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Antarctic circumpolar waves: An indication of ocean-atmosphere coupling in the extratropics

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Abstract. Recent observational studies have detected coherent oceanic and atmospheric Antarctic circumpolar waves (ACWs) which propagate eastward at speeds of 6–8 $\mathrm{cm\,s^{-1}}$ with a circumpolar wavenumber of 2. Analysis of global wind data suggests that the ACWs are not remotely forced by tropical El Niño activity through atmospheric teleconnection. Rather, a coupled instability of the Antarctic Circumpolar Current and its overlying atmospheric motion is proposed to be responsible for the generation of the ACWs. The phase relationship among the observed oceanic and atmospheric variables supports the coupled instability theory. This study provides one clear example that extratropical ocean-atmosphere dynamical coupling is essential in generating climate variations on time scales of several years and longer.

Many observational and theoretical studies have successfully demonstrated that the ocean-atmosphere interaction in the tropics plays an essential role in the interannual climate variability dominated by the El Niño-Southern Oscillation (ENSO) phenomenon. The significance of the ocean-atmosphere interaction in the extratropics, however, remains speculative. While the interannual variability in the extratropical oceans has long been regarded as passive to atmospheric disturbances, recent findings of decadal-to-interdecadal climate variability in the North Pacific and North Atlantic have renewed interest in the notion that the ocean-atmosphere coupling is important in the extratropics as well [e.g., *Nitta and Yamada*, 1989; *Trenberth*, 1990; *Trenberth and Hurrell*, 1994; *Deser et al.*, 1996; *Nakamura et al.*, 1997]. Existing theories on the causes of decadal variability in the North Pacific can be largely grouped into two categories. One attributes the observed decadal variability to tropical forcing which, through the atmospheric teleconnection, modulates the midlatitude atmospheric and oceanic circulations [*Graham et al.*, 1994; *Kumar et al.*, 1994; *Gu and Philander*, 1997]. The alternative theory, based on coupled ocean-atmospheric models [*Latif and Barnett*, 1994; *Robertson*, 1996; *Jin*, 1997], proposes that the local coupling within the extratropical North Pacific itself can give rise to the observed decadal-to-interdecadal signals. So far, a solid explanation of this coupling has been lacking, partially because the observational data are still too limited to clarify the detailed phase relationships among oceanic and atmospheric variables so as to verify the proposed theories.

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In this study, we present observational evidence and a theoretical model to illustrate that regional ocean-atmosphere dynamical coupling in the extratropics is responsible for the Antarctic circumpolar waves (ACWs). The ACWs are the large-scale, eastward-propagating waves observed in the high-latitude Southern Hemisphere around the Antarctica [*White and Peterson*, 1996; *Jacobs and Mitchell*, 1996]. Their existence has been detected both in the oceanic variables (sea surface temperature, sea surface height, and sea-ice extent) and in the atmospheric variables (surface wind and sea-level pressure). The ACWs propagate eastward at average speeds of 6–8 $\mathrm{cm\,s^{-1}}$ and have a dominant wave period of about 4-5 years and a predominant circumpolar wavenumber of 2. The existence of the ACWs has been suggested to be linked to the ENSO activity in the equatorial Pacific, which has a predominant time scale of around 4 years [*White and Peterson*, 1996]. In this scenario, the ACWs in the Southern Ocean are the forced oceanic response to the overlying wind, whose interannual variability is induced through atmospheric teleconnection. To test this hypothesis, we examined the global 500 hPa wind data around the 4-to-5 year frequency band (Plate 1). The result shows that the eastward and westward propagating variances in the power spectra of the extratropical Northern Hemisphere atmospheric variability are nearly symmetric, decreasing in wavenumber from $k = 5 \sim 6$ around 20°N to $k = 4$ around 45°–65°N. This is attributable to the standing oscillatory tropical forcing associated with ENSO through the atmospheric teleconnection [*Wallace and Gutzler*, 1981]. The symmetric variance originating from the low-latitude is also discernible in the extratropical Southern Hemisphere. Around 45°–60°S over the Southern Ocean, the *symmetric* variance has a dominant wavenumber $k = 3$ and a spectral peak which is smaller than that in the same latitude band of the Northern Hemisphere. This result confirms the previous findings that the atmospheric response to the ENSO forcing over the Southern Ocean is dominated by the *stationary*, wavenumber 3 anomalies and that the teleconnected wavetrain amplitude in the Southern Hemi-

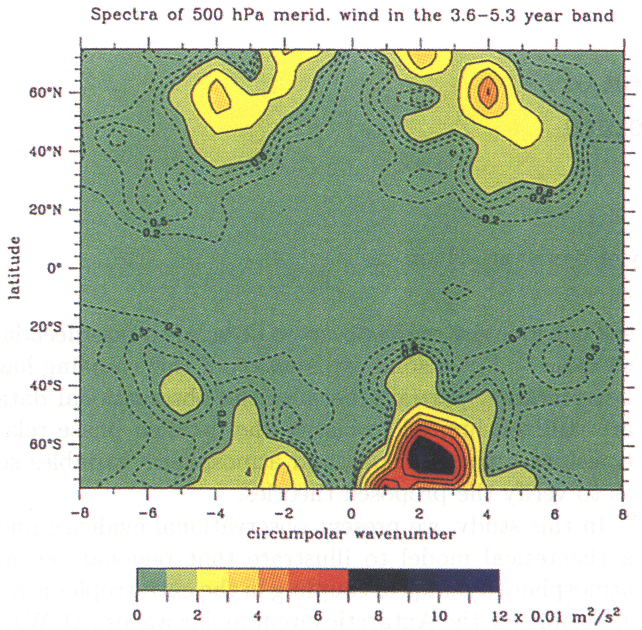

Plate 1. Latitudinal dependence of the auto-spectral amplitudes of the 500-hPa meridional wind data in the frequency band of 3.6–5.3 years. The result is based on the monthly wind analysis data of ECMWF from 1985 to 1995.

sphere is typically smaller than that in the Northern Hemisphere [*Mo and White*, 1985; *Karoly*, 1989]. As the ACWs are clearly associated with the *asymmetric*, eastward propagating $k = 2$ signals, this result does not support the hypothesis that the ACWs are forced and maintained by the remote, tropical ENSO events.

In this study, we hypothesize that the local ocean-atmosphere interaction is responsible for the dominant, eastward propagating ACWs. To test this hypothesis, we adopt a coupled ocean-atmospheric model in a zonally periodic channel suitable for the high-latitude Southern Hemisphere. For the oceanic motion, we use the wind-driven, 2-layer, quasi-geostrophic (QG) model:

$$\left(\frac{\partial}{\partial t} + U_1 \frac{\partial}{\partial x}\right)(\phi_1 - \phi_2) - (c_1 + U_s)\frac{\partial \phi_1}{\partial x} = -\frac{g' \nabla \times \vec{\tau}}{\rho_o f_o} \tag{1}$$

$$\left(\frac{\partial}{\partial t} + U_2 \frac{\partial}{\partial x}\right)(\phi_1 - \phi_2) + (c_2 - U_s)\frac{\partial \phi_2}{\partial x} = 0, \tag{2}$$

where U_i denotes the mean zonal flow in the ocean's i-th layer, $U_s = U_1 - U_2$, ϕ_i the i-th layer's geopotential, g' the reduced gravity, f_o the Coriolis parameter, ρ_o the reference water density, and $\vec{\tau}$ the surface wind stress vector. The baroclinic Rossby wave speed $c_i = \beta g' H_i / f_o^2$, where β is the meridional derivative of f and H_i the mean depth of the i-th layer. Unlike in the low-latitude oceans where upper ocean fluctuations are decoupled from the deep ocean signals, changes in the Antarctic Circumpolar Current (ACC) have been observed to have a strong barotropic component. The 2-layer QG model is the simplest dynamic model to capture this component.

The SST anomalies, which affect the overlying atmospheric motions, are governed by the heat conservation equation for the ocean's upper layer:

$$\left(\frac{\partial}{\partial t} + U_1 \frac{\partial}{\partial x}\right) T' + \frac{\bar{T}_y}{f_o}\frac{\partial \phi_1}{\partial x} = Q', \tag{3}$$

where $\bar{T}_y$ is the mean, meridional SST gradient across the ACC, and Q' is the anomalous surface heat flux related to the SST anomaly by $Q' = -\kappa_o T'$. Although Q' can be parameterized in a more complicated fashion involving, for example, $|\vec{\tau}|$ and T'_a, we have assumed it is dominated by the SST anomaly T', with the coefficient κ_o describing collectively all air-sea thermodynamic feedback processes.

For the overlying atmosphere, we assume its anomalous circulation is in an equilibrium state with the ocean. The anomalous heat balance equation for the lower atmosphere can then be written as follows:

$$U_a \frac{\partial T'_a}{\partial x} + v'_a \frac{\partial \bar{T}_a}{\partial y} + w'_a \frac{\partial \bar{T}_a}{\partial z} = -\kappa_a T'_a - b\,Q', \tag{4}$$

where T'_a is the atmosphere temperature anomaly, U_a the mean zonal wind, v'_a (w'_a) the anomalous meridional (vertical) wind, $\bar{T}_a$ the mean air temperature, κ_a the thermodynamical damping time scale, and $-b\,Q'$ the anomalous heat input from the ocean. Under the equivalent barotropic assumption (which is well valid for the high-latitude atmosphere in the Southern Hemisphere; *Karoly*, 1989), the low-level atmosphere pressure anomaly is proportional to the low-level temperature anomaly, $p = \lambda T'_a$, and through geostrophy, $v'_a = (\partial p/\partial x)/f_o\rho_a$, where ρ_a is the air density. In the following analysis, we will neglect the vertical advection term in Eq.(4), which has been shown by *Hoskins and Karoly* [1981] through the scaling analysis to be much smaller than the horizontal advection terms for the

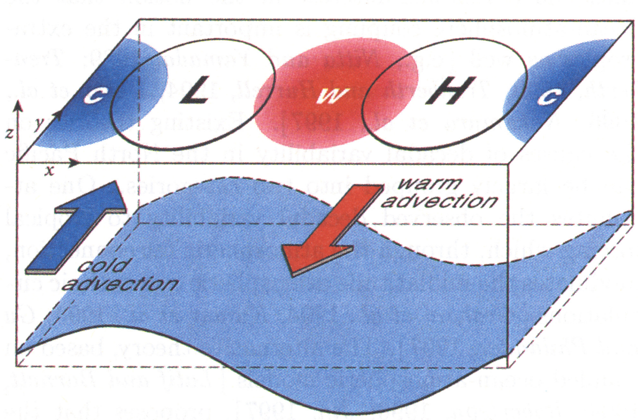

Plate 2. Schematic illustration for the coupled, unstable mode of the ACWs. In the figure, the air-sea interface is located at $z = 0$ and the wavy interface below the sea surface denotes the deformed thermocline due to Ekman convergence. Background zonal flows include U_a in the atmosphere, U_1 in the ocean's upper layer and U_2 in the ocean's lower layer.

high-latitude atmosphere. Coupling between the ocean and the atmosphere is provided by the surface wind stress, $\vec{\tau}/\rho_o = -\epsilon(\vec{k} \times \nabla p)/\rho_a f_o$, which acts as a body force on the upper layer of the model ocean, and by the thermal forcing of the SST to the atmosphere, $-bQ'$, where ϵ is the drag coefficient and b the conversion coefficient.

Assuming the mean flows U_i and U_a are meridionally uniform and the dependent variables ϕ_i, T', and p have wavelike solutions proportional to $\exp i(kx - \omega t)\cos l(y - y_o)$ within the channel (π/l being the channel width; $y_o = 55°$S), we obtain the following dispersion relation between ω and k:

$$(\omega - kc_r)(\omega - kU_1 + i\kappa_o) + A(k^2 + l^2)(\omega + kc_2 - kU_1) = 0, \quad (5)$$

where

$$c_r = (H_1U_1 + H_2U_2)/(H_1 + H_2) - c_1c_2/(c_1 + c_2),$$
$$A = -g'\epsilon\lambda b\kappa_o\bar{T}_y/\rho_a f_o^3(kU_a - k\lambda\bar{T}_{ay}/\rho_a f_o - i\kappa_a).$$

Notice that two modes are present in this coupled system: a "dynamic mode" associated largely with the propagation of oceanic baroclinic Rossby waves and a "SST mode" associated with the rate-of-change of the SST field. Based on the available climatological data of *Levitus* [1982], *Gordon et al.* [1978] and the ECMWF reanalysis, parameters appropriate for the Southern Ocean and its overlying atmosphere are $f_o = -1.19 \times 10^{-4}$ s^{-1}, $\beta = 1.32 \times 10^{-11}$ s^{-1}m^{-1}, $H_1 = 500$ m, $H_2 = 4500$ m, $g' = 0.015$ m s^{-2}, $\rho_o = 1000$ kg m^{-3}, $\rho_a = 1.23$ kg m^{-3}, $U_1 = 0.12$ m s^{-1}, $U_2 = 0.08$ m s^{-1}, $U_a = 10.0$ m s^{-1}, $\pi/l = 10°$lat., $\bar{T}_y = \bar{T}_{ay} = 0.4°$C/°lat., $\kappa_o^{-1} = 12$ months, $\kappa_a^{-1} = 2$ weeks, $\epsilon = 0.9 \times 10^{-5}$ m s^{-1}, $b = 134$, and $\lambda = 200$ Pa °C^{-1}. Under these parameters, the dynamic mode is unstable (Fig.1), whereas the SST mode is stable (not shown). The most unstable dynamic mode has a circumpolar wavenumber $k = 2$, an eastward phase speed of 7.8 cm s^{-1}, and a corresponding wave period of 4.6 years. These wave characteristics agree well with those of the observed ACWs. Notice that the presence of the most unstable wave at $k = 2$ does not depend sensitively on the parameters in the range around the above chosen values (a more detailed, parametric study will be reported separately).

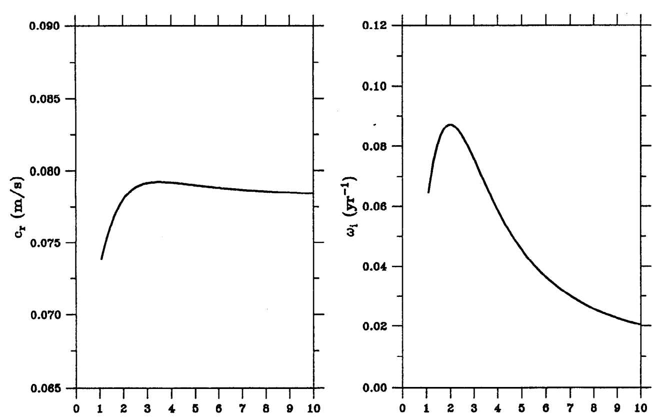

Figure 1. Phase speed c_r and growth rate ω_i as a function of the circumpolar wavenumber k for the dynamic mode in the coupled system of Eqs.(1)–(4). For parameters appropriate for the Southern Ocean and its overlying atmosphere (see the text), this mode becomes evanescent when $k < 1.05$.

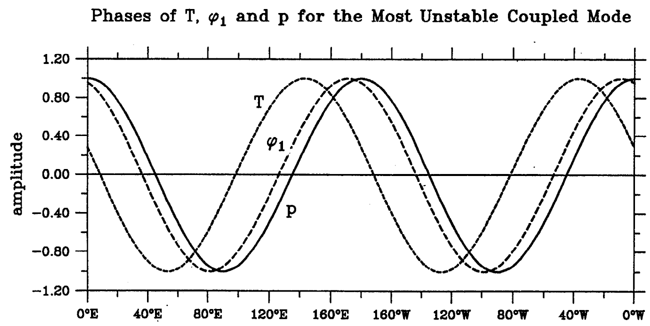

Figure 2. Relative phases of the atmospheric pressure anomaly p, the ocean's upper layer geopotential anomaly ϕ_1, and the SST anomaly T' for the most unstable dynamic mode at $k = 2$.

Figure 2 shows the relative phases of p, T', and ϕ_1 for the most unstable mode. The theory predicts the same phase lag (82 deg.) between the SST and the sea level pressure anomalies as that revealed by the observations [*White and Peterson*, 1996]. The observed phase lag between the SST anomalies and the sea level height anomalies ($\sim \phi_1$) ranges from 0 to 70 degrees (see Plate 3 of *Jacobs and Mitchell*, 1996) and its values are larger, as predicted by our theory (55 deg.), in regions away from the continental land masses such as the Drake Passage. While a more complicated model including realistic Antarctic land masses would simulate the phases quantitatively better, our proposed instability mechanism for the ACW is valid as long as the SLH anomalies fall between the SST and SLP anomalies (see discussion below). It is worth emphasizing that the SLH anomalies falling between the SST and SLP anomalies is a robust observed feature.

The physics underlying this coupled, unstable "dynamic mode" is illustrated in Plate 2. Assume an initial warm SST anomaly is present in the surface layer of the ACC. This provides a heat source for the overlying atmosphere, and the atmosphere, due to its short memory, promptly adjusts. As the zonal heat advection dominates the meridional heat advection in the high-latitude Southern Hemisphere, this adjustment creates a high-pressure anomaly east of the warm SST anomaly and a low-pressure anomaly west of it. Beneath the high-pressure (low-pressure) anomaly, Ekman convergence (divergence) deepens (shallows) the upper layer of the ocean. Because of $\bar{T}_y > 0$ in the Southern Ocean, the surface geostrophic flow resulting from the deformed upper layer works to advect warm (cold) water to strengthen the initial warm (cold) SST anomaly. As the warm SST anomaly is being advected eastward by the ACC, it is amplified through meridional warm water advection, providing a positive feedback mechanism for the growth of the oceanic and atmospheric anomalies in the coupled system. It is worth emphasizing that the warm/cold water advection is a crucial process leading to the growing instability of the dynamic mode; assuming $\bar{T}_y = 0$ results in $A = 0$ in (5) and leads to a *neutral* dynamic mode. Also note that the exponential growth of the coupled waves is likely to be suppressed when these waves attain finite amplitudes and when the nonlinear processes become dominant.

In conclusion we note that the agreement between the above theory and the observations of the ACW's characteristics, namely its zonal length scale, its phase speed, and the relative phases among the oceanic and atmospheric variables, suggests that the ACWs can be understood as a coupled, unstable mode unique to the Southern Ocean and its overlying atmosphere. Equally important, the existence and the dynamics of the ACWs as presented in this study provide one clear example that like in the tropics, ocean and atmosphere can be locally coupled in the extratropics. Since the time scale of the variability resulting from this coupled interaction resides in the long-term memory of the ocean dynamic adjustment, this air-sea coupling mechanism is likely generating decadal-to-interdecadal changes in the other extratropical oceans as well.

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