convergence zone (SPCZ). Van Loon and Shea (1987) noted changes in the annual cycle of MSLP between the year of a warm event ($year_0$) and the previous year ($year_{-1}$) over lower and mid-latitudes, and over much of the Southern Hemisphere. A further analysis, using the Australian fields, was undertaken by Mo and White (1985) for the period 1972–80. They found strong teleconnections between monthly anomalies in 500 hPa height during the summer season and an index of the southern oscillation. The correlations were largest across the subtropical Pacific, but extended into the Southern Ocean. Other studies have also emphasized the links between ENSO and the temporal evolution of MSLP anomalies over the Southern Ocean and the Antarctic coastal region, and identified a decadal propagation of pressure anomalies between the south and north polar regions (Krishnamurti *et al.*, 1986). They traced the pressure reversals in the El Niño years of 1965 and 1969 back to the Antarctic, suggesting a possible role for the high-latitude areas of the Southern Hemisphere in modulating ENSO. In addition, they found that 30% of the pressure variance close to the Antarctic was on ENSO time scales of 30–50 months.

These investigations provided compelling evidence for a signal of ENSO being transmitted to high-latitude areas of the Southern Hemisphere, although a theoretical basis for such linkages was not available. However, Hoskins and Karoly (1981), in their theoretical investigation, showed that an area of deep convection close to the equator can act as a generator of Rossby waves through the vorticity generated by diabatic heating. These Rossby wave trains travel polewards in both hemispheres and provide a means for the establishment of teleconnections between ENSO and the climates of the mid-latitude areas. As discussed by Held *et al.* (1989), the tropically forced wave train need only have modest amplitude to displace the downstream portion of the oceanic storm track so that relatively small changes in tropical SSTs may have a large extra-tropical response.

As the main area of convection in the tropical Pacific moves from over Indonesia during La Niña events to close to the date line during the El Niño phase so the mid-latitude response to ENSO at a particular location will change. In the North Pacific there is a clear signal of ENSO warm events transmitted northwards via

Rossby wave trains known as the Pacific North American (PNA) pattern (Rasmusson, 1991; Hoerling and Kumar, 1997); however, the evidence for ENSO in the South Pacific is not as clear as north of the equator. Karoly (1989) carried out a superposed epoch analysis of South Pacific upper-air fields and showed that a wave train was present during the austral winter in 'warm' events, which became known as the Pacific South American (PSA) connection. This wave train affects the synoptic conditions across the southern part of South America (Ruttlant and Fuenzalida, 1991), as well as the Antarctic Peninsula (Harangozo, 2000). Karoly's work covered only three El Niño events that occurred during the period 1972–83, which raises the question of how stable the results are in the long term. Figure 2 shows the upper tropospheric height anomalies during winter for the three El Niño events examined by Karoly. The Rossby wave train can be seen clearly as an arc of high–low–high anomalies extending in a southeasterly direction from the area of increased convective activity in the central Pacific towards the Antarctic. Figure 3 shows comparable winter-season 500 hPa height anomaly maps, but derived separately for both El Niño and La Niña events, and based on all events during the period 1979–99. This period was chosen since high-quality fields are available from the European Centre for Medium-Range Weather Forecasts reanalysis project (Gibson *et al.*, 1996). The mean anomaly for El Niño events is very similar to that found by Karoly, with an anticyclonic anomaly over the Amundsen–Bellingshausen Sea (ABS) and a slightly weaker, cyclonic anomaly centred on 48 °S, 140 °W. The anomalies found during La Niña events are in almost exactly the same locations as those for El Niño, but of opposite sign.

Further evidence for the varying nature of the wave train between different phases of the ENSO cycle was found by Houseago *et al.* (1998), who carried out a Hovmöller analysis on the Australian daily analyses for the period 1973–94. They found a propagation of negative height and temperature anomalies from the subtropics to high-latitude areas of the South Pacific during 'cold' events up to the peak of the event, with a persistence of the anomaly for about a year at subpolar latitudes. For all 'warm' events, except 1982, there

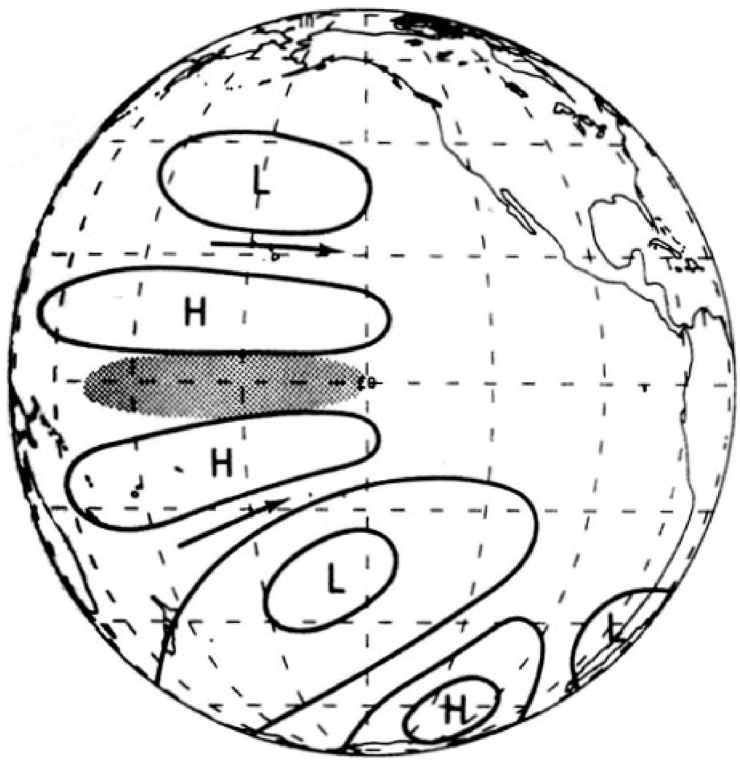

Figure 2. Schematic illustration of the pattern of upper tropospheric height anomalies over the Pacific Ocean during the early stage of an ENSO event in the Southern Hemisphere winter (June–August). The stippling shows the region of enhanced convection over the central equatorial Pacific and the arrows indicate the westerly wind anomalies in the jet streams. From Karoly (1989)

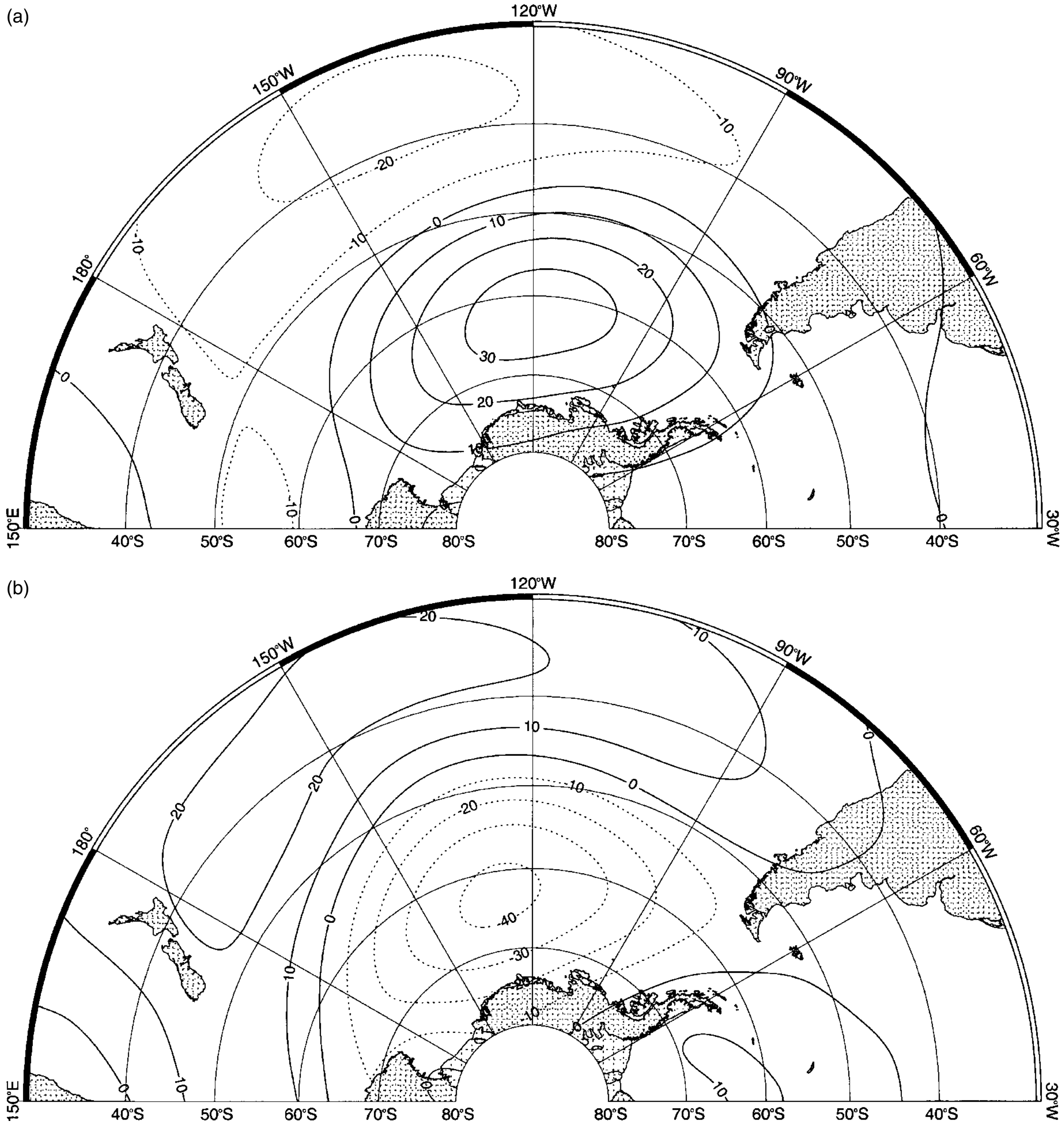

Figure 3. Winter-season (June–August) 500 hPa height anomalies across the South Pacific during (a) eight El Niño and (b) six La Niña cases in the period 1968–99. The figures were derived from NCEP–NCAR reanalysis data

was a propagation of positive height anomalies along the same track up to the peak of the event, along with a strong meridional contrast between subpolar and subtropical pressure and temperature anomalies. The unusual nature of the 1982 'warm' event will be discussed later, but the Houseago *et al.* (1998) study showed that, instead of the usual wave train of positive anomalies poleward, there was a train of negative anomalies extended *equatorwards* in this event.

A continuing problem in understanding the ENSO phenomenon and how its signals are transferred to higher latitudes is the large differences between individual events and the effects of varying atmospheric circulation conditions outside the tropics. At tropical latitudes, the same SSTs in different years may not produce the same deep, tropical convection, because of varying vertical wind shear or differences in the atmospheric convergence field (Graham and Barnett, 1987; Chang, 1993). This suggests that the climatological links between tropical SSTs and Rossby wave generation may sometimes breakdown. Harangozo (personal communication) has suggested that additional information on variations in tropical convection and, therefore, the generation of Rossby waves can be inferred from *temporal variations* in SSTs, and this suggestion will be considered in the following. At higher latitudes the response via Rossby wave activity can be modulated by variations in the westerlies, further complicating the interpretation of Antarctic *in situ* observations during different phases of ENSO.

The variability in the nature of the high-latitude response to ENSO events was further highlighted by the work of Houseago-Stokes and McGregor (2000) in their investigation of 500 hPa height anomalies during different phases of ENSO. They examined the first harmonic of 500 hPa height anomalies based on 24 month periods centred on 'warm' and 'cold' events using Australian analyses covering the period 1973–94. They found a linkage of pressure anomalies close to 60 °E across the Indian Ocean–East Antarctic region during 'cold' events, but no clear signal at this longitude or close to the Antarctic Peninsula during the 'warm' phase. Again, this highlighted the variability in the high-latitude atmospheric circulation between 'warm' events.

White and Peterson (1996) suggested that the ocean circulation could play a role in the transmission of ENSO signals to high southern latitudes via the Antarctic Circumpolar Wave (ACW). They suggested that the ACW in SST originated in the western subtropical South Pacific, and then spread south and east into the Southern Ocean, where subsequent eastward propagation took place via the Antarctic Circumpolar Current (ACC). The SST anomalies develop in response to ENSO events along the equator and move south in parallel with MSLP anomalies. Peterson and White (1998) then investigated in more detail the links between ENSO and the ACW. That investigation, via a case study of the 1982–94 period, confirmed the important role of SST anomalies in the western subtropical South Pacific area in generation of the ACW signal and allowed them to gain insight into the progression of the signal around the globe. A large part of the tropical SST anomaly was found to be advected northwards into the South Atlantic and Indian Oceans, eventually reaching the tropical areas of each basin approximately 6–8 years after their appearance in the equatorial Pacific (Figure 4). There is, therefore, strong evidence for a *northwards* propagation of ENSO signals from the edge of the Antarctic region, following the southbound transfer of the ENSO signal to the ACC and propagation eastwards. However, it is not known to what extent the ACW signal is stable in the long term, since the existing data sets are relatively short.

In a recent study, Cai and Baines (2001) carried out further work into the relationship between the ACW and ENSO via an empirical orthogonal function (EOF) and complex EOF analysis of Southern Ocean SSTs. They found that the observed ACW could be considered as the sum of two linearly independent complex EOFs, with zonal wavenumber 2 and 3 patterns. During the 1981–97 period the wavenumber 2 pattern was dominant and they demonstrated that the wavenumber 2 component was forced by the PSA teleconnection, described earlier. They suggested that the atmospheric wind stress and heat flux anomalies induced by the PSA wave train caused SST changes of around 1 °C over large areas of the South Pacific, which were then advected eastwards by the ACC. So they suggest that the principal component of the ACW is initiated by atmospheric teleconnections via the PSA, propagated by the ACC and maintained by air–sea interactions.

Further work on the links between various ocean parameters and ENSO was carried out by Yuan and Martinson (2000), who found that the highest correlation was with the extra-tropical Indian Ocean SSTs. The structure of the correlation revealed a coherent propagating pattern in which the area of peak correlation migrated to the east at a rate of approximately 45° per year. The eastward migration speed was found to be consistent with the phase propagation speed of the ACW (White and Peterson, 1996). However, it was noted that the propagating pattern was discontinuous in the Indian Ocean sector of the Antarctic, so that the ACW may be more representative of subpolar gyre variations.

4. EVIDENCE FOR SIGNALS OF ENSO IN THE ANTARCTIC METEOROLOGICAL RECORDS

A recurring problem in all Antarctic meteorological research is the relatively short record that is available, which extends back only about 50 years on the Antarctic Peninsula and much less in many other locations.

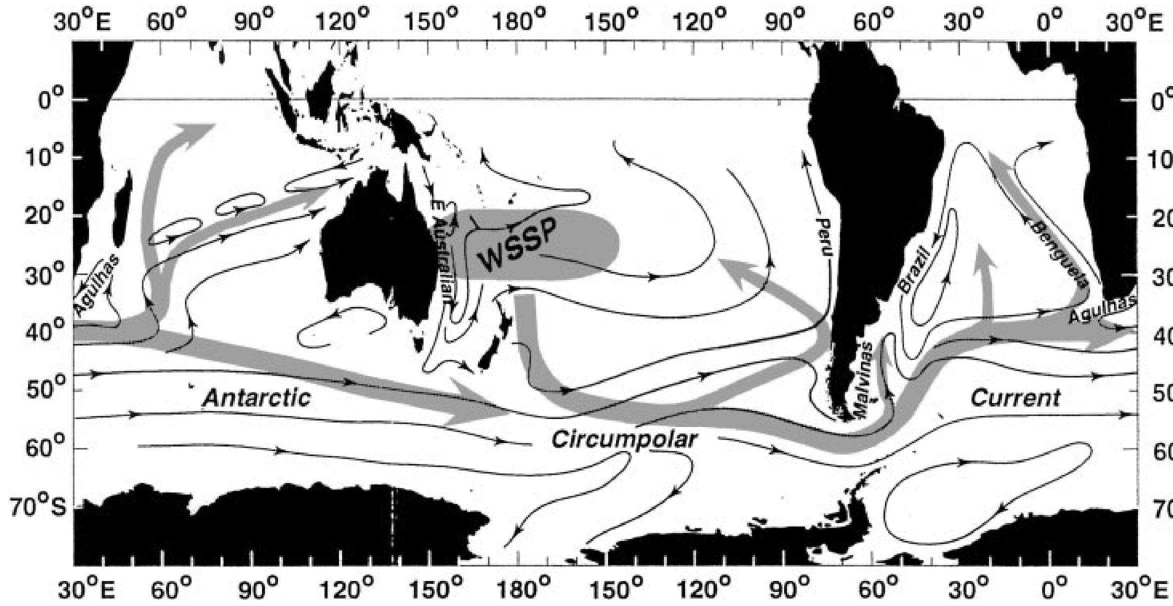

Figure 4. Schematic representation of the principal off-equatorial spreading routes (shaded lines) of inter-annual anomalies in SST from the source region in the western subtropical South Pacific (WSSP; shaded oval). Solid lines show the mean geostrophic circulation patterns at the sea surface. From Peterson and White (1998)

Table I. The El Niño and La Niña events since 1968 (after Trenberth (1997))

El Niño events		La Niña events	
Beginning	End	Beginning	End
Sep 1969	Mar 1970	Jul 1970	Jan 1972
Apr 1972	Mar 1973	Jun 1973	Apr 1976
Aug 1976	Jan 1978	Sep 1984	Jun 1985
Oct 1979	Apr 1980	May 1988	Jun 1989
Apr 1982	Jul 1983	Sep 1995	Mar 1996
Aug 1986	Feb 1988	Jul 1998	Dec 1999
Mar 1991	Mar 1995		
Apr 1997	Apr 1998		

In addition, the reanalysis fields, although starting in the 1950s in the case of the National Centers for Environmental Prediction (NCEP) reanalysis (Kalnay *et al.*, 1996), are poor in the Antarctic before 1968, since few Antarctic observations were used in the reanalysis process before this date (Marshall and Harangozo, 2000). A list of the El Niño and La Nina events, according to the Trenberth (1997) definition, that have occurred since 1968 is given in Table I.

4.1. The Bellingshausen Sea–Antarctic Peninsula region

Since ENSO is a phenomenon of the tropical Pacific the largest signals of the cycle can be expected in the Pacific sector of the Antarctic. This was confirmed by Kwok and Comiso (2002), who investigated the nature of climate anomalies during different phases of the SOI using NCEP–National Center for Atmospheric Research (NCAR) reanalysis fields.

As discussed above, the PSA pattern of upper-air height and surface pressure anomalies will result in higher pressures (greater ridging or blocking) to the west of the Antarctic Peninsula during the winter season within ‘warm’ events as a result of the Rossby wave train (Sinclair, 1996; van Loon and Shea, 1987; Renwick, 1998). Blocking episodes around the Antarctic are rather short-lived (Jones and Simmonds, 1994), with individual anticyclones tending to exist for less than about 4 days. So the anticyclonic anomaly is often manifested as a general lack of cyclones over the ABS rather than a period of marked blocking.

The net result of positive MSLP anomalies over the Bellingshausen Sea region is a circulation pattern that gives a greater southerly component to the winds over the peninsula during ‘warm’ events. It could be expected that such a synoptic pattern would result in negative temperature anomalies at the research stations on the western side of the peninsula. However, the near-surface temperatures here are very heavily influenced by the amount of sea ice present over the eastern Bellingshausen Sea (Jacobs and Comiso, 1997) and the sea ice extent is affected by atmospheric and oceanic factors in a highly non-linear way, so that in the period since 1968 there is no statistically significant correlation between the SOI and the temperature anomalies on the western side of the peninsula. This can be seen from the annual mean temperatures for Faraday station and the annual mean SOI shown in Figure 5 (correlation 0.3). However, Kwok and Comiso (2002) found that surface temperatures over the eastern Bellingshausen Sea from the NCEP–NCAR re-analyses were positively correlated with the SOI, i.e. temperatures were colder than normal during El Niño events.

The El Niño event of 1982–83, which was the most pronounced ‘warm’ event of the 20th century, had a rather anomalous signature in the Antarctic. During this event the ABS blocking pattern did not appear, despite the SOI having a value of −3 (Houseago *et al.*, 1998). It has been suggested that the variable circulation response during different ‘warm’ events could be a result of modulation of the Rossby wave propagation by the mean zonal circulation (Renwick, 1998). However, Harangozo (personal communication) suggests that the variability in the westerlies is a result of changes in the climate of the tropical Pacific, especially tropical convection, a result supported by Mo and Higgins (1997).

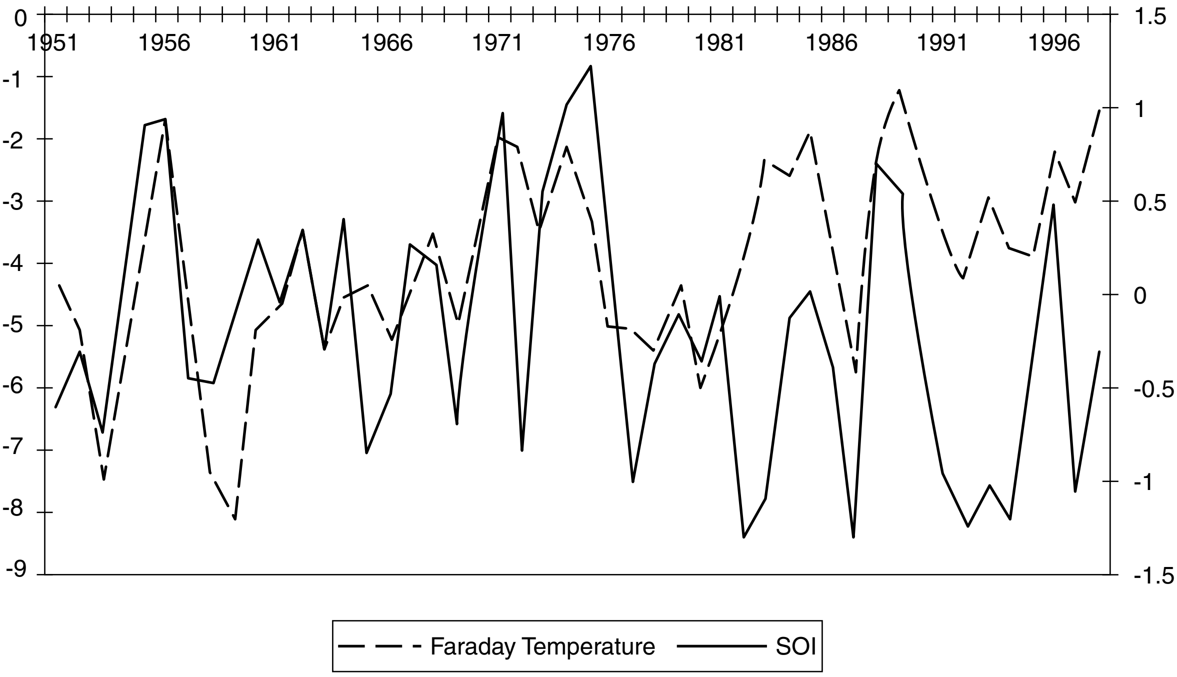

Figure 5. Annual mean Faraday temperatures and annual mean SOI for the period 1950–98

Harangozo (personal communication) has found statistically significant correlations between MSLP in the southeast Pacific and SSTs in the tropical Pacific. His analysis concentrated on 16 austral winters (June–August) when central Pacific SSTs were in transition (essentially when, during a 10 month period prior to an SST maximum or minimum, the SST anomaly change was greater than 1 °C) during the period 1973–93. His analysis showed that SSTs in the Niño 3.4 area correlated well with MSLP around the Antarctic Peninsula with statistically significant positive (negative) anomalies over the southeast Pacific Ocean (east of the Drake Passage). This can be seen in Figure 6, which shows the zero lag winter cross-correlation field for Niño 3.4 SST minus MSLP differences for the 16 winters when central tropical Pacific SSTs were in transition during 1973–93. In Figure 6, the areas where a statistically significant correlation at better than the 95% level have been found are shaded. The dipole of high correlation values on either side of the Antarctic Peninsula suggest higher (lower) MSLP values on the western (eastern) side of the area, which is consistent with the model of the Rossby wave train.

The study by Harangozo (personal communication) also concluded that central tropical Pacific SST values are not a reliable guide to the atmospheric circulation in the high-latitude areas of the South Pacific, so that pressures over the Bellingshausen Sea are not always high during all ENSO warm events, e.g. 1982; this result was confirmed by Renwick (1998), who also found little blocking in the 1982 winter. He also found

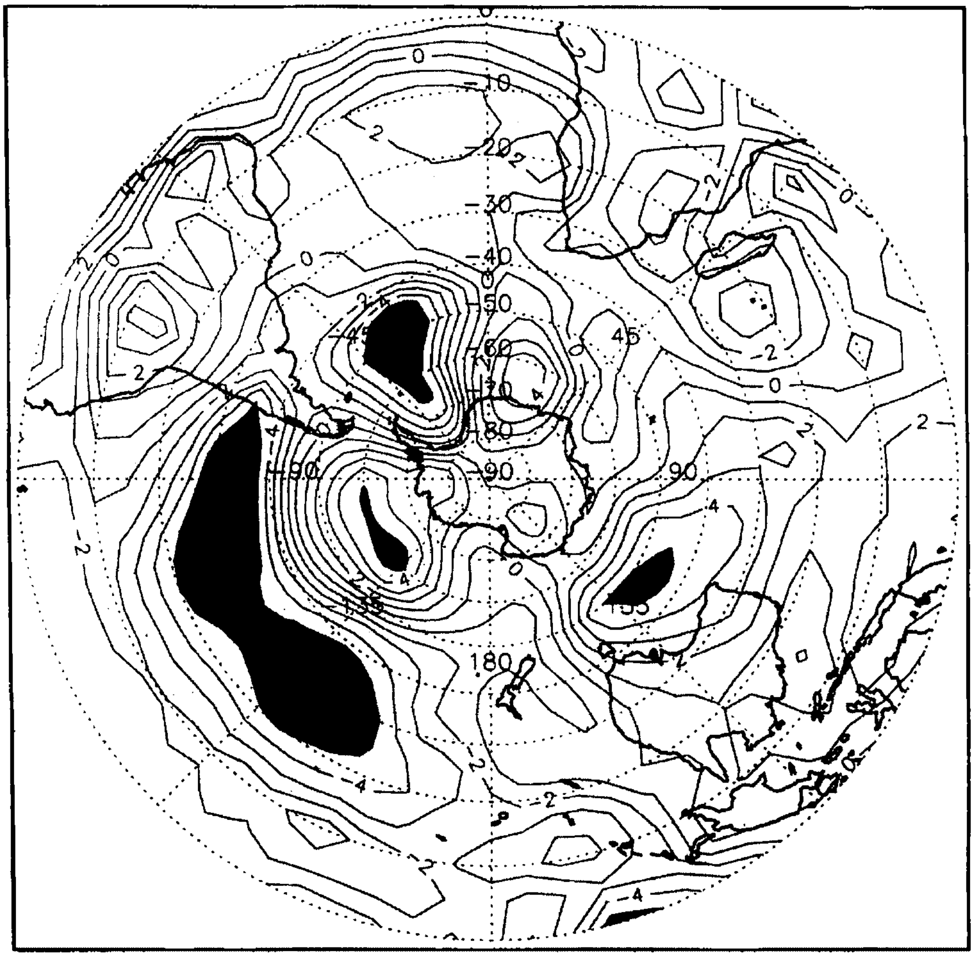

Figure 6. Zero lag winter (June–August) cross-correlation ($r \times 10$) field for Niño ΔSST – MSLP for 16 austral winters when central tropical Pacific SSTs were in transition in the 1973–93 period. Light shading denotes statistically significant correlations at ≤5% level (using a Student's t-test). From Harangozo (2000)

that winters with the least SST cooling in the tropical Pacific were associated with reduced westerlies in the central-eastern Pacific and greater southerlies to the west of the peninsula.

4.2. The split jet region of the western Pacific

One of the most important atmospheric features of the western Pacific is the split nature of the jets at the longitude of New Zealand, where the subtropical (polar front) jet is located close to 30 °S (60 °S). The split jet is not present year-round, but begins to develop in March and is most pronounced in the winter and early spring (Trenberth, 1975a). The reason for this particular form of the jets is not fully understood, although the fact that the main landmass of the Antarctic is displaced from the South Pole may play a role (James, 1988). In addition, prior to the 1982 warm event, Mo *et al*. (1987) found that cold air outbreaks from Antarctica were partially responsible for maintaining the block at mid-tropospheric levels, and thus a split jet. Some research has suggested that signals of ENSO may be propagated southwards via this feature (Kitoh, 1994), with the variability of the jets affecting the tracks of depressions and blocking (Chen *et al*., 1996; Sinclair, 1996).

Chen *et al*. (1996) suggested that the subtropical jet strengthens during warm ENSO events, with the polar front strengthening during cold or normal stages of the cycle. This is consistent with Karoly (1989), who found that across the central South Pacific there is often a weakening of the westerlies during warm events. However, this was not the case during other warm-event winters, such as 1977 and 1979. Harangozo (2000) concludes that the westerlies of the central South Pacific are strongly modulated by Rossby wave activity but that they are unlikely to be the same in individual warm events.

In the Chen *et al*. (1996) study the more vigorous polar front jet (PFJ) was evident in terms of greater cyclonic activity at the latitude of the circumpolar trough with a deeper Amundsen Sea low. This is in agreement with the work of Shiotani (1990), who showed that, on seasonal time scales, when geopotential heights at high latitude are low and the split jet is pronounced, transient eddy activity near the Antarctic coast is vigorous near the Ross Ice Shelf.

The Chen *et al*. (1996) investigation analysed the momentum budget during the event to understand how the upper tropospheric flow and jets over the South Pacific evolved, and found that, during the warm event, the strong subtropical jet (STJ) was associated with advection of the mean flow momentum from the Australian region, which was approximately balanced by a large negative ageostrophic term. The PFJ was primarily associated with eddy momentum convergence, which was partially counterbalanced by the ageostrophic term. During the 1989 cold event the weak STJ was related to an increasingly negative ageostrophic term and a less positive mean flow momentum convergence. The stronger PFJ was associated with an increase in the convergence of eddy momentum flux that was mainly composed of 2.5–6 day poleward momentum transport from mid-latitudes and 7–30 day equatorward momentum transport from high latitudes. They also found that the impacts of eddy stress on the STJ and the mean momentum divergence on the PFJ in the sector were small. The anti-correlation between the strength of the STJ and PFJ during the 1987 warm and 1989 cold events can be appreciated via the schematic meridional cross-section shown in Figure 7.

The Chen *et al*. (1996) study only considered the period 1986–89, and the situation is different if longer periods are considered. Figure 8 shows the 300 hPa anomalies for the zonal wind component for all warm and cold events over the period 1968–97. The light and dark shading indicates statistical significance at the 95% and 99% levels respectively. There is a clear reversal of the anomalies in the STJ with, as noted by Chen *et al*. (1996), a stronger jet during warm events and weaker conditions in the cold event. However, conditions are less clear at the latitude of the PFJ. During El Niño conditions there is evidence of an overall strengthening of the jet south of New Zealand, with this being statistically significant at the 95% level to the west of 165 °E. For La Niña conditions the negative anomalies are very small and not statistically significant at latitudes south of 50 °S. Across the ABS there are no statistically significant links between the zonal wind and the phase of ENSO. One final point to note from Figure 8 is the statistically significant anomalies close to 40 °S over New Zealand and the oceanic region to the east. This is the region of the climatological block and split jet, where zonal winds are usually weak between the STJ and PFJ. Figure 8 shows that there is a significant change in the nature of the block through the ENSO cycle, with a stronger (weaker) block as a

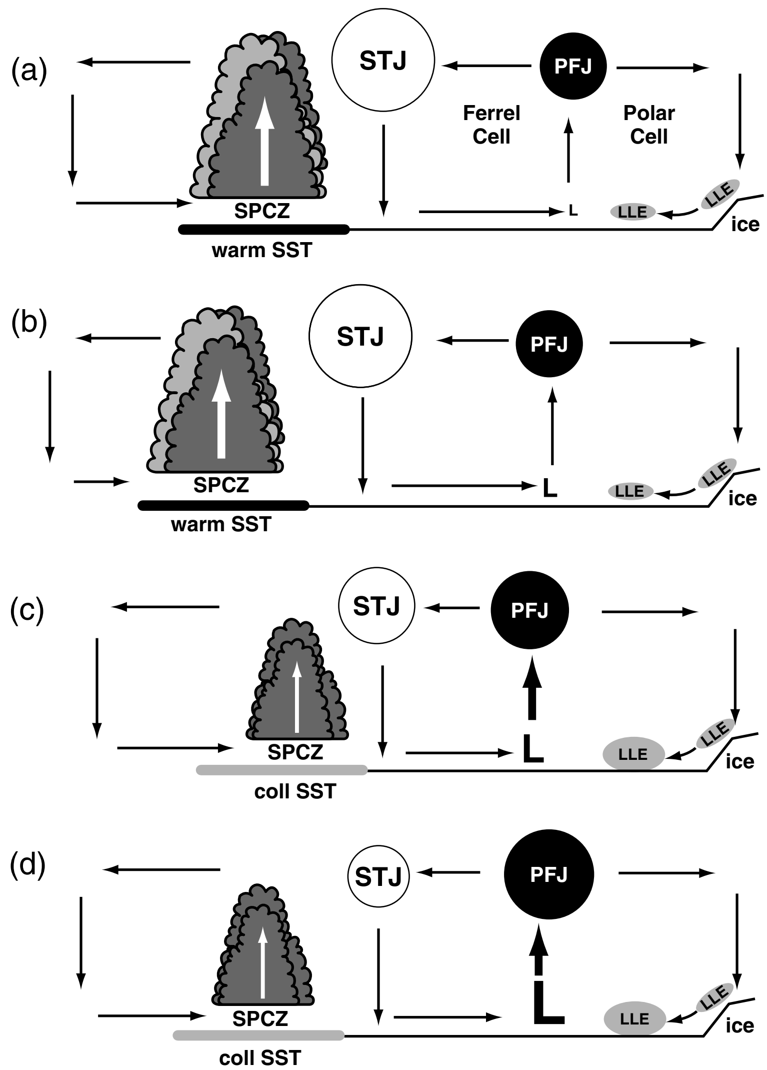

Figure 7. Schematic meridional cross-section for the South Pacific sector for (a) the year before and (b) the year after the 1987 SOI minimum and (c) the year before and (d) the year after the 1989 SOI maximum. The size of the clouds represents the strength of the SPCZ and the size of the circles represent the strength of the subtropical (STJ), polar front jet (PFJ), and low-level easterly (LLE) jet. The size of the 'L' represents the strength of the Amundsen Sea low. From Chen *et al*. (1996)

result of the weaker (stronger) zonal winds during the El Niño (La Niña) phase of ENSO. As can be seen from Figure 8, the most significant changes in the zonal winds (the high-latitude westerlies) are over the central South Pacific close to 140 °W, although during the El Niño event significant anomalies are also found from this location towards the tip of South America.

The split jet region of the western South Pacific clearly plays a role in the southward transmission of ENSO signals towards the Antarctic, but the different responses of the atmosphere here to different warm and cold events complicates the search for robust relationships between tropical and Antarctic atmospheric conditions. In a study of the split jet covering the period 1958–98, Bals-Elsholz *et al*. (1999) found only a weak correlation between a winter-season index of the split jet and the SOI, and that was because of the inclusion of the STJ. In the same study the PFJ was found to have little correlation to the SOI. Clearly, more work is needed on how the split jet varies in relation to ENSO and, in particular, on the mechanisms by which ENSO signals are transmitted southwards.

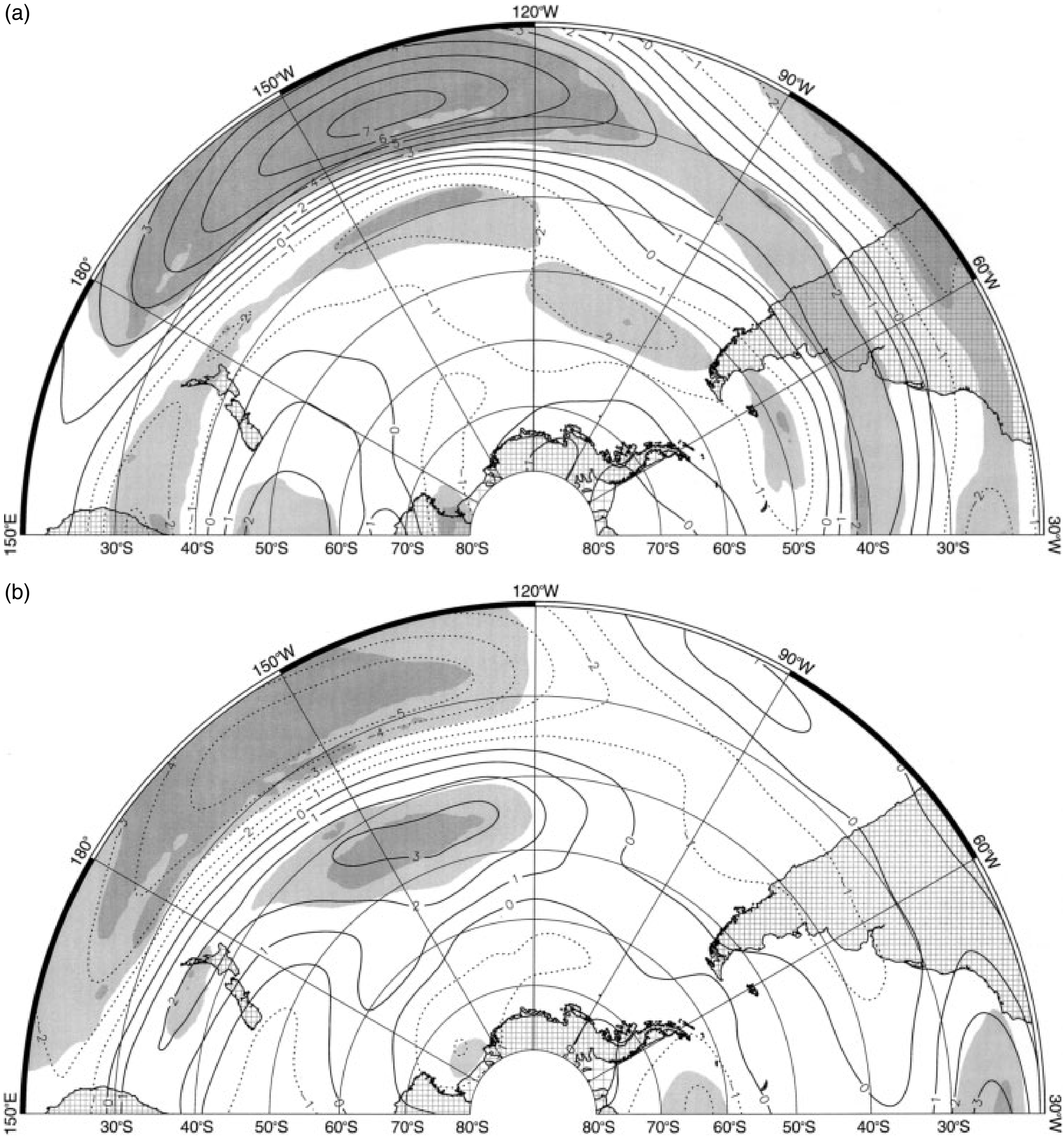

Figure 8. Winter-season (June–August) 300 hPa zonal wind anomalies for (a) El Niño and (b) La Niña years during the period 1968–97. The anomalies were derived from NCEP–NCAR reanalysis data

4.3. *Over the Antarctic continent*

One of the first meteorological studies into ENSO effects on the continent was carried out by Savage *et al*. (1988) using observations from Amundsen–Scott Station at the South Pole, along with the available automatic weather station (AWS) observations. They reported a correlation between El Niño events and low temperatures at the South Pole in the year following a minimum in the SOI. This was most pronounced after the 1982 'warm' event, when the mean annual temperature at the South Pole was $-51\,^{\circ}$C or 3.3 standard deviations below the 1957–86 mean. However, the study only covered the period 1955 to 1985 and the inter-annual variability of temperatures at the South Pole is small (standard deviation 0.65 °C). When the period 1955 to 1998 is considered, which covers several more El Niño events, it is found that there is no statistically significant link between annual mean SOI values and following year mean temperatures during El Niño events (correlation coefficient 0.35). Although South Pole temperatures were colder than normal in the years following the 1982, 1992, 1993, 1994 and 1997 warm events, this was not the case in 1977, 1987 or 1997, and, significantly, cold years have also followed La Niña events. From the previous discussion on how ENSO signals may be transmitted from the tropics to the Antarctic coastal region, it is very unlikely that there will be a simple, linear relationship between the SOI and South Pole temperatures, although in some years an El Niño signal may reach the South Pole. However, a general cooling at high southern latitudes at a minimum in the SOI was also suggested by Pan and Oort (1983). They found that the zonally averaged atmospheric temperature difference between 'warm' and 'cold' events was significantly negative south of 60 °S and through most of the troposphere poleward of 70 °S. But again, it should be noted that temperature data, such as AWS observations, at these latitudes are of debatable quality and this study needs to be repeated with more up-to-date data sets.

The Savage *et al.* (1988) study also examined the effects of the 1982–83 warm event on the anomalies of temperature and surface wind at occupied stations and AWS sites in the Ross Sea–Ross Ice Shelf area. The analysis suggested that the anomalously cold conditions in this area in the year following the warm event were associated with an enhanced surface layer inversion and stronger katabatic drainage flow from the interior. However, because of the anomalous nature of the 1982 El Niño event it is unclear how applicable this result is in general.

A broader ranging investigation into the relationship between the atmospheric flow across the continent and the phase of the SOI by Smith and Stearns (1993a,b) also used the available station observations of temperature and pressure, but not data from the AWSs. They examined the horizontal pressure and temperature patterns across the Antarctic in relation to the phase of the cycle and created composite pressure and temperature fields for warm and cold years using data from 24 stations that had records covering at least 30 years. The data used covered the period 1957 to 1984, which included six SOI minima. They reported a distinct switch in sign of the pressure anomalies from positive to negative across the minimum in the SOI. This can be seen in Figure 9, which shows composite annual pressure and temperature anomaly maps before and after the minimum of the SOI. The anomaly patterns have to be used with some care, since no AWS data were used and there are, therefore, very few observations over the interior of the continent. It can be seen that the anomalies are rather small down the length of the Antarctic Peninsula, but increase towards East Antarctica. In the area south of the ABS, where we might expect a larger impact of ENSO, the switches in anomalies are quite large, but it should be noted that the only station that has operated in this area was Russkaya, and then for only a few years in the 1980s, so that the long-term nature of the anomalies here cannot be estimated with any degree of confidence. The largest anomalies in Figure 9 are over Scott base, close to the Ross Ice Shelf, and near Mawson on the coast of East Antarctica. The smallest anomalies extend across the continent from Casey station to the South Pole, and then towards the Antarctic Peninsula. Such pressure anomaly changes from year_{-1} to year_0 have previously been noted by Carleton (1988; for the Weddell Sea area) and van Loon and Shea (1985, 1987), but it is currently unclear as to the mechanism that is behind them. The fact that there is such a large gradient in pressure anomaly around a relatively small sector of the coast of East Antarctica from Mawson to Casey is surprising, and suggests that more than ENSO is responsible for these changes. If ENSO is playing a major role here it is also surprising that the anomaly gradients are consistent, since we know that very large differences in atmospheric circulation anomaly are found with different El Niño events. Clearly, much more work is required on this very interesting line of research.

Obtaining a knowledge of the precipitation that falls across the Antarctic is essential for the interpretation of ice cores, so an important aspect of ENSO research is to determine whether the precipitation that falls across the continent is affected in a consistent way by different phases of the cycle. Cullather *et al.* (1996) carried out important work in this area by examining the precipitation variability across the whole Antarctic, and especially the South Pacific sector (120–180 °W) of the continent. This latter region is very important, with the study showing that nearly 40% of the moisture flux into

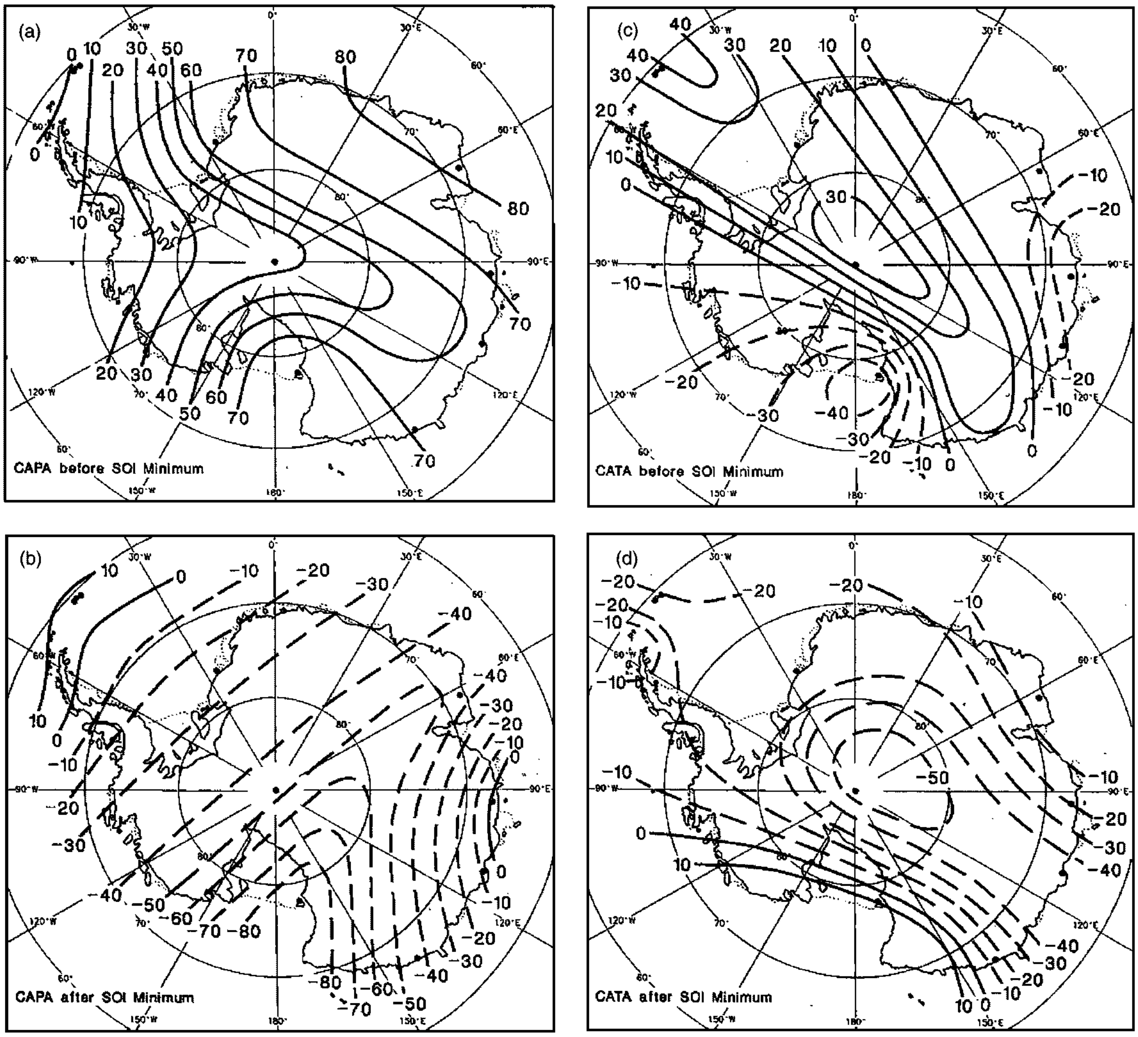

Figure 9. Composite maps for the SOI minimum. (a) The composite annual pressure anomaly (CAPA) before minimum. (b) The CAPA after minimum. (c) The composite annual temperature anomaly (CATA) before minimum. (d) The CATA after minimum. Negative anomalies are shown by dashed contours. From Smith and Stearns (1993a)

Antarctica occurred along the West Antarctic coast. Their study examined Antarctic precipitation variability (determined via a consideration of moisture fluxes) over the period 1980–94 in relation to the ENSO cycle. They found that over the period up to 1990 the moisture convergence for the whole continent was generally in phase with the SOI, with smaller values during warm events and larger values during normal or cold events. For the West Antarctic sector, which is just south of the split jet, the precipitation variability correlated well with ENSO over the period 1980–90 (Figure 10) with more (less) moisture convergence during warm (cold/normal) events. However, the relationship switched to an anti-correlation after 1990, which the authors linked with a strong ridging pattern over East Antarctica. This change further illustrates the fact that simple linear relationships between the SOI and high-latitude measures usually break down because of the apparently complex modes of variability of the Antarctic climate system.

In summary, the various forms of meteorological data present some indications of ENSO signals at high southern latitudes, but the lack of consistency in the response between events indicates the importance of other factors and atmospheric/oceanic processes at these latitudes.

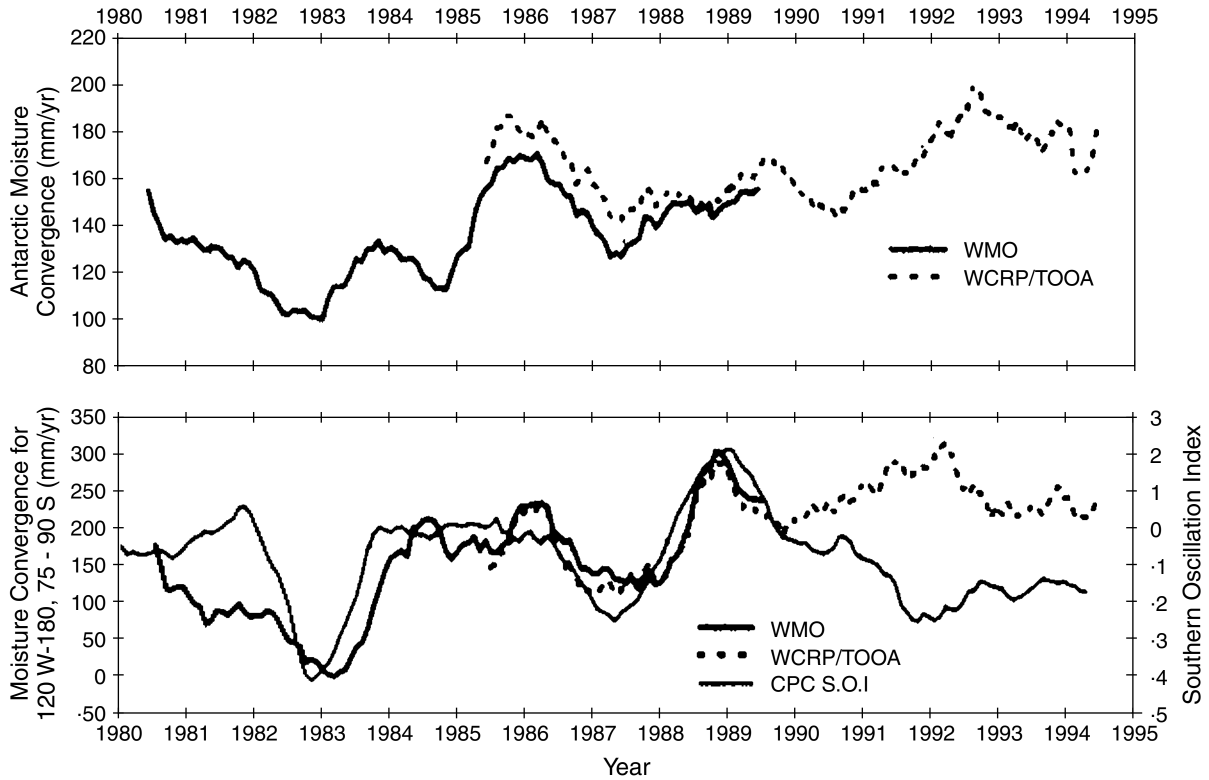

Figure 10. The 12 month centred running mean moisture convergence (mm year^{-1}) for Antarctica (top) and West Antarctic sector (bottom) in comparison with the SOI. From Cullather *et al.* (1996)

4.4. The southern annular mode

Recent research has shown that the principal mode of atmospheric variability of the extra-tropical regions of the Southern Hemisphere has a zonally symmetric or annular structure, which has been referred to as the southern annular mode (SAM; Limpasuvan and Hartmann, 1999; Thompson and Wallace, 2000). The SAM can be observed as an expansion and contraction of the polar vortex and resultant meridional shift of the mid-latitude jet, along with changes in the surface pressures and westerly winds in the Antarctic coastal region. In recent decades the SAM has shifted into its positive phase, with a decrease in surface pressures over the Antarctic and an increase in the westerlies in the coastal region. The reasons for this are still being debated, but increases in greenhouse gases and the depletion of stratospheric ozone could both have played a role in the changing nature of the SAM. The SAM provides a means of coupling the Antarctic climate with that of lower latitudes, and changes in ENSO will undoubtedly affect the SAM, but in a highly non-linear way. Because of the importance of the SAM to the Antarctic climate, more research is needed on its relationship to ENSO.

4.5. Mesoscale weather systems

Mesoscale weather systems or mesocyclones (those with a horizontal length scale of less than about 1000 km and a lifetime of less than 24 h) are a common feature of the ice-free ocean areas around the Antarctic (Carleton and Carpenter, 1990; Heinemann, 1990; Turner and Thomas, 1994; Rasmussen and Turner, 2003). Their development is strongly influenced by the synoptic-scale flow and they are particularly common during cold air outbreaks, especially to the west of deep depressions. Since changes in tropical circulation affect the extra-tropical jet patterns and, therefore, storm tracks, we may expect to find variations in mesocyclone development during different phases of the ENSO cycle. This possibility was investigated by Carleton and Carpenter (1990) as part of their study of mesocyclone variability across the whole of the Southern Hemisphere. Their work covered the seven winters over the period 1977–83 and was based on infra-red satellite imagery. They concluded that the inter-annual variability of mesocyclones was, at least in part, linked to ENSO. With the amplified seasonal cycle found during ENSO 'warm' events (e.g. 1982), large numbers of mesocyclones were found southeast of Australia and around New Zealand (Figure 11). In the previous year (El Niño year$_{-1}$) the annual cycle was suppressed and there were fewer mesocyclones. In addition, there was a longitudinal displacement of the peak of mesocyclone activity between the two years,

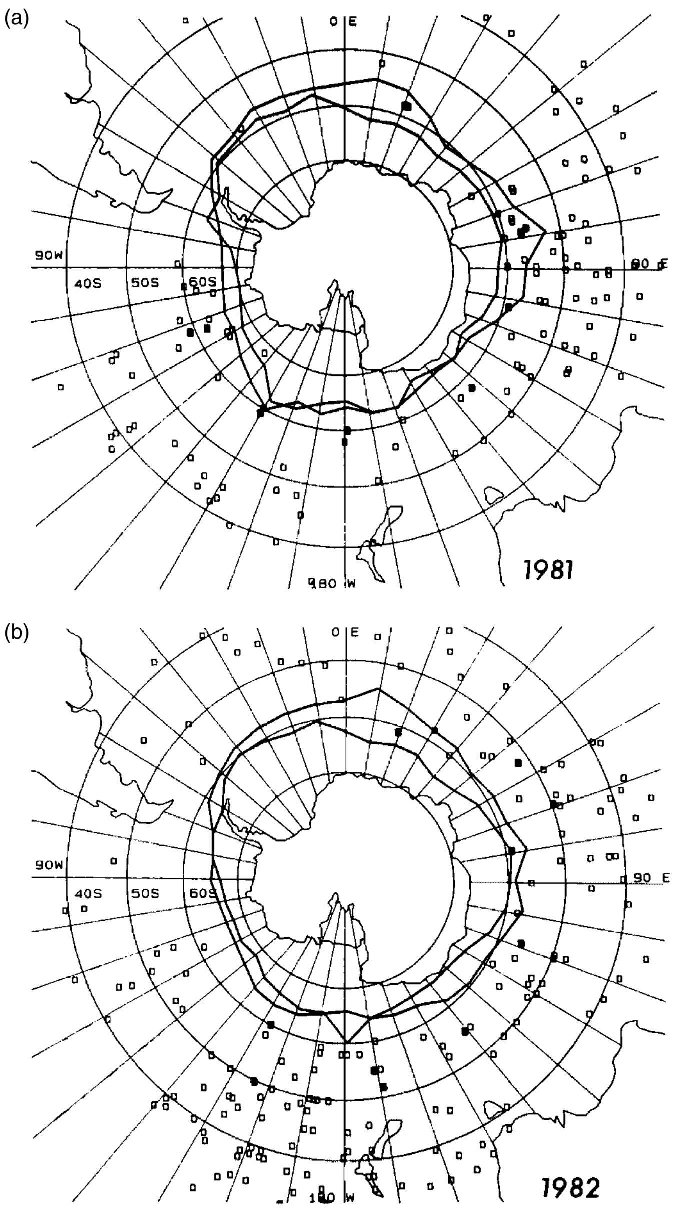

Figure 11. Locations of winter-season (June–September) polar low cloud vortices for (a) 1981 and (b) 1982. Comma clouds are shown as open squares; spiraliform systems are shown as filled squares. The June and September sea ice extents are also shown. From Carleton and Carpenter (1990)

with more (fewer) systems near 90 °E (southwest of New Zealand) in 1981, with the reverse being true in 1982. The occurrence of the lows was linked to the frequency of cold air outbreaks in the New Zealand sector and, as discussed above in connection with the New Zealand split jet, such outbreaks appear to be related to the maintenance of the split jet, which in turn is affected by ENSO. Fitch and Carleton (1992) also found large changes in the polar low regimes between the winter of 1988 (the peak of a cold event) and the following winter.

All these investigations covered relatively short periods, and studies examining longer data sets need to be undertaken before firmer links between mesocyclones and ENSO can be confirmed.

4.6. Possible precursors of ENSO at high latitudes

An important question regarding ENSO is whether we can find statistically significant signals of circulation change or reversal *in advance* of major events. Possible precursors of ENSO 'warm' events at high southern latitudes have, therefore, been investigated by a number of workers. Van Loon and Shea (1985) revealed the existence of some strong and statistically significant reversals of sign of winter MSLP anomaly fields associated with ENSO for the central South Pacific and Australia–eastern Indian Ocean areas. Their analysis only extended to 50 °S, but there was a suggestion for the Weddell Sea that significant pressure reversals occurred between the austral winter preceding the event ($year_{-1}$) and the winter of the event ($year_0$). Trenberth and Shea (1987) also suggested that the extra-tropical regions of the Southern Hemisphere were involved in the period up to the peak of 'warm' events and they identified changes in storm tracks near New Zealand as preceding opposite changes in the Indonesian centre of action by one or two seasons.

Since ENSO is principally a phenomenon of the tropics and most of the atmospheric and oceanic energy is at these latitudes, it is difficult to believe that robust precursors of ENSO events will be found in high southern latitudes. When only a small number of events are considered, some such high-latitude signals will inevitably be found; but, as longer data sets are considered, it seems probable that markedly different conditions prior to events will be found.

5. SIGNALS OF ENSO IN THE OCEAN ENVIRONMENT AND ANTARCTIC SEA ICE

5.1. The ocean

Pittock (1984) was the first to suggest that ENSO could affect high-latitude oceanic conditions via atmospheric forcing of the ACC. His proposal was that changes in cyclonic activity across the ocean could affect the location of the ACC and result in oceanic changes, such as SST anomalies propagating northwards via the Peru Current. Data were rather sparse at the time of that work, but the later use of combined global ship and satellite-derived SST data sets allowed a more detailed investigation of the impact of tropical Pacific SST anomalies on conditions in other parts of the world. Klein *et al*. (1999) found that positive SST anomalies occurred in a number of ocean basins approximately 3–6 months after SSTs peaked in the tropical Pacific. These were ascribed to changes in cloud cover and evaporation, following changes in atmospheric circulation that took place during the El Niño events, which in turn increased the net heat flux entering these areas. They called this concept of a link between SSTs in the tropical Pacific and remote ocean areas an 'atmospheric bridge'. Their work was concerned only with the tropical latitudes, but Li (2000) examined the links between ENSO and SSTs across the Southern Ocean through a modelling investigation. That work showed that, during El Niño events, positive SST anomalies were found over the western part of the South Pacific during summer, which would be consistent with the greater degree of blocking, reduced cloud and more shortwave radiation received during 'warm' events. In this phase of the cycle there were also negative anomalies over the eastern South Pacific. It was suggested that the SST changes at high latitudes took place through modification of the Walker–Hadley circulation, resulting in changes to the surface heat fluxes at the ocean surface. The results indicated reduced surface wind stress to the west of the tip of South America in the experiment with the El Niño tropical SST anomalies, which is again consistent with higher pressure over the southeast Pacific

Ocean. Other SST anomalies were also suggested by the modelling work, including a cold tongue from northern Australia to the tip of South America and warm anomalies west of the Drake Passage.

At higher latitudes, Ledley and Huang (1997) examined SST data for the Ross Sea sector covering the period 1982–94 and found a statistically significant (at the 95% level when autocorrelation was considered) relationship with ENSO, with higher Ross Sea SSTs during El Niño events. Over this period the Ross Sea SSTs lagged the ENSO signal by 2–4 months, with the highest correlations occurring with a lag of 3 months. This can be seen in Figure 12, which shows the Niño 3 SST and Ross Sea SST anomaly. Their work also examined the sea ice concentrations over the same period, which also indicated a relationship between the concentration in the area and ENSO, with reduced values (by between 3 and 10%) in warm-event years. This resulted in a reduction of the seasonal range of the total sea ice area during El Niño events, since the variations in the minimum ice area are small. The higher SSTs and reduced sea ice during 'warm' events are consistent with the observed greater moisture flux convergence into West Antarctica during this phase of ENSO, implying strong northerly flow in the atmosphere. However, it should be noted that the Ledley and Huang (1997) study only included three El Niño events; with our current knowledge of the variability in the high-latitude response to 'warm' events, this needs be extended to deal with the more recent events.

5.2. Sea ice

In many ways, the Antarctic sea ice provides one of the most attractive means of investigating the high-latitude impact of ENSO, since the vast majority of the ice melts each year and the development of the sea ice through the winter season is very dependent on the atmospheric circulation. However, our understanding of how the oceanic conditions affect the development and maintenance of the sea ice is poor. In studies such as that of Carleton (1988), in which signals of ENSO in northern Weddell Sea sea ice were examined, no account could be taken of the role or variability of the cyclonic oceanic gyre that is found in the area and the variability in ice extent/concentration was ascribed to the atmospheric flow only. We know that the atmospheric impact on sea ice conditions is large in the short term, but we cannot through observational data alone determine the effects of ocean variability, because of the lack of suitable data.

Examination of the record of Antarctic sea ice extent in relation to the atmospheric circulation (Harangozo, 1997) has shown that winter sea ice coverage in the eastern Ross Sea and the Bellingshausen Sea is strongly influenced by the atmospheric flow through direct forcing of the wind field on the sea ice. This also came out in the Kwok and Comiso (2002) investigation. Therefore, we may expect to find some evidence of ENSO in

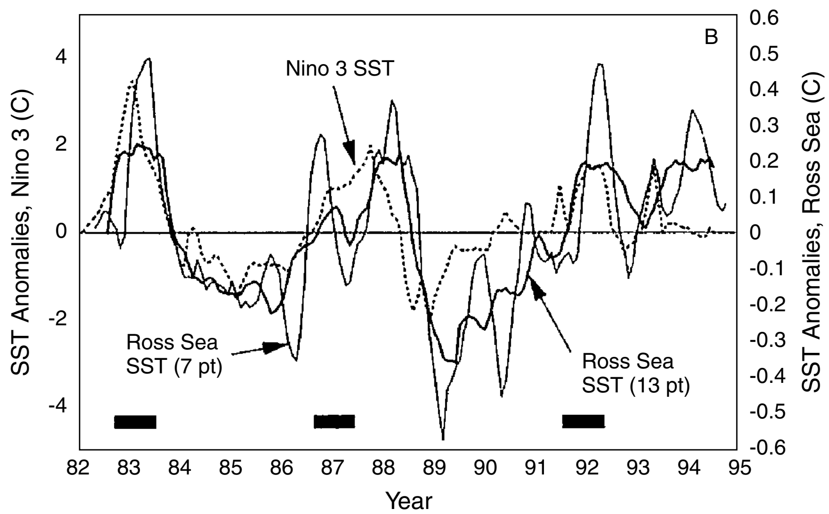

Figure 12. Monthly mean smoothed Ross Sea SST anomalies (13-point smoothed, thick solid line; seven-point smoothed, thin solid line) and Nino 3 region (150–90 °W, 5 °N–5 °S) SST anomalies (dashed line) from the 13 year monthly means. The horizontal bars are each 1 year long and indicate the approximate times of ENSO events. From Ledley and Huang (1997)

the Antarctic sea ice on seasonal time scales. However, the relationship between the circulation and sea ice extent is complex.

The earliest study to imply a link between Antarctic sea ice and ENSO was that carried out by Heap (1965). He found a 2 to 4 year periodicity in the ice coverage in the Weddell Sea sector, but did not explicitly suggest a connection with ENSO. Once the satellite-based Antarctic record of sea ice was analysed, it became clear that there were eastward-propagating signals of inter-annual anomaly on ENSO time scales (Zwally *et al*., 1983). Later work has concentrated on the South Pacific area, and especially the ABS, where the PSA Rossby wave train has its greatest impact, since this is where a robust signal may be expected. With climatologically anomalous southerly flow to the west of the peninsula during El Niño events, positive anomalies in the winter sea ice can develop to give significantly colder surface air temperatures at a research station on the coast, such as Faraday/Vernadsky (Weatherly *et al*., 1991; King, 1994; King and Harangozo, 1998). However, Harangozo (2000) found that, in addition to anomalous southerly flow, it was also necessary to have extensive pre-winter sea ice (pre-conditioning) in order to get the coldest winter temperatures on the western side of the peninsula. Therefore, as discussed above in relation to the role that ENSO plays in modulating the atmospheric circulation over the ABS, the response of the winter sea ice to the different phases of the cycle can be very variable, since it is not only affected by the meridional flow, but also the pre-winter ice extent. However, the study by Harangozo (2000) developed a regression model that captured most cold winters over the western Antarctic Peninsula (which are dictated primarily by the extent of the sea ice of the eastern ABS), indicating the degree to which ENSO-related SST anomalies in the tropics can influence Antarctic sea ice.

Work by van Loon and Shea (1985) pointed to possible ENSO-related signals in the atmospheric circulation of the Weddell Sea region, and Carleton (1988) elected to examine this area for possible ENSO signals in the sea ice record because of the availability of ship data and sea ice reports. Within that study he created composite analyses of Weddell Sea sea ice at extremes of the ENSO cycle from ship- and shore-based observations of ice extent collected over the period 1929–62. He found that there were lower ice concentrations, and therefore more open water, in the December to January period of El Niño events (year$_0$) compared with the previous summer (year$_{-1}$). He ascribed this to the more cyclonic atmospheric circulation found over the high-latitude parts of the South Atlantic during El Niño events, giving enhanced southerly flow and divergence of the pack. This is consistent with the Scotia Bay, South Orkney Islands (Orcadas) sea ice record covering the period 1903–74. Here, a statistically significant (at the 95% level) later clearance of the ice was found in warm-event summers (mean date 9 December) compared with the previous summer (mean date 12 November). Greater cyclonic activity in this area, coupled with the greater blocking over the ABS, would imply more southerly flow over the western Weddell Sea and advection of sea ice towards the north. Such conditions occurred during the major warm event of 1997–98, when sea ice was extensive around the South Orkney Islands. Murphy *et al*. (1995) reported links between ice duration at Signy station and ENSO. However, their figure 6 shows that there is considerable variability in the response during particular events. The 1988 ‘warm’ event coincided with a long ice duration at the station, whereas during the 1982–83 event the duration was short. Overall, the study showed that the fast ice duration in the South Orkneys lagged the ENSO signal by 1–2 years.

Hao *et al*. (1990) also considered the relationship between ENSO and the sea ice in the Weddell Sea. They suggested that ENSO precursors could be found in the sea ice extent some time before events, with there being a negative correlation with ‘warm’ events 1–2 years previously and a positive correlation 3–4 years prior to SST anomalies in the eastern tropical Pacific. In addition, they pointed to similar precursors in the Ross Sea, and suggested that the ice in year$_{-1}$ had a strong negative correlation with SST anomalies in the Niño 4 areas and positive correlations with the SOI in year$_0$. Sea ice 2.5–3 years before had a strong negative correlation with SST anomalies in the Niño 3 area and a strong negative correlation existed for sea ice 3–3.5 years before to SST anomalies in the western tropical Pacific and the SOI. They suggested that ice in the Ross Sea first affected the western part of the tropical Pacific and then gradually expanded its influence to the eastern part a half year later. This investigation was later broadened (Xie *et al*., 1994) to cover a wider time period and to consider interactions in both directions between ENSO and Antarctic sea ice. They again concentrated on the Ross Sea sector, where they found strong ENSO–sea ice interactions, with the highest correlations suggesting both leads and lags between the quantities. They found that the sea ice in the Ross

Sea had a quasi-11 year periodicity and had a strong cross-correlation with the SST anomalies in the Niño 4 area. Their common period produced a resonance from 96 months leading to 36 months lagging, causing a sine-shaped correlation variation with a strong positive SST anomaly from 87 to 50 months leading and a strong negative correlation from 20 months leading to 24 months lagging. A similar result was found for the Weddell Sea sea ice and SST anomalies of the central-eastern tropical Pacific, but with a common period of around 5 years. The feedback of sea ice to SST anomalies in the western tropical Pacific was also significant with a quasi-5 year period, but was very weak for SST anomalies in the central-eastern tropical Pacific. SSTs of the central tropical Pacific had a quasi-contemporary oscillation relationship with Ross Sea sea ice and a 1.5 year lag oscillation relationship with Weddell Sea ice. They called this oscillation relationship between Antarctic sea ice and ENSO events the Southern Oceanic oscillation. The sea ice data used in this investigation covered only the period 1973–89 and, therefore, only contained four 'warm' events. Therefore, it would be very useful to extend the work to cover more recent events.

Regional signals of ENSO in the sea ice have also been noted in a number of sectors of the Antarctic away from the Antarctic Peninsula. As part of the investigation into the effects of ENSO on mesocyclone occurrence described above, Carleton and Carpenter (1990) found that there was increased offshore flow from the Ross Ice Shelf sector of the continent towards New Zealand during the 1982 warm event, although, as noted above, the high-latitude response during the 1982–83 'warm' event was anomalous. As can be seen in Figure 11, there are some indications that the mesocyclone outbreaks are associated with a more northerly sea ice edge in this sector, especially earlier in the winter. Similarly, the more northerly ice edge to the southwest of Australia in 1981 appears to be linked with mesocyclones. However, it must be stressed that sea ice anomalies can be dictated by many other factors than ENSO, and establishing statistical relationships that are robust in the long term has proved very difficult.

Early in the 20th century, Walker (1923, 1924) suggested that ENSO may be, at least in part, a response to forcing from high southern latitudes and examined the relationship between sea ice and pressure conditions. Variability in Antarctic sea ice conditions does seem to influence atmospheric conditions over a reasonably large area, although the impacts are not clear. Peng and Wang (1989) suggested that it had an influence on the northwest Pacific subtropical high, which they believed may be affected via the movement of ocean currents in the South Pacific region. Chiu (1983a,b) investigated further this possibility of ENSO having its origins in the Antarctic and found that there was a significant association between the SOI in March–April and the area of the sea ice in the following July–December period for the 8 years 1973–80, and also suggested that there was an apparent lag of the SOI with the sea ice area in the latter part of the year. The study suggested that a stronger zonal Walker circulation was associated with larger sea ice extent. This work was then carried forward by Carleton (1989), who took into account the autocorrelations in the sea ice and SOI, both of which can be large. His work also examined the associations on a regional basis and in connection with various synoptic pressure indices. The study showed that many of Chiu's apparently significant relationships could have occurred by chance, although some consistent associations were found for some months, particularly the major areas of ice creation in the Ross and Weddell Seas. In these areas the sea ice changes tended to lag the SOI, but preceded changes in the eccentricity of the circumpolar vortex, or wavenumber 1, the so called transpolar index (TPI). The TPI is determined from the MSLP anomaly difference between Hobart, Australia, and Stanley, Falkland Islands, and provides a useful indication of changes in the longest atmospheric waves around the Antarctic. Pittock (1984) was the first to suggest a connection between the TPI and the SOI; in the Carleton (1989) paper this possibility was investigated further, with the TPI–ENSO relationship over the period 1973–82 being represented schematically by a figure, reproduced in Figure 13. During this period, a greater ice extent in late winter was linked to a displacement of wavenumber 1 towards the Australia–New Zealand area (TPI negative) and an amplification of the Tasman Sea trough between August and September. These changes occurred about 5 months prior to changes in the tropical Walker circulation and the SOI. Clearly, this work needs extending to examine a longer period, but the results are in line with those from other studies (Fletcher *et al.*, 1982; van Loon and Shea, 1985; Trenberth and Shea, 1987).

As the record of sea ice extent and concentration derived from passive microwave data became longer, so it was possible to carry out investigations of ice-extent–ENSO relationships covering more 'warm' and 'cold' events. Gloersen (1995) examined data for the period 1978–87 and used a multiple-window harmonic

TPI, $\Delta MI_{(SEP.AUG)}$, $ICE_{(SEP)}$, $SOI_{(FEB+1)}$

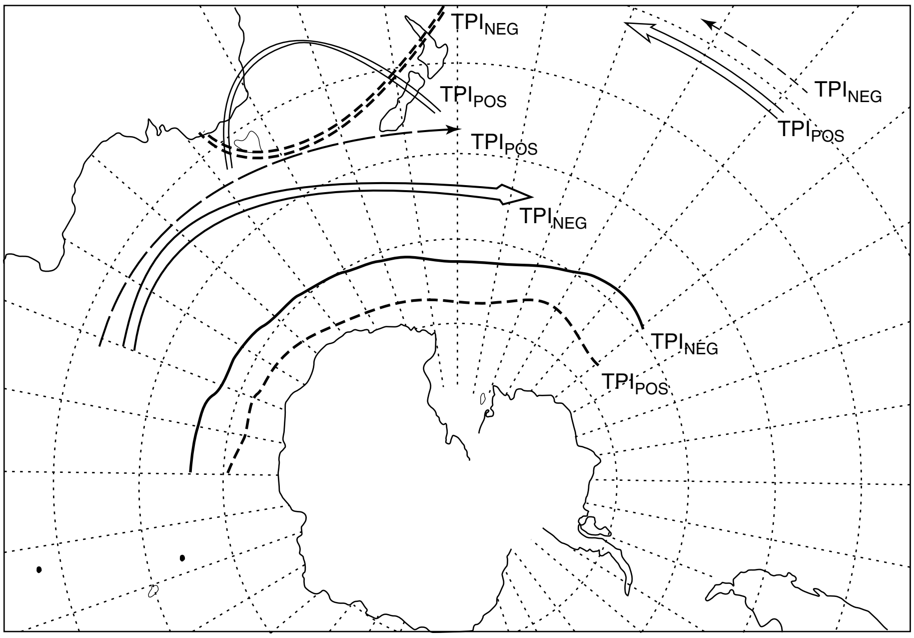

Figure 13. Schematic showing the interrelationship between the sea ice extent and atmospheric circulation in the southwest Pacific, inferred from statistical analyses for the 1973–82 period. Increasing (decreasing) ice occurred in late winter when the TPI was negative: wave displaced towards Australia (positive: wave displaced towards South America) and the Tasman Sea trough amplified (weakened) between August and September. These apparently preceded, by about 5 months, changes in the Walker circulation. From Carleton (1989), reproduced by permission of Springer

analysis technique to show that the time series of Antarctic sea ice cover contained statistically significant quasi-biennial and quasi-quadrennial periodicities that agreed well with variations in the SOI. However, the response of sea ice to these two frequency components varied greatly for different regions. In the Weddell Sea sector, the quasi-quadrennial component had a significant link with ENSO, confirming the earlier work of Carleton (1989), and indicating that sea ice in this area has a significant link with the phase of the SOI. However, the Gloersen (1995) study also showed that the quasi-biennial periodicities were strongest in the ABS, and that in the Ross Sea the quasi-triennial cycle was strong but the quasi-biennial rather weak.

Another longer term investigation into ENSO and sea ice variability was carried out by Simmonds and Jacka (1995) using a sea ice data set that covered the period 1973–92. This built on the work of Chiu (1983a,b) and Carleton (1989) described above. They calculated the correlation of sea ice and SOI for all pairings of calendar months, as well with the SOI from the previous and subsequent years. Most of the largest correlations were found when the SOI led the anomalies in the sea ice, but the results were found to vary according to the sector of the Antarctic examined. Overall, the study revealed a positive correlation between the zonally averaged sea ice extent in the months April to July with the SOI during most of the previous 12 months. The extent of the sea ice in the Indian Ocean in the months April to July was positively correlated with the SOI during most of the previous 12 months. Large correlations over the Pacific sector were found only for the spring season and only with the SOI during the year of the anomalies. No significant correlation was found between sea ice in the Atlantic sector and the previous year's SOI in any month. Some indications were found of sea ice anomalies preceding the SOI values, but these were not as marked as those relating to current or past SOI values. Using the results of earlier investigations, many of which are described above, Simmonds and Jacka (1995) re-examined the links between the SOI and sea ice extent in four key sectors of the Antarctic: the southwest Indian Ocean, the southwest and southeast Pacific Ocean and a sector to the west of the Ross Sea. It was found that the correlations for these areas were stronger than those found over the entire ocean basins, which they interpreted as indicating that there are key parts of the Antarctic sea ice zone where the ENSO links are strongest. One of the strongest relationships that this study found was between ice extent in the southeast Indian Ocean over the period April to October and the SOI in the previous 12 months. In the Australian sector, to the west of the Ross Sea, the largest correlations are associated with sea ice conditions leading the SOI. It was found that the sea ice in May–July was consistently related to the SOI over much of the previous year, a result that is consistent with the work of Carleton (1989).

In a recent study, Yuan and Martinson (2000) examined the temporal cross-correlations between the detrended Antarctic sea ice edge position anomaly and various climate indices, including several associated with ENSO. They found that the sea ice-edge anomaly in the Amundsen–Bellingshausen–Weddell Seas sector showed significant correlation with ENSO parameters, including the Niño 3 and Indian Ocean SSTs and tropical precipitation. As in earlier investigations, the link between climate indices and sea ice-edge anomalies extended over a full range of lead–lag relationships. They make the important point that the temporal–spatial quasi-periodic nature of the ice anomalies and the temporal quasi-periodic nature of the climate indices prevents any identification of particular lags as representative of a direct link between the ice anomaly and extrapolar climate indices, preventing identification of the direction of causality. The correlation of ice-edge anomaly with detrended global surface air temperatures revealed four teleconnection patterns. Firstly, an ENSO-like pattern in the tropics with strong correlations in the tropical Indian Ocean and North America. Secondly, a teleconnection pattern between the eastern Pacific region of the Antarctic and the western-central tropical Pacific. Thirdly, an Antarctic dipole across the Drake Passage and, fourthly, meridional banding structures in the central Pacific and Atlantic extending from the polar regions to the tropics, and a signal even in the Northern Hemisphere. An EOF analysis of the temperature anomaly fields with the correlation fields was carried out to attempt to distinguish the areas representing physically meaningful links. There were particularly strong links between the sea ice fields in the eastern-central Pacific and the Weddell Gyre region, with the regions influenced by ENSO, e.g. the tropical Pacific, western Pacific and Indian Ocean. The strongest correlations associated with the ENSO signal appear to occur with the sea ice extent anomaly at 120–132 °W (ABS area) lagging the temperature anomaly by 6 months. The authors point out that an austral spring–summer El Niño event would have a corresponding signal in the ice extent 6 months later during the early ice growth period in the autumn–winter.

Coastal polynyas can also be induced by anomalous atmospheric flow associated with ENSO events. During the major El Niño of 1997–98 the coastal polynya just north of the Ronne Ice Shelf reached an extent not seen previously in the sea ice record derived from passive microwave satellite imagery, which extends back to the early 1970s (Renfrew *et al.*, 2002). The polynya was so large because of the strong southerly atmospheric flow.

In summary, the work to date on SOI–sea ice links has shown the complex nature of the connections that are a function of season and sector of the Antarctic. But the best evidence to date suggests that the SOI leads the anomalies in the sea ice (Simmonds and Jacka, 1995). However, as the record of sea ice extent and area become longer, it should become possible to establish more reliable statistical relationships than have been developed to date.

6. SIGNALS OF ENSO IN ANTARCTIC ICE CORES

The shortness of the meteorological record from occupied stations and AWSs can potentially be extended through the use of proxy data derived from ice cores drilled at coastal and interior sites across the Antarctic. Each core provides information on aspects of the climate at the drilling site, such as annual surface mass balance and annual mean temperature, through analysis of electrical conductivity (acidity), oxygen isotopes and stratigraphy, as well as chemical species in the snow. Such data can then be used to infer aspects of the atmospheric and oceanic changes on ENSO time scales, assuming that the data from the cores are representative of the broad-scale environment. However, there are a number of problems in the use of ice-core data. It is possible to see some sub-annual signals in ice cores from high-accumulation sites (Marshall *et al.*, 1998), such as on the western side of the Antarctic Peninsula. However, at most sites away from the coast, where accumulation is relatively low, mixing of the snow in the upper layers means that it is necessary to apply running smoothing over 3–5 years so that the data are not dominated by local depositional 'noise'. This is unfortunately the time scale on which ENSO operates, so that it makes the identification of the signals of individual events very difficult, unless unsmoothed data are used. Nevertheless, a number of studies have been carried out to examine possible ENSO signals in ice cores and to try and understand how such signals could be transferred from the tropical Pacific to high, interior sites in the Antarctic.

One of the most powerful tools used is the analysis of methylsulphonic acid (MSA) in ice cores. MSA is an ionic species produced from the photo-oxidation of the marine biogenic sulphur species dimethylsulphide (DMS) and provides an indication of the track of air masses, since we know that air masses passing over the ocean will contain more MSA than those arriving at a site after crossing more land or sea ice. Such information can be used to determine storm tracks and air masses arriving at a site on the monthly to seasonal scales, as well as provide data on sea ice extent. However, such analyses can only be carried out using data from high-deposition sites, where generally only relatively short records can be obtained.

Welch *et al*. (1993) showed that MSA extracted from ice cores and pits at a coastal site on the Newell Glacier in the Ross Sea sector could be used as a proxy for sea ice extent, although the relationships investigated were only on the annual time scale, since it was not possible to extract sub-annual MSA amounts at this location. However, as discussed above, the sea ice extent during the winter is related to the atmospheric flow over the season, but in a non-linear way, and major sea ice anomalies can be established over a relatively short period. The MSA in coastal cores will, therefore, reflect the combined effects of sea ice extent and the atmospheric flow that resulted in the sea ice anomalies.

Most ice cores are collected at high, interior sites where there has been little horizontal movement of the ice, so that the ice sampled fell as precipitation at these locations. Unfortunately, the annual accumulation at such sites is small. A further complication is that whereas at most coastal sites the precipitation comes from synoptic and mesoscale weather systems, in the interior the snowfall comes from both clear-sky precipitation and active weather systems (Noone *et al*., 1999), with the clear-sky contribution being dominant at locations well away from the coast. On the high interior plateau, at sites such as Amundsen–Scott station at the South Pole and Vostok station, weather systems are rather rare and most precipitation falls as diamond dust. At such interior locations Legrand *et al*. (1992) have examined the MSA to non-sea salt sulphate (nss-SO_4) molar ratio R and they found significant spatial variations. Overall, low R values were found on the high plateau, with higher values nearer the coast. Low R values are generally observed at low latitudes, but with a higher ratio towards the polar regions. So the ratios found in the Antarctic coastal region are in agreement with atmospheric observations in high southern latitudes. Legrand *et al*. (1992) consider a number of possibilities for the lower ratios found on the plateau, but conclude by agreeing with Saigne and Legrand (1987) that the interior is mainly supplied by the end products of DMS emitted at temperate latitudes and transported southwards at mid-tropospheric levels.

Such a picture is consistent with the nature of the precipitation across the Antarctic, with synoptic factors being dominant in the coastal region and diamond dust well inland. However, it makes the interpretation of ice cores from intermediate regions very difficult.

The papers by Legrand and Feniet-Saigne (1991) and Legrand *et al*. (1992) both considered the question of possible ENSO signals in MSA and nss-SO_4 data from the Amundsen–Scott station ice core. Legrand and Feniet-Saigne (1991) examined the MSA and nss-SO_4 records from a South Pole ice core (collected in January 1984) in relation to ENSO events over the period 1922–84. Although the nss-SO_4 record was fairly constant, except for during two major volcanic eruptions, the MSA data showed high variability on ENSO time scales. The MSA increases were two to ten times above the background level, with the increases lasting for 0.5 to 2 years. Although there was some uncertainty in the dating of the core, which could be up to 1–3 years, all the MSA events were compared with the list of El Niño 'warm' events covering the last four and a half centuries compiled by Quinn *et al*. (1987). This showed that many of the MSA peaks correspond with major ENSO 'warm' events, including 1925–26, 1941, 1957–58, 1972–73 and 1982–83 (Figure 14). However, care needs to be taken in assuming that all MSA peaks are indicating 'warm' events or that all events are reflected in the core. The problem of dating MSA peaks in the core gives a 1–3 year uncertainty in the interpretation, and there are some difficulties in correlating the signals in the two cores collected at the pole. In addition, with such a data set there is a tendency to try and fit the MSA peaks in the core to the known sequence of El Niño events, rather than to compare the two independent data sets and to see how they correlate. However, a number of the MSA peaks do coincide with 'warm' events, but there are a few that occur at other stages of the ENSO cycle.

A relatively high MSA/nss-SO_4 molar ratio R in coastal cores suggests high-latitude Antarctic precipitation being due to local DMS emission (Saigne and Legrand, 1987). At inland sites on the plateau, R values of precipitation are closer to those obtained in the mid-latitude atmosphere, which acts as its source. In the South Pole cores the background levels of R are generally low, but the MSA peaks are characterized by higher R values. This could reflect more efficient atmospheric transport from marine areas or an enhanced contribution of high-latitude biogenic sources and greater DMS emission from the ocean areas around the Antarctic, rather than a source in tropical latitudes. But the poor correlation between sodium and MSA favours the latter possibility. Legrand and Feniet-Saigne (1991) suggest that this comes about because, during warm events, the southeasterly trades weaken or reverse to weak westerlies that may reduce the equatorial upwelling of Antarctic intermediate water. They suggest that a weakening of intermediate water circulation may cause a stagnation of Antarctic surface water, which may result in enhanced productivity. But the authors point to the problems in having such an oceanic link between high-latitude DMS production and warm events, highlighting the observed time scale, with the observed MSA perturbations being in phase with equatorial El Niño warm events, i.e. tropical SSTs. They also point to the enhanced growth of certain algae in sea ice and suggest that changes in the sea ice cover may provide the means by which the ENSO signals reach the Antarctic, an option considered by other workers later. In addition, high surface wind speeds can lead to more rapid exchange of DMS from the ocean surface to the atmosphere.

The results of a recent study of tropical atmospheric MSA and nss-SO_4 from the Pacific Exploratory Mission-Tropics has shed some doubt on this interpretation of the R ratio data from ice-core signals from the South Pole (Dibb *et al.*, 1999), and the later study of Meyerson (1999) described below did not use such data.

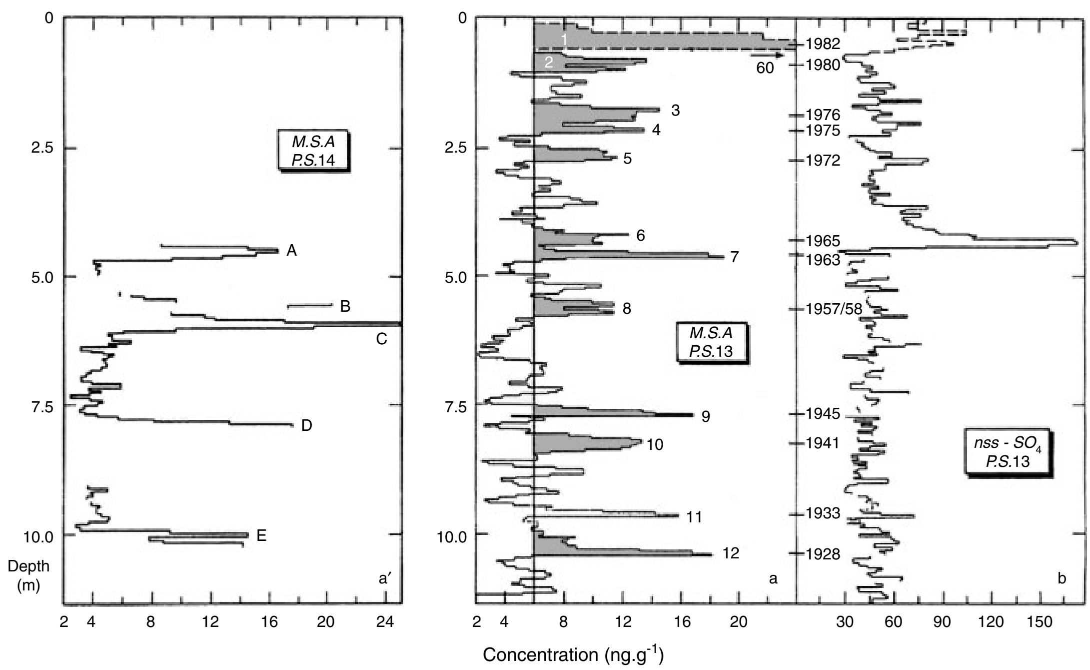

Figure 14. (a) MSA and (b) nss-SO_4 content (ng g^{-1}) along the PS13 south polar firn core (1922–84 time period). Major MSA spikes numbered 1 to 12 are indicated by shading. (c) Discontinuous MSA profile observed along the PS14 core over the same time period. From Legrand and Feniet-Saigne (1991)

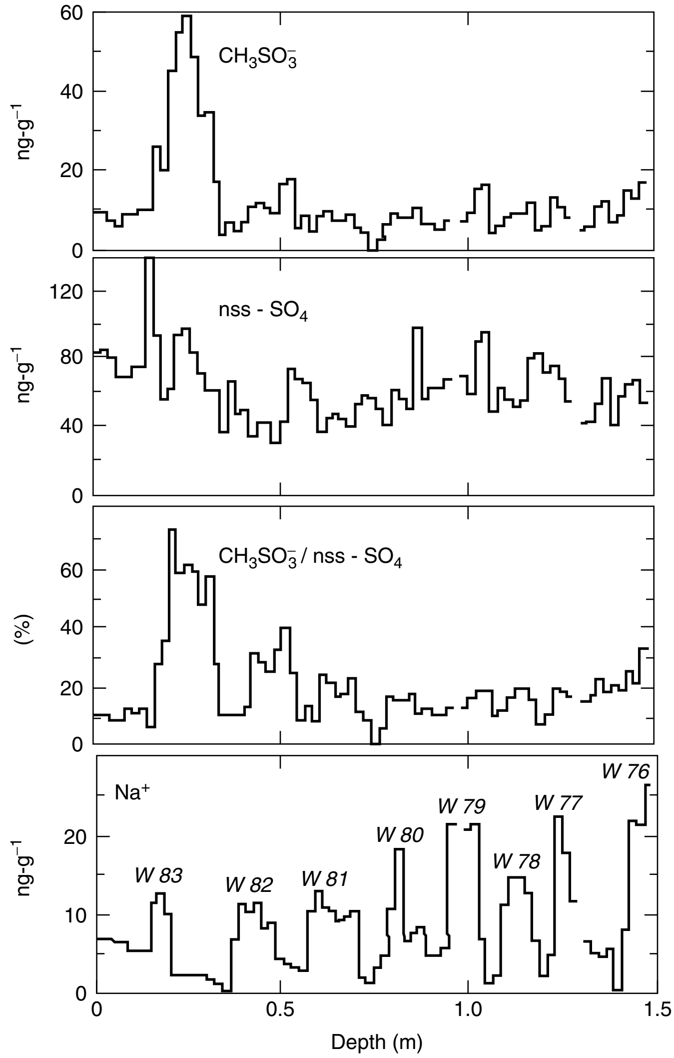

Figure 15. Seasonal variations of MSA, nss-SO_4, R and sodium content of south polar snow deposited between 1976 and 1983. W denotes winter seasons. From Legrand *et al.* (1992)

Legrand *et al.* (1992) presented figures showing the seasonal variations of various chemical species in the same South Pole core covering the period 1976–83 (Figure 15). The sodium profile provided a strong seasonal signal with a peak in the winter, which was attributed to the more frequent advection of marine air masses onto the plateau during winter. This allowed each winter to be clearly delineated and the annual cycle of other quantities and anomalies in particular years examined. Legrand *et al.* (1992) noted that both the El Niño years of 1982 and 1983 had large amounts of MSA, coupled with a small level of nss-SO_4, so resulting in high R values. This suggested an enhanced contribution of high-latitude DMS emissions towards the plateau during this event. They suggested that almost 50% of MSA originated from local DMS emissions, with the other half coming from north of 40 °S. In addition, they proposed that during the 1982–83 event the large increase in MSA was linked to local emission increase, as indicated by the small changes of nss-SO_4. Since the 1982 peak was not accompanied by higher sodium content they concluded, in the same way as Legrand and Feniet-Saigne (1991), that the perturbation reflected enhanced DMS emission at high latitudes. So at the pole, the MSA budget appears to be influenced by sub-Antarctic DMS emissions, especially in some of the El Niño years.

As discussed above, the Welch *et al*. (1993) study showed that the MSA record from the coastal site on the Newell Glacier could be used as a proxy for sea ice extent. However, when the records from other coastal and interior sites, including South Pole, McMurdo Dome and Bearmore Glacier, were compared with sea ice records, perhaps surprisingly, no significant relationships were found. It is believed that the ice-core mass balance data from some of the sites were reflecting only the local precipitation field, rather than that from the broader oceanic region, which would be required in order to have a significant correlation with the sea ice extent. This shows the great care that must be taken in selecting ice-core drilling sites if meaningful information is to be obtained on the broad-scale atmospheric and sea ice environment.

The later study of Meyerson (1999) returned to the question of whether the ice-core records from the Antarctic interior could provide information on sea ice variability and ENSO. That work involved the use of the MSA record from an ice core collected at the pole during 1995, which covered the extended period of 1487–1992. EOFs were used to decompose the variances between the MSA record and data on sea ice extent, the SOI and recent *in situ* meteorological data from the pole. The work suggested that the MSA deposition at the pole was positively related to the extent of the sea ice in the Ross and Amundsen Seas (175–115 °W). This was postulated to take place through the eastward movement during El Niño events of the Amundsen Sea low, which is climatologically located near 150 °W. At its more easterly location there is strong northerly flow on the eastern flank, giving less sea ice in the ABS and strong transport to the pole of air that has passed over a greater amount of open water than normal. However, this is contrary to all observational studies, which suggest greater blocking in this area during warm events.

The Meyerson (1999) analysis of the long-term relationship (1487–1992) between ENSO and the Antarctic suggested that, over this 500 year period, weather systems around the continent were responsible for the signals in the South Pole ice core via the sea ice in the Amundsen–Ross Seas. Although the ENSO–ABS sea ice relationship is one of the more robust links between the tropics and the Antarctic there are still years in which the relationship breaks down. In addition, the atmospheric transport from the ABS to the South Pole is very variable, and a back trajectory analysis really needs to be carried out so that we have more understanding of how air masses arrive at the pole. So, at the moment, there are indications that El Niño signals can reach the interior of the continent, but we need much greater understanding of the paths by which such signals arrive there.

7. ENSO AND HIGH-LATITUDE LIVING ORGANISMS

Although the relationships between ENSO and pelagic ecosystems have received a considerable amount of attention at tropical latitudes (Lehodey, 2001) and even towards the high-latitude areas of the Northern Hemisphere (McKinnell *et al*., 2001), there have been relatively few studies of ENSO effects across the Southern Ocean and in the Antarctic coastal region. This is, perhaps, not surprising, since, as described in earlier sections, there is still considerable debate about the robustness of relationships between ENSO and the atmospheric and oceanic conditions around the Antarctic. Nevertheless, some studies have attempted to link the various phases of ENSO and different components of the ecosystem. Testa *et al*. (1991) examined the occurrence of three species of seal around Antarctica and found quasi-cyclic patterns in some aspects of their biology. The age structure of crabeater seals (*Lobodon carcinophagus*) around the Antarctic Peninsula, juvenile leopard seal (*Hydruga leptonyx*) occurrence at Macquarie Island and Weddell seal (*Leptonychotes weddellii*) reproductive rate in McMurdo Sound have all shown peaks at 4 to 5 year intervals, as demonstrated by complex demodulation used to compare patterns among the three data sets with the SOI. The Weddell seal reproductive rate was generally in phase with the SOI since the Weddell series began in 1970; however, the leopard seals and SOI were in phase in the 1960s, but thereafter the SOI led the leopard seal series by about one-quarter of a cycle. Testa *et al*. (1991) suggest that these results point to large-scale oceanographic variations, possibly related to ENSO, as an important mechanism in Antarctic marine ecosystems.

More recently, Vergani *et al*. (2001) examined the possible effects of ENSO on populations of southern elephant seals (*Mirounga leonine* L.) on King George Island near the tip of the Antarctic Peninsula. This work was based on information on pup weaning mass collected over a 10 year period (1985–94) in relation to the SOI. Weaning mass was found to be higher during the La Niña phase and lower during El Niño events, which the authors suggest may be associated with variations in the seasonal sea ice zone. However, as described earlier, sea ice in the eastern Bellingshausen Sea, though generally being more extensive during El Niño events, can show marked differences between events.

Another ENSO link found in the Antarctic Peninsula region was with the growth rate of upper canine teeth of male Antarctic fur seals (*Arctocephalus gazella*) that died of natural causes at Bird Island, South Georgia. Growth was found to be greater in some years, and particularly poor years for growth were closely related to years in which reproductive performance was also observed to be low. Variations in growth from 1967–68 to 1987–88 were correlated significantly ($P < 0.008$) with the SOI.

The Antarctic Peninsula region is undoubtedly one of the best areas in the Antarctic to try and find biological signals of ENSO. However, we are clearly still at an early stage of understanding how ENSO affects the ecosystems of the Antarctic and the Southern Ocean. Progress in this area will need to take place in parallel with work on sea ice, ocean circulation and atmospheric flow.

8. DISCUSSION AND FUTURE WORK REQUIRED

The observation-based research discussed above has indicated that there is evidence of signals of ENSO in environmental data collected in the Antarctic. However, many of the studies have been concerned with relatively short periods and some of the results are conflicting. The clearest signal of an ENSO teleconnection in the Antarctic is via the PSA connection. There is a high degree of inter-warm and inter-cold event variability in the wave train carrying the ENSO signals towards the Antarctic (Houseago *et al*., 1998), so that composite analyses do not capture the complex dynamical nature of the events.

One of the major questions to be addressed regarding the propagation of ENSO signals to the Antarctic is the reasons for the different extra-tropical responses to different El Niño or La Niña events. Work such as that of Chen *et al*. (1996) has provided a great deal of insight into particular events, but there are major differences between events. Simmonds and Jacka (1995) have noted the rather long time scales that ENSO operates over and have made the important point that we should not necessarily expect that the strongest connection between the SOI and climate parameters noted in the Antarctic would occur in the same year. The time scale on which ENSO signals are transmitted poleward and the lags in the system also require further research. Simmonds and Jacka (1995) showed that the SOI and sea ice had little synchronous or lagged correlation. This was ascribed to the fact that considerable inertia was present in the system via the changing ocean circulation and modification to the westerlies.

There are also a number of other areas where research is needed into the relationship between ENSO and other atmospheric cycles. For example, Meehl (1991) has suggested that the semi-annual oscillation (SAO), which is a major factor in influencing the climate of the Antarctic coastal region, may play a role in the evolution of the extremes of ENSO. Similar arguments apply regarding the SAM, which is closely linked to the SAO.

To date, model-based investigations have not played a major part in the study of ENSO in the Antarctic, principally because there are still questions over the representation of the current high-latitude climate in such models (Connolley and Harangozo, 2001) and uncertainty over whether they are able to reproduce correctly the tropical/high-latitude teleconnections. However, the general circulation model (GCM) studies of Mitchell and Hills (1986) and Simmonds and Dix (1986), who forced the models with prescribed, negative winter Antarctic sea ice anomalies, found strong pressure/height responses in the tropics. In the Mitchell and Hills (1986) case, the anomalies had an ENSO-like appearance, with significant anomalies across the tropical Pacific and in the Indian Ocean.

Gaining further theoretical insight into the links between the tropical expression of ENSO and the high-latitude response and understanding how such signals are propagated is a high priority for the future. The work of Hoskins and Karoly (1981) advanced our understanding of such teleconnections in the Pacific, but progress since that time has been rather limited.

Other features of the Southern Hemisphere atmospheric circulation may also affect the poleward transmission of ENSO signals. For example, the SPCZ is a semi-permanent feature of the mid-latitude atmospheric circulation that is involved in the poleward transfer of energy and moisture. This feature links tropical convection and the cyclolysis/cyclogenesis taking place around the Antarctic. It also modulates the higher frequency (30–60 day) intraseasonal variations in outgoing longwave radiation over the Indian Ocean with the lower frequency oscillations further east associated with ENSO. Work linking the SPCZ with ENSO has been limited to date, but deserves further investigation. Additional work is also required on the role of the atmospheric transient eddies, since it has been shown by Carleton and Whalley (1988) that the efficiency of the heat transport by cloud vortices in winter is greater (less) when the SOI is low (high), which affects the poleward heat transport.

Although the meteorological data sets are still very short, it is clear that there are modes of climate variability in the Antarctic acting on a range of time scales and variability in the ENSO–Antarctic climate relationship is also apparent. This was clear in the switch in the relationship between the SOI and West Antarctica precipitation around 1990. In addition, Carleton (1988) noted that the regional signal of ENSO in the atmospheric flow, and possibly the signal in the Weddell Sea sea ice data, may have become more marked since the early 1950s. This is consistent with the findings of Mo and van Loon (1984) and further reflects how research is hampered by the short records that are available.

Ice-core data obviously have great potential value to aid in the understanding of the Antarctic expression of ENSO, but drilling sites must be chosen with care since the regional impact of ENSO may change with time. West Antarctica is one of the most promising areas, but the relatively large annual total of precipitation makes obtaining long time series difficult, especially if they are collected near to the coast. Perhaps one of the biggest questions to be answered is how the signals of ENSO can be transferred from the tropical Pacific to interior sites in Antarctica. Climate models should be able to help in this work, along with the atmospheric reanalysis data sets that are to be produced over the coming decade.

To date, many investigations have been looking for simple linear relationships between indicators of ENSO, such as the SOI, and climatic parameters that can be measured in the Antarctic. However, the coupled atmosphere–ocean–sea ice system is highly non-linear, and this is bound to be reflected in the links between ENSO and the climate of the high-latitude areas of the Southern Hemisphere. Modern analysis tools applied to climate model output and reanalysis data sets should advance our understanding of such links and aid in the interpretation of ice-core data.

ACKNOWLEDGEMENTS

I am grateful to Dr G. J. Marshall for preparing the charts showing the upper-air height anomalies over the Southern Ocean during different phases of ENSO. Dr S. Harangozo provided valuable comments on an early draft of this paper.

REFERENCES

Bals-Elsholz TM, Atallah EH, Bosart LF, Cempa MJ, Lupo AR, Wasula TA. 1999. Climatology of the cool season Southern Hemisphere split jet. In *Preprint Volume, 17th Conference on Weather Analysis and Forecasting*, 13–17 September 1999, Denver, CO; 58–59.

Bigg GR. 1990. El Niño and the southern oscillation. *Weather* **45**: 2–8.

Cai W, Baines PG. 2001. Forcing of the Antarctic circumpolar wave by El Niño–southern oscillation teleconnections. *Journal of Geophysical Research–Oceans* **106**: 9019–9038.

Carleton AM. 1988. Sea ice–atmosphere signal of the southern oscillation in the Weddell Sea, Antarctica. *Journal of Climate* **1**: 379–388.

Carleton AM. 1989. Antarctic sea — ice relationships with indices of the atmospheric circulation of the Southern Hemisphere. *Climate Dynamics* **3**: 207–220.

Carleton AM, Carpenter DA. 1990. Satellite climatology of 'polar lows' and broadscale climatic associations for the Southern Hemisphere. *International Journal of Climatology* **10**: 219–246.

Carleton AM, Whalley D. 1988. Eddy transport of sensible heat and the life history of synoptic systems: a statistical analysis for the Southern Hemisphere winter. *Meteorology and Atmospheric Physics* **38**: 140–152.

Chang C. 1993. Large-scale variability of atmospheric deep convection in relation to sea surface temperatures in the tropics. *Journal of Climate* **6**: 1898–1913.

Chen B, Smith SR, Bromwich DH. 1996. Evolution of the tropospheric split jet over the South Pacific Ocean during the 1986–89 ENSO cycle. *Monthly Weather Review* **124**: 1711–1731.

Chiu LS. 1983a. Antarctic sea ice variations 1973–1980. In *Variations in the Global Water Budget*, Street-Perrot A, Beran M, Ratcliffe R (eds). Reidel Publishing: Dordrecht; 301–311.

Chiu LS. 1983b. Variation of Antarctic sea ice: an update. *Monthly Weather Review* **111**: 578–580.
Connolley WM, Harangozo SA. 2001. A comparison of five numerical weather prediction analysis climatologies in southern high latitudes. *Journal of Climate* **14**: 30–44.
Cullather RI, Bromwich DH, Van Woert ML. 1996. Interannual variations in Antarctic precipitation related to El Nino southern oscillation. *Journal of Geophysical Research* **101**: 19 109–19 118.
Diaz H, Markgraf V (eds). 1992. *El Niño: Historical and Palaeoclimatic Aspects of the Southern Oscillation*. Cambridge University Press: Cambridge.
Dibb JE, Talbot RW, Scheuer EM, Blake DR, Blake NJ, Gregory GL, Sachse GW, Thornton DC. 1999. Aerosol chemical composition and distribution during the Pacific Exploratory Mission (PEM) tropics. *Journal of Geophysical Research* **104**: 5785–5800.
Fitch M, Carleton AM. 1992. Antarctic mesoscale regimes from satellite and conventional data. *Tellus A* **44**: 180–196.
Fletcher J, Radok U, Slutz R. 1982. Climatic signals of the Antarctic Ocean. *Journal of Geophysical Research* **87**: 4269–4276.
Gibson JK, Hernandez A, Kållberg P, Nomura A, Serrano E, Uppala S. 1996. Current status of the ECMWF re-analysis project. In *Proceedings of the Seventh Conference on Global Change Studies*. American Meteorological Society: Boston; 112–115.
Gloersen P. 1995. Modulation of hemispheric sea-ice cover by ENSO events. *Nature* **373**: 503–506.
Graham NE, Barnett TP. 1987. Sea surface temperature, surface wind divergence and convection over tropical oceans. *Science* **23**: 657–659.
Hao CJ, Zhang L, Xue H, Xie SM. 1990. Antarctic sea ice and ENSO event. *Acta Oceanologia Sinica* **12**: 549–561 (in Chinese).
Harangozo SA. 1997. Atmospheric meridional circulation impacts on contrasting winter sea ice extent in two years in the Pacific sector of the Southern Ocean. *Tellus* **49**(3): 388–400.
Harangozo SA. 2000. A search for ENSO teleconnections in the west Antarctic Peninsula climate in Austral winter. *International Journal of Climatology* **20**: 663–679.
Heap JA. 1965. Antarctic pack ice. In *Antarctica*, Hatherton T (ed.). Praeger: 187–196.
Heinemann G. 1990. Mesoscale vortices in the Weddell Sea region (Antarctica). *Monthly Weather Review* **118**: 779–793.
Held IM, Lyons SW, Nigam S. 1989. Transients and the extratropical response to El Niño. *Journal of the Atmospheric Sciences* **46**: 163–174.
Hoerling M, Kumar A. 1997. Origins of extreme climate states during the 1982/1983 ENSO winter. *Journal of Climate* **10**: 2859–2870.
Hoskins BJ, Karoly DJ. 1981. The steady linear response of a spherical atmosphere to thermal and orographic forcing. *Journal of the Atmospheric Sciences* **38**: 1179–1196.
Houseago-Stokes RE, McGregor GR. 2000. Spatial and temporal patterns linking southern low and high latitudes during South Pacific warm and cold events. *International Journal of Climatology* **20**: 793–801.
Houseago R, McGregor GR, King JC, Harangozo SA. 1998. Climate anomaly wave train patterns linking southern low and high latitudes during South Pacific warm and cold events. *International Journal of Climatology* **18**: 1181–1193.
Jacobs SS, Comiso JC. 1997. Climate variability in the Amundsen and Bellingshausen Seas. *Journal of Climate* **10**: 697–709.
James IN. 1988. On the forcing of planetary scale Rossby waves by Antarctica. *Quarterly Journal of the Royal Meteorological Society* **114**: 619–637.
Jones DA, Simmonds I. 1994. A climatology of Southern Hemisphere anticyclones. *Climate Dynamics* **10**: 333–348.
Kalnay E, Kanamitsu M, Kistler R, Collins W, Deaven D, Gandin L, Iredell M, Saha S, White G, Woollen J, Zhu Y, Chelliah M, Ebisuzaki W, Higgins W, Janowiak J, Mo KC, Ropelewski C, Wang J, Leetmaa A, Reynolds R, Jenne R, Joseph D. 1996. The NCEP/NCAR 40-year reanalysis project. *Bulletin of the American Meteorological Society* **77**: 437–471.
Karoly DJ. 1989. Southern Hemisphere circulation features associated with El Niño–southern oscillation events. *Journal of Climate* **2**: 1239–1252.
Kiladis GN, Diaz H. 1989. Global climatic anomalies associated with extremes in the southern oscillation. *Journal of Climate* **2**: 1069–1090.
King JC. 1994. Recent climate variability in the vicinity of the Antarctic Peninsula. *International Journal of Climatology* **14**: 357–369.
King JC, Harangozo SA. 1998. Climate change in the western Antarctic Peninsula since 1945: observations and possible causes. *Annals of Glaciology* **27**: 571–575.
King JC, Turner J. 1997. *Antarctic Meteorology and Climatology*. Cambridge University Press: Cambridge.
Kitoh A. 1994. Tropical influence on the South Pacific double jet variability. In *Proceedings of the NIPR Symposium on Polar Meteorology and Glaciology 8*. NIPR: Tokyo; 34–45.
Klein SA, Soden BJ, Lau N-C. 1999. Remote sea surface temperature variations during ENSO: evidence for a tropical atmospheric bridge. *Journal of Climate* **12**: 917–932.
Krishnamurti TN, Chu SH, Iglesias W. 1986. On the sea level pressure of the southern oscillation. *Archiv für Meteorologic, Geophysik und Biokl imatologic, Serie A* **34**: 385–425.
Kwok R, Comiso JC. 2002. Southern Ocean climate and sea ice anomalies associated with the southern oscillation. *Journal of Climate* **15**: 487–501.
Ledley TS, Huang Z. 1997. A possible ENSO signal in the Ross Sea. *Geophysical Research Letters* **24**: 3253–3256.
Legrand M, Feniet-Saigne C. 1991. Methanesulfonic acid in south polar snow layers: a record of strong El Nino? *Geophysical Research Letters* **18**: 187–190.
Legrand M, Feniet-Saigne C, Saltzman ES, Germain C. 1992. Spatial and temporal variations of methanesulfonic acid and non-sea-salt sulfate in Antarctic ice. *Journal of the Atmospheric Sciences* **14**: 245–260.
Lehodey P. 2001. The pelagic ecosystem of the tropical Pacific Ocean: dynamic spatial modelling and biological consequences of ENSO. *Progress in Oceanography* **49**: 439–468.
Li ZX. 2000. Influence of tropical Pacific El Niño on the SST of the Southern Ocean through atmospheric bridge. *Geophysical Research Letters* **27**: 3505–3508.
Limpasuvan V, Hartmann DL. 1999. Eddies and the annular modes of climate variability. *Geophysical Research Letters* **26**: 3133–3136.
Marshall GJ, Harangozo SA. 2000. An appraisal of NCEP/NCAR reanalysis MSLP viability for climate studies in the South Pacific. *Geophysical Research Letters* **27**: 3057–3060.

Marshall GJ, Turner J, Miners WD. 1998. Interpreting recent accumulation records through an understanding of the regional synoptic climatology: an example from the southern Antarctic Peninsula. *Annals of Glaciology* **27**: 610–616.
McKinnell SM, Brodeur RD, Hanawa K, Hollowed AB, Polovina JJ, Zhang CI. 2001. An introduction to the Beyond El Nino Conference: climate variability and marine ecosystem impacts from the tropics to the Arctic. *Progress in Oceanography* **49**: 1–6.
Meehl GA. 1991. A reexamination of the mechanism of the semiannual oscillation in the Southern Hemisphere. *Journal of Climate* **4**: 911–925.
Meyerson EA. 1999. Amundsen–Ross Sea ice variability and the extratropical expression of ENSO as recorded in a South Pole glaciochemical series 1. University of New Hampshire.
Mitchell JF, Hills TS. 1986. Sea-ice and the Antarctic winter circulation: a numerical experiment. *Quarterly Journal of the Royal Meteorological Society* **112**: 953–969.
Mo KC, Higgins RW. 1997. The Pacific–South American modes and tropical convection during the Southern Hemisphere winter. *Monthly Weather Review* **126**: 1581–1596.
Mo KC, van Loon H. 1984. Some aspects of the interannual variation of mean monthly sea level pressure on the Southern Hemisphere. *Journal of Geophysical Research* **89**: 9541–9546.
Mo KC, White GH. 1985. Teleconnections in the Southern Hemisphere. *Monthly Weather Review* **113**: 22–37.
Mo KC, Pfaendtner J, Kalnay E. 1987. A GCM study on the maintenance of the June 1982 blocking in the Southern Hemisphere. *Journal of the Atmospheric Sciences* **44**: 1123–1142.
Murphy EJ, Clarke A, Symon C, Priddle J. 1995. Temporal variation in Antarctic sea-ice: analysis of a long term fast-ice record from the South Orkney Islands. *Deep Sea Research* **42**: 1045–1062.
Noone D, Turner J, Mulvaney R. 1999. Atmospheric signals and characteristics of accumulation in Dronning Maud Land, Antarctica. *J. Geophys. Res.* **104**(D16): 19 191–19 212.
Pan YH, Oort AH. 1983. Global climate variations connected with sea surface temperature anomalies in the eastern equatorial Pacific Ocean for the 1958–1973 period. *Monthly Weather Review* **111**: 1244–1258.
Parker DE. 1983. Documentation of a southern oscillation index. *Meteorological Magazine* **112**: 184–188.
Peng G, Wang PK. 1989. Influence of the Antarctic sea-ice on the north-west Pacific subtropical high and its background of ocean–atmosphere circulation. *Chinese Science Bulletin* **34**: 1463–1465.
Peterson RG, White WB. 1998. Slow oceanic teleconnections linking the Antarctic circumpolar wave with the tropical El Niño–southern oscillation. *Journal of Geophysical Research* **103**: 24 573–24 583.
Philander SG. 1983. El Nino southen oscillation phenomena. *Nature* **302**: 295–301.
Philander SG. 1989. *El Niño, La Niña, and the Southern Oscillation*. Academic Press: New York.
Philander SG, Rasmussen EM. 1985. The southern oscillation and El Nino. *Advances in Geophysics A* **28**: 197–215.
Pittock AB. 1984. On the reality, stability and usefulness of Southern Hemisphere teleconnections. *Australian Meteorological Magazine* **32**: 75–82.
Quinn WH, Neal VT, Antunez de Mayolo SE. 1987. El Niño occurrences over the past four and a half centuries. *Journal of Geophysical Research* **92**: 14 449–14 461.
Rasmusson EM. 1991. Observational aspects of ENSO cycle teleconnections. In *Teleconnections Linking Worldwide Climate Anomalies*, Glantz MH, Katz RW, Nicholls N (eds). Cambridge University Press: Cambridge; 308–343.
Rasmussen EA, Turner J (eds). 2003. *Polar Lows: Mesoscale Weather Systems in the Polar Regions*. Cambridge University Press: Cambridge.
Rasmussen EM, Wallace JM. 1983. Meteorological aspects of the El Niño/southern oscillation. *Science* **222**: 1195–1202.
Renfrew IA, King JC, Markus T. 2002. Coastal polynyas in the southern Weddell Sea: variability in the surface energy budget. *Journal of Geophysical Research–Oceans* **107**: C6. DOI:10.1029/2000JC000720.
Renwick JA. 1998. ENSO-related variability in the frequency of South Pacific blocking. *Monthly Weather Review* **126**: 3117–3123.
Ropelewski CF, Halpert MS. 1987. Global and regional scale precipitation patterns associated with the El Niño/southern oscillation phenomena. *Monthly Weather Review* **115**: 1606–1626.
Ruttlant J, Fuenzalida H. 1991. Synoptic aspects of the central Chile rainfall variability associated with the southern oscillation. *International Journal of Climatology* **11**: 63–76.
Saigne C, Legrand M. 1987. Methanesulfonic acid in Antarctic ice. *Nature* **330**: 240–242.
Savage ML, Stearns CR, Weidner GA. 1988. The southern oscillation signal in Antarctica. In *Preprint Volume Second Conference on Polar Meteorology and Oceanography*, Madison WI, 29–31 March. AMS: Boston; 141–144.
Shiotani M. 1990. Low-frequency variations of the zonal mean state of the Southern Hemisphere troposphere. *Journal of the Meteorological Society of Japan* **68**: 461–471.
Simmonds I, Dix M. 1986. The circulation changes induced by the removal of Antarctic sea ice in a July general circulation model. In *Preprints Volume, Second International Conference on Southern Hemisphere Meteorology*, Wellington, New Zealand, 1–5 December. AMS: Boston; 107–110.
Simmonds I, Jacka TH. 1995. Relationships between the interannual variability of Antarctic sea ice and the southern oscillation. *Journal of Climate* **8**: 637–647.
Sinclair MR. 1996. A climatology of anticyclones and blocking for the Southern Hemisphere. *Monthly Weather Review* **124**: 245–263.
Smith SR, Stearns CR. 1993a. Antarctic climate anomalies surrounding the minimum in the southern oscillation index. In *Antarctic Meteorology and Climatology: Studies Based on Automatic Weather Stations*, Bromwich DH, Stearns CR (eds). Antarctic Research Series, Volume 61. AGU: 149–174.
Smith SR, Stearns CR. 1993b. Antarctic pressure and temperature anomalies surrounding the minumum in the southern oscillation index. *Journal of Geophysical Research* **98**(D7): 13 071–13 083.
Testa JW, Oehlert G, Ainsley DG, Bengtson JL, Siniff DB, Laws RM, Rounsevell D. 1991. Temporal variability in Antarctic marine ecosystems — periodic fluctuations in the phocid seals. *Canadian Journal of Fisheries and Aquatic Sciences* **48**: 631–639.
Thompson DWJ, Wallace JM. 2000. Annular modes in the extratropical circulation. Part I: month-to-month variability. *Journal of Climate* **13**: 1000–1016.

Trenberth KE. 1975a. A quasi-biennial standing wave in the Southern Hemisphere and interrelations with sea surface temperature. *Quarterly Journal of the Royal Meteorological Society* **101**: 55–74.
Trenberth KE. 1975b. Reply to comment on 'A quasi-biennial starting wave in the Southern Hemisphere and interrelations with sea surface temperature' by K. E. Trenberth. *Quarterly Journal of the Royal Meteorological Society* **101**: 174–176.
Trenberth KE. 1976. Spatial and temporal variations of the southern oscillation. *Quarterly Journal of the Royal Meteorological Society* **102**: 639–653.
Trenberth KE. 1997. The definition of El Niño. *Bulletin of the American Meteorological Society* **78**: 2771–2777.
Trenberth KE, Hoar TJ. 1996. The 1990–1995 El Niño–southern oscillation event: longest on record. *Geophysical Research Letters* **23**: 57–60.
Trenberth KE, Shea DJ. 1987. On the evolution of the southern oscillation. *Monthly Weather Review* **115**: 3078–3096.
Turner J, Thomas JP. 1994. Summer-season mesoscale cyclones in the Bellingshausen–Weddell region of the Antarctic and links with the synoptic-scale environment. *International Journal of Climatology* **14**: 871–894.
Van Loon H, Shea DJ. 1985. The southern oscillation part IV: the precursors south of 15 °S to the extremes of the oscillation. *Monthly Weather Review* **113**: 2063–2074.
Van Loon H, Shea DJ. 1987. The southern oscillation, VI, anomalies of sea level pressure on the Southern Hemisphere and of Pacific sea surface temperature during the development of a warm event. *Monthly Weather Review* **115**: 370–379.
Vergani DF, Stanganelli ZB, Bilenca D. 2001. Weaning mass variation of southern elephant seals at King George Island and its possible relationship with 'El Nino' and 'La Nina' events. *Antarctic Science* **13**: 37–40.
Walker GT. 1923. Correlation in seasonal variations of weather VIII. Memorandum of the Indian Meteorological Department. Indian Meteorological Department; 75–131.
Walker GT. 1924. Correlation in seasonal variations of weather IX. Memorandum of the Indian Meteorological Department. Indian Meteorological Department.
Ward MN. 1998. Diagnosis and short-lead time prediction of summer rainfall in tropical North Africa at interannual and multidecadal timescales. *Journal of Climate* **11**: 3167–3191.
Weatherly JW, Walsh JE, Zwally HJ. 1991. Antarctic sea ice variations and seasonal air temperature relationships. *Journal of Geophysical Research* **96**: 15 119–15 130.
Welch KA, Mayewski PA, Whitlow SI. 1993. Methanesulfonic acid in coastal Antarctic snow related to sea-ice extent. *Geophysical Research Letters* **20**: 443–446.
White WB, Peterson RG. 1996. An Antarctic circumpolar wave in surface pressure, wind, temperature and sea-ice extent. *Nature* **380**: 699–702.
Xie S, Bao C, Xue Z, Zhang L, Hao C. 1994. Interaction between Antarctic sea ice and ENSO events. In *Proceedings of NIPR Symposium on Polar Meteorology and Glaciology*; 95–110.
Yuan XJ, Martinson DG. 2000. Antarctic sea ice extent variability and its global connectivity. *Journal of Climate* **13**: 1697–1717.
Zwally HJ, Parkinson CL, Comiso JC. 1983. Variability of Antarctic sea ice and changes in carbon dioxide. *Science* **220**: 1005–1012.

