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The effect of solar output, infrared cooling and latitudinal heat transport on the evolution of the earth's climate

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ABSTRACT

An attempt is made to reconstruct some features of the past history of the earth's climate using a one-dimensional planetary model. Starting at 250 million years ago to the beginning of the quaternary glaciation and taking into consideration the evolution of the sun's luminosity and paleoclimatic as well as geophysical data, possible evolutionary pathways are drawn. The main point of the results is that regardless of the details of the pathway followed, the infrared cooling coefficients and the heat transfer coefficient necessarily must have been evolving. Some arguments are advanced, determining the direction of this evolution.

1. Introduction

The influence of the solar output on surface temperature of the earth has been analyzed by Budyko (1969) and Sellers (1969) on the basis of the *ice-albedo feedback*. It was found that a slight variation of the solar constant in the interval 1.5 to 2% can induce climatic catastrophes associated with the emergence of an ice-covered or an ice-free earth. Recently it has become increasingly clear that the sun's energy output has systematically increased over the past 10^9 years (Newmann and Rood, 1977). The apparent discrepancy between this fact and the absence of glaciation during the mesozoic and early cenozoic era has been pointed out. In an attempt to resolve the difficulty, Sagan and Mullen (1972) invoke the possibility of an enhanced greenhouse effect in the framework of a global energy balance model at the planetary scale.

In this paper an attempt is made to analyze the global trends of climatic evolution in the past 250×10^6 years up to the beginning of quaternary glaciations, using a model involving latitudinal energy transfer. The model is briefly presented in Section 2. It incorporates the effect of evolving solar output, infrared cooling, and energy transfer. In Section 3 we compile the main results, whereas Section 4 is devoted to some comments on the implications of the results. Other factors that might have influenced the climate are not considered here (see for instance Bernard, 1974; Nicolet, 1972).

2. The model

The starting point is the energy equation for the earth as a whole

$$\frac{\partial(\text{energy})}{\partial t} = (\text{incoming solar flux}) - (\text{infrared cooling}) + (\text{latitudinal transfer}) \qquad (1)$$

Using the thermodynamic relation $dE = C\,dT$ where T is the temperature and C the thermal inertia, as well as a Fourier type law for the heat transfer, one arrives at the more explicit form (North, 1975a, b)

$$C\frac{\partial T}{\partial t} = \lambda \nabla^2 T - I(T) + \bar{Q}S(\mathbf{r})[1 - \alpha(\mathbf{r}, \mathbf{r}_s)] \qquad (2)$$

λ is the heat conductivity coefficient, I the infrared cooling rate, $\bar{Q}$ the solar constant, $S(\mathbf{r})$ the percentage of incident flux at a position $\mathbf{r}$, and $\alpha(\mathbf{r}, \mathbf{r}_s)$ the albedo. Following Budyko (1969) and North (1975a, b) α is approximated by a discontinuous function around $\mathbf{r}_s$, the locus of the ice boundary.

It is customary to express $I(T)$ starting from the Stefan–Boltzmann law and expanding T around some reference temperature. Keeping linear terms one obtains

$$I(T) = A + BT(\mathbf{r}) \qquad (3)$$

where T now is measured in degrees centigrade.

We are interested in the climatic history of the

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past quarter billion years or so, up to the quaternary period. To this end, it is legitimate to restrict eq. (1) to an ice-free earth

$$\alpha(\mathbf{r}, \mathbf{r}_s) = \alpha_0 \tag{4}$$

where α_0 is space-independent. Next we regard the quantity $\bar{Q}$ as slowly varying in time according to the law suggested by Newmann and Rood (1977)

$$\frac{1}{L}\frac{dL}{dt} \approx \frac{12.5 \times 0.01}{1 + 1.66\,X_0 - 1.66 \times 10^{-2}\,t} \tag{5}$$

where L is the luminosity of the sun, X_0 is the initial hydrogen mass fraction and t is the time in billions of years.

It is convenient to express eq. (2) in spherical coordinates. Performing a longitudinal average and using definition (3) we obtain

$$\frac{C}{B}\frac{\partial I(x,t)}{\partial t} = D\frac{\partial}{\partial x}(1-x^2)\frac{\partial}{\partial x}I(x,t) - I(x,t) + Q(t)\,S(x)(1-\alpha_0) \tag{6}$$

x is the sine of the latitude, and

$$D = \frac{\lambda}{r_0^2 B} \tag{7}$$

where r_0 is the earth's radius. The mean annual distribution of radiation at each latitude $S(x)$ can be expended in Legendre polynomials as follows (North 1975a):

$$S(x) \simeq 1 + S_2 P_2(x) = 1 - 0.482\,\frac{3x^2-1}{2} \tag{8}$$

where the factor S_2 is fitted from astronomical data.

The next point amounts to realizing that the time scale of evolution of I due to planetary factors is much shorter than that arising by the evolving solar output. Hence we may regard the long-term evolution of $I(x)$—or equivalently of $T(x)$—as a sequence of quasi-steady states each one corresponding to the value of $\bar{Q}$ appropriate for a given epoch. Relation (6) therefore yields

$$\frac{d}{dx}(1-x^2)\frac{d}{dx}I(x) - \frac{I(x)}{D} + \frac{3\bar{Q}(1-\alpha_0)}{2D}S_2 x^2 = \frac{\bar{Q}(1-\alpha_0)}{D}\left(\frac{S_2}{2}-1\right) \tag{9}$$

subject to the boundary conditions

$$(1-x^2)^{1/2}\frac{dI}{dx}\bigg|_{x=1} = 0 \tag{10}$$

$$(1-x^2)^{1/2}\frac{dI}{dx}\bigg|_{x=0} = 0$$

expressing the absence of heat transport at the pole and across the equator. The general solution of eq. (9) is of the form

$$I(x) = I_h(x) + I_p(x)$$

where I_p is a particular solution and I_h is the general solution of the homogeneous equation

$$\frac{d}{dx}(1-x^2)\frac{d}{dx}I_h(x) - \frac{I_h(x)}{D} = 0$$

A particular solution satisfying both boundary conditions (10) is easily constructed. It is a quadratic function of x of the form

$$A + BT(x) = I(x) = \frac{\bar{Q}(1-\alpha_0)}{2D+\frac{1}{3}} \times \left(2D + \tfrac{1}{3} - \tfrac{1}{6}S_2 + \tfrac{1}{2}S_2 x^2\right) \tag{11}$$

This solution is meaningful provided $D \neq -\frac{1}{6}$. Here we are interested in positive D's as we expect a heat transfer directed from the equator to the poles.

Now, since eq. (9) is linear and subject only to two boundary conditions, expression (11) will be the *unique* solution of the problem. The situation changes in the presence of ice. In this case more boundary conditions would be needed to determine the ice boundary. The homogeneous solution I_h (which is a superposition of Legendre functions) would then contribute to the solution of eq. (9) (see North, 1975a). We recall however that our analysis deals exclusively with an ice-free earth.

In the sequel we will also use the following equivalent form of eq. (11):

$$2D = \frac{\bar{Q}(1-\alpha_0)\left(\frac{1}{3} - \frac{S_2}{6} + \frac{S_2}{2}x^2\right) - \frac{1}{3}I}{I - \bar{Q}(1-\alpha_0)} \tag{12}$$

3. Results

According to eqs. (11) and (12), the physical conditions prevailing at a given epoch depend on five parameters: the incoming solar flux $\bar{Q}$, the albedo of the earth-atmosphere system α_0, the infrared cooling coefficients A and B, and the latitudinal energy transfer coefficient D. All these parameters are expected to have varied over time. The incoming solar flux $\bar{Q}$, increased systematically, according to eq. (5). The albedo α_0 must have been a variable quantity owing, for instance, to the fluctuations of past level of aerosols arising from volcanic activity (Budyko 1977, 1978). As regards A and B, Sagan and Mullen (1972) suggest an enhanced greenhouse effect through the presence of NH_3 in the early atmosphere, whereas Budyko (1977, 1978), attributes this effect to a high level of CO_2. In particular, it is believed that during the Paleozoic and Mesozoic eras the percentage amount of CO_2 present in the atmosphere varied in the range of 0.1–0.4. From the end of the Mesozoic the CO_2 concentration began decreasing, first at a slow rate and then, faster to reach several hundredths of a percent, at the end of the Pliocene. Finally, the changes in the relative positions of the continents and/or the sea-level (Hays, 1977) must have affected the pole–equator heat transfer coefficient D. Of particular importance is the progressive isolation of the Arctic ocean (Budyko, 1969; Tarling, 1978).

We are interested here in the evolution of polar temperature caused by the evolution of $\bar{Q}$ in conjunction with the following constraints:

(i) Paleotemperature records suggest that for the last 250 myr, equatorial temperatures (T_{eq}) were similar to those at present times (25 °C), while polar temperatures (T_p) may have been much higher, for instance 10–15 °C at 250 myr ago (Hays, 1977).

(ii) As time evolves toward the beginning of the quaternary period, the temperature at the

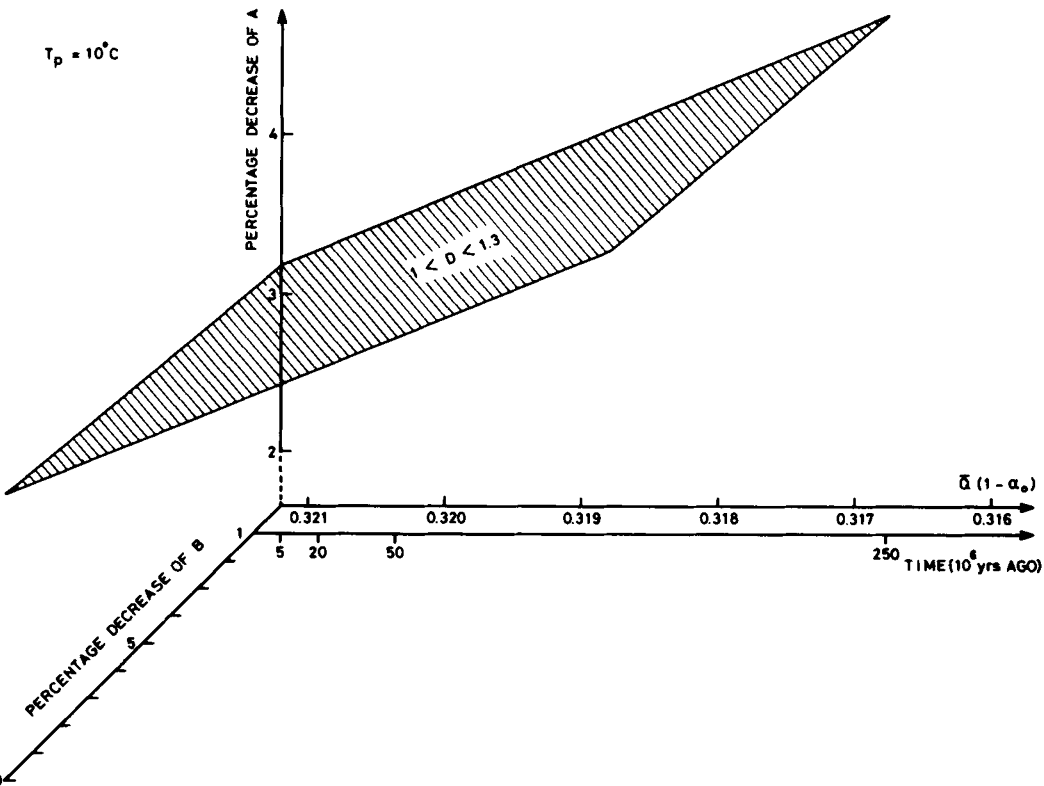

Fig. 1. Dependence of heat influx $\bar{Q}(1 - \alpha_0)$ on percentage decrease of A and B (with respect to their present-day values) corresponding to a value of polar temperature fixed at 10 °C, with T_{eq} = 25 °C. The range of values of the heat transfer coefficient is indicated on the surface.

poles approaches freezing values. Note that as soon as T_p approaches $-10\,°C$ the model leading to eqs. (11) and (12) breaks down and must be replaced by the discontinuous albedo model, eq. (2).

(iii) The emergence of the contrast between polar and equatorial temperature occurs at a slow scale, of the order of several tens of million of years (Budyko, 1974).

For each epoch of interest, constraint (i) enables us to reduce the number of parameters by one. For instance, expressing D through eq. (12), after inserting a given value of T_{eq} one is left with $\bar{Q}$, α_0, A, B. Actually $\bar{Q}$ and α_0 appear only through the combinations $\bar{Q}(1 - \alpha_0)$. Thus the effective parameters reduce further to three, namely $\bar{Q}(1 - \alpha_0)$, A, B. Fig. 1 represents the way these parameters must be coupled in order to maintain a fixed polar temperature.

We now come back to the problem of polar temperature evolution. We consider four characteristic epochs, namely 250, 50, 20 and 5 myr ago. We use for reference the present-day values of $\bar{Q}$, A, B given by North (1975a): $\bar{Q} = 1337.6$ W m^{-2}, $A = 201.4$ W m^{-2} and $B = 1.45$ W m^{-2} °C^{-1}. We then let past values of $\bar{Q}$ vary according to eq. (5), and decrease A, B *simultaneously* up to 10% from present values. From eqs. (11) and (12) we then obtain the corresponding variations of D and T_p. All calculations are carried out with a 6-digit accuracy.

A characteristic feature of the results is their high sensitivity on small variations of the parameters as illustrated in Fig. 2. This fact, which is due to the nonlinear dependence of T_p and D on some of these parameters, limits severely the range of acceptable values of the latter at various epochs. To illustrate further this sensitivity we plot, in Fig. 3, the polar temperature in terms of two of the parameters, namely A and $\bar{Q}(1 - \alpha_0)$ (or time). From both Fig. 2 and Fig. 3 one is led to the general conclusion that A and B, must have been increasing in time, while D is decreasing approaching the present-day value $D = 0.310$.

A more detailed view of the evolution is provided by Fig. 4, which represents two plausible pathways of the polar temperature determined from the previously described procedure for an equatorial temperature fixed at 25 °C and for $\alpha_0 = 0.33$. The values of A, B, D corresponding to the various epochs are given on the curves. We see that if D decreases from 1.637 to 0.492 (i.e. fairly close to its present value) the polar temperature decreases from about 14 °C to freezing values while the infrared cooling increases by a few percent. It is difficult to determine at this time the factors associated with this increase. It would be interesting to examine the results of radiative calculations of the CO_2 contribution to infrared cooling in conjunction with the secular variations of CO_2 content in the atmosphere (see Budyko, 1977). This would give an idea of the role of CO_2 in the decrease reported here. The choice of the relatively high value of α_0 used in drawing Fig. 4 was motivated by some ideas advanced recently (Budyko, 1978) regarding the past level of atmospheric aerosol and volcanic activity. We repeated the numerical simulations for lower values of α_0 and found that the values of T_p and D tended to be unrealistically high, although the general trends remained unchanged.

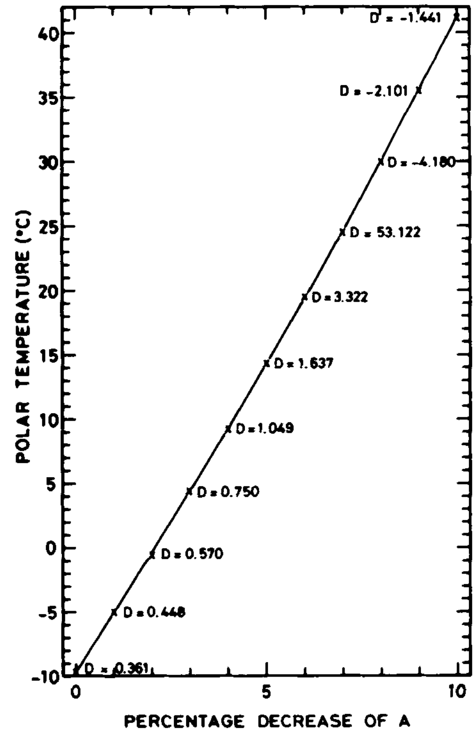

Fig. 2. Polar temperature plotted against the percentage decrease of the infrared cooling coefficient A at 250 myr ago for an equatorial temperature equal to 25 °C and for albedo $\alpha_0 = 0.33$. All values of T_p except that corresponding to a 5% decrease of A from its present value are seen to be unrealistic. The corresponding values of D determined from eq. (12) are also indicated.

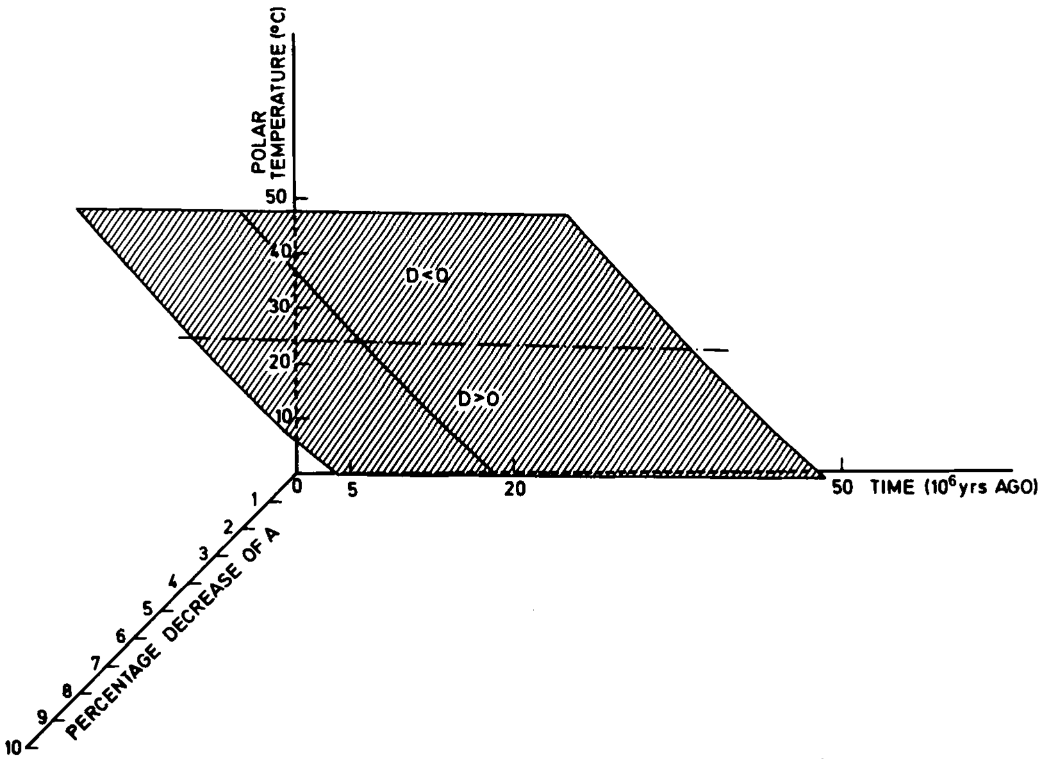

Fig. 3. Polar temperature as a function of the percentage decrease in A and the heat influx $\bar{Q}(1 - \alpha_0)$ (or equivalently, of the time in myr ago). The equatorial temperature is taken equal to 25 °C.

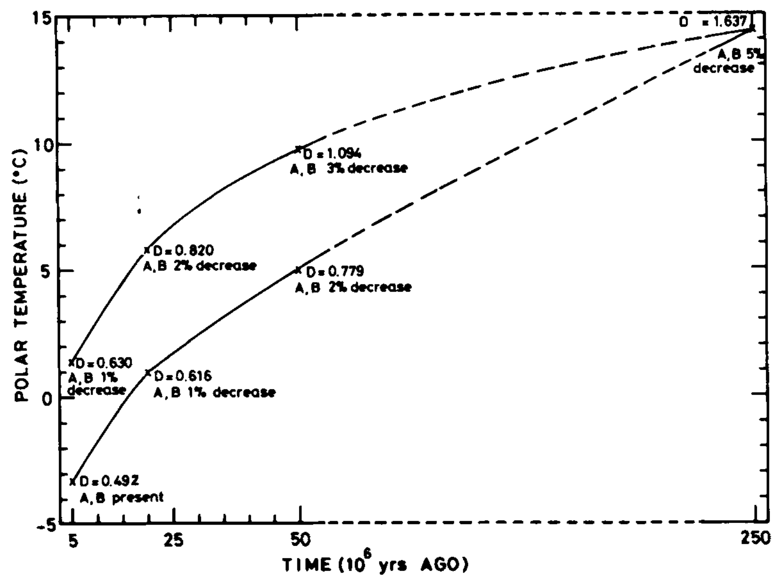

Fig. 4. Two plausible evolutions of polar temperature for equatorial temperature fixed at 25 °C and for albedo $\alpha_0 = 0.33$. D = heat transfer rate. A, B = values of infrared cooling coefficients. At the points on the curves corresponding to 5, 20, 50 and 250 myr ago the values of these parameters resulting from the analysis of Section 3 are indicated.

4. Conclusions

We have seen that it is possible to reconstruct some gross features of the past history of the earth's climate using a longitudinally averaged energy balance model. The analysis reported gives some hints on the factors that may have influenced climate; it also gives some confidence in the use of global modelling for such complex geophysical problems. Obviously, a similar approach can be used to predict some future trends. Work in this direction is in progress.

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ВЛИЯНИЕ СВЕТИМОСТИ СОЛНЦА, ИНФРАКРАСНОГО ВЫХОЛАЖИВАНИЯ И ШИРОТНОГО ПЕРЕНОСА ТЕПЛА НА ЭВОЛЮЦИЮ КЛИМАТА ЗЕМЛИ

С помощью одномерной планетарной модели делается попытка реконструкции некоторых особенностей истории климата Земли. Начиная с периода 250 миллионов лет тому назад до начала четвертичного оледенения, намечаются возможные эволюционные пути, принимая во внимание эволюцию светимости Солнца, палеоклиматические, а также географические данные. Основной вывод из результатов состоит в том, что несмотря на детали эволюционных путей коэффициенты инфракрасного выхолаживания и переноса тепла также с необходимостью должны эволюционировать. Представлены некоторые аргументы для определения направления этой эволюции.