

## Seasonality in the Local and Remote Atmospheric Response to Sea Surface Temperature Anomalies

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### ABSTRACT

Two apparently contradictory situations are provided by observations of the atmospheric response to sea surface temperature anomalies. These are: (i) The extratropical regions of the *winter* hemisphere appear to possess strong teleconnections with equatorial forcing but weak or non-existent connections with local (extratropical) heating anomalies. (ii) The extratropical regions of the *summer* hemisphere are quite sensitive to local thermal forcing but apparently unaffected by the remote forcing from equatorial regions. An attempt is made to provide a consistent physical picture which simultaneously embraces these two situations.

With the aid of a simple linear model it is shown that the summer hemisphere is more sensitive to local forcing than the winter hemisphere because it is closer to the *diabatic limit* of Webster (1981), thus allowing an efficient energy generation. The winter hemisphere is much closer to the *advective limit*. The sensitivity of the midlatitudes to remote (equatorial) forcing is shown to be a function of the relative location of the SSTA (sea surface temperature anomaly) to the zeros of the basic flow and the magnitude of the midlatitude westerlies. A hemisphere will become excited by remote forcing if at least part of the low-latitude sea surface temperature anomaly is located in the *weak* subtropical westerlies. Given that the latter criterion is met it is shown that the amplitude of the response and the latitude to which a particular mode is transmitted depends upon the distribution of westerly winds. The specific situation of El Niño sea surface temperature forcing is considered relative to realistic seasonal mean zonal wind fields. The model response is compared with the gross features of the observed anomalous atmosphere during El Niño years and a correspondence found.

Finally, it is argued that the explanation of seasonality in atmospheric response offered in this paper will allow seasonal climate forecasting to be approached with an *a priori* physical expectation.

### 1. Introduction

The study of the effect of sea surface temperature anomalies (SSTA) on the weather and climate of the atmosphere has received considerable attention during the last few years. The popularity of the subject is not surprising as there were significant indications in many early studies that real links existed between the anomalous state of the ocean and an anomalous response by the atmosphere, conveniently lagged to allow inferences regarding the future climate state of the atmosphere (e.g., Bjerknes, 1966; Namias 1973, 1976a; etc.).

The early observational studies provided somewhat of an ambiguous picture of the interaction of the ocean and the atmosphere. Namias (1973) and Davis (1976) reached contrary conclusions regarding the generality of relationships between sea surface temperature anomalies and the anomalous atmospheric response. Investigations using large-scale general circulation models did little to clarify the contradictory observational evidence. Although it is difficult to classify the results of many experiments in a simple manner, perhaps two main areas of results emerged. Models either did not show a significant response

when an SSTA was placed in the high and middle latitudes (e.g., Chervin and Schneider, 1976; Kutzbach *et al.*, 1977) or they managed to achieve an appreciable response when the anomaly was placed near the equator. Thus, at the same time, the general circulation models were substantiating the Bjerknes hypothesis but finding little support for the higher latitude Namias correlations. Either the general circulation models were behaving inappropriately for middle latitude SSTA's or the observational studies were providing faulty hypotheses to be tested by the models.

A solution to the paradox was sought by Webster (1981) who used a linear model which possessed a scheme which allowed self-determination of the total diabatic heating field relative to a prescribed SSTA. He was able to show that the general circulation models were probably behaving in a manner consistent with the real atmosphere. Anomalies inserted into the strong high-latitude equinoctial<sup>1</sup> basic flow were shown to effect only a small local response be-

<sup>1</sup> In Webster (1981), the "equinoctial" field was referred to as the "annual" field.

cause of the relative importance of advective effects. The weak response results from the inability of the system to generate kinetic energy from the potential energy produced by the anomaly. The advective effects were sufficiently strong to allow the phases of the perturbation temperature field and the vertical velocity field to approach quadrature. On the other hand, the atmospheric response to a low-latitude SSTA produces a near-optimal energy conversion; the relatively weak zonal winds allow the heating anomaly to remain in phase with the vertical motion response. These two limits of response were referred to by Webster (1981) as the *advective* and *diabatic* limits, respectively. Similar conclusions can be inferred from Opsteegh and Van den Dool (1980), and Hoskins and Karoly (1981), although the latter study presents arguments to show that the magnitudes of the meridional and vertical advection as well as the zonal advection also should be considered. That is, other aspects of the basic flow, besides its magnitude, may be of importance. It should be pointed out that Egger (1977) was possibly the first to indicate the necessity of westerly flow for a strong remote response.

The two limits were most likely simulated by the general circulation models in an adequate manner. Indeed, if the models had failed to emulate this dynamic character of the atmosphere, which depends on the most basic of balances, it would seem that there would be little hope for them to simulate almost any other climatic relationship. The differences between the observations and the model results for the middle- and high-latitude anomalies emanates, not from deficiencies in the model, but from the use of an inappropriate basic state or, at worst, an inability to interpret correctly the model results. The results obtained by Opsteegh and Van den Dool (1980), Webster (1981) and Hoskins and Karoly (1981) indicate the usefulness of simple models in parallel with studies using large-scale general circulation models.

The observational studies of Namias (1976b) and Davis (1978) managed to reconcile some of the earlier differences by recognizing hints of seasonality in the SSTA-surface pressure anomaly correlations. Specifically they suggested that significant correlations existed between *summer* middle- and high-latitude SSTA's and subsequent local (i.e., middle- and high-latitude) atmospheric indices. Little or no correlation appeared between the winter SSTA's and an atmospheric response either local or remote. As most of the general circulation models mentioned above were using annual mean or winter insolation values which would have resulted in seasonal mean states similar to the equinoctial or winter Northern Hemisphere, it is not surprising that the experiments were producing a minimal response to surface forcing.

In addition to the distinct seasonality of the local

response to midlatitude thermal forcing, there appears to be a second form of seasonal response sensitivity. Bjerknes (1966) and Namias (1973, 1976a) have shown that the maximum higher latitude response to equatorial SSTA (specifically, the remote response to central and eastern Pacific Ocean warm anomalies) occurred in the winter hemisphere. These observations have been corroborated by Horel and Wallace (1981).

In summary, the observational data provides two seemingly paradoxical situations. These are (i) the extratropical regions of the *winter* hemisphere appear to possess strong teleconnections with equatorial forcing but possess weak or non-existent connections with local (extratropical) boundary heating anomalies, and (ii) the extratropical regions of the *summer* hemisphere are significantly sensitive to local thermal forcing but apparently less affected by the remote forcing from tropical latitudes. In the following we will attempt to establish a consistent physical picture which embraces both (i) and (ii). In this manner we seek to extend the study of Webster (1981) in an attempt to comprehend the effects of seasonal structures in the basic flow.

## 2. Method

The method by which we seek the seasonal atmospheric response to anomalous surface heating follows that of Webster (1981). The model used is the simplest baroclinic prototype of the atmosphere which retains the ability to include an arbitrary baroclinic basic state and thermal forcing. In essence, the model is a linear, spherical, baroclinic (two-layer) primitive equation model which is solved for the steady-state response of a given basic state to a given initial forcing function distribution.

A major philosophical difference exists between this study (and that of Webster, 1981) and those of Opsteegh and Van den Dool (1980) and Hoskins and Karoly (1981). This is the incorporation of a scheme which allows the model to determine its own diabatic heating relative to a prescribed (initial) SSTA. It was argued by Webster (1981) that such a scheme is necessary because, ultimately, the major component of the diabatic heating resulting from the imposition of a SSTA will be the associated latent heat release rather than the direct sensible heat input by the SSTA. *As latent heating is a strong function of the dynamics of the system, the resultant heating need not necessarily exist on the same scale as the SSTA.* Indeed, there appears to be much observational evidence for such differences in scale. In the tropical Pacific Ocean, for example, the regions of large-scale precipitation (and presumably the region of ascending motion and latent heating) is somewhat smaller than the area of positive SSTA. The dynamics-diabatic heating feedback is described fully by Webster (1981); because the feedback is inherently

nonlinear it requires an iterative technique for solution.

However, there are limitations in the manner in which the scheme is used by Webster (1981) and the present study. The feedback on the heating depends strictly on the dynamic model excited in the system so that the associated vertical velocity distribution basically determines the diabatic heating field at the next iteration. However, we have used a two-layer system with a perturbation vertical velocity field (and the diabatic heating) prescribed only at the mid-level of the model! One may argue that such an assumption is valid for the tropical atmosphere but it must be considered as a severe restriction for the extratropical atmosphere, where, we may imply from observations, the response to boundary forcing is quite variable. Obviously, for a more general study, a model with more degrees of freedom in the vertical should be employed.

The restriction can be interpreted as an inflexibility of the model to the testing of the response sensitivity of the extratropical atmosphere to various vertical structures of diabatic heating, as considered, for example, by Hoskins and Karoly (1981). But such questions fall beyond the scope of this study. Here we seek merely to explain the relatively greater magnitude of the local response in the tropics and in the summer extratropics compared to the extratropics in winter. As the two-layer limitation placed on the generality of the heating scheme must surely result in an overestimation of the local extratropical heating, it should be considered a "worst-case" assumption. In addition, it would seem that it is better to apply the same heating scheme to both low and high latitudes, as we have done, than to *prescribe* total diabatic heating fields which, in a sense, pre-supposes a dynamic-diabatic heating feedback.

### 3. Basic states and forcing functions

The basic state used by Webster (1981) was an analytic function similar to the mean equinoctial observed structure. The 250 and 750 mb zonal wind fields, which correspond in locations to the two model layers, are shown in the upper panel of Fig. 1. In addition, the global distribution of the observed zonally averaged fields for the Northern Hemisphere summer (i.e., the mean winds for June, July and August; hereafter referred to as the JJA case) and for the Northern Hemisphere winter (the DJF case) and also shown in Fig. 1. Magnitudes in the winter hemispheres are much larger than those located at similar latitudes in the summer hemispheres. The mean tropospheric zonal winds are shown as dashed lines. The principal comparisons will be made between the remote and local atmospheric responses in the summer and winter hemispheres, particularly in JJA and during the equinox.

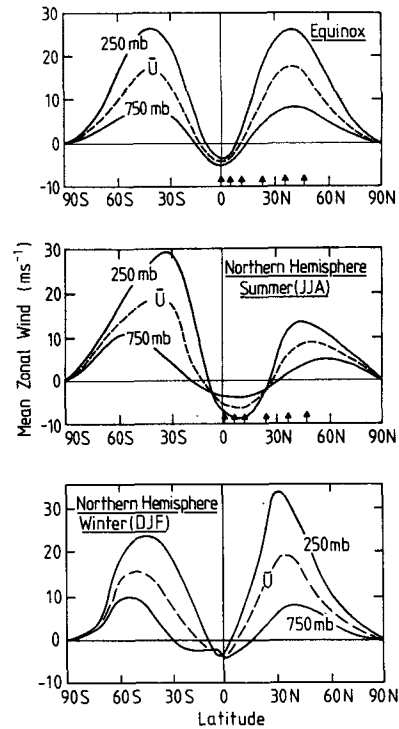


FIG. 1. Latitudinal distribution of the zonally symmetric basic wind field (units:  $m s^{-1}$ ) at 250 and 750 mb levels for the equinox, JJA and DJF cases. Dashed line represents the mean troposphere wind fields and the arrows the locations of the sea surface temperature anomalies.

The experimental format follows Webster (1981). SSTA's of magnitude  $2.5^{\circ}C$  were placed at  $0, 5.7, 11.5, 23.6, 36.9$  and  $44.4^{\circ}N$ . In terms of sine of latitude ( $\mu$ ) the anomalies were located at  $0, 0.1, 0.2, 0.4, 0.6$  and  $0.7$ . Except where stipulated the SSTA was assumed to have a longitudinal wavenumber 1 scale and an  $e$ -folding latitude scale of  $7^{\circ}$  at the equator. The analytic form of the forcing function provides the same heat input to the atmosphere irrespective of the latitude of its location. The iterative scheme of Webster (1981) was used until convergence occurred, usually after a few iterations. All results shown refer to iteration number 11.

### 4. Results

In order to assess quantitatively the seasonality of the atmospheric response to SSTA's, the various experiments will now be detailed. Experiments were performed using the JJA basic state the results of which were compared with parallel experiments using equinoctial conditions. The physical processes which determine the seasonally varying local and remote effects uncovered in this section are discussed in detail in Sections 5 and 6. Details of the equinoctial season case are given by Webster (1981).

Fig. 2 illustrates the variation of the converged

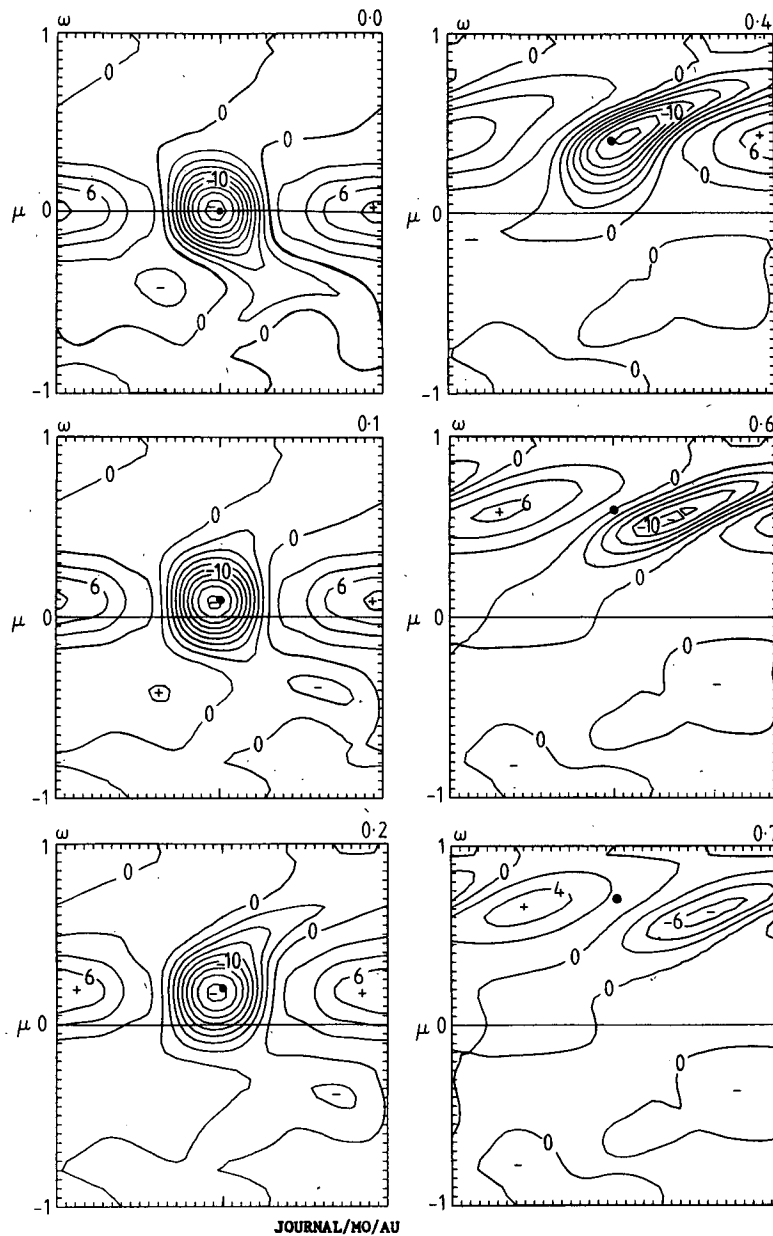


FIG. 2. Vertical velocity distributions relative to the location of the initial sea surface temperature anomaly indicated by the solid circle (units:  $10^{-3}$   $\text{N m}^{-2} \text{s}^{-1}$ ).

vertical velocity response for the six cases described above for the JJA case. In each diagram the center of the initial wavenumber 1 SSTA is marked as a solid circle. With the SSTA located at the equator the response is large and constrained to the equatorial region except for a weaker remote response into the Southern Hemisphere subtropics. The vertical velocity response at high latitudes is very weak in both hemispheres. As the anomaly is moved poleward the local vertical velocity response remains relatively strong and only when the anomaly is located

at high latitudes (e.g.,  $\mu_0 = 0.6$  or  $0.7$ ) does the vertical velocity weaken considerably. This is accompanied by a considerable phase shift of the vertical velocity field relative to the initial anomaly location.

A comparison between the vertical velocity response for the JJA and equinoctial cases is presented in Fig. 3. The figure plots the maximum vertical velocity at a particular latitude as a function of SSTA location ( $\mu_0$ ). The diagonal dashed line denotes the location of the anomalous surface heating. Comparing the *local* responses we find that the vertical ve-

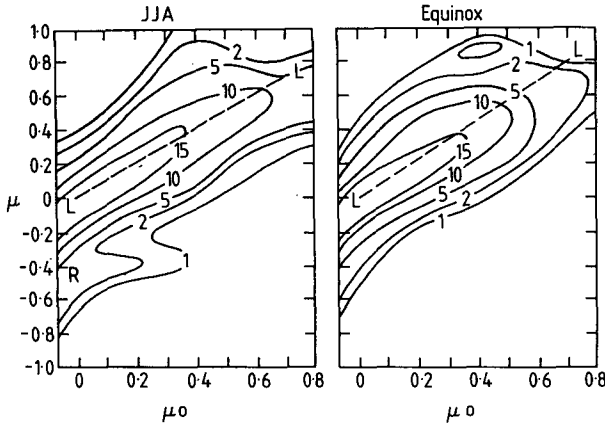


FIG. 3. Vertical velocity response extrema (units  $10^{-3} \text{ N m}^{-2} \text{ s}^{-1}$ ) plotted as a function of  $\mu$  (sine of latitude) for various Northern Hemisphere anomaly locations. Diagonal dashed line denotes the anomaly position.

locity is much stronger in the JJA Northern Hemisphere than during the equinox. Neither season shows a marked *remote* response which probably is a result of the barotropicity of the remote response rather than real similarities between the seasonal response. The equivalent barotropic nature was a marked property of the remote response observed in the studies of Webster (1981) and Hoskins and Karoly (1981).

The nature of the seasonal differences in the vertical velocity response is underlined by a comparison of the magnitudes of the converged heating fields in Fig. 4. At low latitudes the heating field is slightly greater during the equinox than during JJA. Both seasons show a general reduction of diabatic heating toward the pole, reducing at very high latitudes to the initial magnitude of SSTA heat input referred to as the "base" heating. However, at moderately high latitudes the JJA diabatic heating is substantially larger; in fact by over 50% for the  $\mu_0 = 0.6$  case.

The most important feature of the atmospheric response is the comparative location of the heating anomaly and the vertical velocity maximum as is their relative phase which will determine the magnitude of the local response (Webster, 1981). Specifically, the phase difference is an indication of whether the response will be closer to the "diabatic limit" or the "advective limit". For equinoctial conditions the longitudinal phase differences increased rapidly as the SSTA was moved progressively poleward. The  $\omega - \dot{Q}$  phase relationships are shown for both seasons in Fig. 5. In the Northern Hemisphere during JJA  $\omega$  and  $\dot{Q}$  remain much more in-phase when the anomaly is located at high latitudes than during the equinox.

Examples of the geopotential and velocity response for the JJA case are shown in Fig. 6. Velocity scales

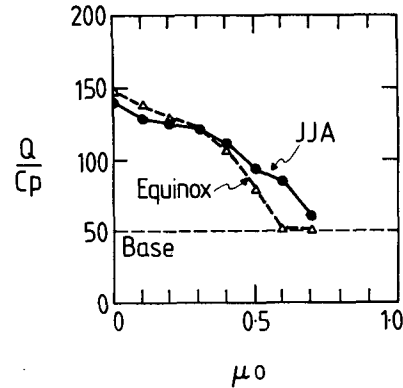


FIG. 4. A comparison between the converged total heating (units:  $10^{-4} \text{ J s}^{-1} \text{ kg}^{-1}$ ) for the JJA (solid line) and equinox (dashed line) cases as a function of latitude. Horizontal dashed line denotes the base maintained anomaly heating.

are indicated on each diagram. The fields may be summarized as follows. With the anomaly at low latitudes a large *local* velocity response occurs in a manner similar to the equinoctial case. At low latitudes the geopotential response is small while the magnitude of the velocity field is large; a relationship which we may have anticipated from traditional scaling arguments. However, the *remote* geopotential response is only large in the winter (southern) hemisphere and nearly absent in the summer (northern) hemisphere. As  $\mu_0$  increases the local response in the Northern Hemisphere grows in intensity but the re-

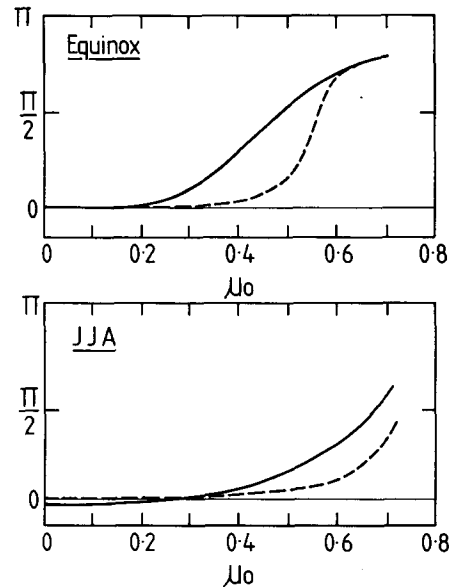


FIG. 5. A comparison between the equinox and JJA cases of the phase lead of the vertical velocity over the heating anomaly. The solid line indicates the phase lead relative to the initial sea surface temperature anomaly and the dashed line shows the phase lead relative to the converged total diabatic heating.

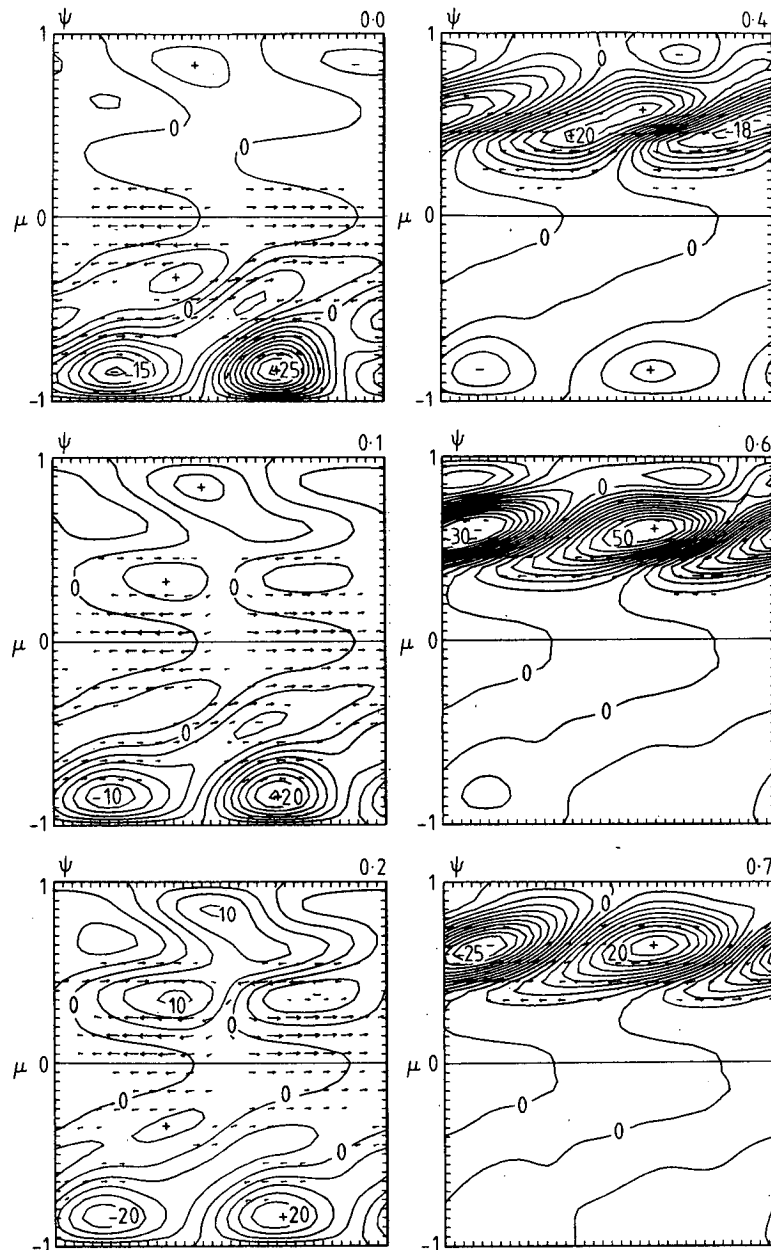


FIG. 6. The fields of perturbation velocity and geopotential response at the 250 level for the JJA case for the SSTA located at  $\mu_0 = 0, 0.2, 0.4$  and  $0.6$ . Vectors less than  $0.5 \text{ m s}^{-1}$  are omitted.

mote (winter hemisphere) response gradually declines.

A comparison between the JJA and equinoctial geopotential responses is made in Fig. 7. It is most evident that the local geopotential response for the middle- and high-latitude locations of the anomaly are considerably stronger in JJA. Furthermore, the *remote response* in the Northern Hemisphere to equatorial forcing is weaker than in the equinoctial case. However, the magnitude of the remote response in the Southern Hemisphere for equatorial anomalies

is substantially larger than the forcing during the equinox. In that sense the model is showing similar characteristics to the observed response structures which we noted in the Introduction.

## 5. Interpretations

### a. Seasonality in local response

An explanation of the seasonality of the local response noted in the last section follows directly from

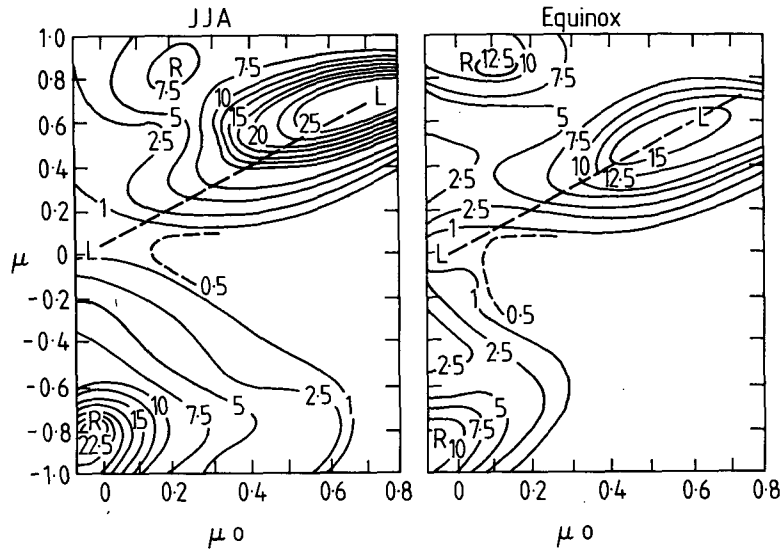


FIG. 7. Geopotential response extrema (units: m) for the JJA and equinox cases plotted as functions of  $\mu$  for various anomaly locations. Diagonal line denotes the anomaly location. R denotes region of strong remote response.

interpretations of the conclusions of Webster (1981). As summarized in the Introduction, Webster emphasized the importance of the magnitude of the basic wind in determining the magnitude of the response. Crudely, the relationship is inverse. With the SSTA placed in the strong midlatitude westerlies, as in the equinox case, the local response is weak because an advective limit is approached. Consequently, when the basic zonal winds lose intensity as the hemisphere approaches summer, the effect of the SSTA is magnified as the atmospheric response changes from a strong basic-state advective response to a weak basic-state diabatic response.

A schematic of the local response of the atmosphere to the imposition of a SSTA is shown in Fig. 8. Examples of the relative phase of the variables for the low-latitudes (either season), the midlatitudes during summer (weak zonal winds) and the midlatitudes during winter (strong zonal winds) are presented. The top and bottom diagrams illustrate the diabatic and advective limits of response discussed by Webster (1981). Such configurations correspond to a strong direct circulation in the diabatic limit ( $\omega$ ,  $T$  and  $\dot{Q}$  in phase) and to a very weak or marginally direct circulation in the advective limit ( $\omega$ ,  $T$  and  $\dot{Q}$  nearly out of phase). In the former case, the configuration allows the development of the secondary heating (via diabatic heating-dynamics feedback) in phase with the perturbation temperature. In the latter case, the secondary heating cannot develop as the quantities are advected out of phase by the strong zonal winds. During summer (middle diagram), with relatively weak zonal westerlies, the

phase relationships allow the development of a moderately strong direct circulation and a correspondingly strong local response.

The variation between an advective response in winter to a near-diabatic response in summer pro-

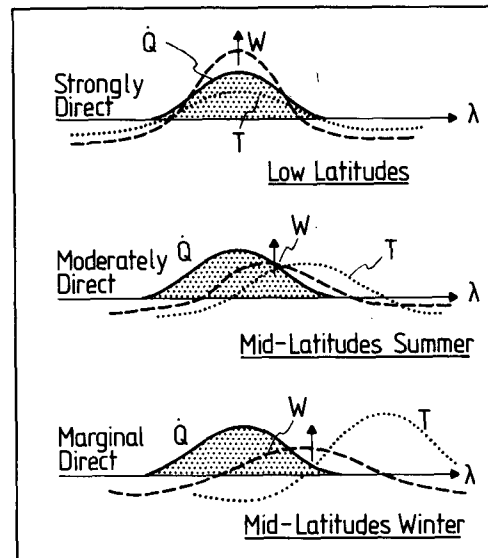


FIG. 8. A schematic representation of the relative locations of the initial SSTA (marked  $\dot{Q}$ ) and the perturbation vertical velocity field ( $W$ ) and the temperature field ( $T$ ) plotted as functions of longitude. The relative locations determine the vigor of the generation of the eddy available potential energy and its conversion to kinetic energy and, consequently, the magnitude of the local response of the atmosphere. Typical low-latitude, midlatitude summer and midlatitude winter examples are shown.

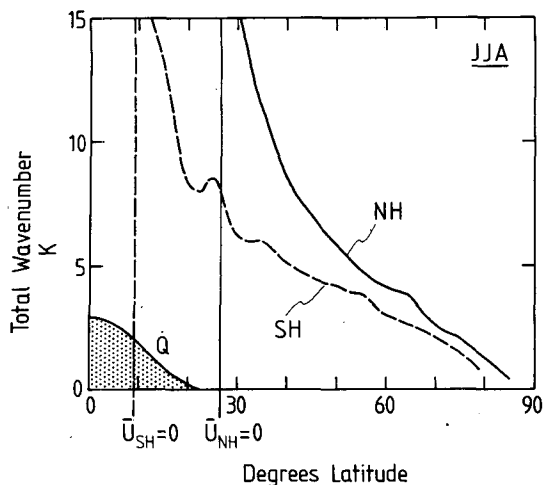


FIG. 9. The distribution of total wavenumber  $K$  of Hoskins and Karoly (1981) which defines turning latitude of a poleward propagating model plotted as a function of latitude for the Northern and Southern Hemisphere of the JJA case. Vertical lines denote the locations of the mean tropospheric basic wind zeroes.

vides some clues to the Namias-Davis correlations. In effect, an anomaly located in the midlatitudes during summer will perturb the atmospheric structure to a greater degree than in winter because the dynamic feedback will keep the initial anomaly and response reasonably in phase.

#### b. Seasonality in remote response

The amplitude of the *remote* response appears to depend upon a number of factors. The most important of these are 1) the location of the heating source relative to the zeros of the basic zonal wind, and 2) the magnitude of the basic zonal wind.

The location of the anomaly relative to the basic state zero is critical as it determines whether or not the mode response is oscillatory or exponential in latitude. An anomaly embedded completely within the equatorial easterlies may only excite equatorial Kelvin modes which possess a vanishingly small meridional velocity component (Webster, 1972). In the inviscid limit the latitudinal structure of the modes is purely exponential so that the frequency of the Kelvin wave and, consequently, its group velocity vector, is independent of latitude. Due to its non-dispersive nature viewed from a stationary perspective, the group velocity is independent of longitudinal aspect, and energy may radiate away from the source only in the vertical direction. In this way, the maximum response remains local to the region of forcing. In a viscous system the Kelvin wave phase surfaces become functions of latitude (Chang, 1977) but are still strongly damped so that even then the mode remains confined to the equatorial regions. In summary, an anomaly located within the easterlies can only force a local equatorial response.

If the anomaly resides either partially or wholly within the weak subtropical westerlies, the form of the forced modes will change substantially. Instead of the latitudinally trapped Kelvin mode, the anomaly may force Rossby modes which possess three-dimensional phase surfaces. Thus not only may energy be radiated vertically (depending on the vertical structure of the basic state) but also longitudinally (the modes are dispersive) and, most importantly for present considerations, latitudinally. Thus assuming the westerlies are sufficiently *weak* to allow a significant *local* response (see Section 5), the modes may provide a substantial *remote* response at higher latitudes as well. Propagation toward lower latitudes is limited, of course, by the existence of a  $\bar{U} = 0$  critical line for stationary modes.

In order to understand the strong scale dependence of the propagating modes into the winter hemisphere we adopt Karoly's (1978) and Hoskins and Karoly's (1981) total (or stationary) wavenumber ( $K$ ) which is defined as the ratio of the meridional gradient of the absolute vorticity and the angular velocity; i.e.,  $K = (B_m/\bar{U})^{1/2}$ , which sets the turning latitude of poleward propagating Rossby modes.

Fig. 9 summarizes the JJA  $K$  distribution due to an equatorial forcing with a distribution shown as the stippled region about the equator. In the Southern Hemisphere the heating resides both in the equatorial easterlies and the subtropical westerlies (cf. Fig. 1). However in the Northern Hemisphere the anomaly is enclosed by the easterlies and being equatorward of the critical lines only exponential solutions in latitude may result. Thus it is only in the winter Southern Hemisphere that the propagating Rossby modes may evolve. Furthermore, Fig. 9 shows that modes of longitudinal wavenumber 5 emanating from near the critical lines may propagate to about  $42^\circ\text{S}$  and  $55^\circ\text{N}$ , whereas the gravest longitudinal modes may propagate to  $76^\circ\text{S}$  and  $85^\circ\text{N}$ .

The effect of the strength of the basic state on the magnitude of the response enters in the following manner. Hoskins and Karoly have shown via WKB methods that the streamfunction amplitude of the remote response goes as  $(K^2 - k^2)^{-1/4}$ , where  $k$  is the longitudinal wavenumber. That is, for small  $k$  the remote response goes as  $(\bar{U}_m/B_m)^{1/4}$  which emphasizes the magnitude of the zonal flow. As a consequence, the amplitude of the winter hemisphere high-latitude remote response to low-latitude forcing should be considerably larger than either the equinox or summer hemisphere high-latitude response. Furthermore, we may infer that the gravest longitudinal modes will possess maximum amplitudes at high latitudes in the winter hemisphere. It should be remembered, though, that a high-latitude response to equatorial forcing will only occur if the weak subtropical westerlies are directly perturbed by the forcing.

Fig. 10 shows the distribution of the spectral power



as a function of longitudinal wavenumber of the geopotential response for the  $\mu_0 = 0$  (equator) case but with equal forcing at all wavenumbers. Maximum amplitudes are denoted by dashed lines. The results appear to follow the anticipations of the linear theory quite well. The WKB theory predicts the turning latitudes for the  $k = 1$  mode to be near  $75^\circ\text{S}$  (see Fig. 9) with the maximum amplitude at about  $60^\circ\text{S}$ . Strong correspondence is given by the numerical model though there are resolution problems at high latitudes; the model is stepped in increments of  $\mu$  (i.e.,  $\sin\phi$ ) rather than  $\phi$ .

**6. Seasonal response to El Niño forcing**

In the preceding sections we have discussed the processes which produce the seasonal differences in the atmospheric response to thermal forcing. We will now consider a specific situation—the seasonal response of the atmosphere to El Niño type ocean warming in the central and eastern Pacific Oceans. We choose as a guide the 1972–73 El Niño which is perhaps the best-documented event of its type. The anomalous surface temperature distribution was taken from Wooster and Guillen (1974) and the general description of the meteorology from Ramage (1975). The aim of the exercise is to examine the seasonal variation in the local and remote response to the El Niño event.

The response of the JJA and DJF basic wind fields described in Fig. 1 were considered. A warm anomaly of initial maximum magnitude of  $3^\circ\text{C}$  was placed to the east of  $180^\circ$  longitude with a longitudinal scale approximating that of an isolated wavenumber 3. The same latitudinal scale as in the other experiments ( $e$ -folding in  $7^\circ$  of latitude) was assumed. The location and areal extent of the initial anomaly is indicated by stippling in Figs. 11–14.

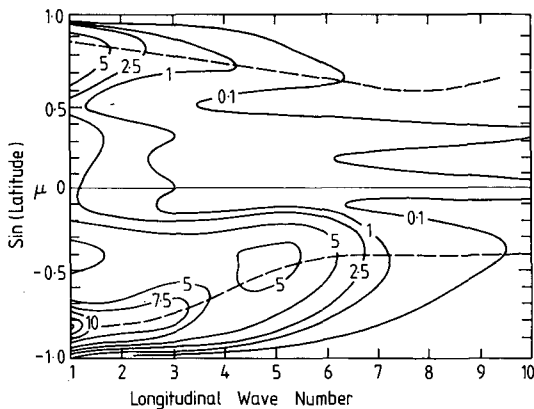


FIG. 10. The latitudinal distribution of the geopotential spectral power of the northern summer basic state as a function of longitudinal wavenumber. Forcing from an anomaly located at  $\mu_0 = 0$  but with a white noise spectral distribution.

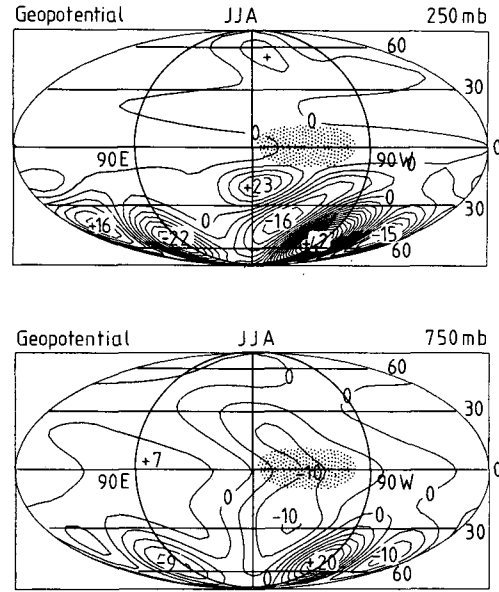


FIG. 11. The 250 mb (upper panel) and 750 mb (lower panel) geopotential (units: m) response of the JJA basic state to forcing by an El Niño-type thermal anomaly indicated by stippling. A Mollweide projection is used.

*a. The geopotential response*

The geopotential fields are presented to illustrate the remote response of the equatorial forcing. Although used often in discussions of model results the fields are poor indicators of local low-latitude effects. In any event the geopotential fields for 250 and 750 mb for the JJA and DJF cases are shown in Figs. 11 and 12.

As we should have anticipated from the preceding section it is the winter hemispheres of both cases which are most excited by the equatorial forcing; the summer hemispheres are only moderately affected and these perturbations are mainly confined to the subtropics. Furthermore, the maximum response in geopotential is distinctly downstream from the source region, as well as far poleward; features which may have been anticipated from Section 5(b). In the Northern Hemisphere the maximum high-latitude perturbation will occur over the Aleutians and Canada. The largest response in the Southern Hemisphere occurs over the southeast Pacific Ocean with secondary centers in the southern Atlantic and Indian Oceans.

The anomalous geopotential fields appear in agreement in a gross sense with those observational fields which are available. For example, the locations of the remote Northern Hemisphere winter geopotential maxima appear consistent with the results of Namias (1976b) and Horel and Wallace (1981). Comparisons are more difficult for the Southern Hemisphere winter due to the lack of data in the

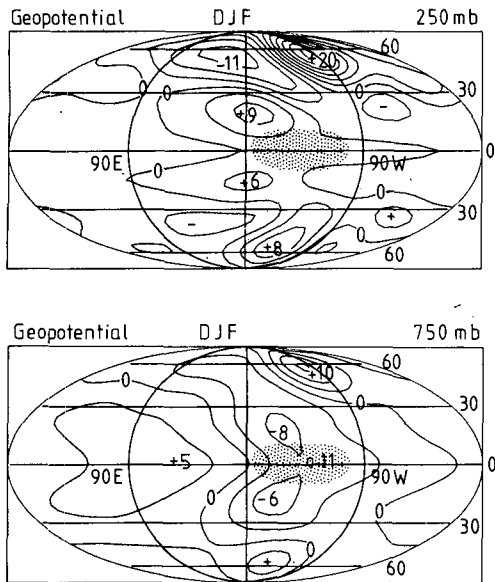


FIG. 12. As in Fig. 11 except for the geopotential response of the DJF basic state to El Niño-type forcing.

southern Pacific Ocean. However, some consistency with observation may be seen in the lower tropospheric geopotential fields in the equatorial regions and the Southern Hemisphere subtropics. The relatively higher pressures over the Australasian region and lower pressures over the eastern South Pacific Ocean during JJA and the relative eastward movement of the lower pressures during DJF appear consistent with the Southern Oscillation variation during El Niño events.

#### b. The streamfunction response

The streamfunction fields allow a greater insight into the local response of the low latitudes as well as the remote response at high latitudes. Furthermore, it is one of the fields presented by Ramage (1975) to show the circulation differences occurring during the 1972–73 event compared to other years. Specifically, Ramage calculated the anomalous streamfunction fields for the 1972 summer (during the El Niño year) and the 1971 summer (non-El Niño) relative to a 4-year mean field. At low latitudes Ramage found changes in the sign of the streamfunction field over the western and central Pacific Oceans and over the Atlantic. Fig. 13, which shows the streamfunction for the JJA case, indicates anomaly patterns which were very similar to those of Ramage. In particular, near the equator in the central Pacific Ocean, the flow is consistent with a Walker Circulation with its ascending branch moved considerably eastward. Both the observations and theoretical results show a wave propagation train away to higher latitudes in the manner described in Section 5b.

## 7. Concluding remarks

The study utilized a simple linear numerical model which sought the steady-state response of the zonally symmetric basic state of the atmosphere during the Northern Hemisphere summer to the influence of surface heating anomalies located at various latitudes. The model response was interpreted in terms of simple theory developed by Hoskins and Karoly (1981) and Webster (1981). The following conclusions were reached:

(i) The influence of *local* thermal anomalies in the *winter* hemisphere is small because the form of the response approaches the “advective limit” (Webster, 1981). That is, the magnitude of the winter hemisphere extratropical basic zonal winds are sufficiently strong so that the energetics necessary to create a significant local response (i.e., the generation of eddy available potential energy and the conversion of the potential energy to kinetic form) are not allowed to develop.

(ii) The influence of *local* thermal anomalies in the *summer* hemisphere [or the near-equatorial subtropics of the winter hemisphere; see (iii) below] is moderately large because the advective terms are of less importance with the smaller basic zonal winds. Thus the response is closer to the “diabatic limit” (Webster, 1981) which allows the vertical velocity, the heat source and the temperature field to remain more in phase than in the winter hemisphere which, in turn, provides conditions which allow a more ef-

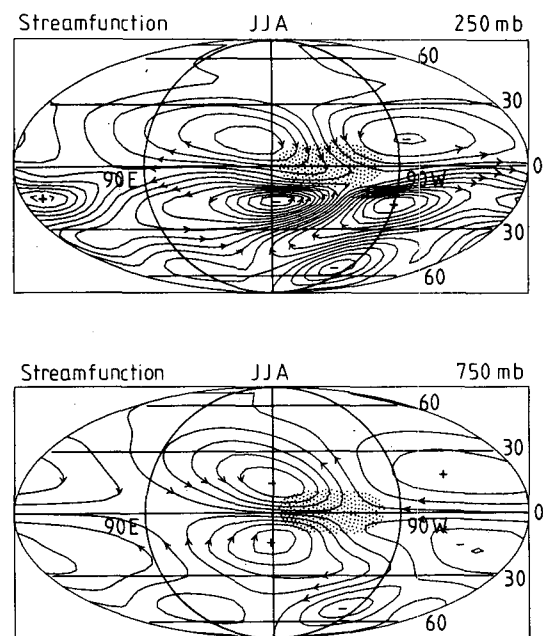


FIG. 13. As in Fig. 11 except for the streamfunction response (contour intervals:  $20 \times 10^5 \text{ m}^2 \text{ s}^{-1}$ ) of the JJA basic state to El Niño-type forcing.

ficient energy conversion and a substantial local response.

(iii) Strong teleconnections will exist between the equatorial regions and the *winter* high latitudes simply because the westerlies, closer to the equator in the winter hemisphere, may lie over part of the anomaly. If this is the case modes with meridional components to the group velocity vector will be excited which will allow a substantial response at higher latitudes. The effect is constrained by (i) and (ii) above and depends on the strength of the westerlies in the subtropical regions being sufficiently small to allow a sizable local response to provide a large enough secondary energy source for propagation.

(iv) The weak or non-existent teleconnections between the equatorial forcing and the *summer* high-latitudes results from the extent of the deep easterlies in the summer hemisphere. Viewed from a summer hemispheric perspective the equatorial heat source is embedded in the tropical easterlies. The only modes which will result in a substantial local response are the Kelvin waves. As the *stationary* Kelvin waves do not possess a horizontal component of the group velocity vector relative to the heat source the response remains local to the source.

The conclusions reached regarding the atmospheric remote response to equatorial forcing were tested using realistic seasonal basic zonal wind fields and SSTA distributions consistent with (in a gross sense) the 1972-73 El Niño event. The fields appeared to agree, at least in character, with the observation of Ramage (1975) and with the longer term anomalous fields generated by Horel and Wallace (1981), at least in the Northern Hemisphere.

In viewing the response to an El Niño sea surface temperature distribution, the following point should be remembered. At all times, for any given equatorial boundary forcing, there will be a high-latitude response. *Thus the anomalous response which will be noted at higher latitudes during the anomalous El Niño periods will be the difference between the normal high-latitude response and the response typified in Figs. (12) and (14).*

With the above conclusions in mind it is possible to interpret many of the apparent paradoxical relationships which have appeared from observational and numerical modeling studies alike. In a sense the conclusions allow statistical studies to attack the boundary forcing problems of climate research armed with *a priori* expectations rather than hindsight *a posteriori* realizations. An understanding of the atmospheric response, as outlined in this paper as well as by Opsteegh and Van den Dool (1980), Webster (1981) and Hoskins and Karoly (1981), probably removes some of the problems inherent in the observational studies. Not only does it *lower* the statistical significance levels required of such studies,

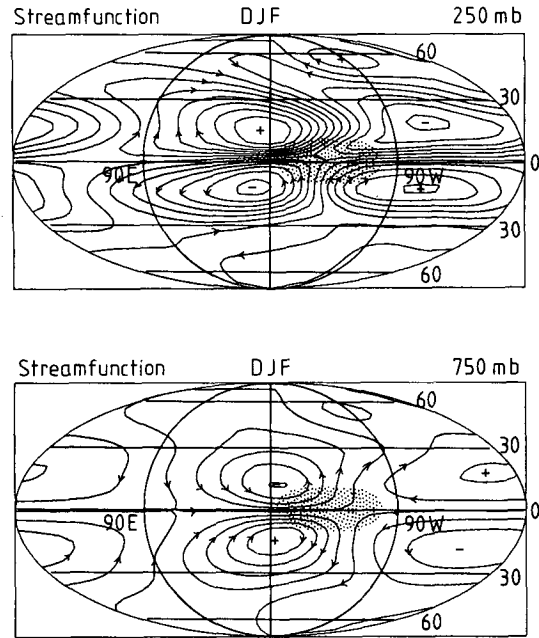


FIG. 14. As in Fig. 11 except for the streamfunction response (contour intervals:  $20 \times 10^5 \text{ m}^2 \text{ s}^{-1}$ ) of the DJF basic state to El Niño-type forcing.

it allows a prefiltering of a data set as both the location of a response and the time of year in which it will be maximized may be anticipated.

It should be emphasized that the results emerge from an extremely simple facsimile of the slowly evolving climate system. Indeed, it would be an error to suppose or pretend that complete answers will emerge from so simple a system. Such answers will probably only arrive through joint studies using more data, more sophisticated models, and with attendant phenomenological models such as discussed in this paper. With this in mind we should view the results relative to a number of caveats:

(i) The conclusions reached in this study (and also those by Opsteegh and Van den Dool, 1980; Webster, 1981; Hoskins and Karoly, 1981) are constrained by the assumption of zonal symmetry in the basic state. Observational studies of general circulation statistics indicate that quasi-stationary modes in extratropical regions *and* the tropics provide significant longitudinal perturbations in the mean state of the atmosphere. Such quasi-stationary perturbations will probably alter the character of both the teleconnection propagation and the local form of the atmospheric response. A study which considers the effect of a medium possessing a variable refractive index at low latitudes has been undertaken by Webster and Holton (1982).

(ii) The remote response of the model to (for example) equatorial El Niño forcing resembles only the character of the real response, at least as described

by Horel and Wallace (1981). The magnitudes are at least a factor of 2 too small. The reason for this is unclear and may be a consequence of the simplified basic state used in this study. Whereas we may anticipate with some measure of success the response of a fluid possessing deep easterlies and westerlies, we have little knowledge of the effect of anomalies on regions of strong vertical shear such as in the eastern Pacific Ocean (moderate westerlies atop easterlies) or in the Indian Ocean (strong easterlies above moderate westerlies). A study which considers the sensitivity of the response of such basic states to boundary forcing is underway.<sup>2</sup>

(iii) It must be anticipated that a considerable part of the variance of the interannual variability of the real atmosphere will not be associated with systematic variations in boundary forcing. For example, the various "centers of action" of middle and high latitudes (e.g., the Aleutian low region) appear to possess considerable interannual variability. However, only about 40–50% of the variability may be attributed directly to the effect of recognizable remote forcing. It is intriguing to note that the regions of maximum interannual variability appear to be spatially coherent and that these regions appear to coincide with the major teleconnection patterns. That is, the centers of action of both forced teleconnections and of "natural variability" appear to occupy the same geographical locations.

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<sup>2</sup>The author would like to thank the reviewer for suggesting this series of experiments.