

The Antarctic Circumpolar Wave: A Beta Effect in Ocean–Atmosphere Coupling over the Southern Ocean

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ABSTRACT

The Antarctic circumpolar wave (ACW) is a nominal 4-yr climate signal in the ocean–atmosphere system in the Southern Ocean, propagating eastward at an average speed of $6\text{--}8\text{ cm s}^{-1}$, composed of two waves taking approximately 8 years to circle the globe. The ACW is characterized by a persistent phase relationship between warm (cool) sea surface temperature (SST) anomalies and poleward (equatorward) meridional surface wind (MSW) anomalies. Recently, White and Chen demonstrated that SST anomalies in the Southern Ocean operate to induce anomalous vortex stretching in the lower troposphere that is balanced by the anomalous meridional advection of planetary vorticity, yielding MSW anomalies as observed. In the present study, the authors seek to understand how this atmospheric response to SST anomalies produces a positive feedback to the ocean (i.e., an anomalous SST tendency displaced eastward of SST anomalies) that both maintains the ACW against dissipation and accounts for its eastward propagation. To achieve this, we couple a global equilibrium climate model for the lower troposphere to a global heat budget model for the upper ocean. In the absence of coupling, the model Antarctic Circumpolar Current (ACC) advects SST anomalies from initial conditions to the east at speeds slower than observed, taking 12–14 years to circle the globe with amplitudes that become insignificant after 6–8 years. In the presence of coupling, eastward speeds of the model ACC are matched by those due to coupling, together yielding a model ACW of a nominal 4-yr period composed of two waves that circle the globe in approximately 8 years, as observed. Feedback from atmosphere to ocean works through the anomalous zonal surface wind response to SST anomalies, yielding poleward Ekman flow anomalies in phase with warm SST anomalies. As such, maintenance of the model ACW is achieved through a balance between anomalous meridional Ekman heat advection and anomalous sensible-plus-latent heat loss to the atmosphere. This balance requires the alignment of covarying SST and MSW anomalies to be tilted into the southwest–northeast direction, which accounts for the spiral structure observed in global SST and sea level pressure anomaly patterns around the Southern Ocean. Eastward coupling speeds of the model ACW derive from a beta effect in coupling that displaces a portion of the anomalous meridional Ekman heat advection, and its corresponding anomalous SST tendency, to the east of SST anomalies. Therefore, the ACW is an example of self-organization within the global ocean–atmosphere system, depending upon the spherical shape of the rotating earth for its propagation and the mean meridional SST gradient for its maintenance, and producing a net poleward eddy heat flux in the upper ocean that tends to reduce this mean gradient.

1. Introduction

The space–time evolution of interannual sea surface temperature (SST) anomalies associated with the Antarctic circumpolar wave (ACW) displays general eastward propagation around the Southern Ocean (Fig. 1). White and Peterson (1996) found the phase relationship between SST and sea level pressure (SLP) anomalies such that associated poleward (equatorward) meridional surface wind (MSW) anomalies approximately coincide with warm (cool) SST anomalies over most of the Southern Ocean. In the presence of an apparent advection of SST anomalies by the eastward flow of the Ant-

arctic Circumpolar Current (ACC), the persistent phase relationships between SST, SLP, and MSW anomalies suggested to White and Peterson (1996) that maintenance of the ACW involves coupling between ocean and atmosphere. Here, we propose that this occurs because SST anomalies drive MSW anomalies in the overlying atmosphere and corresponding zonal surface wind (ZSW) anomalies drive an anomalous SST tendency through anomalous meridional Ekman heat advection.

Recently, White and Chen (1998; hereafter WC98) began testing this coupling hypothesis by simulating the atmospheric response to interannual SST anomalies in the Southern Ocean with an equilibrium climate model for the lower troposphere. From analysis of European Centre for Medium-Range Weather Forecasts (ECMWF) and National Centers for Environmental Prediction (NCEP) datasets, they found anomalous vortex stretching of the lower troposphere over the Southern

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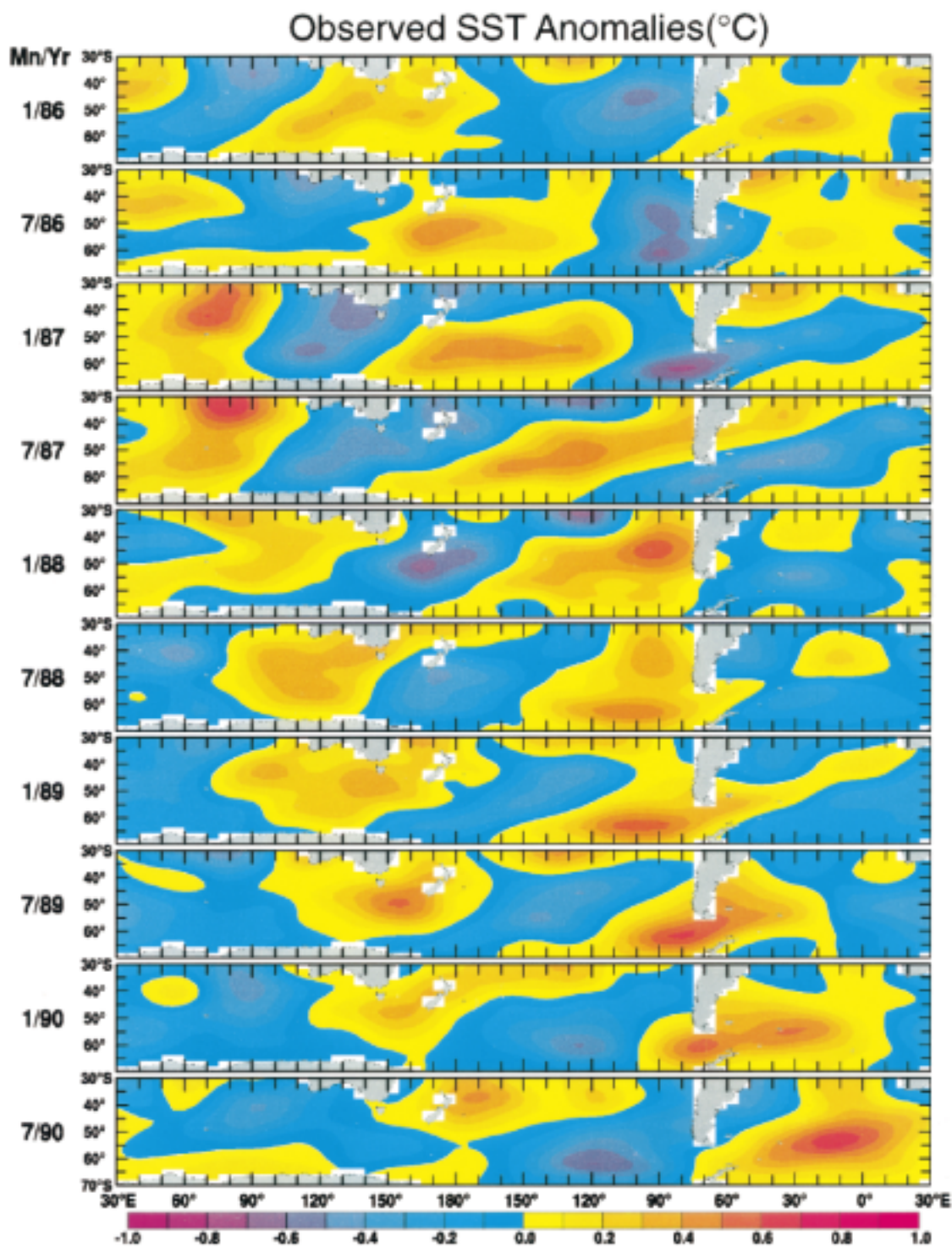


FIG. 1. Time sequence of interannual SST anomalies associated with the Antarctic circumpolar wave (ACW), extending from 30°S to 70°S and zonally around the globe each January and June from 1986 to 1990. Anomalies were formed from SST variables provided by the NCEP Reanalysis Project for the 15-yr period from 1982 to 1996 (Kalnay et al. 1996). Interannual anomalies are formed by bandpassing time sequences of SST anomalies from 1982 to 1996 with a 2 to 7-yr period admittance window (Kaylor 1977). In each figure, zonal means were removed and spatial smoothing applied to enhance the structure of the ACW. Contours are given by color bars ranging from $\pm 1.0^{\circ}\text{C}$.

Ocean occurring in response to SST-induced midlevel diabatic heating anomalies on interannual timescales (see the appendix). They found this leading to low-level anomalous convergence balanced by anomalous meridional advection of planetary vorticity, yielding poleward (equatorward) MSW anomalies over warm (cool) SST anomalies, as observed. Since the relationship between extratropical SST anomalies and midlevel diabatic heating anomalies holds for winter and not summer, this relationship extends to interannual timescales because of the former's dominance over the latter in the annual average. In the present study, we demonstrate how this atmospheric response to anomalous SST forcing can produce a positive feedback onto the ocean that yields the characteristic phasing of the observed ACW.

We begin by examining the NCEP Reanalysis for 15 years from 1982 to 1996 (Kalnay et al. 1996) to determine more precise phase relationships between SST, SLP, and ZSW anomalies associated with the ACW. This gives us clues as to the nature of coupling in the ACW. Second, we conduct a scale analysis that isolates specific processes believed responsible for the anomalous upper-ocean heat balance associated with the ACW. This simplifies the coupling, allowing us to construct an analytical model that yields a dispersion relation for free coupled waves of the system that match many of the wave characteristics observed in the ACW. Finally, we construct a numerical coupled model by linking a global equilibrium climate model for the lower troposphere (WC98) to a global heat budget model for the upper ocean. The latter ocean model balances the anomalous SST tendency with the advection of SST anomalies by the mean broadscale flow of the ACC and with the advection of the background SST field by anomalous meridional Ekman flow, the latter driven by the anomalous ZSW response to SST anomalies. The numerical coupled model allows us to simulate much of the space-time evolution exhibited by the ACW in Fig. 1.

2. Spiral structure in global SST and SLP anomaly patterns associated with the ACW

Recently, Robertson (1996) demonstrated how coupling between the extratropical ocean and atmosphere on interdecadal timescales requires the spatial scales of covarying SST and SLP anomalies to become anisotropic and their spatial orientation to become tilted at some angle from the zonal direction. He finds this tilt arising from the maintenance of covarying ocean and atmosphere signals against dissipation, as we shall demonstrate here in this study. As such, this tilt in the spatial orientation of covarying oceanic and atmospheric anomaly patterns becomes a signature not only for coupling but for the type of coupling between ocean and atmosphere. A similar tilt can be seen in the spatial orientation of SST anomalies associated with the ACW in Fig. 1, where the spatial orientation of elongated positive and negative SST anomalies can be seen

aligned in the southwest–northeast direction as the ACW propagates eastward across the Indian and Pacific sectors of the Southern Ocean. A more realistic view of this is given on a polar stereographic projection (Fig. 2), where this tilt in spatial orientation in Fig. 1 can be seen to be a manifestation of a spiral structure in global spatial patterns of covarying SST and SLP anomalies around the Southern Ocean. Within this spiral structure, SLP anomalies can be seen straddling the boundary between warm and cool SST anomalies, consistent with poleward (equatorward) MSW anomalies aligned with warm (cool) SST anomalies, as explained by WC98. Most importantly, the spiral structure allows westward (eastward) ZSW anomalies to come into approximate alignment with the warm (cool) SST anomalies as well.

If the spiral structure in global spatial patterns of covarying SST and SLP anomalies in the Southern Ocean is a signature for ocean–atmosphere coupling in the ACW, we should be able to use it to gain some insight into how covarying SST and SLP anomalies interact to produce ACW characteristics. Closer inspection in Fig. 2 of the SLP anomalies along the spiraling boundaries between warm and cool SST anomalies finds their extrema displaced poleward from the latitude connecting maximum SST anomalies. In particular, this can be seen with regard to the high SLP anomaly in the Pacific sector and with the two low SLP anomalies south of Africa and Australia. In these cases, associated ZSW anomalies between 45° and 65°S are in position to generate anomalous meridional Ekman advection of the background SST field (in addition to anomalous sensible-plus-latent heat flux and/or anomalous vertical mixing at the base of the near-surface mixed layer), yielding an anomalous SST tendency displaced poleward and eastward from SST anomalies of the same sign. Not only do these ZSW anomalies suggest a source for eastward (and poleward) propagation of the ACW, but that portion of the resultant anomalous SST tendency that is in phase with the SST anomalies suggests a source for maintaining the ACW against dissipation (i.e., anomalous heat loss to the atmosphere).

Therefore, we now ask whether an ocean–atmosphere coupled model can be constructed that will simulate this observed displacement between covarying SST and SLP anomalies. At the same time, we ask whether these displacements can explain the eastward propagation of the ACW observed in Fig. 1, while accounting for maintenance of the ACW against dissipation and its attendant global spiral structure in Fig. 2.

3. Scale analysis for the ACW

A necessary approach toward understanding coupling between ocean and atmosphere in the ACW lies in identifying the salient terms in anomalous heat and vorticity budgets in both the upper ocean and lower troposphere. White and Chen have already conducted a scale analysis

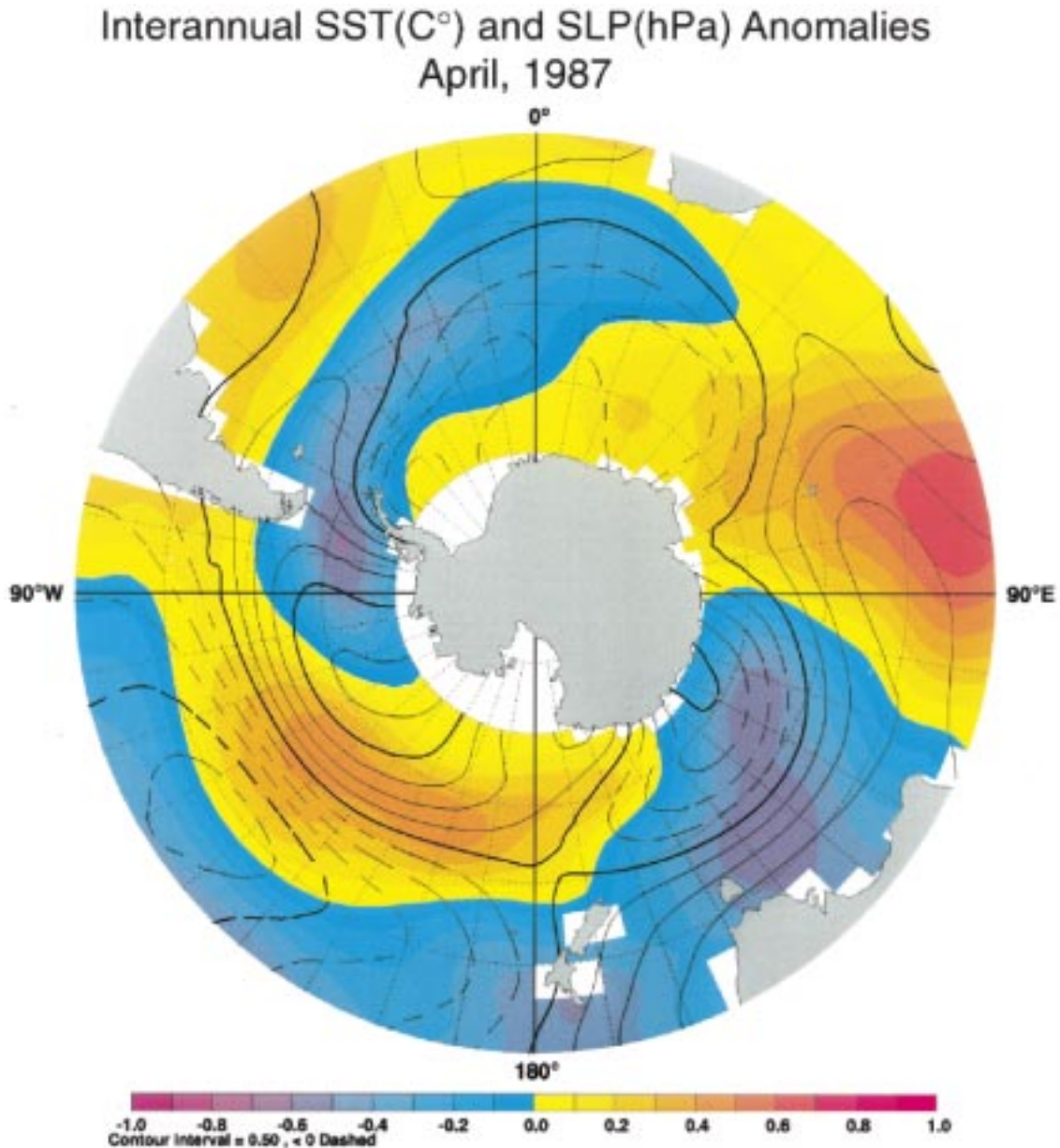


FIG. 2. Interannual SST and SLP anomalies for April 1987 plotted contemporaneously on a polar-stereographic projection centered at the South Pole and extending to 30°S. Anomalies were formed from SST and SLP variables provided by the NCEP reanalysis project for the 15-yr period from 1982 to 1996 (Kalnay et al. 1996). The zonal mean was removed from both SST and SLP anomalies and spatial smoothing applied to enhance the structure of the ACW. Contours of SST are given in the color bar, ranging from $\pm 1.0^\circ\text{C}$. Contours of SLP are 0.5 hPa, with positive anomalies given by thin solid lines and negative anomalies given by thin dashed lines, with the zero contour given by a thick solid line.

for the equilibrium response of the lower troposphere to SST-induced midlevel diabatic heating on large zonal spatial scales like those of the ACW (see the appendix). Here we do the same for oceanic heat and vorticity budgets, where scales and model parameters for both are summarized in Table 1.

The anomalous wind-driven potential vorticity budget for the upper layer of a two-layer ocean is given by Pedlosky (1996, p. 108), where relative vorticity is neg-

ligible compared to planetary vorticity on the global spacescales of the ACW; that is,

$$\partial h / \partial t + c_R \partial h / \partial x = -\text{curl}(\tau) / f, \quad (3.1)$$

where h is the depth anomaly of the interface representing the mean depth of the main pycnocline, positive downward; c_R is the nondispersive Rossby wave speed, negative westward (i.e., $-\beta g' H_o / f^2$); β is the meridional derivative of the Coriolis parameter (f); g' is the

TABLE 1. Summary of scales and model parameters.

Observed scales	Model parameters	Model (observed) variables	Constants
$L_T = 1.4 \times 10^8$ s	$g' = 0.01$ m s ⁻¹	T (anomalous SST)	$g = 9.8$ m sec ⁻¹
$L_X = 1.1 \times 10^7$ m	$c_R = -0.004$ m s ⁻¹	v_A (anomalous MSW)	$f = -1.2 \times 10^{-4}$ s ⁻¹ at 55°S
$L_Y = 4.4 \times 10^6$ m	$C_D = 1.3 \times 10^{-3}$	u_A (anomalous ZSW)	$\beta = 1.1 \times 10^{-11}$ m ⁻¹ s ⁻¹ at 55°S
$\partial T / \partial y = 7.5 \times 10^{-6}$ °C m ⁻¹	$\gamma = 1.3 \times 10^{-5}$ m s ⁻¹	P (anomalous SLP)	$C_{PA} = 1.0 \times 10^3$ m ² sec ⁻² K ⁻¹
$\partial T / \partial z = -2.0 \times 10^{-2}$ °C m ⁻¹	$\alpha = 2 \times 10^{-3}$ m ² s ⁻³ °C ⁻¹		$(C_{PA}/C_{PO}) = 0.24$
$\bar{u}_O = 0.05$ m s ⁻¹	$K = 1.3 \times 10^{-7}$ s ⁻¹		$\rho_A = 1$ kg m ⁻³
$\bar{u}_A = 10$ m s ⁻¹			$(\rho_A/\rho_O) = 1 \times 10^{-3}$
$H_O = 500$ m			
$H_P = 100$ m			
$H_A = 5000$ m			
$\Theta = 280$ K			
$N_A^2 = 10^{-4}$ s ⁻²			
$C_A^2 = 2.5 \times 10^3$ m ² s ⁻²			

scale of reduced gravity for the upper layer of the ocean (i.e., 0.01 m s⁻²); H_O is the mean depth scale of the main pycnocline (i.e., 500 m); and τ is anomalous kinematic horizontal wind stress at the sea surface (i.e., in units of m² s⁻²). The much thicker lower layer is assumed to be at rest. We obtain characteristic wavelength scales of the ACW from inspection of Figs. 1 and 2, yielding a zonal wavelength scale of 180° (i.e., $L_X = 1.1 \times 10^7$ m near 55° S) and a meridional wavelength scale of 40° (i.e., $L_Y = 4.4 \times 10^6$ m). As such, the average slope of the spiral structure in covarying global SST and SLP anomaly patterns from 45° to 65° latitude in Fig. 2 is approximately $\Delta y / \Delta x = -k / \lambda = (L_Y / L_X) = 0.4$, where k is the zonal wavenumber (i.e., positive eastward) and λ is the meridional wavenumber (negative southward). We obtain the characteristic period scale $L_T = 4.5$ yr from White and Peterson (1996), representing observed periodicity ranging from 4 to 5 yr over the 16 years of record from 1979 to 1994. These scales, together with -1.2×10^{-4} s⁻¹ for f and -0.004 m s⁻¹ for c_R near 55° S, finds the relative scale of the stretching term (i.e., $\partial h / \partial t$) on the lhs of Eq. (3.1) dominating that of the meridional advection of planetary vorticity (i.e., $C_{PX} \partial h / \partial x$) by more than an order of magnitude. Therefore, Eq. (3.1) reduces to the expression for Ekman pumping; that is,

$$\partial h / \partial t = (\partial \tau^x / \partial y) / f, \quad (3.2)$$

where $\partial \tau^x / \partial y$ also dominates $\partial \tau^y / \partial x$ in $\text{curl}(\tau) = (\partial \tau^y / \partial x - \partial \tau^x / \partial y)$, with τ^x and τ^y zonal and meridional components, respectively, of the anomalous surface wind stress. So, whatever influence SST anomalies have upon the atmosphere in the ACW, the feedback upon the main pycnocline in the ocean is through Ekman pumping in Eq. (3.2), principally through $\partial \tau^x / \partial y$, the latter proportional to the meridional gradient of the anomalous ZSW.

The anomalous heat budget for a near-surface mixed layer embedded in a two-layer ocean can be derived from Kraus and Businger (1994) on interannual timescales; that is,

$$\begin{aligned} \partial T / \partial t + KT + \bar{u}_O \partial T / \partial x \\ = -(\partial T / \partial y)(v_G + v_E) - (\partial T / \partial z)w_E, \end{aligned} \quad (3.3)$$

where T is the uniform anomalous temperature in this layer, representing anomalous SST; $\bar{u}_O$ represents the mean zonal geostrophic flow of the broadscale ACC, uniform over this layer; KT is dissipation representing the anomalous sensible-plus-latent heat loss to the atmosphere; v_G and v_E represent anomalous meridional geostrophic and Ekman flow, respectively, both uniform over the near-surface mixed layer, the latter proportional to anomalous ZSW; and w_E represents the vertical velocity anomaly at the base of the near-surface mixed layer. By assuming that climatological atmospheric forcing during winter deepens the near-surface mixed layer to the top of the main pycnocline at a scale depth, H_P , of 100 m, then fluctuations from year to year in the depth of the main pycnocline from anomalous Ekman pumping will yield cool (warm) SST anomalies when the pycnocline is displaced upward (downward) from its long-term mean depth. For scaling purposes, we propose that w_E in Eq. (3.3) is less than or equal to $\frac{1}{2} \partial h / \partial t$ from Eq. (3.2) (where $w_O = 0$ at the sea surface) and that $(\partial T / \partial z)$ in Eq. (3.3) is less than or equal to the mean vertical temperature gradient in the upper half of the main pycnocline (i.e., -2×10^{-2} °C m⁻¹).

In subsequent development of the coupled model, we will continue to assume winter conditions because WC98 found anomalous midtropospheric level diabatic heating in the Southern Ocean responding to SST anomalies exclusively during winter, not during summer. This derives from the influence that SST anomalies have upon the location and/or intensity of aggregate synoptic storm activity during winter. White and Chen inferred from this that the phase relationship between SST and MSW anomalies observed on interannual timescales by White and Peterson (1996) represents an annual average of an inherent winter (and possibly autumn and spring) mechanism.

Additional assumptions have been made in the heat

budget model in Eq. (3.3). In the feedback from atmosphere to ocean, ZSW anomalies influence more than just anomalous meridional Ekman heat advection represented in this model; they also influence anomalous wind mixing at the base of the near-surface mixed layer and the anomalous sensible-plus-latent heat loss at the sea surface, both of which are in phase with the meridional Ekman heat advection in their collective influence on the anomalous SST tendency (Miller et al. 1994). But, for the sake of simplicity, we have ignored the latter two mechanisms in favor of the former, recognizing that they may contribute to effects attributed in this model solely to the anomalous meridional Ekman heat advection. Also, we have ignored the influence that anomalous cloud cover has upon the anomalous net radiation flux in its contribution to the anomalous SST tendency. Furthermore, we have ignored the near-surface Ekman deflection of anomalous geostrophic winds in the near-surface planetary boundary layer of the lower troposphere, assuming that anomalous oceanic Ekman flow (i.e., vertically averaged over the near-surface mixed layer of the ocean) is directed to the left of anomalous surface wind stresses in the Southern Hemisphere, the latter aligned with anomalous surface geostrophic winds.

From the description of horizontal and vertical temperature distributions and zonal geostrophic surface currents (0/4000 db) in the Southern Ocean (Olbers et al. 1992), we estimate the scales and model parameters used in Eq. (3.3) from spatial averages taken over the Southern Ocean from 45° to 65°S, summarized in Table 1. The dissipation parameter K corresponds to an e-folding timescale for anomalous sensible-plus-latent heat loss to the atmosphere of 3 months. This estimate for K^{-1} is obtained from $4(\rho_A/\rho_O)(C_{PA}/C_{PO})(C_D\bar{u}_A/H_p)$, balancing the dissipation of anomalous heat storage in the near-surface mixed layer of depth H_p with the anomalous sensible-plus-latent heat loss to the atmosphere, where the factor of 4 assumes that anomalous latent heat loss is three times the anomalous sensible heat loss, the latter directly related to anomalous SST. This factor could be larger or smaller depending upon location, season, and year of interest.

Therefore, given these assumptions and estimates of scale and parameters, the scale analysis for the heat budget in the upper ocean on the lhs of Eq. (3.3) finds relative scales for the three terms to be approximately 2, 4, and 1, respectively, where derivative terms have been multiplied by 2π radians. As such, each term is important in the anomalous upper ocean heat budget of the ACW and cannot be ignored. On the rhs of Eq. (3.3), relative scaling between $v_E \partial T/\partial y$ and $w_E \partial T/\partial z$ depends upon their mutual relationship to the pycnocline depth anomaly (h) for the two-layer model ocean; that is,

$$v_G = (g'/f)\partial h/\partial x, \quad w_E = \frac{1}{2}\partial h/\partial t. \quad (3.4)$$

Substituting Eq. (3.4) into Eq. (3.3), and applying the scales and parameters evaluated previously and summarized in Table 1, finds relative scales for $w_E \partial T/\partial z$ and $v_G \partial T/\partial y$ to be approximately 1 and 1, respectively. This indicates that, if either advective process is important in the anomalous upper ocean heat budget of the ACW, then both processes are important.

Finally, relative scaling between $v_E \partial T/\partial y$ and $w_E \partial T/\partial z$ on the rhs of Eq. (3.3) depends upon their mutual relationship to the ZSW anomaly (u_A); that is,

$$v_E = -(\gamma/fH_p)u_A, \quad w_E = (\gamma/(2f))\partial u_A/\partial y, \quad (3.5)$$

where γ is the coefficient (i.e., $1.3 \times 10^{-5} \text{ m s}^{-1}$) linearly relating anomalous ZSW to anomalous zonal surface wind stress, estimated from $(\rho_A/\rho_O)(C_D/\bar{u}_A)$. In Eq. (3.5), v_E is meridional Ekman flow (e.g., Gill 1982) while the vertical velocity w_E is half the Ekman pumping velocity given in Eq. (3.2). Substituting Eq. (3.5) into Eq. (3.3), and applying the scales and parameters evaluated previously and summarized in Table 1, finds relative scales for $v_E \partial T/\partial y$ and $w_E \partial T/\partial z$ to be approximately 8 and 1, respectively, where the derivative term in Eq. (3.5) has been multiplied by 2π radians. This indicates that meridional Ekman heat advection dominates both vertical Ekman heat advection and meridional geostrophic heat advection on the rhs of Eq. (3.3) in the feedback from atmosphere to ocean in the ACW. This is consistent with the meridional Ekman heat advection scenario discussed previously in the examination of the phase differences between ACW SST and SLP anomaly patterns in Fig. 2.

4. An analytic coupled model for the ACW

The foregoing scale analysis given for the upper ocean and that given by WC98 for the lower troposphere greatly simplifies modeling of ocean-atmosphere coupling in the ACW. In the lower troposphere, it allows us to relate SST anomalies to vertical-averaged zonal geostrophic wind anomalies (assumed equivalent to ZSW anomalies), and in the upper ocean it allows us to relate ZSW anomalies to the anomalous SST tendency through anomalous meridional Ekman heat advection. The question now becomes whether such a simple coupled model can simulate the wave characteristics of the ACW in a manner consistent with observed phase relationships in Figs. 1 and 2.

Hence, we utilize results from the foregoing scale analysis to simplify the anomalous heat budget for the upper ocean in Eq. (3.3), yielding

$$\partial T/\partial t + KT + \bar{u}_O \partial T/\partial x = (\partial T/\partial y)(\gamma/(fH_p))u_A, \quad (4.1)$$

where the anomalous SST tendency ($\partial T/\partial t$) following a fluid parcel traveling zonally with the mean speed $\bar{u}_O$ of the broadscale ACC and damped by the anomalous turbulent loss of heat to the atmosphere (i.e., parameterized by KT), is balanced by anomalous heat advection stemming from anomalous meridional Ekman flow [see

Eq. (3.5a)] operating on the mean meridional SST gradient (i.e., $\partial T/\partial y$). Here, to keep the model linear, we assume both γ and K to be constant, neglecting their dependency upon wind speed.

Now, the anomalous sensible-plus-latent heat loss (KT) in Eq. (4.1) influences the overlying atmosphere and instigates the feedback to the ocean. How this works has been examined by White and Chen and is summarized here in the appendix. There, we learn that some portion or all of KT instigates midlevel anomalous diabatic heating (Q) in Eq. (A.2), which leads to the anomalous MSW (v_A) response in Eq. (A.5). Yet coupling between ocean and atmosphere in association with the ACW requires understanding of the anomalous ZSW (u_A) response to SST anomalies. This is determined by substituting anomalous MSW from Eq. (A.5) into the expression for the conservation of mass in the lower troposphere in Eq. (A.4), yielding

$$\partial u_A/\partial x = -(f\alpha/(\beta C_A^2))[\partial T/\partial y + (2\beta/f)T], \quad (4.2)$$

where the factor of 2 in the second term on the rhs of this expression stems from the fact that anomalous MSW in Eq. (A.5) varies as f/β and the low-level convergence is proportional to SST in Eq. (A.4). The resulting expression indicates that $\partial u_A/\partial x$ derives both from $SST(T)$ and its meridional derivative. Therefore, the influence of anomalous ZSW upon the anomalous heat budget of the upper ocean in Eq. (4.1) can only be achieved by taking the zonal derivative of the latter, yielding

$$\begin{aligned} \partial^2 T/\partial t \partial x + \bar{u}_O \partial^2 T/\partial x^2 \\ + [\overline{\partial T/\partial y}(\gamma\alpha/(\beta C_A^2 H_p))(2\beta/f)]T + KdT/dx \\ + [\overline{(\partial T/\partial y)}(\gamma\alpha/(\beta C_A^2 H_p))]\partial T/\partial y = 0, \end{aligned} \quad (4.3)$$

where a term involving the zonal derivative of $\overline{\partial T/\partial y}$ is negligible in this broadscale development. This latter expression in $SST(T)$ constitutes the coupled partial differential equation for the model ACW.

We can compute the dispersion relation for the model ACW expressed in this coupled PDE by setting scales to constant values representing spatial and temporal averages over the Southern Ocean from 45° to 65°S (and summarized in Table 1), and by substituting an idealized wave solution $T = T_O \exp(k_M x + \lambda_M y - \sigma_M t)$ into it, yielding

$$\begin{aligned} \sigma_M = \bar{u}_O k_M - [\overline{\partial T/\partial y}(\gamma\alpha/(\beta C_A^2 H_O))(2\beta/f)]k_M^{-1} \\ - \{K + [\overline{\partial T/\partial y}(\gamma\alpha/(\beta C_A^2 H_p))](\lambda_M/k_M)\}i, \end{aligned} \quad (4.4)$$

where σ_M is the complex frequency, k_M is the zonal wavenumber, and λ_M is the meridional wavenumber of the model ACW. The zonal phase speed of the model ACW (i.e., $c_{ACWX} = \sigma_M/k_M$) depends upon the broadscale flow of the ACC, augmented by a zonal coupling speed, (c_{CX}); that is,

$$\begin{aligned} c_{ACWX} = \bar{u}_O + c_{CX}, \\ c_{CX} = -[\overline{\partial T/\partial y}(\gamma\alpha/(\beta C_A^2 H_p))(2\beta/f)]k_M^{-2}, \end{aligned} \quad (4.5)$$

where in the Southern Ocean $[\overline{\partial T/\partial y}/f] < 0$; so, the zonal coupling speed (c_{CX}) is eastward, in the same direction as the broadscale mean flow ($\bar{u}_O$) of the ACC.

The dispersion relation in Eq. (4.4) also contains a growth term depending upon the sign of $\{K + [\overline{\partial T/\partial y}(\gamma\alpha/(\beta C_A^2 H_p))](\lambda_M/k_M)\}$. For the coupled model of the ACW to maintain its amplitude against dissipation, this expression is set to zero, thereby defining the magnitude and direction of the meridional wavenumber for stable coupled waves; that is,

$$\lambda_M/k_M = -[\overline{\partial T/\partial y}(\gamma\alpha/(\beta C_A^2 H_p K))]^{-1}. \quad (4.6)$$

For $\overline{\partial T/\partial y} > 0$ and for $k_M > 0$, then $\lambda_M < 0$. So, as the coupled model ACW propagates eastward around the Southern Ocean, its amplitude remains stable if it has a poleward wavenumber component λ_M given by Eq. (4.6). Therefore, the meridional phase speed of the ACW is obtained by substituting Eq. (4.6) for λ_M into the expression for $c_{ACWY} = \sigma_M/\lambda_M$ in the real part of Eq. (4.4), yielding

$$\begin{aligned} c_{ACWY} = \bar{u}_O(k_M/\lambda_M) + c_{CY}, \\ c_{CY} = K(2\beta/f)\lambda_M^{-2}. \end{aligned} \quad (4.7)$$

This meridional coupling speed (c_{CY}) is highly dispersive depending upon λ_M^{-2} , also depending upon the beta effect (i.e., β/f) and the dissipation timescale K^{-1} . Note that this coupling phase speed is poleward and decreases with latitude through the inverse Coriolis parameter (i.e., negative in the Southern Hemisphere).

The magnitude of the zonal coupling speed c_{CX} can be estimated from Eq. (4.5) for scales and model parameters estimated previously and summarized in Table 1. We can see that c_{CX} is highly dispersive, depending upon k_M^{-2} , making it larger for longer wavelengths. From the dispersion relation in Eq. (4.4), we can compute the zonal wavenumber in terms of frequency {i.e., $k_M = 0.5(\sigma_M/\bar{u}_O)[1 + (1 + 4\overline{\partial T/\partial y}(2\gamma\alpha)(\bar{u}_O/\sigma_M^2)/(fC_A^2 H_p))^{1/2}]$ }, yielding $k_M = 0.7(\sigma_M/\bar{u}_O) = 6 \times 10^{-7} \text{ m}^{-1}$ (i.e., $L_{XM} = 2\pi/k_M = 1.1 \times 10^7 \text{ m}$) at 55°S for a model frequency ($\sigma_M = 4.5 \times 10^{-8} \text{ s}^{-1}$) corresponding to a period of 4.5 yr. This model zonal wavelength L_{XM} is the same as that observed earlier (i.e., $L_X = 1.1 \times 10^7 \text{ m}$ at 55°S) from consideration of Figs. 1 and 2. From knowledge of this model zonal wavenumber, the zonal coupling speed (c_{CX}) can be computed from Eq. (4.5b), yielding $c_{CX} = 0.04 \text{ m s}^{-1}$. Take note that both the zonal wavenumber and the zonal coupling speed are independent of the dissipation timescale K^{-1} . The zonal coupling speed is approximately the same as the scale magnitude for the broadscale flow of the ACC (i.e., $\bar{u}_O = 0.05 \text{ m s}^{-1}$). Summing the two speeds in Eq. (4.5a) yields an estimate for c_{ACWX} of approximately 0.09 m s^{-1} , similar to the observed speed of 0.08 m s^{-1} (i.e., L_X/L_T in Table 1).

The meridional coupling speed (c_{CY}) in Eq. (4.7) and

its associated meridional wavenumber (λ_M) in Eq. (4.6) depend upon the zonal wavenumber (k_M). The meridional wavelength (L_{YM}) and the meridional coupling phase speed are linearly related to the dissipation time-scale K^{-1} . For scales and model parameters estimated previously and summarized in Table 1, we find stable coupled waves requiring $\lambda_M = -2.2k_M = -1.2 \times 10^{-6} \text{ m}^{-1}$ (i.e., $L_{YM} = 2\pi/\lambda_M = -5.2 \times 10^6 \text{ m}$), directed poleward. This model meridional wavelength is similar to that observed earlier (i.e., $L_Y = -4.4 \times 10^6 \text{ m}$) from inspection of Figs. 1 and 2. Moreover, it yields a southwest–northeast alignment in crests and troughs of the model ACW, with an average slope of $\Delta y/\Delta x = -k_M/\lambda_M = 0.5$, similar to that observed earlier (i.e., $-k/\lambda = 0.4$) in the spiral structure displayed in Figs. 1 and 2. This value for λ_M allows meridional coupling speed (c_{CY}) to be computed from Eq. (4.7b), yielding $c_{CY} = -0.02 \text{ m s}^{-1}$ at 55°S . This is approximately one-half the zonal coupling phase speed of c_{CX} in Eq. (4.5b). When added to the advection of the coupled wave by the broadscale flow of the ACW [i.e., $\bar{u}_O k_M/\lambda_M$ in Eq. (4.7a)], their sum $c_{ACWY} = -0.04 \text{ m s}^{-1}$ again approximately one-half that of c_{ACWX} in Eq. (4.5a). Therefore, for the model ACW to remain stable it must propagate poleward at approximately one-half its eastward phase speed.

5. Model phase relationship between pycnocline depth and SST anomalies

The phase relationship between pycnocline depth (h) and SST(T) anomalies can now be determined by taking the zonal derivative of Eq. (3.2), allowing $\partial u_A/\partial x$ to be represented by Eq. (4.2), yielding

$$\partial^2 h/\partial t \partial x = -[(\gamma\alpha)/(\beta C_A^2)]\partial^2 T/\partial y^2 + [(\gamma\alpha)/(\beta C_A^2)](2\beta/f)\partial T/\partial y. \quad (5.1)$$

Therefore, if $T = T_O \cos(k_M x + \lambda_M y - \sigma_M t)$, then the solution for h can be written

$$h = h_O [\cos(k_M x + \lambda_M y - \sigma_M t) + (2\beta/f)(1/\lambda) \sin(k_M x + \lambda_M y - \sigma_M t)], \quad (5.2)$$

where $h_O = T_O[(\gamma\alpha\lambda^2)/(\beta C_A^2 \sigma_M k_M)]$. With λ_M defined by Eq. (4.6) to be $-2.2k_M$ for stable coupled waves, then the ratio of the cosine to the sine term in Eq. (5.2) is approximately 5 to 1. Therefore, since $(2\beta/f)(1/\lambda_M) > 0$ in the Southern Hemisphere, positive h anomalies (i.e., positive downward) are displaced approximately 10° of phase to the east of positive SST anomalies. For magnitudes of $T_O = 0.5^\circ\text{C}$, the magnitude of h_O is approximately 10 m, corresponding to a sea-level height amplitude [i.e., $h_O(\rho'/\rho_O)$] of approximately 1 cm. This amplitude is consistent with that observed by Jacobs and Mitchell (1996), who also observed no significant lag between SST and sea-level height anomalies associated with the ACW. Moreover, this anomalous vertical

displacement of the pycnocline yields an anomalous SST change in Eq. (3.3) of $\Delta T = -(\partial T/\partial z)(\Delta h/2) = 0.1^\circ\text{C}$. As already discussed in section 3 and as we shall demonstrate in the next section, this influence is too small to significantly influence SST anomalies associated with the ACW.

6. Model phase relationships leading to eastward and poleward phase propagation

To understand how phase relationships between model SST, MSW, SLP, and ZSW anomalies are related to eastward and poleward phase propagation of the model ACW in Eqs. (4.4)–(4.7), we employ a graphical approach. This allows us to examine phase relationships as SST anomalies generate ZSW anomalies and as the latter, in turn, generate an anomalous SST tendency through anomalous meridional Ekman heat advection. Moreover, this graphical approach allows us to demonstrate the importance of the beta effect [i.e., $(2\beta/f)$ in Eqs. (4.5) and (4.7)] in this coupled interaction.

We begin by displaying an idealized zonal- and meridional-varying SST wave (top panel, Fig. 3) where $T = T_O \sin(k_M x + \lambda_M y)$, with $k_M = 2\pi/L_{XM}$ and $\lambda_M = -2\pi/L_{YM}$ assumed to be uniform over a schematic ocean extending from 30° to 60°S , 30° to 120°E with no continental boundaries. We evaluate $L_{XM} = 180^\circ$ longitude (i.e., $1.3 \times 10^7 \text{ m}$ at 45°S), $L_{YM} = 60^\circ$ latitude (i.e., $6.0 \times 10^6 \text{ m}$), and $T_O = 0.5^\circ\text{C}$, similar to model scales computed in the previous section, with constant f and β evaluated at 45°S , and model parameters and scales otherwise given in Table 1. These parameters and scales are taken as uniform over the domain, consistent with conditions allowing for the idealized wave solution to Eq. (4.3); yet, f and β are allowed to vary with latitude when we choose. Next, we use the relationship between MSW and SST anomalies in Eq. (A.5) in the appendix to compute the corresponding anomalous SLP(P) wave; that is,

$$\partial P/\partial x = \rho_A [f^2 \alpha/(\beta C_A^2)] T. \quad (6.1)$$

Integrating this expression analytically yields $P = -T_O \rho_A [f^2 \alpha/(\beta C_A^2)] (1/k_M) \cos(k_M x + \lambda_M y)$, which is displayed (second panel, Fig. 3) for variable Coriolis parameter (left) and constant Coriolis parameter (right). Since SLP anomalies in Eq. (6.1) vary as the square of the Coriolis parameter (f), their response to SST anomalies is poleward intensified, as discussed by WC98. By comparing left and right panels in Fig. 3, we find the SLP amplitude [i.e., $P_O = T_O \rho_A [f^2 \alpha/(\beta C_A^2)] (1/k_M) = 6 \text{ hPa}$] at approximately 55°S computed with variable Coriolis parameter. So, maximum SLP anomalies occur along the sloping boundary between warm and cool SST anomalies but are displaced southward by approximately 10° latitude from that centered at 45°S computed on the basis of constant Coriolis parameter. This southward displacement is similar to that observed in Fig. 2.

Negative ZSW anomalies near 45°S , 90°W (i.e., geo-

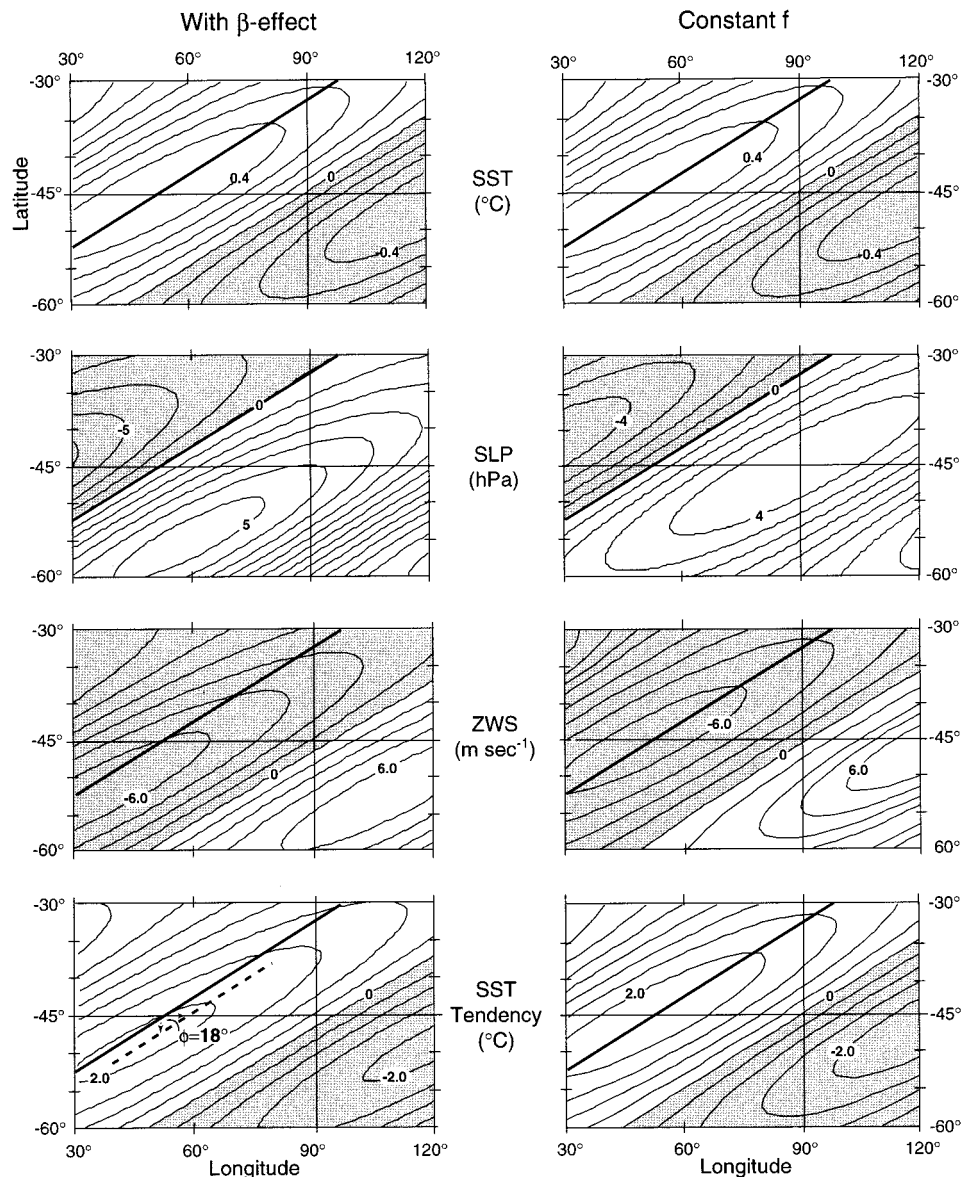


FIG. 3. A schematic representation is given of the model phase relationships between anomalous SST, SLP, ZSW, and SST tendency, extending over the domain from 30° to 60°S, 30° to 120°E. Top panel: The model SST wave is assumed to be proportional to $\sin(k_M x + \lambda_M y)$, with wavenumbers and amplitudes given in the text. Second panel: The model SLP wave in response to the SST wave then is proportional to $-f^2 \cos(k_M x + \lambda_M y)$, with amplitudes given in the text. Third panel: The model ZSW wave is related to the SLP wave through the geostrophic relationship, proportional to $(f/\beta)[(2\beta/f) \cos(k_M x + \lambda_M y) - \lambda_M \sin(k_M x + \lambda_M y)]$, with amplitudes given in the text. Bottom panel: Finally, the model SST tendency wave due only to anomalous meridional Ekman heat advection, driven by the model ZSW wave, becomes proportional to $[(2\beta/f) \cos(k_M x + \lambda_M y) - \lambda_M \sin(k_M x + \lambda_M y)]\beta^{-1}$, with amplitudes given in the text. In the left column are results based upon a variable Coriolis parameter in the atmosphere; in the right column are results based upon a constant Coriolis parameter where terms within brackets containing β have been set to zero. Negative anomalies are shaded; positive anomalies are unshaded. In each panel, amplitudes have been weighted by latitude with a Gaussian filter that allows zonal and meridional phase shifts between right and left panels to be discerned more easily. The sloping heavy black line is the same in each panel, used as a reference to discern shifts in phase between right and left panels. See text describing how model waves associated with each variable can be followed from one panel to the next and how wave amplitudes are computed.

strophically inferred from peak SLP anomalies near 55°S) hold the key to understanding how the ACW propagates to the east. To see how this works, we compute an idealized ZSW wave analytically from Eq. (4.2), yielding $u_A = T_o[f\alpha/(\beta C_A^2)](1/k_M)[(2\beta/f) \cos(k_M x + \lambda_M y) - \lambda_M \sin(k_M x + \lambda_M y)]$, which again is displayed (third panel, Fig. 3) for variable Coriolis parameter (left) and constant Coriolis parameter (right). This finds the ZSW wave computed with variable Coriolis parameter occurring nominally along the sloping boundary between warm and cool SST anomalies, but displaced southward and eastward by approximately 5° latitude and longitude, respectively, from that computed on the basis of constant Coriolis parameter. This displacement stems from the cosine term in the expression for the ZSW wave, which is located 90° of phase eastward and poleward of peak SST anomalies. As such, the amplitude of the ZSW wave occurring in phase with the SST wave (i.e., the sine term in the solution) is $u_{AO} = T_o[f\alpha/(\beta C_A^2)](\lambda_M/k_M) = -6 \text{ m s}^{-1}$, while that shifted 90° of phase to the east of the SST wave (i.e., the cosine term in the solution) is $u_{AO} = -T_o[f\alpha/(\beta C_A^2)](2\beta/f)(1/k_M) = -1.8 \text{ m s}^{-1}$. The former is responsible for growth of the SST wave in Eq. (4.3), while the latter is responsible for its poleward and eastward phase propagation. This propagation depends upon the beta effect stemming from the coupling through the coefficient $2\beta/f$.

To see this, we next compute the anomalous SST tendency due strictly to the influence of the ZSW wave upon anomalous meridional Ekman heat advection in Eq. (4.1), yielding $\Delta T/\Delta t = T_o[\gamma\alpha/(\beta C_A^2 H_p)](\partial T/\partial y)(1/k_M)[(2\beta/f) \cos(k_M x + \lambda_M y) - \lambda_M \sin(k_M x + \lambda_M y)]$, which is displayed (bottom panel, Fig. 3) for variable Coriolis parameter (left) and constant Coriolis parameter (right). For constant Coriolis parameter, the $\Delta T/\Delta t$ wave is in phase with the SST wave in the top panel, in which case the feedback from atmosphere to ocean solely produces a growth tendency for the SST wave. On the other hand, for variable Coriolis parameter, the $\Delta T/\Delta t$ wave is displaced eastward and poleward by approximately 18° of phase and, therefore, not only gives the SST wave a growth tendency but causes it to propagate poleward and eastward as well. This propagation stems from the cosine term in the expression for the $\Delta T/\Delta t$ wave (i.e., in concert with that in the expression for the ZSW wave) and is due to the poleward intensification of the anomalous SLP response to anomalous SST in Eq. (6.1), the latter translated into a beta effect (i.e., $2\beta/f$) when the ocean and atmosphere are coupled.

The corresponding anomalous meridional Ekman flows are relatively weak on these interannual time-scales, computed to be $v_E = -(\gamma/fH_p)u_A = -2.3 \times 10^{-3} \text{ m s}^{-1}$ for $u_A = -1.8 \text{ m s}^{-1}$. However, over a one-year period even this relatively weak magnitude of Ekman flow will create an anomalous SST tendency of $\Delta T = -v_E (\partial T/\partial y) \Delta t = 0.5^\circ\text{C}$ displaced 90° of phase eastward and poleward of the original SST wave. Since this portion of the anomalous SST change is the same as

the amplitude of the SST wave, this magnitude of anomalous Ekman flow can account for eastward and poleward phase propagation of the ACW. The total anomalous SST change (i.e., due to anomalous meridional Ekman heat transport both in phase and shifted 90° of phase eastward and poleward of original SST wave) is approximately 2.1°C, with that portion in phase with the SST wave (i.e., 2.0°C) accounting for a growth tendency in the amplitude of the wave. This growth tendency is balanced by the anomalous sensible-plus-latent heat loss to the atmosphere, thereby keeping the wave amplitude stable.

Now we can see that the ACW is also associated with a net poleward transport of heat in the Southern Hemisphere between subtropical and polar latitudes. This does not occur in any linear sense, but the poleward eddy flux of heat (i.e., $v_E \overline{T}$) is nonzero when averaged over time at a particular location or along a latitude extending around the globe at a particular time. We can demonstrate this by beginning with $T = T_o \sin(k_M x + \lambda_M y)$ and following the sequence of interactions given above to arrive at the meridional Ekman flow wave $v_E = -(\gamma/fH_p)u_A = -T_o[\alpha\gamma/(\beta H_p C_A^2)](1/k_M)[(2\beta/f) \cos(k_M x + \lambda_M y) - \lambda_M \sin(k_M x + \lambda_M y)]$. This has a constituent common with T (i.e., the sine term), which together yield a net poleward eddy heat flux $\text{EHF} = \rho_o C_{po} H_p 2L_{XM} v_E \overline{T}$ integrated around the globe at 45°S; that is,

$$\begin{aligned} \text{EHF} &= \rho_o C_{po} T_o^2 (\alpha\gamma/(\beta C_A^2)) (\lambda_M/k_M) L_{XM} \\ &= -2 \times 10^{13} \text{ watts.} \end{aligned} \quad (6.2)$$

This estimate is two orders of magnitude less than that for the total oceanic heat transport at 45°S (i.e., approximately -1.5×10^{15} watts), established indirectly from budget considerations (Niiler 1992). The portion of the v_E wave responsible for this net poleward eddy heat flux is that leading to the anomalous meridional Ekman heat advection that maintains the ACW against dissipation, not that leading to eastward and poleward phase propagation. This is also the portion of the v_E wave that leads to the tilt in the orientation of the covarying SST and SLP anomaly patterns in Fig. 1. As such, the spiral structure in the ACW anomaly patterns in Fig. 2 can be considered a signature not only of the coupling between ocean and atmosphere, but also of the net poleward eddy heat flux in the upper ocean by the ACW, weak as it is.

7. Numerical simulation of the coupled ACW

To demonstrate that this simple coupled model for the ACW is capable of simulating most of its observed wave characteristics in Figs. 1 and 2, we construct a numerical version of this coupled model by linking the numerical version of the global equilibrium climate model of WC98 with a numerical version of the anomalous heat budget model for the near-surface mixed layer

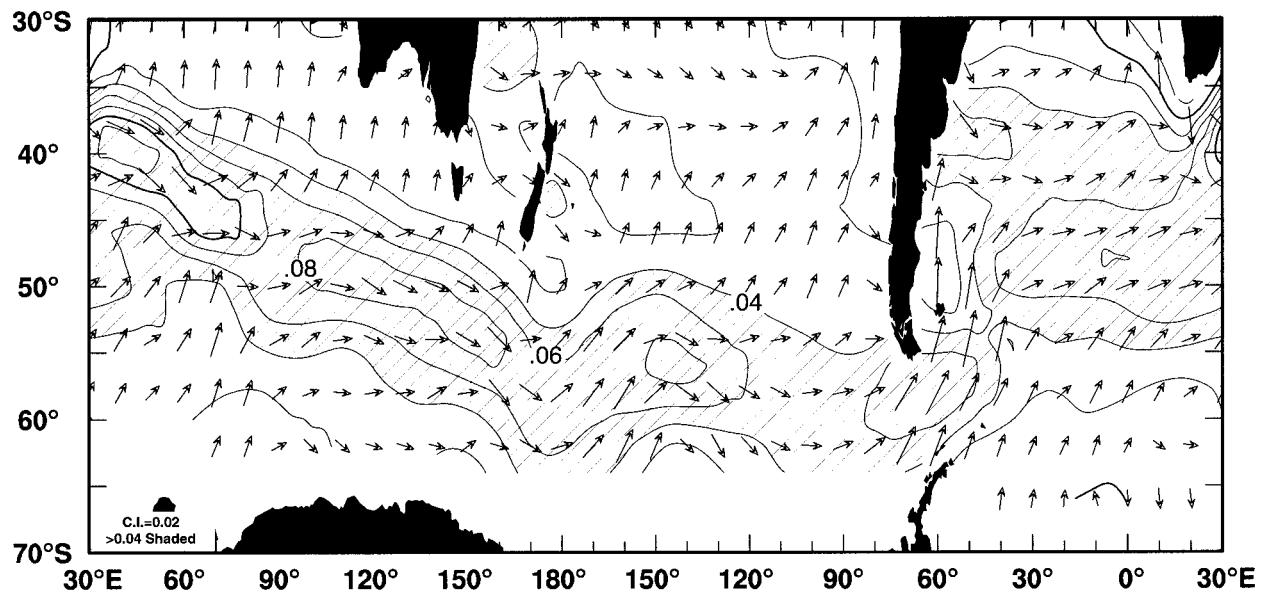


FIG. 4. Global distribution of the surface geostrophic plus Ekman flow over the Southern Hemisphere from 30° to 70°S. Vectors indicate flow direction and intensity, while contours indicate speed of the zonal component. The geostrophic component is made relative to 2000 db, computed from the long-term annual-mean geopotential anomaly computed from all available hydrographic data collected from 1950 to the present. The vertical-averaged Ekman component (i.e., directed 90° to the left of the surface wind stress in the Southern Hemisphere) was computed from the long-term annual mean wind stress (i.e., 1982–96) obtained from the NCEP Reanalysis Project (Kalnay et al. 1996), acting over the winter near-surface mixed layer depth scale of 100 m. Contour intervals are 0.02 m s⁻¹. Shading is for effect.

in the Southern Ocean given by Eq. (4.1). We extend the latter by including both meridional and zonal components of the mean background flow of the ACC in the Southern Ocean. The finite difference analogue for this coupled model is constructed on a 2° by 5° latitude–longitude grid with realistic coastline topography of the Southern Ocean and spatial-varying estimates of background surface currents and SST gradients. We choose numerical model parameters to be the same as analytic model parameters given above with one exception; that is, K^{-1} is evaluated with an e -folding timescale of 1 year. This is larger than 3 months utilized in the analytic model required to keep the numerical coupled model stable. In the analytic solution to Eq. (4.3), increasing K^{-1} in this way would have had no influence upon the eastward phase speed of the model ACW, but it should have increased the poleward phase speed by a factor of 4. Yet, as we shall see, the poleward phase speed in the numerical model is that predicted by the analytic coupled model. This suggests that increasing K^{-1} to balance some numerical noise and/or dispersion preserved the necessary thermodynamic balance. We initialize this numerical coupled model with an observed pattern of SST anomalies from April 1987 (Fig. 2) and determine whether it can simulate the space–time evolution of the ACW and maintain its amplitude against dissipation. We compare the observed ACW to the model ACW after 13 years of model integration, after which the model generates its own repeating evolution of the ACW independent of the April 1987 initial conditions.

In this numerical model, the mean surface flow is

composed of both geostrophic and Ekman components. The former is made relative to 2000 db, computed using the dynamic method from the available hydrographic data from the period 1950–89, gridded onto the model 2° lat. by 5° long. grid. The annual-mean Ekman flow is computed from 15 years (1982–96) of surface wind stress estimates from the NCEP Reanalysis Project (Kalnay et al. 1996) utilizing a mean near-surface mixed-layer depth scale in autumn–winter of $H_p = 100$ m, with vertical-averaged Ekman flow directed 90° to the left of the surface wind stress in the Southern Hemisphere. When this Ekman flow is added onto the geostrophic flow, the total surface flow (Fig. 4) displays the relatively narrow core of the eastward flowing ACC (i.e., with speeds ranging from 0.05 to 0.12 m s⁻¹) extending around the Southern Ocean between 45° and 60°S, embedded in the general eastward broadscale flow of the ACC (i.e., with speeds of approximately 0.05 m s⁻¹) occurring throughout the Southern Hemisphere from approximately 40° to 70°S. The Ekman flow produces a broadscale equatorward flow directed away from Antarctica over most of the Southern Ocean.

Given this broadscale surface flow pattern, we conduct two numerical experiments: one in the absence of coupling and another in its presence. In the absence of coupling (not shown), anomalous SST initial conditions from April 1987 are advected eastward by the broadscale flow of the ACC and equatorward by mean Ekman flow in all three oceans, consistent with Fig. 4, with amplitudes decreasing over time to insignificance after 6–8 years. Even so, a remnant of the initial SST signal

is advected by the ACC completely around the Southern Ocean after 12–14 years, nearly twice the time it takes (i.e., approximately 8 years) for the observed ACW to circle the globe (White and Peterson 1996). This indicates that the broadscale flow of the ACC is too slow (i.e., by nearly a factor of 2) to account for the eastward propagation of the ACW.

In the presence of coupling and beginning 13 years after initiation of model integration, an animation of SST anomalies from the coupled model (Fig. 5) reveals a model ACW with a 4-yr period and the required two wavelengths circling the globe in the Southern Ocean, similar to that observed in Fig. 1. This arises because the zonal coupling speed in Eq. (4.5b) approximately doubles the broadscale flow of the ACC. Moreover, magnitudes of the interannual SST anomalies are maintained against dissipation and dispersion of the initial SST signal. This occurs in response to the balance between anomalous meridional Ekman heat advection and anomalous sensible-plus-latent heat loss to the atmosphere expressed in Eq. (4.3). In addition, model SST anomalies in the Indian and Pacific sectors of the Southern Ocean display the characteristic elongation and tilt of their orientation into the southwest–northeast direction observed in Fig. 1 and predicted by Eq. (4.6). It also displays significant poleward propagation, where SST anomalies in the vicinity of New Zealand can be seen propagating eastward and poleward toward Drake Passage as observed in Fig. 1. The speed of this southward propagation is approximately 0.04 m s^{-1} , the same as that evaluated analytically in Eq. (4.7a). However, the model simulation differs significantly from that observed in the Atlantic sector of the Southern Ocean where the mean geostrophic flow of the ACC in Fig. 4 is underresolved, unable to simulate the strong advection of observed SST anomalies through and east of Drake Passage observed in Fig. 1.

We also integrated a version of the numerical coupled model where in the model atmosphere the Coriolis parameter southward of 30°S was set to a constant value (i.e., $f = f_o$ at 45°S) except where its meridional gradient is taken explicitly. From the analytical model in connection with discussion of Fig. 3, this step would be expected to eliminate eastward and southward coupling speeds but retain maintenance of the initial signal against dissipation. Indeed, this is what happened, with the model ACW maintained against dissipation and characterized by the southwest–northeast tilt in the alignment of the anomalies but propagating eastward only with the broadscale flow of the ACC in Fig. 4. This model experiment demonstrates that the fundamental physics of coupling in the numerical model is similar to that in the analytical model, with eastward and poleward propagation stemming from the f -squared dependency in the SLP response to SST anomalies in Eq. (6.1), which translates into a beta effect when the ocean and atmosphere are coupled.

We can also demonstrate the effect of this f -squared

dependency in the numerical simulation by examining the phase relationship between covarying SST and SLP anomalies in the model ACW (Fig. 6) as was done with observations in Fig. 2 and with the analytical coupled model in Fig. 3. With the latter we showed that the impact of the f -squared dependency upon SLP anomalies along the sloping boundary between warm and cool SST anomalies is to displace them southward from the latitude line connecting maximum SST anomalies, as observed in Fig. 2. This is also observed in the snapshot from the numerical simulation in Fig. 6.

This numerical coupled model is also capable of simulating the global spiral structure in covarying SST and SLP anomaly patterns associated with the ACW observed in the polar-stereographic projection in Fig. 6. This indicates that the global spiral structure is a signature for coupling between ocean and atmosphere. It is a manifestation of the southwest–northeast tilt in the orientation of the anomalies in the ACW in Eq. (4.6) and derives from the balance between anomalous meridional Ekman heat advection and anomalous sensible-plus-latent heat loss to the atmosphere. It does not depend upon the beta effect due to coupling. The latter is only responsible for eastward and poleward coupling speeds.

8. Discussion and conclusions

In the present study, covarying SST and SLP anomaly patterns in the ACW were observed to take on a global spiral structure in the Southern Ocean around Antarctica (Fig. 2). Earlier, White and Peterson (1996) had found poleward (equatorward) MSW anomalies to be aligned with warm (cool) SST anomalies in the ACW. Recently, WC98 had explained this association in terms of a balance between anomalous meridional planetary vorticity advection and SST-induced anomalous vortex stretching of the lower troposphere (see appendix). This indicated that SLP anomalies should straddle the boundary between warm and cool SST anomalies and be poleward intensified. Here, we observed this to be so, with the resulting pattern of SLP anomalies yielding ZSW anomalies that are displaced eastward of peak SST anomalies. Therefore, we hypothesized that associated anomalous meridional Ekman heat advection, and/or anomalous vertical mixing and anomalous sensible-plus-latent heat loss with which it is in phase, induces an anomalous SST tendency that accounts for the eastward propagation of the ACW.

To test this hypothesis, we constructed analytic and numerical models of this coupled system, with MSW anomalies in balance with underlying SST anomalies (WC98) and with associated ZSW anomalies providing a positive feedback to the ocean, forcing the anomalous SST tendency with anomalous meridional Ekman advection of the background SST field. The analytic model yielded the dispersion relation for free waves of the coupled system [i.e., Eqs. (4.4)–(4.7)], while the numerical model yielded a simulation for the evolution of

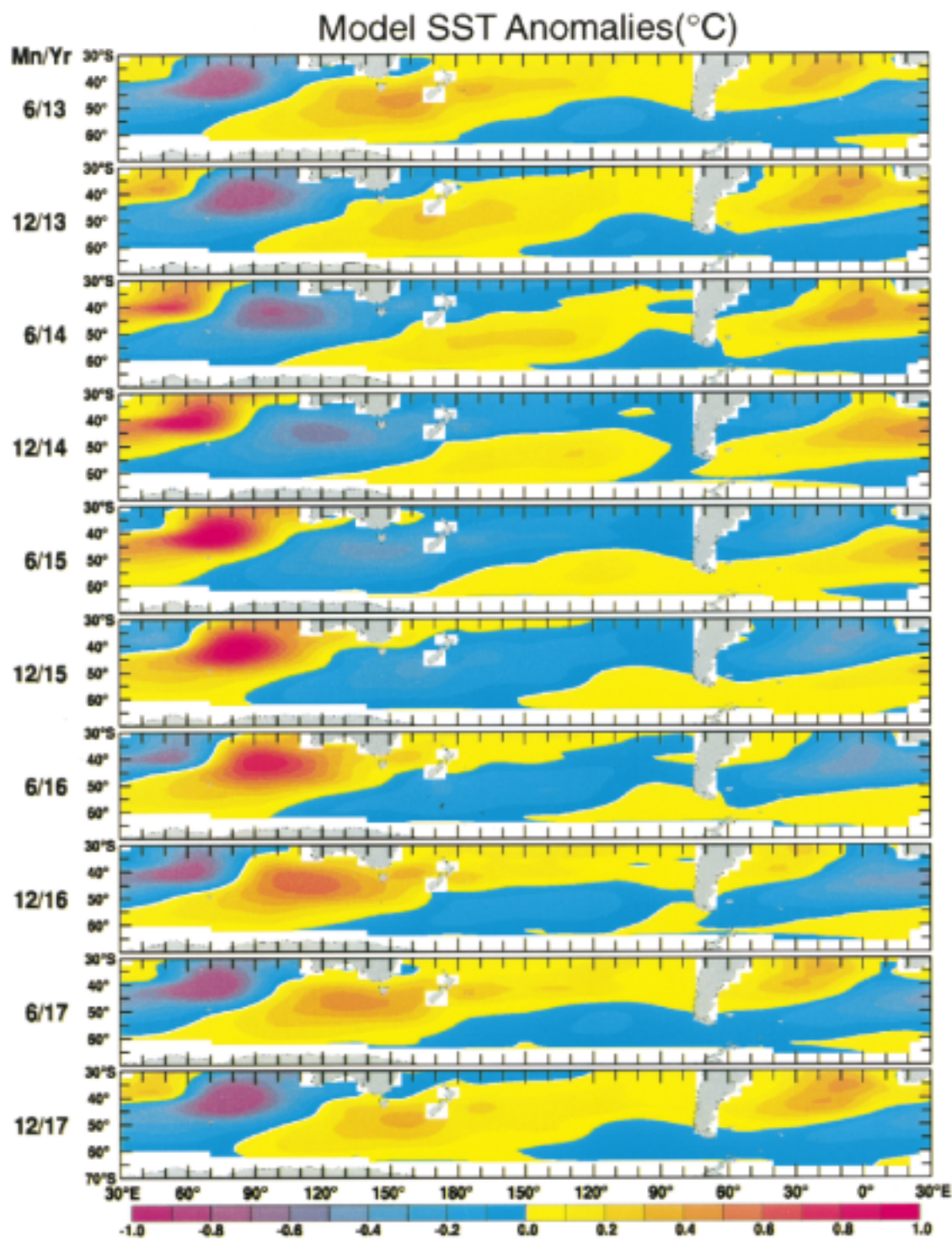


FIG. 5. Time sequence of interannual SST anomalies from the numerical coupled model of the ACW in the Southern Ocean, extending from 30°S to 70°S and zonally around the globe. The time sequence displays each January and June for 4 years beginning 13 years after the start of model integration from SST initial conditions given for April 1987 in Fig. 2. In each figure, the zonal mean was removed to enhance the structure of the ACW. Contours are given by the color bar ranging from $\pm 1.0^{\circ}\text{C}$.

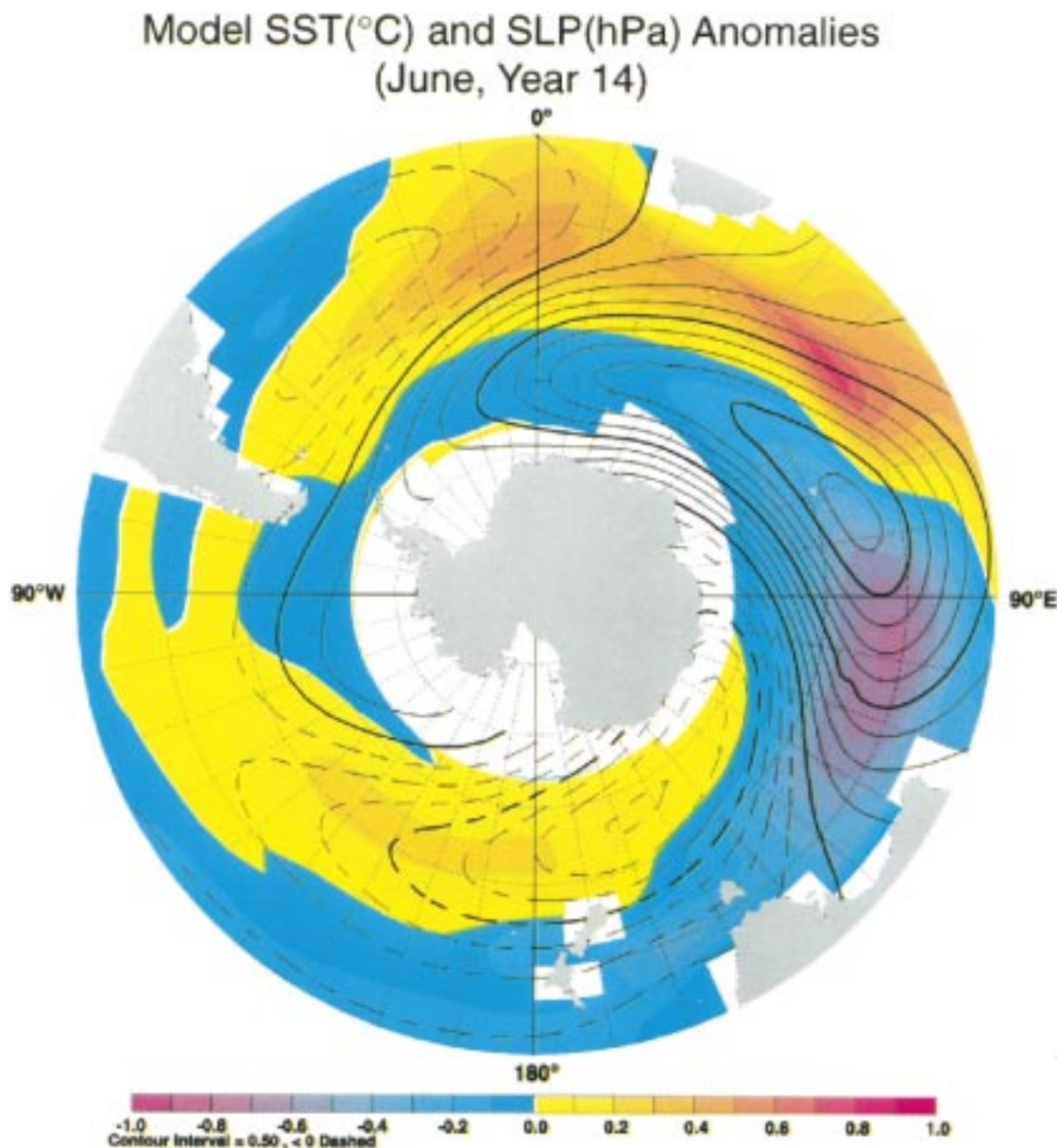


FIG. 6. Model SST and SLP anomalies for June in year 14 of the numerical model integration plotted contemporaneously on a polar stereographic projection centered at the South Pole and extending to 30°S. Anomalies were formed intrinsically in the model, but the zonal mean was removed to enhance the structure of the ACW. Contours of SST anomalies are given in the color bar, ranging from $\pm 1.0^{\circ}\text{C}$. Contours of SLP anomalies are at 0.5 hPa intervals, with positive anomalies given by thin solid lines and negative anomalies given by thin dashed lines, with the zero contour given by a thick black line.

the model ACW in the presence of spatially varying background SST, geostrophic plus Ekman flow, and coastline topography (Figs. 5 and 6). Both modeling efforts established the presence of eastward and poleward phase speeds of approximately 0.09 cm s^{-1} and 0.04 cm s^{-1} , respectively, with zonal wavelengths of 180° longitude (i.e., $1.1 \times 10^7 \text{ m}$ at 55°S) and meridional wavelengths of 50° latitude (i.e., $5.2 \times 10^6 \text{ m}$), all similar to those observed (summarized in Table 1). In the analytic model, these speeds and wavelengths stem from

requiring the coupled system to fluctuate with a period of 4.5 yr observed in the ACW by Peterson and White (1996). In the numerical model, a nominal 4–5 yr periodicity was generated intrinsically, with both frequency and wavenumber arising together from model scale and parameter estimation. We conclude that this particular frequency and wavenumber set arises as a global normal mode of the particular thermodynamics contained in this coupled ocean–atmosphere model of the Southern Ocean.

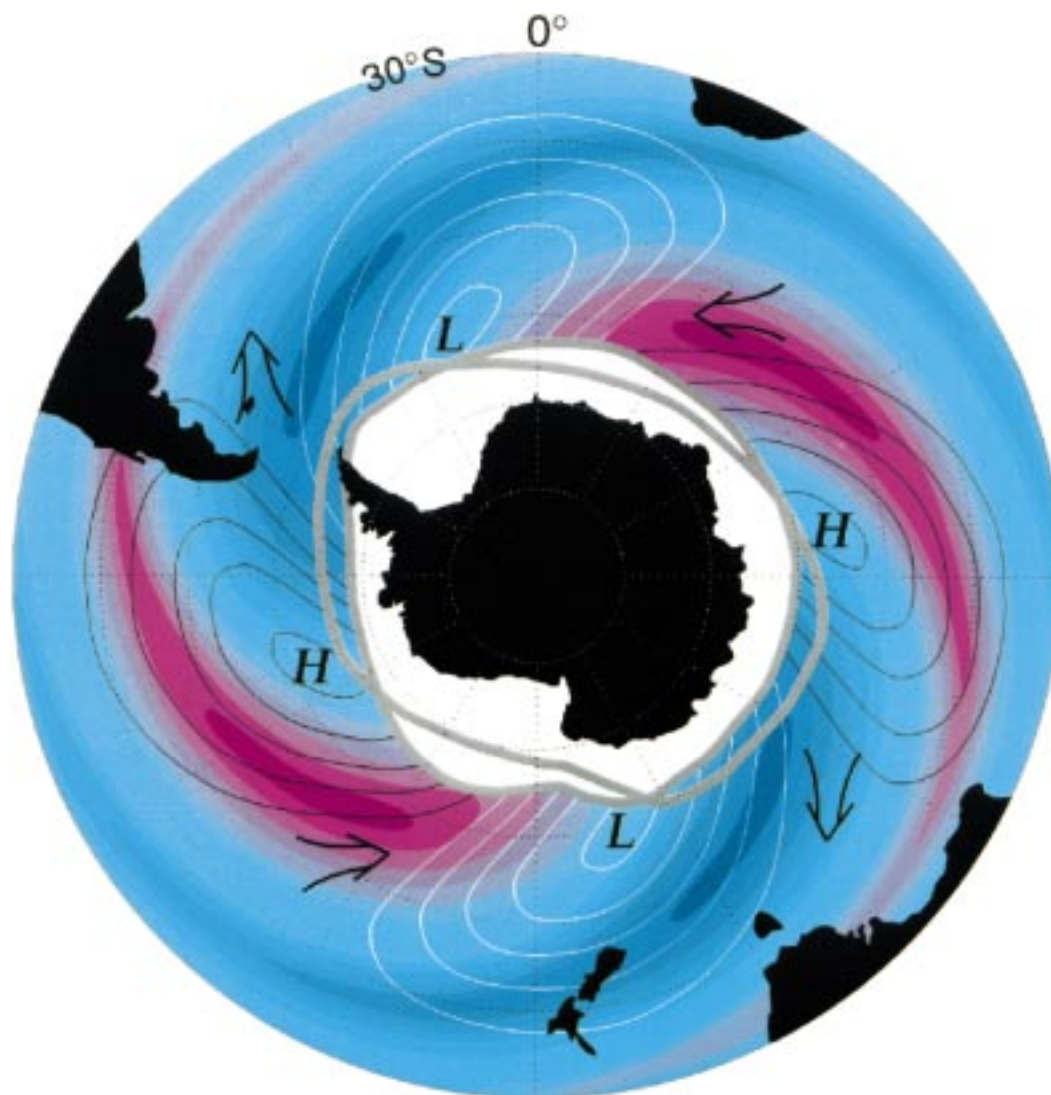


FIG. 7. Schematic diagram of interannual SST anomalies (red: warm, blue: cool) and SLP anomalies (black lines: high; white lines: low) observed in Figs. 1 and 2, representing their spatial patterns and phase relationships in the ACW over the Southern Ocean. Schematic meridional geostrophic wind anomalies are given by the broad arrows; anomalous sea ice extent is given by the gray lines. Information about sea ice extent is based upon a 13-yr dataset analyzed by White and Peterson (1996).

The thermodynamics of coupling leading to the eastward and poleward phase propagation of the ACW, and to its maintenance against dissipation, were explained through examination of the phase relationship between SST and SLP anomalies in both observations (Fig. 2) and numerical simulations (Fig. 6). These relationships are summarized in the schematic diagram given in Fig. 7, which is a significant improvement over that lacking the spiral structure introduced initially by White and Peterson (1996). In this schematic diagram, maximum SLP anomalies occur along the sloping boundary between warm and cool SST anomalies, displaced poleward from the latitude connecting maximum SST anomalies. In the analytic coupled model, we found this

displacement arising from the f -squared dependency in the anomalous SLP response to SST anomalies, yielding poleward intensification of the response. The spiral structure in this SLP anomaly pattern acts to bring anomalous westward (eastward) ZSW anomalies into phase with warm (cool) SST anomalies, allowing anomalous meridional Ekman heat advection to balance the anomalous sensible-plus-latent heat loss to the overlying atmosphere and thereby maintain the ACW against this form of dissipation. The poleward intensification of this SLP anomaly pattern yields a portion of the ZSW anomalies that drives anomalous meridional Ekman heat advection, and its corresponding anomalous SST tendency, eastward and poleward of SST anomalies. This accounts

for the eastward and poleward propagation of the covarying SST and SLP anomalies.

We also demonstrated that the model ACW is associated with a net poleward eddy heat flux from subtropical to polar latitudes, deriving from the coincidence of the anomalous poleward (equatorward) Ekman flow and warm (cool) SST anomalies in the near-surface mixed layer of the Southern Ocean. Yet, it is two orders of magnitude smaller than the net poleward transport of heat by the ocean (Niiler 1992). We demonstrated that it derives from the same coupled interactions that give rise to the maintenance of the ACW against dissipation, both of which are linked to the spiral structure in the global SST and SLP anomaly patterns in Fig. 7. Therefore, the latter becomes a signature not only for the coupling between ocean and atmosphere but also for the net poleward eddy flux of heat associated with the ACW, albeit weak.

Even though this coupling is restricted to the near-surface mixed layer of the ocean, the analytical model found downward displacements in the main pycnocline nearly in phase with warm SST anomalies, displaced only 10° of phase to the east. This is consistent with that observed by Jacobs and Mitchell (1996), where pycnocline displacements were inferred from altimetric sea-level height anomalies. But the interpretation given here is different from that given by Jacobs and Mitchell (1996). They proposed that anomalous vertical Ekman pumping and associated meridional geostrophic heat advection is the dominant advective influence upon the anomalous SST tendency. In our analytic model, we found this mechanism too weak; instead, we found anomalous meridional Ekman heat advection responsible for the feedback from atmosphere to ocean.

While the numerical version of the coupled model was able to simulate much of the observed evolution of the ACW in the Indian and Pacific sectors of the Southern Ocean, it was unable to simulate the observed jetlike advection of the ACW through Drake Passage and east of there in the Atlantic sector (Fig. 1). We ascribed this to the inability of the mean geostrophic flow (0/2000 db) used in the model to resolve the jetlike core of the ACC throughout much of this domain. To simulate the behavior of the ACW in this domain correctly will require high-resolution modeling of the ACC there. Moreover, the numerical model did not simulate maximum interannual variability in SST and SLP in the vicinity of the sea ice edge in the Pacific sector of the Southern Ocean (Peterson and White 1998). This may require the inclusion of model interactions between ocean, atmosphere, and sea ice anomalies. Furthermore, this model did not contain remote wind stress forcing arising from meridional atmospheric teleconnections (i.e., quasi-stationary Rossby waves in the atmosphere), which may have projected tropical ENSO signals into the Southern Ocean domain (Karoly 1989).

A number of assumptions need to be confirmed before the thermodynamics expressed in this coupled model can be accepted. Principal among these is the assumption

that the response of the atmosphere to anomalous SST forcing produces MSW and ZSW anomalies that follow the geostrophic winds in the lower troposphere. In this study, we have ignored the Ekman deflection of geostrophic wind anomalies in the near-surface planetary boundary layer of the lower troposphere. This might have yielded a component of anomalous Ekman heat advection in the near-surface mixed layer that significantly alters the brand of coupling between ocean and atmosphere presented here. Moreover, for the sake of simplicity this model assumed permanent winter conditions, when the near-surface mixed layer coincides with the top of the main pycnocline and when SST anomalies instigate midlevel diabatic heating in the overlying troposphere. We might expect that, had the numerical model included the annual cycle of atmospheric feedback and mixed layer dynamics, the resulting ACW would have exhibited at least a winter dominance and at most a chaotic element that would have transformed the periodic behavior of the model ACW into the quasiperiodic behavior of the limit cycle.

Finally, while anomalous meridional Ekman heat advection (i.e., linear relation to ZSW anomalies) was chosen here to represent the principal feedback mechanism from the atmosphere to the ocean, this feedback could also have derived from anomalous wind mixing at the base of the near-surface mixed layer and/or from anomalous sensible-plus-latent heat flux, both of which are also linearly related to ZSW anomalies. Also, anomalous cloud cover might have affected the anomalous SST tendency through its influence on anomalous air-sea radiation fluxes. These additional mechanisms, and others, will need to be sorted out quantitatively before we understand fully the coupling mechanisms responsible for the ACW.

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APPENDIX

Lower-Tropospheric Response to Extratropical SST Anomalies

White and Chen (1998) recently examined the global ECMWF monthly analysis for 8 years from 1985 to 1992, finding two robust statistical results on interannual timescales. First, they found low-level anomalous convergence in the extratropical troposphere associated with low-level anomalous meridional wind, consistent

with the tropospheric Sverdrup balance of Hoskins and Karoly (1981) and Palmer and Sun (1985), that is,

$$\beta v_A = -fD_A, \quad (\text{A.1})$$

where β is the meridional derivative of the Coriolis parameter f , and v_A and $-D_A$ are meridional wind and convergence anomalies, respectively, both vertically averaged over the lower troposphere. Second, they found the low-level convergence anomalies associated with warm SST anomalies over most of the global ocean.

To account for the latter association in the extratropics, White and Chen analyzed the anomalous tropospheric heat budget utilizing synoptic data from the NCEP Reanalysis Model (Kalnay et al. 1996) operated at SIO. They found maximum diabatic heating differences from austral winter 1992 to winter 1993 in the eastern Pacific sector of the Southern Ocean occurring at midtroposphere-levels and directly related to changes in underlying SST; that is,

$$Q = \alpha_A T, \quad (\text{A.2})$$

where Q is the midlevel diabatic heating anomaly, T is the SST anomaly, and α_A is an empirically derived coefficient equating the two variables. Moreover, they found these year-to-year changes in winter midlevel diabatic heating arising from changes in the midlevel latent heat release stemming from year-to-year changes in aggregate synoptic storm activity. This was consistent with an earlier hypothesis proposed by Namias (1985) that extratropical SST anomalies destabilize the overlying troposphere during the autumn–winter, thereby modifying the location and/or intensity of aggregate synoptic storm activity.

Subsequently, White and Chen argued that these midlevel diabatic heating anomalies are balanced chiefly by anomalous vertical heat advection in conserving potential temperature, with anomalous horizontal advection terms tending to cancel one another, yielding

$$w_A = Q/[g/(\rho_A N^2 C_{pA} H_A \bar{\Theta})], \quad (\text{A.3})$$

where w_A is the anomalous vertical velocity at midlevel; N^2 is the mean Väisälä frequency at midlevel [i.e., $(g/\bar{\Theta}) \partial \bar{\Theta} / \partial z$]; H_A is the mean height of the lower troposphere; $\bar{\Theta}$ is the mean potential temperature; ρ_A is the mean density of air; and C_{pA} is the specific heat of air. By applying conservation of mass to the lower troposphere (i.e., where $w_A/H_A = -D_A = -\partial u_A / \partial x - \partial v_A / \partial y$), the potential temperature balance in Eq. (A.3) can be linked with the empirical relationship in Eq. (A.2), yielding

$$D_A = \partial u_A / \partial x + \partial v_A / \partial y = -(\alpha / C_A^2) T, \quad (\text{A.4})$$

where C_A is the internal gravity wave speed (i.e., $NH_A = 50 \text{ m s}^{-1}$) and α is a coefficient [i.e., $(\alpha_A g) / (\rho_A C_{pA} \bar{\Theta}) = 2 \times 10^{-3} \text{ m}^2 \text{ s}^{-3} \text{ }^\circ\text{C}^{-1}$]. This relationship is consistent with the association observed between warm SST anomalies and low-level anomalous convergence.

This foregoing expression allows vertical-averaged zonal wind anomalies (u_A) in the lower troposphere to

be computed from knowledge of extratropical SST anomalies (T) and vertical-averaged meridional wind anomalies (v_A). The latter can be related directly to SST anomalies by linking anomalous mass and potential temperature conservation in Eq. (A.4) to anomalous vorticity conservation in Eq. (A.1), yielding

$$v_A = [(f\alpha) / (\beta C_A^2)] T. \quad (\text{A.5})$$

White and Chen (1998, manuscript submitted to *Quart. J. Roy. Meteor. Soc.*) demonstrated that this latter diagnostic relationship allows both magnitude and phase of global SLP anomaly patterns [i.e., $v_A = f^{-1} \partial P / \partial x$, where the vertical-averaged low-level pressure anomaly (P) is assumed equivalent to the SLP anomaly] to be diagnosed from observed global SST anomaly patterns for the 8 years of analysis from 1985 to 1992. This included the diagnoses of MSW anomalies (i.e., assumed equivalent to v_A) from SST anomalies in the ACW observed by White and Peterson (1996).

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