JOURNAL OF GEOPHYSICAL RESEARCH, VOL. 87, NO. C9, PAGES 7295–7302, AUGUST 20, 1982

The Faint Young Sun-Climate Paradox: Continental Influences

ANDREW S. ENDAL

Applied Research Corporation, Landover, Maryland 20785

KENNETH H. SCHATTEN

Laboratory for Planetary Atmospheres, Goddard Space Flight Center, Greenbelt, Maryland 20771

We examine the various mechanisms which have been proposed to compensate for the climatic effects of a 30% increase in the solar luminosity over the past 4½ billion years. Although atmospheric greenhouse effects have received most attention, other mechanisms may have played a role of comparable importance. In particular, we note that the development of continents during the past 2½ billion years could have had a significant secular effect on the atmosphere-ocean heat transport system. As a result, past climates may have been less susceptible to complete freeze-over. A simple energy balance model is used to demonstrate the magnitude of this effect. Because the CO_2 greenhouse effect is not the only means of compensating for solar evolution, the faint-young-sun problem should not be used to infer past levels of atmospheric CO_2.

1. INTRODUCTION

Stellar evolution theory [see *Schwarzschild,* 1958] predicts that the sun was initially about 30% fainter than at present. The luminosity then increased in a roughly linear fashion to its present value some 4.7 billion years later. Since the earth's climate is determined by a balance between incoming solar radiation and outgoing infrared radiation, this evolution would have produced a 9%, or 25°C, change in the earth's effective radiating temperature, if all other factors remained constant. Due to the positive feedbacks provided by the ice-albedo and water-vapor greenhouse effects, energy balance models predict that the corresponding change in the surface temperature would be much greater [*Budyko,* 1969]. Since annual-mean surface temperatures now range from −49°C at the south pole to 27°C near the equator [*Warren and Schneider,* 1979], the direct inference is that the earth was totally frozen for much of its history. This conflicts with the geological record of glaciations [cf. *Frakes,* 1979]. In addition, reflection of solar radiation by snow and ice is so efficient that a totally glaciated earth would have remained frozen even for a solar luminosity 30% larger than its present value [cf. *North,* 1975].

This 'faint young sun' problem has stimulated a number of investigations into terrestrial effects which could compensate for the changing solar luminosity and maintain climatic conditions closer to those of the present. Following the early work of *Sagan and Mullen* [1972], most of these investigations have exclusively considered an enhanced greenhouse effect due to some postulated change in the composition of the earth's atmosphere. The recent consensus [*Owen et al.,* 1979; *Henderson-Sellers and Schwartz,* 1980; *Wigley,* 1981; *Walker et al.,* 1981] seems to favor the greenhouse effect due to a larger abundance of carbon dioxide in the early atmosphere. While this may be an entirely reasonable solution, consideration should be given as to whether it is the complete solution. This is particularly true considering recent trends to reverse the reasoning and use the faint-young-sun problem to infer the chemical history of the earth's atmosphere [e.g., *Wigley and Brimblecombe,* 1981].

Paper number 2C0941.
0148-0227/82/002C-0941$05.00

In this paper, we examine the faint-young-sun problem for the specific purpose of determining whether the CO_2 greenhouse solution is necessary and/or unique. We begin in section 2 by considering the chronology of solar and terrestial evolution in order to develop constraints on greenhouse and other mechanisms for compensating for solar evolution. Specific compensatory mechanisms are examined in section 3. This examination indicates that independent geochemical arguments on the CO_2 greenhouse are ambiguous and that other factors may have played a comparable role in determining the climate. These and related conclusions are discussed in section 4.

2. SOLAR EVOLUTION AND TERRESTRIAL EVENTS

The solid curve in Figure 1 shows the evolution of the solar luminosity (in units of the present luminosity L_S). This curve was computed with the stellar evolution code described by *Endal and Sofia* [1981]. The age of the earth is generally placed at 4.5 to 4.65 billion years (b.y.). We adopt 4.7 b.y. as the age of the sun, based on the expectation that the more massive sun contracts to a near-equilibrium state faster than the earth. The other important parameters which must be specified to calculate the evolution are the total mass and the initial composition. The mass can be directly measured by orbital mechanics while the initial composition is set equal to the spectroscopically determined surface composition of the present sun [*Ross and Aller,* 1976]. The latter procedure is internally consistent in that the models do not predict any significant nuclear processing of the surface layers and this is confirmed by observations of solar-type stars with a range of ages. The solar helium abundance cannot be directly determined, due to the absence of suitable absorption lines in the solar spectrum. The initial helium abundance is, therefore, adjusted to give the observed solar luminosity at an age of 4.7 b.y. To a limited degree, this can be checked by observations of other types of stars where suitable helium lines are present in the spectra.

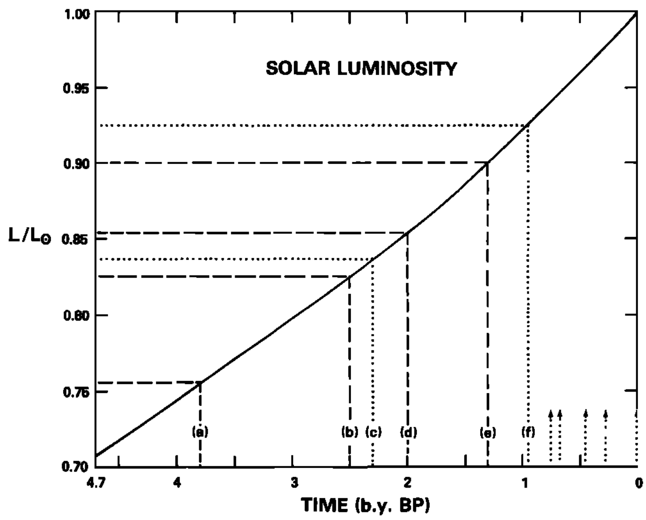

Fig. 1. Predicted solar luminosity as a function of time. The dotted lines indicate times of major glaciations, as summarized by *Tarling* [1978*a*]. The dashed lines refer to the following events: (*a*) oldest known rocks, earliest evidence of life and liquid H_2O [*Cloud*, 1976]; (*b*) Proterozoic interval [*Sutton and Watson*, 1974], (*d*) beginning of accumulation of substantial free oxygen [*Cloud*, 1976], and (*e*) formation of rigid crustal plates [*Sutton and Watson*, 1974].

Because of the freedom in choosing the initial helium abundance, there are no simple direct checks that the solar models accurately portray the present state and history of the sun. However, as summarized by *Iben* [1967], the basic predictions of stellar evolution theory (including the luminosity prediction) have been confirmed by a wide variety of observational tests applied to other stars. Furthermore, simple scaling arguments [*Newman and Rood*, 1977; *Endal*, 1981] show that the luminosity prediction is "robust" in the sense that the result is not sensitive to uncertainties in the physical data used to construct solar models. This is very different from the prediction of the solar neutrino flux, which is very sensitive to detailed physical data. For this reason, uncertainties regarding the solar neutrino controversy do not transfer to the solar luminosity prediction. We emphasize this point because it has apparently been a source of confusion [e.g., *Pollack*, 1979].

According to our calculations, the initial luminosity of the sun (once it settled to an equilibrium state) was $L = 0.71\ L_S$. The subsequent increase in luminosity is not quite a linear function of time. For instance, 41% of the luminosity increase takes place during the first half of the evolution, while 59% of the increase occurs in the second half. The simple formula

$$L(t) = [1 + 0.412(1 - t/t_S)]^{-1} L_S \qquad (1)$$

gives an excellent fit to the curve. In equation (1), t is the elapsed time since the formation of the sun and t_S is the present age (4.7 b.y.). Scaling arguments predict a similar functional form for $L(t)$ and show that this form has a basic physical explanation [*Endal*, 1981]. Complicating factors such as the initial rotation rate and the possibility of a strong magnetic field in the solar core contribute, at most, a 5% uncertainty to these predictions (i.e., the true luminosity increase is somewhere between 25 and 35%). Considering the other uncertainties of the faint-young-sun problem, this is entirely negligible.

We now turn to the corresponding history of the earth, for which the vertical and horizontal lines in Figure 1 indicate a number of important events and boundaries. The dotted lines indicate the major glacial events, following *Tarling* [1978*a*]. The impression obtained from Figure 1 is that the frequency of glacial events tends to increase with time. Prior to the Proterozoic interval, event (b), the absence of known glaciations could be explained either by the lack of continent-sized land masses to act as suitable foundations for glaciers or, perhaps more simply, by the sparseness of the geological record. However, neither of these explanations applies subsequent to the Proterozoic interval. The general trend toward more frequent glaciations is consistent with the decreasing surface temperatures found by *Knauth and Epstein* [1976] from isotopic analysis of cherts formed over the past 3 b.y. This runs contrary to the naive expectation from the solar luminosity curve and points, once again, to the need for additional influences to compensate for the solar evolution.

The Huronian glaciation, event (c), is of some interest with regard to possible compensatory mechanisms. The most extensive evidence for the Huronian glaciation is found in the Canadian shield, but glacial deposits have also been found in South Africa and India [*Tarling*, 1978*a*]. The glaciations at the widely distributed locations were not necessarily synchronous and the Huronian glaciation may actually refer to a series of more local events. This picture is supported by the occurrence of three distinct levels of glacial strata in the Canadian shield. The evidence points to a fairly extended period with conditions conducive to glacier formation [*Frakes*, 1979].

The relevance to compensatory mechanisms is that the Huronian glaciation implies that a transition to a lower mean surface temperature was possible at a very early time. This event (or series of events) was followed by a much longer period (of at least 1 b.y.) of apparently warmer climates, devoid of permanent snow or ice cover. Such behavior argues against a simple monotonic compensatory mechanism such as the decrease in atmospheric CO_2 due to the rise of biological activity, though fluctuations in the atmospheric CO_2 concentration due to fluctuations in the level of volcanic activity [cf. *Budyko and Ronov*, 1979] offer a possible explanation.

For the purpose of examining possible compensatory mechanisms, it would be useful to have values for the mean surface temperatures over a considerable fraction of the earth's history. We might, for instance, adopt the ground water temperatures determined by *Knauth and Epstein* [1976] from cherts collected in the middle and western United States. However, these temperatures refer to specific locations and the inferred temperature history could be severely influenced by changes in latitude due to continental migration. Until more extensive data are available, we can only require that potential compensatory mechanisms prevent a total glaciation of the earth. Considering that the climate is now more glacial than the long-term average [cf. *Tarling*, 1978*a*], we will require that compensatory mechanisms have maintained mean surface temperatures near or above the present value.

3. COMPENSATORY MECHANISMS

To counteract the monotonically increasing solar luminosity, the magnitude of a potential compensatory mechanism must evolve (more or less) steadily over a substantial fraction of the earth's history. We will examine three possibilities: (1) a decreasing rate of heat flow from the earth's interior, which supplements the solar input, (2) an atmospheric greenhouse effect which decreases in magnitude due to the chemical evolution of the atmosphere-lithosphere system, and (3) the effect of the evolution of the lithosphere on heat transport by the atmosphere-ocean system. The first mechanism directly alters the energy budget of the earth, while the second and third mechanisms affect primarily the temperature gradients in the vertical and meridional directions, respectively. The latter two mechanisms may also alter the energy budget by their effect on the planetary albedo (ice-albedo feedback).

3.1. *Heat Flow From the Interior*

The present solar irradiance is 1368.3 W m^{-2} [*Willson et al.*, 1981] and, with a planetary albedo of 0.30 [*Ellis et al.*, 1978], the heat input from the sun is 1.2×10^{17} W. The present rate of heat flow from the earth's interior is estimated to be 4.2×10^{13} W [*Sclater et al.*, 1980], i.e., 3.5×10^{-4} times smaller than the solar influence. At earlier times, the heat flow was certainly larger and so the crucial question is whether it could have been large enough to compete with the solar influence for a substantial fraction of the earth's history. This can be ruled out by considering the relevant energetics.

If we assume that the heat flow from the interior was sufficient to maintain the total (solar plus interior) heat input to the climate system at its present (solar) value, and we integrate the interior component over the age of the earth, then the energy to be extracted from the interior amounts to 3×10^{33} J. The largest potential source of energy is the gravitational binding energy released in the formation of the earth. The total binding energy of the earth is 2.5×10^{32} J [*Allen*, 1973], which is too small by an order of magnitude. Even if this energy was released at a rate optimal for the faint-young-sun problem, it could only compensate for 10% of the solar luminosity change. In reality, most of the energy was released very early during the formation of the earth [*Hanks and Anderson*, 1969]. The present source of heat from the earth's interior is radioactive decay; this energy reservoir is smaller than the gravitational energy by yet another order of magnitude.

3.2. *Greenhouse Effects*

As mentioned in section 1, the greenhouse effect has received much attention as a mechanism for compensating for solar evolution. Briefly, atmospheric constituents such as H_2O, CO_2, and (to lesser extents) O_3 and N_2O absorb infrared radiation from the earth's surface. As a result, the surface temperature is raised above the effective radiating temperature. With the present atmospheric composition, the greenhouse effect increases the mean surface temperature by some 30 K [cf. *North et al.*, 1981]. Doubling the infrared absorption of the atmosphere would be sufficient to compensate for solar evolution so it is entirely conceivable that the earth was kept warm by such an effect.

Specific scenarios using the greenhouse mechanism to compensate for the changing solar luminosity have been described by *Sagan and Mullen* [1972], *Hart* [1978], and *Owen et al.* [1979]. *Owen et al.* [1979; see also, *Kuhn and Atreya*, 1979, and *Kasting*, 1982] have offered a number of arguments against the ammonia- and methane-based greenhouse effects suggested by *Sagan and Mullen* [1972] and *Hart* [1978]. These arguments are fairly compelling and restrict the roles of reduced-gas greenhouses to a small fraction of the earth's history, so we will consider only the effects of H_2O and CO_2. In particular, we will focus on the question of whether a CO_2 greenhouse effect sufficient to compensate for solar evolution necessarily follows from the current picture of the evolution of the atmosphere.

For a given solar luminosity, the atmospheric CO_2 content required to maintain the present mean surface temperature can be estimated from radiative-convective atmosphere models. *Augustsson and Ramanathan* [1977] find that doubling the CO_2 concentration has roughly the same effect as a 2% change in the solar luminosity (both producing a 2–3 K change in the model surface temperature). However, this result cannot be linearly extrapolated to estimate the effects of larger changes in CO_2 concentration because of saturation of the infrared absorption bands. Calculations for very large CO_2 concentrations have been presented by *Owen et al.* [1979], in conjunction with their greenhouse scenario. From their Table 1, we find that a factor of 27 increase in CO_2 is required to maintain present surface temperatures when the sun was 11% fainter than now. Assuming a linear evolution of the solar luminosity, Owen et al. estimate that the sun was 11% fainter some 2 b.y. BP. From our equation (1) (or Figure 1), we estimate that this point was reached only 1.4 b.y. BP. It appears that at least a factor of 30 change in atmospheric CO_2 concentration is required to totally compensate for solar evolution during the past 2 b.y. Much larger changes are required for earlier times. We can examine the physical processes which have led to the present atmospheric composition to see whether such large (and recent) changes are to be expected.

The present atmosphere accumulated by outgassing from the surface of the earth after the primordial atmosphere escaped the earth's gravitational field. The most conspicuous example of such outgassing is now associated with volcanic fumaroles. Although fumarolic gasses are partially recycled, we may use data on their composition as, at least, a qualitative guide to the products of earlier outgassing [*Walker*, 1977]. The major constituents of fumarolic gases are H_2O (~97% by weight) and CO_2 (~2%), with N_2 present at a level of ~0.1% [*White and Waring*, 1963]. Since the atmosphere is now dominated by N_2, substantial sinks of H_2O and CO_2 are required to produce the present atmosphere by outgassing.

The obvious sink for water vapor is condensation. The existence of very old sedimentary rocks indicates that condensation occurred at an early epoch, at least 3.8 b.y. BP [*Moorebath et al.*, 1973]. From this point on, the partial pressure of water vapor was controlled by equilibrium between evaporation and condensation. If this equilibrium maintains roughly constant relative humidity [*Manabe and Wetherald*, 1967], the H_2O greenhouse effect acts as a positive feedback for solar luminosity changes and, therefore, is unsuitable as a compensatory mechanism.

Once the oceans formed, they also provided a sink for CO_2 by solution of bicarbonate ions in water and subsequent precipitation of solid carbonates. The chemistry of this process is described in detail by *Walker* [1977, chapter 3], so we will consider only a schematic representation. Weathering of surface silicate minerals by reactions such as

$$CaSiO_3 + 2CO_2 + H_2O \rightarrow Ca^{++} + 2HCO_3^- + SiO_2 \quad (2)$$

removes carbon dioxide from the atmosphere. The products of such weathering are carried by rivers to the oceans, where carbonate precipitation occurs via reactions such as

$$Ca^{++} + 2HCO_3^- \rightarrow CaCO_3 + H_2O + CO_2 \quad (3)$$

The net result of this (nonbiological) process is removal of CO_2 from the atmosphere-ocean system and conversion of silicate minerals to carbonate minerals.

As pointed out by *Rasool and de Bergh* [1970], the critical difference between Venus and the earth—leading to a massive CO_2-dominated atmosphere on the former and a moderate N_2-dominated atmosphere for the latter—is that the earth's atmosphere passed through the region of the phase diagram where H_2O exists as a liquid. The early formation of a hydrosphere, rather than later biological activity, is thought to be responsible for removal from the earth's atmosphere of most of the outgassed CO_2. In terms of compensatory mechanisms, the critical issue concerns the time scale for removal of the excess atmospheric CO_2.

The rate of CO_2 removal is presently controlled by the rate of silicate weathering. At the current weathering rate, 2 b.y. would be required to build up the carbonate deposits to their present level of 5×10^{21} moles of CO_2. (This is derived from *Walker*'s [1977] estimate that 2×10^5 yr would be required to convert the 6×10^{17} moles of CO_2 contained in the fossil-fuel reservoir into carbonate minerals.) At this rate, equilibrium partial pressures of CO_2 could have been established during the first half of the earth's evolution. The CO_2 greenhouse effect would have played a dominant role during the first 2 b.y. after formation of the oceans, but other influences would still be required to compensate for the substantial increase in the sun's luminosity during the second half of the earth's history.

Several factors could have substantially increased the time required to reduce the high atmospheric CO_2 levels produced by volcanic emanations:

1. Scenarios for the chemical evolution of the oceans [e.g., *Garrels and Perry*, 1974; *Schopf*, 1980, chapter 5] generally start with a slightly acidic mixture (pH $\sim$ 6.5), reflecting the chemistry of volcanic gases. Since carbonates dissolve for pH < 7.0, reaction (3) would have been prevented until reactions on basic igneous rocks raised the oceanic pH closer to the present values of 7.5 to 8.5. The existence of chemically precipitated carbonate deposits in the 3.8 b.y.-old Isua, Greenland formation [*Moorebath et al.*, 1973] shows that the required pH values were reached (at least locally) at an early epoch;

2. Geological processes leading to volcanic recycling and weathering of carbonate deposits return CO_2 to the atmosphere. This reduces the net rate of removal of atmospheric CO_2. On the other hand, the rate of carbonate deposition should have been greater in the past when atmospheric CO_2 concentrations were larger, and recycling would only compete effectively with deposition once the equilibrium state was approached;

3. In the present carbon budget, roughly 20% is contained as organic sedimentary deposits. Even if the carbon cycle approached equilibrium very early, the equilibrium partial pressure of CO_2 may have been higher in the prebiology era than at present. It is difficult to estimate how large an effect this would have on the atmospheric CO_2 level since both the inorganic and organic carbon reservoirs, as well as the oceanic reservoir, are much larger than the atmospheric content. The atmosphere thus plays only a minor role in total carbon storage. *Berkner and Marshall* [1966] suggest that it is unlikely that the equilibrium partial pressure of CO_2 was ever greater than 10 times its present value. *Budyko and Ronov* [1980] suggest similar increases above the present value due to departures from carbon-cycle equilibrium caused primarily by variations in the level of volcanic activity.

Perhaps most important, the present rate of weathering of surface silicate minerals is influenced by a variety of physical processes which have not been even approximately constant throughout the history of the earth. Such processes include the rate of surface water run-off and the rate of exposure of fresh silicate minerals by geological uplift and erosion. Given that the time scale estimated above is marginal (a factor of 2 increase or decrease would substantially alter the conclusions), the inherrent uncertainties preclude a firm conclusion regarding the importance of the CO_2 greenhouse as a compensatory mechanism. Simulations of the chemical evolution of the atmosphere [e.g., *Hart*, 1978] are not much help, either. The number of assumptions and adjustable parameters in such simulations is at least as large as the number of available constraints. In particular, *Hart* [1978] assumed that a CO_2 greenhouse effect maintained constant surface temperatures during the second half of the earth's history to constrain the possible solutions. Such models may indicate that compensation for the increasing solar luminosity by a decreasing CO_2 greenhouse effect is plausible, but they cannot prove the case. Direct arguments against this mechanism are likewise inconclusive. For this reason, it may be useful to estimate the magnitude of possible competing influences.

3.3. *Changes in Meridional Heat Transport*

The greenhouse effect alters the relationship between the surface temperature and the effective radiating temperature of the earth. This can be viewed as a change in the vertical distribution of temperature in the atmosphere. We should also consider the possible influence of changes in the horizontal (latitudinal) temperature distribution, near the surface. Such changes may not directly alter the mean surface temperature, but they will affect the conditions for onset of glaciation. As noted in section 2, the absence of significant glaciation throughout most of the earth's history provides the strongest evidence for the existence of compensatory mechanisms.

The latitudinal temperature distribution is determined by the distribution of incoming solar radiation and the poleward transport of energy by the atmosphere-ocean system. The changing topography of the earth's surface may alter the efficiency of poleward energy transport and, hence, the latitudinal temperature gradient. Before considering specific mechanisms by which surface topography may affect energy transport, we will briefly review the major stages in the evolution of the earth's crust. This will serve to identify long-term trends relevant to the faint-young-sun problem.

Recent interpretations of the geological record favor an evolutionary development with changing tectonic patterns, as opposed to the uniformitarian view that the modern pattern of plate tectonics has dominated for most of the earth's history. *Sutton and Watson* [1974] divide the evolution of the earth's surface into three distinct stages, based on the qualitative features (style) of orogenic activity. Variations on this theme have been developed from models of mantle convection [*McKenzie and Weiss,* 1975], from models for accumulation of sedimentary deposits [*Hargraves,* 1976], from paleomagnetic data [*Embleton and Schmidt,* 1979], and from interpretations of rock types and their tectonic settings [*Goodwin,* 1981]. Although these interpretations differ in detail, they all agree that evolution, rather than a steady state, has been the rule up to relatively recent times.

Following *Sutton and Watson* [1974], we specify three major regimes: (1) the Archean regime, prior to 2.5 b.y. BP; (2) the early Proterozoic regime, 2.5 to 1.3 b.y. BP; and (3) the late Proterozoic and Phanerozoic regime, from 1.3 b.y. BP to the present. The Archean regime contains the most dramatic changes, including differentiation and consolidation of most of the crust during the first 1.0 b.y. [*Hargraves,* 1976] and accumulation of almost all of the oceans during the first 2.0 b.y. [*Schopf,* 1980]. Permanent crustal masses were small and highly mobile. The second regime starts at the Proterozoic interval (event (b) in Figure 1), which marks the first appearance of large and stable continental masses. Common pole wandering curves indicate the existence of a few stable 'supercontinent' sheets from 2.3 b.y. BP to, at least, 1.6 b.y. BP [*Embleton and Schmidt,* 1979]. The characteristic orogenic features are the complex systems of mobile belts within the large sheets [*Sutton and Watson,* 1974]. These belt systems resulted from internal deformations of the relatively flexible crustal sheets. This is in sharp contrast to the tectonic style of the present regime, which is dominated by disruptions and collisions of rigid plates, with orogenic activity concentrated at the leading and trailing edges of the plates. Within this last period (starting at (e) in Figure 1), independent drifting of continent-sized plates has been the rule since, at least, 1.0 b.y. BP [*Embleton and Schmidt,* 1979].

The common feature in these successive regimes is the emergence and thickening of continental plates, separated by ocean basins. This resulted from progressive segregation of the sialic (granitic) crust into continental formations. During the Archean regime, this was restricted to localized massifs which, during the early Proterozoic regime, became connected by extensive, shallow, and deformable sheets of sial. Continued thickening of these sheets led to rigidity, causing the sheets to break up into the smaller (continental) plates of the late Proterozoic and Phanerozoic regime.

An important physical constraint on the timing of this process has been pointed out by *Hargraves* [1976], based on the requirement that the temperature at the base of the sial must be less than 750°C to maintain a solid state. As the heat flux from the interior diminished, and the geothermal gradient decreased, the depth below the surface of the 750°C isotherm increased. This depth can be calculated from cooling models, such as those of *McKenzie and Weiss* [1975], once the initial abundances of radionuclides in the crust and mantle are specified. For chondritic initial abundances, the depth of the 750°C isotherm (in granite) is 16 km at 2.5 b.y. BP, 25 km at 1.3 b.y. BP, and 38 km at 0.5 b.y. BP. For the initial abundances inferred from surface rocks on the earth [*Wasserburg et al.,* 1964], the corresponding depths are 21, 35, and 41 km, respectively. By comparison, the continental crust is now 30 to 40 km thick and, if spread uniformly over the earth, would form a sialic layer approximately 12 km deep.

With this geological background, we can consider possible effects on the global climate. In general, we expect that the progressive development of large variations in surface elevation (both on land and in the ocean basins), impedes meridional heat transport. Changes in the total surface area of the oceans may be particularly important. Using an isostatic model, with constant oceanic and sialic volumes, *Hargraves* [1976] finds that the mean continental elevation would have been below sea level until the thickness of the continental crust approached 30 km. The 750°C isotherm would have reached this depth between 1.4 and 0.9 b.y. BP. Prior to this time, highland regions would have existed above sea level (as required by fluvial deposits in early Proterozoic formations), but the oceans would have been less constrained to well-defined ocean basins. This may have allowed freer ocean circulation, more closely reflecting the zonal atmospheric driving. The sensitivity of ocean circulation to changes in the land-sea boundaries has been demonstrated by the laboratory simulations of *Luyendyk et al.* [1972]. At present, the oceans carry an average of 40% of the poleward heat flux in the latitudes 0–70°N [*Vonder Haar and Oort,* 1973]. Circulation is constrained by ocean-basin boundaries and almost all of the water resides in deep basins, with very long turnover times. With the postulated greater ocean circulation of the Archean and early Proterozoic regimes, the oceans may have dominated meridional energy transport.

The flatter topography of the earlier surface may also have influenced atmospheric circulation. However, numerical simulations of the general circulation, with and without mountains, suggest that orographic effects on global climate may be minimal [*Karahara and Washington,* 1971; *Manabe and Terpstra,* 1974]. In this case, the effect of the rise of continents on ocean circulation would be of primary interest.

We will not attempt a direct calculation of the effects on climate of greater ocean circulation. Rather, we will use an idealized climate model to explore the general effects of more efficient heat transport by the atmosphere-ocean system. We choose the one-dimensional (latitudinal) energy balance model derived by *North* [1975]. This model describes meridional heat transport as a diffusive process characterized by a global diffusion coefficient, D. Energy balance is obtained by setting energy transport into a given latitude zone equal to the net radiative loss (outward infrared radiation minus absorbed solar radiation). The model allows for the temperature dependence of the infrared absorption (using the *Budyko* [1969] parameterization with constant cloudiness) and the dependence of the albedo on ice cover. A annual-mean latitudinal distribution of solar radiation is used, so seasonal variations are suppressed. While such a simple model cannot be expected to give highly accurate results, the physical reasons for the response of the climate to external forcing are clearly brought out.

The equations and data which specify this model are given in the Appendix. The free parameters are D, the diffusion coefficient, and Q, the solar radiation incident on the top of the atmosphere. For given values of D and Q, the model predicts the surface temperature (as a function of X = sine (latitude)) and X_s, the location of the ice-cap edge. The edge of the ice cap is placed at the latitude with a (annual-mean) temperature of −10°C, as observed for the present climate [*Budyko*, 1969].

The diffusion coefficient which characterizes the present atmosphere-ocean system is determined by matching predicted surface temperatures to the observed climate, using the present solar radiation ($=Q_0$). We find that a dimensionless diffusion coefficient $D = 0.267$ gives a reasonable fit. With this fixed value for D, the location of the ice cap as a function of Q/Q_0 is shown in Figure 2. The portions of the curve indicated by dotted lines are unstable and, therefore, inaccessible to a steady-state climate. These solutions (with negative slope) suggest that the ice caps grow in response to an increase in the solar radiation, which is not physical. Mathematical instability of such solutions has been demonstrated by *North* [1975] for this model and, for more generalized models, by *Cahalan and North* [1979].

The solutions shown in Figure 2 indicate that, if the incident solar radiation is decreased, the ice caps expand from $X_s = 0.95$ (at $Q/Q_0 = 1$) to $X_s = 0.60$ (at $Q/Q_0 = 0.95$). The high albedo of the ice caps provides a positive feedback to changes in Q since expansion of the ice in response to a decrease in incident solar radiation further decreases the fraction of that radiation which is absorbed. For $Q/Q_0 < 0.95$, this feedback dominates to the point that the only possible solution corresponds to an ice-covered earth ($X_s = 0.0$). Given that the solar luminosity has changed by considerably more than 5% over the history of the earth, this model clearly demonstrates the need for compensatory mechanisms to prevent total freeze over. According to equation (1), the solar luminosity was lower than the critical value ($L/L_S = 0.95$) for times earlier than 0.6 b.y. BP. Unless other climatic factors intervened, the earth would have been totally glaciated and the high albedo would have maintained this state up to some future time when $L/L_S > 1.3$ (see Figure 2). The stability of the 'white earth' solution, with the present solar input, has been demonstrated with general circulation models [*Wetherald and Manabe*, 1975], as well as with the simpler energy balance models.

We can qualitatively explore the effect of changes in the atmosphere-ocean heat transport by varying the global diffusion coefficient in the climate models. Models for various values of D are shown in Figure 3. We have not differentiated between stable and unstable portions of the solutions since stability can be easily determined by examining the slopes of the curves. We have also deleted the ice-free and ice-covered branches. The dashed lines indicate the limits of no transport and infinite transport. Within this range, it is difficult to prescribe direct limits on D since the global diffusion coefficient characterizes the integrated effects of many local transport subsystems. An indirect upper limit can be set by noting that, for $D \geq 1.5$, the equatorial temperature drops below 0°C, even for $X_s = 1.0$. However, this limit is not very restrictive in the present context since solutions for such large values of D converge rapidly to the infinite diffusion solution.

Examination of Figure 3 reveals several interesting aspects of this climate model. First, the lower limit on Q/Q_0 for an ice-free solution occurs where the curves intersect $X_s = 1.0$. As D is increased (possibly representing past climates with greater free ocean circulation) this lower limit decreases, suggesting that a smaller solar luminosity can be compensated by increased energy transport to prevent a freeze over. The limit on such compensation corresponds to $Q/Q_0 = 0.81$ for the infinite diffusion case and to $Q/Q_0 = 0.84$ for $D = 1.5$. Second, we find that for $D \gtrsim 0.05$ the curves have negative slopes everywhere. In this case, the only stable solutions correspond to the ice-free and ice-covered branches. This suggests that an ice age which stops short of total freeze-over is not possible in a vigorously circulating climate. It is interesting that the early Proterozoic regime, which we have tentatively identified as a period of enhanced oceanic heat transport, roughly coincides with the long period between 2.3 and 0.9 b.y. BP which is remarkable for its lack of ice ages (see Figure 1). Finally, it appears that changes in the global transport efficiency can initiate ice ages in this model. Consider, for example, the solutions for $Q/Q_0 = 1.0$. With $D = 0.4$, an ice-free solution is possible but there are no stable solutions corresponding to partial ice cover. If the diffusion coefficient drops to 0.2, the only stable solutions correspond to an ice cap extending to $X_s = 0.81$ and the ice-covered solution.

Of course, conclusions drawn from such a simplistic

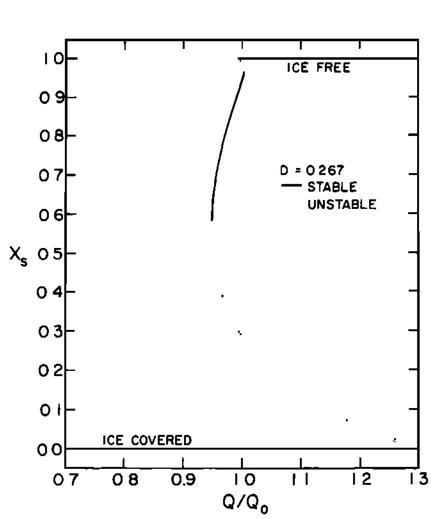

Fig. 2. Position of the ice-cap boundary as a function of the solar radiation input for a model with the diffusion coefficient for the present climate.

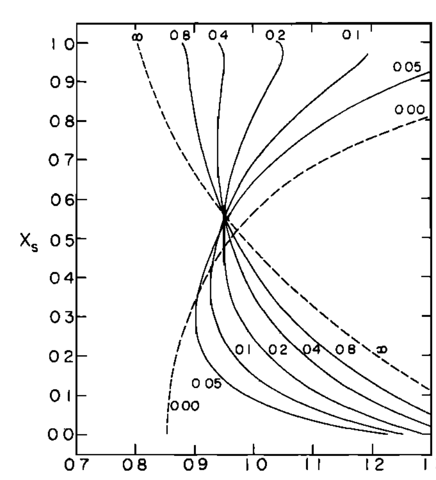

Fig. 3. Position of the ice-cap boundary as a function of the solar radiation input for models with various diffusion coefficients. The values of the dimensionless diffusion coefficient (see appendix) are given next to the curves.

model must be treated with caution. Although attempts have been made to derive energy balance models capable of quantitative predictions [cf. *Barron et al.*, 1981], the model we have used is much simpler and was designed to study qualitative and mathematical properties of climate models. Further, it is not clear that diffusion, which refers to a small-scale, random-walk type of transport, can provide an adequate description of the effects of the large-scale ocean circulation. Nevertheless, this model clearly demonstrates the important influence of meridional heat transport.

4. CONCLUSIONS

As stated by *Budyko* [1977], 'the climate is determined by the solar radiation incident on the upper boundary of the atmosphere, by the atmospheric composition, and by the structure of the earth's surface.' Models of the sun predict that the solar radiation has monotonically increased by 0.3 L_S over the past 4.7 b.y. Since the earth's surface temperature has remained remarkably constant during this time, one or both of the other climatic factors stressed by Budyko must have compensated for the change in the solar input. Traditionally, only changes in atmospheric composition (producing a greenhouse effect) have been considered as a compensatory mechanism. Although this remains an interesting area of investigation [e.g., *Walker et al.*, 1981], we believe that changes in the surface structure should also be considered. We suggest that such changes affect the climate primarily by altering the effectiveness of meridional heat transport, particularly the component of heat transport due to the oceans. Note that the increased heat transport is not a direct heating mechanism like the greenhouse effect, but rather a moderating mechanism. Nevertheless, by preventing the formation of ice cover and a concomitant increase of the earth's albedo, heat transport can greatly affect the transition to a frozen-over state and maintain moderate surface temperatures.

In section 3.3, we estimated that changes in the structure of the earth's surface can, at most, compensate for a reduction in the solar luminosity to 0.84 L_S, through changes in meridional heat transport. This represents only 55% of the predicted solar change and the actual compensation by this mechanism may be less. Other mechanisms are still required. It is probably best to assume that the greenhouse effect and changes in poleward heat transport have both played roles in the long-term evolution of the climate. We should also note that the geological changes described in section 3.3 will have important effects on the rate of removal of CO_2 from the atmosphere. This will affect the time scale estimated in section 3.2.

Finally, we emphasize the speculative nature of the entire faint-young-sun problem. At present, the only statements which can be made with some certainty are (1) that the solar radiation incident on the atmosphere has increased by approximately 30% (of its present value) over the history of the earth, (2) that this change in solar energy input cannot be compensated by energy sources arising from the interior of the earth, and (3) that the absence of a paleoclimatic signature of the increasing solar luminosity requires some compensatory mechanism(s). Because of the possible operation of more than one compensatory mechanism, it is particularly important that the changing solar luminosity not be used to estimate past concentrations of atmospheric CO_2.

APPENDIX

In the diffusive energy balance climate model derived by *North* [1975], latitudinal energy transport balances the net radiative loss (or gain) in each latitude zone. The equation expressing this balance is

$$D \frac{d}{dx} (1 - X^2) \frac{d}{dx} T(X) = I(X) - QS(X)a(X) \quad \text{(A1)}$$

where D, X, and Q are defined in section (3.3), $T(X)$ is the surface temperature, $I(X)$ is the outward infrared radiation, $S(X)$ specifies the latitudinal distribution of incoming solar radiation, and $a(X)$ is the coalbedo. For $a(X)$ and $I(X)$, we used the forms suggested by *Budyko* [1969]:

$$\begin{aligned} a(X) &= 0.38 \qquad X > X_s \\ &= 0.68 \qquad X < X_s \end{aligned} \quad \text{(A2)}$$

with the ice-cap boundary X_s determined by the condition $T(X_s) = -10°\text{C}$, and

$$I(X) = A + BT(X) \quad \text{(A3)}$$

with $A = 201.4$ W m^{-2} and $B = 1.45$ W m^{-2} °C^{-1} [*North*, 1975]. Recent calibrations of the infrared parameterization suggest that the coefficient B should be increased to ~2 W m^{-2} °C^{-1} [*North et al.*, 1981]. Use of this larger value would decrease somewhat the sensitivity of our models to changes in solar radiation. The values of the diffusion coefficient discussed in 3.3 have been put in dimensionless form by dividing by B.

$S(X)$ was expanded as a series in even order Legendre polynomials:

$$S(X) = 1 + \sum_{n,\text{even}}^{N} S_n P_n(X) \quad \text{(A4)}$$

describing the seasonally averaged distribution of solar radiation. The models discussed in section 3.3 were computed using the $N = 8$ expansion of *Coakley* [1979]. Trial runs with the $N = 2$ expansion suggested by *North* [1975] gave very similar results. For given values of D and Q, equations (A1) to (A4) can be solved analytically, as described by *North* [1975].

For the present solar radiation we used

$$Q_0 = S_0/4 = 342.1 \text{ W m}^{-2} \quad \text{(A5)}$$

where S_0 is the solar constant, based on the satellite measurements of *Willson et al.* [1981]. This value is 2.3% larger than that used by *North* [1975]. As a result, the diffusion coefficient we require to fit the present climate differs from the value obtained by North.

Acknowledgments. We thank G. R. North for some very helpful suggestions regarding climate models and R. F. Cahalan and T. J. Crowley for pointing out a number of important references. This research was supported in part by funds from the Laboratory for Planetary Atmospheres/Goddard Space Flight Center/NASA and in part by National Science Foundation grant AST79-19688 at Louisiana State University.

REFERENCES

Allen, C. W., *Astrophysical Quantities*, Athlone, London, 1973.
Augustsson, T., and V. Ramanathan, A radiative-convective model study of the CO_2 climate problem, *J. Atmos. Sci.*, *34*, 448–451, 1977.

Barron, E. J., S. L. Thompson, and S. H. Schneider, An ice-free Cretaceous? Results from climate model simulations, *Science, 212,* 501–508, 1981.

Berkner, L. V., and L. C. Marshall, Limitation on oxygen concentration in a primitive planetary atmosphere, *J. Atmos. Sci., 23,* 133–143, 1966.

Budyko, M. I., The effect of solar radiation variations on the climate of the earth, *Tellus, 21,* 611–619, 1969.

Budyko, M. I., *Climatic Changes,* AGU, Washington, D. C., 1977.

Budyko, M. I., and A. B. Ronov, Chemical evolution of the atmosphere in the Phanerozoic, *Geochem. Int., 16,* 1–9, 1979.

Cahalan, R. F., and G. R. North, A stability theorem for energy-balance climate models, *J. Atmos. Sci., 36,* 1178–1188, 1979.

Cloud, P., Beginnings of biospheric evolution and their biogeochemical consequences, *Paleobiology, 2,* 351–387, 1976.

Coakley, J. A., Jr., A study of climate sensitivity using a simple energy balance model, *J. Atmos. Sci., 36,* 260–269, 1979.

Ellis, J. S., T. H. Vonder Haar, S. Levitus, and A. H. Oort, The annual variation in the global heat balance of the earth, *J. Geophys. Res., 83,* 1959–1962, 1978.

Embleton, B. J. J., and P. W. Schmidt, Recognition of common Precambrian polar wandering reveals a conflict with plate tectonics, *Nature, 282,* 705–707, 1979.

Endal, A. S., Evolutionary variations of solar luminosity, in *Variations of the Solar Constant,* edited by S. Sofia, pp. 175–183, NASA, Greenbelt, Md., 1981.

Endal, A. S., and S. Sofia, Rotation in solar-type stars, I, Evolutionary models for the spin-down of the sun, *Astrophys. J., 243,* 625–640, 1981.

Frakes, L. A., *Climates Throughout Geologic Time,* Elsevier, New York, 1979.

Garrels, R. M., and E. A. Perry, Cycling of carbon, sulfur, and oxygen through geologic time, in the *The Sea,* edited by E. Goldberg, vol. 5, pp. 303–336, John Wiley, New York, 1974.

Goodwin, A. M., Precambrian perspectives, *Science, 213,* 55–61, 1981.

Hanks, T. C., and D. L. Anderson, The early thermal history of the earth, *Phys. Earth Planet. Int., 2,* 19–29, 1969.

Hargraves, R. B., Precambrian geologic history, *Science, 193,* 363–371, 1976.

Hart, M. H., The evolution of the atmosphere of the earth, *Icarus, 33,* 23–39, 1973.

Henderson-Sellers, A., and A. W. Schwartz, Chemical evolution and ammonia in the early earth's atmosphere, *Nature, 287,* 526–528, 1980.

Iben, I., Jr., Stellar evolution within and off the main sequence, *Ann. Rev. Astron. Astrophys., 5,* 571–626, 1967.

Kasahara, A., and W. M. Washington, General circulation experiments with a six-layer NCAR model, including orography, cloudiness and surface temperature calculations, *J. Atmos. Sci., 28,* 657–701, 1971.

Kasting, J. F., Stability of ammonia in the primitive terrestial atmosphere, *J. Geophys. Res., 87,* 3091–3098, 1982.

Knauth, L. P., and S. Epstein, Hydrogen and oxygen isotope ratios in nodular and bedded cherts, *Geochim. Cosmochim. Acta, 40,* 1095–1108, 1976.

Kuhn, W. R., and S. K. Atreya, Ammonia photolysis and the greenhouse effect in the primordial atmosphere of the earth, *Icarus, 37,* 207–213, 1979.

Luyendyk, B. P., D. Forsyth, and J. D. Phillips, Experimental approach to the paleocirculation of the oceanic surface waters, *Geol. Soc. Am. Bull., 83,* 2649–2664, 1972.

Manabe, S., and T. B. Terpstra, The effects of mountains on the general circulation of the atmosphere as identified by numerical experiments, *J. Atmos. Sci., 31,* 3–42, 1974.

Manabe, S., and R. T. Wetherald, Thermal equilibrium of the atmosphere with a given distribution of relative humidity, *J. Atmos. Sci., 24,* 241–259, 1967.

McKenzie, D., and Weiss, N., Speculations on the thermal and tectonic history of the earth, *Geophys. J. R. Astron. Soc., 42,* 131–174, 1975.

Moorbath, S., R. K. O'Nions, and R. J. Pankhurst, Early Archaean age for the Isua iron formation, Greenland, *Nature, 245,* 138–139, 1973.

Newman, M. J., and R. T. Rood, Implications of solar evolution for the earth's early atmosphere, *Science, 198,* 1035–1037, 1977.

North, G. R., Analytical solution to a simple climate model with diffusive heat transport, *J. Atmos. Sci., 32,* 1301–1307, 1975.

North, G. R., R. F. Cahalan, and J. A. Coakley, Jr., Energy balance climate models, *Rev. Geophys. Space Phys., 19,* 91–121, 1981.

Owen, T., R. D. Cess, and V. Ramanathan, Enhanced CO_2 greenhouse to compensate for reduced solar luminosity on early earth, *Nature, 277,* 640–642, 1979.

Pollack, J. B., Climatic change on the terrestial planets, *Icarus, 37,* 479–553, 1979.

Rasool, S. I., and C. De Bergh, The runaway greenhouse and the accumulation of CO_2 in the Venus atmosphere, *Nature, 226,* 1037–1039, 1970.

Ross, J. E., and L. H. Aller, The chemical composition of the sun, *Science, 191,* 1223–1229, 1976.

Sagan, C., and G. Mullen, Earth and Mars: Evolution of atmospheres and surface temperatures, *Science, 177,* 52–56, 1972.

Schopf, T. J. M., *Paleoceanography,* Harvard University Press, Cambridge, 1980.

Schwarzschild, M., *Structure and Evolution of the Stars,* Princeton University Press, Princeton, New Jersey, 1958.

Sclater, J. G., C. Jaupart, and D. Galson, The heat flow through oceanic and continental crust and the heat loss of the earth, *Rev. Geophys. Space Phys., 18,* 269–311, 1980.

Sutton, J., and J. V. Watson, Tectonic evolution of continents in early Proterozoic times, *Nature, 247,* 433–435, 1974.

Tarling, D. H., The geological-geophysical framework of ice ages, in *Climatic Change,* edited by J. Gribben, pp. 3–24, Cambridge University Press, New York, 1978*a*.

Tarling, D. H., Plate tectonics: present and past, in *Evolution of the Earth's Crust,* edited by D. H. Tarling, pp. 361–408, Academic, New York, 1978*b*.

Vonder Haar, T. H., and A. H. Oort, New estimate of annual poleward energy transport by northern hemisphere oceans, *J. Phys. Oceanogr., 3,* 169–172, 1973.

Walker, J. C. G., *Evolution of the Atmosphere,* Macmillan, New York, 1977.

Walker, J. C. G., P. B. Hays, and J. F. Kasting, A negative feedback mechanism for the long-term stabilization of earth's surface temperature, *J. Geophys. Res., 86,* 9776–9782, 1981.

Warren, S. G., and S. H. Schneider, Seasonal simulation as a test for uncertainties in the parameterizations of a Budyko-Sellers zonal climate model, *J. Atmos. Sci., 36,* 1377–1391, 1979.

Wasserburg, G. J., G. J. F. MacDonald, F. Hoyle, and W. A. Fowler, Relative contributions of uranium, thorium and potassium to heat production in the earth, *Science, 143,* 465–467, 1964.

Wetherald, R. T., and S. Manabe, The effects of changing the solar constant on the climate of a general circulation model, *J. Atmos. Sci., 32,* 2044–2059, 1975.

White, D. E., and Waring, G. A., Volcanic emanations, *U.S. Geol. Surv. Prof. Pap. 440-K,* 1963.

Wigley, T. M. L., Climate and paleoclimate: what can we learn about solar luminosity variations, *Solar Phys., 74,* 435–471, 1981.

Wigley, T. M. L., and P. Brimblecombe, Carbon dioxide, ammonia, and the origin of life, *Nature, 291,* 213–215, 1981.

Willson, R. C., S. Gulkis, M. Janssen, H. S. Hudson, and G. A. Chapman, Observations of solar irradiance variations, *Science, 211,* 700–702, 1981.

(Received October 8, 1981;
revised June 11, 1982;
accepted June 18, 1982.)