Evaporation-Limited Tropical Temperatures as a Constraint on Climate Sensitivity

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ABSTRACT

Studies of paleoclimate and modern observations indicate that evaporative effects limit thermal response in equatorial regions. We develop a latitude-resolved, steady-state energy balance model which incorporates the effect of an evaporative constraint on the variation of equatorial temperature with solar luminosity. For a diffusive model of surface heat transport the constraint requires the diffusion coefficient to vary with insolation. We find that the movement of the iceline with insolation is four times larger than in standard energy balance models with a constant thermal diffusion coefficient. This is a consequence of the global energy balance which forces temperature changes to occur at high latitudes when they are evaporatively buffered at the equator. Nonlinear temperature–ice albedo feedback at high latitudes then amplifies the response leading to greater sensitivity in the vicinity of current climate.

1. Introduction

Latitude-resolved energy balance models (EBMs) pioneered by Sellers (1969) and Budyko (1969) have proven useful in recent years for assessments of climate sensitivity, particularly the feedback between temperature and albedo associated with freezing of bright, high-latitude icesheets (Schneider and Gal-Chen, 1973; Held and Suarez, 1974; North, 1975; Ghil, 1976; Lindzen and Farrell, 1977, 1980; Hartmann and Short, 1979; Suarez and Held, 1979; Held *et al.*, 1981). Comprehensive reviews of such models were given by Schneider and Dickinson (1974) and more recently by North *et al.* (1981). Typically, EBMs formulate equations for zonally-averaged surface temperature at the Earth's surface $T(x)$, based on the local balance between absorbed insolation, infrared cooling to space and meridional (poleward) heat transfer by the atmosphere and oceans, where $x = \sin\phi$ is a coordinate proportional to surface area from the equator to latitude ϕ.

Despite the complexity of transport in the atmosphere and oceans, standard EBMs reproduce reasonably well the currently-observed annual mean $T(x)$ distribution by assuming total heat flux across latitude circles is given by a simple Fourier heat conduction law of the form $F(x) = -2\pi R^2 D(1 - x^2)dT/dx$, where R is the planetary radius, D is a phenomenological thermal diffusion coefficient and the factor $(1 - x^2)$ arises from the sphericity of the Earth. The coefficient D in energy balance models serves as a surrogate for specific knowledge of dynamical energy transport which the model cannot predict. Stone (1978) showed that this is due to radiative constraints on dynamical energy transports on a spherical planet. Most models take D (and an equivalent coefficient for meridional water vapor transport in some models) independent of latitude, although some recent work has been done with models which prescribe $D = D(x)$ from predictions of atmospheric circulation models at the current climate (Held *et al.*, 1981). More importantly, all current EBMs make assumptions equivalent to holding D [or $D(x)$] constant during changes in external forcing, e.g., changes in the solar constant. This assumption we hereafter call the Transport Limit.

The canonical EBM problem is to find the $T(x)$ distribution, or "sensitivity," of the global climate when the forcing changes. The well-known "White Earth catastrophes" from runaway temperature–ice–albedo feedback exhibited by the original Sellers–Budyko models, in which decreases in solar output of only a few percent freeze the entire Earth, are much less severe in current formulations, owing mainly to more realistic albedo prescriptions (Hartmann and Short, 1979). On the other hand, standard EBMs produce results inconsistent in many ways with climatic records:

1) They are too insensitive in the vicinity of current climate to explain the large movement of the iceline during the last (Wisconsin) Ice Age (North *et al.*, 1981).

2) They produce too much equator-to-pole temperature difference to explain reconstructed ice-free climates like the Cretaceous 65–140 M year B.P. (Barron *et al.*, 1981).

3) They do not adequately describe the weak seasonal and long-term variability of the tropics (Lindzen and Farrell, 1977, 1980; Warren and Schneider, 1979).

In this paper we suggest an alternate EBM formulation based on a constraint on tropical temperatures associated with evaporative (latent) heat flux as the dominant term in the equatorial surface energy balance. Because evaporation rates depend exponentially on surface temperature, small changes in surface temperature can balance large changes in heating in the tropics, as recognized by Priestly (1966) and more recently by Newell (1979), who suggested equatorial temperatures are buffered against large variations by evaporation. Newell and Dopplick (1979, 1981) further suggested that such damping might limit the global warming from atmospheric carbon dioxide increases, but they did not explore the impact of tropical evaporation damping in the context of a latitude-resolved climate model. Moreover, some recent studies which addressed the weak global response implications of Newell and Dopplick (Ramanathan, 1981; Kandel, 1981) also used horizontally-averaged models, so they, too, did not address surface evaporative damping in a latitude-resolved context. A basic feature of our analysis is that the latitudinal-dependence of the surface energy balance terms is critical.

In what follows, we formulate a Tropical Evaporation Constraint (TEC) as an alternative to the Transport Limit in an EBM, compare the results of numerical solutions to the energy balance equation with the two approximations and discuss the implications of the model in terms of explicit formulations of poleward heat flux by the atmosphere and oceans. Constraining the equatorial temperature in this way increases the sensitivity of climate worldwide (in contrast to the Newell-Dopplick reduced CO_2 global warming predictions), produces a more realistic distribution for $T(x)$ during the Cretaceous, and inherently produces weak variability of the tropics.

2. Formulation of transport- and TEC-limited energy balance models

a. Standard energy balance model

We use a conventional formulation (North *et al.*, 1981) to model the steady state distribution of annual mean surface temperature $T(x)$ today. It includes heating by sunlight $S(x)$ at the top of the atmosphere, where a fraction $a(x, T)$ is actually absorbed ($1 - a$ is the albedo), cooling by infrared radiation $I(T)$, and a Fourier heat conduction law for meridional heat flux with a latitude-independent thermal diffusion coefficient D,

$$\frac{1}{2\pi R^2}\frac{dF}{dx} = -D\frac{d}{dx}\left[(1-x^2)\frac{dT}{dx}\right] = a(x)S(x) - I(T). \quad (1)$$

Generally, $S(x)$ is derivable from astronomical considerations in terms of the solar constant and spin axis tilt (see, e.g., Hoffert *et al.*, 1981), and $a(x)$ is known for the present climate from satellite observations (Ellis and Vonder Haar, 1976). This data also supports the usual approximation that longwave flux to space is approximately a linear function of the surface temperature of the underlying air column (Warren and Schneider, 1979). The feature of the albedo prescription generating temperature–ice albedo feedback is a discontinuous switch to low absorptance (highly-reflective) values typical of ice and snow whenever surface air temperature falls below some reference value T_s associated with surface freezing (Budyko, 1969; Sellers, 1969; North, 1975; etc.). More realistic results are obtained when the ice-free part of the planetary absorptance also decreases with increasing latitude corresponding to increased backscattering of insolation to space at larger solar zenith angles (Hartmann and Short, 1979). For simplicity, and in view of the many well-known uncertainties involved, the effect of cloudiness feedback on both planetary albedo and longwave flux is neglected. Based on these considerations, we model insolation, absorptance and longwave flux to space by

$$S(x) = Q(s_0 - s_2x^2), \quad (2)$$

$$a(x, T) = a_0 - a_2x^2, \quad T > T_s, \quad (3a)$$

$$a(x, T) = b_0, \quad T < T_s, \quad (3b)$$

$$I(T) = A + BT, \quad (4)$$

where Q is ¼ of the solar constant.

Mathematically, Eq. (1), describing energy balance, is a second-order ordinary differential equation for $T(x)$ that is nonlinear because of temperature–ice albedo feedback. Given the symmetrical nature of the source terms (a hemispherically-symmetric planet is assumed in the albedo formula), the condition of local energy balance establishes boundary conditions on the solution for $T(x)$. At the equator at $x = 0$:

$$dT/dx = 0. \quad (5a)$$

At the pole at $x = 1$:

$$\frac{dT}{dx} = \frac{[a(x, T)S(x) - I(T)]}{2D}. \quad (5b)$$

A numerical code was written to solve the associated two-point boundary value problem for $T(x)$. As indicated earlier, observations and theory provide estimates of the coefficients appearing in Eqs. (1)–(5), but they can be tuned within allowable ranges to match $T(x)$ today. In particular D is an adjustable parameter, since the poleward heat transfer model is *ad hoc*. In Table 1 we list values of the coefficients adopted here. They result in an iceline at $x = 0.95$, equatorial temperature $T_{eq} = 302$ K (29°C) and polar temperature $T_{pole} = 259$ K (−14°C). Despite its simplicity, the model reproduces $T(x)$ rather well, which is the principal reason for accepting the diffusive model for poleward heat flux.

Figure 1 shows the total heat flux across a latitude circle versus latitude computed in this way for the current climate corresponding to $D = 0.623$ W m^{-2} K. Note that the flux peaks at $x \approx 0.58$ ($\phi \approx 35°$) in agreement with Stone's (1974) estimates, and is roughly twice the magnitude of poleward fluxes measured in the atmosphere (Oort and Vonder Haar, 1976; Trenberth, 1979). Presumably, the oceans carry the remainder.

A fundamental assumption of sensitivity studies using standard Transport-Limited EBMs is that D remains constant at its currently-calibrated value while a new $T(x)$ distribution is found for a change in, say, Q. Note that there is no freedom in this formulation to constrain the equatorial temperature $T(0)$, since the entire $T(x)$ distribution is determined for D constant. We next show $T(0)$ is constrained by evaporation cooling at the surface in ways which are inconsistent with the Transport Limit.

b. *The TEC model for climate change with varying insolation*

To derive the effects of evaporation on tropical temperatures when solar forcing varies we make use of observations that in the tropics today: 1) air and sea surface temperatures are nearly identical (Reinking and Barnes, 1981) and relax to small air–sea temperature differences on timescales of less than a month (Hoffert *et al.*, 1980); and 2) the flux of latent heat dominates the exchange of energy between the water and air (Newell, 1979).

Consider the energy balance at the surface of the tropical ocean given by the heat balances of water

TABLE 1. EBM coefficients tuned to current climate.

Insolation	$Q_0 = 334$ W m^{-2}	$s_0 = 1.246$ $S_2 = 0.738$
IR radiation	$A = 205$ W m^{-2}	$B = 2.23$ W (m^2 °C)$^{-1}$
Co-albedo	$a_0 = 0.785$ $b_0 = 0.380$	$a_2 = 0.263$ $T_s = 263.15$ K
Thermal diffusivity	$D = 0.623$ W m^{-2} K^{-1}	
Earth radius	$R = 6.37 \times 10^6$ m	

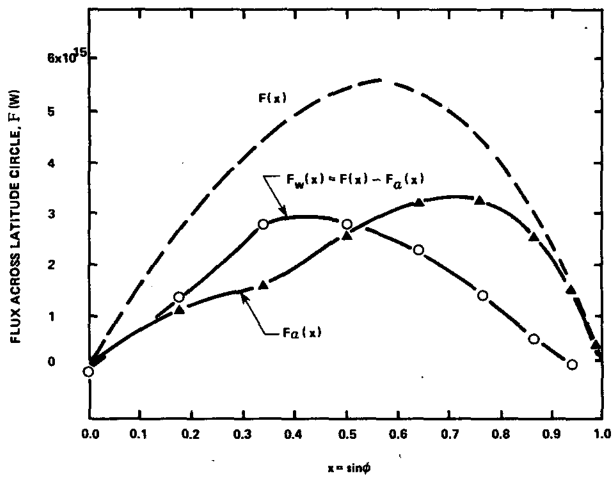

FIG. 1. Poleward heat flux with latitude for the present climate. $F(x)$ is the total heat flux across latitude circles calculated from the EMB. The total atmospheric transport $F_a(x)$ comes from averaging Northern and Southern Hemisphere data from Oort and Vonder Haar (1976) and Trenberth (1979). The implied oceanic transport $F_w(x)$ is computed as a residual, $F_w = F - F_a$.

and air, characterized by temperatures T_w and T_a. For simplicity, the effect of land surfaces is omitted. Under steady state conditions, the derivatives per unit surface area of horizontal heat fluxes in the atmosphere and oceans are

$$dF_a/dA = S_a - I + I_{wa} + H_{wa} + LE_{wa}, \quad (6a)$$

$$dF_w/dA = S_w - I_{wa} - H_{wa} - LE_{wa}, \quad (6b)$$

where subscripts a and w stand respectively for air and water, S_a and S_w are the insolation fluxes absorbed by the atmosphere and oceans, I_{wa}, H_{wa} and LE_{wa} are the net vertical fluxes from water to air by longwave radiation, sensible heat transfer and latent heat transfer across the surface boundary layer, respectively. As before, I is the net radiative loss to space. A differential surface area of the spherical Earth is $dA = \pi R^2 \cos\phi d\phi = 2\pi R^2 dx$. If equations (6a) and (6b) are summed the interreservoir fluxes cancel to reproduce Eq. (1).

In principle, however, the temperature of a reservoir will adjust until cooling balances heating. In Eq. (1) losses from radiation to space depend linearly on T, but the evaporative flux from the surface depends exponentially on temperature. Using the usual formulations for latent heat flux from a water-saturated surface with relative humidity r_a on the air side and constant temperature $T_{eq} \approx T_a \approx T_w$ gives

$$LE_{wa} \approx 1.7 \times 10^4 (1 - r_a) q_{sat} \text{ [W m}^{-2}\text{]}, \quad (7)$$

where $q_{sat}(T)$ is the saturation mixing ratio of water in air, and the constant of proportionality is evaluated for a mean windspeed of 3.6 m s^{-1}, and the GATE-recommended bulk exchange coefficient of the tropical boundary layer (Reinking and Barnes, 1981). The saturation mixing ratio at the calibrated "current climate" equatorial temperature of 302 K is 0.025 kg water vapor/kg air. Since q_{sat} is proportional to the saturation vapor pressure, its temperature-variation is expressible by the Clausius–Clapeyron equation (Hess, 1959, p. 49),

$$\ln[q_{sat}(T)] = \text{constant} - L/RT, \qquad (8)$$

where L and R are the latent heat of vaporization and gas constant of water vapor respectively. In the tropics where the climatological mean specific humidity at the surface is $q_a = 0.015$ (Houghton, 1977, p. 167) corresponding to $r_a = 0.60$, the flux of latent heat is $LE_{wa} = 170$ W m^{-2}. However, since $LR^{-1} = 5270$ K, LE_{wa} decreases by about 50% when surface temperature decreases only 10 K from its equatorial value of 302 K.

We now estimate the size of other terms in equation (6a) to show that LE_{wa} dominates heat loss from the tropical ocean. The flux of sensible heat depends on the air–sea temperature difference and has been measured to be some two orders-of-magnitude smaller than latent heat flux over tropical oceans (Reinking and Barnes, 1981), so H_{wa} can be neglected. From the results of the tuned model shown in Fig. 1, we find the total loss by heat transport at the equator to be $dF/dA \approx 60$ W m^{-2}, with an estimated oceanic component of $dF_w/dA \leqslant 40$ W m^{-1}.

The net radiative flux at the surface is the difference between radiation from the ocean surface and backwarming by the atmosphere, $I_{wa} = \sigma(\epsilon_w T_w^4 - \epsilon_a T_a^4)$. The optically thick ocean radiates very nearly like a blackbody with $\epsilon_w = 0.94 \pm 0.02$ (Sellers, 1965, p. 41), but ϵ_a must be calculated from a detailed model because the atmosphere is transparent at some frequencies. A commonly cited value for the global average air emissivity at the surface is $\epsilon_a = 0.70$ (Paltridge and Platt, 1976, p. 8). However, using a detailed radiative transfer model, one of us (W.W.) finds that under the conditions of high specific humidity in the tropics the e-continuum of water vapor becomes virtually opaque in the otherwise optically thin 8–12 μm window, as well as in the 12–18 μm interval already partly blocked by carbon dioxide absorption, so that $\epsilon_a = 0.92 \pm 0.02$. Thus, in equatorial zones both air and water radiate nearly like black bodies at the same temperature with a small residual flux, $I_{wa} = 9 \pm 18$ W m^{-2}. Similar results were obtained recently by Kiehl and Ramanathan (1982) in a study of carbon dioxide radiation effects. At high latitudes, they found that the increase of atmospheric water vapor with rising surface temperature contributes strongly to increased radiative backwarming, but in the tropics the effect is much less because water vapor has already become thick. This would seem to contradict, in the tropics, the earlier findings of Ramanathan (1981) from a horizontally-averaged model that back radiation from water vapor feedback compensates for the strong cooling constraint of latent heat loss. The loss rates, from the foregoing considerations, sum to a total cooling rate of 220 W m^{-2}, which agrees well with estimates of solar heating of the tropical ocean (Budyko, 1978). Clearly, evaporative heat flux LE_{wa} dominates oceanic cooling in the tropics.

To estimate effects of changing insolation we assume that solar heating is entirely balanced at the equator by evaporative cooling and that heating is proportional to Q, so that $\Delta S_w/S_w = \Delta Q/Q_0 = \Delta LE_{wa}/LE_{wa}$. If air and water temperatures remain equal and windspeeds and relative humidities do not change, then $\Delta LE_{wa}/LE_{wa} \approx \Delta T d[\ln(q_{sat})]/dT$. As a consequence of these assumptions, heat transfer in the surface boundary layer must adjust to maintain nearly equal air and sea temperatures and poleward heat transfer by the atmosphere/ocean system must adjust to satisfy a constraint on the equatorial temperature.

We now return to Eq. (1) to study the sensitivity of climate with respect to changes in solar output. In the standard Transport-Limited sensitivity studies D is held constant while Q varies. Here we enforce the evaporative constraint on tropical ocean temperature, and assume air and sea temperatures remain equal near the equator. From our previous discussion, this gives the TEC (Tropical Evaporation Constraint) boundary condition

$$T_{eq}(Q) \approx 302 + 17.31 \ln(Q/Q_0) \text{ [K]}. \qquad (9)$$

This condition can only be satisfied by allowing the thermal diffusion coefficient to be a function of Q. Mathematically, $D(Q)$ is an adjustable parameter of the problem. Physically, variation of D corresponds to an adjustment in the dynamics or energetics associated with meridional transport accompanying the change in solar forcing.

c. *Limits of the TEC model*

In general the TEC boundary condition [Eq. (9)] cannot be satisfied for all values of Q. We describe solutions to the TEC model in the next section, but limits on their range of validity can be established analytically. Relative to the Transport-Limited models, equatorial temperature changes less rapidly with Q in the TEC model. Since in the TEC model T_{eq} does not fall as rapidly with decreasing Q, $D(Q)$ must decrease to reduce heat loss. Similarly, when Q increases, $D(Q)$ rises to inhibit equatorial warming. Outside a limited range for Q, the TEC model requires diffusive transport to act in a countergradient (refrigeration) mode, with $D < 0$, which is physically implausible. (Starr, 1968, gives examples of locally negative eddy diffusion coefficients in the atmosphere and oceans, but globally negative values are ruled out on thermodynamic grounds.) As Q and D decrease, the model resembles one in local radiative balance.

If Eq. (9) required a temperature above that in radiative balance, the solution would require $D < 0$. Q_{min} can be obtained by substitution for T from Eq. (9) into Eqs. (1)–(3a) with $D = 0$:

$$A + BT_{eq}(Q_{min}) = Q_{min}a_0s_0. \qquad (10)$$

In the limit of large Q, as D increases, the model becomes isothermal. If we express the source terms and temperature in Eq. (1) as expansions in Legendre polynomials, so that $T(x) = \Sigma\ T_iP_i(x)$, then in the isothermal limit $T(x) = T_0$. This gives a limit for Q_{max}:

$$A + BT_{eq}(Q_{max}) = Q_{max}(as)_0, \qquad (11)$$

where $(as)_0$ is the leading coefficient in the expression of $a(x)s(x)$ as an expansion in Legendre polynomials. If Q exceeded Q_{max} the equatorial temperature would have to be lower than the average temperature, which again would require $D < 0$. (At this limit D switches sign from positive to negative infinity as Q increases through Q_{max}.) For the coefficient values adopted here we find the limits $0.8 < Q/Q_0 < 1.18$.

Obviously, the solutions become unphysical long before these limits, since the approximation of using only a boundary condition at the equator to mimic evaporative effects becomes unreliable. At the high Q extreme, on an isothermal Earth, evaporative effects are equally important at all latitudes. On a low Q earth, evaporation might be unimportant, even at the equator. However, we believe the TEC approximation is useful near current conditions, since evaporative effects are important, but dominant primarily in the tropics.

3. Numerical solutions for TEC and transport-limited models

To study the effects on climate sensitivity, we calculated numerical solutions to both the Transport-Limited and TEC EBMs as a function of variable insolation. Figure 2 illustrates the sensitivity of the iceline at $x = x_s$ to changes in solar forcing. As is well known (Budyko, 1969; Sellers, 1969; etc.), the nonlinearity introduced by temperature–ice albedo feedback allows the standard, Transport-Limited EBM to exhibit multiple solutions near $Q/Q_0 = 1$. Those solutions between $0.87 \leqslant Q/Q_0 \leqslant 1.3$, where $0.0 \leqslant x_s \leqslant 0.58$, in which $dx_s/dQ < 0$ are unstable (Held and Suarez, 1974). They would not persist in an evolving situation. In the TEC model the variation of the ice line with changing Q is much steeper near the current climate: at $Q/Q_0 = 1$, dx_s/dQ (TEC) $= 4.5dx_s/dQ$ (Transport-Limited). This is the principal result of our study. If horizontal transport must adjust during changes in climate, then sensitivity can be altered. In the case described here, transport increases with enhanced solar forcing to maintain the evaporative limit on equatorial temperature. Larger heat flux to the poles magnifies the movement of the iceline.

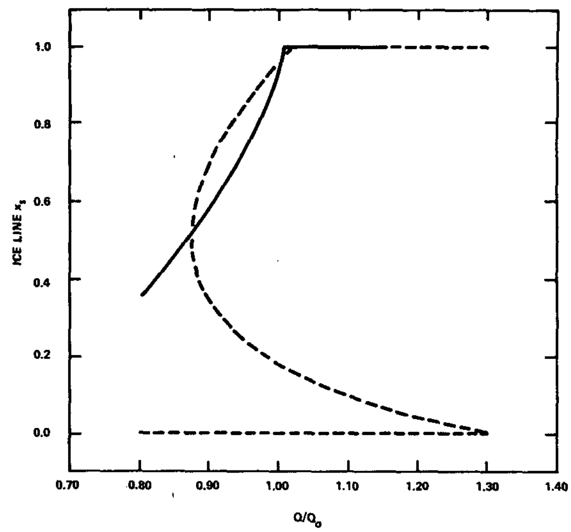

FIG. 2. The variation of the iceline location x_s with solar forcing, where $Q/Q_0 = 1$ is the present climate. The solid line gives the results with the TEC model, the dashed curve shows the results from the standard, transport-limited model. By construction, the two models agree at the present climate where $x_s = 0.95$. But the iceline varies more rapidly with insolation in the TEC model near present conditions. Note also the well-known result that the standard model admits multiple solutions at $Q/Q_0 = 1$.

Figure 3 shows the variation of equatorial, polar and average temperatures versus insolation Q/Q_0. (Equator-to-pole distributions are easily estimated from these.) For clarity, we only show results from the stable branch of the standard model with high-latitude ice caps. The two models agree at Q/Q_0

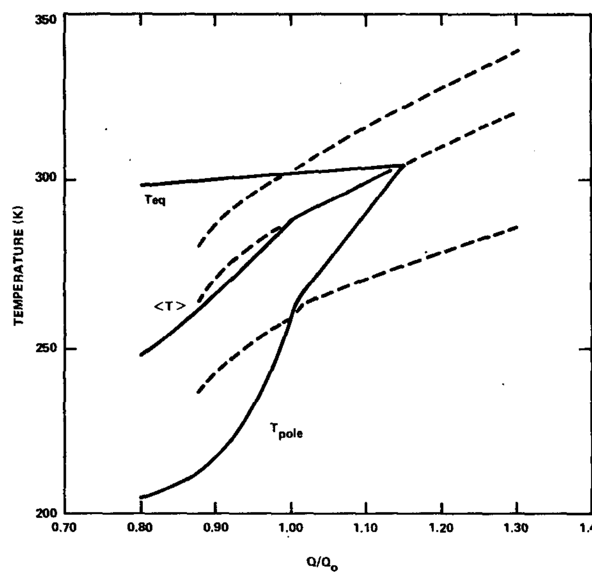

FIG. 3. The variation of polar temperature T_{pole}, equatorial temperature T_{eq} and average temperature $\langle T \rangle$ with insolation in the TEC model (solid curve) and the transport-limited model (dashed curve).

= 1 by construction, and T_{eq} of the TEC model obeys the boundary constraint of Eq. (9). Notice that whenever the icelines are identical, the mean temperature $\langle T \rangle = \int_0^1 T dx$ agrees. This can be understood mathematically from an analysis based on expressing $T(x)$ as an expansion of Legendre Polynomials, as in the discussion of Eq. (11). Here $\langle T \rangle$ depends on the value of $(as)_0$ which depends only on the location of the iceline. Since the only nonlinear process affecting $\langle T \rangle$ is temperature–ice albedo feedback at glacial boundaries, it follows physically that global mean temperatures of the two models are equal when the icelines are at the same latitude. In particular, $\langle T \rangle$ agrees in both models whenever the icecaps disappear, because the temperature–ice albedo feedback is absent altogether. However, the equator-to-pole temperature distribution is markedly different in the two models, even for small perturbations from the present solar constant. From the discussion of TEC model limits this is readily understood. In particular, as Q rises the TEC model becomes isothermal, and latitudinal temperature differences vanish. This behavior contrasts with the Transport-Limited model where diffusive effects become relatively less important with increasing Q, so the temperature gradient grows.

The variation of D with Q shown in Fig. 4 also illustrates the previous discussion on the limits of the TEC model. As Q decreases, to maintain the warm equatorial temperature of the TEC model D must decrease to retain heat in the tropics. At $Q/Q_0 = 0.80$ the model would be in radiative balance with $D = 0$. Any additional reduction in Q would require heat transfer from the pole with $D < 0$. As Q rises, the model becomes isothermal with $D \rightarrow \infty$ at $Q/Q_0 = 1.18$. This is further illustrated in Fig. 5 which shows the peak poleward flux carried by meridional transport normalized to the total solar flux intercepted by the Earth's surface, $\pi R^2 Q_0$. Notice that the flux varies more steeply in the vicinity of the current climate, and monotonically over the 80–118% insolation range considered, in the TEC approximation.

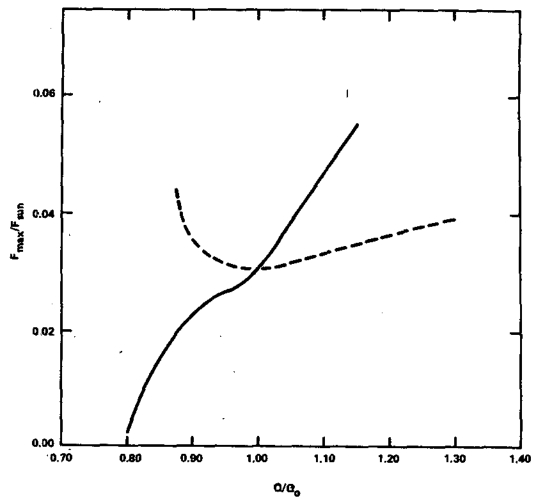

FIG. 5. Peak meridional energy flux as a function of insolation in the TEC (solid curve) and transport-limited (dashed curve) models. The surface flux is normalized to the incident solar flux intercepted by the Earth today.

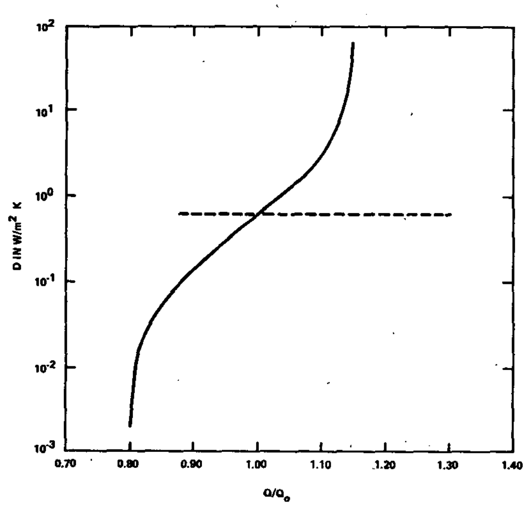

FIG. 4. The variation of thermal diffusion coefficient with insolation in the TEC model (solid curve). The dashed line is the fixed diffusion coefficient in the transport-limited model. At $Q/Q_0 = 0.80$ the TEC model is in local radiative balance; at $Q/Q_0 = 1.18$ the TEC model becomes isothermal.

As a final example of the difference between the Transport-Limited and TEC approximations we show in Fig. 6 the equator-to-pole temperature distributions for $\langle T \rangle = 298$ K corresponding to $Q/Q_0 = 1.09$. The symbols represent estimates of sea surface temperatures during the Cretaceous era which had a similar mean temperature. Global mean temperatures some 10 K above current values have been reconstructed for the Cretaceous (Barron *et al.*, 1981), but a solar output 9% greater 65–140 M year B.P. is not predicted by current models of stellar evolution. It seems more useful in this context to regard the larger solar constant as a proxy for other types of climatic forcing such as higher planetary absorptance or greenhouse heating by atmospheric gases. Neither model can itself be regarded as a description of climate evolution to Cretaceous-like conditions, since neither prescribes the forcing mechanism. However, the TEC model is clearly in much better agreement with reconstructed surface temperature, displaying a diminished equator-to-pole gradient relative to the present. One possible global warming mechanism is continental drift, which left polar zones bereft of continents during this period, thereby preventing highly-reflective continental glaciers from freezing (Donn and Shaw, 1977). But Barron *et al.* (1981) found the redistribution of continents in the Cretaceous was insufficient by itself to get the necessary warming. An interesting possibility is suggested by Budyko's (1982)

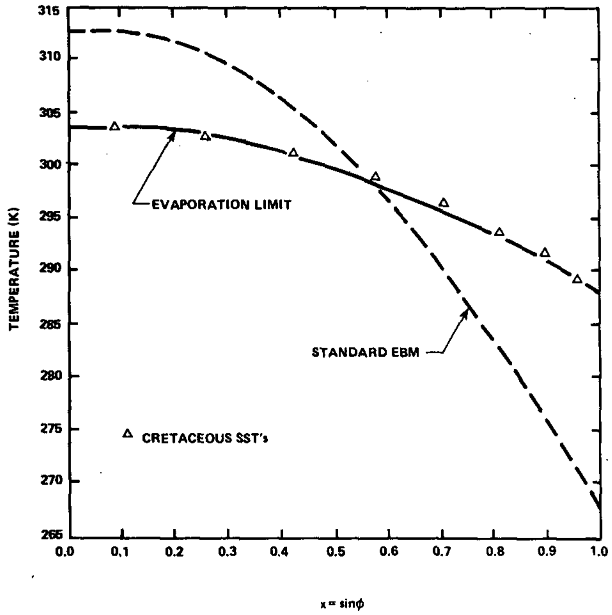

FIG. 6. Comparison of the surface temperature distribution with latitude in the TEC (solid curve) and transport-limited (dashed curve) models when $Q/Q_0 = 1.09$ corresponding to an average surface temperature of 298 K. The symbols show sea surface temperatures during the Cretaceous era (65–140 M year B.P.) from data in Frakes (1979), Schopf (1980) and Barron *et al.* (1981).

recent reconstructions of atmospheric carbon dioxide concentration during the Cretaceous. His Fig. 2.1, p. 43, derived by scaling ambient CO_2 levels to the amount fixed in sedimentary rocks deposited at the time, indicates values of atmospheric carbon dioxide some 8 times current values, corresponding to a greenhouse warming of 7–9 K from logarithmic scaling of carbon dioxide doubling estimates (Hoffert *et al.*, 1980). If these CO_2 levels were present, it is possible that they, in combination with reduced planetary albedo from ice-free poles, could have produced the Cretaceous global warming. In any event, the small equator-to-pole gradient has to be explained; and that feature, in particular, is much better produced by the TEC approximation than the Transport Limit.

4. The "surplus transport" hypothesis

Clearly, an evaporation constraint on equatorial surface temperature, of the type proposed here, is only realistic if the atmosphere/ocean fluid dynamic system has the capacity to carry the implied transport. This implies the surface temperature distribution is not limited by meridional transport, at least to first order, but attains equilibrium by a kind of horizontal "advective adjustment." The plausibility of this hypothesis is discussed below.

Suppose, for simplicity, that the surface temperature distribution is given by a quadratic of the form $T(x) = T_{eq} - T_2x^2$, where T_2 is the equator-to-pole temperature difference. This is a close approximation to our numerical results and is equivalent to the $P_2(x)$–Legendre polynomial expansion approximation of North (1975). Stone (1978) showed for this case that meridional flux across latitude circles is constrained by the insolation and albedo distributions versus latitude, and by the longwave flux versus temperature relation, to the form: $F(x) = (3\sqrt{3}/2)F_m(x - x^3)$, which peaks at $x_m = \sqrt{3}(\phi_m = 35°)$, at the present value $F_m = 5.54 \times 10^{15}$ W from our "current climate" calibrations. In this approximation, the location of the peak meridional flux, which occurs where $a(x)S(x) - I(T) = 0$, is independent of the solar constant, but the peak value is not (cf. Fig. 5). Substituting the quadratic $T(x)$ distribution into the Fourier heat conduction law on a sphere gives

$$F(x) = -2\pi R^2D(1 - x^2)dT/dx = 4\pi R^2DT_2(x - x^3),$$

corresponding to a phenomenological thermal diffusion coefficient $D = 3\sqrt{3}F_m/(8\pi R^2T_2) = 0.64$ W m^{-2} K—reasonably close to the value calibrated for the present climate from numerical solutions of Eq. (1) (cf. Table 1). This "D" depends only on surface temperature and the radiation distributions, but is independent of the specific mechanisms that produce the transport, a point already made by Stone (1978).

The actual mechanisms producing the total meridional flux can be seen in the decomposition of expressions for the air and water components ($F = F_a + F_w$),

$$F_a = 2\pi R(1 - x^2)^{1/2} \int_0^\infty \times (\rho c_p\langle v\rangle\langle T\rangle + \rho c_p\langle v'T'\rangle + \rho L\langle vq\rangle)_a dz,$$

$$F_w = 2\pi Rf_w(1 - x^2)^{1/2} \int_0^{h_w} \times (\rho c_p\langle v\rangle\langle T\rangle + \rho c_p\langle v'T'\rangle)_w dz,$$

where ρ is density, c_p is specific heat at constant pressure, L latent heat of vaporization (water), q specific humidity, v meridional velocity, f_w the fraction of Earth covered by water, h_w is oceanic depth, angular brackets denote a (zonal) average over all longitudes, and the primes denote fluctuations from the zonal mean. The terms in the atmospheric part were derived from atmospheric circulation statistics for both hemispheres by Oort and Vonder Haar (1976) and Trenberth (1979). They include contributions from the mean meridional circulation $\langle v\rangle\langle T\rangle$, baroclinic eddies and forced standing waves $\langle v'T'\rangle$, and poleward humidity (latent heat) flux. In Fig. 1, we plotted total meridional flux $F(x)$ and the atmospheric part obtained by averaging Oort and Vonder Haar's Northern and Trenberth's Southern Hemisphere data, consistent with our neglect of hemispheric asymmetries in the annual mean $T(x)$. The residual $F_w(x) = F(x) - F_a(x)$ is plausibly attributed to the oceans. While direct measurements of the world oceanic meridional flux are difficult and generally unavailable, transports of the right order are computed from ocean general circulation models for a combination of thermohaline overturning $\langle v \rangle \langle T \rangle$, and western boundary currents plus mesoscale eddies $\langle v'T' \rangle$ (Bryan, 1978; Bryan and Lewis, 1979). Some independent estimates of the Atlantic basin contribution to $F_w(x)$ were also made by treating the oceanic poleward flux as a residual in the surface energy balance (Lamb, 1981). Thus the ocean carries roughly half the transport in the real world, possibly more at low latitudes, with the atmospheric part distributed among several mechanisms.

Nevertheless, there has been a tendency in recent years to attribute a governing role in poleward heat transport to atmospheric baroclinically unstable eddies. Lindzen and Farrell (1980), who recognized the somewhat arbitrary nature of the "constant D" version of the Transport Limit, suggest $T(x)$ adjusts to satisfy a criteria for self-neutralization of baroclinic instabilities (along with a tropical Hadley Cell adjustment). This is in fact another kind of Transport Limit. Although we are proposing that $T(x)$ adjusts to a fundamentally different kind of constraint, it is worth reemphasizing the finding of Lindzen and Farrell (1980) and earlier of Stone (1978), that current theory indicates the baroclinic instability mechanism in the atmosphere can itself account for all the needed transport.

Suppose, for example, the only mechanism of significance in poleward transport were the baroclinic instability, generating a mean fluctuation product $\langle v'T' \rangle = -KdT/Rd\phi$ across latitude circles, where K is its effective eddy diffusivity. (As a baroclinic wave grows in the presence of a zonal mean wind, it gives rise to eddy meridional heat and momentum fluxes proportional to the spared amplitude of the wave; these transports feed back on the structure of the mean flow so as ultimately to limit the growth of the waves.) Ignoring the contribution of the mean circulation, forced standing waves, condensational heating by water vapor transported to higher latitudes, and the oceanic flux, the poleward heat flow from zonally averaged baroclinic $v'T'$ eddies only is

$$F(x) = -2\pi(p_0 c_p K/g)(1 - x^2)dT/dx,$$

where p_0 is surface pressure = 1.01×10^5 Pa, c_p is specific heat of air at constant pressure = 1000 J kg^{-1} K^{-1}, and g the gravitational acceleration = 9.81 m s^{-2}. The corresponding thermal diffusion coefficient is $D = p_0 c_p K/(gR^2) = (2.54 \times 10^{-7})K$ (SI units). Models of baroclinic eddy heat flux typically yield horizontal eddy diffusivities of the form (Schneider and Dickinson; 1974) $K = (0.4 \pm 0.2)(R/f)^2 \cdot (\langle T \rangle \langle \gamma \rangle / g)^{1/2}(2T_2/3R)$, where R is the atmospheric gas constant = 287 m^2 s^{-2} K^{-1}, f is the midlatitude Coriolis parameter = 10^{-4} s^{-1}, $\langle T \rangle$ is the global mean surface temperature = 287.3 K, $\langle \gamma \rangle = \langle g/c_p + \partial T/\partial z \rangle$ is the global mean thermal stability = 3.3×10^{-3} K m^{-1} (Stone, 1974). The corresponding thermal diffusivity is $D = 1.2 \pm 0.6$ W m^{-2} K^{-1} for the "present climate" equator-to-pole temperature difference T_2 = 44 K. Remarkably, this mechanism alone seems capable of supplying all the poleward flux needed at present (D = 0.623 W m^{-2} K^{-1} from Table 1), perhaps much more, despite the fact that atmospheric baroclinic eddies in the real world account for only of the order of ¼ or less of the total (Oort and Vonder Haar, 1976). This suggests a "surplus transport" hypothesis in which some factor other than large-scale dynamics of the atmosphere and oceans is the limiting factor in establishing $T(x)$.

As a final consideration in the surplus transport argument, we note EBMs in which D is fixed, or scaled by T_2 as in baroclinic eddy transport models, encounter serious problems when called upon to model $T(x)$ distributions reconstructed for the Cretaceous (Section 3). At that time T_{eq} was at most a few degrees above today's value, while equator-to-pole temperature differences T_2 were some 20–25 K, roughly half of the current value (Frakes, 1979, p. 178; Schopf, 1980, p. 142, Figs. 12–14). Barron *et al.* (1981) tried to recover this $T(x)$ with an EBM, but found ". . . total heat transport must be maintained at close to present-day values despite the fact that meridional temperature gradient was considerably reduced." This they could only accomplish by artificially adding transport to the model-predicted $F(x)$. It is easy to see why. Since transport scales with $D \times T_2$, it is only possible to maintain $F(x)$ at close to present levels when T_2 halves when D doubles. If a constant-D version of the transport limit is used in an EBM, the needed meridional flux simply cannot be maintained without adding some artificially. This is equivalent to saying transport is not limiting. A baroclinic eddy transport model fares even worse. Here D scales with T_2 but F with T_2^2, so only ¼ of the needed transport is predicted for the Cretaceous for D calibrated to the current climate, as opposed to ½ in the constant-D version.

5. Concluding remarks

The considerations and model calculations presented here suggest that the equator-to-pole surface temperature distribution of the Earth is strongly influenced by evaporative cooling in the tropics. The possibility of an evaporative limit on tropical surface temperatures was raised earlier by Priestly (1966) and Newell (1979). What is new in this paper is the incorporation of a Tropical Evaporation Constraint on equatorial temperature in a latitude-resolved Energy Balance Model. This leads to an explicit formulation of how meridional heat transfer changes when global climatic variations are induced by solar forcing.

The crucial result of the TEC model, compared with standard EBMs, is that constraining equatorial temperature actually increases climate sensitivity as measured by movement of the ice line and polar temperature. For nearly identical change in global mean temperature, the standard, transport-limited models produce a quite flat distribution of temperature rise with latitude, compared with the polar amplification and equatorial anchoring of the TEC model (Fig. 3). Polar amplification found in the TEC model agrees better with the general circulation model (GCM) results of Manabe and Wetherald (1980) for solar forcing. However, the GCM study shows that equatorial temperature rises by an amount intermediate between that of the TEC and transport-limited models. Although the GCM calculation explicitly includes those effects of evaporation proposed by Newell and Dopplick (1979) to limit tropical thermal response, the GCM shows a temperature rise much larger than would be expected from the Newell and Dopplick analysis. This has been attributed to non-linear feedback mechanisms unaccounted for by Newell and Dopplick, especially those associated with warming from enhanced atmospheric water vapor emission (Kandell 1981, Ramanathan 1981). In Section 2b, we noted that line-by-line models of backwarming in tropical regions show that water vapor feedback is less effective than found in other studies relevant to mid-latitudes. The effect is smaller because H_2O absorption bands are more nearly saturated than in mid-latitudes, see also Kiehl and Ramanathan (1982). Thus, the true situation in tropical latitudes might be closer to the TEC model than indicated by GCM results, while the TEC model definitely produces better accord than the standard, transport limited EBMs for polar warming.

In the TEC model ice line boundaries move with considerably less insolation change during periods when high latitudes are covered by continental glaciers (as presently). The inability of standard EBMs to respond strongly enough to astronomical solar radiation oscillations to recover historical Ice Ages (Hays *et al.*, 1976) is well known, and may be related to the Transport Limit incorporated in even the most sophisticated models of this type (e.g., Suarez and Held, 1979). We have also discussed the better performance of the TEC model under conditions like the warm Cretaceous when the poles may have been largely ice-free. Moreover, the weak variability of surface temperature in the tropics over seasonal cycles, as well as paleoclimatically, is consistent with an evaporative damping model.

Nevertheless, these considerations are only a first step toward understanding the relationship between surface and global energy balances, and the specific ways the various poleward transport processes adjust. However, if the findings of this study are borne out in more stringent tests, such as seasonal climate simulations, the TEC concept could lead to a fundamentally new picture of how global climate responds to external forcing. It would be one in which horizontal dynamical processes in the Earth's fluid envelopes play the secondary role of adjustment to another constraint, rather than limiting the $T(x)$ distribution themselves. This is already the situation in the "convective adjustments" which determine $T(z)$ in the troposphere, and in certain zones of stellar atmospheres. A more detailed model will necessarily involve term-by-term analysis of the surface energy balance, and distinct thermal reservoirs for air, sea, land, and possibly ice.

More work is also needed to understand tropical evaporation buffering in the context of atmospheric and oceanic general circulation models (GCMs) which compute transports explicitly from dynamics. The present state-of-the-art is that equilibrium states computed with coupled atmosphere-ocean circulation models are only beginning to emerge, and most atmospheric GCMs constrain sea surface temperature fields either by specifying them or by computation from surface energy balances without oceanic transport. Since roughly half the poleward heat flow is carried by oceans, the satisfaction of realistic surface boundary conditions means the modeled atmosphere must carry roughly twice its actual poleward heat flux. The GCM-modeled atmosphere therefore has the property of adjusting its dynamics to supply the total transport requirement, in accord with our Surplus Transport hypothesis.

Finally, we observe that the tropical oceans of the Earth are not the only place in the solar system where surface temperatures are buffered by latent heat absorbed during phase changes. The CO_2 polar icecaps on Mars apparently perform the analogous function of stabilizing polar temperatures to the sublimation frost point of the carbon dioxide atmosphere (see, e.g., Hoffert *et al.*, 1981). Again, one sees the localized effect of a planetary surface in near vapor-pressure equilibrium with an atmospheric gas strongly constraining the local surface temperature (to about 150 K at the Martian caps). On both planets, surface radiant energy levels, controlled by orbital distance and gas composition, in combination with thermodynamics of phase transition from the liquid or solid to the gaseous state, may dominate meridional transport effects in establishing the equator-to-pole distribution of surface temperature.

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