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The Effect on Earth's Surface Temperature From Variations in Rotation Rate, Continent Formation, Solar Luminosity, and Carbon Dioxide

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Proposed evolutionary histories of solar luminosity, atmospheric carbon dioxide amounts, Earth rotation rate, and continent formation have been used to generate a time evolution of Earth's surface temperature. While speculative because of uncertainties in the input parameters, such a study does help to prioritize the areas of most concern to paleoclimatic research while illustrating the relationships and mutual dependencies. The mean temperature averages about 5 K higher than today over most of geologic time; the overall variation is less than 15 K. The evolution of Earth's rotation rate makes a significant contribution to the surface temperature distribution as late as 0.5 b.y. ago. While there is little change in equatorial temperatures, polar temperatures decrease, being some 15 K lower 3.5 b.y. ago than with present day rotation. The effect of continent growth on albedo is of secondary importance.

INTRODUCTION

The surface temperature of the early earth has evoked much interest recently. The geologic record tells us that at least to 3.8 b.y. ago liquid water was present, and there is no indication that the temperature has varied on a long time scale by more than about 10°C from the present temperature. Yet it is generally believed that solar luminosity has increased over geologic time and that 4.5 b.y. ago the solar radiation reaching Earth was only about 70 to 80% of that of today. Under present conditions, this would certainly have caused a "frozen planet." And while the amount of carbon dioxide in the atmosphere is only 0.035% today, it may have been several bars a few billions of years ago before silicate weathering converted much of the carbon dioxide to carbonate rocks. It has been suggested that the larger greenhouse heating which would have occurred with more carbon dioxide may have countered the cooling from lesser amounts of radiation from the sun and so caused the temperature to be much as it is today [*Owen et al.*, 1979; *Walker et al.*, 1981; *Kuhn and Kasting*, 1983; *Marshall et al.*, 1988]. But we also know that the surface was covered mostly by a global ocean with the only land being small and possibly isolated embryonic continents [*Taylor and McLennan*, 1985; *Nisbet*, 1987]. Certainly a water-covered planet has much different radiative characteristics than Earth today; for example, the ocean albedo is different from land albedo, and polar ice caps are more difficult to form if there is no high-latitude land mass. And finally, the rotation of the Earth has slowed; early in the planet's history, this more rapid rotation rate would have lowered the atmospheric transport of heat poleward and would thus have changed the latitudinal temperature gradient.

Most studies of the temperature of the early Earth have considered these processes separately, although there have been a few that have included the coupling between carbon dioxide amounts and reduced solar luminosity. There have been no studies of the latitudinal distribution of temperature that have coupled these various processes. This work does so.

DESCRIPTION OF THE MODEL AND NUMERICAL PROCEDURES

The latitudinal distribution of surface temperature is calculated from an energy balance model similar to that of *North* [1975], as revised by *Lin* [1978],

$$-\frac{d}{dx}\left[D(1-x^2)^{\frac{1}{2}}\left|\frac{dT}{dx}\right|\left(\frac{dT}{dx}\right)\right]+I(p_{co_2},\ell,T)$$

$$=QS(x)\left[1-\mathcal{A}(T,\ell)\right]/4 \qquad (1)$$

where x is the sine of latitude, $I(p_{co_2},\ell,T)$ is the outward infrared radiation at the top of the atmosphere which depends on surface temperature T, and the atmospheric partial pressure of carbon dioxide p_{co_2}, D is the nonlinear heat transport coefficient, Q is the solar constant (1380 W m^{-2}), S is the annual meridional solar radiation distribution, and $\mathcal{A}$ is the albedo for solar radiation; ℓ is the fraction of land area in a latitude zone. The second term is a loss of energy (cooling) because of longwave radiation escaping to space; the last term represents a heating due to absorption of solar radiation, and the first term is the gradient of the meridional transport of heat from low to high latitudes which can either cool or heat locally.

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Heat Transport

We have used a nonlinear version of the model in which the heat transport is proportional to the square of the temperature gradient. This allows for the inclusion of the rotation rate and provides for better agreement with atmospheric data than does the linear version of the model [*Stone and Miller*, 1980]. Also, the derivation of the nonlinear form is based on baroclinic instability by methods of some rigor [*Stone*, 1974], whereas the linear transport law relies on a mixing length hypothesis.

The dependence of heat transport on the rotation rate is not surprising because of the strong dependence of the mid-latitude eddies (which accomplish most of the transport) on rotation rate. From baroclinic instability theory, it is well known that the amount of energy received by an unstable eddy from the zonal flow is determined by the meridional and vertical components of the flow and the angle subtended with the isentropes. Since the rotation rate partially dictates the critical angle and hence the degree of instability, the growth rate and size and spectrum of the baroclinically unstable waves depend upon the rotation rate.

Stone [1972] has derived the dependence of the heat transport on rotation rate via baroclinic instability theory,

$$\text{heat transport} \propto \left(\partial\theta/\partial y\right)^2/f^2 \tag{2}$$

where $\partial\theta/\partial y$ is the meridional gradient of the potential temperature θ, and f is the Coriolis parameter. The theory is a quasi-geostrophic one and thus is not correct for small rotation rates, but throughout geologic time the Earth's rotation has always been fast enough for geostrophy to apply. This form of the transport also assumes eddies that are of depths comparable to the troposphere but these are the ones of primary importance to the large-scale transport we are considering. There is no provision for transport by latent heat.

The parameterization of heat transport in simple energy balance climate models usually includes latent heat transport in a crude implicit way by tuning to the present climate [*e.g., North*, 1975]. We assume the transport of latent heat has the same functional relation to the rotation rate as does the sensible heat transport. *Mullen* [1978] has suggested a relation that directly relates the meridional flux of sensible heat and the flux of latent heat. The implication from that study is that both forms of meridional heat flux have the same dependence on rotation rate, which is what we have assumed here. The effect of ocean heat transport is too unknown to include; however, some of the oceanic transport is due to baroclinic eddies and would have a rotational dependence similar to that used here.

We have chosen the nonlinear diffusive heat transport $D = 1.2$ W m^{-2} K^{-2} to give a latitudinal temperature gradient in agreement with that of the present day (see Figure 8).

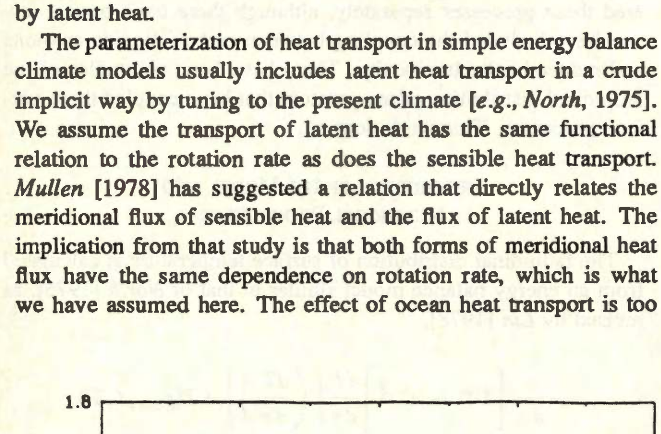

Fig. 1. Variation of Earth's rotation rate over geologic time relative to present day.

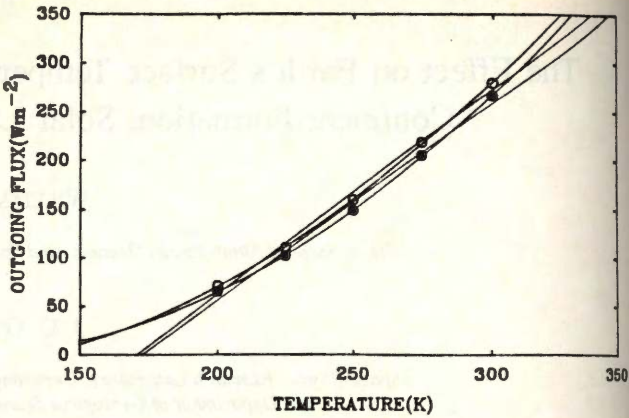

Fig. 2. Calculated outgoing infrared flux for a cloud cover of 36% (open circles) and 56% (solid circles). These cloud covers correspond to those at 25° latitude for land (open circles) and ocean (solid circles). The straight line fit is a linear approximation to the data, while the curved line represents a Laguerre polynomial fit.

Rotation Rate

The assumed variation of rotation rate over Earth's history is shown in Figure 1 [*Zahnle and Walker*, 1987]. The rotation has slowed because of tidal friction with the moon and may have been 70% faster (day length of 14 hours) 4 b.y. ago.

Outgoing Planetary Radiation

In most energy balance studies of this type, the escaping planetary radiation is expressed as a linear function of surface temperature, i.e., $I = A + BT$, where A and B are constants and are approximately -350 and 2, respectively (with temperature in degrees Kelvin and the outgoing radiation in W m^{-2}). For our purposes, however, a linear approximation is not acceptable because we distinguish between the radiation escaping over land and that over the oceans; the differences in these fluxes are comparable to the errors in the linear approximation. A typical example is shown in Figure 2. We have specified land and ocean separately because we allow the land and ocean fractions to vary over geologic time and use the present-day cloud cover appropriate for land and ocean for that latitude to determine the escaping planetary radiation. Total cloud covers are given in Table 1. Since we are assuming the hemispheres are symmetrical, the cloud covers in Table 1 represent averages of the northern and southern hemispheres for the specific latitudes.

The radiative model used to calculate the outgoing planetary radiation is one that has been used by Kuhn in the comparison of radiation codes for use in climate models [*Luther*, 1984; *Hummel and Kuhn*, 1981]. The atmosphere was divided into 1-km layers extending from the ground to 70 km. For the present study the only radiative gases considered were carbon dioxide and water vapor. The spectral data used to calculate the transmission functions were from the 1983 Air Force Geophysical Laboratory data tape [*McClatchey et al.*, 1973]; there are 13 infrared spectral intervals

TABLE 1. Zonally Averaged Annual Cloud Cover Over Land and Ocean

Latitude	Land	Ocean
85	0.43	0.66
75	0.54	0.76
65	0.64	0.82
55	0.64	0.82
45	0.53	0.74
35	0.46	0.64
25	0.36	0.56
15	0.49	0.54
5	0.65	0.58

(extending from 500 to 1420 cm^{-1}) for carbon dioxide and 11 intervals (0 to 2450 cm^{-1}) for water vapor.

The outgoing radiation was determined for specific surface temperatures varying from 200 K to 300 K at 25 K intervals; carbon dioxide partial pressures were varied from the present amount, designated 1 PAL (present atmospheric level, 345 ppm), to 2500 PAL.

The lapse rates, relative humidities, and cloud cover used in the calculation of the outgoing radiation are the same as the present-day atmosphere. Surface relative humidity is 77%, and the relative humidity varies with height as given by *Manabe and Wetherald* [1967] to 150 mbar, above which the mixing ratio is fixed at 10^{-6}. The lapse rate is from *Kuhn and Kasting* [1983] and is 6.7 K km^{-1} from the ground to 10 km, and 0.9 K km^{-1} from 10 to 20 km; the atmosphere is isothermal above 20 km. The clouds are located at 500 mbar.

The calculated fluxes for clear and overcast skies for each of five surface temperatures and eight different amounts of carbon dioxide are given in Table 2. The outgoing flux for any specified cloud cover and surface temperature is found by weighting the clear and overcast fluxes by the fractional cloud amount. The fluxes for arbitrary temperature are found from a five-term Laguerre interpolation.

A comparison of the calculated outgoing planetary radiation with observations by *Stephens et al.* [1981] is given in Figure 3. The surface temperatures were from *Crutcher and Meserve* [1970] and *Jenne et al.* [1969] and the cloud cover for land and ocean from *Warren et al.* [1986] and *London* [1988], respectively.

Carbon Dioxide

Two scenarios (Figure 4) have previously been developed for the time history of carbon dioxide. *Hart* [1978] related an exponentially decreasing degassing rate and concomitant release of carbon dioxide to the weathering rate and found that carbon dioxide would decrease approximately exponentially with time from an initial amount of 1500 PAL 4.5 b.y. ago. *Kasting* [1987], on the other hand, determined the amount of carbon dioxide that would be necessary to maintain global mean temperatures above freezing; very early in Earth's history, he requires 2500 PAL of carbon dioxide. The variations in Kasting's carbon dioxide amounts reflect glacial periods when presumably there would be less carbon dioxide. We have used the more recent carbon dioxide scenario from Kasting, although we discuss how surface temperature differs for Hart's carbon dioxide amounts.

Albedo

In many energy balance climate models, the latitudinal extent of an annual averaged polar ice cap is an important result. The "climate" will locally be different over a frozen ice-covered surface, but ice will also affect the overall climate because of its higher albedo, with concomitant reduction in absorbed solar radiation which alters the transport of heat.

Models which attempt to predict perturbations from the present climate generally assume that the ice line occurs when surface temperature reaches -10°C, based on climatological studies of *Budyko* [1969]. For models with present day polar geography this is certainly reasonable. For our study, however, which extends several b.y. back in time, the polar oceans, indeed the entire planet, have little or no land masses and it is not clear that the ice line would be the same as that of today. Rather than arbitrarily defining an ice line, we have related the albedo to temperature based on observational data from the present day climate for land and ocean separately.

The albedos, taken from the measurements of *Stephens et al.* [1981] are given in Figures 5*a* and 5*b*. We determined from their annual albedo map average albedos for each 10° latitude belt for land and ocean separately. The latitudinal variation is due to many factors, primarily solar zenith angle, cloud cover, type of surface,

TABLE 2. Clear and Overcast (in Parentheses) Outgoing Fluxes (W/m^2) for Selected Total CO_2 Amounts (Units of PAL)

	Temperature, deg				
	200	225	250	275	300
1 PAL	81.8 (52.4)	125.8 (83.2)	179.8 (124.1)	242.3 (175.8)	303.3 (237.4)
6.67	79.9 (51.5)	121.4 (81.6)	172.4 (121.1)	232.2 (171.1)	292.0 (230.7)
35.1	77.6 (50.7)	116.8 (80.0)	164.7 (117.8)	221.2 (166.0)	278.5 (223.5)
100.	75.8 (50.0)	113.8 (78.8)	159.4 (115.5)	213.0 (162.5)	267.4 (218.6)
333.3	73.7 (49.1)	110.1 (77.3)	152.5 (113.0)	201.0 (158.4)	250.7 (211.9)
666.7	72.7 (48.6)	107.7 (76.4)	147.5 (111.5)	192.4 (155.6)	239.0 (207.0)
1000.	72.2 (48.3)	106.1 (75.9)	144.3 (110.6)	186.8 (153.7)	231.6 (203.7)
1338.	71.8 (48.2)	104.9 (75.6)	141.8 (109.8)	182.6 (152.3)	226.2 (201.1)

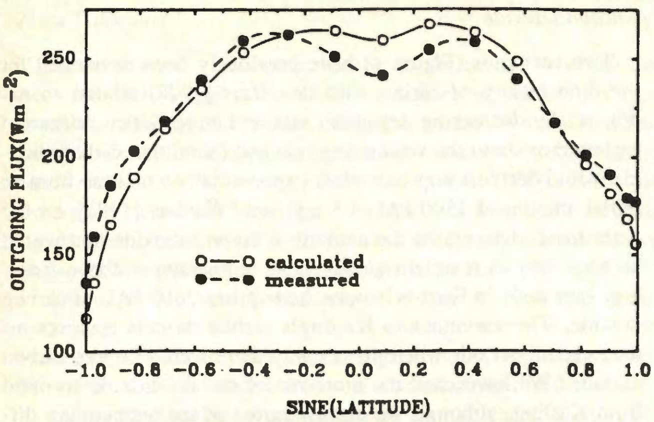

Fig. 3. A comparison of calculated outgoing infrared radiation with observations by *Stephens et al.* [1981].

atmospheric water vapor, and presence of snow or ice. The effect of temperature on albedo manifests itself primarily through the formation of snow and ice and to some extent by water vapor's absorption of solar radiation.

Note that the trends for the northern and southern hemispheres are very similar, so that a single function can be used to represent both hemispheres. However, the variation over land is not the same as the variation over water, so use of a single albedo averaged over land and ocean would cause error. In equatorial and subtropical latitudes, the land albedo is larger than the ocean albedo even though the cloud amount is larger over the oceans; this occurs because of the low ocean albedo when the sun is "high." In the mid-latitudes, the land albedo increases more slowly with latitude. Much of the effect is due to the larger ocean cloud cover.

To ascertain the effect of temperature on albedo, we calculated the albedo using the formulation of *Thompson and Barron* [1981], which includes solar zenith angle, ozone and water vapor absorption, scattering, ocean albedo including roughness, cloud cover, and land albedo. For each latitude, the surface temperatures were varied over $\pm 30°$ at $10°$ intervals about the mean latitude temperature [*Crutcher and Meserve*, 1970; *Jenne et al.*, 1969]. As long as snow or ice did not form, the effect of temperature on albedo over the range considered was very small, being 0.02 or less. The calculated albedos are also plotted in Figures 5*a* and 5*b*. Note that they agree well with the observed data until a latitude

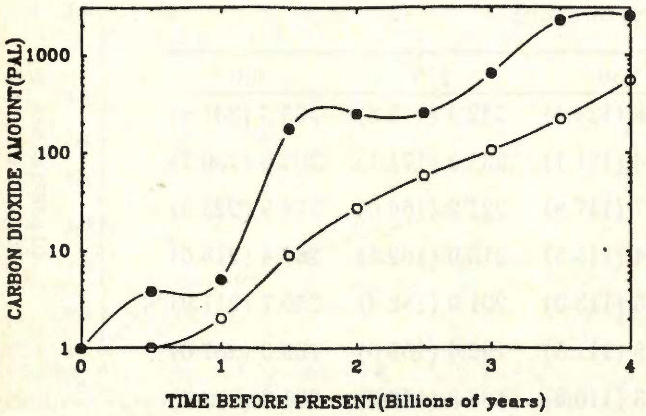

Fig. 4. Variation of atmospheric carbon dioxide relative to present day over geologic time. The topmost curve is modified from *Kasting* [1987], while the lower curve is from *Hart* [1978].

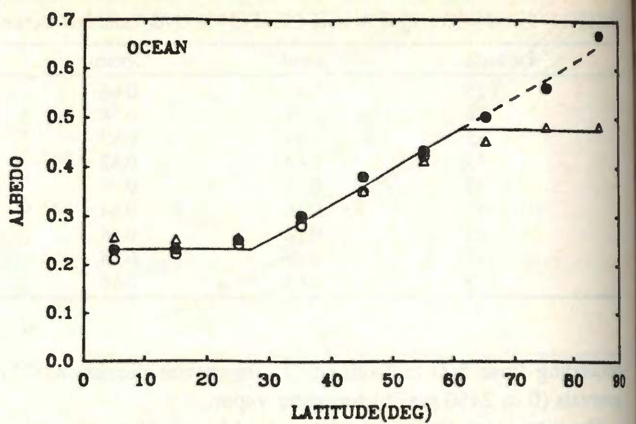

Fig. 5*a*. The albedo of the ocean (including clouds) for the northern (solid circles) and southern (open circles) hemispheres as a function of latitude [from *Stephens et al.*, 1981]. The albedos calculated from *Thompson and Barron* [1981] are shown as triangles. The solid line refers to the assumed variation of albedo with latitude when snow and ice are not present. The difference between the solid and dashed curves is the additional contribution to the albedo when temperature is less than 268 K, i.e., when snow and ice are present.

of about 65° is reached when snow (land) or ice (ocean) begin to form.

The "ice line" for the ocean appears a few degrees lower than that for land, but certainly our analysis cannot justify this small difference. It is not clear how the ice line should differ over land and water. Wave motion would tend to inhibit ice formation, but on the other hand, to have ice and snow over land there must be adequate precipitation. We take the ice line as 268 K corresponding to the annual average temperature at 65° latitude.

For latitudes θ in our model where the surface temperature is higher than 268°K, the albedo $\mathcal{A}$ is independent of temperature and varies with latitude according to

Ocean

$$\theta \leq 27° \qquad \mathcal{A} = 0.225$$

$$27° < \theta \leq 64° \qquad \mathcal{A} = 0.00730\theta + 0.0312$$

$$\theta > 65° \qquad \mathcal{A} = 0.475$$

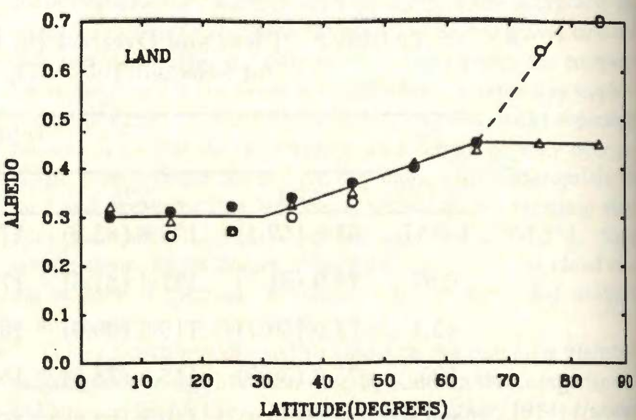

Fig. 5*b*. The albedo of the land (including clouds) for the northern (solid circles) and southern (open circles) hemispheres as a function of latitude [from *Stephens et al.*, 1981]. The albedos calculated from *Thompson and Barron* [1981] are shown as triangles. The solid line refers to the assumed variation of albedo with latitude when snow and ice are not present. The difference between the solid and dashed curves is the additional contribution to the albedo when temperature is less than 268 K, i.e., when snow and ice are present.

Land

$\theta \leq 30°$ $\quad \mathcal{A} = 0.300$

$30° < \theta \leq 65°$ $\quad \mathcal{A} = 0.00420\theta + 0.1725$

$\theta > 65°$ $\quad \mathcal{A} = 0.45$

The difference in the two curves in each of the Figures 5*a* and 5*b* for latitudes greater than 65° is a measure of the effect of temperature on albedo, i.e., when the change in albedo is plotted against surface temperature, there results for temperature $T <$ 268°K,

Ocean

$$\mathcal{A} = 3.497 - 0.013047T$$

Land

$$\mathcal{A} = 1.9069 - 0.0071154T$$

subject to the condition that the albedo cannot exceed 0.7, which is the albedo over the Antarctic continent and should represent a reasonable upper limit. Pack ice, for example, has an albedo in the range of 60 to 70% [*Kondrat'ev*, 1973].

A deficiency in our parameterization is of course that we have not separated the albedo due to clouds from that due to the surface. Implicit in this is that as the land and ocean fraction change, the cloud contribution to albedo at that latitude changes only in proportion to the change in land or ocean. This problem has not been solved even for perturbations to the present-day system and thus we can only speculate. The rotation rate has not changed so much that the general characteristics of the tropospheric circulation would be grossly different. On the other hand, we do find latitudinal temperature gradients that may have been substantially larger in Precambrian times, so precipitation patterns could have been different.

Continentality

The land area of the Earth's surface has increased over geologic time. Prior to 3 b.y. ago most of the Earth was covered with water. As crustal fractionation and cratonization proceeded, the continents formed as small, mobile plates transformed into the rigid lithospheric plates of today. The evolutionary history of dry land that we use is shown in Figure 6 and is modified from *Goodwin* [1981].

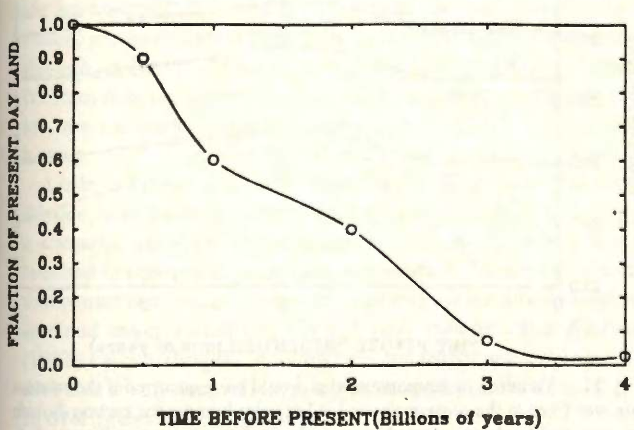

Fig. 6. Variation in growth of continental mass relative to present day over geologic time. Modified from *Goodwin* [1981].

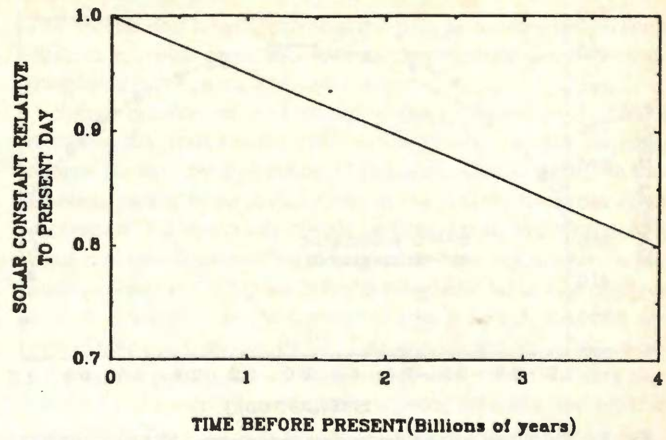

Fig. 7. Variation in the solar constant over geologic time.

Solar Luminosity

Models of stellar evolution indicate that the solar luminosity has increased [*Newman and Rood*, 1977] uniformly over geologic time. As hydrogen is converted to helium in the solar core, the rate of nuclear burning increases, causing the "solar constant" to be some 20% to 30% higher today than it was at the time of Earth's formation, some 4.5 b.y. ago. The time evolution of solar luminosity is shown in Figure 7; the present-day value we take as 1374 W m^{-2}.

Numerical Procedure

A modification of the control volume method was used to solve (1) [*Patanker*, 1975]. Gauss's theorem was applied to (1) which was written in integral form. Boundary conditions are $(1 - x^2)^{1/2} dT/dx = 0$ for $x = 0, 1$.

The latitudinal variation of temperature and the derivatives were originally expanded in Legendre functions. However, there was poor spectral convergence beyond the first three terms in the expansion when the ice line moved to lower latitudes. This same problem occurs in the original *North* [1975] model, although it is not generally realized because of the low-order truncation used in both the temperature and albedo expansion. It should be noted that simply adding more terms in the expansion of the temperature will not resolve this problem; a further expansion of the albedo must also be included. Thus we have expressed the one derivative in finite difference form (Gauss's theorem eliminates the other). We used 120 "volumes" equally spaced in $x(= \sin\theta)$.

RESULTS AND DISCUSSION

A comparison of our computed present day surface temperatures with the actual annual temperature is given in Figure 8. The model was tuned by adjusting the solar constant and the heat transport coefficient to reproduce observed global average temperature and the first moment of temperature against sine(latitude). The agreement is good with the exception of the Antarctic region, where the mean temperature profile we used to generate the outgoing radiation does not adequately represent such a high-altitude region. Thus we are confident our radiative transfer scheme adequately determines the outgoing radiation.

A convenient parameter used to estimate the sensitivity of "the climate" to various feedback mechanisms is the sensitivity parameter $\beta = QdT/dQ$. *Cess* [1976] has employed annual and seasonal climatology and finds a value of $\beta = 145°$C. *North et al.* [1981] has estimated that for the actual climate $80° < \beta < 340°$.

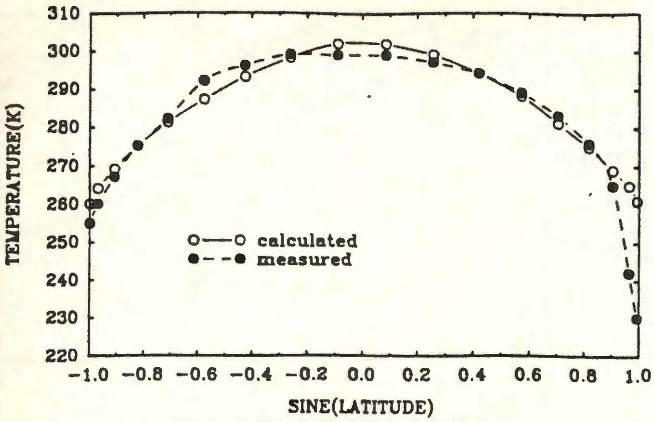

Fig. 8. Calculated present day surface temperature. Measured temperatures are from *Crutcher and Meserve* [1970] and *Jenne et al.* [1969].

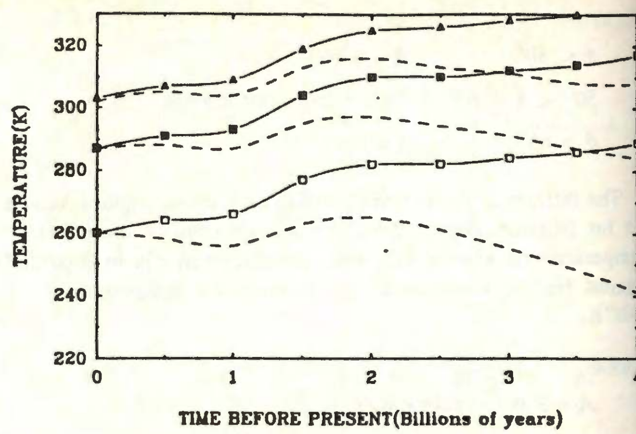

Fig. 10. Variation in temperature over geologic time that would have occurred if atmospheric carbon dioxide varied according to *Kasting* [1987] (see Figure 4), but the solar constant, rotation rate and land fraction are held constant at today's values. The dashed curves are those from the standard model, i.e., Figure 9.

Most climate models, be they simple one-dimensional (1D) radiative convective models, or general circulation climate models, fall in this range. We find for our model $\beta = 133°C$.

The surface temperature evolution which we shall use to ascertain the significance of variations in rotation, carbon dioxide concentrations, solar luminosity, and land coverage is shown in Figure 9. This we determined from the evolutionary sequences given in Figures 1, 4, 6, and 7. We obviously cannot say with certainty that this represents the actual evolution of Earth's surface temperature over geologic time because of uncertainties in the aforementioned variables; we would expect, however, that the trends are correct so that as the uncertainties in these parameters are reduced, this study may help modelers prioritize the need for complexity in their representations of the variables.

In Figure 9 and the following figures, the top and bottom curves represent the time sequence of equatorial and polar temperatures, while the mean global temperature [sine(latitude) weighted temperature] is the middle curve. The temporal variation is less than about 15 K and the mean temperature averages about 5 K higher than today over most of geologic time. Our results are consistent with present-day thinking, but this is to be expected since carbon dioxide strongly affects the temperature and we have used *Kasting*'s [1987] carbon dioxide results which were tuned to be consistent with paleoclimate data.

We have also included in Figure 9 the mean temperature over geologic time from *Hart*'s [1978] data. The temperatures average some 3 to 20 K lower than with Kasting's carbon dioxide amounts, which are much larger than those of Hart (see Figure 4). One might expect that the temperature differences would be even greater given the differences in carbon dioxide in the two studies, but the carbon dioxide bands are saturating with such large amounts as occur in both models early in Earth's history so that the difference in the greenhouse effect is small.

In a previous study, *Kuhn and Kasting* [1983] used a 1D radiative convective model and Hart's carbon dioxide amounts to compute a temperature history over geologic time; temperatures were about 290 K, some 3 to 13 K higher than the mean temperature from Hart's carbon dioxide as shown in the present model (Figure 9). The 1D results did not allow for a cloud amount that depends on the land fraction. Since the cloud amount is higher over ocean than land (see Table 1), and the ocean fraction increases as one goes back in geologic time, the albedo will increase, causing a lower temperature in the present study. Also, there is no ice albedo feedback in the 1D model. In the present model, as the ice line moves to lower latitudes the albedo increases, causing a lower temperature.

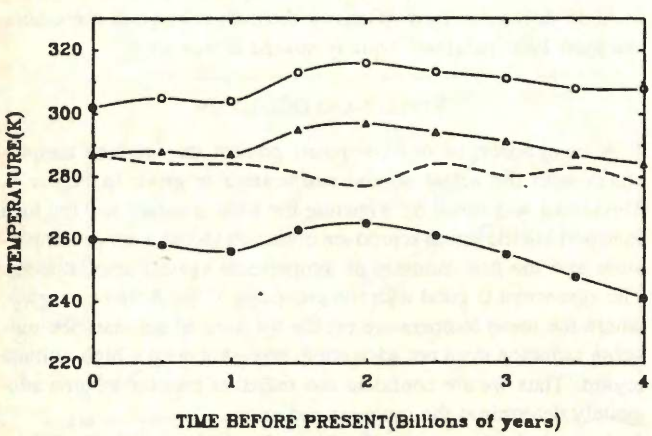

Fig. 9. Variation of surface temperature over geologic time. The uppermost and lower curves refer to the equator and pole, respectively; the middle curve is the average surface temperature and corresponds to 35° latitude. The dashed line refers to the mean temperature if *Hart*'s [1978] carbon dioxide scenario is used instead of *Kasting*'s [1987].

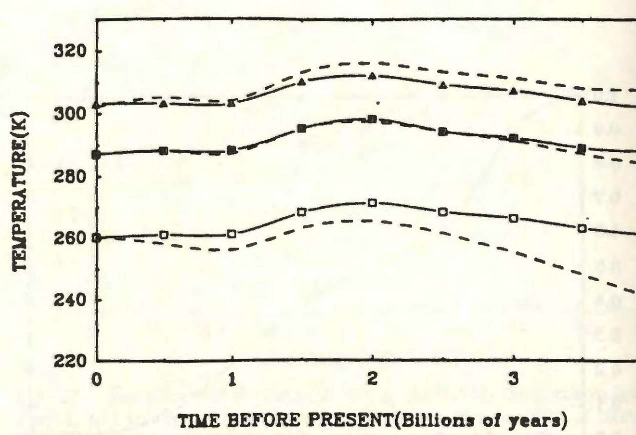

Fig. 11. Variation in temperature that would have occurred if the rotation rate was fixed at the present-day value but solar luminosity, carbon dioxide amount, and land mass varied over geologic time as given in Figures 4, 6, and 7. The dashed curves are those from the standard model, i.e., Figure 9.

It is well known that as the carbon dioxide amount increases, all else remaining the same, the temperature increases because of the greenhouse effect. This is shown in Figure 10 where the mean temperature is 30 K higher 4 b.y. ago when the assumed carbon dioxide amounts were 2500 PAL (see Figure 4). There is no change in the pole to equator temperature difference, since there is no ice albedo feedback because polar temperatures are high enough to prevent formation of ice.

The change in rotation rate affects the pole to equator temperature difference as early as 0.5 b.y. ago (Figure 11); this 13% increase in rotation rate has decreased the polar temperature about 5 K. The gradient continues to increase as we go back in time, and 3.5 b.y. ago the polar temperature is 17 K less than it would be if the rotation rate were fixed at the present value. Note that the change in temperature at the equator is less than 5 K. The small change in equatorial temperature relative to the large decrease in polar temperature occurs because a small change in poleward heat transport has a greater effect on polar temperatures. Energy is distributed over a much smaller area in high relative to low latitudes. Changing rotation causes little change in mean global temperature.

Hunt [1979] has also examined with a general circulation model the effect of Earth's rotation on the latitudinal temperature gradient. Present atmospheric conditions were used so that the cloud cover, surface albedo, and carbon dioxide amount were not allowed to vary. He finds, e.g., with a 14-hour day (corresponding to our rotation rate of 4 b.y. ago) that the pole to equator temperature difference is about 50 K. We have run our model with present-day conditions and find a difference of 67 K, i.e., our polar temperature is lower. This is to be expected since we have an ice-albedo feedback. At higher rotation rates, less energy is transported poleward, causing a lowering of polar temperatures and a larger albedo, which further reduces the temperature.

If the present day land fraction is used throughout Earth history, then the computed temperatures are lower by 2 K early in Earth's history. The lower temperatures for the larger land amount occur because the land has a higher albedo than does the ocean. If at present there were no land, then our calculations indicate that the mean temperature would increase some 2 K. The change in land mass is likely of secondary importance insofar as its effect on albedo. However, as we have shown previously [*Marshall et al.*, 1988], its effect on carbon dioxide weathering is significant, and the latitudinal distribution of land is important. The carbon dioxide-greenhouse weathering feedback stabilizes the climate system. More study of this effect is presently under way.

Of the features considered, i.e., solar luminosity, carbon dioxide amount, rotation, and land fraction, the first two, in combination, are most important to maintaining the surface temperature throughout Earth's history. And while the variation of the solar constant is better understood than that of changes in carbon dioxide, the uncertainty in solar constant may be 20% in early Earth history.

Little is known about the time history of atmospheric carbon dioxide over geologic time; while *Kasting* [1987] has developed a scenario, his study is based on the amount of carbon dioxide required to produce the expected temperature. He uses a 1D radiative convective model which does not include ice albedo feedback or cloud cover variations. We can only conclude that *Kasting*'s [1987] carbon dioxide amounts are consistent with current wisdom that mean temperature has not varied by more than 10 K or so over much of geologic time.

Perhaps the most interesting conclusion of this work is that variation in rotation rate should be included for Precambrian studies. The shorter day length increases the pole to equator temperature difference, which could influence the tropospheric circulation, atmospheric lapse rates, and cloud cover.

Simple one-dimensional latitudinal energy balance models likely represent the most complex model that can be justified for early climate studies; the application of global circulation models to this problem awaits better data. Even in our model, there are other parameters that eventually should be considered. We have used a lapse rate that does vary with height but have not allowed a latitude dependence. At present the tropospheric lapse rate changes from 6 K km^{-1} near the equator to less than 3 K km^{-1} at the pole [*Stone and Carlson*, 1979]. This would have the effect of decreasing the pole to equator temperature difference. We have also used a fixed relative humidity assumption; typically, the relative humidity in the polar atmosphere is lower than at the tropics [e.g., *Manabe and Wetherald*, 1967]. If this applied also to the early Earth, and there is no a priori reason to expect that it would, then the pole to equator temperature difference would be larger than we have shown. And finally, there is the ubiquitous cloud cover, which even today is not well parameterized. Because of the large albedo associated with clouds, they have a major effect on surface temperature. And while we have included the latitudinal distribution of clouds and their variation over land and ocean, the infrared formulation uses a fixed cloud height and present-day amount. A decrease in cloud amount and/or increase in cloud height causes an increase in surface temperature and vice versa.

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