Baroclinic Adjustment

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ABSTRACT

Two-layer models of baroclinic instability predict that there is a critical temperature gradient separating conditions which are stable from those which are baroclinically unstable. In continuous models this critical gradient corresponds to a transition from conditions where the dominant baroclinic instabilities are inefficient at transporting heat to conditions where they are efficient. Zonal mean meridional temperature gradients in the atmosphere are compared with this critical gradient. For averages over periods longer than a few months the observed mid and mean tropospheric gradients never appreciably exceed the critical gradient. In fact they coincide remarkably closely with it in mid and high latitudes in all seasons in spite of strong seasonal changes in the forcing. This behavior shows that a very rapid transition must exist between conditions where eddy fluxes are inefficient to conditions where they are highly efficient. Thus, the primary effect of baroclinic eddies on the meridional temperature structure is to limit the gradients from becoming appreciably supercritical. This behavior allows one to take into account quite accurately the effect of the eddy fluxes on temperature structure without calculating the eddy fluxes explicitly, simply by adjusting the temperature gradients so that they never exceed the critical value. This baroclinic adjustment process is illustrated by incorporating it into a one-dimensional energy balance climate model. The results show that the process enhances the stability of the current climate to changes in the solar constant.

1. Introduction

The problem of parameterizing atmospheric eddy fluxes has received increasing attention in recent years because it offers the possibility of constructing simple models of the atmosphere for use in studying climate. The starting point for attempts to parameterize the eddy flux of heat has generally been baroclinic stability theory with β-effects neglected (Green, 1970; Stone, 1972). However, β-effects are clearly important in the dynamics of atmospheric eddies (Phillips, 1954), and attempts to include these effects in parameterizations have not been too successful (Stone, 1974).

The simplest model available for studying β-effects is the two-level model (Phillips, 1954). In this model there is a critical value of the vertical shear of the mean zonal flow corresponding to neutral stability. When the shear exceeds this critical value, the two-level model is baroclinically unstable, and when it does not the model is stable. This critical value of the shear can be expressed as a critical value of the meridional temperature gradient by applying the thermal wind relation. Analysis of the continuous analog of the two-level model (Green, 1960) shows that the physical significance of this critical value of the shear is actually somewhat different. Rather than separating unstable conditions from stable conditions, this critical shear represents a transition between "supercritical" conditions where the dominant unstable waves are long, deep waves and "subcritical" conditions where the dominant unstable waves are short, shallow waves. Held (1978) has shown with scaling arguments that one can expect the long, deep waves to be much more efficient at transporting heat than the short, shallow waves.

Phillips (1954) compared the actual shear in the atmosphere with the local value of the critical shear given by the two-layer model and found that in mid-latitudes the shear was 50% supercritical in winter, and approximately critical in summer. More recently, Moura and Stone (1976) made a similar comparison but using more up-to-date and much more complete data sets. They found that the annual mean shear was remarkably close to the local value of the critical shear throughout mid and high latitudes. Such a coincidence suggests that the local value of the critical shear may have special significance with respect to the effect of baroclinic eddy fluxes on local atmospheric structure in mid and high latitudes. However, the comparison by Moura and Stone made two approximations which may well affect the result—namely, it neglected latitudinal variations of the static stability in calculating the critical shear and

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it included the stratosphere in calculating the mean shear of the observed zonal flow.

In this paper we present a more careful and more detailed comparison of the actual shear in the atmosphere with the critical shear given by the two-layer model. The results will lead us to suggest a very simple parameterization of the effect of eddy fluxes on atmospheric temperature structure, a parameterization which automatically includes β-effects. We will then illustrate the parameterization by applying it in a one-dimensional heat-balance climate model.

2. Comparison of atmospheric temperature gradients with the critical gradient

According to Phillips' (1954) analysis the critical shear for the two layer model is

$$u_c=\frac{\beta R(\theta_1-\theta_3)}{f^2}, \tag{1}$$

where f is the Coriolis parameter, $\beta=df/dy$, y is the meridional coordinate, R the perfect gas constant, θ potential temperature, and the subscripts 1 and 3 refer to the upper and lower layers, respectively. In evaluating this expression we chose 500 mb as the reference pressure in the definition of θ. If u is the zonal wind, then the model predicts instability if $u_1-u_3>u_c$, and stability if $u_1-u_3\leqslant u_c$.

The multiplicative constant in Eq. (1) does depend on the particular pressure levels one chooses in defining the model's two layers. Therefore, the precise magnitude of the critical shear cannot be defined unambiguously. However, there is no similar ambiguity in the parameter dependences of u_c. Any other choice of reference levels would give the same dependences as those shown in Eq. (1). Consequently, we will be particularly interested in comparing variations in u_1-u_3 with variations in u_c.

Also we are primarily concerned with the temperature structure of the atmosphere. Therefore we will use the thermal wind relation to replace u_1-u_3 and u_c by corresponding meridional temperature gradients. The thermal wind relation for the two level model is

$$f\frac{u_1-u_3}{z_1-z_3}=-\frac{g}{T_2}\frac{\partial T_2}{\partial y}, \tag{2}$$

where T is temperature, z geopotential height, g the acceleration of gravity, and the subscript 2 refers to the interface between the two layers. In Eq. (2) it is understood that the meridional derivative is evaluated at constant pressure. If we set $u_1-u_3=u_c$ and substitute from Eq. (1) in Eq. (2) we find for the critical temperature gradient

$$\left.\frac{\partial T_2}{\partial y}\right|_c=-H_2\frac{\beta}{f}\frac{\theta_1-\theta_3}{z_1-z_3}. \tag{3}$$

Here H_2 is the scale height at the interface. This expression can be simplified further if we introduce the definitions of f and β and write the meridional gradient in terms of the latitude ϕ. The critical gradient in the two-layer model is then

$$\left.\frac{\partial T_2}{\partial \phi}\right|_c=-H_2\frac{\theta_1-\theta_3}{z_1-z_3}\cot\phi. \tag{4}$$

It should be noted that the critical gradient is determined by the vertical structure of the atmosphere.

Since we will be using data specified at latitude intervals of 5°, we will have to write the meridional gradient in finite difference form. Specifically, we will define a temperature difference across a 5° latitude belt,

$$\Delta T(\phi)=T(\phi-2\tfrac{1}{2}°)-T(\phi+2\tfrac{1}{2}°), \tag{5}$$

which is, therefore, related to the gradient by

$$\frac{\partial T}{\partial \phi}\approx-\frac{36}{\pi}\Delta T. \tag{6}$$

The critical value of ΔT_2 is thus

$$\Delta T_2|_c=\frac{\pi}{36}H_2\frac{\theta_1-\theta_3}{z_1-z_3}\cot\phi. \tag{7}$$

We now wish to compare $\Delta T_2|_c$ with the observed ΔT for the atmosphere's zonal mean state. The data we will use is that given by Oort and Rasmusson (1971), which includes monthly, seasonal and annual mean values of the zonal means of T, θ and z, based on the five-year period May 1958–April 1963. These data are given at standard pressure levels, and we must decide what pressure levels to use in evaluating quantities with subscripts 1, 2 and 3. Since the greater static stability of the stratosphere tends to confine eddies to the troposphere we will identify the upper layer of the two-layer model with the upper half of the troposphere and the lower layer with the lower half. Since the 200 mb level approximates the tropopause level in mid and high latitude, we will identify the upper layer with the pressure levels 200–600 mb, and the lower layer with the levels 600–1000 mb. Quantities referred to one layer in the two-layer model, i.e., θ_1, z_1, etc., will then be identified with the mass-weighted mean values of the respective atmospheric variables, averaged over the appropriate pressure levels. Quantities evaluated at the interface of the two-layer model will be identified with the respective atmospheric variables at the 600 mb level.

Fig. 1 shows $\Delta T_2|_c$ and the observed ΔT_2 as a function of latitude, calculated in the above way from Oort and Rasmusson's (1971) data for annual mean conditions. In addition the figure includes ΔT cal-

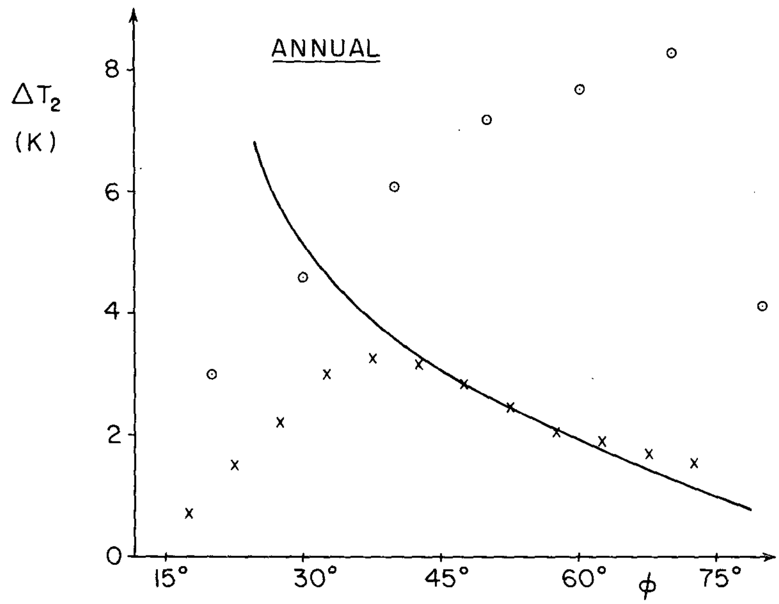

FIG. 1. Various zonal mean meridional temperature gradients at 600 mb vs latitude, evaluated for annual mean conditions. The solid curve gives the critical gradient [Eq. (7)], the X's indicate the observed gradient and the O's indicate the radiative equilibrium gradient.

culated from the local effective radiative temperature

$$T_e=\left(\frac{A}{s}\right)^{\frac{1}{4}}, \tag{8}$$

where A is the zonal mean absorption of solar radiation per unit surface area and s the Stefan-Boltzman constant. The values of A were taken from Ellis and Vonder Haar (1976). These values of ΔT are a rough approximation of what the 600 mb gradients would be if the atmosphere were in local radiative equilibrium. The differences between these values and the observed values are a measure of the effect of the dynamical fluxes on the meridional temperature gradient. Fig. 1 corroborates the result found by Moura and Stone (1976). In mid and high latitudes the

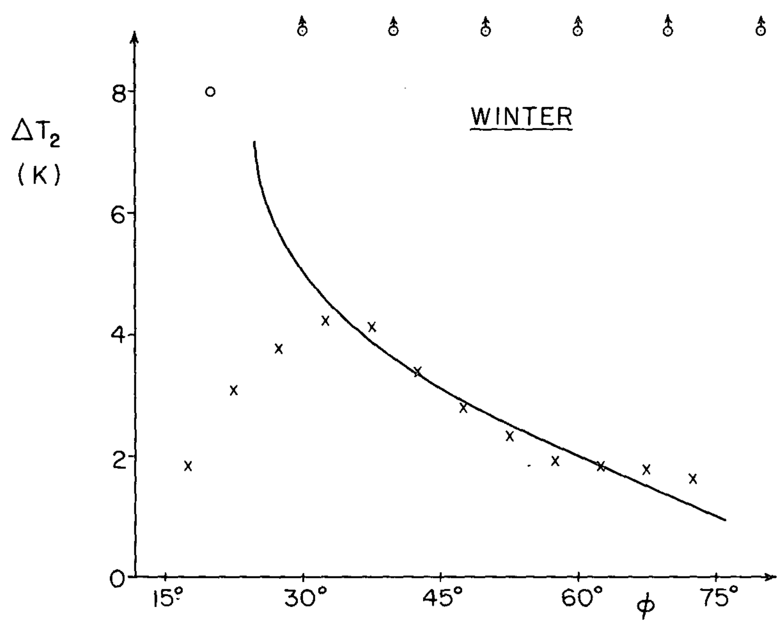

FIG. 2. As in Fig. 1 except for winter mean conditions (December, January, February).

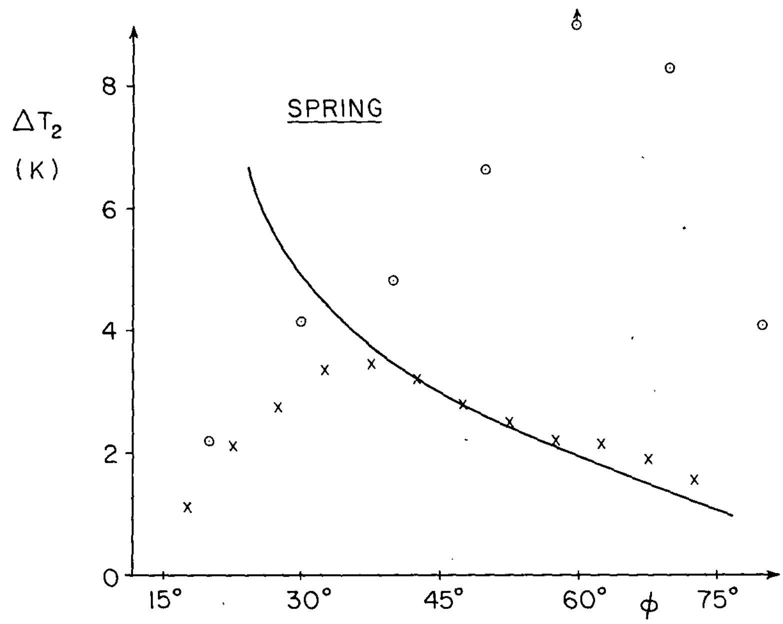

FIG. 3. As in Fig. 1 except for spring mean conditions (March, April, May).

mid-tropospheric gradients are considerably less than they would be in the absence of dynamical fluxes, and in fact follow closely the critical gradient of the two-layer model.

In order to check whether this coincidence persists under different conditions, the same gradients were calculated for the different seasons from Oort and Rasmusson's (1971) and Ellis and Vonder Haar's (1976) data. The results are shown in Figs. 2–5, for the winter, spring, summer and fall seasons, respectively. The same result holds in all seasons. Figs. 2–5 show that the primary feature of the seasonal change at 600 mb is the shift in latitude of the maximum meridional temperature gradient. This maximum, whose position corresponds closely to the position of the mean jet stream, shifts from 33° in winter to 45° in summer. On the poleward side of this maximum, the observed zonal mean gradient remains close to the local value of the critical gradient as defined by Eq. (1) in all seasons, even though the radiative equilibrium gradient changes by a factor of 4.

Figs. 2–5 do indicate small changes in the observed

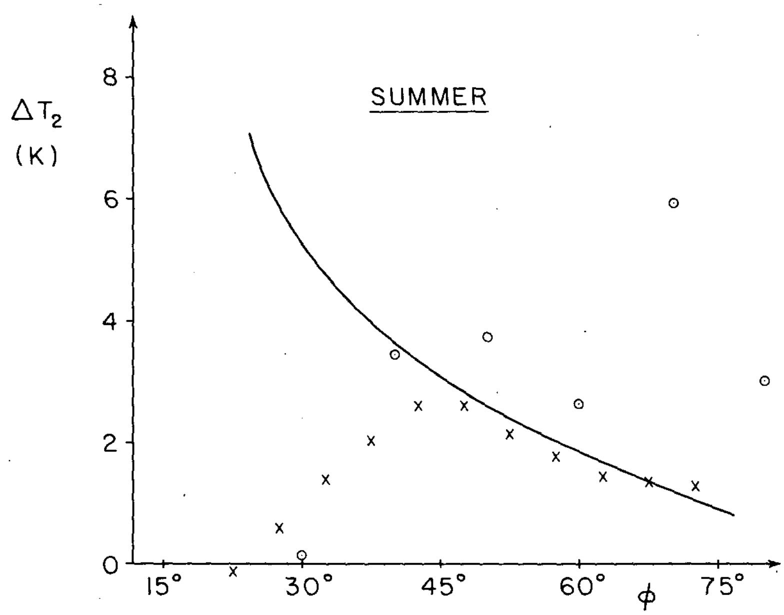

FIG. 4. As in Fig. 1, except for summer mean conditions (June, July, August).

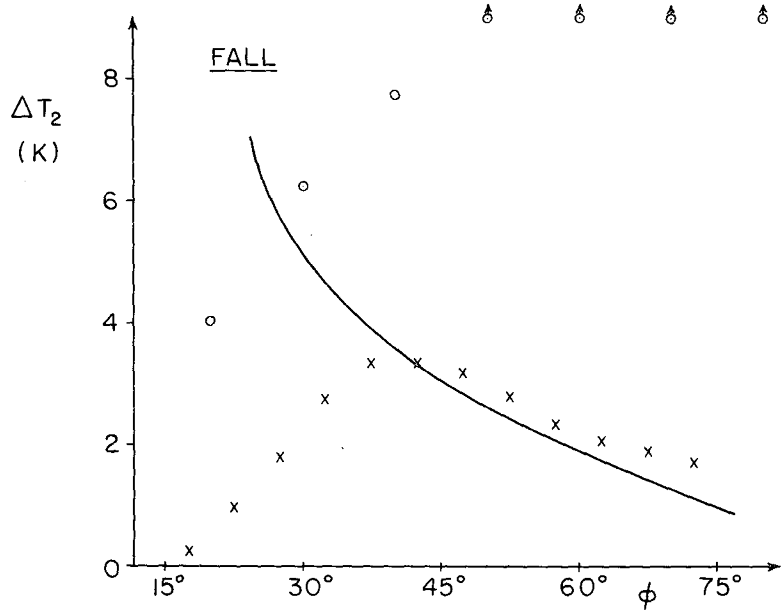

FIG. 5. As Fig. 1 except for fall mean conditions (September, October, November).

mean gradients relative to the critical gradient. These seasonal changes are about 25%. It is not clear how much of these changes is real and how much is due to the statistical variability of 15-month averages. In any case the changes are much smaller than the changes in the radiative equilibrium gradients.

In the annual case, the observed zonal mean gradients do appear to deviate from the critical gradient by a small but significant amount in high latitudes. This behavior is present in all seasons, and is not surprising if we recall that the critical gradient in the two-layer model is associated with a specific zonal wavelength. In sufficiently high latitudes, the length of a latitude circle becomes smaller than this wavelength, and the critical gradient must lose any special significance.

If we assume that the meridional and zonal scales of the critical wave are equal, then the critical zonal wavelength is given by (Phillips, 1954)

$$L=2^{\frac{1}{4}}(2\pi/f)[R(\theta_1-\theta_3)]^{\frac{1}{2}}. \tag{9}$$

Thus, the latitude ϕ_0 at which L is equal to the length of the latitude circle is given by

$$\phi_0=(\pi/2)-\tfrac{1}{2}\arcsin\left\{\frac{2^{\frac{1}{4}}}{\Omega a}[R(\theta_1-\theta_3)]^{\frac{1}{2}}\right\}, \tag{10}$$

where Ω is the rotation rate and a the radius of the earth. If we evaluate θ_1 and θ_3 for annual mean conditions at 75°, we find $\phi_0=83°$. The fact that the relation between the observed and the critical gradients changes at 70° indicates that the changeover from latitudes where the critical gradient is significant to those where it is not is not sharp. This is to be expected in view of the finite meridional extent of atmospheric eddies.

3. Baroclinic adjustment

The results of the preceding section suggest that the local value of the critical gradient as defined by Eq. (1) is an effective upper bound to the local value of the meridional gradient that can occur in the atmosphere. Such a role for the critical gradient can be explained if the long, deep waves which dominate under supercritical conditions are indeed much more efficient at transporting heat poleward than the short, shallow waves which dominate under subcritical conditions. Held (1978) has recently used scaling arguments to conclude that such will be the case. According to his analysis the vertically integrated kinetic energy and eddy flux of sensible heat in a baroclinic wave are proportional to the cube of the wave height. Thus the long, deep waves should be much more efficient at transporting heat than the short, shallow waves. Held's analysis is supported by Gall's (1976) numerical study which showed that the long, deep baroclinic waves do attain much greater amplitudes than the short, shallow waves. Because of the cubic dependence on wave height, one can expect a considerable sensitivity of the flux to small changes in gradient when the gradient is near the critical value.

The degree of sensitivity can be estimated from Figs. 2–5. For example, the apparent seasonal changes in the mean gradient in mid-latitudes are no more than about 25%, whereas the seasonal change in the eddy flux of sensible heat is about 280% (Oort and Rasmusson, 1971). In fact, the seasonal change in the flux corresponds closely to the seasonal change

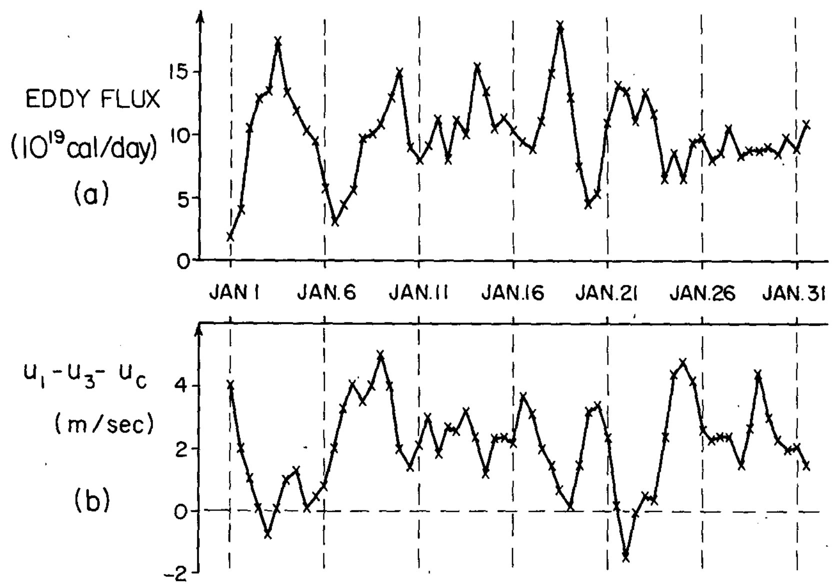

FIG. 6. Eddy flux of sensible heat (a) and excess of the mean shear over the critical shear (b) at 50°N during January, 1974.

in the differential radiative heating, i.e., just enough flux is produced to reduce the mean gradients from the radiative equilibrium values to the critical value. According to Held's argument such large changes in the flux should be accompanied by large changes in the amount of energy in the long, deep waves. Spectral analyses of the atmosphere do in fact show such a change (Saltzman, 1970; Julian *et al.*, 1970). According to Saltzman's analysis, the kinetic energy in planetary wavenumbers $\geqslant 6$ shows virtually no seasonal change, while the kinetic energy in wavenumbers <6 is twice as much in the winter half of the year as in the summer half.

A clearer picture of the relation between the eddy flux and the meridional temperature gradient can be obtained by examining time series of the two quantities. For this purpose the total tropospheric eddy flux of sensible heat across a latitude belt, the zonal mean zonal wind shear, and the zonal mean of the local value of the critical shear were calculated from data from the National Meteorological Center's analyses for synoptic times for three Januaries (1973–75). Fig. 6 shows a typical mid-latitude time series for the eddy flux and for the difference between the mean shear and the critical shear. (This difference is of course proportional to the difference between the mean meridional gradient and the critical gradient.) This particular series is for 50°N during January 1974. In each January the time and zonal mean eddy flux peaked at 50°N.

Fig. 6 illustrates a property found throughout mid-latitudes in all three Januaries—namely, on short time scales there is a clear negative correlation between the eddy flux and the excess of the shear over the critical shear. This negative correlation is apparently the same as the negative correlation between the eddy flux and the meridional temperature gradient found recently by E. N. Lorenz (personal communication) in a much more extensive analysis covering 10 years of data. For this particular time series the correlation coefficient is -0.49. The mean correlation coefficient at 50°N in the three Januaries was -0.38. This behavior suggests the following picture of how the gradient is maintained near the critical value. When the gradients are supercritical, the eddy flux is weak, and conditions are favorable for the growth of the long, deep waves. In a few days these waves grow to large amplitude and produce a strong eddy flux, which reduces the gradient to subcritical values. Now the main heat transporting waves are stable and decay. As a result the eddy flux decreases and, this allows the diabatic processes to build the temperature gradient up to supercritical values. Then the cycle repeats. This process can be thought of as one of *baroclinic adjustment*, i.e., over long periods of time baroclinic waves adjust the gradient so as to keep it from being appreciably supercritical. We note that individual months can be appreciably more supercritical than the mean seasons illustrated in Figs. 2–5, each of which contain 15 months of data. For example, the three Januaries studied had mean gradients at 50°N equal to 0.97, 1.19 and 1.32 times the critical gradient in 1973, 1974 and 1975, respectively.

All of the preceding results suggest that one does not need an explicit calculation of the eddy flux in order to take into account its effect on the zonal mean meridional temperature structure. The effect appears to be closely analogous to the effect of Bénard convection on atmospheric lapse rates, i.e., in subcritical conditions the eddy fluxes are negligible and the gradient is set by the other mechanisms present, while in supercritical conditions the eddy fluxes are so efficient that they do not allow the gradient to become significantly supercritical. Thus one need only solve for the meridional temperature structure with the eddy fluxes neglected, and then adjust the meridional temperature gradient so that

$$\frac{\partial T_2}{\partial \phi} \leqslant \left.\frac{\partial T_2}{\partial \phi}\right|_c \quad \text{when} \quad \phi < \phi_0. \tag{11}$$

Provided one is concerned with time periods of at least a few months, this baroclinic adjustment should give an excellent approximation to the effect of the eddies.

For example, in mid and high latitudes the temperature gradient must be adjusted so that

$$\frac{\partial T_2}{\partial \phi} = \left.\frac{\partial T_2}{\partial \phi}\right|_c = -H_2 \frac{\theta_1 - \theta_3}{z_1 - z_3} \cot\phi. \tag{12}$$

The restriction $\phi < \phi_0$ is of little consequence for the earth, because ϕ_0 is near 90°, and because the gradients are in any case small in high latitudes. Furthermore, the scale height and static stability are slowly varying compared to $\cot\phi$. Therefore, we can replace Eq. (12) by

$$\frac{\partial T_2}{\partial \phi} \approx -k \cot\phi, \tag{13}$$

where k is a typical value of the coefficient in Eq. (12). Now we can integrate (13) to obtain an approximate asymptotic form for the temperature field in mid and high latitudes,

$$T_2 \approx T_2(\pi/2) - k \ln \sin\phi. \tag{14}$$

The integration constant is determined by joining this solution to the solution in low latitudes.

Fig. 7 shows how well this simple solution fits the atmosphere's temperature structure in mid and high latitudes in January, July and the annual mean. In making these comparisons in each case, k was evaluated at 55°N latitude, and the integration constant was determined by fitting Eq. (14) to the correct temperature at 55°N. All the necessary data were again taken from Oort and Rasmusson (1971). We note that the temperatures found by using the baroclinic adjustment parameterization have even less relative error than the temperature gradients, because the meridional temperature contrasts are much smaller than the absolute temperatures.

In all of our discussion so far, we have focused on the mid-tropospheric temperature T_2. Since atmospheric eddies must respond to temperatures throughout the troposphere, and not just to temperatures at the mid-tropospheric level, one might expect the physically relevant temperature to be some appropriate tropospheric mean temperature rather than T_2. Fig. 8 compares the meridional temperature gradient at mid-troposphere (600 mb) with the tropospheric mean temperature gradient (the mass-weighted gradient from 200 to 1000 mb) in winter and summer. In fact there is very little difference between these two gradients in all seasons in mid and high latitudes. Thus all of our preceding discussion can be applied to the

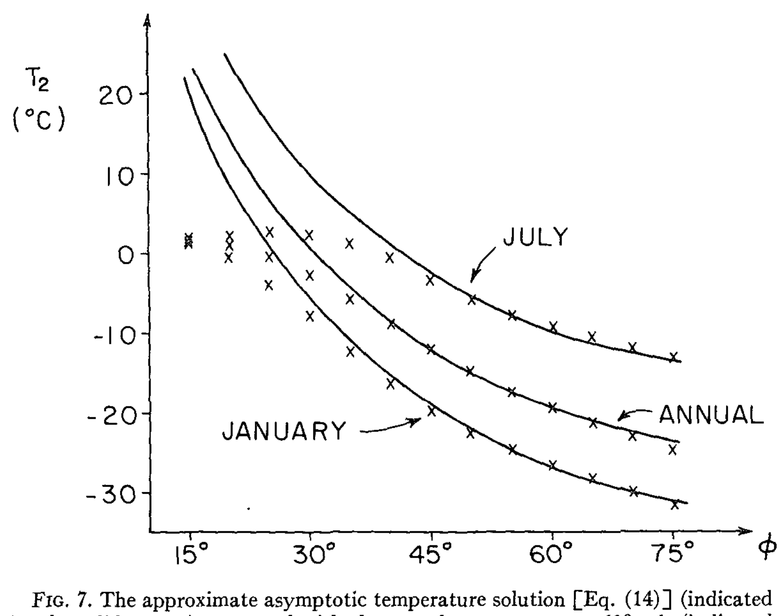

FIG. 7. The approximate asymptotic temperature solution [Eq. (14)] (indicated by the solid curves) compared with the actual temperature at 600 mb (indicated by the X's) for January, July and the annual mean.

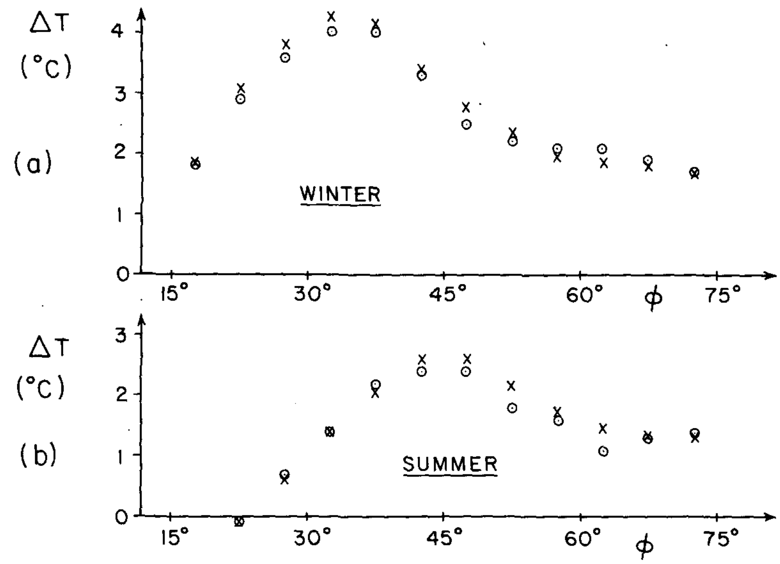

FIG. 8. Comparison between the mid-tropospheric temperature gradient (X's) and the mean tropospheric temperature gradient (O's) (a) in winter and (b) in summer.

tropospheric mean gradient, with almost as great accuracy as when applied to the mid-tropospheric gradient.

4. Effect in an energy balance climate model

In order to illustrate the application of the baroclinic adjustment mechanism, we will use it in a one-dimensional heat balance model for the annual mean latitudinal temperature structure. We will use the particularly simple formulation of this type of model given by North (1975). In this formulation the dynamical heat transport is represented by a linear diffusion law, and the temperature structure is approximated by the first two Legendre polynomials in a series expansion. The flux I of infrared radiation to space is taken to be the prime dependent variable. The edge of the polar ice cap is placed at the latitude where the surface temperature equals $-10°C$, and the sine x_s of this latitude is determined implicitly.

Because of the two-mode approximation, the solution for the temperature structure in North's (1975) formulation can be reduced to the solution of three coupled algebraic equations. The unknowns in these equations are the two coefficients of I in the truncated expansion, I_0 and I_2, and x_s. The details are given by North (1975). The three equations are

$$I_0=(q/4)H_0(x_s), \quad (15)$$

$$6DI_2=(q/4)H_2(x_s)-I_2, \quad (16)$$

$$I_s=I_0+\tfrac{1}{2}I_2(3x_s^2-1), \quad (17)$$

where q is the solar constant, D the thermal diffusivity corresponding to the present climate ($D=0.382$), and I_s the outgoing flux of infrared radiation corresponding to a surface temperature of $-10°C$ ($I_s=195.7$ W m^{-2}). The functions H_0 and H_2 are defined by Eq. (12) in North (1975). H_0 is a fifth-order polynomial in x_s and H_2 a seventh-order polynomial.

Eq. (15) corresponds to the statement that the global absorption of solar radiation is balanced by the global emission of infrared radiation; Eq. (16) to the statement that the poleward diffusion of heat compensates the net radiative heating in low latitudes and the net radiative cooling in high latitudes; and Eq. (17) is the definition of the position of the edge of the polar ice cap. Eqs. (15)–(17) can be combined into a single ninth-order polynomial equation for x_s, i.e.,

$$I_s=(q/4)\left[H_0(x_s)+\frac{H_2(x_s)(3x_s^2-1)}{2(6D+1)}\right]. \quad (18)$$

This equation defines a function $x_s(q)$ which describes how the extent of the polar ice cap changes as the solar constant changes. The result is illustrated in Fig. 2 of North (1975).

We now wish to determine $x_s(q)$ when the dynamical transports are controlled by the baroclinic adjustment process in mid-latitudes. The radiative equilibrium gradient in mid-latitudes is much larger than the actual gradient—about twice as large according to Fig. 1 and about three times as large according to Eq. (16). Thus we anticipate that baroclinic adjustment will keep the mid-tropospheric gradients in mid-latitudes near the critical value even for substantially different climates. We can, therefore, parameterize baroclinic adjustment by requiring that Eq. (12) hold in mid-latitudes.

The one-dimensional energy balance models implicitly neglect any changes in the vertical structure of the atmosphere. Therefore, according to Eq. (12), the critical gradient must be assumed to be the same in different climates, and we must hold the mid-tropospheric gradients fixed in mid-latitudes. Fixing the vertical structure also implies that the surface temperature gradient is constant whenever the mid-tropospheric gradient is. In North's formulation, the surface temperature gradient has a fixed distribution which peaks at 45° latitude, and its magnitude is proportional to I_2. Therefore, we can introduce the baroclinic adjustment process to this model by simply fixing I_2 at its present value ($I_2 = -44.0$ W m^{-2}).

If I_2 is fixed, then Eq. (16) is decoupled from Eqs. (15) and (17), and (15) and (17) can be combined directly into a single equation for x_s, i.e.,

$$I_s = (q/4)H_0(x_s) + \tfrac{1}{2}I_2(3x_s^2 - 1). \qquad (19)$$

Eq. (16) remains valid, but now it must be considered an equation for D rather than I_2. As q changes, both D and the dynamical flux necessary to maintain the mid-latitude gradient at the same value must change. Eq. (19) is only a fifth-order polynomial equation, in contrast to the ninth-order equation that results when a linear diffusion law is used.

The solution for $x_s(q)$ found from Eq. (19) is presented in Fig. 9. The present value of the solar constant q_0 was chosen to be 1337.6 W m^{-2}. The solution for the linear diffusion law is also shown for comparison. The qualitative behavior of the x_s vs q curve is the same for both parameterizations, since this behavior results from the albedo-temperature feedback, which has not been changed. However, when baroclinic adjustment is included the current climate ($x_s = 0.95$) is about twice as stable. For example, if the solar constant drops by 1%, with the linear diffusion law the ice cap moves 11.3° southward, and the global mean temperature drops by 4.4 K; but with the baroclinic adjustment parameterization the ice cap only moves 5.3° southward and the mean temperature drops by only 2.6 K.

Fig. 10 shows how the thermal diffusivity D calculated from Eq. (16) changes with solar constant. The behavior of D provides us with a check on the consistency of our parameterizing the baroclinic adjustment process by keeping the mid latitude gradients and I_2 constants. D has a minimum value D_0, corresponding to the thermal diffusivity in the absence of eddy fluxes, i.e., the variations in D illustrated in Fig. 10 must correspond to variations in the eddy flux, and D cannot fall below D_0. Thus, the complete parameterization of the baroclinic adjustment process in this model would be

$$I_2 = -44.0 \text{ W m}^{-2}, \quad \text{if} \quad D > D_0, \qquad (20)$$

or

$$D = D_0, \qquad \text{if } |I_2| < 44.0 \text{ W m}^{-2}. \qquad (21)$$

We don't know the value of D_0, but we can be sure that it is considerably less than the current value since most of the flux in mid-latitudes is currently carried by eddies. Fig. 10 shows that D does not fall very much below its current value for values of $q \approx q_0$; therefore, Eq. (20) rather than Eq. (21) is indeed appropriate for these climatic states.

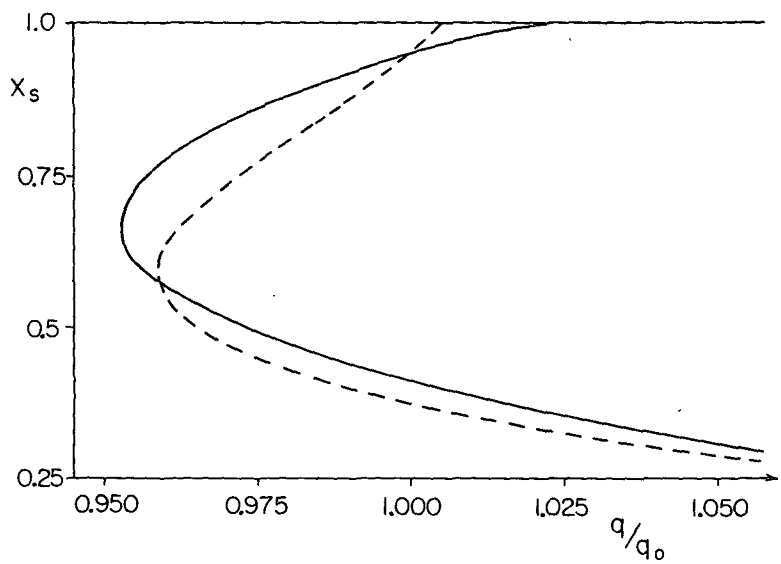

FIG. 9. Position of the edge of the polar ice cap versus solar constant when the dynamical transports are parameterized by the baroclinic adjustment process (solid curve) and by a linear diffusion law (dashed curve).

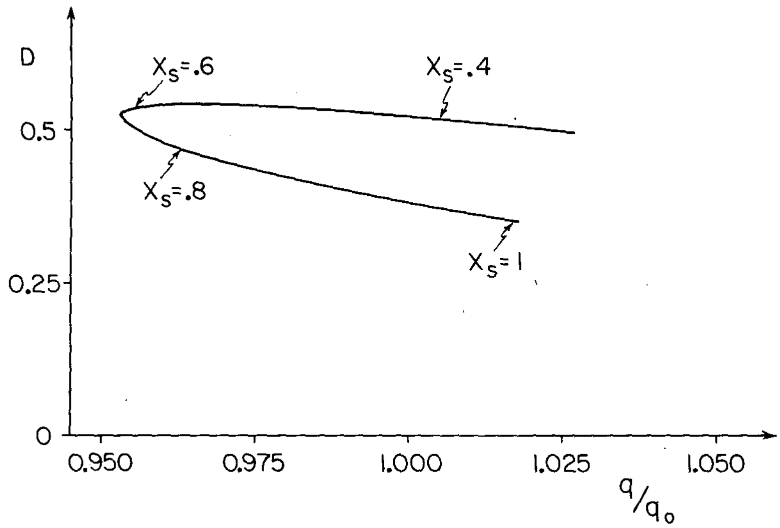

FIG. 10. Thermal diffusivity versus solar constant.

5. Discussion

Our analysis of the observations shows that the zonal mean mid-tropospheric and mean tropospheric meridional temperature gradients on the poleward side of the mean jet stream coincide closely with the local value of the critical gradient as defined by Eq. (1). The agreement is closest over long-time periods and occurs under a wide range of conditions, i.e., for all seasons. Qualitatively this behavior can be attributed to the enhancement of the eddy flux in a continuous atmosphere under supercritical conditions. This enhancement leads to a negative feedback between the meridional eddy flux of heat and the meridional temperature gradient. This feedback restricts gradients to values near the critical value, a process which we have termed *baroclinic adjustment.*

The remarkable closeness of the observed and critical gradients indicates that the eddy flux is particularly sensitive to changes in the gradient when the gradient is near the critical value. This sensitivity is associated with a corresponding sensitivity of the kinetic energy spectrum. An explanation of this extreme sensitivity, therefore, will require the development of a theory for how the spectrum depends on the meridional temperature structure. In any case, this great sensitivity enables us to take into account the effect of the eddies on the zonal mean temperature structure in climate models without explicitly calculating the eddy flux. The effect can be modeled quite accurately simply by restricting the local values of the zonal mean meridional temperature gradients so that they do not exceed the local values of the critical gradient. This adjustment automatically includes the effect of eddy fluxes of both sensible and latent heat and automatically includes β-effects. Our analysis of an energy balance climate model illustrates that this approach leads to a simpler mathematical formulation and that baroclinic adjustment enhances the stability of the current climate.

One particularly interesting feature of the baroclinic adjustment process is that it strongly couples the meridional and vertical temperature structure [see Eq. (12)]. This should make it quite easy to construct simple climate models which incorporate feedbacks associated with both the meridional and vertical temperature structure. Also the fact that the process is present in the two-layer model, even if in modified form, supplies a strong justification for using the two-layer model when modeling baroclinic eddy processes. Finally, the closeness of the observed gradient to the critical gradient gives a strong motivation for simplifying solutions of the two-layer model in mid and high latitudes by making expansions about the critical gradient.

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