Dynamics of Atmospheric Teleconnections during the Northern Summer

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ABSTRACT

In this paper, the mechanisms of northern summertime teleconnections are investigated using a barotropic model. In a series of numerical experiments we study the atmospheric response over the eastern Pacific–North America to an idealized local divergence source corresponding to the northward displacement of the ITCZ in the eastern Pacific. It is found that the response is much stronger in June than in May and is strongest when the forcing is located north of about 10°N. This can be explained in terms of the refractive properties of the climatological summertime subtropical jet stream over North America. In another series of experiments we examine the global response as a function of the longitudinal location of the tropical forcing. A wave train emanating from the subtropics of the western Pacific near the Philippines, arching across the Aleutians and the Gulf of Alaska, and terminating with a high anomaly over the continental United States appears over a wide longitudinal range of local forcing, suggesting the existence of a normal mode for the northern summertime climatological flow. The normal-mode concept is supported by further experiments using extratropical forcings as well as free-mode integrations. The upstream anomalous low over the Gulf of Alaska is found to be essential for the development of the anomalous high over the continental United States. These results indicate that an above-normal high over the continent may occur when the anomalous forcing (both tropical and extratropical) acts to amplify the normal-mode structure. The caveats and implications of the present results to the possible linkage between tropical forcing and United States droughts are also discussed.

1. Introduction

It has been hypothesized that the 1988 drought over North America may be related to anomalous tropical latent heating resulting from a northward displacement of the intertropical convergence zone (ITCZ) associated with the formation of cold water in the tropical Pacific (Trenberth et al. 1988). The crux of this hypothesis is that such anomalous heating forces an upper-level wave-train pattern stretching from the tropical eastern Pacific to North America with a high over the continental United States. The development of the high results in a northward shift of the summertime jet stream and storm tracks from their climatological positions, leading to enhanced dry conditions over the continent. Trenberth et al. (1988) showed that such a wave-train pattern can be simulated using a linear baroclinic model under climatological June mean-flow conditions.

Atmospheric teleconnections linking tropical convection and extratropical circulations are well known for wintertime circulations (Wallace and Gutzler 1981). Indeed, most of the past observational and theoretical studies of teleconnections are confined to the wintertime circulation. Dynamically, much of the observed characteristics of wintertime teleconnection patterns can be attributed to two basic mechanisms: (i) tropical influence through Rossby wave dispersion and its interaction with the basic-state flow (Hoskins and Karoly 1981; Webster 1981; Lau and Lim 1984; Branstator 1983), and (ii) extratropical normal mode arising from the barotropically unstable basic-state flow (Simmons et al. 1983). Theoretical and GCM studies have shown that these two mechanisms are not necessarily mutually exclusive, but often act in concert to produce large extratropical response initiated by tropical forcing (cf. Branstator 1985; Blackmon et al. 1987). Most important, both mechanisms are strongly influenced by the strength and location of the wintertime jet stream. However, it is not clear to what extent these two mechanisms can be applied to the northern summertime teleconnection. In contrast to the wintertime mean flow, the summertime mean flow has very different large-scale circulation patterns, characterized by a relatively weak, and northward-displaced, westerly jet stream. The dynamical properties of the summertime basic flow have not been documented. In this paper, we use a barotropic model to investigate the mechanisms of summertime telecon-
nection and its connection with tropical forcing. The possible link to United States drought will be discussed insofar as the upper-level high over the continent is taken as a signal of enhanced dry conditions over the same region. Because of the model limitation, only idealized forcings are used. As a starting point, we shall focus on the response over the eastern Pacific–North America sector to an idealized local divergence (heat) source that corresponds to the northward displacement of the ITCZ in the eastern Pacific. Then we expand the analysis to include the global extratropical response as a function of the longitude of the local forcing to explore the idea of extratropical normal modes. We conclude with a discussion of the caveats and implications of the present results to the possible linkage between tropical forcing and droughts in the United States. Currently, we are conducting a further study with a higher-resolution version of the present model and including more realistic forcings. Results of this further investigation will be reported in a separate paper.

2. Model description and diagnostics

Choosing a model for studying a regional yet globally connected phenomenon such as droughts in the United States and teleconnection patterns presents somewhat of a dilemma. For physical and dynamical completeness or for realistic simulation, a high-resolution general circulation model would be most appropriate. For easy understanding of the basic mechanisms and economy to carry out a large number of simulations, a simple model would be preferred. Barotropic models have often been used in simulating and diagnosing large-scale circulation patterns derived from observation or from general circulation models. In spite of model built-in shortcomings (see section 7 for discussions), barotropic models have been extremely useful in elucidating the basic mechanism of teleconnections (e.g., Simmons et al. 1983; Lau and Lim 1984; Branstator 1985; Held and Kang 1987; Sardeshmukh and Hoskins 1988). Focusing on this theme, we used a nonlinear spectral barotropic model with rhomboidal truncation at wavenumber 15 for this study. The basic equation of the model is the barotropic vorticity equation with a specified divergence field consisting of a climatological and an anomalous component. Thus,

\[
\frac{\partial \zeta}{\partial t} = -v_y \cdot \nabla (\zeta + f) - \nabla \cdot [(\zeta + f) (\nabla_x + v'_x)]
\]

\[
- a \zeta + b \nabla^2 \zeta + R \tag{1}
\]

where \( \zeta \) is vorticity and \( f \) the Coriolis parameter. The wind is decomposed into rotational \( v_y \) and divergent \( v'_x \) components. An overbar denotes a climatological quantity and a prime the deviation from the climatology. The terms involving the climatological divergent wind \( \nabla_x \) are related to climatological mean forcing, and those involving \( \nabla_x \) represent the specified forcing. They will be referred to as the source terms. A Rayleigh friction is included for dissipation and a biharmonic diffusion term for preventing energy accumulation at the truncated end of the spectrum. The \( R \) term represents an imposed forcing function to be explained shortly. Experiments were carried out with initial conditions \( (\psi) \) and climatological divergent flows \( (\nabla_x \psi) \) obtained from the globally gridded 10-year (1975–84) monthly mean 250-mb wind data of the National Meteorological Center (NMC) and idealized divergence fields \( (\nabla \cdot \nabla_x \psi) \) representing anomalous forcing. The model atmospheric response to the forcing was obtained by integrating (1) to reach steady state. The sensitivity of the response to the structure of the basic flow and to the location of the forcing were examined. Since the model is nonlinear and since climate states, particularly the zonal mean flow, cannot be maintained by barotropic process alone, the model must be constrained to preserve climate invariance in the absence of anomalous forcing, in order to have a well-defined response to specified forcing. To do so, two procedures were employed. First, a constant (in time) eddy forcing function \( (R) \) is specified. This is obtained by computing the initial tendency of the climatological flow in the model and then reversing its sign. The eddy forcing is then held fixed throughout the integration and acts to restore the mean flow to its specified climatology. Second, the zonal mean flow is also kept fixed throughout the experiment by setting to zero the tendency for the wavenumber-zero component at each time step. These procedures have little to do with realistic forcings needed to maintain the observed flow, but were used here only to keep the model climatological flow approximately invariant and to provide a quasi-nonlinear system that allows wave–wave interaction as well as extraction of wave energy from the mean flow. Unless otherwise stated, the model uses a Rayleigh friction coefficient \( (10\text{-day})^{-1} \) and a biharmonic diffusion coefficient \( 10^{-16} \text{ m}^4 \text{ s}^{-1} \).

We used two model diagnostics to aid in the interpretation of the results of experiments. The first is based on the computation of a refractive index for wave propagation in a zonal flow with nonzero meridional vorticity gradient. The second is the use of wave-flux vectors to depict the generation and dissipation of wave action for quasi-stationary flow (Plumb 1985). The theoretical bases for these two diagnostics are discussed briefly in the following.

a. Refractive index

For stationary flow, the barotropic vorticity equation linearized about a zonally symmetric basic flow \( U(\phi) \) is governed by the following equation:

\[
\frac{U}{a \cos \phi \partial \lambda} \nabla^2 \psi + \frac{\beta \partial \psi}{a \cos \phi \partial \lambda} = 0
\]
where
\[ \beta_m = \frac{1}{a} \frac{df}{d\phi} - \cos\phi \frac{d^2 (U \cos\phi)}{a^2 d(sin^2\phi)}; \]

\( U \) is the zonal mean flow, and \( \psi^* \) the steady-state perturbation streamfunction. Assuming \( \psi^* = \text{Re} \{ \psi^* \exp(\text{i} k\lambda) \} \), the equation becomes the following,
\[ \frac{1}{a^2 \cos\phi} \frac{\partial}{\partial \phi} \left( \cos\phi \frac{\partial \psi^*}{\partial \phi} \right) + r^2 \psi^* = 0 \tag{2} \]

where
\[ r^2 = \frac{\beta_m}{U} \frac{k^2}{a^2 \cos^2\phi}; \]
k is zonal wavenumber. Equation (2) is either oscillatory or evanescent depending on the sign of the refractive index \( r^2 \). Thus, the zero contour of \( r^2 \) will provide, in a linear sense, the boundary between the propagating and trapped solution to (2).

It should be noted, however, that the application of the concept of refractive index to the model results is only in a qualitative and comparative sense. Since the model is nonlinear and the mean flow is zonally asymmetric, leakage through the zero index line is possible.

b. Wave-flux vector

The presence of barotropic energy exchange arising from wave–wave and wave–(zonal) mean interaction in the present model can be regarded as extra vorticity sources/sinks on the linear version of the vorticity equation in addition to the imposed tropical divergence forcing. In the present model, the interaction between the wave and mean flow is allowed in the direction of mean flow to wave only because the zonal mean is fixed (see earlier discussion in this section). These secondary sources and sinks can be described by the wave-flux vector devised by Plumb (1985) as an extension of the Eliassen–Palm flux vector. For quasi-stationary and quasigeostrophic flow, the wave action \( A_z \), the wave-flux vector \( F_z \), and the source or sink \( C_z \) are governed by the following equation:
\[ \frac{\partial A_z}{\partial t} + \nabla \cdot F_z = C_z. \tag{3} \]

In the plane-wave representation, the wave-flux vector \( F_z \) is parallel to the group velocity \( c_g \) and indicates the direction of energy flow:
\[ F_z = c_g A_z. \]

In general, a divergence of \( F_z \) indicates export of wave activity (source), while a convergence of \( F_z \) indicates dissipation of wave activity (sink). The horizontal component of the flux vector \( F_{zH} \) can be computed from the steady-state eddy streamfunction \( \psi^* \) as
\[ F_{zH} = p \cos\phi \left( \frac{1}{2a^2 \cos^2\phi} \left[ \left( \frac{\partial \psi^*}{\partial \phi} \right)^2 - \psi^* \frac{\partial^2 \psi^*}{\partial \lambda^2} \right] - \psi^* \frac{\partial^2 \psi^*}{\partial \phi^2} - \psi^* \frac{\partial^2 \psi^*}{\partial \lambda \partial \phi} \right) \tag{4} \]

where symbols are conventional. On an equivalent barotropic level, (4) approximately represents the barotropic component of the wave flux. For a steady state in a barotropic model, it is the total wave flux. Due to the quasigeostrophic assumption, we limit our application of Eq. (4) to the region outside deep tropics.

3. Monthly mean flow and eastern Pacific heating

a. Summertime climatological 250-mb flow

Figure 1 shows the global distribution of the 250-mb wind averaged for the period May–August based on the NMC 10-year (1975–84) climatology. Although the global distribution is shown, the following discussion will be focused on the regional features near the Pacific–North America sector. Over the eastern Pacific and North America region, the most prominent feature is an extended trough stretching from the tropical central Pacific to the northwest coast of North America and a ridge over the continent. This is accompanied by a band of westerly winds that stretches from the central tropical Pacific to the southern part of the United States. The westerly band is sandwiched between the main jet stream near 45°N and a band of weak easterly wind near the equator. Over the North American continent, the jet stream is displaced northward due to the formation of the continental high.

From May to August, the climatological planetary circulations are broadly similar to Fig. 1, but the regional circulation for the individual months are quite different. Figure 2 shows the mean eddy streamfunction (zonal mean removed) for the months May, June, July, and August, respectively. In May, the circulation over the eastern Pacific is characterized by two deep troughs, one near 60°N and one between 20° and 30°N. The former diminishes and the latter intensifies as the season progresses. A trough similar in extent and intensity to the Pacific subtropical trough is also found over the subtropical Atlantic. The June pattern features one major trough over the subtropics of the eastern Pacific and a midcontinental high over North America. This trough–ridge system appears to be coupled to the trough over the western Atlantic along the coast of North America in the form of a wave train. By July, the Pacific trough/midcontinent ridge/Atlantic trough system is most pronounced. The pattern remains basically unchanged in August. It appears that the largest change of the regional circulation occurs during the transition from late spring to early summer (May to June) and that the dominant circulation features are quite persistent from June to August. In the following,
we examine the sensitivity of the extratropical response to tropical forcing to these basic-state circulations.

b. Responses to eastern Pacific tropical heat source

In this section, we study the model response to a prescribed forcing in the tropical eastern Pacific for different monthly mean basic flows. The tropical forcing consists of a meridional pair of divergence/convergence patterns with the center of the divergence situated at 13.3°N, as shown in Fig. 3. The maximum value of the divergence center is $2 \times 10^{-6}$ s$^{-1}$. This forcing is to mimic the anomalous heating due to a northward shift of the eastern Pacific ITCZ as measured by the averaged May–June 1988 OLR anomalies (cf. Trenberth et al. 1988). Figure 4 shows the model steady-state response to such a forcing for the May, June, July, and August climatological mean flow, respectively. Overall, the responses for June, July, and August are similar but the May response is quite different. For June to August, the common structure of the responses consists of an elongated upper-level high between the equator and 30°N near the forcing center, an enhanced low near the Gulf of Alaska extending southeast toward the Gulf of Mexico, and a high over the midsection of the continental United States. For June, this structure is connected eastward to a low near Greenland and a high over the northeast Atlantic (outside the range of Fig. 4). The appearance of an arching wave train emanating from the tropical source region across North America to the Atlantic is most conspicuous. Figure 4b is quite similar to Fig. 8 of Trenberth et al. (1988), which is simulated with a global primitive-equation baroclinic five-level stationary planetary-wave model, forced by a heat source similar to that in Fig. 3. For July and August, while the relative positions of the anomalies are similar to that in June, the anomalies seem to have broken up into individual centers, and the wave train appearance is less pronounced. In contrast, the May response is generally weaker everywhere. The subtropical high can still be found, though it is much diminished compared to the June–August response. This high is part of a wave-train pattern connecting the subtropical high to a low centered over the southern part of the continent and a weak high in the northeast corner of the continent. This pattern resembles that obtained from linear Rossby wave dispersion in the absence of mean flow (Hoskins and Karoly 1981). More discussion on the differences between the response for May and for the other months will be presented in the next section.

Before we examine the sensitivity of the response to the meridional location of the above forcing, an interesting question will first be addressed. How much of the above extratropical response can be described as simply an enhancement of the seasonal planetary-scale pattern? To examine this, Table 1 shows the correlation between the extratropical response and its corresponding climatological eddy circulation (see Fig. 2) for different domain sizes covering the continental United States and adjacent oceanic regions. Clearly the correlation increases monotonically from May to July regardless of domain size. When the domain includes the adjacent ocean, only the August correlation is significant (>0.6). When the domain size is about the size of the North American continent (23°–65°N, 130°–65°W), the correlations are all significant (>0.5) for June, July, and August, but not for May. The cor-
FIG. 2. The zonally asymmetric part of the climatological mean streamfunction at 250 mb for (a) May, (b) June, (c) July, and (d) August. Contour interval is in $2 \times 10^6$ m$^2$ s$^{-1}$. 
relations increase significantly when the region north of around 55° is excluded. When the domain is reduced to 23°–53°N, 130°–65°W, the correlation for June and July is 0.616 and 0.666, respectively. Here the overall correlation is highest for July and decreases slightly for August. The increasing correlation from May to August suggests that the extratropical response is increasingly phase locked to the wavy pattern in the monthly mean. Since the correlation suggests that about 25%–40% (correlation squared) of the spatial variance can be explained by the monthly mean pattern alone for the June to August cases, it may be argued that for these months the extratropical response is somewhat phase locked to, and partially represents, an enhancement of climatological regional circulation, particularly the early summer development of the United States high. For May, the response bears little resemblance to, and therefore appears to be decoupled from, the structure of the climatological flow. This suggests a different mechanism of response in May compared with those in June, July, and August.

c. Sensitivity to meridional location of heat source

Next we investigate the sensitivity of the model response to the meridional location of the tropical forcing. Because the results are similar for all three months from June to August, only results for the June case will be discussed. Figure 5 shows the June response of the model to the same dipole forcing placed at the same longitudes but with the positive divergence center located at 15.5°N, 13.3°N, 8.9°N, and 4.4°N, respectively. The magnitude of the forcing used is the same as described earlier. When the tropical divergence is placed between 9° and 15°N, the response pattern is quite similar. The subtropical high, which represents the direct response to the tropical forcing, moves equatorward with the source, but the extratropical low–high–low structure remains essentially unchanged as long as the forcing lies between 8°–16°N. The response is largest when the divergence center is at 13.3°N and reduces rapidly as the forcing is moved equatorward. When the anomalous forcing is south of 8°N, the response over the North Pacific and continental United States is drastically reduced. Yet the wave-train pattern resembles that expected from damped linear Rossby wave dispersion, similar to that for the May response but with the sign reversed (see Fig. 4a). Thus, for forcing in the tropical eastern Pacific, there appears to be a threshold latitude. If the tropical forcing is poleward of this latitude, the extratropical response is independent of the exact location of the forcing and hence strongly controlled by the basic flow. When the tropical forcing is equatorward of this latitude, the extratropical response is somewhat decoupled from the basic flow and quite well approximated by linear Rossby wave dispersion.

It should be noted here that the extratropical low–high–low response can be traced back across the ridge at the date line to a low–high structure in the western Pacific (outside the range of Fig. 5), suggesting that the enhanced low near the Gulf of Alaska and the United States high may be the result of superposition of two merging wave trains, one directly forced from the tropical eastern Pacific and another excited upstream from the western Pacific. This feature together with the global response to tropical forcing will be further discussed in section 6.

4. Teleconnection mechanisms

In this section, we examine the mechanism of summertime teleconnection by interpreting the results presented in the previous section using the two diagnostic tools discussed in section 2.

a. Refractive index analysis

In the following, "zonal mean" is taken to mean the average from 180° to 60°W. Figure 6 shows the line of zero refractive index or "turning latitude" for the zonal mean U, as a function of wavenumber for the May and June mean flow, respectively. Regions to the
Fig. 4. Model streamfunction response to specified dipole forcing in the tropical eastern Pacific (Fig. 3) with climatological mean flow for (a) May, (b) June, (c) July, (d) August. Contour interval in $10^6$ m$^2$ s$^{-1}$. 
left (right) of the zero line indicate propagating (non-
propagating) solutions for (2). The mean zonal wind
for each of these two months is also shown on the right
of the figure. There are two turning latitudes for the
June mean flow and one for the May mean flow. Ac-
cording to linear theory, these turning latitudes play a
key role in the scale selection for the extratropical re-
sponse to tropical forcing because it determines the
region in wavenumber domain in which propagating
or decaying solutions to the barotropic equation (2)
can exist for a given latitude.

For the June mean flow, propagating solutions are
roughly divided into three regimes. Between 10° and
30°N, the propagation region decreases monotonically
in the wavenumber domain so that at 30°N propaga-
tion is possible only for wavenumbers smaller than 5.
Between 30° and 45°N, the propagation region ex-
spands somewhat beyond wavenumber 5, but decreases
monotonically to 60°N where propagation is limited
to wavenumber 1 or less. The latitudes 30°–45°N and
45°–60°N correspond approximately to the equator-
ward and poleward flank of the seasonal mean sub-
tropical jet over the eastern Pacific and North America,
where the mean meridional vorticity gradient is quite
different. Propagation is restricted to only wavenumber
1 in the region just south of 60°N. There is a slight
increase in the propagation region to wavenumber 2
at 70°N, beyond which propagation is again restricted
to wavenumber 0 and 1. In contrast, the southern
turning latitude, located at about 10°N, is almost in-
dependent of wavenumber. This is due to the presence
in June of a zero wind line separating easterly mean
flow in the equatorial regions and westerly mean flow
in the extratropics. The presence of the southern zero
contour indicates a nonpropagating solution equator-
ward of 10°N and suggests that any source placed en-
tirely equatorward of this latitude will not produce
propagating solutions poleward. This appears to be
consistent with the large reduction in extratropical re-
sponse as the dipole forcing is moved south of the
threshold latitude, as noted in the discussion of Fig. 5.
It is noted that there is also a critical line (where the
zonally averaged \( \bar{U} \) is zero) near 10°N for the June
mean flow (not shown). Although the general idea of
the linear wave refraction is found useful here, the
singular behavior related to absorption at a critical line
predicted by linear theory is unimportant in a nonlinear
model with a wavy mean state.

<table>
<thead>
<tr>
<th></th>
<th>18°–75°N</th>
<th>18°–75°N</th>
<th>23°–70°N</th>
<th>23°–67°N</th>
<th>23°–58°N</th>
<th>23°–53°N</th>
</tr>
</thead>
<tbody>
<tr>
<td>May</td>
<td>−0.164</td>
<td>−0.048</td>
<td>0.055</td>
<td>0.065</td>
<td>0.048</td>
<td>−0.005</td>
</tr>
<tr>
<td>June</td>
<td>0.054</td>
<td>0.384</td>
<td>0.416</td>
<td>0.438</td>
<td>0.564</td>
<td>0.616</td>
</tr>
<tr>
<td>July</td>
<td>0.440</td>
<td>0.494</td>
<td>0.481</td>
<td>0.504</td>
<td>0.616</td>
<td>0.666</td>
</tr>
<tr>
<td>August</td>
<td>0.503</td>
<td>0.620</td>
<td>0.597</td>
<td>0.573</td>
<td>0.547</td>
<td>0.528</td>
</tr>
</tbody>
</table>

While the zonal mean flow for May and June pole-
ward of the jet stream (centered near 45°N) is essen-
tially the same, on the equatorward flank of the jet
stream the zonal flow between the two months differs
significantly. This is reflected in the changes in the zero-
refractive index curve between May and June (Fig. 6).
For wavenumbers larger than 3, the turning latitudes
for May are shifted equatorward, reducing the regions
of extratropical propagation. As can be seen in Fig. 4,
the dominant scales for the extratropical response to
an eastern Pacific heat source is around wavenumber
4 to 5. The zero–refractive index line for May (dotted
line) indicates that disturbances with scales of wave-
number 4 to 5 are largely evanescent north of 30°N.
This means that disturbance of these scales generated
from the south will not be able to penetrate these lati-
itudes. Further, there is a complete cutoff of a propa-
gating solution at 60°N for all wavenumbers, indicating
that it is highly unlikely to generate any significant ex-
tratropical response beyond this latitude from tropical
forcing. Another significant difference is the absence
of a southern zero–refractive index line for May, due
to the presence of westerly winds down to the equatorial
region. Therefore, in May, medium- to small-scale dis-
turbances generated in the equatorial region are favored
to propagate within and confined to the sub tropics.
Although the preceding analyses are based on linear
dynamics, they seem to be consistent with the model
response shown in Figs. 4 and 5.

b. Wave-flux analysis

The wave-flux analysis offers a somewhat different
perspective on the mechanisms of the extratropical re-
sponse compared with the refractive index analysis.
The former is based on the analysis of the response pat-
tern itself and is concerned with what maintains the
steady-state (final) response, while the latter is based
on the analysis of the zonal mean flow and is more
appropriate for a description of the initial tendency of
the response. The two analyses will yield similar results
for linear but not for nonlinear systems.

To better understand extratropical wave sources in
contributing to the extratropical response shown in Fig.
4, the possible role of stationary waves in the main-
tenance of the climatological summertime circulation
pattern is first discussed. Figures 7a and 7b show the
wave-flux vectors for May and June climatological
Fig. 5. Model streamfunction response to specified dipole forcing for the June mean flow, when the center of the forcing is placed along 120°W at (a) 9°N, (b) 7°N, (c) 2°N, and (d) 2°S. The positive (negative) center of the dipole forcing is indicated by cross (dot).
The wave vectors have been recomputed from the tropical forcing experiments for May and June (see discussions for Figs. 4a and 4b). Figure 8 shows the difference in the flux vector between the tropical forcing mean flow, respectively. Most noticeable in both cases is the concentration of large equatorward flux vectors in the midoceanic trough region of the central to eastern Pacific directed toward the tropics. This indicates that much of the wave activity in this region is generated from the extratropics, possibly from energies extracted from the jet stream poleward and upstream of the mid-Pacific trough (see Fig. 1). In contrast, the wave fluxes over the continental United States and the Aleutians are much weaker. Since no flux vectors are seen directed poleward from the tropics, no evidence is apparent of direct tropical influence on the extratropical climatological summertime flow. However, this does not preclude indirect tropical effects that may influence the evolution of the large-scale circulation, leading eventually to the observed climatological flow.

Fig. 6. Zero-refractive line as a function of latitude and wave-number for the May and June mean flow in the Northern Hemisphere. Also shown on the right is the latitudinal profile of the zonal mean flow averaged between 180° and 60° W.

Fig. 7. Wave-flux vector for the climatological monthly mean flow at 250 mb for (a) May and (b) June. The scale of the wave-flux vector is shown at the upper right corner. Units in $1.25 \times 10^9$ m$^2$ s$^{-2}$ Pa.
experiment and the climatology for May and June, respectively. The May anomalous flux vectors, which correspond to the anomaly field shown in Fig. 4a, show a weak meridional propagation of wave activity from the forced region into North America, consistent with the refractive index (linear) analysis. Elsewhere, there is a very weak secondary wave source (flux divergence) off the coast of northern California, exporting wave energy northward. In contrast, the June flux vector indicates much stronger wave activity over the United States and adjacent regions. As expected on the basis of linear dynamics, a major wave source is found emanating from the tropical eastern Pacific between 130° and 90°W. Interestingly, a pronounced second wave source appears over the subtropical eastern Pacific (40°N, 140°W), near the Gulf of Alaska, exporting wave energy eastward into the continental United States. These two branches appear to converge over the continent at 45°N and are presumably responsible for the maintenance of the steady response over this region.

5. Vorticity dynamics

In this section, the importance of secondary wave source(s) in leading to the difference between the May and June response is examined in terms of the different dynamical processes inherent in Eq. (1), which may be rewritten as follows:

\[
\frac{\partial \zeta'}{\partial t} = A + S + F
\]

where \(F\) represents the specified forcing function including friction; \(A\) is the vorticity advection by the rotational part of the wind, which is further grouped into linear (I) and nonlinear (II) terms

\[
A = - (\nabla \cdot \nabla \hat{\zeta}' + \nabla \cdot \nabla \hat{\eta}) - \nabla \cdot \nabla \hat{\zeta}'
\]

(I) \hspace{1cm} (II)

\(\hat{\eta} = \hat{\zeta} + f\); and \(S\) is the source or sink of vorticity anomaly due to the divergent part of the specified cli-
matological or anomalous winds, which consists of the advection (III) and stretching (IV) terms

\[ S = - (v'_k \cdot \nabla \eta + v_k \cdot \nabla \xi' + v'_l \cdot \nabla \xi') \]

III

\[- (\eta D' + \xi' D + \xi' D') \] (7)

IV

\[(D \text{ is divergence}) \]

It is noted that, even though the anomalous divergence forcing is specified only in a small area and generates vorticity anomalies locally, its effect can be far reaching through the following processes: (i) The locally generated vorticity anomaly disperses and interacts with the mean vorticity through the processes represented by \( A \), (ii) the dispersed vorticity anomaly interacts with the mean divergence to generate new sources or sinks through \( \xi' D \) and/or to produce anomalies through \( v_k \cdot \nabla \xi' \), and (iii) the continuity requirement for the locally specified divergence forcing induces a widespread divergent velocity, which modifies the vorticity distribution through \( v'_k \cdot \nabla \xi \). The importance of processes (i), (ii), and (iii) depends upon the fields of the climatological vorticity (\( \xi \)) and divergence (\( D \)).

To focus on the difference of the June and May modeled response, Fig. 9 shows June minus May contributions to \( S \) and \( A \) computed from the model. The inverse Laplacian of \( S \) and \( A \) has been taken to convert from vorticity to streamfunction to facilitate direct comparison with Fig. 4. The forcing \( S \) due to terms III and IV is shown in Fig. 9a. This consists of a dipole source/sink in the region of the forcing, and a positive tendency in the far field near Alaska. However, the amplitude of \( S \) is much smaller compared to that of \( A \) (Fig. 9b). Consistent with the extratropical response shown in Figs. 4a and 4b, the major difference in the effect of \( A \) between June and May is a tendency for a (streamfunction) high over the midcontinent and lows off the coasts. The similarity between Fig. 9b (\( A \)) and Fig. 9c (\( A + S \)) indicates that much of the difference in the June and May response is due to \( A \) or the \( v'_k \) advection.

Further decomposition of \( A \) indicates that a large part of the trough–ridge pattern over the United States and adjacent oceanic regions is due to the linear vorticity advection by the rotational part of the wind—term I in (5) (Fig. 10a). In particular, the low over the Gulf of Alaska and the high over the eastern half of the United States are due to this effect. The nonlinear interaction (term II, Fig. 10b) in most of the area produces anomalies of the opposite sign to that produced by the linear advection (term I). Large cancellations between these two terms occur across the continent. The magnitude and distribution of the vorticity anomaly would be significantly different if the nonlinear interaction is neglected. Comparing Fig. 10b with Fig. 9c, it can be seen that a large part of the net (\( S + A \)) in the region of the Gulf of Alaska, which coincides with the secondary wave source previously noted, is due to the linear vorticity advection (Fig. 10a). To confirm the importance of \( v_k \) advection in leading to the June response shown in Fig. 4b, the above experiment is rerun with the June mean flow, but with the \( v_k \) advection—term III in (6) suppressed. The response obtained (not shown) is almost identical to that in Fig. 4b. The relatively minor role played by the \( v_k \) advection represents a marked deviation of the behavior of the summertime circulation to that of the wintertime circulation (Sardeshmukh and Hoskins 1988). This point will be discussed further in section 7.

6. Extratropical normal modes

So far, we have focused our discussion on the forced model responses in the eastern Pacific–North America sector, particularly the pattern of Rossby wave dispersion in May and the pronounced low–high pattern in June, which consists of the high over the United States and the strong low off the northwest coast of North America near the Gulf of Alaska. Let us now turn our attention to another teleconnection mechanism mentioned in the Introduction, namely, normal-mode excitation. As shown earlier, the pronounced extratropical low–high pattern in June is independent of the exact latitude location of the tropical forcing as long as it is northward of 10°N (Fig. 5a–c), and the low coincides with the secondary wave source region (Fig. 8b). It has also been noted in the discussion in section 3c that the enhanced low–high pattern may be a result of the superposition of the directly forced mode with a wave train originated from the western Pacific. Is this wave train a normal mode of the June mean flow? Is the secondary wave source, which is almost due north of the specified forcing, related to normal-mode excitation? How important is the role of the secondary wave source? The following experiments are carried out to explore the concept of normal modes and the relative importance of the normal mode (internal) versus the direct forcing (external) mechanism.

a. Sensitivity to longitudinal position of tropical forcing

In this section, we present the results of a set of experiments to study the global response to tropical forcing. Figures 11b–h show a sequence of the steady-state global response for a tropical divergence centered, respectively, at 5°N and at 30° longitude intervals from 60°E through the date line to 120°W for the June climatological mean flow. As the forcing is moved from west to east, three organized features can be discerned in all of the response patterns. The most obvious one is a pair of elongated anticyclones straddling the equa-
tor, characteristic of the Gill-type equatorially trapped solution. This feature moves eastward with the forcing. The other organized features are the two arching wave train patterns, which do not move with the forcing. They are roughly symmetric about the equator, one stretching from the subtropical western Pacific near the Philippines across the Aleutians to North America, the other in the Southern Hemisphere stretching from the central Pacific to Chile and Argentina. When the forcing is west of 150°E, the response is mainly the Gill type. As the forcing is moved east of 150°E, the wave-train response becomes quite pronounced and phase-locked to the climatological mean eddy flow. The insensitivity of the extratropical pattern to the longi-

Fig. 9. Decomposition of the difference between June and May response. (a) "source" term $S$, (b) advective term $A$, and (c) $S + A$. Contour interval in 2 m$^2$ s$^{-2}$. Zero contour is omitted.
the perturbed flow is allowed to evolve without anomalous forcing. If an unstable normal mode with linear growth rate greater than the model dissipation rate exists, it should emerge eventually and its structure should be virtually independent of the initial conditions. Such a normal mode was not found or did not survive for the 10-day dissipation time scale used in all previous experiments. When the dissipation time scale is doubled to 20 days, the arching wave-train structure emerges over the north Pacific and the continental United States after 20–30 days. This wave train is not stationary but evolves slowly. Figure 12 shows the wave-train structures 40 and 60 days, respectively, after a positive streamfunction perturbation is introduced over the Tibet region in the June mean flow. They are similar to those found in Figs. 11e–h, which are forced by anomalous divergence in the tropical eastern Pacific, and to those shown in Figs. 5a–c, which are forced by divergence dipoles at 120°W. Other results of our experiments (not shown) indicate that certain geographic locations and patterns of initial perturbation are favorable for the excitation of such a mode, for example, a positive streamfunction perturbation over East Asia.
Fig. 11. (a) Pattern of the imposed divergence forcing and (b) the global response of streamfunction to the forcing centered along 5°N at 60°E, (c) 90°E, (d) 120°E,
FIG. 11. (Continued) (e) 150°E, (f) 180°, (g) 150°W, and (h) 120°W.
Contour interval for streamfunction in $2 \times 10^6$ m$^2$ s$^{-1}$. 
and the North Pacific along \(\sim 30^\circ\)N. Briefly, our results suggest that a marginally unstable mode exists for the June climatological flow and that heating over the tropical eastern Pacific may amplify and anchor this mode.

Next, we introduce a fixed extratropical divergence source in the region of the maximum secondary wave source, that is, the Gulf of Alaska. The steady-state response is compared with previous results. Here the sign of the forcing is consistent with those derived from the vorticity dynamics analysis of section 5. Figure 13 shows the response for a negative divergence forcing in the region of the Gulf of Alaska for the May and June basic flow, respectively. The response, almost identical for May and June, features the characteristic dipole pattern with a low over the Gulf of Alaska and a high over the continental United States. This pattern appears to agree with the eastern portion of the normal-mode wave train but without the upstream and subtropical features (cf. Fig. 11f–h and Fig. 12). Since the
zonal mean flows in midlatitudes are quite similar in May and June, the marked resemblance between the response for these two months merely indicates that it is a midlatitude signal that is not affected significantly by the transmission properties of the tropical and subtropical mean flow. Undoubtedly, the (streamfunction) high over the United States in the experiment is not a local phenomenon but is closely related to the upstream low in the Gulf of Alaska region. This suggests the possibility that a preexisting anomaly in the Gulf of Alaska region may provide an anchoring mechanism to an approaching wave train and thereby affects the downstream development over the continental United States.

7. Discussion and conclusions

The results we have presented so far suggest that during northern summer, there are two plausible mechanisms by which tropical forcing may lead to atmospheric teleconnections over the eastern Pacific and North American region. First is the direct response related to Rossby wave dispersion, which is a function of the mean zonal wind. Here, we found that the response is strongest when the forcing is located north of approximately 10°N in the eastern Pacific. The amplitude of the extratropical response is consistent with the refractive properties of the climatological summertime subtropical jet stream over North America.

Second is the excitation of a normal mode in a wavy basic state. This normal mode consists of an arching wave train stretching from the subtropical western Pacific across the Aleutians to North America. This wave train is also found in a higher-resolution (R30) version of the present model when forced by a realistic distribution of tropical heating (Lau 1992). Nitta (1987, 1989) has noted a similar wave train forced by tropical heating over the northern Philippines. However, Nitta's pattern has only very weak amplitude over North America, and the spatial scale is somewhat smaller than the pattern noted here. Our results suggest that this normal mode may have a significant impact on the summertime teleconnection pattern over the eastern Pacific and the North American region. Extratropical response to heating in the tropical eastern Pacific con-
sists of both direct response as well as normal-mode excitation.

The wave-flux computation suggests that tropical forcing has little direct influence on the maintenance of the climatological summertime planetary-scale circulation over the eastern Pacific–North America region. As for the anomalies, the high over the United States appears to be linked to a low over the Gulf of Alaska as part of the aforementioned normal-mode structure. The response over the United States is sensitive to a wave source upstream in the Gulf of Alaska region. Analysis of the barotropic vorticity equation indicates that a large part of this anomalous trough-ridge pattern is due to the anomaly-mean vorticity advection by the rotational wind. The same results were obtained by suppressing the $v_\ast$ advection term. This is quite different from the results of Sardeshmukh and Hoskins (1988), which showed that the advection by the divergent component of the wind is important for the wintertime flow because of the nearly perpendicular direction between the mean vorticity gradient and the divergent wind in the region of the East Asian jet. For the summertime basic flow over the eastern Pacific, the jet stream is much weaker and the zonality of the basic vorticity gradient is less pronounced. Here the “apparent” Rossby wave source is due to anomaly-mean flow interaction largely involving the rotational part of the wind.

While the above results provide some basic understanding of the dynamics of the summertime teleconnection in the eastern Pacific, there are some caveats that deserve special caution. The use of a barotropic model has inherent drawbacks or uncertainties, such as neglecting the contribution of baroclinic processes, the choice of equivalent barotropic level, and the necessity of using a proxy for heating. The present approach is defensible only because it is a preliminary study to provide guidance for interpretation of observation and results from general circulation models. The resemblance of our results for the tropical eastern Pacific to Trenberth et al.’s based on a global primitive-equation baroclinic five-level model is reassuring. The more serious problem is probably the adequacy of the model resolution to represent details of the forcing. The model used is a low-resolution model (R15) and the imposed tropical forcing all have meridional scales much larger than those attributed to the observed displacement of the ITCZ in the tropical eastern Pacific. Thus, it is questionable if the dipole forcing used here and that used in Trenberth et al. actually corresponds to the heating observed during United States drought episodes. For example, the anomalous outgoing longwave radiation during May–June 1988 indicates that the strongest heating anomaly is in the tropical western Pacific and that the shift of the eastern Pacific ITCZ corresponds to a relatively small contribution to the total anomalous heating. Therefore, we should be warned against making direct reference of the present results to specific cases of United States drought, especially when meridionally small-scale heating is involved. The results here reveal plausible fundamental mechanisms only for the summertime teleconnection and its relationship to tropical forcings. Further, there is uncertainty that the extent to which the extratropical normal modes can be excited may be dependent on model resolution and the magnitude of the dissipation used. These limit the application of the present model to idealized tropical forcing only. Work is now being undertaken to carry out experiments with a higher-resolution version (R30) of the model to study the atmospheric response to specified forcings that are better approximations to the observed heating such as occurred during the Northern Hemisphere summer of 1988.

REFERENCES


