ANNALI DI GEOFISICA, VOL. 41, N. 4, October 1998

On baroclinic adjustment of a radiative convective atmosphere

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Abstract
In this paper we study the implication of the hypothesis that the radiative convective equilibrium climate is neutral with respect to baroclinic eddies. If such neutral state is achieved by tropopause height readjustment, we find that the sensitivity of the climate equilibrium to baroclinic eddies is comparable to the sensitivity to water vapor profile. Multiple solutions to the readjusted tropopause are found by decreasing stratospheric static stability.

Key words *climate – equilibrium – baroclinic*

1. Introduction

In the next century planet Earth faces the potential hazard of climate changes, such as climate warming, rising sea level, deforestation, desertification, ozone depletion, acid rain, and reduction in biodiversity. However, many related important scientific questions remain unanswered. For example, while a significant global climate change, man induced or not, is likely, its magnitude and timing (both at the global and regional level) are quite uncertain. Additional information on the rate, causes, and effects of global change is essential in order to develop an understanding of a such important physical process. An overreaching goal of a program to determine the extent, causes, and regional consequences of global climate change consists in the understanding of how the change in the average temperature and the time scale over which it will occur may cause variations which prove to be most detrimental to the environment. For these reasons studying Climate is an unpostponable research program.

Lastly, we will understand climate changes through a combined effort between careful monitoring of climate behaviour and theoretical studies, which in turn may help the design of a complete and efficient observational network.

The present paper, which falls in the latter category of studies, discusses how climate equilibrium may be achieved through the interaction of the adiative properties of the atmosphere and the dynamical processes.

Although it is not longer common practice, we will try to approach the question on the ground of simple calculations that explicitly avoid detailed interactions. Nevertheless, we hope that the conclusions which we may draw will be helpful for planning future studies or measurement needs.

We intend to study the sensitivity of a radiative-convective equilibrium surface temperature on the baroclinicity of the resulting atmosphere. The observed climate statistics do not show baroclinic active eddies. In fact, the long term mean removes these features. However,

Dedicated to Prof. M. Caputo on his 70th birthday.

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as is known, the heat and momentum transport implied by these disturbances modify the climate equilibrium especially in midlatitude, probably leaving the symmetric circulation near a neutral state.

In studying these effects, we solve in Section 2 for a radiative-convective atmosphere where the zonally symmetric circulation is determined by a simple prescribed profile of the radiation reaching the top of the atmosphere. In Section 3 we present a condition for neutrality with respect to the baroclinic (Eady) atmosphere which is parameterized through the tropopause height variation. The overall question whether baroclinic unstable eddies acted to bring the atmosphere to a neutral state has been discussed for many years (see Stone and Brascombe, 1992 and reference therein).

These studies focus on the Charney-Stern theorem (Charney-Stern, 1962) following which the condition for the flow to be unstable involves the existence of a surface temperature gradient. In fact, the potential vorticity gradients are generally positive in the interior flow, so that the Dirac delta implied by the boundary condition destabilizes the fluid. For the Eady model, being the potential vorticity gradient zero in the entire domain, baroclinic instability is due to the interaction of two waves generated at the two vertical boundaries.

Thus, we can have surface temperature gradient and stable eddies provided that the two edge waves are uncoupled. As a consequence, in the Eady problem neutrality can be achieved by sufficient displacement of the model top layer which is generally assumed to be a rigid surface. In Section 4 we calculate the sensitivity to model parameter changes. Conclusions and a few speculations are offered in the final section.

2. A simple model for radiative equilibrium

In this section we develop a «toy model» for radiative-convective processes in the atmosphere.

Let us consider the simplest equilibrium which an atmosphere can attain, *i.e.* a hydrostatic equilibrium with no motion. In this case, assuming a gray atmosphere, the equations of radiative transfer in the two stream approximations are

$$\begin{aligned} \frac{d}{d\tau}U &= \frac{3}{2}(U-B) \\ \frac{d}{d\tau}D &= \frac{3}{2}(B-D) \end{aligned} \tag{2.1}$$

where U and D are the upward and downward infrared fluxes respectively; $B=\sigma T^4$. $\tau(z)$ is the optical depth. Boundary conditions are $D(0)=0$ and $B(\tau(0))=\sigma T_s^4$ where T_s is the surface temperature.

We suppose that the atmosphere is transparent to the solar radiation I and that it is in radiative equilibrium. Moreover, we consider I as a linear function of latitude y. With these assumptions we may determine the temperature structure in the meridional plane. In fact, since the atmosphere is supposed to be in radiative equilibrium we have

$$U-D=I. \tag{2.2}$$

Thus

$$T(z,y)=\left[\frac{I}{\sigma}\left[\frac{1}{2}+\frac{3}{4}\tau\right]\right]^{-\frac{1}{4}}. \tag{2.3}$$

The corresponding lapse rate

$$\Gamma(z,y)=\frac{dT}{dz}=-\frac{3}{16}\left[\frac{I}{\sigma}\left[\frac{1}{2}+\frac{3}{4}\tau\right]\right]^{-\frac{3}{4}}\frac{I}{\sigma}\frac{d\tau}{dz}. \tag{2.4}$$

For an hydrostatic, motionless atmosphere we must have

$$\Gamma_d-\Gamma\geq 0 \tag{2.5}$$

where $\Gamma_d = 9.8$ °K/km is the dry adiabatic lapse rate. The optical depth τ depends on the atmospheric constituents, among which the most important absorber, both for solar and longwave radiation, is water vapor.

We will present the case in which a detailed profiling of this constituent, especially in the upper troposphere, is of a paramount importance in determining the equilibrium structure of the climate. For now, we pose (Goody and Yung, 1989)

$$\tau = \tau_0 \exp(-z/z_0) \tag{2.6}$$

where τ_0, z_0 are, respectively, the maximum optical depth and the scale height for water vapor. Typical values of τ_0, z_0 for an equilibrium atmosphere are 4 and 2 km respectively.

Taking for I_0 at the equator a typical solstice value of 298 W/m^2 and

$$I = I_0 - 1.2\,y$$

we get, in the meridional plane, the temperature structure shown in fig. 1.

Of course, this solution cannot be stable since the lapse rate exceeds the dry adiabatic one, hence also the moist adiabatic lapse rate. Therefore, convection will occur which will reduce the tropospherical lapse rate under critical values. The new radiative-convective equilibrium can be calculated as follows.

Suppose that H_T is the height above which radiative equilibrium exists. Below such a height assurance that no convection occurs requires that the lapse rate decreases less than the dry adiabatic one. The observed lapse rate γ in the troposphere averages around 6.5 °K/km, so we choose such a value in our model.

Assuming temperature continuity at $z = H_T$ we have

$$T(z, y) = \left[\frac{I}{\sigma}\left[\frac{1}{2} + \frac{3}{4}\tau(z)\right]\right]^{\frac{1}{4}} \quad \text{for} \quad z \geq H_T \tag{2.7a}$$

and

$$T(z, y) = \left[\frac{I}{\sigma}\left[\frac{1}{2} + \frac{3}{4}\tau(H_T)\right]\right]^{\frac{1}{4}} + \gamma(H_T - z) \quad \text{for} \quad z \leq H_T. \tag{2.7b}$$

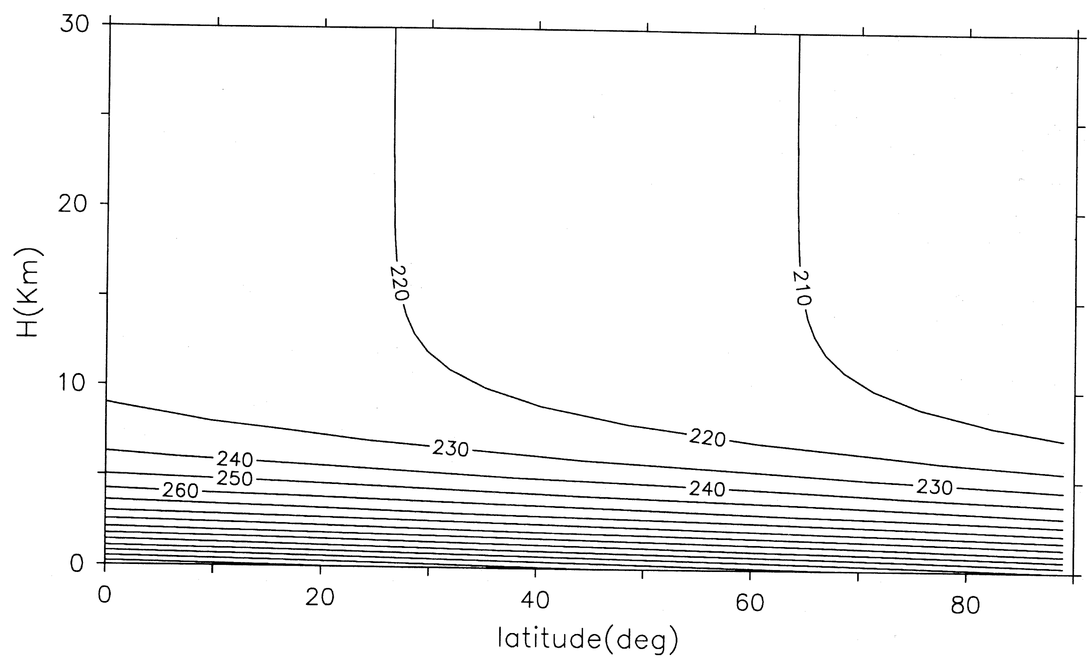

Fig. 1. The radiative equilibrium solution $T(z, y)$ for: $\tau_0 = 4$; $z_0 = 2$ km; $I_0 = 298$ W/m^2.

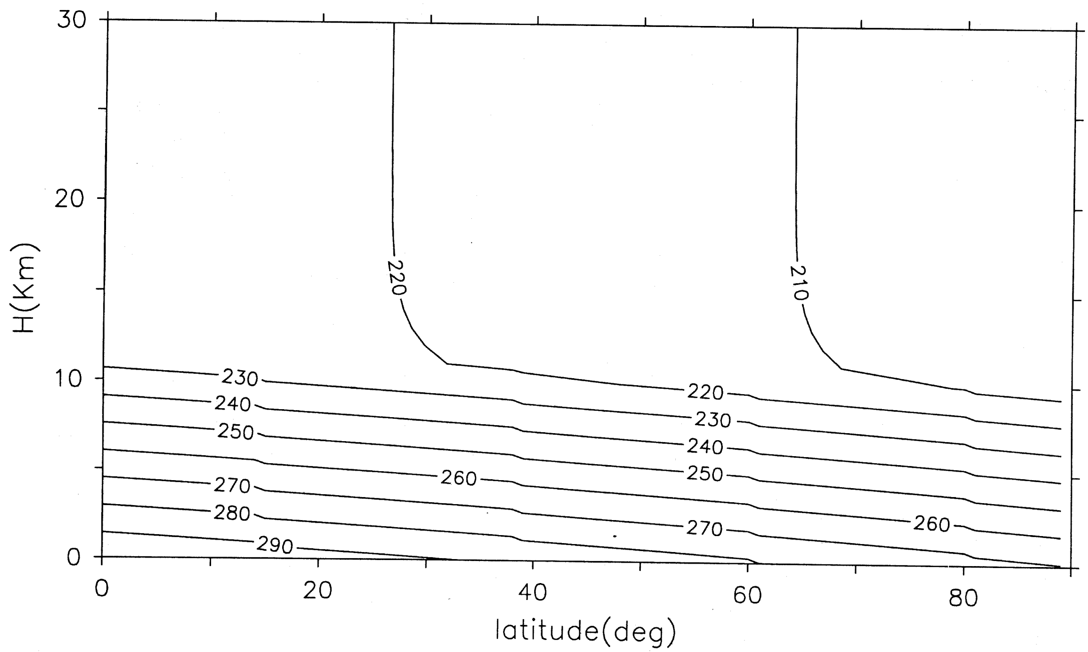

Fig. 2. The radiative convective equilibrium temperature $T(z, y)$ for the values: $\tau_0 = 4$; $z_0 = 2$ km; $I_0 = 298$ W/m^2; $\gamma = 6.5$ °K/km.

To determine H_T, we can solve (2.1) by increasing H_T from a given value up to the one for which the upper boundary condition on U is satisfied. Such a solution is shown in fig. 2.

This solution is stable with respect to ordinary convection as long as the atmosphere is not saturated. However, there are other processes which may destabilize the above solution; among them, certainly baroclinic instability processes account for a large portion of heat transport, at least in midlatitude. As we know, the time scale of baroclinic instability processes is much shorter than the one on which climate reaches a statistical equilibrium. In fact, the time mean flow at midlatitudes (say over winter) is the result of the interaction of radiation and heat transport associated with many instabilities and other dynamic stationary processes. However, the time mean flow does not show traces of these shorter time phenomena. Thus, to calculate their effects on the climate equilibrium we must parameterize these instabilities. The total effect will be a «renormalization» of the circulation which, in our simple model, will be reflected solely by a modification of the tropopause height H_T.

3. Baroclinic instability of radiative convective equilibrium

The meridional gradient implied by the radiative-convective solution previously discussed will drive a zonal circulation. In fact, the pressure gradient associated with the temperature field must be balanced by the Coriolis force. To compute the wind it is convenient to write the equation of motion in $\ln(p)$ coordinate system, p being the pressure.

If ζ is the new vertical coordinate defined as

$$\zeta = -H_0 \ln\left(\frac{p}{p_0}\right) \tag{3.1}$$

where H_0 is an arbitrary scale height and p_0 an arbitrary reference pressure the β-channel equations of motion are

$$
\begin{aligned}
&\frac{d}{dt}u - fv = -\frac{\partial}{\partial x}\Phi \\
&\frac{d}{dt}v + fu = -\frac{\partial}{\partial y}\Phi \\
&\frac{\partial}{\partial \zeta}\Phi = \frac{RT}{H_0} \\
&\left(\partial_t + \vec{u}_h\right)\frac{\partial}{\partial \zeta}\Phi + \dot{\zeta}\frac{\partial}{\partial \zeta}\left(\frac{\partial}{\partial \zeta}\Phi + \kappa\Phi\right) = \kappa Q \\
&\frac{\partial}{\partial \zeta}\dot{\zeta} - \dot{\zeta} + \left[\operatorname{div}\vec{u}_h\right]_{\zeta} = 0
\end{aligned}
\tag{3.2}
$$

where $\vec{u}_h = (u, v)$, u, v are the horizontal wind vector, and its zonal and meridional components respectively; Φ, R, T, Q are the geopotential, the perfect gas constant, temperature and heating rate. Finally $\kappa = R / c_p$, c_p being the specific heat at constant pressure, and

$$f = f_0 + \beta y,$$

with f_0, $\beta = \left[\frac{\partial}{\partial y}f\right]_{y=y_0}$ the Coriolis parameter and its derivative calculated at a central latitude y_0. The geostrophic balance and the hydrostatic equation imply that

$$
\begin{aligned}
&(\vec{u}_h)_G = (u_G, 0) = \left(-\frac{1}{f}\frac{\partial}{\partial y}\Phi, 0\right) \\
&\frac{\partial}{\partial \zeta}u_G = -\frac{R}{f}\frac{\partial}{\partial y}T.
\end{aligned}
\tag{3.3}
$$

Thus, although the system is assumed to be convectively stable, a slanted convection is still possible. Heat and moisture transport being allowed, correction to the radiative-convective equilibrium must be calculated.

If the sought equilibrium has to resemble the long-term multiyear time average, we can postulate that the atmosphere must be neutral with respect to baroclinic instability process of the sort implied by (3.3). In fact, observational and theoretical studies (Speranza and Malguzzi, 1988) show that the multiyear time mean of winter circulation has a very weak residual baroclinic wave structure mostly concentrated on the ultralong waves (Lindzen, 1993, Stone and Nemet, 1996; Hall and Sardeshmukh, 1998).

The way in which the atmosphere achieves such a state of baroclinic neutrality is matter of ongoing research and no firm results are yet available (Lindzen, 1990). Apparently, interaction with the meridional structure of the jet is a good candidate. However, a direct interaction with the tropopause height can be implied (Egger, 1995 and references therein).

Let us suppose here, just to see how far we can go, that baroclinic neutrality is achieved only by the variation of the tropopause height. To calculate the needed variation the relevant dynamics is quasi-geostrophic, and the equations of motion reduce to the conservation of the quasi-geostrophic potential vorticity

$$q = \nabla^2\Phi_G + f + \frac{1}{\rho_0}\frac{\partial}{\partial \zeta}\left(\varepsilon\rho_0\frac{\partial}{\partial \zeta}\Phi_G\right) \tag{3.4}$$

and the thermodynamic energy equation

$$\left(\frac{\partial}{\partial t} + \vec{u}_G\right)\frac{\partial}{\partial \zeta}\Phi_G + \dot{\zeta}N^2 = 0 \tag{3.5a}$$

where

$$
\begin{aligned}
&N^2 = \frac{R}{H}\left(\frac{\partial}{\partial \zeta}T + \kappa\frac{T}{H}\right) \\
&\varepsilon = \frac{f_0^2}{N^2}.
\end{aligned}
\tag{3.5b}
$$

The subscript G denotes geostrophic variables, while ρ_0 is the basic density profile. Suppose that the fluid is Boussinesq, then ρ_0 is a constant and drops from (3.4). As a further simplification let us consider the basic field deduced from the radiative-convective model as independent on y, *i.e* we consider the latitudinal average of the geostrophic basic wind and N^2. Moreover, in simplifying the following calculations let us assume that

$$u_b^1 = \lambda_1 \zeta \qquad N = N_1 \qquad \zeta \leq H_T$$
$$u_b^2 = \lambda_2 \zeta \qquad N = N_2 \qquad \zeta \geq H_T.$$

The above constants can be calculated as appropriate estimates from the radiative-convective solution (see next section).

With these assumptions, plus the condition $\beta = 0$, our baroclinic instability problem reduces to Eady problem and can be readily solved.

We distinguish two cases depending on whether the «vertical velocity» perturbation field $\dot{\zeta}$ is set to zero (rigid lid) or not at the tropopause.

For the rigid lid case, the linearized vorticity and thermodynamic equations are

$$(\partial_t + u_b^1 \partial_x)(\nabla^2 \phi_1 + \varepsilon_1 \partial^2_{\xi\xi} \phi_1) = 0$$
$$(\partial_t + u_b^1 \partial_x)\partial_\xi \phi_1 - \partial_x \phi_1 \, \partial_\xi u_b^1 = 0 \quad \text{at } \xi = 0, H_T \tag{3.6}$$

where ϕ_1 is the geopotential perturbation. As lateral boundary conditions we use rigid walls.

The normal mode solution is obtained by setting

$$\phi_1 = \phi_1(z) \cos ly \times \exp(ik(x - ct)). \tag{3.7}$$

Thus from (3.6) and (3.7) we get the following eigenvalue problem:

$$\frac{d^2}{d\zeta^2} \varphi_1 - \alpha_1^2 \varphi_1 = 0$$

$$(\lambda_1 \zeta - c) \frac{d}{d\zeta} \varphi_1 - \lambda_1 \varphi_1 = 0 \quad \text{at} \quad \zeta = 0, H_T$$

$$\alpha^2 = (k^2 + l^2)/\varepsilon$$

$$k = \frac{2\pi}{L_x}, \; l = \frac{\pi}{L_y}$$

which leads to dependence of c on H_T

$$c = \frac{\lambda_1 H_T}{2} \pm \frac{\lambda_1 H_T}{2} \left[1 - \frac{4 \cosh \alpha H_T}{\alpha H_T \sinh \alpha H_T} + \frac{4}{\alpha^2 H_T^2} \right]^{.5}.$$

Neutral perturbation is achieved for H_T such that the square bracket is ≥ 0 for any α, which occurs for

$$\alpha H_T / 2 = \coth(\alpha H_T / 2) \quad \text{or}$$

$$\alpha H_T \cong 2.4. \tag{3.8}$$

Next, consider that a $\zeta = H_T$ the variables are continuous but their vertical gradients are discontinuous functions of ζ. It implies that at $\zeta = H_T$ we must have

$$\varphi_1 = \varphi_2 \tag{3.9a}$$

$$\dot{\zeta}_1 = \dot{\zeta}_2. \tag{3.9b}$$

Moreover, we must require that

$$\varphi_2(\zeta) \to 0 \qquad \zeta \to \infty. \tag{3.9c}$$

Set

$$\alpha_1^2 = (k^2 + l^2)/\varepsilon_1$$

$$\alpha_2^2 = (k^2 + l^2)/\varepsilon_2$$

we get the following equations:

$$(\partial_t + u_b^1 \partial_x)(\nabla^2 \varphi_1 + \varepsilon_1 \partial^2_{\zeta\zeta} \varphi_1) = 0$$
$$(\partial_t + u_b^2 \partial_x)(\nabla^2 \varphi_2 + \varepsilon_2 \partial^2_{\zeta\zeta} \varphi_2) = 0. \tag{3.10}$$

The normal mode solutions are

$$\phi_1 = \varphi_1(\zeta) \cos ly \times \exp(ik(x - ct))$$

$$\phi_2 = \varphi_2(\zeta) \cos ly \times \exp(ik(x - ct)).$$

After substitution in (3.10) we get the following eigenvalue problem:

$$\frac{d^2}{d\zeta^2}\varphi_1 - \alpha_1^2\varphi_1 = 0$$
$$\frac{d^2}{d\zeta^2}\varphi_2 - \alpha_2^2\varphi_2 = 0 \qquad (3.11)$$

$$(\lambda_1\zeta - c)\frac{d}{d\zeta}\varphi_1 - \lambda_1\varphi_1 = 0 \text{ at } \zeta = 0 \qquad (3.12)$$

and

$$\frac{1}{N_1^2}\{(\lambda_1 H_T - c)\,\partial_\zeta\varphi_1 - \lambda_1\varphi_1\} =$$
$$= \frac{1}{N_2^2}\{(\lambda_2 H_T - c)\,\partial_\zeta\varphi_2 - \lambda_2\varphi_2\} \text{ at } \zeta = H_T. \qquad (3.13)$$

Boundary conditions require that

$$\varphi_2 = A_2 \exp(-\alpha_2\zeta)$$

while

$$\varphi_1 = A_1 \sinh \alpha_1\zeta + B_1 \cosh \alpha_1\zeta.$$

Substitution in (3.12) and (3.13) plus continuity (3.9a) gives

$$\Delta_1 c^2 + \Delta_2 c + \Delta_3 = 0 \qquad (3.14)$$

where the coefficients of the characteristic eq. (3.14) are lengthy expressions of the parameters of the problem which we will not report here. Neutrality requires that the discriminant of (3.14) be ≥ 0.

With the mathematical apparatus described above, we will analyze the effect of the baroclinic instability on the troposphere height as a function of z_0, the height scale of water vapour.

4. Sensitivity studies

Consider our midlatitude channel to be 30° wide, say from y_1 = 30°N to y_2 = 60°N.

Let us consider the dependence of H_T, λ_1, λ_2, N_1 and N_2 as a function of z_0 (from 1 to 3 km) for the case of the radiative-convective equilibrium.

N_2 and λ, defined by eqs. (3.5b) and (3.3), have a weak dependence on the latitude which we averaged out across the channel and along the troposphere for N_1 and λ_1, and from H_T to H_{top} = 30 km for the stratosphere

$$\lambda_1 = \frac{1}{\Delta y H_T}\int_{y_1}^{y_2}\int_0^{H_T}\left(-\frac{R}{f_0}\frac{\partial}{\partial y}T_1\right)dy\,d\zeta,$$

$$\lambda_2 = \frac{1}{\Delta y(H_{\text{top}} - H_T)}\int_{y_1}^{y_2}\int_{H_T}^{H_{\text{top}}}\left(-\frac{R}{f_0}\frac{\partial}{\partial y}T_2\right)dy\,d\zeta,$$

$$N_1^2 = \frac{1}{\Delta y H_T}\int_{y_1}^{y_2}\int_0^{H_T}\frac{R}{H}\left(\frac{\partial}{\partial\zeta}T_1 + k\frac{T_1}{H}\right)dy\,d\zeta,$$

$$N_2^2 = \frac{1}{\Delta y(H_{\text{top}} - H_T)}\int_{y_1}^{y_2}\int_{H_T}^{H_{\text{top}}}\frac{R}{H}\left(\frac{\partial}{\partial\zeta}T_2 + k\frac{T_2}{H}\right)dy\,d\zeta.$$

Here T_1 and T_2 are the temperature profiles of radiative-convective equilibrium in the troposphere and in the stratosphere respectively (see eqs. (2.7a), (2.7b)), Δy is the width of the channel. The results are summarized in table I.

Apparently, the detailed water vapour profile has a profound influence on the tropopause height H_T which, in turn, determines the surface temperature through the convective adjustment discussed in Section 2.

It remains to impose that the climate state just found must be neutral with respect to baroclinic instability. Lindzen (1993), using the Eady model of Section 3 with the rigid lid condition, found that H_T could be determined, if the meridional scale of the Eady wave was specified. In fact, if L_y is such meridional scale

Table I. Values of H_T, λ_1, λ_2, N_1, N_2, T_s as function of $z_0 = 1, 2, 3$ km for the case of radiative convective equilibrium. T_s is the resulting surface temperature using the lapse rate $\gamma = 6.5$ °K/km in the troposphere.

H_T (km)	z_0 km	$\lambda_1 * 10^{-3}$ s^{-1}	$\lambda_2 * 10^{-3}$ s^{-1}	$N_1 * 10^{-2}$ s^{-1}	$N_2 * 10^{-2}$ s^{-1}	T_s (°K)
9	1	2.93	2.93	0.91	1.47	272.97
11	2	2.41	2.40	0.76	1.19	287.23
13	3	2.07	2.05	0.65	0.99	303.03

then the total wave number is $K_T^2 = k_x^2 + \frac{1}{L_y^2}$, $k_x = \frac{s}{r\cos\phi}$, s = zonal wave number and r is the Earth's radius. By considering Lindzen's choice of parameters, we found his results, namely, that to neutralize with respect to baroclinic disturbances the tropopause should be about 16 km. We can apply its argument to our case by considering a channel 30° wide and centered at 45°N. For the rigid lid case the results are summarized in table II.

The impact of the baroclinic waves on the climatic equilibrium seems to be relevant and comparable to the one introduced by the water vapour radiative effect. One would be tempted to conclude that, at variance with the tropical atmosphere, the climate equilibrium in midlatitude is determined by the baroclinic neutralization process here envisioned.

However, when we consider the effect of including a stratosphere, *i.e.* the region of radiative equilibrium, the neutralization process effect is substantially reduced. In fact in table III we report the results obtained by removing the rigid lid assumption and using the parameter values presented in table I. The table shows that the previous effect has been reduced and it is no longer comparable to the one associated to water vapour content. Nevertheless it is comparable to the effect that is implied by CO_2 doubling.

So far we have discussed how climate can achieve equilibrium by neglecting some important properties of the stratospheric radiation balance and transport. Here, in fact, the stratosphere is simply a region of strict radiative balance which is determined, in the absence of the solar heating due to ozone and the cooling associated with CO_2, by the upper boundary condition on the upwelling radiation. These assumptions, together with no meridional transport, do not allow a correct description of stratospheric circulation in the meridional plane. To improve the representation of the stratosphere, the proper approach would be to replace our radiative and stratospheric dynamics model with one where the above effects are explicitly calculated. Here we decided, instead, to call for observations and thus to empirically correct for the shortfalls of the model.

In particular, for the lack of the observed decreasing wind vertical shear above the

Table II. New values of tropopause height, H_{Tn}, as function of z_0 for which the atmosphere is neutral with respect to baroclinic instability for the rigid lid case. ΔT_s is the corresponding variation of the surface temperature in Kelvin degrees with respect to the radiative convective case.

H_T (km)	z_0 (km)	H_{Tn} (km)	ΔT_s (°K)
9	1	13.4	28.6
11	2	16.1	33.2
13	3	18.7	37.1

Table III. As in table II for the Eady two levels model.

H_T (km)	z_0 (km)	H_{Tn} (km)	ΔT_s (°K)
9	1	9.8	5.2
11	2	11.6	3.9
13	3	13.3	1.9

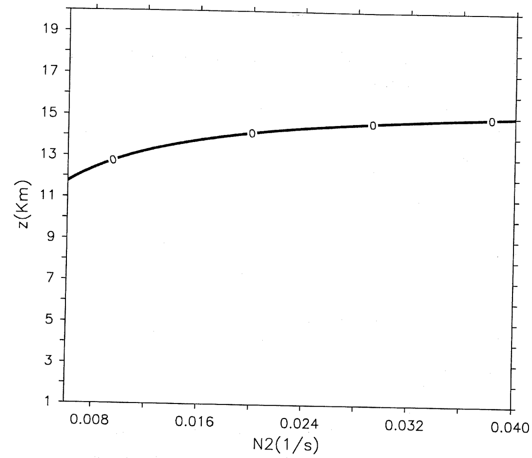

Fig. 3a. The height of tropopause for a neutral atmosphere with respect to baroclinic instability as a function of N_2; solid lines correspond to $Im(c) = 0$ for the case $z_0 = 2$ km and $\lambda_2 = 0$.

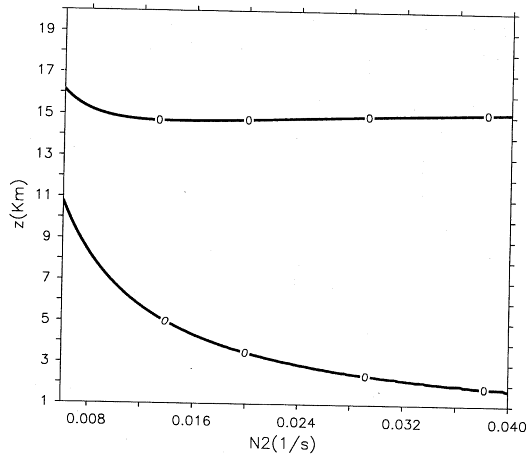

Fig. 3b. The height of tropopause for a neutral atmosphere with respect to baroclinic instability as a function of N_2; solid lines correspond to $Im(c) = 0$ for the case $z_0 = 2$ km and $\lambda_2 = -\lambda_1$.

tropopause we just assume that it exists. For the lack of the mechanism associated with ozone heating and CO_2 cooling (balance responsible for the increase of temperature with height in the stratosphere), we will consider bulk parameter values for static stability. This translates into λ_2 being negative or just zero, while, for N_2 we choose a range of variability included in between the one corresponding to the isothermal stratosphere previously computed and some reasonable asymptotic value suggested by observations.

As an example, we show in figs. 3a and 3b, two cases for $z_0 = 2$ km.

First, in fig. 3a we consider the case $\lambda_2 = 0$. It can be seen that by considering a constant wind, it is possible that a variation of the stratospheric static stability may be associated to a reduction of the tropopause height. The effect in terms of surface temperature changes is comparable to those which may be induced by CO_2. This effect is solely due to the baroclinic adjustment process.

Next, we can calculate the same effect in the presence of a decreasing wind in the stratosphere, again, as a function of the static stability in this layer.

Setting $\lambda_1 = -\lambda_2$, the height of the tropopause required for a neutral atmosphere with respect to baroclinic instability as a function of N_2 is presented in fig. 3b.

Surprisingly enough, the neutralization condition changes drastically. The tropopause height has decreased further with the appearance of two possible solutions with the same value of N_2. One of these solutions implies that surface temperature increases a great deal, while the other solution would lead just to the opposite result.

The appearance of this degeneracy is a novel result that should, however, be confirmed by more complex models. Nevertheless, in the limits of validity of our calculations, it seems possible to argue that a decreasing stratospheric static stability (for instance, due to the more cooling expected by increasing CO_2 content) may lead in some circumstances to a decrease of the surface temperature.

5. Conclusions

In the present paper we have studied the consequences of possible water vapour increase on the climate radiative-convective equilibrium and its impact on baroclinic eddies heat transport.

We have shown that the stratospheric thermal state may have a profound impact on the surface temperature because of the processes which operate in neutralizing the atmosphere with respect to these eddies.

We found that a declining stratospheric lapse rate does not necessarily imply a warming of the surface temperature of the model.

Moreover, we found that two drastically different tropopause heights exist which lead to a radiative-convective and baroclinic equilibrium.

In the near future we plan to verify the findings of the present paper in a more detailed model environment.

Acknowledgements

The authors are grateful for financial support to the EC, through ENV4-CT96-0344 Grant to Camerino University, and the Italian Space Agency (ASI).

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(received May 15, 1998;
accepted September 10, 1998)