

Equable Climate Dynamics

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ABSTRACT

As the record of past climate becomes clearer, the existence of regimes has emerged as a primary characteristic of the climate system. Present climate is now known to represent one regime among others including glacial climates such as characterized recent intervals of the Pleistocene and the much warmer equable climates of the Eocene and Cretaceous. An important test of climate theory is the ability to explain the record of climate change in terms of atmosphere-ocean dynamics, and the response of the dynamics to internal and external changes.

The hypothesis is advanced that variation in the strength and extent of the symmetric circulation is an important mechanism mediating climate change. Using a simplified model we explore dynamical variables that influence the symmetric circulation and the physical processes that may have modified these dynamical variables to maintain the warm equable climates.

1. Introduction

The depth of the troposphere, 10 to 15 km, is about one part in a thousand of the earth's diameter, and yet this thin film of gas is highly effective in transferring heat between the equator and pole. Consider the equilibrium Energy Balance Climate Model (EBCM) (Budyko 1969):

$$Qs(y)\mathcal{A} - I + F[T] = 0 \quad (1.1)$$

with y denoting sine of latitude, T surface temperature ($^{\circ}\text{C}$), Q the solar constant divided by four (334.4 W m^{-2}), $s(y)$ the latitudinal distribution of solar radiation, and \mathcal{A} an absorptivity of the surface to incoming solar radiation. The three terms represent, respectively, incoming solar radiation, outgoing infrared radiation, and divergence of the heat flux carried by the atmosphere.

Taking the annual average distribution of solar radiation,

$$s(y) = 1 - 0.241(3y^2 - 1) \quad (1.2)$$

and the ice free absorptivity, $\mathcal{A} = 0.7$, together with the present atmospheric infrared flux linearization,

$$I = A + BT. \quad (1.3)$$

With $A = 211.2 \text{ W m}^{-2}$ and $B = 1.55 \text{ W m}^{-2} \text{ C}^{-1}$, gives an annual average equator temperature of 51°C and pole temperature of -58°C , in the absence of heat flux divergence, that is with $F[T] = 0$. This temperature difference of 109°C indicates the importance of dy-

namic heat flux both in reducing the equator to pole temperature drop to the present value near 40°C and in reducing the equatorial temperature to the present value near 27°C .

Parameterization of heat flux in EBCMs usually involves a relaxation toward the global mean (Budyko 1969) or diffusion the strength of which is chosen to match present gradients (Sellers 1969; Donn and Shaw 1977). These devices are useful in study of the present climate and small perturbations around it, but are inadequate for understanding climate regimes far from present conditions. An attempt to improve the heat flux parameterization by appeal to the underlying physics of the transport was made by Lindzen and Farrell (1980). Their model recognized two dynamical regimes: a symmetric circulation in the tropics and subtropics, and a baroclinic wave regime in midlatitudes. Parameterization of heat flux by the symmetric circulation was based on redistribution of heat to enforce thermal wind balance with the angular momentum conserving wind in the poleward branch of the circulation. These constraints give the extent of the symmetric circulation heat flux and the temperature gradient in the tropics and subtropics (Held and Hou 1980). The gradient at midlatitudes was obtained in the model from a baroclinic adjustment to neutrality of unstable waves.

Paleoclimate data is providing an increasingly accurate picture of climate regimes far removed from present conditions. As these regimes become more clearly delineated, they present a challenge to our theoretical understanding and an opportunity to evaluate models. One intriguing paleoclimate regime is referred to as the equable climate. Equable climates prevailed during the Cretaceous (65 to 140 million years ago)

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and Eocene (50 million years ago). During these times there is persuasive evidence for a much reduced equator to pole temperature difference compared to present values while at the same time equatorial temperature remained near today's, implying much stronger poleward heat flux than is accounted for by present climate models. Evidence for equable climate regimes has been extensively reviewed (Barron 1989; Hallam 1985; Donn 1982). The argument for equable climates relies on many intersecting datasets including evidence of flora during the Eocene that were intolerant of even episodic frost thriving on Spitsbergen at paleolatitude 79°N (Schweitzer 1980). Also during the Eocene, alligators (Dawson et al. 1976) and mammals currently restricted to tropical habitats, such as flying lemurs (McKenna 1980) were found on Ellesmere Island at paleolatitude 78°N. During the Cretaceous an annual mean polar temperature of 10°C is estimated by Parrish and Spicer (1988) from North Slope Alaska flora at paleolatitude 85°N. At that time palm trees grew near 60°N in the interior of the Asian continent (Vakhrameev 1975). Benthic ocean temperatures, now indicative of the coldest polar water, were near 15°C during the Cretaceous and Eocene (Douglas and Woodruff 1981).

The dynamic problem posed by these observations has been reviewed by Barron (1983). Reduced zonal winds implied by thermal wind balance with the observed reduced temperature gradients makes it difficult to obtain sufficient heat transport to maintain the warmth of high latitudes. Ocean currents driven by weaker wind stresses were likely to have been weak and could not account for warmth in the interior of continents where the heat flux must be carried by the atmosphere. Even the imposition of infinite effective poleward ocean heat transport failed to maintain warm values for continental interior temperature in the GCM simulations of Schneider et al. (1985). Moreover, it is important to notice that an increased greenhouse effect, due, for example, to an increase of CO₂, raises the global average temperature but cannot directly affect the equator to pole temperature difference which is the primary problem.

In model studies it is often remarked that high-latitude temperature response to climate forcing is enhanced, suggesting that increased forcing alone could effectively lower the equator to pole temperature difference. However, this result arises largely from the ice-albedo feedback and the effect of permanent ice caps on near surface temperature inversions. The equable climates were essentially ice-free according to present reconstructions so that these feedbacks could not have worked to increase polar climate sensitivity.

An explanation alternative to increased heat transport by the symmetric circulation is more effective baroclinic wave flux. Baroclinic adjustment (Stone 1978; Lindzen and Farrell 1980) models baroclinic wave transport by enforcing a temperature gradient

proportional to the tropospheric static stability. Reducing static stability to between one-half and one-third its present value would, according to this theory, decrease the midlatitude temperature gradient by the same factor, sufficient to give an average equator to pole temperature difference in accord with Cretaceous reconstructions. This is certainly a possible explanation for average polar warmth but equability seems to require, in addition, a climate without the inevitable large variations in temperature that would arise from frequent passage of cold and warm fronts associated with strong baroclinic wave transport. As we shall see, the influence of static stability is opposite in the Hadley regime. A more stable troposphere with a small lapse rate favors transport by the symmetric circulation while a less stable troposphere with higher lapse rates favors a baroclinic adjustment mechanism for maintaining the equable climate. Study of mountain flora assemblages are expected soon to provide information on lapse rates prevailing during equable climate regimes (Parrish 1990, personal communication).

Although the vertical extent of the atmosphere below the tropopause is small compared to the radius of the earth, nevertheless, considered from first principles, it is not surprising that the atmosphere efficiently transports heat. Consider the hypsometric relation for the vertical spacing between two pressure surfaces arising from hydrostatic balance and the perfect gas law (Holton 1975):

$$\delta Z = \frac{R\bar{T}}{g} \ln(p_1/p_2) \quad (1.4)$$

where R is the gas constant for air, \bar{T} is the pressure weighted average temperature between the pressure surfaces p_1 and p_2 and g is the gravitational acceleration. If one latitude of the atmosphere is heated relative to another, the pressure surface rises and a gradient of pressure increasing with height is established. Because a motionless gas is unable to sustain a shearing stress, a circulation will immediately arise to level the pressure surfaces and, consequently, also the temperature gradient. The time scale for this adjustment in the absence of rotation is given by the width of the region of imposed temperature gradient divided by the external gravity wave speed (hundreds of meters per second). From this it can be concluded that radiative driving with typical time scales of weeks is unable to maintain a temperature gradient of any significant magnitude in a nonrotating atmosphere. The surprise is that this thin film of tropospheric air is not even more efficient as a "heat pipe" between the equator and pole. The puzzle is not that equable climates occur, it is that they are not the norm.

Consider a poleward moving zonal ring of air at the tropopause, height H , which is sufficiently decoupled from the surface that it approximately conserves its angular momentum. The air accelerates to the east as it moves north obtaining a zonal wind speed $u(y, H)$

sufficient to balance the typical midlatitude temperature gradient, $d\bar{T}/dy$, according to the thermal wind relation (Holton 1979):

$$u(y, H) - u(y, 0) = -\frac{R}{f} (\partial\bar{T}/\partial y)_p \ln(p_0/p_H) \quad (1.5)$$

where f is the Coriolis parameter, p_0 is the surface pressure, p_H is the tropopause pressure, and $u(y, 0)$, the surface wind, can be assumed small relative to $u(y, H)$. This is such a familiar balanced flow in the present earth atmosphere that it seems a necessity. However, it arises from a special combination of circumstances: rapid rotation and approximate conservation of angular momentum on the poleward branch of the Hadley cell (for historical reasons symmetric circulations are commonly referred to as Hadley cells).

Whereas proxy observations of paleoclimate constrain the surface temperature, there are fewer constraints on the thermal structure of the deep interior of the atmosphere. Because of the aforementioned impasse in understanding equable climates, we explore in this work the hypothesis that these climates are characterized by a symmetric circulation extending into polar regions.

The most obvious way to obtain a global symmetric circulation would be to have the earth rotating less rapidly than is the case today. This would result in a circulation similar to that of the first few scale heights of the Venus atmosphere, where a 243-day rotation period enforces, by the mechanism discussed above, an equator to pole difference in temperature of at most a few degrees (Seiff 1983). This example illustrates the efficiency of heat transport in the absence of a strong rotational constraint. We will concentrate in what follows on the circumstances under which a global Hadley cell could arise from a combination of increased angular momentum loss in the poleward branch of the circulation and an increased height of the tropopause. A simple model of the symmetric circulation is employed and an attempt is made to obtain observable consequences of the theory to make it testable with present and future datasets.

2. The symmetric circulation model

We begin with a brief review of the symmetric circulation model of Held and Hou (1980) and Hou (1984). The Boussinesq equations for steady nondivergent axisymmetric flow on a sphere are the zonal momentum equation:

$$\nabla \cdot (\mathbf{V}u) - 2\Omega yv - \frac{uv}{a} \frac{y}{(1-y^2)^{1/2}} = F[u], \quad (2.1)$$

the meridional momentum equation:

$$\begin{aligned} \nabla \cdot (\mathbf{V}v) + 2\Omega yu + \frac{u^2}{a} \frac{y}{(1-y^2)^{1/2}} \\ = -\frac{(1-y^2)^{1/2}}{a} \frac{\partial\phi}{\partial y} + F[v], \quad (2.2) \end{aligned}$$

the equation of hydrostatic balance:

$$\partial\phi/\partial z = g\theta/\theta_0, \quad (2.3)$$

the thermodynamic equation:

$$\nabla \cdot (\mathbf{V}\theta) = \frac{\theta_E - \theta}{\tau_R} + F[\theta], \quad (2.4)$$

and the equation of continuity:

$$\nabla \cdot \mathbf{V} = 0, \quad (2.5)$$

where (u, v, w) are the zonal, meridional, and vertical components of velocity, $\mathbf{V} = (v, w)$, ϕ is the pressure, y the sine of latitude, z the height, a the earth radius, Ω the rotation rate, g the gravitational acceleration, ∇ the meridional gradient operator:

$$\nabla \equiv \left[\frac{1}{a} \frac{\partial(1-y^2)^{1/2}}{\partial y}, \frac{\partial}{\partial x} \right], \quad (2.6)$$

θ the potential temperature, θ_E the radiative-equilibrium potential temperature, θ_0 the global mean potential temperature, and τ_R the radiative relaxation time. Flux divergences are parameterized by a coefficient of turbulent mixing, ν , so that

$$F[\theta, u, v] = \nu(\partial^2/\partial z^2)(\theta, u, v).$$

Boussinesq equations permit a simple presentation of symmetric circulation dynamics; more accurate solutions for a perfect gas atmosphere in pressure coordinates require changing the hydrostatic relation. Experience has shown that physical insights gained using the Boussinesq equations are not essentially modified.

We adopt the annual mean radiative driving of Held and Hou (1980):

$$\frac{\theta_E}{\theta_0} = 1 - \delta_h \left(y^2 - \frac{1}{3} \right) + \delta_v \left(\frac{z}{H} - \frac{1}{2} \right) \quad (2.7)$$

where H is the depth of the circulation. In (2.7) δ_h is the ratio of the equator to pole potential temperature difference to the global mean temperature. In a similar way, δ_v is the ratio of the surface to height H potential temperature difference to the global mean temperature.

Boundary conditions require vanishing heat and mass fluxes at all boundaries, vanishing momentum flux at the upper boundary, linear drag at the lower boundary, and symmetry about the equator:

$$w = \frac{\partial \mathbf{V}_h}{\partial z} = \frac{\partial \theta}{\partial z} = 0 \quad z = H \quad (2.8a)$$

$$w = \frac{\partial \theta}{\partial z} = 0, \quad \frac{\partial \mathbf{V}_h}{\partial z} = \frac{C \mathbf{V}_h}{\nu} \quad z = 0 \quad (2.8b)$$

$$v = 0 \quad y = 0, 1 \quad (2.8c)$$

in which C is a drag coefficient with dimensions of velocity which is assumed constant, and ν is the coefficient of turbulent mixing, and $\mathbf{V}_h = (u, v)$.

The circulation satisfies thermal wind balance obtained from the meridional momentum equation (2.2) and the hydrostatic relation (2.3):

$$\frac{\partial}{\partial z} \left[2\Omega y u + \frac{u^2}{a} \frac{y}{(1-y^2)^{1/2}} \right] = - \frac{g(1-y^2)^{1/2}}{a} \frac{\partial}{\partial y} \left[\frac{\theta}{\theta_0} \right]. \quad (2.9)$$

Angular momentum, $M = \Omega a^2(1-y^2) + u(0, H)a(1-y^2)^{1/2}$, is conserved along streamlines of the meridional flow if the momentum flux divergence vanishes, $F[u, v] = 0$. With this assumption a vortex ring having zonal velocity $u(0, H)$ at the equatorial tropopause would, on being displaced poleward, obtain the angular momentum conserving zonal wind:

$$u_M(y, H) = \frac{u(0, H) + \Omega a y^2}{(1-y^2)^{1/2}}. \quad (2.10)$$

Accounting for momentum loss reduces this wind in the poleward branch of the circulation. Multiplying (2.1) by $(a(1-y^2)^{1/2})$ we obtain the equation for angular momentum, M :

$$\frac{1}{a} \frac{\partial}{\partial y} [vM(1-y^2)^{1/2}] + \frac{\partial}{\partial z} (wM) = F(u)a(1-y^2)^{1/2}. \quad (2.11)$$

Applying the vertical averaging operator $\overline{(\quad)} = 1/H \int_0^H (\quad) dz$ to (2.4), (2.9), and (2.11), respectively, the thermal wind, angular momentum, and thermodynamic equations, gives the dynamic equations:

$$\frac{2\Omega a y [u(y, H) - u(y, 0)]}{(1-y^2)^{1/2}} + \frac{y[u(y, H)^2 - u(y, 0)^2]}{(1-y^2)} = -gH \frac{d}{dy} \left[\frac{\bar{\theta}}{\theta_0} \right], \quad (2.12)$$

$$\frac{1}{a} \frac{d}{dy} [\overline{vM}(1-y^2)^{1/2}] = \overline{F(y)}a(1-y^2)^{1/2}, \quad (2.13)$$

$$\frac{1}{a} \frac{d}{dy} [\overline{v\theta}(1-y^2)^{1/2}] = \frac{\bar{\theta}_E - \bar{\theta}}{\tau_R}. \quad (2.14)$$

For simplicity, momentum loss in the averaged angular momentum equation (2.13) is parameterized as due solely to surface drag $\overline{F(y)} = -[u(y, 0)/\tau_c]$ with $\tau_c \equiv H/C$. The poleward and equatorward circulation is modeled as confined to upper and lower horizontal boundary layers of width δ confined near $z = 0$ and $z = H$ so that by continuity:

$$\bar{v}(y) \equiv v_u(y) + v_l(y) = 0, \quad (2.15)$$

where $(\quad)_u \equiv 1/H \int_{H-\delta}^H (\quad) dz$ and $(\quad)_l \equiv 1/H \int_0^\delta (\quad) dz$.

Recalling $F(u) = \nu(\partial^2 u/\partial z^2)$ and noting by (2.8a) that the stress vanishes at $z = H$, the boundary layer approximation requires divergence of this stress between $z = H - \delta$ and $z = H$ so that

$$F_u(y) \approx - \frac{\nu}{H} \left[\frac{u(y, H) - u(y, 0)}{H} \right] \approx - \frac{u(y, H)}{\tau_\nu} \quad (2.16)$$

where $\tau_\nu \equiv H^2/\nu$.

Fluxes of quantities X not confined to the boundaries are approximated in terms of the mass flux $V_b = \nu_u(1-y^2)^{1/2}$ as:

$$\overline{X V_b} = V_b(y)[X(y, H) - X(y, 0)]. \quad (2.17)$$

In particular the heat equation becomes:

$$\frac{1}{a} \frac{d}{dy} (V_b S) = \frac{\bar{\theta}_E - \bar{\theta}}{\tau_R \theta_0}, \quad (2.18)$$

in which the total static stability parameter $S(y) = [\theta(y, H) - \theta(y, 0)]/\theta_0$ is assumed to be constant and determined by processes external to the model.

If the surface winds are small compared to the upper branch winds and vertical advection of angular momentum can be ignored compared with horizontal advection, three equations are obtained in the variables $u(y) \equiv u(y, H)$, $V_b(y)$, and $\bar{\theta}(y)$. These are the thermal wind relation:

$$\frac{2\Omega a y u(y)}{(1-y^2)^{1/2}} + \frac{y u^2(y)}{1-y^2} = - \frac{gH}{\theta_0} \frac{\partial \bar{\theta}}{\partial y}, \quad (2.19)$$

the heat equation:

$$\frac{1}{a} \frac{\partial V_b(y)}{\partial y} = \frac{\bar{\theta}_E - \bar{\theta}(y)}{S \tau_R \theta_0}, \quad (2.20)$$

and the angular momentum equation:

$$\frac{V_b(y)}{a} \left\{ \frac{d}{dy} [u(y)(1-y^2)^{1/2}] - 2\Omega a y \right\} = - \frac{u(y)}{\tau_\nu} (1-y^2)^{1/2}. \quad (2.21)$$

With the nondimensionalization $u = \tilde{u}\Omega a$, $V_b = \tilde{V}_b(a\delta_h/S\tau_R)$, and scaling $\bar{\theta}/\theta_0$ by the radiative equator-pole difference, $\bar{\theta}/\theta_0 = \delta_h(\tilde{\theta}/\theta_0)$, we obtain a coupled set of nonlinear equations:

$$\frac{d}{dy} \left[\frac{\tilde{\theta}}{\theta_0} \right] = - \frac{1}{R} \left[\frac{2y\tilde{u}(y)}{(1-y^2)^{1/2}} + \frac{y\tilde{u}^2(y)}{1-y^2} \right] \quad (2.22)$$

$$\frac{d}{dy} \tilde{V}_b(y) = \frac{\tilde{\theta}_E - \tilde{\theta}(y)}{\theta_0}, \quad (2.23)$$

$$\frac{d}{dy} [\tilde{u}(y)(1-y^2)^{1/2}] = -\Gamma \frac{\tilde{u}(y)(1-y^2)^{1/2}}{\tilde{V}_b(y)} + 2y. \quad (2.24)$$

Tildes denoting nondimensional quantities are dropped in sequel.

Solutions to these equations include a Hadley regime lying between the equator where $V_b(0) = 0$ and a poleward termination latitude where again $V_b(y_H) = 0$, and a radiative equilibrium regime in thermal wind balance with the radiative equilibrium temperature for $y > y_H$. For $y_H < 1$, there is the additional requirement of continuity in temperature at the boundary between the Hadley regime and the radiative equilibrium regime, $\theta(y_H) = \theta_E(y_H)$.

In general, solutions are determined by two nondimensional parameters, the thermal Rossby number $R = gH\delta_h/\Omega^2 a^2$ and a measure of the relative dominance of the radiation and momentum time scales, $\Gamma = S\tau_R/\delta_h\tau_v$. However, in the inviscid limit $\tau_v \rightarrow \infty$, solutions depend only on R , and $u(y)$ obtains the angular momentum conserving profile (2.10) between the equator and y_H . Over the width of the Hadley cell a meridional circulation arises to decrease the temperature gradient from the radiative equilibrium profile to that which is in thermal wind balance with the angular momentum conserving upper-level zonal wind. This latter is considerably flatter near the equator. In the limit $\tau_v \rightarrow \infty$, Held and Hou (1980) found the poleward extent of the Hadley circulation y_H as the solution of:

$$\frac{1}{3}(4R - 1)y_H^3 - \frac{y_H^5}{1 - y_H^2} - y_H + \frac{1}{2} \ln \left[\frac{1 + y_H}{1 - y_H} \right] = 0, \quad (2.25)$$

a plot of which is shown in Fig. 1.

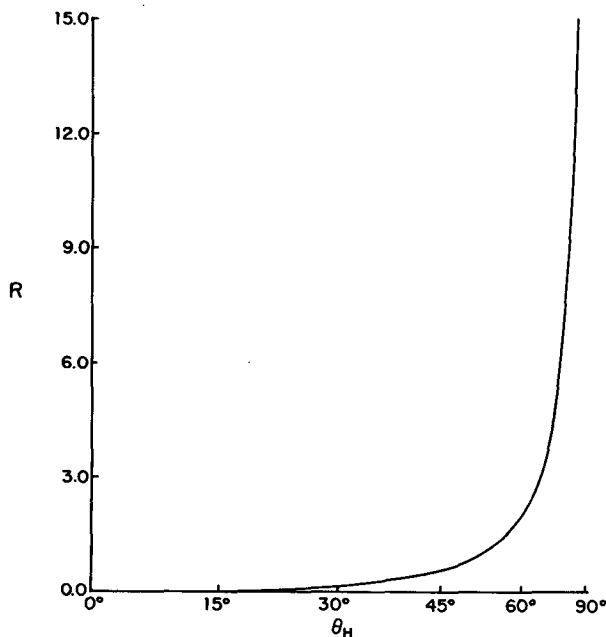


FIG. 1. The latitudinal extent of the inviscid Hadley circulation as a function of thermal Rossby number $R = gH\delta_h/\Omega^2 a^2$.

The associated temperature structure equatorward of y_H is

$$\frac{\theta}{\theta_0} = 1 + \delta_h \left[\frac{1}{3} - y_H^2 + \frac{1}{2R} \left(\frac{y_H^4}{1 - y_H^2} - \frac{y^4}{1 - y^2} \right) \right] \quad (2.26)$$

resulting in a temperature difference between the equator and $y < y_H$ of

$$\theta(0) - \theta(y) = \frac{\theta_0 \delta_h}{2R} \left[\frac{y^4}{1 - y^2} \right]. \quad (2.27)$$

The inviscid circulation must reach radiative equilibrium at y_H . Nevertheless for $R \gg 1$, except in a narrow boundary layer near the pole, the Hadley circulation effectively eliminates the temperature gradient between the equator and high latitudes. While now the annual average temperature difference between the equator and 60° is 30°C , limits of Cretaceous reconstructions give between 10°C and 20°C for this difference (Barron 1983). With $\theta_0 = 300\text{ K}$ and radiative equilibrium equator to pole temperature contrast $\delta_h\theta_0 = 100\text{ K}$, a temperature difference of 20°C at latitude 60° requires by (2.27) the value of thermal Rossby number $R = 5.6$. A high R Hadley circulation controls the surface climate of slowly rotating planets such as Venus where the lower atmosphere supports only a few degrees of temperature difference between the equator and pole; for Venus, Hou (1984) estimated the value $R = 10^3$.

Evaluating $R = gH\delta_h/\Omega^2 a^2$ for the present earth with $\Omega = 7.29 \times 10^{-5}\text{ s}^{-1}$, $a = 6.37 \times 10^6\text{ m}$, $g = 9.81\text{ m s}^{-2}$, $H = 15.0 \times 10^3\text{ m}$, and $\delta_h = 1/3$ we find $R = 0.227$. Therefore, to obtain the maximum gradient compatible with Cretaceous reconstructions an inviscid Hadley circulation with R increased by a factor of 25 is required. Examining the terms in R we note that a change in δ_h , the equator to pole radiative forcing, could arise from albedo changes associated with variation in continent positions and cloud distribution. However, the increase possible is only of order one and clearly the effect on the meridional extent of the Hadley circulation would be dominated by direct forcing of the gradient. Moreover, in model studies, decreases of δ_h were found to be insufficient to account for the reconstructed gradient (Barron 1983; Douglas and Woodruff 1981). Changes in the distribution of radiation arising from variation in orbital parameters such as the tilt axis beyond the current limits set by orbital mechanics are unlikely. It is accepted that the rotation rate of the earth was greater in the Cretaceous by approximately 3% because of tidal torques. There remains only the effective height, H , as a candidate for extending the Hadley circulation in the inviscid limit.

There are reasons for believing that the height of the tropopause may have been considerably increased in the Cretaceous. While current understanding of the ra-

diative-dynamic mechanisms responsible for variations in tropopause height is incomplete, it is known that the height is highly sensitive to surface temperature. Tropopause potential temperature increases approximately 7.5°C for each 1°C increase in surface-ocean temperature for temperatures near 30°C (Chimonas and Rossi 1987). Given that equatorial ocean temperatures possibly increased by a few degrees (Baron 1983; Douglas and Woodruff 1981), and that high-latitude oceans were certainly warmer, there is potential for a substantial rise in tropopause temperature. An increase from today's equatorial average surface-ocean temperature of 27°C to 32°C would raise the potential temperature of the tropical tropopause by 37°C , nearly doubling the present value of the static stability parameter, S , in this region. Although the rate of increase of tropopause temperature with surface temperature decreases at midlatitude temperatures, a greater difference in midlatitude temperature between equable regimes and present conditions compensates for this. For example, at 60° the annual average temperature increases from approximately 0°C at present to 20°C in the equable regime. There is a 56°C difference in potential temperature between air saturated at 20°C and air saturated at 0°C .

The effect of this increase on H depends on what we assume about the radiative equilibrium structure of the stratosphere, which in turn is determined by the distribution of absorbing and emitting gases and particulates. Increased concentrations of CO_2 in the Cretaceous (Lasaga et al. 1985) would result in cooling of the stratosphere by longwave emission, thus, tending to destabilize the stratosphere. Decreases in O_3 concentration in the lower stratosphere (Chimonas 1987) would reinforce this effect by eliminating the stabilizing influence of lower stratospheric heating (Rind et al. 1990). Taken together, for plausible paleoconcentrations of gases, there results a radiative equilibrium temperature structure without a transition to higher stability at what is now the tropopause (Vardavas and Carver 1985; Gerard 1989; Manabe and Strickler 1964). The effective height of the circulation under these circumstances would be greater than in the present atmosphere as well as being potentially more sensitive to surface temperature changes owing to the attenuation or perhaps elimination of the radiatively forced inversion in the lower stratosphere. Our knowledge of the composition of the Cretaceous atmosphere and of the effect of trace gases on O_3 concentrations is insufficient to place close bounds on these processes. However, we think the increase in surface temperature and the destabilization of the stratosphere could result in a doubling of H and in the global stability S . This factor of increase in these quantities is found between midlatitudes and the tropics today. Reconstructions of 18 000 BP lapse rates from mountain snow lines show low to middle troposphere stability approximately half the present value, a result considered firm by Broecker

and Denton (1989). Failure by GCM models to agree with this observation (Rind and Peteet 1985) suggests important processes affecting atmospheric stability are not correctly parameterized. Emergence of stability and effective height as important constraints on paleoclimate will, we hope, encourage further study of mechanisms determining these dynamical variables.

Reviewing influences on the thermal Rossby number, we conclude that change in the effective height of the circulation could result in a doubling of R . However, this increase falls far short of the factor of 25 needed to account even for the highest temperature gradient found in reconstructions of the Cretaceous. Physically this result follows from angular momentum being perfectly conserved in the poleward branch of the inviscid circulation. Thermal wind balance requires the temperature gradient rise to balance a rapidly increasing zonal wind as the flow approaches the pole resulting in steep temperature gradients in midlatitudes. However, zonal winds are found to have far from angular momentum conserving velocity; observations require effective decelerations of approximately $2 \text{ m s}^{-1} \text{ d}^{-1}$ (Palmen and Newton 1965). The implied torque continually reduces angular momentum in the poleward branch of the circulation and therefore the temperature gradient required by thermal wind balance. The observed torque arises from a combination of small scale diffusion, cumulus momentum flux (Schneider and Lindzen 1977; Palmen and Newton 1969), gravity wave drag (Palmer et al. 1986), and the net westward force arising from potential vorticity mixing by large-scale waves (Tung 1989). The extent to which a given torque decreases angular momentum in the poleward branch of the Hadley cell depends on the mass flux V_b , which in turn is set for a given radiative driving by the product of the static stability and the radiative relaxation time scale, $S\tau_R$. This factor is irrelevant in the case of a completely inviscid circulation because no matter how slowly such a circulation proceeds to the pole it conserves angular momentum and requires, for thermal wind balance with finite H , a divergent temperature gradient. It is the inability of the radiative driving to supply this gradient that terminates the circulation at y_H .

We wish to explore dynamics in which the upper branch of the Hadley circulation proceeds so slowly that angular momentum sinks reduce the zonal wind so that it is in thermal wind balance with reconstructions of the Cretaceous temperature gradient. The important nondimensional parameter is $\Gamma = S\tau_R/\delta_h\tau_\nu$ in the angular momentum Eq. (2.24). This parameter is set to zero in the inviscid calculations. Terms in Γ for present conditions are $S = 1/8$, $\tau_R = 20 \text{ d}$, $\delta_h = 1/3$, and $\tau_\nu = 15 \text{ d}$. This last is a momentum time scale sufficient to reduce a 30 m s^{-1} jet by $2 \text{ m s}^{-1} \text{ d}^{-1}$. In the model the entire burden of decreasing the zonal flow is born by vertical diffusion but it is understood that this is a simplified parameterization for the complex of mo-

mentum transports described above. Putting in the numbers for present conditions we find $\Gamma = 0.5$.

Because of the increase in average surface temperature, which for reasons discussed above increases the static stability, we allow for a doubling of S . The Cretaceous atmosphere is often considered to have contained from 6 to 12 times the present CO_2 (Lasaga et al. 1985) and because of increased average surface temperature would likely have contained more water vapor. There are other trace gases in addition to CO_2 and H_2O . For example, CH_4 and N_2O are important today and given our lack of information on concentration of these gases during the Cretaceous, we are at liberty to speculate that abundances may have been much higher. To be specific, we allow a doubling of the radiative time scale τ_R . It is within the bounds of current observations to half τ_ν and it is likely that at least cumulus momentum transport increased substantially over the large warm oceans and inland seas of the Cretaceous. Taken together, these factors result in an eightfold increase in the radiative-diffusive time scale to $\Gamma = 4.0$.

While we have attempted to argue physically for the plausibility of these changes influencing the value of Γ it is not our purpose, nor is it possible at this stage, to obtain firm values. Our purpose is to construct a well-posed physical model for a symmetric circulation regime theory for equable climates. Having constructed

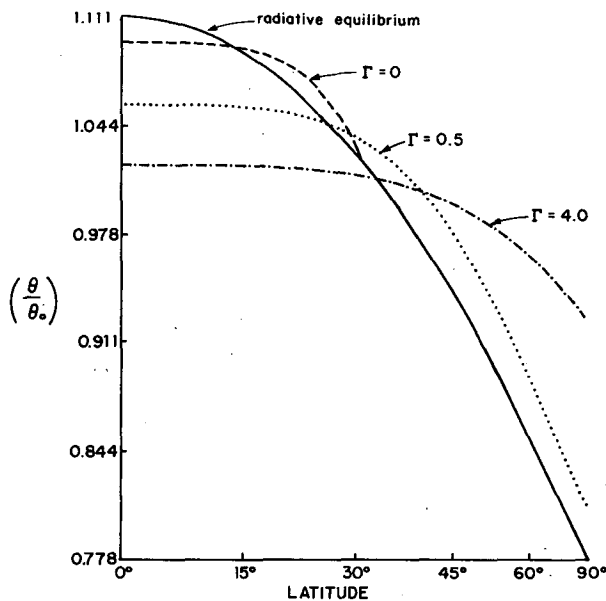


FIG. 2. Temperature normalized by its global average as a function of latitude for tropopause height $H = 15$ km. The solid line is the radiative equilibrium temperature. The dash line is temperature for the inviscid value of the radiative-diffusive parameter $\Gamma = 0$; the dot line is temperature for the present climate value $\Gamma = 0.5$; and the dash-dot is temperature for the equable climate value $\Gamma = 4.0$. There is a 20°C temperature difference between ordinate ticks for $\theta_0 = 300$ K.

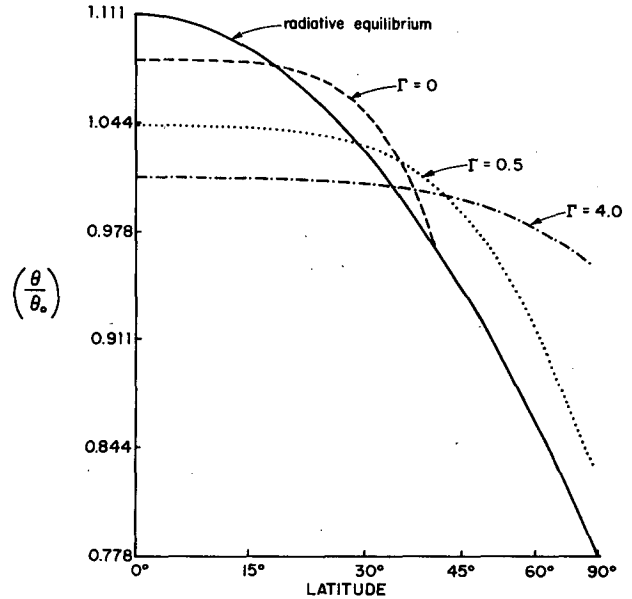


FIG. 3. Temperature normalized by its global average as a function of latitude for tropopause height $H = 30$ km. The solid line is the radiative equilibrium temperature. The dash line is temperature for the inviscid value of the radiative/diffusive parameter $\Gamma = 0$; the dot line is temperature for the present climate value $\Gamma = 0.5$; and the dash-dot is temperature for the equable climate value $\Gamma = 4.0$. There is a 20°C temperature difference between ordinate ticks for $\theta_0 = 300$ K.

such a theory we can then rationally evaluate data and theoretical constraints that bear on it.

Figure 2 shows θ/θ_0 , the temperature normalized by its global average, as a function of latitude with circulation height $H = 15$ km and for the inviscid circulation value ($\Gamma = 0$), the present value ($\Gamma = 0.5$), and the Cretaceous value ($\Gamma = 4.0$). Figure 3 shows similar results for $H = 30$ km. Absolute temperature is not predicted by the model but taking the convenient value $\theta_0 = 300$ K results in a 100 K radiative equilibrium equator to pole temperature difference so that each increment in Figs. 2 and 3 corresponds to 20°C . The case $\Gamma = 4.0$, $H = 30$ km in Fig. 3, which results in an equator-to-pole temperature difference of 16°C , is compatible with the reconstructed Cretaceous, perhaps warmer by a few degrees than required by Parrish and Spicer (1988). While no account has been taken in this calculation of the heat flux arising from baroclinic waves, this flux would decrease temperature gradients near the pole just as it does in the present atmosphere, where baroclinic wave fluxes exceed symmetric circulation fluxes at high latitudes leading to formation of a thermally indirect Eulerian circulation. Our results suggest that in equable climates wave fluxes are confined to even higher latitudes, leaving most of the atmosphere under the influence of the symmetric circulation. This circulation can be viewed as an extension to large Γ of the Hadley cell obtained in a global

nonlinear two-dimensional simulation by Hou et al. (1990).

Zonal winds for the example in Fig. 3 are shown in Fig. 4. The zonal wind maximum is displaced poleward in the Cretaceous simulation with $\Gamma = 4.0$, and its strength is comparable to present annual average jet velocity. Instability on this jet would have provided abundant precipitation to high latitudes given the absolute humidities expected at these temperatures.

A variable of interest for comparison with reconstructions of Cretaceous circulation is the surface wind, particularly the latitude of transition between low-latitude easterlies and high-latitude westerlies. The distribution of surface winds is related among other things to the wind stress that drives the Sverdrup circulation of the oceans. Surface winds are obtained diagnostically from the averaged angular momentum Eq. (2.13). Assuming $u(y, H) \gg u(y, 0)$,

$$u(y, 0) = -\frac{\tau_c}{a(1-y^2)^{1/2}} \frac{d}{dy} \times [V_b u(y, H)(1-y^2)^{1/2}]. \quad (2.28)$$

The surface wind given by (2.28) is proportional to the divergence of the average momentum flux. Because this flux vanishes at the equator and pole, the integrated surface torque also vanishes.

Surface winds for $\tau_c = 5$ d for the examples in Fig. 3 are shown in Fig. 5. In the Cretaceous case with $\Gamma = 4.0$ and $H = 30$ km weak surface easterlies prevail in a broad trade wind belt extending from equator to 41° with substantially greater velocity in the surface westerlies poleward of this latitude. Notice that global angular momentum conservation requires relatively enhanced surface westerlies to balance the torque exerted by the extended trade wind belt. This pattern of

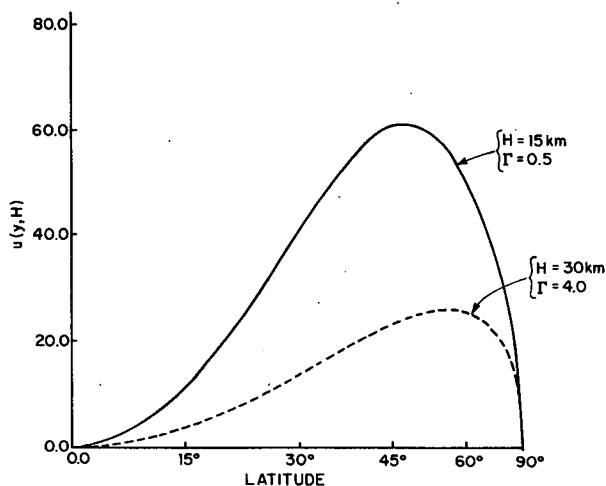


FIG. 4. Upper-level zonal wind $u(y, H)$ (m s^{-1}) as a function of latitude for present climate parameter values; $H = 15$ km, $\Gamma = 0.5$ (solid), and for equable climate parameter values; $H = 30$ km, $\Gamma = 4.0$ (dash).

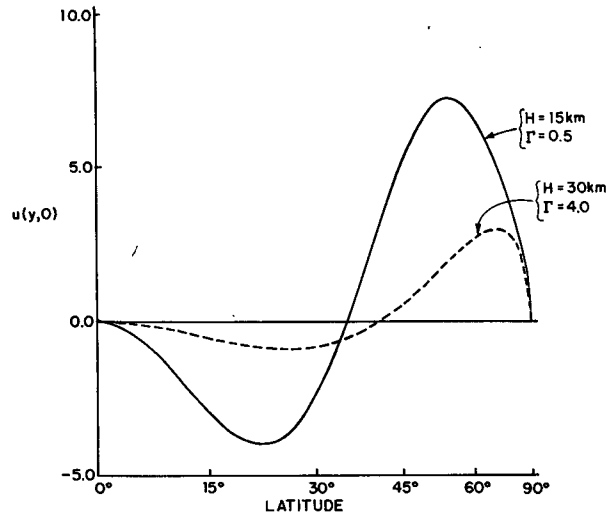


FIG. 5. Surface wind $u(y, 0)$ (m s^{-1}) as a function of latitude for present climate parameter values; $H = 15$ km, $\Gamma = 0.5$ (solid), and for equable climate parameter values; $H = 30$ km, $\Gamma = 4.0$ (dash).

surface winds constitutes a prediction of the theory that can be tested using proxy data. Examples of such proxy datasets include eolian sediments and floral indicators of precipitation distribution that provide indications of wind patterns.

3. Discussion

The earth appears to support three distinct climate regimes. Examples of these regimes are given by the 18 000 BP glacial climate, the present climate, and the equable climate that characterized the Cretaceous and Eocene. Because orbital parameters today are nearly identical to those of 18 000 BP and equable climates spanned many cycles of orbital variation, an explanation for the maintenance of these climate regimes must lie in the internal dynamics of the atmosphere-ocean system. While the role of heat transport by the oceans can be significant, in equable climates evidence for warm continental interiors argues strongly for fundamental alteration of the function of the atmosphere in transporting heat.

The present climate regime includes a symmetric circulation extending to about 30° latitude with a region dominated by baroclinic waves at higher latitudes. While it is conceivable that equable climates resulted from more active baroclinic wave transport, we concentrated in this work on the possibility that these climates resulted from transition to a dynamic regime dominated by a Hadley cell extending to polar latitudes. Because latitudinal temperature contrasts in a gaseous atmosphere result in latitudinal pressure gradients, and in the absence of motion a gas is not able to support a shearing stress, one might expect to find small temperature differences at the surface of planets with atmospheres. Indeed this is the case for Venus. It is only

the peculiar circumstance of approximate angular momentum conservation in a rapidly rotating and nearly inviscid atmosphere that allows the Coriolis force to support a pressure gradient and its attendant temperature gradient. Changes in the composition of the atmosphere and in its interaction with boundaries that result in enhanced angular momentum loss in the poleward branch of the Hadley circulation reduce the latitudinal temperature gradient. Sufficient loss of angular momentum requires the temperature gradient to approach that of a nonrotating planet. We argue this is the case in equable climates. Parameters determining the temperature structure are the height of the tropopause and $\Gamma = S\tau_R/\delta_h\tau_v$, which is controlled for fixed radiative driving by the ratio of the product of the radiative time scale and the total static stability to the momentum sink time scale. The most likely mechanism for decreasing angular momentum is not a simple increase in momentum sinks but alteration of the radiative-convective structure of the atmosphere and in particular the vertical stability and radiative time scale.

Mechanisms for raising global mean temperature, for instance increasing CO_2 or insolation, do not address the geographic and temporal constancy of equable climates. The delicacy and tropical nature of high-latitude flora and fauna in equable climates argues for a highly restricted range of temperature variation. This equability in variance is not characteristic of climates maintained by baroclinic wave transport in which the passage of warm and cold fronts results in large and rapid temperature changes. Equability results directly from a global Hadley circulation which responds on short dynamic time scales to enforce a nearly constant temperature over global space scales.

The model predicts that the Hadley circulation would collapse on the radiative time scale of order one month if the radiative-convective structure maintaining it were abruptly altered, for instance by a decrease in the effective height and stability of the tropopause arising from a sudden change in the radiatively important constituents of the upper troposphere and stratosphere. Then high latitude temperature maintained far from its radiative equilibrium value by the Hadley circulation would be exposed to a far colder climate on time scales likely to be insufficient for evolutionary adaptation. Alternatively, variation in these constituents on geological time scales would result in slow transition from an equable to a glacial regime. Paleoclimate reconstructions suggest that rapid transitions are frequently superposed on secular trends as would be expected if both gradual and occasional abrupt variations occur.

We have attempted to provide an explanation for the large-scale structure of the global circulation in equable climates. Obviously the real climate would have its own peculiar localized weather patterns superposed on the Hadley cell. For example, the high-latitude baroclinic jet would likely support extensive

precipitation owing to baroclinic wave generation in the presence of high, absolute humidity consistent with high temperatures in these regions. In addition, we must expect monsoonal and Walker type circulations to result from seasonal changes in land and ocean heating.

An increase in the efficiency of heat transport to the poles as the atmosphere becomes more opaque to thermal infrared radiation might be expected to delay the onset of a runaway greenhouse by removing tropical heat to higher latitudes where it could be radiated to space more effectively. This mechanism is analogous to the cumulus heating theory of Lindzen et al. (1980) in which a more efficient vertical transport was found to delay the greenhouse effect by removing heat from lower regions of the troposphere to higher regions where radiation to space is more effective.

Paleodata bearing on the radiatively active constituents of the Cretaceous and Eocene atmospheres are of importance to the theory of equable climates advanced in this work. For instance, the results of Lasaga et al. (1985) are inferential and need to be corroborated by more direct proxy data. Mountain flora data that constrain lapse rates are of particular importance because these would discriminate between the baroclinic adjustment and symmetric circulation theories. Eolian deposits, ocean current and upwelling indicators, and precipitation distribution data from flora assemblages can be interpreted to constrain surface and upper-level winds. A poleward shift of the jet stream with strong high-latitude surface winds coupled with a broad poleward extension of the trade winds is predicted by this theory.

A consistent GCM study of the equable climate regime would have to include a correct model of the radiative, chemical, and dynamic interactions that mutually determine the vertical stability and tropopause height. In addition the radiative time scale and the dynamics of the symmetric circulation must be correct and sufficient vertical resolution included to accommodate the expected increase in tropopause height. A one-dimensional radiative-chemical model of the vertical thermal structure with parameterized dynamics may be a more realistic next step in modeling equable climates.

Taking a wider view, the theory of climate change advanced here is that variation of atmospheric constituents controls climate indirectly. The direct effect of a change in radiatively active gases is to alter the vertical temperature structure of the atmosphere. The climate is not profoundly altered by this change alone; the major climate effect results from alteration of poleward heat transport arising from changes in the global circulation regime that attend the changes in vertical structure.

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