A Diagnostic Model for Equatorial Wave Disturbances: The Role of Vertical Shear of the Mean Zonal Wind

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(Manuscript received 19 August 1970)

ABSTRACT

A model is developed to diagnose the response of the atmosphere to a known distribution of diabatic heating. The linearized primitive equations in spherical coordinates are reduced to a single partial differential equation relating the perturbation geopotential to the diabatic heating pattern. The model diagnoses the atmospheric perturbation due to a heating pattern of specified zonal wavenumber, frequency, and distribution in the meridional plane.

The model is used to test the hypothesis that the observed westward propagating wave disturbances in the equatorial Pacific are Rossby waves driven by the latent heat release in the cloud clusters embedded within the waves. For a diabatic heating pattern designed to model the heating in cloud clusters the model duplicates many features of the observed waves. The computed perturbation meridional wind field has maxima in the upper and lower troposphere separated by a relatively undisturbed region in the mid-troposphere. The structure of the disturbance is quite sensitive to vertical shear of the mean zonal wind. In particular, with westerly shear in the lower troposphere the precipitation occurs to the east of the surface trough, but with easterly shear the precipitation zone is west of the trough. These features are all in qualitative agreement with observations in the western and central Pacific.

1. Introduction

The existence of westward propagating wave disturbances in the equatorial Pacific first described by Riehl (1948) has been confirmed by several recent observational studies. Most of these studies are based on spectral analysis of conventional radiosonde data (Yanai et al., 1968; Wallace and Chang, 1969; Nitta, 1970; Chang et al., 1970). In addition to the spectral studies, however, other types of evidence are available. In particular, Chang (1970) used time-longitude sections constructed from successive daily satellite photographs cut into thin zonal strips to demonstrate the existence of westward propagating cloud patterns in the zone from 5-15N.

That these westward moving cloud patterns are associated with the wave disturbances in the wind field has been shown by Reed (1970). Reed intensively analyzed the disturbances in the western Pacific for the period July–September 1967 by the combined use of spectral analysis, satellite cloud photographs, time-height sections of the meridional wind, conventional synoptic analyses of the meridional wind field, and surface observations of cloudiness and precipitation.

Together these several studies indicate beyond reasonable doubt that westward propagating synoptic disturbances do exist in the tropical Pacific and that much of the precipitation in that area is associated with the cloud clusters which are embedded within these waves. At the same time it should be said that the observed average amplitude, period and wavelength of these disturbances vary from year to year, and that the structure of the disturbances also depends on the season and longitude. The range of periods reported in the above spectral studies is ~4–7 days while the reported zonal wavelengths range from 2000–10,000 km.

This wide range in zonal wavelengths is apparently due to the existence of two types of wave modes (Chang et al., 1970). The first type with wavelength range of 6000–10,000 km has its maximum amplitude in the upper troposphere. The axes of waves of this mode generally have a substantial tilt with height. Nitta (1970) has suggested that this mode is the mixed Rossby-gravity mode first observed in the stratospheric data by Yanai and Maruyama (1966). This mode has been discussed theoretically by Lindzen and Matsuno (1968). The second type of wave mode with a wavelength range of 2000–5000 km is most prominent in the lower troposphere at the western Pacific stations. Holton (1970) has shown that this latter mode may be theoretically interpreted as a forced equatorial Rossby wave.

Because these two families of waves may exist simultaneously with one predominating in the statistics at the lower levels and the other predominating at the higher levels, spectral studies often indicate little coupling between the lower and upper tropospheric disturbances. However, both types of waves appear to be associated with convection, and at least some of the

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1 Contribution No. 226, Department of Atmospheric Sciences, University of Washington.
waves extend through the entire depth of the troposphere. Thus, the spectral results of Chang et al. (1970) indicate that at times there is relatively high coherence between the wave disturbances in the upper and lower troposphere. In that case the upper and lower level disturbances are generally out of phase, while the largest temperature fluctuations occur in the intervening layer with the warmest air above the low-level trough. However, it is dangerous to generalize these results to other seasons and longitudes. Reed (1970) has shown that the vertical structure of the disturbances is quite sensitive to the vertical profile of the mean zonal wind in which the disturbances are embedded. Fig. 1, reproduced from Reed's paper, gives an indication of the different vertical profiles of amplitude of the disturbance meridional wind observed at stations with rather different mean zonal wind distributions. In summary, the observational evidence proves the existence of westward propagating wave disturbances in the equatorial Pacific; however, a number of questions remain concerning their origin, maintenance and structure.

Three processes have been suggested as likely energy sources for the formation and maintenance of these waves:

1) Barotropic instability of the mean zonal flow in the presence of strong lateral shear. This has been studied theoretically by Nitta and Yanai (1969).

2) Lateral forcing by mid-latitude disturbances. This process has been investigated by Mak (1969).

3) Latent heat release in convection driven by frictional moisture convergence in the conditionally unstable tropical atmosphere. This process (CISK) has been studied theoretically by Yamasaki (1969).

Although each of these mechanisms may occasionally be responsible for initiating equatorial wave disturbances, the observational studies cited above strongly indicate that the third mechanism is of primary importance at least for waves which form along the ITCZ. In addition to the evidence previously mentioned (i.e., the correspondence between westward propagating cloud and precipitation patterns and wave disturbances in the wind field), the spectral studies of Chang et al. (1970) have shown that there is a strong correlation between the 300-mb temperature and relative humidity. Thus, the strongest convective activity with its associated upward motions occurs when the mid-tropospheric temperatures are warmer than average. As a result these waves convert potential energy derived from the release of latent heat to kinetic energy. That this is an important feature of the energetics of the general circulation has been emphasized by Riehl (1969).

In this paper the hypothesis that the observed waves are forced Rossby waves driven by latent heat release is tested by examining the structure of the atmospheric

![Figure 1](https://example.com/figure1.png)

**Fig. 1.** Variance of meridional component in 3–10 day band and time-averaged zonal wind as functions of height for selected stations (after Reed, 1970). The upper portion of the variance profile at Johnston Island is plotted to half scale.
disturbance excited by a specified heating pattern. In this
diagnostic model no attempt is made to determine
theoretically how the small-scale convective cells
interact cooperatively to provide a large-scale heating
pattern or how this heating should be distributed in the
vertical. Nor is any attempt made to deduce the
influence of the synoptic wave disturbance on the
structure of the heating pattern. Rather, the latent
heating by the cloud cluster is prescribed as an external
parameter, and the resulting structure of the forced
wave is deduced.

2. Formulation of a diagnostic model

The basic equations of the model are the linearized
primitive equations in spherical coordinates\(^1\) \((\lambda, \phi, z)\),
where \(\lambda\) is the longitude, \(\phi\) latitude, and \(z\) \(-H \ln(p/p_s)\)
the vertical coordinate. Here \(H\) is a constant scale
height, \(p\) the pressure and \(p_s\) a standard sea level
pressure. In this system the horizontal momentum
equations, hydrostatic equation, continuity equation,
and first law of thermodynamics may be written as
follows:

\[
\begin{align*}
\frac{\partial u}{\partial t} + \frac{\partial u}{\partial \lambda} + \frac{\partial \bar{v}}{\partial \phi} + u \frac{\partial \bar{v}}{\partial \phi} &= -\frac{\partial \Phi}{\partial \phi} + 2\Omega \sin \phi - \kappa u, \quad (1) \\
\frac{\partial \bar{v}}{\partial t} + \frac{\partial u}{\partial \lambda} + \frac{\partial \bar{v}}{\partial \phi} &= -\frac{\partial \Phi}{\partial \phi} - 2\Omega \sin \phi - \kappa v, \quad (2) \\
\frac{\partial \Phi}{\partial z} &= \frac{RT}{H}, \quad (3)
\end{align*}
\]

where we have neglected the vertical advection term in
(1) and the meridional advection of the mean tempera-
ture in (5) since both terms are small for the motions
of interest.\(^2\) Here \(u\) and \(v\) are the perturbation zonal,
meridional and vertical velocity components, \(\phi\) the
perturbation geopotential, \(\bar{u}\) and \(\bar{v}\) the zonally
averaged zonal velocity and temperature fields, \(\epsilon_p\) the
specific heat at constant pressure, and \(Q\) the rate of condensation
heating per unit mass. In (1) and (2) \(\kappa\) is a Rayleigh
drag coefficient, and in (5) it is a rate coefficient for
Newtonian cooling. Thus, (1), (2) and (5) all contain
linear damping terms. As we shall see below, these
damping terms are necessary to prevent the occurrence
of singularities in the final diagnostic equation.

To nondimensionalize (1)–(5), we let the radius \(a\)
of the earth be the horizontal scale, the reciprocal of twice
the angular velocity of the earth, \((2\Omega)^{-1}\), be the time
scale, and the scale height \(H\) be the depth scale. We can
then define nondimensional variables as follows:

\[
\begin{align*}
\bar{u}^* &= \frac{u}{\cos \phi (2\Omega)^{-1}} \\
\bar{v}^* &= \frac{v}{\cos \phi (2\Omega)^{-1}} \\
\bar{u}^* &= \frac{u}{\cos \phi (2\Omega)^{-1}} \\
\bar{w}^* &= \frac{w}{w(2\Omega)^{-1}} \\
\Phi^* &= \frac{\Phi}{(2\Omega)^{-1}} \\
T^* &= \frac{RT}{(2\Omega)^{-1}} \\
\bar{T}^* &= \frac{T}{(2\Omega)^{-1}} \\
Q^* &= \frac{KQ(4\Omega^2a^2)^{-1}}{(2\Omega)^{-1}}
\end{align*}
\]

where the asterisks denote nondimensional variables.
Letting \(y = \sin \phi\) be the meridional coordinate, the non-
dimensional form of the system (1)–(5) may be
written as

\[
\begin{align*}
\frac{\partial \bar{u}}{\partial t} + \bar{u} + \bar{v} \frac{\partial \bar{u}}{\partial \lambda} + \bar{v} \frac{\partial \bar{u}}{\partial \phi} &= -\frac{\partial \Phi^*}{\partial \phi} + \frac{1}{1-y^2} \frac{\partial \Phi^*}{\partial y} - Du, \quad (1a) \\
\frac{\partial \bar{v}}{\partial t} + \bar{u} \frac{\partial \bar{v}}{\partial \lambda} + \bar{v} \frac{\partial \bar{v}}{\partial \phi} &= -\frac{\partial \Phi^*}{\partial \phi} - \frac{1}{1-y^2} \frac{\partial \Phi^*}{\partial y} - Dv, \quad (2a) \\
\frac{\partial \bar{u}}{\partial z} &= T, \quad (3a) \\
\frac{\partial \bar{T}}{\partial t} + \bar{u} \frac{\partial \bar{T}}{\partial \lambda} + \bar{v} \frac{\partial \bar{T}}{\partial \phi} &= -\frac{\partial \epsilon_p H}{\partial \lambda} - \frac{\partial \bar{T}}{\partial \phi} - \frac{\partial T}{\partial \phi} - \kappa T, \quad (5a)
\end{align*}
\]

where we have dropped the asterisks since all variables
are now dimensionless.\(^4\) The two dimensionless parameters
which appear here are the dissipation coefficient
\(D = \kappa(2\Omega)^{-1}\)
and the static stability parameter
\(S = \frac{K}{\epsilon_p} \frac{d\bar{T}}{dz}\)

\(^3\) An equatorial \(z\)-plane model proves unsatisfactory because a
heat source with narrow latitudinal extent will excite a large
number of latitudinal modes, and, as shown by Lindzen (1967),
only the lowest modes decay rapidly enough away from the equator
to be valid approximations to the forced Rossby waves on a
sphere.

\(^4\) The meridional advection of the mean temperature by the
perturbation velocity is, of course, the process which supplies
the energy for baroclinically unstable waves. Since this term is
negligible compared to the diabatic heating, it is clear that the
observed equatorial waves are not the result of ordinary baroclinic
instability.

\(^5\) It should be noted that the scaling used here does not pro-
duce a nondimensional set in which all the variables are order
unity. It is chosen solely to facilitate the mathematical develop-
ment.
We next assume a solution in the form of a wave propagating in the zonal direction with zonal wavenumber \( k \) and frequency \( \omega \). In that case, \( Q \) can be written as

\[
Q = Q'(y,z)e^{i(kx + \omega t + z\phi^2)}.
\]

The system (1a)-(5a) will then have forced solutions of the form

\[
\begin{pmatrix}
  u' \\
  \nu' \\
  \omega'
\end{pmatrix} = e^{i(kx + \omega t + z\phi^2)}.
\]

Substituting the above into (1a)-(5a) and eliminating \( T \) between (3a) and (5a), we have

\[
i\phi' + \left( \frac{\partial \phi}{\partial y} - y \right)\nu' = -i k \Phi', \tag{6}
\]

\[
i\phi' + y\nu' = -(1 - y^2) \frac{\partial \phi'}{\partial y}, \tag{7}
\]

\[
\frac{iku'}{1 - y^2} + \frac{\partial \phi'}{\partial y} + \left( \frac{\partial}{\partial z} - \frac{1}{2} \right)w' = 0, \tag{8}
\]

\[
i\phi' \left( 1 + \frac{\partial}{\partial z} \right) + \frac{1}{2} \Phi' + w' = Q', \tag{9}
\]

where

\[
\phi = \omega + \frac{k\nu}{1 - y^2} - iD. \tag{10}
\]

In this study we assume that \( \bar{u} = u_0(z)(1 - y^2) \) so that at any height the angular frequency of the wave relative to the mean zonal wind is independent of latitude. Thus, if the dissipation depends only on height, then \( \bar{\phi} = \bar{\phi}(z) \).

Since for realistic zonal wind distributions \( u_0 \ll 1 \) we may neglect \( \partial \bar{u}/\partial y \) compared to \( y \) in (6). With these restrictions, (6) and (7) may be solved for \( u' \) and \( \nu' \) in terms of \( \Phi' \) as

\[
u' = \left[ \frac{k\Phi'}{1 - y^2} \right] / (y^2 - \bar{\phi}^2), \tag{11}
\]

\[
u' = \left[ i k \Phi' - i\bar{\phi}(1 - y^2)\frac{\partial \Phi'}{\partial y} \right] / (y^2 - \bar{\phi}^2). \tag{12}
\]

Eliminating \( u' \), \( \nu' \) and \( w' \) between (8)-(12), we obtain a single partial differential equation in \( \Phi' \), i.e.,

\[
\frac{(y^2 - \bar{\phi}^2)}{S} \left[ \frac{\partial^2}{\partial x^2} - \frac{1}{4} \left( \frac{\partial + 1}{\partial z} \right) \right] \Phi' + (y^2 - \bar{\phi}^2) \Phi'
\]

\[
= -i(y^2 - \bar{\phi}^2) \left( \frac{1}{\partial_x} + \frac{1}{\partial z} \right) Q', \tag{13}
\]

where

\[
\sigma = \frac{S}{\partial_x} \left( \frac{\partial}{\partial z} \right) S.
\]

If \( \phi \) is constant the \( y \) and \( z \) dependencies may be separated in (13). The resulting equations are just the usual equations of atmospheric tidal theory (Lindzen, 1967). However, in the present case where \( \bar{\phi} = \bar{\phi}(z) \), (13) is not separable. Therefore, to compute the response of the atmosphere to a specified forcing \( Q \), we write (13) in finite-difference form and solve for \( \Phi' \) numerically using the direct inversion technique suggested by Lindzen and Kuo (1969).

In this study we have used a grid with latitudinal resolution of approximately 2° in the tropics and vertical resolution of 1 km. In all cases discussed here the perturbation geopotential field is assumed to be symmetric about the equator.

3. Structure of the forced wave with zero mean zonal wind

In order to compute the response of the atmosphere to a specified heat input using (13) we must specify: 1) the zonal wavenumber \( k \), 2) the frequency \( \omega \), relative to the ground, 3) the damping rate \( D \), 4) the static stability \( S(z) \), 5) the heating pattern \( Q'(y,z) \), and 6) the mean zonal wind \( \bar{u}(z) \). Since the main object of this study is to examine the influence of mean wind shear on the structure of the waves, we choose values for all parameters which are appropriate for the observed equatorial waves and examine the effect of varying only \( \bar{u}(z) \). Thus, in the case discussed in this Section and the three cases treated in Section 4, all parameters have identical values except the mean zonal wind profile. The values chosen for the parameters of the model are as follows:

1) \( k = 10 \), corresponding to a zonal wavelength of \( \sim 4000 \) km.

2) \( \omega = 0.1 \), corresponding to westward propagation with a period of 5 days relative to the ground.

3) \( D = 0.03 \), corresponding to a damping time of \( \sim 20 \) days.

4) \( S = \begin{cases} 0.01, & z < 16 \text{ km} \\ 0.025, & z > 16 \text{ km} \end{cases} \)

corresponding to a lapse rate of \( 6 \)C km\(^{-1}\) in the troposphere \((z < 16 \text{ km})\) and an isothermal stratosphere.

5) \( Q'(y,z) = Q_0 \exp \left[ -\left( \frac{y - y_0}{\alpha} \right)^2 \right] \)

\[
= \begin{cases} \exp^{-z/12 - e^{-z}}, & z < 12 \text{ km} \\ 7 - z/2, & 12 \leq z \leq 14 \text{ km} \\ 0, & z > 14 \text{ km} \end{cases}
\]

In the above \( z \) is dimensional and \( Q_0 \) a constant chosen so that the vertically integrated heating rate at latitude
\( y_b \) is \( \sim 4 \text{C day}^{-1} \). (This corresponds to a rainfall rate of \( \sim 2 \text{ cm day}^{-1} \).) In the cases to be discussed here we have let \( y_b = 0.15 \) and \( \alpha = 0.05 \), corresponding to a lateral distribution of heating with the maximum at 9N latitude and a half-width of about 3° latitude. Above 900 mb the vertical distribution of condensation heating specified here is a close approximation to the heating function derived by Kuo (1965) to represent the largescale heating of the atmosphere by deep cumulus convection. We have, however, added a cooling term which decays exponentially away from the boundary. This surface cooling which is clearly shown in the spectral results of Chang et al. is thought to be primarily due to descent of low equivalent potential temperature air in downdrafts occurring in the rain areas (Zipser, 1969). Such surface cooling appears to be important for the energetics of equatorial waves.

Figs. 2–5 show the computed atmospheric response for values of the parameters specified above for Case I where the mean zonal flow is set to zero. In Fig. 2 the

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**Fig. 2.** Amplitude (nondimensional) of the diabatic heating \( Q \), geopotential perturbation \( \Phi \), and meridional velocity \( v \) vs latitude at a height of 13 km for Case I (zero mean zonal flow). Unit amplitude corresponds to a heating rate of 4°C day\(^{-1}\), a geopotential height of 10 m\(^2\) sec\(^{-2}\), and a velocity of 10 m sec\(^{-1}\).

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**Fig. 3.** Meridional cross section of the amplitudes (°C day\(^{-1}\)) of diabatic heating [heavy lines, shaded region] and the meridional velocity (m sec\(^{-1}\)) [light lines] for Case I.
Fig. 4. Vertical profiles of amplitude and phase of meridional velocity v, temperature perturbation T, and vertical velocity w for Case I. Unit amplitude corresponds to \( v = 10 \text{ m sec}^{-1} \), \( w = 2 \text{ cm sec}^{-1} \), \( T = 1 \text{C} \). The phase is shown relative to the heat source.

The amplitudes of the perturbation geopotential and meridional velocity fields as a function of latitude are compared with the latitudinal distribution of the heat source. It can easily be seen that the half-width of the geopotential perturbation is nearly double that of the source, but that the perturbation decays uniformly away from the source in both directions. Solutions (not shown) for a narrower source and without the constraint of symmetry in \( \Phi \) about the equator are qualitatively similar to that shown. In particular, there is no evidence of either a symmetric or antisymmetric induced perturbation in the Southern Hemisphere which should be present if the wave were simply one of the pure equatorial \( \beta \)-plane modes discussed by Matsuno (1966).

A meridional plane cross section of the distributions of amplitude for the heat source and the perturbation meridional velocity is shown in Fig. 3. Note that the atmospheric response is characterized by maxima in \( v \) in the upper and lower troposphere with a broad region of small response in the mid-troposphere. If the surface cooling term is omitted, the upper tropospheric response turns out to be unaltered, but the response in the lower troposphere is considerably reduced, and the maximum occurs right at the ground.

In Fig. 4 the amplitudes and phases of \( v \), \( w \) and \( T \) are plotted vs height for latitude 9N. The heat source \( Q \) is not plotted here because between 1 and 13 km its profile is virtually identical to that of \( w \). Referring back to the thermodynamic energy equation (9) we see that the approximate proportionality between \( w \) and \( Q \) implies that to the first order adiabatic cooling must balance the condensation heating. Indeed, the computed local temperature change is nearly an order of magnitude smaller than the diabatic heating. It is interesting to note that in this case the heating and meridional velocity fields are nearly out of phase in the lower troposphere so that the maximum heating occurs in the northerly flow to the west of the trough. While this is contrary to the equatorial wave model of Riehl (1948) in which the precipitation occurs east of the surface trough, it is consistent with Reed's (1970) analysis of waves in the western Pacific.

Fig. 5. Geopotential height and velocity fields at 200 mb for Case I.
It may be easily verified that above the low-level maximum in $v$ the amplitude and phase of the temperature oscillation are qualitatively in agreement with the thermal wind relationship. Thus, the maximum positive temperature perturbations are coincident with the upper level ridge where it is increasing in intensity with height, and the maximum negative temperature perturbations (neglecting the surface layer) are coincident with the upper level ridge above the height of maximum intensity. However, it should be stressed that the temperature oscillation is everywhere less than $1^\circ C$ in amplitude.

The relationship of the perturbation geopotential and velocity fields on an isobaric surface for this case is shown in Fig. 5. It appears that north of $\sim 12^N$ the wind is in approximate geostrophic equilibrium. However, equatorward of that latitude, there is no simple relationship between the wind and pressure field. In fact, it appears that the maximum vorticity actually occurs equatorward of the trough near $9^N$ which is the latitude of maximum heating.

The horizontal divergence pattern for this case is not explicitly shown. However, the vertical profile of divergence can be estimated from the profile of $w$ in Fig. 3. Thus, there is a deep region of convergence into the cloud cluster region from the surface to $\sim 11$ km, topped by a region of strong divergence from 11 to 14 km. This pattern is in reasonable agreement with the observational results of Williams (1970) who used data composited from many tropical Pacific cloud clusters to derive an average divergence profile. The upper level outflow region is, however, somewhat deeper than that computed here. This is perhaps an indication that the level of maximum heating in the present model is slightly too high since, as pointed out above, the vertical profiles of $Q$ and $w$ are virtually identical.

4. Structure of the forced waves in the presence of vertical shear of the mean zonal wind

In the first section we suggested that the observational evidence indicates that vertical shear of the mean zonal wind may profoundly affect the structure of the atmospheric response to a given heating pattern. In this section we present results of three solutions with
rather different profiles of $\mathbf{u}(z)$, which are, to some extent, representative of the types of profiles found in the central and western Pacific.

Case II, shown in Figs. 6 and 7, is typical of the easterly trade wind regime with strong easterlies near the surface and westerly shear with height throughout the lower and middle troposphere. It should be emphasized that with the exception of $\mathbf{u}$, all parameters are the same as in Section 3. It is apparent from Fig. 6 that one effect of the westerly shear is to shift the lower tropospheric perturbation equatorward, and the upper tropospheric perturbation poleward. For this reason, as shown in Fig. 7, the vertical profile of $v$ at a given station will depend on the latitude of the station. For stations poleward of 9N the strongest response is at the upper levels, while for stations equatorward of 9N it is in the lower levels. The phase of the perturbation, however, does not depend much on latitude. In this case, the heating maximum occurs in the southerly flow to the east of the surface trough. There is also an indication of eastward phase tilt with height. These features are both in agreement with Riehl's easterly wave model. Thus, it appears the westerly shear of the mean zonal wind is necessary to produce the "classical" easterly wave.

This hypothesis is supported by Case III shown in Figs. 8 and 9. In this case there is easterly shear of the mean zonal wind from the surface to 12 km. This profile is typical of summer conditions in the western equatorial Pacific. The response in the lower troposphere is enhanced and the response in the upper troposphere is shifted equatorward and diminished in amplitude. As in Case I, the heat source maximum lies to the west of the trough at the surface. However, the trough axis tilts westward with height sufficiently so that the heating maximum is at the center of the trough at 4 km and east of the trough above that level. Qualitatively, this solution is remarkably similar to the structure of the observed waves in the western Pacific reported by Reed (1970).

As a final example (Case IV) in Figs. 10 and 11 we show the solution for a situation in which there is a broad region in the lower troposphere where the zonal wind speed is faster than the phase speed of the waves. In this region (bracketed by two "critical levels" where the phase speed and mean zonal wind speed are equal), the waves are actually moving slowly eastward relative to the wind. The atmospheric response in this case is much stronger in the upper troposphere than it was in
the previous two cases. The general amplitude pattern is quite similar to the structure of the $\tilde{u}=0$ case (Figs. 3 and 4). However, the phase of the response in the lower troposphere is quite different. In the present case the heating maximum is east of the trough, whereas in the $\tilde{u}=0$ case it was west of the trough. A simple physical interpretation of these differences is certainly not obvious.

5. Summary and conclusions

In this study we have presented evidence that the westward propagating waves of the equatorial Pacific can be interpreted as Rossby waves driven by the latent heat release due to precipitation in the cloud clusters embedded within the waves. We have shown that many of the observed structural features of these waves can be simulated in a simple diagnostic model based on the linearized primitive equations. In this model the heating pattern is specified as an external parameter and the response of the atmosphere to the heating is computed.

The most important conclusion from this study is that the structure of the atmospheric perturbation in response to a given heating pattern is very sensitive to the vertical shear of the zonal mean wind. This result is in good agreement with the analysis of observed waves in the equatorial Pacific by Reed. However, it should be pointed out that the zonal mean wind in this model represents a longitudinally averaged zonal wind, whereas the mean zonal winds shown in Fig. 1 are time averages of the zonal wind at different longitudes. Therefore, a detailed comparison of the model results with Reed's analysis is not warranted. The fact that there is reasonable qualitative agreement between the results of this model and observation is perhaps evidence that the waves rather quickly adjust to the local mean zonal wind at any longitude as they move westward.

It is also possible that lateral shear of the mean zonal wind may alter the structure of the waves. This possibility is currently being investigated with a similar diagnostic model and will be reported later.

Acknowledgments. I wish to thank my colleagues Profs. R. J. Reed and J. M. Wallace for many stimulating discussions and helpful suggestions. The research was supported by the Atmospheric Sciences Section, National Science Foundation, under Grant GA-23488.
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