Teleconnections between the tropical Pacific and the Amundsen-Bellinghausens Sea: Role of the El Niño/Southern Oscillation

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[1] Tropical Pacific–high southern latitude teleconnections are shown to be caused by Rossby wave dynamics and are sensitive to the exact pattern of sea surface temperature (SST) anomalies forcing anomalous ascent. Further, the signal becomes obscured by local natural variability at higher southern latitudes. Here it is shown that a Rossby wave with a source in the tropical Pacific can explain the general pattern of teleconnection found during the southern winter. Upper level divergence associated with deep convection in the tropical Pacific plays a major role in the generation of these Rossby waves. SST anomalies associated with El Niño/Southern Oscillation force the tropical deep convection. However, the relationship between deep convection and SST anomalies is complex and modeling experiments show that in areas of large-scale subsidence positive SST anomalies would have to be larger than any observed anomalies (>4°C) before they have any effect on the vertical motion. Large variations are observed between the high-latitude response to El Niño with apparently similar tropical forcings. This makes accurate prediction of the high-latitude response to tropical change difficult. Ensemble GCM runs show that variations due to the natural variation of the zonal flow in the Southern Hemisphere can swamp the signal resulting from changes in the Rossby wave source region.

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1. Introduction

[2] The teleconnection between El Niño–Southern Oscillation (ENSO) sea surface temperature (SST) variability in the tropical Pacific and the atmospheric circulation in the Amundsen–Bellingshausen Sea (ABS) is well known [*Chen et al.*, 1996; *Karoly*, 1987; *Liu et al.*, 2002; *Turner*, 2004; *Harangozo*, 2000]. In most of these studies the temperature in the El Niño 3.4 region [*Trenberth*, 1997] has been correlated with a variety of parameters in the ABS region. Taking the ERA40 data set, a positive correlation (0.5) is found between the El Niño 3.4 SST and the pressure in the center of the ABS that is significant at the 95% level, while a weak negative correlation (−0.23) is found between the SST in the El Niño 3.4 region and the 1.5 m air temperature on the west coast of the Peninsula during the winter, which is not significant at the 90% level. These two observations are consistent with each other as higher pressures in the ABS mean a more southerly flow close to the west side of the Peninsula, resulting in more cold air being advected over the area. However, although the correlation with pressure is significant climatologically over several ENSO cycles, in individual years the connection between tropics and high latitudes is often weak or missing [*Turner*, 2004].

[3] Two mechanisms have been proposed in recent years to explain this teleconnection. *Liu et al.* [2002], *Yuan* [2004], and *Chen et al.* [1996] explain the connection in terms of the circulation in the Hadley and Ferrel cells. They suggest that with increased convection, during El Niño events, in the tropics the Hadley circulation intensifies and this leads to an increase in the speed of the subtropical jet, which moves equatorward. This in turn results in an increased Ferrel circulation resulting in an increased poleward eddy heat flux and the storm tracks moving equatorward. So the storm activity in the ABS decreases leading to higher pressure in the region.

[4] The other mechanism considers the connection in terms of Rossby waves generated by the tropical SST anomaly [*Karoly*, 1989; *Hoskins and Karoly*, 1981; *Renwick and Revell*, 1999]. Figure 4 in the work of *Trenberth et al.* [1998] shows an idealized Rossby wave in the Northern Hemisphere with a wave train of high and low pressure propagating poleward from a source linked to increased upper level divergence over an SST anomaly on the equator. The geometry is such that when the wave propagates into the Southern Hemisphere from a source in the midtropical Pacific, a high-pressure anomaly forms over the ABS. Using the theory outlined in the work of *James* [1994], it can be shown that a source in the tropics will produce a wave that propagates through in a southeasterly direction, in the troposphere, until it reaches a critical latitude close to the edge of the Antarctic continent where it is reflected. In the Northern Hemisphere the climatological stationary planetary

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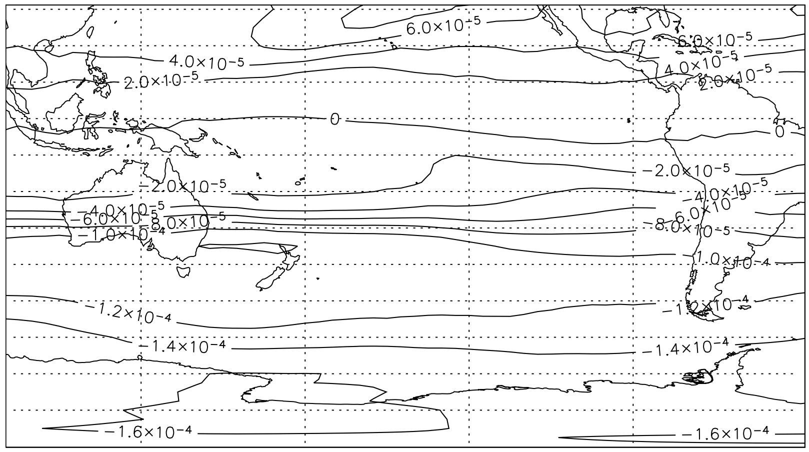

Figure 1. The absolute vorticity at 300 hPa for the Southern Hemisphere winter months (JJA) from the Hadley Centre atmosphere-only model (HadAM3) forced with yearly repeating SSTs.

wave, associated with the large continental land masses, can make the total Rossby wave sources somewhat insensitive to the position of the tropical heating and can create preferred response patterns, such as the Pacific-North American pattern that are quite repeatable [*Trenberth et al.*, 1998]. Similar patterns in the Southern Hemisphere are less stable, because of the lack of large continents but still produce repeatable climatological teleconnections.

[5] In this paper we investigate the source region of the Rossby waves found in the Southern Hemisphere winter (JJA). It has been suggested that the source of the Rossby waves is not the area of highest SST anomaly associated with an El Niño [*Harangozo*, 2004]. We look closely at the relationship between SST anomalies and vertical motion in the tropical Pacific and use Rossby wave theory to explain some of the observed variability of the ENSO−ABS circulation teleconnection.

2. Models and Model Data

[6] For the case studies the analysis carried out by the European Centre for Medium-range Weather Forecasting (ECMWF) reanalysis project (ERA-40) have been used. Since the late 1970s, when satellite sounder data became available, ERA-40 closely follows Southern Hemisphere observations showing more skill than the National Centers for Environmental Prediction-National Center for Atmospheric Research (NCEP-NCAR) reanalysis [*Bromwich and Fogt*, 2004; *Connolley and Harangozo*, 2001].

[7] For the climate model runs carried out for this study we use the atmosphere-only version of the Hadley Centre General Circulation Model (GCM) [*Pope et al.*, 2000] known as HadAM3. This model is similar to earlier versions, but has improved physics. The model has 19 levels in the vertical and a $2.5^\circ \times 3.75^\circ$ horizontal grid. It has been forced at the lower boundary by imposing the sea surface temperature and sea ice concentration derived from the Global sea-Ice and Sea Surface Temperature (GISST) data set [*Rayner et al.*, 1996]. This data set stretches back to the 19th century but it is most accurate in polar regions in the satellite era (after 1978). This limitation is one of the reasons this study does not consider any El Niños before 1983. HadAM3 reproduces the circulation found in ERA-40 well after 1979 [*Turner et al.*, 2006]. This study only considers the Austral winter (JJA) and during this season the interannual variability of HadAM3 is similar to that of ERA-40 [*Lachlan-Cope et al.*, 2001].

3. Theory

3.1. Rossby Wave Source

[8] The vorticity equation can be written as

$$\frac{\partial \boldsymbol{\zeta}}{\partial t} + \mathbf{v}_\psi \cdot \nabla \boldsymbol{\zeta} = -\boldsymbol{\zeta} D - \mathbf{v}_\chi \cdot \nabla \boldsymbol{\zeta} \qquad (1)$$

where $\boldsymbol{\zeta}$ is the absolute vorticty, v_ψ is the rotational wind, $\boldsymbol{D}$ is the divergence, and v_χ is the divergent wind [*James*, 1994]. The left-hand side of this equation represents the propagation of Rossby waves, while the right-hand side represents the forcing terms. It can be seen that the forcing will be large in areas where the divergence, divergent wind, absolute vorticity, and gradients in absolute vorticity are large. This can happen poleward of heating on the equator where the upper-level divergence, associated with deep convection, is greatest and large gradients of absolute vorticity exist associated with the subtropical jets [see *James*, 1994, Figure 8.5].

[9] We will consider the terms in the Rossby wave source (RWS) separately to investigate how they will affect the generation of Rossby waves.

3.1.1. Absolute Vorticity Gradient

[10] First, we consider the absolute vorticity. Figure 1 show the absolute vorticity for a control run of the atmosphere only version of the Hadley Centre model. On the equator the magnitude of the absolute vorticity is quite small but rapidly increases farther south. This is particularly true in the western Pacific near Australia. In the eastern Pacific the gradient of the absolute vorticity is less. It could be expected then that the western Pacific, to the east of Australia, would be a source region for Rossby waves while the weaker gradient in absolute vorticity in the eastern Pacific would make this area less prone to the generation of Rossby waves.

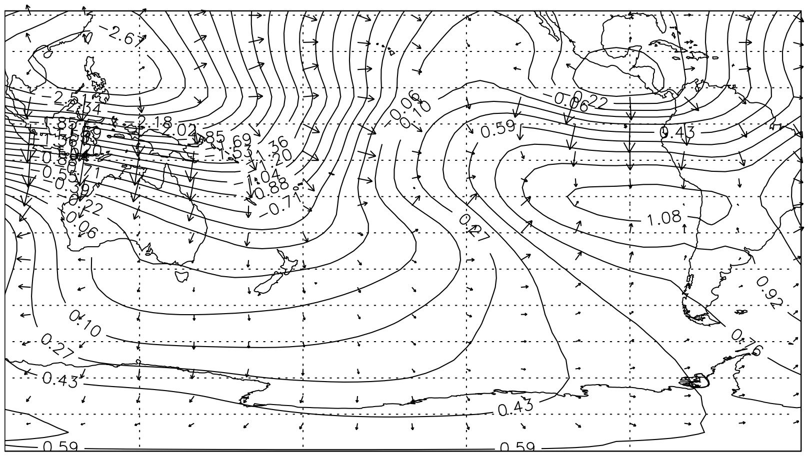

Figure 2. The velocity potential (contours) and divergent component of the wind (arrows) at 300 hPa for the Southern Hemisphere winter months (JJA) from the Hadley Centre atmosphere-only model (HadAM3) as in Figure 1.

3.1.2. Divergent Wind

[11] Now, we consider the upper level divergence and the divergent wind. Figure 2 shows the divergent wind plotted on top of the velocity potential. Generally speaking, areas of ascent are associated with negative velocity potential, while descent is associated with positive values. The divergent wind shows there is strong divergence component in the wind to the east of Australia to the south of the equator. Even though the area of ascent is on, or occasionally to the north of the equator, there is an area of strong divergence some distance to the south of the equator. The area of divergence associated with the ascent is not necessarily collocated with the ascent. This is the same area that has a large gradient in the absolute vorticity making this a favored area for the formation of Rossby waves.

3.2. Impact of SST Anomalies on RWS Terms

[12] The upper level divergence depends on the vertical velocity, in areas of ascent the upper level divergence is positive and it increases as the ascent increases. The correlation between changes in SST and associated changes in vertical velocity is complex. *Graham and Barnett* [1987] reported that SSTs above a critical value of 27.5°C were required for deep convection to occur in the tropical Pacific; however, this value does not hold for all cases and is not a universal constant but depends on the lapse rate, temperature, and circulation in the free troposphere. Also, *Bony et al.* [1997] show that while in some areas increases in SST are associated with increased upward vertical velocity, in other areas there is no correlation. In some areas, increased SSTs are even associated with a decrease in the updraft.

[13] To look at the effect of the SST anomalies associated with ENSO events, it is better to look at the vertical velocity (or omega) than at the upper level divergence as differences are more easily seen. Figure 3 shows omega for the same control run as Figures 1 and 2. In the tropical Pacific it shows a narrow band of ascent (negative omega) on the equator with a large area of descent (positive omega) over the subtropical central and eastern Pacific. A second narrow band of ascent spreads from the equator, close to the dateline, to the southeast, this is known as the South Pacific Convergence Zone (SPCZ). *Bony et al.* [1997] suggests that it should only be in the areas of ascent that we get further anomalous climatological ascent when the SSTs increase. In

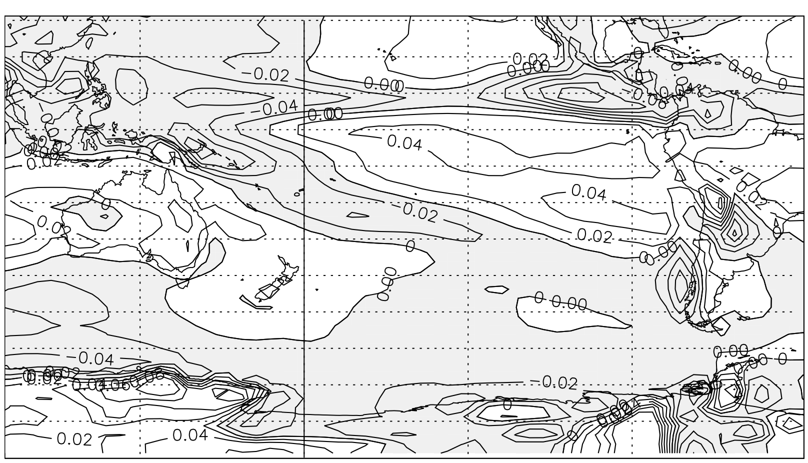

Figure 3. The vertical velocity (omega) at 505 hPa for the Southern Hemisphere winter months (JJA) from the Hadley Centre atmosphere-only model (HadAM3) as in Figure 1.

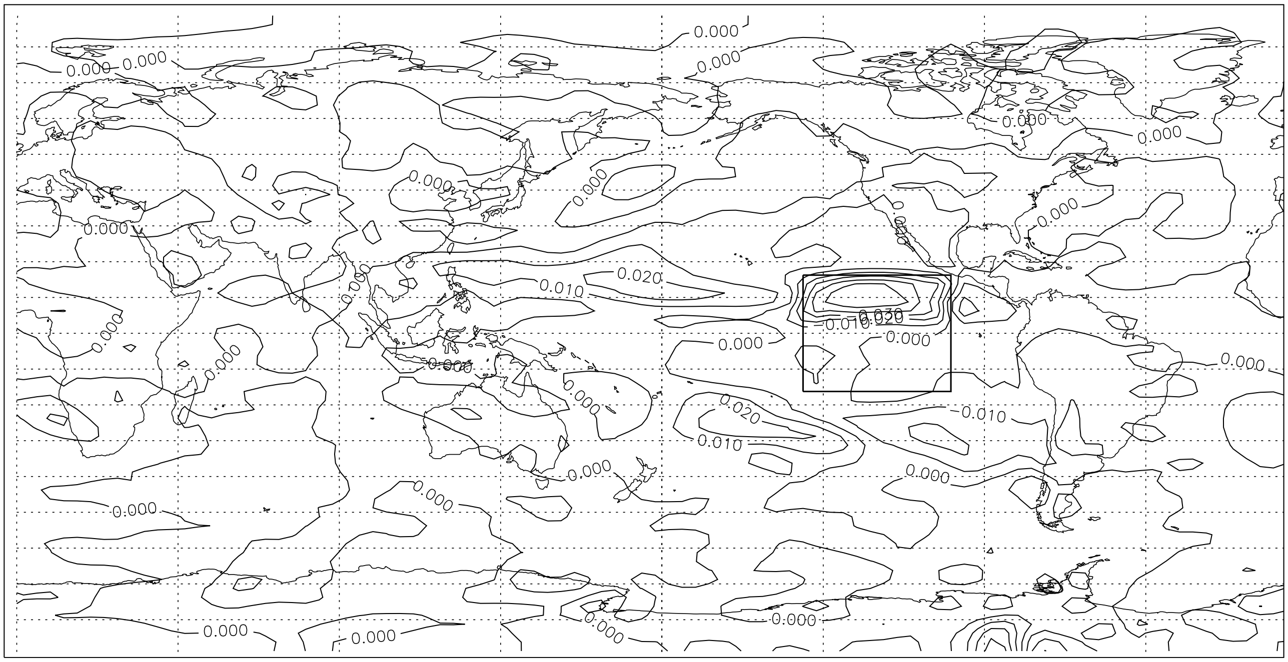

Figure 4. Omega anomalies at 500 hPa for a +2°C SST anomaly in the mid-Pacific (anomaly marked with square).

areas of large-scale subsidence the variations found in the vertical motion are largely independent of local SST changes. In areas where there is large-scale ascent, for example in the ascending branches of the Hadley circulation, a strong correlation is generally found between SST changes and ascent.

[14] To investigate the role of the large-scale vertical velocity in controlling the anomalous vertical velocity, we have used the atmosphere-only version of the Hadley centre climate model (HadAM3) to carry out a series of experiments with SST anomalies imposed in the tropical Pacific. First we put an SST anomaly in a 40 × 30° box centered at 0°S 120°W in an area that includes both climatological ascent and descent. Figure 4 shows the anomalous ascent for this area, for a +2°C anomaly, showing an increased updraft in areas of ascent but very little change in the area of descent.

[15] A further experiment was carried out with the Hadley Centre model to investigate the response in areas of descent. A series of four runs was carried out with anomalies of increasing strength (2, 4, 8, and 16°C) in a 20 × 20° box centered at 15°S 105°W in the area of descent in the eastern Pacific. Figure 5 shows the change in the averaged omega in this box as the size of the SST anomaly is increased. The anomalous ascent is small (less than −0.01 Pa s^{-1} or 5 cm s^{-1}) until the SST anomaly increases to +4°C, an amount that is much greater than any observed ENSO anomaly. After 6°C the ascent increases nearly linearly with increasing temperature.

[16] It appears that deep convection cannot form in the areas of descent in the tropics until the SST anomaly reaches a value that is sufficient for it warm up the boundary layer enough to allow deep convection to form throughout the troposphere. Before this value is reached the boundary layer remains decoupled from the main depth of the atmosphere. This explains the deep convection pattern as observed by satellites in the outgoing longwave radiation (OLR) [*Harangozo*, 2004]. The observed OLR pattern shows deep convection during the winter, in the intertropical convergence zone (ITCZ) and the SPCZ but little deep convection over the area of descent in the Eastern Pacific, even in strong El Niño years.

4. Case Studies

[17] Case studies have been carried out for individual winters during El Niño events using data from the ERA-40 analysis project. Here we look at two winters, 1997 and 1987 that show a clear Rossby wave pattern generated in the tropical Pacific, close to the date line, and propagating into the ABS. We will contrast these two winters with the

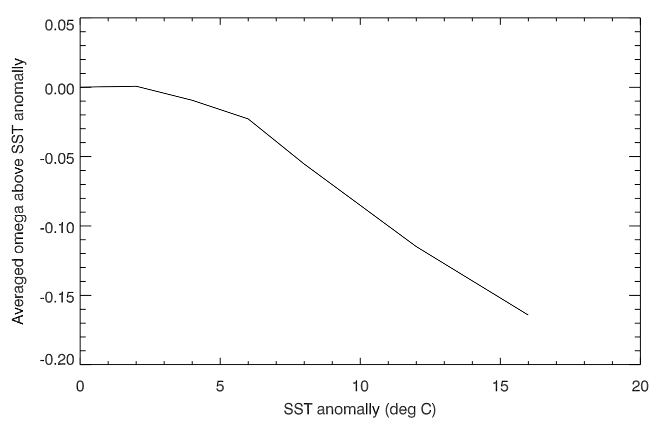

Figure 5. Omega anomalies at 500 hPa plotted against SST anomaly for a series of SST anomalies positioned in the mid-Pacific (15°S 105°W).

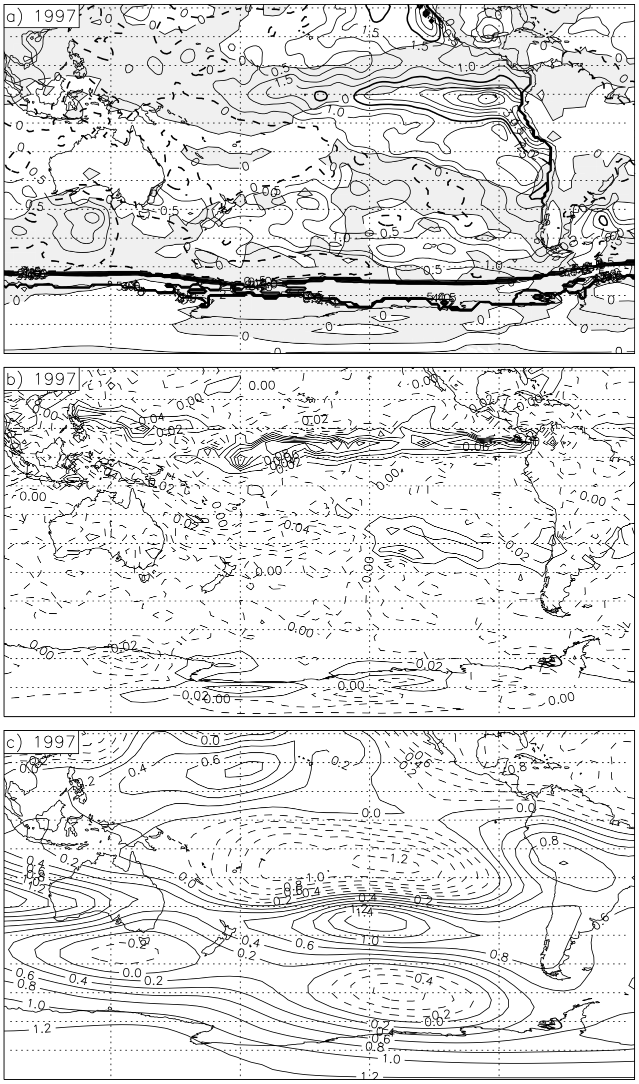

Figure 6. The 1997 winter months (JJA) (a) SST anomalies in the Southern Hemisphere compared to 1979–2000 from the GISST data set; the 0°C contour is dashed, while the 2°C contour is bold. The contour interval for the SST anomalies is 0.5°C Areas of ascent at 500 hPa are shaded. (b) Areas of anomalous ascent (omega) at 500 hPa compared to 1979–2000 from ERA-40. (c) Anomalous stream function at 200 hPa compared to 1979–2000 from ERA-40.

winters of 1983 and 1991 which, although they show some evidence for Rossby wave propagation do not show such clear signals.

4.1. Year 1997

[18] The winter of 1997 (JJA) show a strong El Niño signal in the SST anomaly field (see Figure 6a). A tongue of very warm water extends from the coast of South America to the date line, just south of the equator. Figure 6a also shows the area of ascent at 500 hPa as shaded. Most of this area of warm water is south of the equator, in an area of descent, and so, following the results reported above, does not trigger deep convection. However, a significant part of the warm pool is north of the equator just to the east of the date line, in an area of ascent (shown as shaded in Figure 6a), and here the anomalous SSTs can trigger deep convection. Figure 6b shows that the strongest area of anomalous ascent is just north of the equator stretching across the Pacific in the ITCZ. A weaker area of ascent is seen over the SPCZ further south, while the changes in the areas of descent are rather small. The ascent close to the equator generates the upper level divergence that acts as a source for Rossby wave. The anomalous stream function at 200 hPa (figure 6c) shows a distinct Rossby wave propagating from the tropical mid Pacific south slightly eastward toward the ABS.

4.2. Year 1987

[19] The pattern of surface temperature anomalies in the tropical Pacific for the 1987 winter (Figure 7a) is very different. The spatial extent and magnitude of the El Niño event are smaller (during the winter) than the 1997 event especially in the eastern Pacific close to the coast of South America in the areas of descent (unshaded). However, in the areas of ascent, particularly just to the east of the date line,

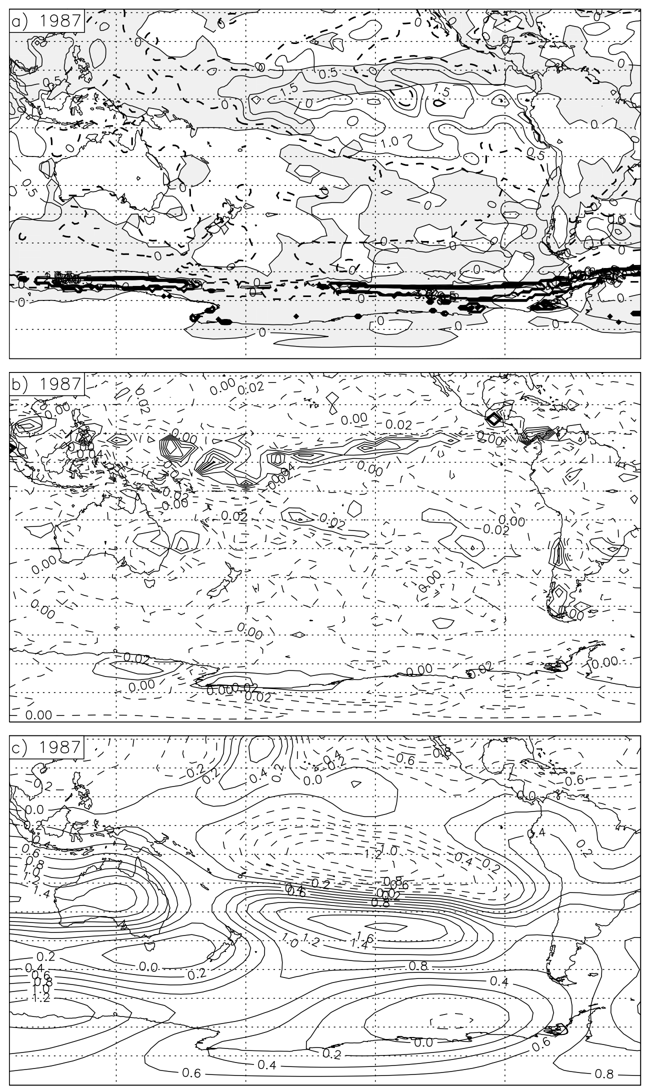

Figure 7. As in Figure 6 but for 1987 winter (JJA) months.

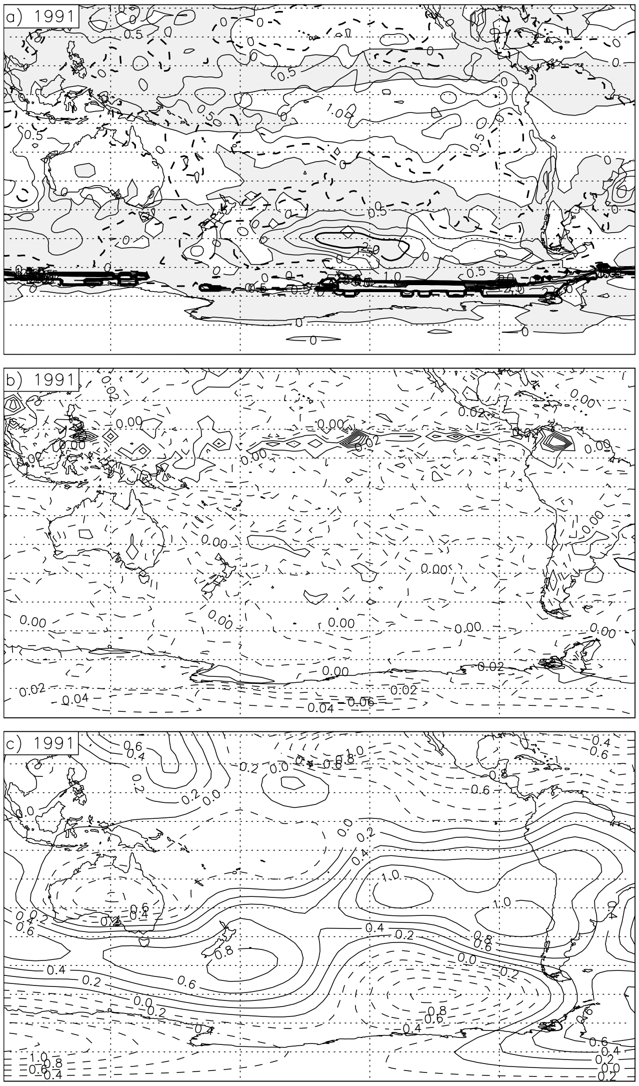

Figure 8. As in Figure 6 but for 1991 winter (JJA) months.

the SST anomalies are very similar to 1997. Figure 7b shows the anomalous ascent during this winter and if it is compared with Figure 6b, which shows the same for 1997, it can be seen that the area of ascent is smaller, and does not extend all the way east across the Pacific. However, the area of ascent in 1987 is equatorward of a region of strong gradient in absolute vorticity and so this area is favored as a Rossby wave source region. The stream function at 200 hPa (Figure 7c) in the 1987 winter shows very similar pattern of Rossby waves to that seen in 1997 with a low in the stream function over the ABS, which corresponds to an area of high pressure at the surface. A strong wave train propagating south from the tropical Pacific toward the ABS can be seen.

4.3. Year 1991

[20] Looking at the SST anomalies in the 1991 winter (Figure 8a) we see that they are generally much smaller than in 1987 or 1997. The anomalous ascent in the 1991 winter (Figure 8b) is also much smaller than in 1987 or 1997 with a thin line of ascent just to the south of the ITCZ. The small area of ascent over mid tropical Pacific may be responsible for the wave train propagating south into the ABS, although the phase of this wave in the tropics is different to that in 1997 and 1987. Also, there is an area of confused ascent to the north of Australia and although it is not clear that this is stronger than in 1987 or 1997, it may be responsible for the wave train propagating from Australia to the ABS. In this winter the observed Rossby wave pattern (Figure 8c) is different to that observed in either 1987 or 1997. The clearest pattern is a wave starting over Australia and curving toward the ABS and Antarctic Peninsula. A weak wave train can be seen starting in the mid-Pacific, but it is not as distinct as in 1987 or 1997.

4.4. Year 1983

[21] In the winter of 1983 the SST anomalies associated with an El Niño event are strong in the eastern Pacific, close to South America (see Figure 9a). However, these anomalies are restricted to the east of the Pacific in an area in which the air is mostly descending. A small part at the north of the SST anomaly is in the area of ascent and in this area

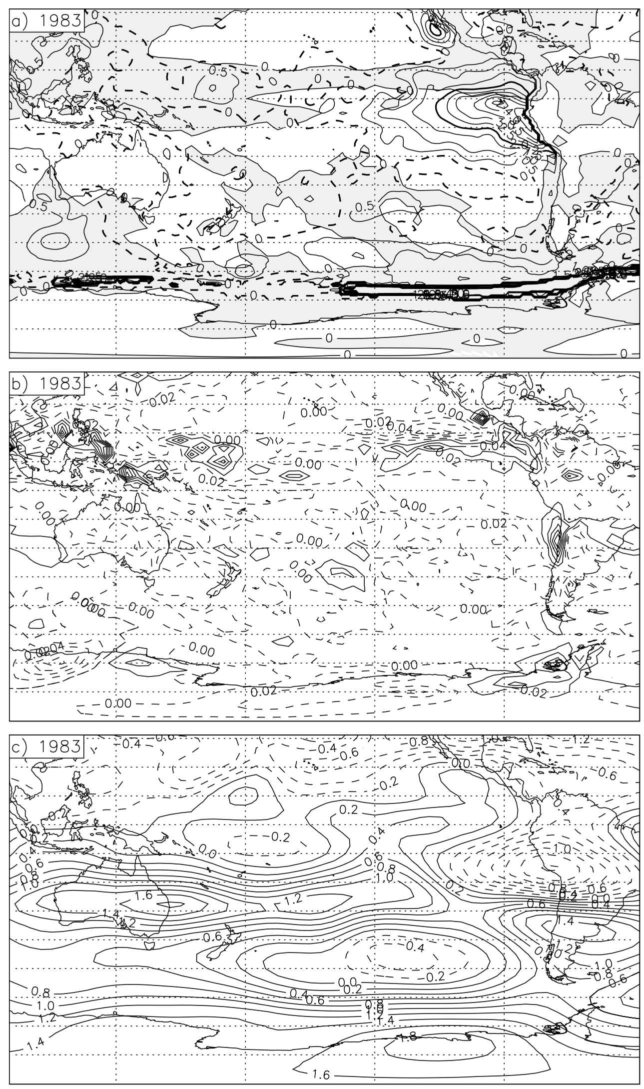

Figure 9. As in Figure 6 but for 1983 winter (JJA) months.

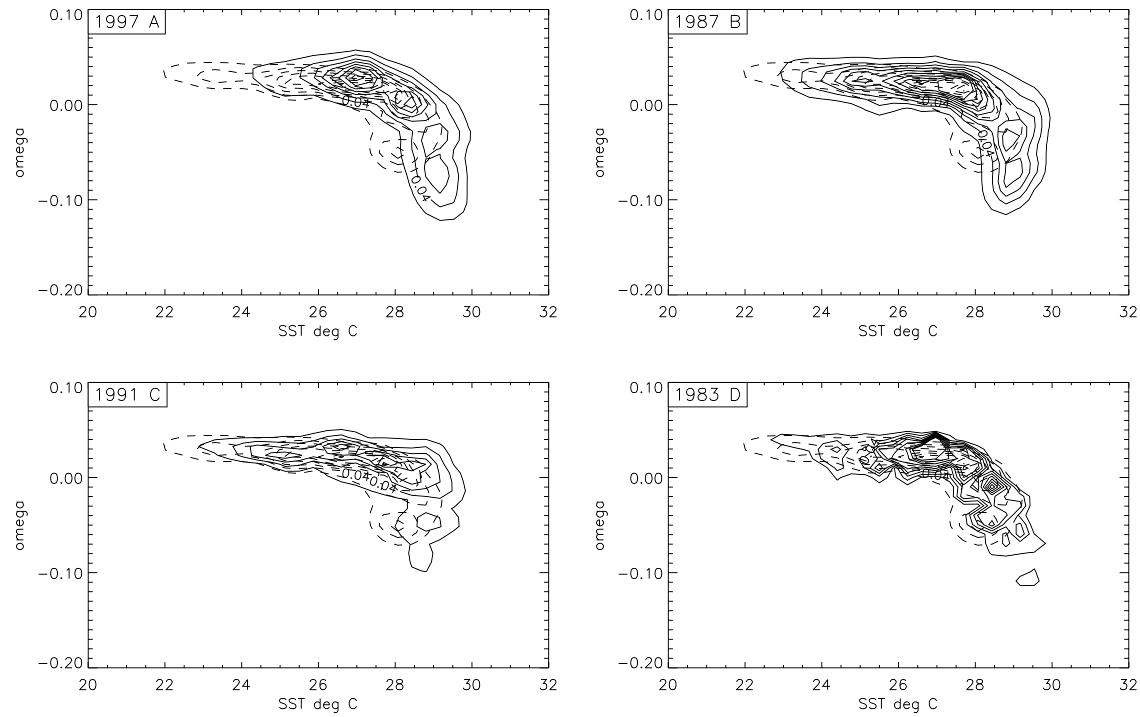

Figure 10. Probability density distribution of SST against vertical ascent at 500 hPa for an area in the tropical Pacific (El Niño 3.4 area) for (a) 1997, (b) 1987, (c) 1991, and (d) 1983.

the anomalous ascent (Figure 9b) is strongest. However, in the eastern Pacific the gradient of absolute vorticity is rather weak and so the Rossby wave source term (equation (1)) will be weak in this area, although a wave can be seen propagating southward across South America toward the Weddell Sea. A weak SST anomaly on the Equator just to the west of the date line is associated with an area of ascent in Figure 9b. Corresponding to the more broken and confused ascent anomalies the streamfunctions show a completely different pattern (Figure 9c) in this year. A wave of smaller magnitude and different phase from 1997 and 1987 can be seen propagating from an area of very weak anomalous ascent in the mid-Pacific toward the ABS.

4.5. Vertical Motion

[22] The results above shows that the generation of Rossby waves in the tropical Pacific is largely controlled by anomalous ascent forced by SST anomalies in areas of large-scale climatological ascent. So the relationship between SSTs and vertical motion is complex and nonlinear. Figure 10 shows the probability distribution of SST against vertical velocity (omega) for an area in the tropical Pacific (170°E to 270°E and 20°S to 20°N) for the four cases considered here. Also plotted for comparison is the probability distribution for the 1979–2000 mean (dashed line). Generally, these plots can be divided into two regions, the first region below a SST of around 27–28°C are areas of descent where changes in SST have little effect on the descent. The other region contains areas of higher SST s and are generally areas of ascent where an increases in SST leads to an increase in ascent. These plots are essentially comparably to distributions of SST and OLR [see *Zhang*, 1993].

[23] Figures 10a and 10b (1987 and 1997) again look very similar. In both cases the whole distribution has moved toward warmer temperatures and the critical temperature at which ascent starts has increased as has the maximum magnitude of the ascent. In 1991 and 1983 (Figures 10c and 10d) although there is some evidence of the distribution moving toward higher SSTs there is little sign of the maximum rate of ascent increasing. The pattern of the whole distribution moving toward warmer SSTs is consistent with observations made above that areas of descent tend to be insensitive to changes of SST while in areas of ascent will tend to increase as SST increases.

4.6. Variability

[24] Although the stream function anomaly pattern for both 1997 and 1987 is similar the effect these anomalies have on the climate of the Antarctic Peninsula is very different. Figure 11 shows the winter ascent in the tropical Pacific (in the El Niño 3.4 area, 5°N to 5°S, 170°–120°W) and the winter temperature at Faraday station plotted against year on the same graph. In 1987 a high-pressure anomaly formed in the ABS allowing a strong southerly flow to form over the Peninsula (see Figure 12a) and so the strong ascent in the tropical Pacific was associated with low Peninsula temperatures. However, this was not true of 1997. Although the ascent in the tropical Pacific was even stronger in 1997 than in 1987 the temperature during the winter at Faraday was not colder than normal. In 1997 the high-pressure anomaly that formed in the ABS (see Figure 12b), associated with the Rossby wave train, was much further to the west and so instead of a strong southerly forming over the Peninsula a weak ridge formed instead.

[25] The question arises if the difference between 1987 and 1997 was due to a difference in the tropical forcing or due to natural variability? The region of the ABS is known to be an area of large variability, even in the absence of

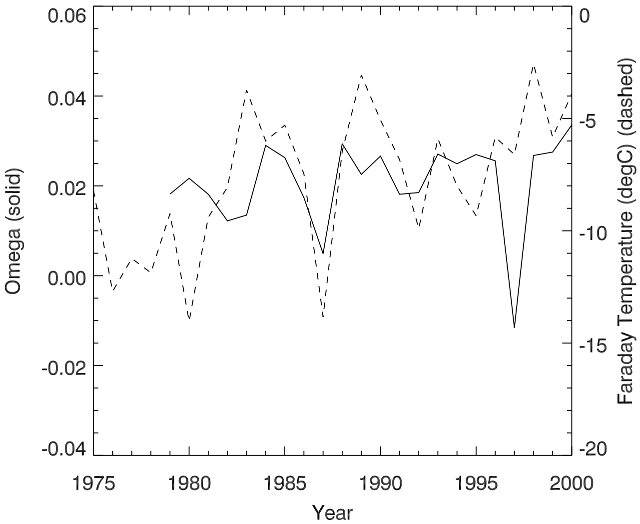

Figure 11. Omega at 500 hPa in the El Niño 3.4 area (in black) from ERA-40 against year and Faraday surface temperature (in red) against year.

tropical SST variations [see *Lachlan-Cope et al.*, 2001; *Connolley*, 1997]. To investigate the variability, an ensemble of nine atmosphere-only model (HadAM3) runs was carried out. The model was forced with observed SSTs for 1996 to 1998 in the tropics (between 20°S and 20°N) and climatology elsewhere. Here we look at the results for winter 1997. Figure 13 shows the standard deviation of the mean sea level pressure between the ensemble members. The variability in the ABS is over 4 hPa in the ABS and, in the area close to the Peninsula, is over twice as large as the pressure anomaly in the same area in Figures 12a and 12b. Hence the noise is larger than the signal we are hoping to observe. The large vacillation in the zonal flow in the Southern Hemisphere [*Hartmann*, 1995] means that it is difficult to predict the exact climatological conditions over the Antarctic Peninsula based just on the propagation of Rossby waves from the tropics. Hence natural variability of the extratropical atmosphere is sufficiently large that attribution of the difference between 1987 and 1997 to forcing terms is not possible.

5. Discussion

[26] The role of winter SST anomalies in the generation of teleconnections at high southern latitudes has been investigated by looking at the generation of Rossby waves by these SST anomalies and the transmission of these waves into the highly variable region of the ABS.

[27] We have shown that Rossby waves are generated in the tropical Pacific as a result of SST anomalies but only if positive SST anomalies occur under areas of climatological ascent in the atmosphere. The area of ascent on the equator, normally known as the ITCZ, is the rising part of the Hadley circulation. In this area, positive SST anomalies force anomalous ascent leading to divergence aloft and hence increasing the Rossby wave source term in equation (1) above. The areas of ascent do not generally coincide with the maximum SST anomalies. In the eastern Pacific, south of the ITCZ in the descending part of the Hadley circulation, modeling studies indicate that increasing SSTs have no effect on vertical motion within the atmosphere, until a threshold value of around 4–5°C is reached. This threshold value is greater than SST anomalies normally observed. The SST anomalies produced by different El Niño events can force Rossby waves in very different ways, despite being of the same magnitude, depending on the location of the anomaly relative to the Hadley circulation.

[28] *Liu et al.* [2002] suggested that the observed teleconnection between the tropical Pacific and ABS are caused by changes in the regional Ferrel call in the southern Pacific that are themselves a consequence of changing meridional eddy heat flux associated with change in the storm tracks. These changes are also associated with changes in the subtropical jets. It is suggested that the climatological changes observed by [*Liu et al.*, 2002] between El Nino and La Nina years are caused by the Rossby waves described in this paper and so it is unnecessary to invoke the complex mechanisms they describe.

[29] Although the generation of Rossby waves by tropical Pacific SST anomalies is relatively straightforward, the

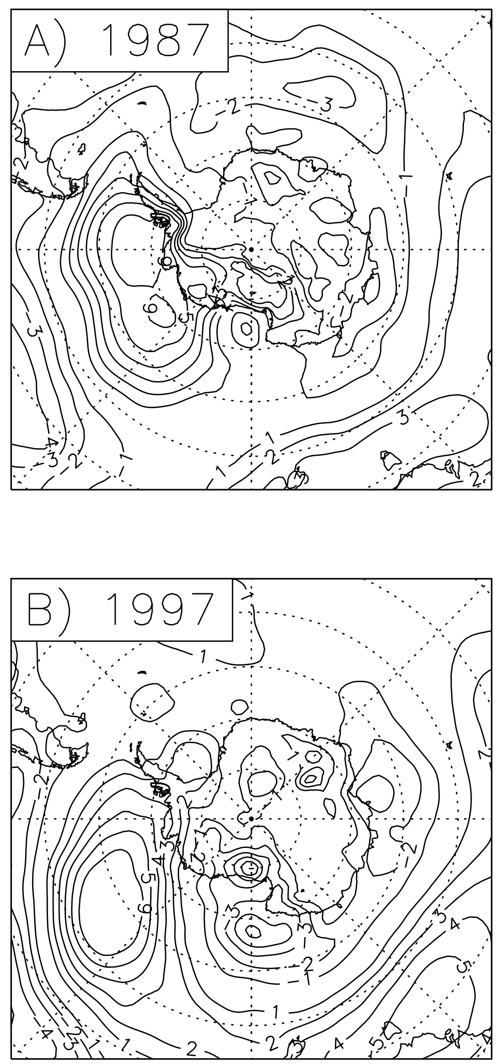

Figure 12. Mean sea level pressure anomaly for (a) winter 1987 from ERA-40 against 1979–2000 mean and (b) winter 1997 from ERA-40 against 1979–2000 mean.

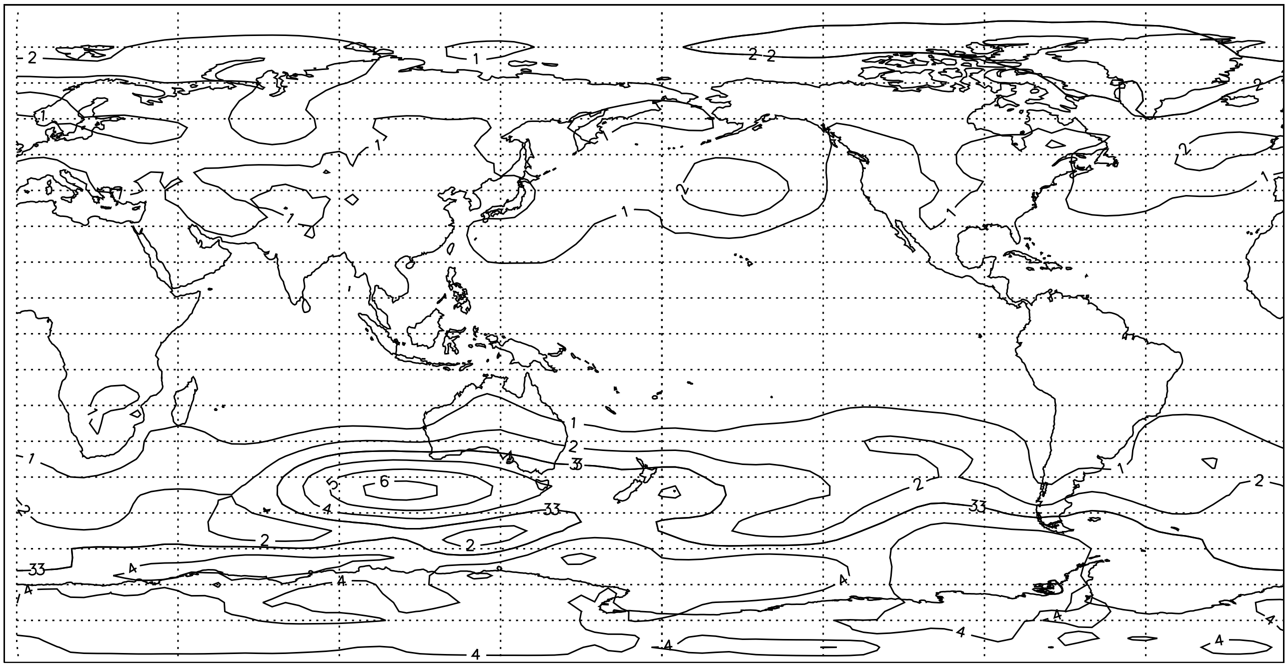

Figure 13. Standard deviation of mean sea level pressure for an ensemble of runs forced with 1997 winter (JJA) tropical SST anomalies.

propagation of the waves into the Southern Hemisphere is dependent on the zonal wind in the South Pacific. In particular, the ABS is an area of extreme variability [*Connolley*, 1997] and it is not surprising that it is difficult to predict the response of the atmosphere at high southern latitudes to tropical forcing. The response of the Northern Hemisphere to forcing in the tropical Pacific is much clearer, with El Niño and tropical convection being linked to the two dominant modes of extratropical atmospheric variability in the Northern Hemisphere, the Pacific/North Atlantic pattern and the North Atlantic Oscillation [*Trenberth et al.*, 1998; *Lin et al.*, 2005]. However in the Northern Hemisphere the large-scale circulation tends to be tied to the distribution of the major land masses and so the vacillation in the zonal flow is much smaller and the propagation of Rossby waves is more predictable.

6. Conclusions

[30] The deep convection associated with upper level divergence that forces Rossby waves in the tropical Pacific occurs in areas away from the greatest SST anomalies. Although the atmospheric response to SST forcing in the tropics is highly nonlinear, modern global climate models can reproduce the observed tropical response well. It is possible to reproduce the tropical part of a Rossby wave train generated by El Niño SST anomalies within a model. However, ENSO events are normally defined by an index such as the El Niño 3.4 index, based on the SST anomaly in a defined area in the tropical Pacific [*Trenberth*, 1997]. However, as the El Niño 3.4 area is largely in an area of climatological descent, SST anomalies in this area do not normally cause deep convection and so do not force Rossby waves. This means that attempts to predict the extra tropical response to El Niño forcing in the ABS by looking at just one area in the tropics are unlikely to succeed. We suggest that it may be possible to achieve greater predictability of the production of Rossby waves by developing an index of SSTs under regions of atmospheric ascent.

[31] The variability of the atmospheric circulation in the southern Pacific is such that the propagation of Rossby waves is very variable and it is this, more than variations in the forcing of the waves, that means that the correlation between tropical SSTs and meteorological parameters close to the west side of the Antarctic Peninsula are low during the southern winter (JJA).

[32] This study has only considered the generation and propagation of Rossby waves during the winter as it is during this time that conditions are more favorable for these waves to be formed and to propagate to high southern latitudes. Further studies should investigate these waves during other seasons.

References

Bony, S., K. M. Lau, and Y. C. Sud (1997), Sea surface temperature and large-scale circulation influences on tropical greenhouse effect and cloud radiative forcing, *J. Clim.*, *10*, 2055–2077.

Bromwich, D. H., and R. L. Fogt (2004), Strong trends in the skill of the ERA-40 and NCEP-NCAR reanalysis in the high and midlatitudes of the Southern Hemisphere, 1958–2001, *J. Clim.*, *17*, 4603–4619.

Chen, B., S. R. Smith, and D. H. Bromwich (1996), Evolution of the tropospheric split jet over the South Pacific Ocean during the 1986–89 ENCO cycle, *Mon. Weather Rev.*, *124*, 1711–1731.

Connolley, W. M. (1997), Variability in annual mean circulation in southern high latitudes, *Clim. Dyn.*, *13*, 745–756.

Connolley, W. M., and S. A. Harangozo (2001), A comparison of five numerical weather prediction analysis climatologies in southern high latitudes, *J. Clim.*, *14*, 30–44.

Graham, N. E., and T. P. Barnett (1987), Sea surface temperature, surface wind divergence and convection over tropical oceans, *Science*, *23*, 657–659.

Harangozo, S. (2004), The relationship of Pacfic deep tropical convection to the winter and springtime extratropical atmospheric circulation of the South Pacific in El Niño events, *Geophys. Res. Lett.*, *31*, L05206, doi:10.1029/2003GL018667.

Harangozo, S. A. (2000), A search for ENSO teleconnections in the west Antarctic Peninsula climate in Austral winter, *Int. J. Climatol.*, *20*, 663–679.

Hartmann, D. L. (1995), A PV view of zonal flow vacillation, *J. Atmos. Sci.*, *52*, 2561–2576.

Hoskins, B. J., and D. J. Karoly (1981), The steady linear response of a spherical atmosphere to thermal and orographic forcing, *J. Atmos. Sci.*, *38*, 1179–1196.

James, I. N. (1994), *Introduction to Circulating Atmospheres*, Cambridge Univ. Press, New York.

Karoly, D. J. (1987), Southern Hemisphere temperature trends: A possible greenhouse gas effect?, *Geophys. Res. Lett.*, *14*, 1139–1141.

Karoly, D. J. (1989), Southern Hemisphere circulation features associated with El Niño-Southern Oscillation events, *J. Clim.*, *2*, 1239–1252.

Lachlan-Cope, T. A., W. M. Connolley, and J. Turner (2001), The role of the nonaxisymmetric Antarctic orography in forcing the observed pattern of variability of the Antarctic climate, *Geophys. Res. Lett.*, *28*, 4111–4114.

Lin, H., J. Derome, and G. Brunet (2005), Tropical Pacific link to the two dominant patterns of atmospheric variabilty, *Geophys. Res. Lett.*, *32*, L03801, doi:10.1029/2004GL021495.

Liu, J. P., X. J. Yuan, D. Rind, and D. G. Martinson (2002), Mechanism study of the ENSO and southern high latitude climate teleconnections, *Geophys. Res. Lett.*, *29*(14), 1679, doi:10.1029/2002GL015143.

Pope, V. D., M. L. Gallani, P. R. Rowntree, and R. A. Stratton (2000), The impact of new physical parametrizations in the Hadley Centre climate model: HadAM3, *Clim. Dyn.*, *16*, 123–146.

Rayner, N. A., E. B. Horton, D. E. Parker, C. K. Folland, and R. B. Hackett (1996), Version 2.2 of the Global sea-Ice and Sea Surface Temperature data set, *Hadley Cent. Clim. Res. Tech. Note*, *CRTN 74*, 43 pp.

Renwick, J. A., and M. J. Revell (1999), Blocking over the South Pacific and Rossby wave propagation, *Mon. Weather Rev.*, *127*, 2233–2247.

Trenberth, K. E. (1997), The definition of El Niño, *Bull. Am. Meteorol. Soc.*, *78*, 2771–2777.

Trenberth, K. E., G. W. Branstator, D. J. Karoly, A. Kumar, N.-C. Lau, and C. Ropelewski (1998), Progress during TOGA in understanding and modeling global teleconnections associated with tropical sea surface temperatures, *J. Geophys. Res.*, *103*, 14,291–14,324.

Turner, J. (2004), The El Niño-Southern Oscillation and Antarctica, *Int. J. Climatol.*, *24*, 1–31.

Turner, J., W. M. Connolley, T. A. Lachlan-Cope, and G. J. Marshall (2006), The performance of the Hadley Centre Climate Model (HadCM3) in high southern latitudes, *J. Clim.*, *26*, 91–122.

Yuan, X. J. (2004), ENSO-related impacts on Antarctic sea ice: a synthesis of phenomenon and mechanisms, *Antarct. Sci.*, *16*, 415–425.

Zhang, C. I. (1993), Large-scale variability of atmospheric deep convection in relation to sea surface temperature in the tropics, *J. Clim.*, *6*, 1898–1913.

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