Mechanisms Determining the Atmospheric Response to Sea Surface Temperature Anomalies

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ABSTRACT

A simple model is used to study the mechanisms which control the local and remote (teleconnection) response of the atmosphere to the thermal forcing resulting from sea surface temperature (SST) anomalies located at various latitudes. The model chosen is a linear baroclinic spherical primitive equation model containing a zonally symmetric basic state with horizontal and vertical shear. An iterative procedure is developed in which the total diabatic heating resulting from the initial heating by the SST anomaly is calculated via feedbacks between the heating and the dynamic response of the system.

Depending on the latitudinal location of the SST anomaly, two major limits of atmospheric response may be identified. The first, the "diabatic limit", occurs with the SST anomaly embedded in weak low-latitude basic flow and results in a strong enhancement of the initial anomaly response through a vigorous positive dynamics-diabatic heating feedback. Strong teleconnections are evident between low and high latitudes. The second domain, the "advective limit", occurs when the SST anomaly is placed at higher latitudes in the vicinity of the westerly maximum. The local response is extremely small due to the creation of an indirect zonal circulation in the vicinity of the anomaly which is related to the strength of the local basic flow and the latitude of the forcing due to rotational limitations on the relative scale of the vertical velocity. In contrast to the diabatic limit, the form of the principal forced mode appears unimportant in determining the final response. That is, the dynamics-diabatic heating feedback is weak and only marginally positive.

The form of the remote high-latitude response in all cases is scale selective and only the largest scale transmitted modes are excited. It is argued that these are closest to resonance in latitudes of strong basic zonal flow. The remote response shows a distinct structural difference on either side of the westerly maximum, being highly baroclinic on the equatorial side but barotropic on the polar side. Limited cross-equatorial propagation occurs due to the existence of critical latitudes on the equatorial side of the forcing.

The model results are used to interpret the experimental results obtained from general circulation models (GCM) and provide a rationale for the existence of teleconnections found between low and high latitudes when SST anomalies were imposed in the equatorial oceans. Furthermore, the results suggest why "super-anomalies" were required in midlatitudes in some GCM experiments in order to produce a response which was measurable above the noise level of the model.

It is shown that it is possible to resolve the apparent paradox between the minimal response of GCMs to the imposition of middle latitude SST anomalies and the observations of Namias (1976a) and Davis (1978) who related atmospheric anomalies at least relative to summer SST anomalies. It is argued that only at times of small basic flow (i.e., summer) will a significant response arise in midlatitudes. Finally, the relevancy of the model results to such features as the South Pacific cloud band and the Southern Oscillation is discussed.

1. Introduction

Sea surface temperature (SST) exhibits a marked seasonal variation which is a strong function of both latitude and longitude. The interactive nature of the ocean with the atmosphere above via heat, moisture and momentum fluxes at the interface is well known. This, together with the zonal asymmetries in the lower boundary condition introduced by the spatial finiteness of the oceans and the existence of land areas define the seasonally evolving climatic envelope within which weather exists (Webster and Lau, 1977). Superimposed on this evolving SST distribution are persistent and large-scale temperature anomalies. The role the anomalies play in producing significant deviations from the climatic envelope is a subject of speculation and, from a climatic forecasting viewpoint, one of considerable anticipation and hope. Large-scale SST anomalies appear as the most easily monitored potential perturbers of the climatic system. It is not surprising that considerable effort has
been expended in attempting to understand the long-term atmospheric response to the imposition of an SST anomaly. Significant empirical evidence exists which promotes the climatic role of the SST anomaly, especially in the equatorial Pacific Ocean (e.g., Bjerknes, 1966; Namias, 1976b; Horel and Wallace, 1981).

The logical extension of the observational studies has been numerical experiments aimed at elucidation by simulation. Hitherto the study of the atmospheric response to SST anomalies has resided in the realm of the GCM, which probably arises from an a priori expectation of an extremely complicated response inherited from the observational studies. Most experiments were performed with the SST anomaly either placed in the midlatitudes or the low latitudes. However, despite the large number of experiments performed with many different models, a clear picture of the atmospheric response has yet to emerge.

Rowntree (1972) performed the first tropical SST anomaly experiment. He found a strong local and remote response, which possessed many of the characteristics noted by Bjerknes (1966), when a 3.5°C anomaly was maintained in the eastern equatorial Pacific Ocean. However, the impact of Rowntree’s study was dulled somewhat by questions relating to the model suitability for the particular experiment. In particular, Ramage and Murakami (1975) argued that the equatorial wall in Rowntree’s model was responsible for the large response. The substantiation of Rowntree’s results had to wait until statistically significant responses were obtained by Julian and Chervin (1978) and by Wells (1979). These encouraging results for equatorial anomalies were not reflected in experiments when the anomalies were placed in midlatitudes. Spar (1973), Kutzbach et al. (1974), Simpson and Downey (1975) and many others all showed similar non-defined results. The response to midlatitude SST anomalies, at least as simulated by GCMs produced a very weak response which did not emerge above the inherent noise level of the model. In fact, in order to produce a statistically significant response Chervin and Schneider (1976) and Kutzbach et al. (1977) required a super anomaly of magnitude ±12°C located in the northern Pacific Ocean. A detailed review of the many model studies of ocean circulation and climate interaction is given by Haney (1979).

The different significance of the response as the latitude of the anomaly is changed introduces an element of quandry into the assessment of the importance of SST anomalies in the actual climatic system. For example, if we believe the results of GCM experiments we may conclude that the system is insensitive to excitation by SST anomalies in midlatitudes but responsive when the anomaly is placed closer to the equator. Such a conclusion would require reconsideration of some observational studies of the effect of midlatitude anomalies (e.g., Namias, 1976a). However, we must consider the other possibility that the current technical state of GCMs is incapable of assessing the effect of SST anomalies. If we adopt this stance we would be forced to omit the Rowntree (1972) and Julian and Chervin (1978) results, together with what would be concluded as inconclusive results for the midlatitudes. It is with this dilemma in mind that we propose the present study where the contention is tested that the selective response of the GCM experimentation is correct and that the location of the anomaly determines the degree of response.

It must be emphasized that despite the fairly universal result from GCM experiments that the influence of the SST anomaly on the atmosphere is minimal when the anomaly is located at higher latitudes, there does appear considerable evidence of the existence of relationships between the SST anomalies and atmospheric pressure anomalies. For example, Namias (1973) suggested relationships between monthly mean SST anomalies and atmospheric indices thus indicating an apparent disparity between GCM and statistical studies. Even though some doubt was cast on the statistical studies by Davis (1976) it was eventually shown by Namias (1976b) and Davis (1978) that the relationships were seasonally selective and that only relationships appear to exist between summer SST anomaly patterns and subsequent autumn and winter atmospheric indices.

Consequently, the testing of the contention regarding the GCM results carries with it a rider. Is it possible to contain within the framework of an explanation of the spatial selectivity of the GCM response an understanding of the temporal selectivity of the response of the atmosphere to the imposition of SST anomalies as indicated by Namias (1976b) and Davis (1978)?

Rather than utilize a complex model we will seek an understanding of the GCM results via a much simpler system. In fact, we use a linear spherical and baroclinic primitive equation model similar to that of Webster (1972). However, we introduce one significant deviation from Webster where the linear perturbation response of a prescribed basic state to an imposed total diabatic heating distribution was sought (see, also, Gill, 1980). In reality the total diabatic heating is made up of radiative, sensible and latent heating components. The sensible heating may be attributed directly to the SST anomaly but the latent heating depends on the dynamic response of the system to the sensible heating. Thus, if the scheme of imposed heating of Webster (1972) is utilized we would need to presuppose the dynamic
response of the system to the SST anomaly. To overcome this major deficiency we will develop an iterative scheme which allows a dynamic response-total diabatic heating feedback process to establish the total diabatic heating.

In the ensuing study, SST anomalies are introduced at the lower boundary of the model at various latitudes with the form, scale and magnitude held constant. The return atmosphere to ocean feedback is purposely omitted even though it is realized that in a real climate system the feedback may cause significant alteration to the initial SST anomaly. The main motivation of this study is to understand the atmospheric response to a fixed and maintained SST distribution. It is felt that such an understanding is a necessary goal prior to tackling the fully coupled system.

2. The basic model

The rudiments of the basic model are given by Webster (1972). The equations which govern the system are written for a primitive, spherical and baroclinic atmosphere. Linearizing about a basic state \( \bar{U}(\phi, p) \), \( \bar{\psi}(\phi, p) \) and assuming steady-state conditions, the following system evolves:

\[
\frac{\bar{U}}{a \cos \phi} u_\lambda + \frac{1}{a \cos \phi} (\bar{U} v) \cos \phi_c = -\frac{1}{a \cos \phi} \psi_\lambda \\
+ (\bar{U} \omega)_p = - \frac{1}{a \cos \phi} \psi_\lambda \\
+ v(f + \tan \phi \bar{U}) + F_1, \quad (1)
\]

\[
\frac{\bar{U}}{a \cos \phi} V_\lambda = - \frac{1}{a} \psi_\phi - fu + \frac{2u\bar{U} \tan \phi}{a} + F_2, \quad (2)
\]

\[
\frac{\bar{U}}{a \cos \phi} T_\lambda + \frac{v}{a \cos \phi} \bar{T}_\phi = \omega \bar{S} + \frac{\bar{Q}}{C_p} + F_3 \quad (3)
\]

\[
\omega_p + \frac{1}{a \cos \phi} (v \cos \phi)_\phi + \frac{1}{a \cos \phi} u_\lambda = 0, \quad (4)
\]

\[
\psi_p = - \frac{RT}{p}, \quad (5)
\]

where

\[
\bar{S} = \frac{R\bar{T}}{C_p p} - \frac{\partial \bar{T}}{\partial p}.
\]

Here \( \lambda, \phi \) and \( p \) represent the three independent variables of longitude, latitude and pressure and the dependent variables \( (u, v, \omega, T, \psi; \lambda, \phi, p) \) are perturbation quantities about the basic state:

\[
- \frac{1}{a} \bar{\psi} + f\bar{U} + \bar{U}^2 \tan \phi \frac{\psi}{a} = 0, \quad (6)
\]

\[
\bar{\psi}_p = - \frac{R\bar{T}}{p}, \quad (7)
\]

where \( \bar{\psi} \) is the geopotential field associated with the prescribed basic zonal flow \( \bar{U} \). The basic field may be thought of as the time-averaged zonally symmetric flow resulting from an unspecified external forcing. \( \bar{Q}(\lambda, \phi, p) \) is the total diabatic heating rate which will be discussed in detail in Section 3.

The formulation of the dissipative processes \( F_i \) also follow Webster (1972) and are expressed as

\[
F_i = \begin{cases}
-K_1u - K_2 \frac{\partial u}{\partial p}, & i = 1 \\
-K_1v - K_2 \frac{\partial v}{\partial p}, & i = 2 \\
-K_3T, & i = 3.
\end{cases} \quad (8)
\]

In this simple representation the dissipative processes are reduced to a Rayleigh friction, a vertical mixing dissipation and a Newtonian-type radiational cooling. The \( K_i \) represent a drag coefficient, a vertical mixing coefficient and a radiative cooling coefficient with values representing 6-, 25- and 40-day decay rates, respectively. It is important to note that the dissipative parameterizations are not scale dependent. That is, the \( K_i \) are independent of scale.

It is convenient to define the following Fourier expansions for perturbation quantities:

\[
u, v, \omega, T(\lambda, \mu, p) = \text{Re} \sum_s [U^s, V^s, \Omega^s, T^s(\mu, p)] \exp(is\lambda), \quad (9)
\]

\[
\bar{Q}(\lambda, \mu, p) = \text{Re} \sum_s [q^s(\mu, p)] \exp(is\lambda),
\]

where \( \mu = \sin \phi, s \) is a non-negative integer and \( \text{Re} \) signifies the real part of the Fourier expansion.

The system is represented in the vertical as a two-layer atmosphere with mid-levels at 750 and 250 mb denoted by subscripts 2 and 1, respectively. The model atmosphere is shown in Fig. 1. As orography is omitted from the present study the vertical boundary conditions are represented identically as

\[
\omega(p = 0, 1000 \text{ mb}) = 0.
\]

The momentum and continuity equations are ascribed to the 250 and 750 mb levels (i.e., levels 1 and 2 in Fig. 1) and the thermodynamic equation at the mid-level. With a vertical differencing format and the Fourier expansion (9) a set of two-dimensional \((\phi, p)\) spectral differential equations evolve. By the use of a staggered spatial centered difference scheme (see Webster, 1972) 6 first-order difference equations are produced, which, by successive
elimination are reduced to the following two coupled, linear, complex, second-order difference equations in $V^*_1$ and $V^*_2$ (i.e., in the Fourier coefficients of the upper and lower level meridional velocity components):

$$
m_1^kV^*_1^{k+1} + m_2^kV^*_2^k + m_3^kV^*_1^{k-1} + m_4^kV^*_2^{k-1} + m_5^kV^*_3^k + m_6^kV^*_4^{k-1} = r^k
$$

$$
n_1^kV^*_1^{k+1} + n_2^kV^*_2^k + n_3^kV^*_1^{k-1} + n_4^kV^*_2^{k-1} + n_5^kV^*_3^k + n_6^kV^*_4^{k-1} = t^k
$$

where the $m_1^k$ and $n_1^k$ are complex coefficients and $r^k$ and $t^k$ are complex non-homogeneous coefficients, possessing the functional form

$$
m_1^k, n_1^k, r_1^k, t_1^k = \text{fn}(\mu, s, \bar{T}, \bar{U}, K_1, K_2, K_3, q^*).
$$

Eq. (10) is expressed at the latitudinal grid points $k = 1, 41$ between the poles producing 156 linear, real simultaneous equations in 164 unknowns. The system is closed by the boundary condition that the meridional velocity vanishes at the poles. The resulting $156 \times 156$ coefficient matrix is inverted via numerical techniques and repeated for each longitudinal wavenumber within the limit $s = 1, 6$.

It would seem that the choice of a two-layer vertical format would have only weak justification. Inviscid calculations (Webster, 1972) indicate that the planetary-scale (horizontal) low-frequency modes of the tropical troposphere possess extremely small vertical scales which were about a factor of 5 smaller than those observed. Chang (1977) was able to reconcile the difference between the predicted (inviscid) and observed vertical scales and showed that it resulted from the neglect of dissipative processes in the theoretical calculations. By including a dissipative time-scale characteristic of cumulus momentum transports, Chang found that the low-frequency modal structure within the troposphere can be identified with a class of waves dominated by viscous processes. The vertical scales of the viscous modes are of the scale of the troposphere. The result was corroborated by Webster (1977) who showed that the largest horizontal scales were stretched in the vertical by about an order of magnitude beyond the inviscid modal limit. Such stretching is scale selective with the vertical scale rapidly approaching the inviscid limit as wave-number increased. However, for the horizontal scales of motion considered here the vertical scales suggested by Chang (1977) and Webster (1977) support the two-layer approximation.

3. Diabatic heating

The major problem in devising a heating parameterization in a steady state system is that the largest component of the heating (the latent heating) depends on a cooperative feedback between the dynamic response of the system to the initial SST anomaly and the subsequent heating. Thus steady-state solutions are difficult to achieve unless the heating function is stipulated. In this section we will devise an alternative technique in which the heating is determined by an iterative procedure which acknowledges the dynamic response of the system.

Latent heating parameterizations describing convective processes often adopt the form

$$
\tilde{Q}_L = B \Delta\{\omega\},
$$

where $B$ represents aspects of the large-scale flow and $\Delta\{\}$ is the Heaviside function which is unity for $\omega < 0$ (ascending air) or zero for $\omega > 0$. In all subsequent calculations $B$ is assumed to be $5 \times 10^2$ m$^3$ kg$^{-1}$. However, $\Delta\{\omega\}$ is a nonlinear function and can only be calculated iteratively or by adopting a time-marching scheme. The iterative procedure we adopt computes $\tilde{Q}_L$ at each iteration utilizing the steady-state system structure determined at the previous iteration.

The iterative scheme is best illustrated using a one-dimensional example. The total diabatic heating is assumed to be the sum of a sensible heating input due to an SST anomaly and latent heating which is related to the dynamic response of the system to the total heating. Furthermore, we assume that the SST anomaly is represented by a cosine function with amplitude $A$ such that

$$
\tilde{Q}_1 = A(1 + \cos\lambda),
$$

where $\tilde{Q}_1$ is the total diabatic heating at iteration 1 ($N = 1$). At $N = 2$ we assume that the total heating ($\tilde{Q}_2$) is made up of the SST anomaly heating ($\tilde{Q}_1$) plus the heat generated by the latent heat release in the ascending air in the vicinity of the SST heating. With (12) we can write

$$
\tilde{Q}_2 = \tilde{Q}_1 + \tilde{Q}_{22} = \tilde{Q}_1 - B \Delta\{\omega_1\},
$$

where $B$ is a constant. In the cases discussed in
subsequent sections \( \omega'_1 \) is determined by the full dynamic model described by the set (10) and (11). However, in this example we assume the simple relationship that the diabatic and adiabatic heating is nearly exactly balanced. That is, from (3) we can write

\[
\dot{\omega}' = -\dot{Q}'_1, \tag{15}
\]

where \( \dot{S} = \kappa \tilde{T} p \) and \( \omega'_1 \) and \( \dot{Q}'_1 \) are perturbation quantities. \( \dot{Q}'_1 \) is calculated by subtracting the zonal mean of (13) from (13) so that \( \dot{Q}'_1 = \cos \lambda \). Then using (15) the total heating at \( N = 2 \) is given by

\[
\dot{Q}_2 = A [1 + \cos \lambda + B \tilde{S}^{-1} \Delta (\cos \lambda)]. \tag{16}
\]

The heating for successive iterations may be calculated in a similar manner; e.g.,

\[
\dot{Q}_3 = \dot{Q}_2 + \dot{Q}_{13} = \dot{Q}_2 - B \{\omega'_2\}, \tag{17}
\]

so that

\[
\dot{Q}_3 = A \{1 + \cos \lambda + 2B \tilde{S}^{-1} \Delta (\cos \lambda) \}
+ B^2 \tilde{S}^{-2} \Delta [\Delta (\cos \lambda) - \Delta (\cos \lambda)]}, \tag{18}
\]

and so on.

It is interesting to study the scale change which occurs with the heating function with successive iterations. Between iterations 1 and 2, the heating function is changed by the addition of \( B \tilde{S} \Delta (\cos \lambda) \) [cf. (13) and (16)]. The effect is to peak the positive part of the cosine function which can only be accomplished by successive contributions from smaller scales. The iterative nature of the heating scheme is shown schematically in Fig. 2a.

Fig. 2b shows the growth of the diabatic heating over 25 iterations with \( \omega \) being calculated by (15). The solid curves show the individual Fourier coefficients for \( s = 1 \) to 4 and the total (i.e., summed) diabatic heating. As \( N \) increases the total diabatic heating grows without bound principally by the increase in amplitude of the higher wavenumber components. To indicate (for later reference) the effect of dynamic constraints in the heating function, two cases in which \( \omega \) was determined by the full dynamic system [i.e., (10) and (11)], rather than by (15), are shown as dashed lines on Fig. 2. With SST anomalies placed at the equator (indicated by \( \mu_0 = 0 \)) and at 23.6° (\( \mu_0 = 0.4 \)) the magnitude of the total heating function rapidly approaches an equilibrium value after about 10 iterations. The bifurcation of the curve in the \( \mu_0 = 0.4 \) case illustrates a change in latitudinal location of the maximum total diabatic heating which will be discussed later together with the merits of the scheme described above.

4. The iterative scheme

The following procedure is adopted. An initial SST anomaly is assumed and the steady-state response to the sensible heating produced by the anomaly is sought within the system described by (10) and (11). The steady-state vertical velocity field is used to calculate the latent heating field using (12). A new steady-state atmospheric response is calculated relative to the updated heating field and the procedure is continued until an equilibrium solution

Fig. 2b. Growth of the diabatic heating over 25 iterations for the total function and for the first four Fourier coefficients. The solid lines indicate the evolution of the heating with no dynamic feedback while the dashed lines show the evolution for SST anomaly placements at the equator (\( \mu_0 = 0 \)) and at 24° (\( \mu_0 = 0.4 \)) with dynamic feedback.
is approached. An equilibrium solution is defined to occur when the steady-state solutions from successive iterations are the same.

5. Experiments

Six experiments were performed. In all cases the basic fields $\bar{U}_1$ and $\bar{U}_2$ were given by expressions which approximate the observed annual mean zonally averaged wind fields at 250 and 750 mb, respectively. These are

$$\bar{U}_i = C_i \sin \{\frac{\pi}{2}(1 + \mu)\} + D_i(1 - \mu^2), \quad i = 1, 2,$$

where $\mu = \sin \phi$. The fields with $C_1 = 18$, $C_2 = 7$, $D_1 = 14$ and $D_2 = 2$ are shown in Fig. 3.

The following initial heating function was used in all experiments:

$$\dot{Q}(\lambda, \mu, p) = A \sin \lambda \times \exp[-b(1 - \mu^2) - \mu_0^2]f(p),$$

where $f(p)$ describes an equipartition (i.e., linear distribution) of the heating with height and $\mu_0$ the latitudinal displacement from the equator of the SST. $b$ sets the latitudinal decay of the anomaly and was chosen to $e$-fold within 10° of $\mu_0$. The magnitude of the anomaly is given by $A$. $A = 50 \times 10^{-4}$ J kg$^{-1}$ s$^{-1}$ corresponds to a surface sensible heat flux of 50 W m$^{-2}$ which would result from an anomaly of 5°C at the center of the distribution given by (20). Experiments were run for values of $\mu_0 = 0$, 0.1, 0.2, 0.4, 0.6 and 0.7 which correspond to latitudinal anomaly locations of 0, 5.7, 11.5, 23.6, 36.9 and

44.4°. The location of the center of the anomaly relative to the basic wind fields shown in Fig. 2 should be noted.

The same heating format is used in each experiment even though it is realized that atmospheric stability conditions at higher latitudes would most likely insist with a different heating partition in the vertical than the equipartition used in (20). However, the use of a tropical heating parameterization at higher latitudes is a worst case choice to test the contention discussed in the Introduction that the apparent diminution of atmospheric response with increasing latitudinal position of the SST anomaly is real.

6. Results

Fig. 4 shows the variation of the maximum value of the diabatic heating as a function of the iteration $N$ for the various $\mu_0$ with $\omega$ determined by the dynamic model. Considerable latitudinal variation is evident. With the anomaly centered at low latitudes ($\mu_0 = 0$, 0.1 and 0.2) the magnitude of the heating converges at values three times greater than the initial SST anomaly heating. For the cases where $\mu_0 > 0.4$, the final magnitude is considerably smaller. In fact, for $\mu_0 = 0.6$ and 0.7 the values of the total heating recombine in the magnitude of the initial anomaly. The dashed line indicates that the maximum value of the forcing has shifted to the indicated location. For example, with the anomaly placed at $\mu_0 = 0.4$, the maximum heating initially resides at the latitude of the anomaly but eventually resides closer to the equator at $\mu = 0.3$. This transition is labeled 0.4* (0.3) and indicated by the dashed

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1 Estimation of the magnitude of the SST anomaly for a given sensible heat flux assumed that the anomalous heating was distributed evenly in the vertical. Climatological values of the mean lower tropospheric wind speed (5 m s$^{-1}$) were chosen.

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Fig. 3. The basic zonal velocity components at 250 mb ($\bar{U}_1$) and at 750 mb ($\bar{U}_2$). Arrows on abscissa indicate the locations of the SST anomalies for various experiments.

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Fig. 4. Growth of the total diabatic heating (units: $10^{-4}$ J s$^{-1}$ kg$^{-1}$) for the SST anomalies placed at various latitudes when full dynamic interactions are permitted. Dashed curves indicate a latitudinal shift of location of the maximum diabatic heating from the initial latitude to the latitude indicated in the parenthesis.
curve. Thus two forms of equilibrium heating functions evolve via the dynamics-diabatic heating feedback. In the low-latitude cases the final heating is large, whereas when the SST anomaly is placed in the middle or high latitudes the heating shows little or no growth.

The global distributions of $\dot{Q}$ and $\omega$ at $N = 1, 5$ and 11 are shown for the $\mu_0 = 0$ and 0.4 cases in Figs. 5 and 6. Both cases indicate a substantial diminution of the longitudinal scale of the rising motion compared to regions of descent. The ratio reduces from 1:1 at $N = 1$ to 1:3.5 at $N = 11$. It should also be noted that the magnitude of the vertical velocity with $\mu_0 = 0.4$ is considerably less than the converged $\mu_0 = 0$ case which is consistent with the latitudinal diminution of $\dot{Q}$.

The latter observation is underlined by Fig. 7 which shows the vertical velocity field for $N = 11$ with $\mu_0$ set at 0.1, 0.2 and 0.6. Whereas the characteristics of the vertical velocity response are similar to those noted in either Fig. 5 or 6, the diminution of vertical velocity with increasing latitude is most evident, even though the magnitude of the initial SST anomaly was the same. However, it is important to note that with increasing latitude the maximum heating and the maximum upward flow become progressively out of phase. To illustrate this the center of maximum heating at iteration 11 is plotted on the $\omega$ fields of Fig. 7 as a solid circle.

Figs. 8–12 show the associated perturbation fields at iteration 11 for the anomaly locations indicated in Fig. 3. All fields are shown in the same format; the appropriate horizontal velocity vectors are plotted on the upper and lower fields of geopotential.

With the anomaly located in low latitudes (e.g., $\mu_0 = 0, 0.1$) Figs. 8 and 9 the model produces a response with considerable latitudinal dependence. At low latitudes the response is primarily in the zonal velocity component. However, correlations between the small meridional velocity component and the geopotential height fields indicate a poleward wave-energy flux which appears to be manifested in a significant remote response at high latitudes. Furthermore, the geopotential response in the region of the anomaly is considerably weaker than that occurring in the remote high latitudes, the character of which is basically barotropic. Also, the high-latitude remote response appears to possess a larger spatial scale than the local response. That is, the remote response is more confined to the lower end of the spectrum than is the local response.

In the subtropical and middle-latitude cases ($\mu_0 = 0.2$ and 0.4), the form of the response changes considerably. The magnitude and latitudinal scale of the perturbation velocity fields increase with the fields showing considerable phase tilt with latitude even though $\omega$ and $\dot{Q}$ have decreased. The local response is baroclinic and possesses a maximum geopotential response in the region of forcing. In the northern high latitudes the upper and lower geopotential fields are similar to the $\mu_0 = 0$ case in that they have adopted a barotropic format.

When $\mu_0 = 0.6$ (Fig. 12) all fields show a considerable decrease in magnitude. For example the zonal perturbation velocity component has decreased by nearly a factor of two from the $\mu_0 = 0.4$ case which matches a similar diminution of the $\omega$ field. The $\mu_0 = 0.7$ case is not shown, the response being extremely similar to that shown in Fig. 12.

The high latitude cases are anomalous for another reason. With $\mu_0 = 0.1$ and 0.2 the local geopotential maxima are poleward of the region of maximum forcing. For example, for $\mu_0 = 0.1$, the maximum upper level response occurs at $\mu = 0.28$ and for $\mu_0 = 0.2$, at $\mu = 0.35$. In contrast, when $\mu_0 = 0.4$ the maximum lies almost on the same latitude line as the maximum forcing, while for $\mu_0 = 0.6$ the maximum lies slightly to the south of the heating. However the location of the maximum remote geopotential response is nearly identical for each case (including when $\mu_0 = 0$) and is located in the vicinity of $\mu = 0.85$.

Figs. 5 and 8 indicate completely different phase variations for the geopotential and vertical velocity fields. Whereas the troughs and ridges of the Northern Hemisphere slope from southeast to northwest, the region of ascending motion slopes from the SST anomaly in a northeast direction. With the anomaly placed at higher latitudes (e.g., $\mu_0 = 0.4$ as displayed in Figs. 6 and 11) the phase orientation of the geopotential fields and the vertical velocity are both to the northeast. However, the vertical velocity maximum is well to the east of the geopotential maxima and the SST anomaly.

7. Interpretation

\[ a. \text{Form of the response} \]

From Webster (1972, 1973b) it is apparent that two major forms of response may be manifested by the steady-state system. These are the Kelvin modes for which stationary solutions are possible in the easterly basic flow of the equatorial regions and the Rossby modes which are excitible in the midlatitude westerlies.

Figs. 5 and 8 show a near-classical rotationally trapped Kelvin mode excited by the equatorial ($\mu_0 = 0$) SST anomaly. The mode is symmetric about $\mu = 0$, nearly two dimensional ($\mu \gg \nu$) with the zonal velocity field and the geopotential in quadrature and the magnitude of the response decaying exponentially away from the equator. Longitudinal phase variations in the response are evident poleward of the local response and these are explicable in terms of viscous effects (Chang,
Fig. 5. Total diabatic heating (units: $10^{-4}$ J s$^{-1}$ kg$^{-1}$) and vertical velocity (units: $10^{-3}$ N m$^{-2}$ s$^{-1}$) at iterations 1, 5 and 11 with the SST anomaly located at the equator ($\mu_0 = 0$).
Fig. 6. As in Fig. 5 but with the SST anomaly located at 23.6°N (μ = 0.4).
FIG. 7. Converged (iteration 11) vertical velocity fields (units: $10^{-3}$ N m$^{-2}$ s$^{-1}$) for SST anomaly locations 5.7, 11.5, and 36.9°N. Solid circle indicates the location of the diabatic heating maximum.

FIG. 8. Associated fields of horizontal velocity and geopotential (units: m) at 250 and 750 mb for the SST anomaly located at $\mu_0 = 0$ (6°N). Velocity vector scale denoted in bottom left corner of panel. Note scale change between upper and lower levels. Vectors of magnitude $< 0.5$ m s$^{-1}$ are omitted.
tropospheric cyclone. The maximum winds which occur on the equatorial side of the heat source are a direct result of the sphericity of the system or, in other words, of the $\beta$-effect. It is interesting to note that this adjustment is probably the reason for the location of the upper tropospheric easterly jet stream on the equatorward side of the summer Asian continent and hence to the south of the mon-

Fig. 9. As in Fig. 8 except for the SST anomaly located at $\mu_0 = 0.1$ (5.7°N).

1977; Webster, 1977) and probably in terms of the shear in the basic flow.

Figs. 6 and 11 describe the higher latitude stationary Rossby wave response. The relative magnitude of $v$ has increased and the response is no longer symmetric about the equator. In the vicinity of the heat source a strong anticyclonic flow has developed in the upper troposphere above a lower

Fig. 10. As in Fig. 8 except for the SST anomaly located at $\mu_0 = 0.2$ (11.5°N).
soon latent heat release (see Webster et al., 1977). The flow is also consistent with the enhanced westerlies in the upper troposphere of the low-latitude western and central Pacific Ocean although this is usually tied to the release of latent heat in the Indonesian region.

In Figs. 9 and 10 the response close to the equator possesses a zonal character indicating the excitation of an equatorial Kelvin wave even when the SST anomaly is removed away from the equator. Only when a critical latitude exists between the forcing and the equator does the equatorial region remain quiescent as shown in Figs. 11 and 12.

Finally, it should be noted that the form of the response in the lower troposphere for all \( \mu_0 \) is smaller and possesses less latitudinal structure than

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**Fig. 11.** As in Fig. 8 except for the SST anomaly located at \( \mu_0 = 0.4 \) (23.6\(^\circ\)N).

**Fig. 12.** As in Fig. 8 except for the SST anomaly located at \( \mu_0 = 0.6 \) (36.0\(^\circ\)N).
TABLE 1. Comparison of the local atmospheric response at iteration 11 to SST anomalies placed at various $\mu_0$. Also shown is the basic flow $\bar{U}$, and the converged equilibrium heating.

<table>
<thead>
<tr>
<th>$\mu_0$</th>
<th>$\bar{U}_1(\mu_0)$</th>
<th>$u_1(\mu_0)$</th>
<th>$\omega_1(\mu_0)$ $\times 10^6$</th>
<th>$Q_{C_0}(\mu_0)$</th>
<th>$\psi_{max}$</th>
</tr>
</thead>
<tbody>
<tr>
<td>0</td>
<td>-5.0</td>
<td>-2.4</td>
<td>-1.9</td>
<td>146</td>
<td>117</td>
</tr>
<tr>
<td>0.1</td>
<td>-4.5</td>
<td>-2.7</td>
<td>-1.8</td>
<td>138</td>
<td>92</td>
</tr>
<tr>
<td>0.2</td>
<td>2.5</td>
<td>-3.3</td>
<td>-1.7</td>
<td>131</td>
<td>55</td>
</tr>
<tr>
<td>0.3</td>
<td>13.0</td>
<td>-4.1</td>
<td>-1.6</td>
<td>125</td>
<td>47</td>
</tr>
<tr>
<td>0.4</td>
<td>19.1</td>
<td>-4.0</td>
<td>-1.3</td>
<td>108</td>
<td>30</td>
</tr>
<tr>
<td>0.5</td>
<td>22.5</td>
<td>-3.1</td>
<td>-1.0</td>
<td>80</td>
<td>19</td>
</tr>
<tr>
<td>0.6</td>
<td>25.4</td>
<td>-1.8</td>
<td>-0.5</td>
<td>50</td>
<td>10</td>
</tr>
<tr>
<td>0.7</td>
<td>25.0</td>
<td>-1.0</td>
<td>-0.2</td>
<td>50</td>
<td>4</td>
</tr>
</tbody>
</table>

* Maximum Southern Hemisphere geopotential.

the flow in the upper troposphere. Presumably, this arises from the distribution of dissipation which is largest in the lower troposphere as indicated in Eq. (8).

b. Limits of atmospheric response

The most striking experimental result is the diminishing magnitude of the total diabatic heating and the vertical velocity as the SST anomalies are placed progressively poleward (see Fig. 7). Associated with the decrease in $\omega$ is a tendency for the maximum vertical velocity to be removed substantially eastward or downstream of the region of maximum heating (see Fig. 7). The magnitude of the response in the Southern (unforced) Hemisphere also decreases simultaneously. These features are summarized in Table 1.

To understand the mechanisms producing the latitudinal diminution of the response, we resort to (3). Using (5)–(7) with the neglect of the curvature term in (6) we may write

$$\frac{\bar{U}}{a \cos \phi} \frac{\partial}{\partial \lambda} \left( \frac{\partial \psi}{\partial p} \right)$$

(A)

$$-\frac{v}{\cos \phi} \frac{p f}{R} \frac{\partial \bar{U}}{\partial p} + \omega S = -\frac{Q}{C_p}$$

(B) (W) (C)

where the terms have been labeled for convenience. Generally, at low latitudes where the Rossby number ($Ro = u/L$, where $u$ and $L$ are velocity and length scales) is large, the diabatic heating is almost balanced by the adiabatic response of the system. That is, $\omega$ may be determined quite accurately from the distribution of diabatic heating as (W) $\sim (C)$ and (A), (B) $\ll (C)$. In other words, the thermal advective effects are usually insignificant. However, as Table 1 indicates that the vertical response appears to decrease as $\bar{U}$ increases, or as

the latitude of the SST anomaly is increased, it is worthwhile to reexamine the system.

The ratio of the terms (A) and (C) (demarked A/C) and (B) and (C) (B/C) are shown in the upper panel of Fig. 13. In the calculations $\bar{U}$ was taken as the mean of $\bar{U}_1$ and $\bar{U}_2$. A longitudinal scale corresponding to $s = 1$ was assumed in the evaluation of $\partial / \partial \lambda (\partial \psi / \partial p)$. The solid curves show the ratios as functions of $\mu_0$ (i.e., anomaly location) when (C) was set equal to the initial amplitude of the forcing $50 \times 10^{-4} J kg^{-1} s^{-2}$). The dashed curves plot the same ratios but with (C) determined by the converged value of the diabatic heating. The lower panel shows the longitudinal phase lead of the vertical velocity response over the initial location of the SST anomaly (solid line) and over the final converged diabatic heating maximum (dashed line). The coincidence of the dashed and solid curves at high latitudes in both panels of Fig. 13 indicate that the initial location and magnitude of the total diabatic heating remains unchanged with increasing iteration in those regions.

Fig. 13 indicates that there are two basic limits of response to forcing by the SST anomaly. These are a diabatic limit and an advective limit. The purely
diabatic response is consistent with the low-latitude response which may be anticipated from conventional scale analysis for moderate Ro. Thus the diabatic limit occurs where the anomaly interacts with the atmosphere in regions of small \( f \) so that advective terms of (20) play little part in the determination of \( \omega \). The fields shown in Figs. 5, 8, 9 and 10 provide examples of this limit. The advective limit is characterized by near parity between (A), (B) and (C) so that the vertical velocity is a function of both diabatic and thermal advective processes. The advective terms are more significant in middle and higher latitudes as \( \tilde{U} \) and \( \psi \) are much larger than in the tropics. The increase of importance of the advective effects is consistent with the decrease of the ratio \( \psi/\psi' \) (\( \psi \) and \( \psi' \) are vertical and horizontal velocity scales) as the Rossby number decreases with increasing latitude.

It is now possible to understand why \( \psi \) decreases so rapidly in higher latitudes. For a given SST anomaly the vertical velocity will develop in the vicinity of the initial heating if the thermal advection terms are small. Subsequent vertical velocity heating feedbacks in which anomalously warm air is raised over the heat source and cold air is lowered downstream from the source will allow the local response to grow until an equilibrium value is approached. Such is the development of the response in the diabatic limit. However, if \( \tilde{U} \) is very large, cold air is rapidly advected across the SST anomaly. The initial upward motion which is tied to the anomaly is forced to vertically advect cold air. In the adjacent downstream region the advected warm air is forced to sink. The raising of cold air and the sinking of warm air diminish the perturbation kinetic energy of the system. Alternatively, as the advective terms are dominant, the SST anomaly is forced to heat relatively cool air which has been advected across it. As there is a finite response of the system, a positive correlation must exist between \( Q \) and \( T \) but it must be extremely weak. Thus, in contrast to the direct circulation developed in the diabatic limit, the response in the advective limit is relatively indirect.

c. Scale and magnitude of the local response

Given the mutual dependency of the total diabatic heating and the dynamic response of the system (see Section 3) it is not difficult to understand the nature of the converged total diabatic heating displayed in Fig. 4 and, consequently, the scale and magnitude of the local response. Fig. 13 provides the key as the dominating characteristic of the response is whether it falls closer to the diabatic limit or the advective limit. If the system falls near the diabatic limit, the nature of the principal mode excitation is of paramount importance. Nearer to the advective limit the nature of the principle mode is irrelevant.

At low latitudes, where \( \tilde{U} \) is small and the diabatic processes are dominant, the principal mode is the Kelvin wave which possesses the property of being closest to resonance at the largest scales (Webster, 1973b; Holton, 1973). Formally, this preference for the largest scale is manifested by an inverse dependency of the steady, forced Kelvin wave amplitude on the square of the longitudinal wavenumber [e.g., Eq. (8), Webster, 1973b]. Consequently, the form of the converged local response depends on competition between two diverse processes; the tendency of the diabatic heating to flow to smaller and smaller scales (Section 4) and the selective large-scale response of the principle mode. The result is a dynamic configuration which has a smaller scale than the initial anomaly and a spectral energy peak biased toward the long-wave end of the spectrum.

In high-latitude regions where the advective processes are dominant or where they approach parity with the diabatic processes, the heating and the dynamic response possess a negative feedback linkage. From Fig. 4 it can be seen that after an initial small growth (in an iterative sense) the heating anomaly reconverges on the initial (and externally maintained) heating of the SST anomaly. The reason for the lack of growth of the system response was outlined previously; the advective processes invoke an indirect longitudinal circulation and the kinetic energy of the perturbation response is reduced rather than amplified as in the low-latitude cases.

d. Forcing and scale of the remote response

One of the major results of the experiments was the identification of significant model response in locations well-removed laterally from the region of forcing. With the SST anomaly located in the low latitudes, the geopotential height response in the remote high latitudes was significantly larger than the local response. Besides responding in a near barotropic mode the model displayed distinctive large-scale mode selection at high latitudes. This may be seen in Fig. 14 which shows isopleths of the inverse magnitude of the response at different latitudes normalized relative to the \( s = 1 \) mode at that latitude. Plots are presented for the \( \mu_0 = 0 \) and \( \mu_0 = 0.4 \) cases. In both cases the diminution of the relative magnitude away from the source region is most evident. That is, the larger scale modes have managed to maintain their magnitude more successfully than the smaller modes.

That a distinct remote response at high latitudes should occur from low-latitude forcing is not particularly surprising. Correlations between the perturbation geopotential and the velocity fields indicate a poleward wave-energy flux away from the source region. Furthermore, in a steady system
those modes with a Doppler-shifted frequency which approaches zero are closest to resonance and in middle and high latitudes these are the largest modes (Webster, 1973a). This is because the Doppler-shifted frequency for stationary modes is merely proportional to $U_0$ (Webster, 1973a). Thus smaller modes (large $s$) are removed further and further from resonance.

Karoly (1980) and Hoskins and Kavoly (1981) have offered an explanation of the sign of the wave propagation (i.e., low to high latitudes) using ray tracing techniques. They found that low-latitude excitation on the equatorial side of the basic westerly maximum will result in the longest waves being ducted to high latitudes, whereas the smaller scale modes are rapidly reflected and constrained to the region of the source. Karoly’s arguments were based on Rossby mode propagation but even with the SST anomaly located at the equator it can be argued that modes in the westerlies are still excited either by wings of the forcing distribution actually being in the weak westerlies (cf. Figs. 3 and 5) or that the effect of dissipation on the equatorial Kelvin mode has allowed an energy propagation to higher latitudes as discussed earlier. Importantly the propagation patterns shown in Figs. 8–12 appear to follow great circle trajectories as indicated by Karoly and by Grose and Hoskins (1979) and Hoskins and Kavoly (1981).

The rapid decay of the modes toward and across the equator is caused by the existence of critical latitudes for stationary modes where $U_i = 0$. Except for the small seepage of energy through these latitudes due to the dissipative nature of the system (Webster, 1973a) the low latitudes and the unforced hemisphere are virtually unexcited by an SST anomaly which does not actually reside in the low latitudes or lies across the equator. However, this interpretation is, to a certain extent, dependent on the simple basic state we choose. In the next section we will discuss implications of the result to more realistic systems.

The quasi-barotropic nature of the high-latitude remote response is rather puzzling. Presumably, the basic state of the system provides further filtering between the barotropic part and the baroclinic part of the mode. Indeed, Mak (1969) showed that the resonant frequencies of the barotropic part of the mode are greater than for the baroclinic part. A steady-state interpretation of Mak’s results suggests that the barotropic part will penetrate further into the westerlies than the baroclinic part. However, Mak’s analytic analysis in which it was possible to separate out the barotropic and baroclinic parts of the flow was for a special system in which the vertical and horizontal shear of the basic flow was neglected. As the barotropic response is confined to the region poleward of the basic westerly maximum in all cases considered, it would appear that shear is an important aspect of the selective response.

3. Vertical velocity distributions relative to low latitude SST anomalies

The vastly different latitudinal structures of the geopotential and the vertical velocity fields produced by a low-latitude SST anomaly (see Section 6) arise from the transition from a “diabatic limit” balance in the vicinity of the anomaly to an “advective limit” at higher latitudes. Returning to (20), (W) balances (C) at low latitudes so that the vertical velocity field, the heating and the geopotential extrema are nearly in phase. Moving poleward the magnitude of (C), the diabatic heating, rapidly decreases and the vertical velocity is determined by the advective terms of (20). As a result the vertical velocity maxima is found increasing further downstream at higher latitudes.

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3 Webster (1973a) suggests that the existence of shear in the basic flow may be partly responsible for energy seepage through critical latitudes. With shear, the discrete two level system possesses different critical latitudes in each level.
The lower tropospheric geopotential fields and the vertical velocity distribution for the \( \mu_0 = 0 \) case are shown on Fig. 16. The solid circle indicates the location maximum diabatic heating and the open circle denotes the location of the geopotential minimum. It is important to note that the maximum heating lies significantly to the west of the geopotential minimum. The physical implications of the vertical velocity and lower tropospheric geopotential distributions will be discussed in the next section.

8. Summary and inferences

The primary aim of the study was to provide a simple platform from which the myriad of results generated by GCM experimentation could be interpreted and understood relative to the statistical studies of Bjerknes (1966), Namias (1976a) and Davis (1978).

a. Interpretation of GCM results

In many aspects, the results of the linear study agree quite well with those of the many GCM experiments. In particular both the GCMs and the linear model exhibit strong local and remote excitation to the placement of an SST anomaly at the equator. The linear model also fails to respond to middle- and high-latitude anomalies which has been explained in terms of the increasing importance of thermal advection in the vicinity of the anomaly. Consequently, the study tends to support the hypothesis that the GCMs were behaving correctly to both tropical and high-latitude forcing. Possibly on the space and time scale considered here midlatitude SST anomalies possess little influence on the evolving climatic state. However, it should be noted that Simpson and Downey (1975) found a substantial alteration to their model response when the polarity of the SST anomaly was reversed which could suggest a sensitivity to the state of the atmosphere where the anomaly is introduced. That is, one may expect (perhaps) an increased sensitivity to the anomaly if it were introduced into a region of (for example) ascending air. Furthermore it is possible that the nonlinearity of the GCM (or the atmosphere) may be able to override the initial dominance of the advective effects but probably only for the unphysical "super anomaly" magnitude of Chervin and Schneider (1976). Such an override is not possible with the present model because of the inherent linearity of the system.

b. The Namias-Davis correlations

It was pointed out earlier that the GCM experiments, while universally predicting a limited response to SST anomalies located at middle and high latitudes, do not reflect the statistical correlations of Namias (1976a) and Davis (1978) who indicated predictability of atmospheric response relative to summer SST anomaly patterns. It should be remembered that these two studies, which indicated a temporal restriction to the predictability, followed an earlier study (Davis, 1976) which questioned a general relationship between ocean and atmospheric midlatitude anomaly patterns.

It has been argued that the reason for the minimal local response was because the model was approaching an "advective limit" in regions of strong westerly flow. Thus if a GCM utilizes either a mean annual or a January insolation distribution, the experiments relate the response of the atmosphere to an SST anomaly cell within the advective limit regime. Thus the GCM experiments appear to confirm the Davis (1976) conclusion. However the basic westerly flow in the summer hemisphere decreases by almost a factor of two in midlatitudes (see Webster 1972, Fig. 7) with the maximum migrating northward to near 45°N compared to the mean annual position of 33° shown in Fig. 7. Thus in the region of \( \mu = 0.6 \) (i.e., 36°) the mean tropospheric wind decreases from 15 m s\(^{-1}\) in the mean annual case to ~6 m s\(^{-1}\) in the summer. Thus the role of advection and the consequent downstream migration of the vertical velocity maximum away from the heat source will be considerably reduced. As a consequence, the response will be considerably larger in summer because, from an
energetic viewpoint, the resulting circulation will be considerably more direct.

In summary, it would appear that the Namias-Davis correlations are physically consistent with the mechanism of dynamic response of the atmosphere developed in this study. A more detailed investigation of the role of seasonality in the atmospheric response to SST anomalies is currently underway.

c. Tropical SST anomalies, cloud bands and the Southern Oscillation

Perhaps the most persistent climatological feature observed from a satellite is the South Pacific cloud band (Streten, 1975). The orientation of the band is northwest-southeast emanating from low latitudes to the east of Indonesia which correspond to the warmest SST region in the Pacific Ocean. During El Niño periods when the SST anomaly is removed significantly to the east, the cloud band appears to move in an eastward direction (Streten, 1975).

Fig. 16 displayed the difference in latitudinal structure of the vertical velocity field and the geopotential relative to a tropical SST anomaly. Quite possibly the cloud band is a manifestation of the vertical velocity distribution. Importantly, the scale and orientation of the band appears to match that of the vertical velocity distribution. Furthermore the sympatetic migration of the band with the SST anomaly observed by Streten (1975) possibly strengthens the relationship.

It is also interesting to note that the lower tropospheric geopotential field distribution is consistent with the anomalous mean sea level pressure pattern associated with the Southern Oscillation (Troup, 1965; Trenberth, 1976). The mechanisms which may relate the Southern Oscillation and the El Niño are currently being investigated.

It is important that the results be viewed relative to the simplicity of the model. For example, the model showed little interhemispheric wave transmission when the forcing was located at middle or high latitudes principally because of the existence of the $\bar{U} = 0$ critical latitudes near the equator. However, in reality the seasonal time averaged flow shows considerable longitudinal variation with mean westerlies at the equator as well as easterlies! That is, true longitudinally independent $\bar{U} = 0$ "critical latitudes" may not exist (Webster et al., 1977; Webster, 1978). The mean upper tropospheric equatorial westerlies in the real system are functions of the steady forcing in the tropical regions and, to complicate matters, functions of the SST anomaly locations themselves. Thus the present model may simplify and underestimate the cross-equatorial teleconnection by not containing longitudinal variations in basic zonal flow. The subject of cross-equatorial wave propagations on the time-scales considered in this paper and the influence of a nonzonal symmetric basic state on the atmospheric response to SST anomalies are subjects of another study.

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