Eddy Heat Fluxes and Stability of Planetary Waves. Part II

CHARLES A. LIN[1]

Canadian Climate Centre, Downsview, Ontario, Canada M3H 5T4

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ABSTRACT

The stability analysis of baroclinic Rossby waves in a zonal shear flow examined in Part I is applied to the atmosphere. The basic state consists of a planetary-scale (wavenumber 1) Rossby wave in a zonal flow with vertical shear close to the minimum critical shear of the two-level model. The most unstable mode grows at the baroclinic time scale and is almost stationary relative to the basic wave. Maximum perturbation amplitude is at wavenumber 3, intermediate between the response scales of the two destabilizing mechanisms—vertical shears of the zonal flow and basic wave. The perturbation meridional scale is close to the radius of deformation. Applications to planetary-scale waves that transport heat are considered.

1. Introduction

In Part 1 of this paper (Lin, 1979; hereafter referred to as I), we presented a parameter study of the stability of baroclinic Rossby waves in a zonal shear flow. The model used was a linear, quasi-geostrophic, two-level, adiabatic and frictionless midlatitude β-plane model. The nature of the most unstable modes was examined for a wide range of parameter values: the zonal scale of the basic wave relative to the radius of deformation, and the barotropic and baroclinic wave amplitudes relative to the amplitude of the zonal flow. In this paper, we will apply the stability analysis to the atmosphere. The primary motivation is to examine the midlatitude planetary scale (zonal wavenumbers 1, 2, 3) transient waves. Although the source of these eddies is unclear, observational studies do show they explain a significant portion of the total eddy variances and covariances (Julian *et al.*, 1970; Kao and Sagendorf, 1970; Willson, 1975). In particular, these eddies account for a major portion of the total meridional heat transport (Kao and Sagendorf, 1970), i.e., they are baroclinic in nature. We are primarily concerned here with these baroclinic, planetary-scale transient waves.

One possible source of planetary-scale transient waves is slowly growing, longwave modes, first discovered by Green (1960). These "Green modes" have been found in many studies of baroclinic instability of a zonal flow. Although Green modes have small growth rates, they can attain large amplitudes. Gall (1976), in a numerical study using a general circulation model, examined the baroclinic instability of realistic zonal wind profiles. He found that long, deep baroclinic waves do attain much greater amplitudes than short, shallow waves. This is because the stabilizing effect of nonlinear wave-mean flow interaction occurs most rapidly in low levels and thus short, shallow waves are affected more by nonlinear effects. Green modes, being long, deep waves, can thus grow to finite amplitude. These finite-amplitude waves will affect the nature of the baroclinic stability problem. Thus, it is of great interest to examine the stability of Green modes to further perturbations.

The two-level model cannot resolve weakly unstable Green modes. But we may identify such modes with a neutral free mode of the two-level model: the baroclinic Rossby wave. This mode is characterized by the upper and lower waves being out of phase by 180°. Fullmer (1979)[2] examined the baroclinic instability of one-dimensional basic states using a β-plane quasi-geostrophic model. His model had 48 levels in the vertical with the troposphere extending from 1000 to 250 mb and the stratosphere from 250 to 0 mb. The upper boundary condition used is vanishing perturbation streamfunction amplitude at the top of the model atmosphere. Green modes were found using a basic state with the static stability of the stratosphere 50 times that of the troposphere and a linear shear zonal wind which vanished at the ground and reached 24 m s^{-1} at 250 mb. The ratio of static stabilities of the strato-

[1] Present affiliation: Department of Physics, University of Toronto, Toronto, Ontario, Canada M5S 1A7.

[2] Fullmer, J. W. A., 1979: The baroclinic instability of simple and highly structured one-dimensional basic states. Ph.D. thesis, MIT, 235 pp.

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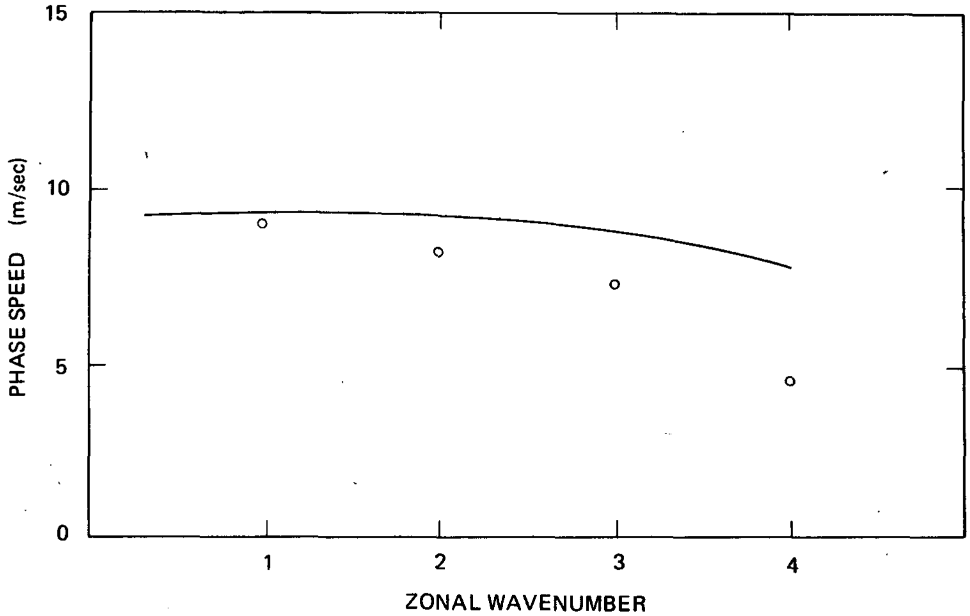

FIG. 1. Comparison of phase speeds of Green modes obtained with linear vertical shear of basic zonal wind and stratospheric static stability 50 times that of the troposphere (circles; from Fullmer, 1979), and phase speeds of neutral baroclinic Rossby wave in two-level model with comparable basic flow parameters.

sphere and troposphere is realistic for midlatitude winter conditions; the zonal wind profile is also realistic in the troposphere. The doubling time of these modes is over 10 days. Fig. 1 shows the phase speed of these modes as a function of zonal wavenumber, together with the phase speed of the baroclinic Rossby wave in the two-level model with the corresponding values of tropospheric static stability and barotropic and baroclinic zonal flow components. We see that the phase speeds agree well for long wavelengths. The vertical structure of the streamfunction phase for the fastest growing Green mode is shown in Fig. 2. There is an abrupt phase shift of approximately π radians at 650 mb, much like the baroclinic Rossby wave which has the upper and lower level streamfunctions out of phase with each other by 180°. The good agreement of the phase speed and vertical structure of the Green mode in a high vertical resolution model and the baroclinic Rossby wave in a two-level model suggests that the latter can be identified with a Green mode. Also, Green modes, being slowly growing, are inefficient at transporting heat. In the two-level model, the baroclinic Rossby wave transports no heat at all. Thus from the point of view of planetary-scale transient waves which transport heat, the stability problem of Green modes is of great interest.

In addition to providing insight to the nature of baroclinic transient eddies (TE's), our study is also of interest to planetary-scale stationary eddies (SE's). Stone (1977) observed that these waves are highly baroclinic in nature and suggested planetary scale TE's and SE's might be two manifestations of the same phenomenon—eddies generated by instability of a baroclinic flow with external forcing. Such eddies would be baroclinic in nature and planetary scale in size, and could well contain both TE and SE components. The external forcing will force SE's and the resulting zonal flow and SE field will be unstable to small wavelike perturbations. As the basic flow is non-axisymmetric, interaction between the basic wave and the perturbation will generate further waves, i.e., a spectrum of waves is realized. Since the baroclinic Rossby wave is a free wave and has to satisfy a dispersion relation, it cannot be identified with a true, forced stationary wave. However, the vertical structure of stationary wavenumber 1 has considerable tilt with height; the results of Muench (1965) and van Loon *et al.* (1973) show that stationary wavenumber 1 has a phase shift of almost 180° in the troposphere. Thus its vertical structure is similar to that of the baroclinic Rossby wave. Moreover, the stability problems of a free Rossby wave and a topographically forced wave in a two-level model are similar in nature (Glenn Flierl, personal communication). The Rossby wave is governed by a dispersion relation which relates the barotropic and baroclinic components of the zonal flow, the wavenumber, and the ratio of upper and lower level wave amplitudes. In order to avoid a resonant response, we need only move away from the free wave curve in parameter space. The topographic forcing, being an inhomogeneous forcing, does not appear in the linearized perturbation equations. Thus the stability problems of a free Rossby wave and a topographically forced stationary wave differ only in the basic states. The scale selection mechanism still operates in both cases to generate a spectrum of waves. Thus, our stability study is also of interest to the hypothesis formulated by Stone.

2. Model and application to atmosphere

The model used is a linear, quasi-geotrophic, two-level, adiabatic and frictionless midlatitude β-plane model. We present an outline of the model here, and further details can be found in I. The governing equations in pressure coordinates are

$$\frac{\partial}{\partial t}\nabla^2\Psi + J(\Psi, \nabla^2\Psi + f) + J(\Psi_T, \nabla^2\Psi_T) = 0, \quad (1)$$

$$\frac{\partial}{\partial t}(\nabla^2 - \mu^2)\Psi_T + J(\Psi, (\nabla^2 - \mu^2)\Psi_T) + J(\Psi_T, \nabla^2\Psi + f) = 0, \quad (2)$$

where $\Psi = \frac{1}{2}(\Psi_1 + \Psi_3)$ and $\Psi_T = \frac{1}{2}(\Psi_1 - \Psi_3)$ denote the mean and thermal streamfunctions, respectively; quantities with subscripts 0, 1, 2, 3, 4 denote values at 0, 250, 500, 750, 1000 mb, respectively; $\mu^2 = 2f_0^2/(R\sigma_2 p_2)$, where f_0, R, σ_2, p_2 are the Coriolis parameter at 45°N, ideal gas constant, and the static stability at level 2 and 500 mb, respectively; μ^{-1} is the radius of deformation, t is time, ∇^2 and J denote the horizontal Laplacian and Jacobian, respectively. The Coriolis parameter is assumed to have a constant meridional gradient β.

Solutions to (1) and (2) exist of the form

$$\Psi = -U_y y + B \sin k(x - ct), \quad (3)$$

$$\Psi_T = -U_T y + B_T \sin k(x - ct). \quad (4)$$

This solution describes zonally propagating Rossby waves of wavenumber k and phase speed c, in a zonal flow with barotropic and baroclinic components U and U_T, respectively. It is an exact solution provided the dispersion relation relating the four nondimensional quantities $\mu^2(U - c)/\beta$, $\mu^2 U_T/\beta$, $K = k/\mu$ and B_T/B are satisfied. These quantities describe the barotropic zonal flow relative to the basic wave phase speed normalized by β/μ^2, the baroclinic zonal flow normalized by β/μ^2, the ratio of the radius of deformation to basic wave scale and the ratio of baroclinic to barotropic basic wave amplitudes, respectively. Specifying any two of these four quantities determine the other two, with an overall wave amplitude being arbitrary. For the Rossby waves to be non-growing, c must be real. This, in turn, requires the zonal shear not to exceed the "critical shear" of the two-level model. The minimum critical shear occurs when $K^2 = 2^{-1/2}$ and has the value $U_T = \beta/\mu^2$.

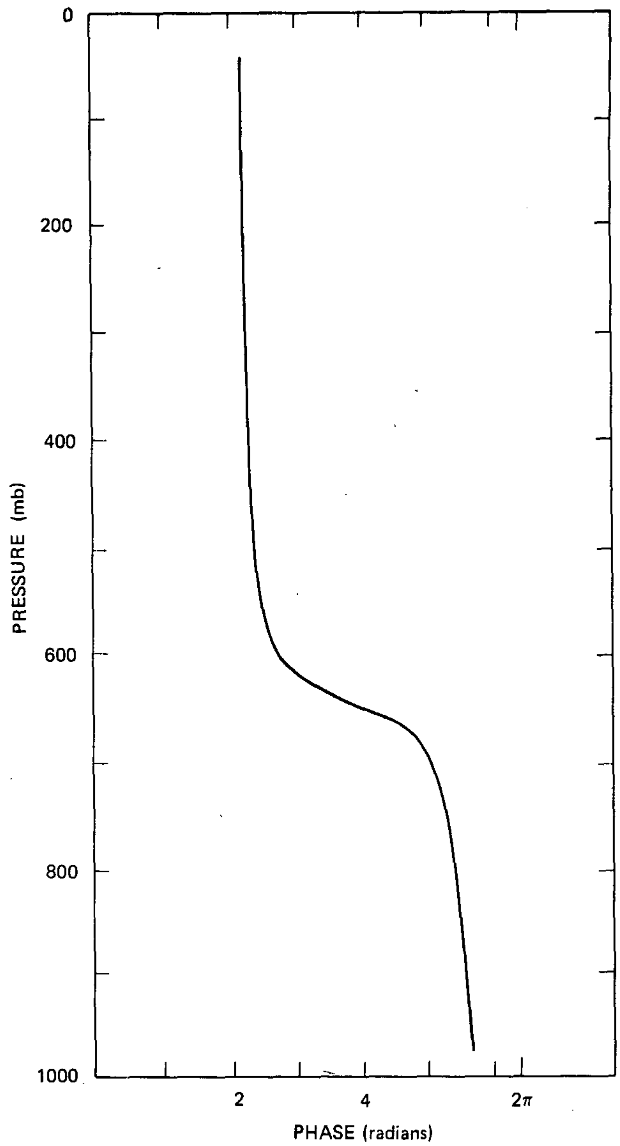

FIG. 2. Vertical structure of streamfunction phase for fastest growing Green mode. Basic flow parameters as in Fig. 1. (from Fullmer, 1979).

The basic flow whose stability we wish to investigate is described by (3) and (4). The basic streamfunctions are independent of time in a coordinate system moving with the basic wave. Thus we let $x_0 = x - ct$ and treat x_0, y, t as the independent variables. We impose small perturbations on this flow by linearizing (1) and (2). Perturbation solutions of the linearized equations exist of the form

$$\Psi' = \sum_n X_n \exp[i(nkx_0 + ly + \sigma t)], \quad (5)$$

$$\Psi_T' = \sum_n Y_n \exp[i(nkx_0 + ly + \sigma t)]. \quad (6)$$

The zonal structure of the perturbations consists of Fourier harmonics of the basic wavenumber; the perturbations are thus cyclic in the zonal direction with the basic wavelength as period. The perturbations are also characterized by a meridional wavenumber l; we will allow a continuous spectrum of l values. Instability exists if the frequency σ has a negative imaginary part. Substitution of (5) and (6) in the linearized equations yields the eigenvalue equations for σ. These equations in nondimensional form are

$$(\lambda + a_n)X_n + c_{n-1}X_{n-1} + c'_{n-1}X_{n+1} + b_nY_n + s_{n-1}Y_{n-1} + s'_{n+1}Y_{n+1} = 0, \quad (7)$$

$$(\lambda + e_n)Y_n + g_{n-1}Y_{n-1} + g'_{n+1}Y_{n+1} + f_nX_n + t_{n-1}X_{n-1} + t'_{n+1}X_{n+1} = 0, \quad (8)$$

where $\lambda = \sigma/[k(U - c)]$, the nondimensional frequency, is the ratio of the phase speed of the perturbation component with the basic wave scale, to the barotropic zonal flow relative to the basic wave. The nondimensional coefficients, a_n, c_n, c_n', b_n, s_n, s_n', e_n, g_n, g_n', f_n, t_n, t_n', are given in I. Other nondimensional parameters of interest are μ/k, the ratio of basic wave zonal scale to radius of deformation; $B_Tk/U - c$, $Bk/U - c$, the velocity amplitudes of baroclinic and barotropic basic wave normalized by the barotropic zonal flow relative to the basic wave; and $L = l/k$, the basic wave zonal scale relative to the perturbation meridional scale. Three basic-state parameters can be specified independently. The stability problem is then solved by finding the nondimensional eigenvalue λ as a function of the nondimensional meridional wavenumber L. This is done by truncating the integral values of n in (7) and (8) to $|n| \leq N$. When N is so large that λ has converged, a good approximation to the "true" solution will have been obtained.

We showed in I that there are two important zonal scales: the basic wave scale and the radius of deformation. The former is important because it is present as an explicit scale while the latter is the natural response scale of a perturbed baroclinic zonal flow. We identified different regimes of instability in parameter space where the dominant energy sources for instabilities are the horizontal or vertical shears of the basic wave, or the vertical shear of the zonal flow.

In the case of interest to the atmosphere, the basic wave will be a planetary-scale Green mode, or a baroclinic Rossby wave in our model. Two-layer models of baroclinic instability of a zonal flow predict that there is a minimum critical zonal shear separating conditions which are stable from those which are baroclinically unstable. This minimum critical shear may be translated to a critical meridional temperature gradient with the aid of the thermal wind equation. Stone (1978) has shown that the observed mean tropospheric temperature gradients coincide closely with this critical temperature gradient in mid and high latitudes in all seasons. His results suggest the zonal shear should be taken to be close to its critical value $\mu^2U_T/\beta = 1$. For typical winter conditions, this corresponds to $U_T = 5.6$ m s^{-1}, and a realistic value of the barotropic component of the zonal flow is $U = 15$ m s^{-1}. Nondimensionally, this gives $\mu^2U/\beta = 2.7$. The dispersion relation for the baroclinic Rossby mode then requires wavenumbers 1 and 2 to propagate eastward at ~10 m s^{-1}. This phase velocity is typical of Green modes. Having specified the basic wavenumber and the zonal flow components, there remains an additional parameter—the amplitude of the basic wave. Wavenumber spectra of travelling planetary scale waves indicate that peak power is at wavenumbers 1–2 with period of about 15–30 days (Arai, 1973; Deland, 1973). The geopotential amplitude at such scales is ~60 m (Arai, 1973). We will take as the basic wave amplitude $B_Tk/U - c = 0.5$. For wavenumber 1, this corresponds to a geopotential amplitude of 140 m, and 70 m for wavenumber 2. The amplitude for wavenumber 1 is large, but time series of geopotential amplitude do show instances when wavenumber 1 attains an amplitude close to 140 m in winter (Arai, 1970). For comparison, typical geopotential amplitudes for planetary-scale stationary waves in winter are about 50–150 m in the troposphere (Muench, 1965; van Loon *et al.*, 1973). For basic wavenumbers 1 and 2, the parameter values $\mu^2U_T/\beta = 1$, $\mu^2U/\beta = 2.7$ and $B_Tk/U - c = 0.5$ mean that the dominant energy sources of instability are the vertical shears of the basic wave and zonal flow. According to the classification presented in I, these parameter values are located intermediate between the Phillips and Kim regimes in parameter space.

3. Results

In Fig. 3, we show the nondimensional growth rate $|\sigma_ik^{-1}/(U - c)|$ of the most unstable mode for basic wavenumber 1 as a function of wave amplitude $B_Tk/(U - c)$ and meridional wavenumber l/k, for $N = 10$, when convergence has been obtained. The meridional wavelength of maximum instability corresponds to $l/k \approx 5$, i.e., a meridional scale close to the radius of deformation. Modes with small values of l/k are nonpropagating relative to the basic wave, while those with larger values are propagating.

For the wave amplitude $B_Tk/(U - c) = 0.5$, the structures and energetics of all the unstable modes with $l/k = 5$, the meridional wavenumber of maximum instability, are shown in Fig. 4. The fastest growing mode is of planetary scale, with kinetic

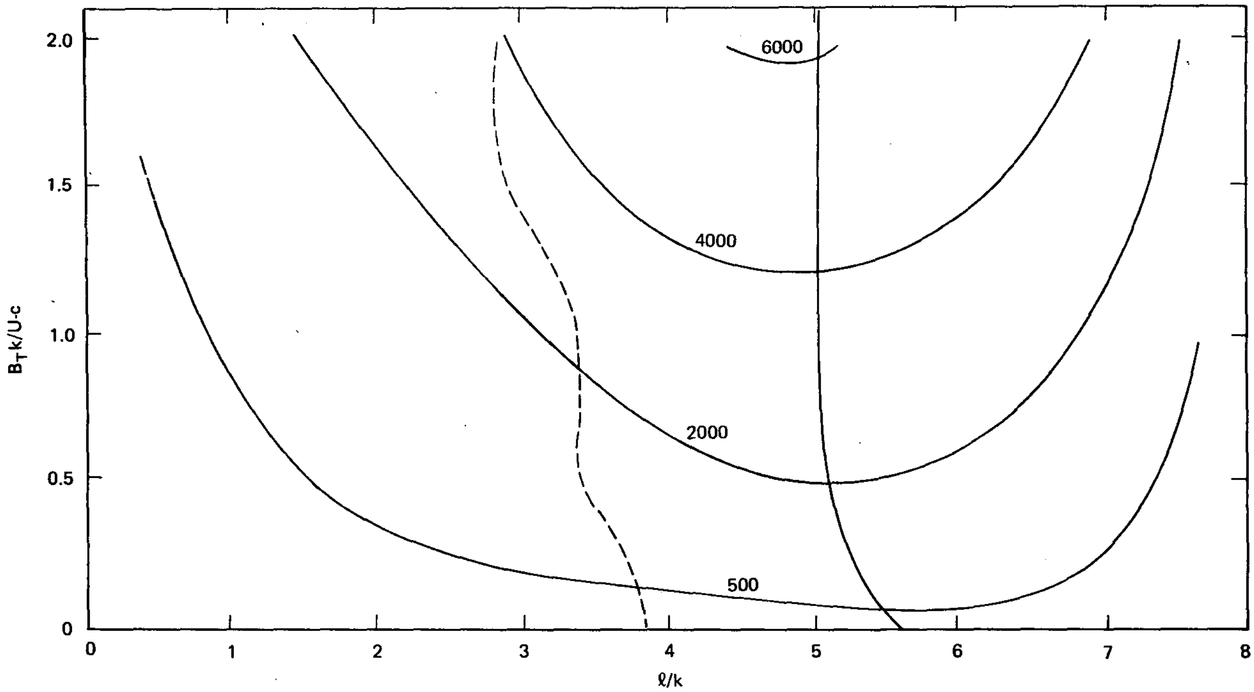

FIG. 3. Growth rate ($|\sigma_i/k(U-c)| \times 10^3$) of most unstable mode as function of wave amplitude ($B_T k/(U-c)$) and meridional wavenumber (l/k) for $N = 10$. Propagating modes (relative to basic wave) exist to right of dashed line. $\mu^2 U_T/\beta = 1$, $k/\mu = 0.13$.

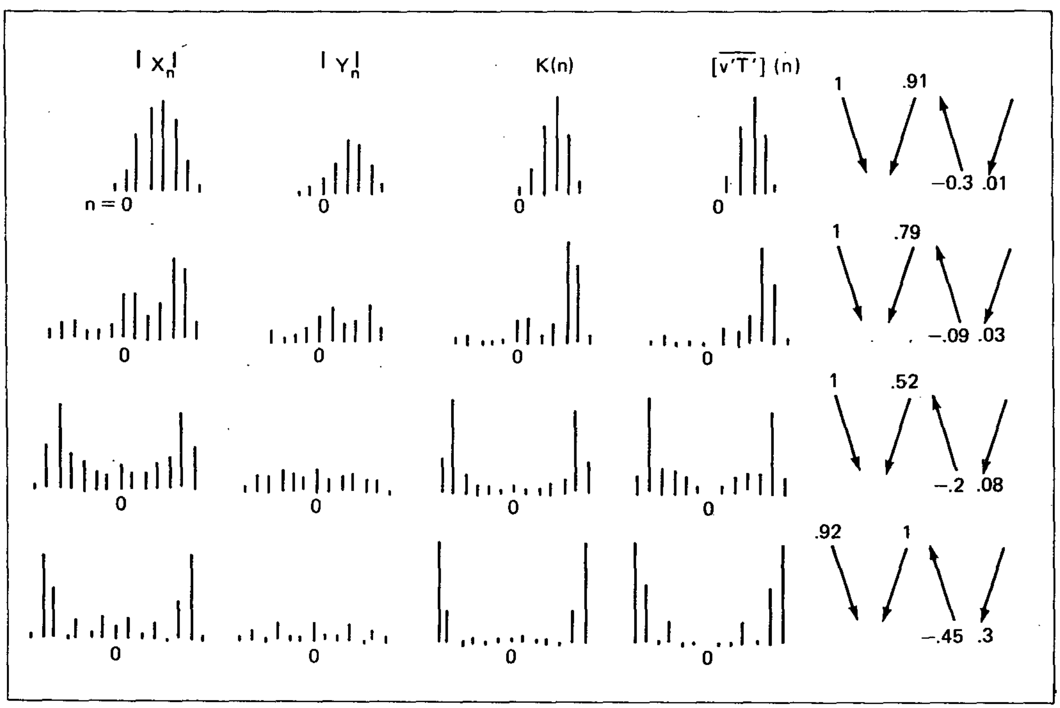

FIG. 4. Structures and energetics of most unstable modes with $l/k = 5$, $B_T k/(U-c) = 0.5$, $\mu^2 U_T/\beta = 1$, $k/\mu = 0.13$. Wavenumber spectra of perturbation barotropic component ($|X_n|$), baroclinic component ($|Y_n|$), kinetic energy ($K(n)$), and heat transport ($[\overline{v'T'}](n)$) are shown. Two left arrows of energetics show conversions of APE of zonal flow and basic wave to perturbation APE; two right arrows show conversions of KE of upper and lower level basic wave to perturbation KE; all for positive values of conversions. Growth rates increase upward.

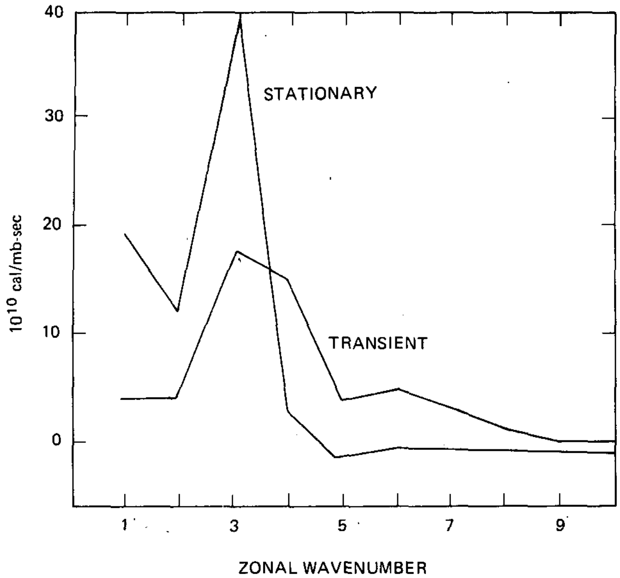

FIG. 5. Observed wavenumber spectra of meridional sensible heat transport by stationary and transient eddies at 850 mb, 60°N in winter (from Kao and Sagendorf, 1970).

energy maximum at $n_0 = 3$. It propagates relative to the basic wave as the spectra are not symmetric about $n = 0$. Examination of the real part of the dimensional eigenvalue (σ_n) shows that the ratio of σ_n/n_0k to c is much less than unity, i.e., the dominant unstable perturbation component is almost stationary relative to the basic wave. This is true of all the unstable modes shown in Fig. 4. Relative to the basic wave, the unstable modes have decreasing phase speeds with decreasing growth rates and the dominant zonal scale shifts toward the radius of deformation. The latter scale corresponds to about wavenumber 6 with our parameter values.

For the most unstable mode, the nondimensional eigenvalue gives an e-folding time of 4.8 days and as remarked earlier, a phase velocity of the dominant component close to that of the basic wave. This represents a disturbance which grows at the baroclinic time scale and propagates at around 10 m s^{-1}. Fig. 5 shows the wavenumber spectra of meridional sensible heat flux for stationary and transient eddies, at 850 mb, 60°N in winter. The data are taken from Kao and Sagendorf (1970). Planetary scales dominate both spectra, with peaks at wavenumber 3 for both cases. The agreement between the observed heat transport spectra and the spectra of the most unstable mode we obtained is fairly good, given the two-level approximation of the model.

Energetics of the unstable modes show that conversions of available potential energy (APE) of the zonal flow and basic wave to eddy APE are the dominant interactions. Thus baroclinic eddy-eddy interaction is comparable to baroclinic eddy-mean flow interaction, while barotropic interactions are small. From the point of view of stationary waves, evidence of this eddy-eddy interaction is found in Lau (1979). In an observational study of tropospheric stationary waves, he found that heat transport by transient eddies acts as a dissipative mechanism of the zonally asymmetric component of the mean temperature field. As the latter is a measure of the APE of stationary waves, this means there is a strong conversion of stationary eddy APE to transient eddy APE.

We have seen the most unstable mode grows at the expense of APE of the zonal flow and basic wave. In the classification terminology of I, this means baroclinic instability mechanisms of the Phillips and Kim regions are dominant. Baroclinic instability describes the release of APE of horizontal temperature gradients associated with vertical shears of horizontal currents. Energy release is accomplished by motion in the direction parallel to the horizontal temperature gradient. In the Phillips regime, characterized by a dominant zonal flow with vertical shear, the associated meridional temperature gradient means that APE of the zonal flow is released by motion in the meridional direction. The most unstable mode has the radius of deformation (L_r) as its zonal scale and has infinite meridional extent (Phillips, 1954; I). Thus the horizontal wavevector of the perturbation $\mathbf{K} = (k,l) = (L_r^{-1},0)$, where k and l denote the zonal and meridional wavenumbers respectively. In the Kim regime, characterized by a strong meridional current with vertical shear, the associated temperature gradient is in the zonal direction, and release of APE is accomplished by motion in the zonal direction. Barotropic interactions are unimportant for a planetary-scale basic wave (I). In this case, the most unstable mode has a meridional scale close to L_r and in the zonal direction, the perturbation zonal component has the largest amplitude (Kim 1978; I). In other words, $\mathbf{K} = (\sim 0, L_r^{-1})$. In both the Phillips and Kim regimes, the perturbation horizontal wavevector $\mathbf{K}$ aligns itself with the basic current. In our case, the basic wave is a planetary-scale baroclinic Rossby wave with the zonal flow near minimum critical shear. We found realistic kinetic energy and heat transport spectra which peak at wavenumber 3 for the most unstable mode; the response is thus maximum at a zonal scale intermediate between those of the Phillips and Kim regimes. The meridional scale remains at L_r. The perturbation wavevector is no longer along the x or y directions, but is now oriented at approximately 60° to the horizontal for the dominant perturbation component. This is because the basic flow now consists of zonal and meridional currents of comparable magnitudes. This orientation of the perturbation wavevector optimizes release of APE of the basic flow.

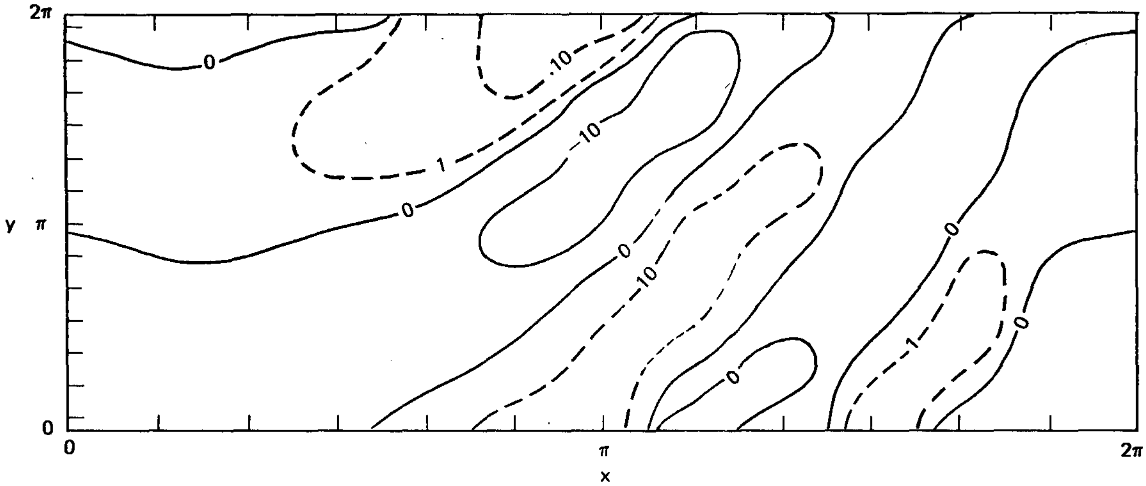

FIG. 6. Amplitude (undetermined to within an arbitrary multiplicative constant) of the upper level perturbation streamfunction at $t = 0$, over one cycle of the basic wave in the zonal and meridional directions.

Our result that the most unstable mode has a zonal and meridional scale both close to the radius of deformation is important for closure assumptions used in parameterizations of heat transport by large-scale eddies (Green, 1970; Stone, 1972; Held, 1978). Green (1970) assumed the parameter dependence for the ratio of meridional to zonal scale to be the planetary scale divided by the radius of deformation. Stone (1972) and Held (1978) implicitly assumed this ratio to be order unity. If the scale selection mechanism in our study is appropriate, our results suggest that Stone and Held's assumption is more valid.

Frederiksen (1979) examined the effects of planetary-scale waves on regions of cyclogenesis using a linear, spherical, two-layer quasi-geostrophic model. The basic state consists of a jet at 30°N and a zonal wavenumber 3 planetary wave in the upper layer. There is no motion in the lower layer. Unspecified external forcing is required to maintain the basic state as it does not satisfy the conservative equations of motion; the effect of the forcing on the perturbations is neglected. He found the presence of the upper layer wave causes little change in the growth rate and phase speed spectra compared to those obtained with only the jet in the basic state. The dominant perturbation wavenumber is wavenumber 7, the same as that of the most unstable mode with the jet only. However, there is a modulation of the wavenumber 7 structure by the basic wavenumber 3. Maximum development took place close to the positions of maximum excess shear, where the difference between the vertical zonal shear and that corresponding to Phillips' (1954) stability criterion of the two-level model is a maximum. The zonal shear between the upper and lower layers due to the jet greatly exceeds that due to the upper layer wave: at 30°N; the former is ~ 40 m s^{-1} while the latter is 8 m s^{-1} (Fig. 2 of Frederiksen). This accounts for the dominance of the jet in the growth rate and phase speed spectra. In our case, the zonal shear is close to the minimum critical shear of the two-level model, which is about 12 m s^{-1} at mid-latitudes. Without the presence of the basic Rossby wave, our basic state would thus be neutral, unlike Frederiksen's case. Also, the amplitude of our basic Rossby wave is comparable to the zonal flow.

We show in Fig. 6 the amplitude of the upper level perturbation streamfunction at $t = 0$, over one cycle of the basic wave in the zonal and meridional directions. Maximum development occurs at the center of the wave domain. Our basic wave represents a meridional current which varies cosinusoidally in the zonal direction as its streamfunction is sinusoidal in x only. Thus the center of the wave domain is a position of maximum meridional shear. As meridional currents corresponding to barotropic or baroclinic Rossby waves are unstable irrespective of their amplitude (Lorenz, 1972; Kim, 1978), there is no nonzero minimum critical shear. Thus, in our case, maximum development occurs at a position where the excess meridional shear is a maximum.

4. Conclusions

We have examined the stability of baroclinic Rossby waves in a baroclinic zonal flow. A parameter study of this problem was presented in I. In this paper, we have applied the stability analysis to the atmosphere: a planetary-scale basic wave (wavenumber 1) in a zonal flow near the minimum critical shear of the two-level model. This study is motivated by an attempt to better understand planetary-scale transient waves that transport heat. Whatever their source, these waves are observed to account for a significant portion of the total transient eddy variances and covariances. Thus, their stability properties are of great interest. We found the resulting most unstable mode consists of a spectrum of waves, with maximum amplitude at wavenumber 3. The response is thus maximum at a zonal scale intermediate between the basic wave scale and the radius of deformation. The latter are characteristic scales of the two destabilizing mechanisms—the vertical shears of the basic wave and the zonal flow. As the magnitudes of the latter are comparable, the response lies between the two corresponding unstable regimes. The most unstable mode grows at the baroclinic time scale and is almost stationary relative to the basic wave. Maximum eddy activity takes place at a position where the vertical shear of the basic wave is largest.

Comparison of the heat transport spectrum of the most unstable mode and the observed spectra of planetary-scale waves shows good agreement. Both the calculated and observed spectra peak at wavenumber 3. We have shown the planetary-scale, neutral baroclinic Rossby wave of the two-level model can be identified with a weakly unstable Green mode found in more refined models. This suggests a plausible source of baroclinic, planetary-scale transient waves: they result from the baroclinic instability of finite-amplitude Green modes. The Green modes, being slowly growing, are inefficient at transporting heat; but once they reach finite amplitude, they give way to baroclinically unstable modes that transport heat. As planetary-scale transient eddies are observed to account for a substantial portion of the total heat transport, this is a plausible source of their presence. Our results also lend support to the hypothesis of Stone (1977) on the generation of efficient heat transporting stationary waves of the midlatitude winter atmosphere. He suggested they are due to instability of a baroclinic flow with external forcing. Our study shows that a finite-amplitude, stationary, planetary-scale Rossby wave, which does not transport heat, can give rise to a heat transporting, unstable mode which is almost stationary, and which has a dominant wavenumber 3 component. Baroclinic instability of the stationary wave and zonal flow are the principal energy sources for the unstable mode. The stationary Rossby wave, being a free wave, cannot be identified with a true, forced stationary wave. However, as we discussed earlier, there are similarities between the two cases. Thus we anticipate destabilizing and scale selection mechanisms similar to those present in our study will also operate, when the basic wave is a forced wave. Our results thus motivate further study of the stability of forced winter stationary waves.

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