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# An Ice-Free Cretaceous? Results from Climate Model Simulations

Eric J. Barron, Starley L. Thompson, Stephen H. Schneider

*Summary.* We have quantitatively investigated the mechanisms that could explain the warm, equable climate that is believed to have been typical of the mid-Cretaceous (100 million years ago). By performing simulations with a climate model based on zonal energy balance, we demonstrate that past changes in geography were important in bringing about climatic change. However, the meridional distribution of Cretaceous temperatures cannot be successfully simulated unless additional physical "feedback mechanisms" are included in the model. These mechanisms may involve cloud and meridional heat transport changes. We also conclude that paleoclimatologists should reexamine carefully both existing data and their interpretations with regard to reconstruction of Cretaceous tropical and polar surface temperatures.

The contrast between the climate of the Cretaceous period (65 million to 140 million years ago) and that of the present epoch is the largest in the history of the earth that has been fairly well documented. A fundamental problem in paleoclimatology is how a globally ice-free climate could be maintained. We discuss here several mechanisms that could explain such a climate and examine these mechanisms quantitatively by performing simulations with a climate model based on zonal energy balance.

This investigation has a dual focus. Because the geography of the earth 100 million years ago can be well specified (*1*), we can use the climate model to examine the importance of changes in paleogeography as a mechanism of climatic change on long time scales. In addition, we can use the paleoclimatic data as observations against which to compare model simulations for a number of assumed physical processes.

Regardless of how well the model simulates the paleoclimatic data, a conclusive explanation of Cretaceous climate cannot be given for two reasons:

1) Our results are constrained by the limited capabilities of present models whose simulation ability can largely be verified only against observations of present-day climate.

2) There are insufficient detailed paleoclimatic data to determine the relative importance of a number of physical mechanisms.

In other words, our ability to verify the model's Cretaceous climate simulation is limited by the assumptions that must be made in any quantitative reconstruction of the climate 100 million years ago. This investigation can, however, help to identify those processes that might deserve further refined investigations; it might also suggest what improvements in accuracy or additions to proxy paleoclimatic data might be needed for future studies.

The method of investigation we use is to vary model parameters (such as land-sea distribution, cloud amount or height) one at a time. Thus, our goal is not a reliable comprehensive simulation but rather a series of sensitivity experiments in which cause-and-effect relationships can be quantified.

## Paleoclimatic Evidence and Interpretation

The Cretaceous climate has been classically described as warm and equable on the basis of "climate-sensitive" sedimentary indicators, characteristics of fossil floras and faunas, and oxygen isotope data. Floras and faunas presently indicative of warm climates extended into polar latitudes, and the "tropical-temperate" boundary was displaced significantly poleward (*2, 3*). The Cretaceous was the acme of exothermic reptiles (*4*). As yet, unequivocal evidence of permanent ice is unknown. On the basis of plankton assemblages, the early Arctic Ocean was not ice-covered at the end of the Cretaceous (*5*), and the first evidence of glacial erosion on Antarctica was approximately 50 million years ago (*6*). Although limited in extent, oxygen isotope data suggest equatorial temperatures were near 30°C and bottom water temperatures were approximately 15°C. Mid-latitude temperatures were apparently distinctively warmer [see summaries (*7, 8*)].

Translation of all the Cretaceous paleoclimatic data into an estimate of annual mean surface air temperature with respect to latitude is not straightforward. However, without permanent polar ice (or even if there were small ice sheets) and with equatorial temperatures apparently similar to (or possibly cooler than) those of the present day, the meridional surface temperature gradient must have been considerably reduced. On the basis of several different aspects of the paleoclimatic data, Cretaceous polar temperatures have been estimated by various authors to be between 5° and 19°C (*9*). The measured equatorial isotopic temperature is about 30°C and, if modern shallow water organisms are indicative, faunal temperature tolerances were not exceeded (*10*).

Because the paleoclimatic data are not globally homogeneous and because

E. J. Barron is a postdoctoral fellow, S. L. Thompson is a visiting graduate student from the Department of Atmospheric Sciences of the University of Washington, and S. H. Schneider is a deputy director of and a scientist in the Advanced Study Program at the National Center for Atmospheric Research, Boulder, Colorado 80307.



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many of the inferred temperatures are based on proxy indicators, we have chosen to bracket the paleoclimatic data with "warmest" and "nominal" estimates of temperatures with respect to latitude. In the "nominal" estimate we assume that the mean annual temperature at the poles must have been near or above freezing because there is no reliable evidence of significant permanent ice. However, we cannot preclude seasonal subfreezing temperatures as a possibility, particularly in the mid-Antarctic continent. In addition, we assume that equatorial temperatures were very similar to those of today. The estimated mean annual temperature profile (Fig. 1, curve A) assumes a smooth, symmetrical, equator-to-pole transition. The "warmest" estimate of mid-Cretaceous temperatures (Fig. 1, curve D) assumes a mean equatorial temperature of 304 K and a mean polar temperature of 288 K. Other intermediate temperature profiles (Fig. 1, curves B and C) used in later calculations, are also given.

Although curve A (Fig. 1) can be considered as a nominally ice-free Cretaceous climate and is considerably cooler than the warmest estimates based on paleoclimatic data, the absence of any glacial activity should not yet be accepted as fact. The Cretaceous record on Antarctica is not extensive and the Antarctic polar ice cap prevents detailed examination of the geologic history. An alternative interpretation of the isotopic data (*11*) could allow for ice accumulation earlier than previously considered (*6*).

## The Climate Model

The model we use for our experiments is a version of that developed by Thompson and Schneider (*12*), which is based on a zonally averaged energy balance. The energy balance of the vertically integrated, zonally averaged, earth-atmosphere system is described by:

$$\frac{\partial}{\partial t}[R(\phi)T_s(\phi,t)] = Q(\phi,t)[1 - \alpha(\phi,t)] - F_{IR}{}^{\uparrow}(\phi,t) - \text{div}\, F(\phi,t) \qquad (1)$$

The variables are listed in Table 1. The formulation of the energy balance components in Eq. 1 constitutes the bulk of the problem of low-resolution climate modeling (*13*). The formulation for planetary albedo is taken directly from Thompson and Barron (*7*). The outgoing infrared radiation is assumed to be a linear function of $T_s$, with a correction for cloud cover (*14*).

The energy transported poleward by the atmosphere may be divided into two components: sensible plus potential energy and latent energy. Both components are approximated as diffusion processes. Sensible plus potential energy transport is proportional to the meridional gradient of $T_s$, and latent heat transport is proportional to the meridional gradient of atmospheric water vapor concentration (*15*). Outside the tropics the diffusion coefficients are assumed to be proportional to the absolute value of the temperature gradient, thus the diffusion is nonlinear. The transport of energy by the oceans is specified as the present annual mean values (*16*) since, to our knowledge, there are no verified means of calculating this component for today's conditions, let alone for a time of substantially altered physical geography.

The thermal inertia, $R$, is a strong determinant of the amplitude of the annual cycle of temperature at a given latitude; $R$ is a weighted average of the thermal inertias of land and ocean (*17*).

With each term specified, Eq. 1 is solved numerically for $T_s$ by integrating stepwise in time until an equilibrium annual climatic cycle is achieved. For the present-day the simulation is accurate to within a few percent of the annual amplitude of temperatures.

Table 1. Definitions of all variables.

| Symbol | Definition |
|---|---|
| Equation 1 | |
| $\phi$ | Latitude |
| $t$ | Time |
| $R$ | Thermal inertia (heat capacity per unit area) |
| $T_s$ | Surface air temperature |
| $Q$ | Incoming solar radiation |
| $\alpha$ | Planetary albedo (earth-atmosphere reflectivity) |
| $F_{IR}{}^{\uparrow}$ | Outgoing infrared radiation to space |
| $F$ | Northward energy transport by atmosphere and oceans |
| Parameterization of $F_{IR}{}^{\uparrow}$ (*14*) and zonally averaged, annually averaged parameters specified in the model | |
| $f_0$ | Ocean fraction |
| $f_{si}$ | Sea ice fraction |
| $f_c$ | Cloud fraction |
| $T_s - T_c$ | Surface temperature minus cloud top temperature |
| $\Delta_p T$ | Present-day $T_s - T_c$ distribution (*26*) |
| $\Delta_c T$ | Hypothetical Cretaceous $T_s - T_c$ distribution (*27*) |
| A, B, C | $-248.1$ W/m$^2$, 1.8 Wm$^{-2}$ K$^{-1}$ and $-1.73$ Wm$^{-2}$ K$^{-1}$ (*14*) |
| ML | Mixed layer depth (*17, 22*) |
| "Transport-adjusted" simulations | |
| $\Delta Q(\phi)$ | Net incremental energy input at each latitude |
| $\overline{\Delta Q}$ | Globally averaged $\Delta Q(\phi)$ |
| $\Delta Q_t(\phi)$ | Component of $\Delta Q(\phi)$ resulting solely from changes in the meridional transport of energy |

## Cretaceous Simulations: Effects of Prescribed Surface Changes

Numerous authors have presented evidence that long time-scale variations in climate (on the order of tens of millions of years) are a function of past continental positions or tectonically related (climate-independent) eustatic sea-level fluctuations (*18, 19*).

Using zonally averaged land distributions derived from paleogeographic reconstructions (*1*), Barron *et al.* (*20*) demonstrated that paleogeography modifies the planetary surface albedo in such a way as to result in a global cooling during the last 65 million years. In a more rigorous model of the planetary albedo for the mid-Cretaceous, Thompson and Barron (*7*) calculated a 2.3 to 3.9 percent increase in total absorbed solar radiation during the Cretaceous compared to the present, resulting from the combination of geographic changes and assumptions of warm polar temperatures (and hence no snow or ice cover) and Cretaceous cloud cover (*21*). This increase in absorbed solar energy alone could explain a large fraction of the global temperature difference between the Cretaceous and the present. Approximately 40 percent of the increase results from the geographic changes.

An important question is whether the increased solar heating due to geographic changes is sufficient, when combined with internal climatic feedback mechanisms, to account for a permanent ice-free climate. Because these planetary albedo calculations are only one component of the planetary energy balance in Eq. 1, the more comprehensive energy balance climate model is used to determine if the change in absorbed solar radiation resulting from paleogeography might be of sufficient magnitude to produce and maintain warm, equable temperatures.

Specified variables for the present-day and Cretaceous are given in Table 2. The fraction of the ocean covered by ice during the Cretaceous is assumed to be 0.0 because there is no evidence for the presence of any significant sea ice at that time. Since the oceanic heat transports at times of different ocean basin configuration are unknown, present values of heat transport by the oceans are specified. The effects of this assumption will be examined later. In the absence of information on mixed-layer depths during the Cretaceous, we specify a time independent value of 100 meters at all latitudes (*22*). These specifications can be used to define a "base" Cretaceous experiment that can be compared with

the present-day control simulations and with a series of other sensitivity experiments with Cretaceous geography.

Compared to the control experiment, the "base" Cretaceous simulation shows an increase in total absorbed solar energy of 1.2 percent (from 239.0 to 241.9 watts per square meter) and a global temperature increase of 1.62 K. The globally averaged temperature given by our "nominal" temperature profile (Fig. 1) is 6°C larger than the present. The "base" experiment (Fig. 2A) increase in temperature is largest at high latitudes, particularly in the Northern Hemisphere where the temperature at 80°N increases by 5°C. However, the polar temperatures remain well below 273 K.

A second simulation (dashed line, Fig. 2A) is performed by specifying a land surface albedo characteristic of lush tropical vegetation or a "wet" surface albedo [Table 2; see also (7)]. This decreases the land albedo equatorward of 45° relative to present values. The total absorbed solar energy increases by 1.6 percent (to 242.9 $W/m^2$) and the global temperature increases 2.18°C over the present-day control experiment. The temperature increase from the "wet" land albedo case over the "base" Cretaceous experiment is about 1°C and, because of the meridional transport formulation, is uniformly distributed with respect to latitude—even though the change in land surface albedo is not at all uniformly distributed. In all the remaining experiments we use the "wet" surface albedos.

The results in Fig. 2A clearly show that these simulations do not reproduce the temperature distribution of the "nominal" Cretaceous (solid line, Fig. 2A). This discrepancy between the model calculations and the nominal Cretaceous temperature profile may be the result of three factors: (i) the structure of the model or physical processes incorporated in the model may be incomplete or incorrect; (ii) the paleoclimatic data may have been misinterpreted; or (iii) other external mechanisms of climatic change, in addition to paleogeography, may have been operating. As for (ii), simple extrapolation of the model results in Fig. 2A indicates that if polar temperatures are to be above 273 K (the "nominal" temperature estimate) then the simulated equatorial temperatures would be excessively high. However, either very high equatorial temperatures or polar temperatures well below 273 K do not appear to be a reasonable interpretation of the paleoclimatic data. As for (iii), plate tectonics is the only quantitatively well-documented mechanism that is known to operate on the appropriate $10^6$- to $10^7$-year time scale. Other forcing mechanisms are possible, but their potential influence would be more speculative. For these reasons we assume, for the time being, that the discrepancy in temperature is due to incomplete or incorrect replication of major physical processes in the model.

## Effects of Prescribed Cloud Changes

In the preceding section we performed experiments by changing boundary conditions for which some geologic evidence exists. Now we examine the effects of some prescribed changes to internal conditions for which there exist no direct evidence of change, but which may nev-

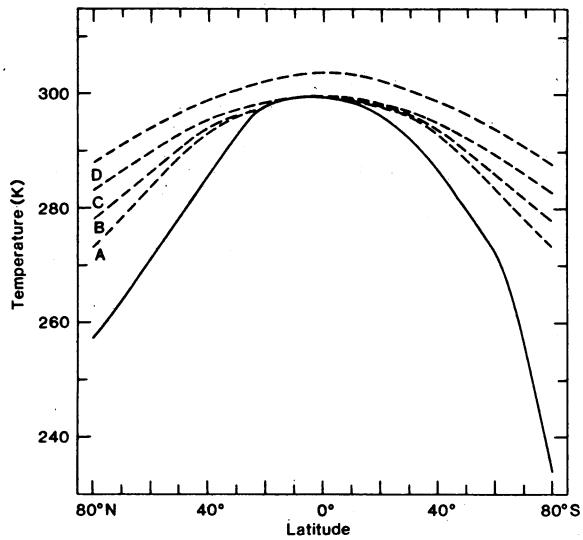


Fig. 1. The present-day (solid line) and hypothetical Cretaceous (dashed lines) distributions of annual mean surface air temperature with respect to latitude. (Curve A) The nominal distribution of temperatures based on a mean annual polar temperature of 273 K and equatorial temperatures similar to those of the present day. (Curve D) The "warmest" estimate of Cretaceous temperatures based on paleoclimatic data. (Curves B and C) Intermediate hypothetical distributions.

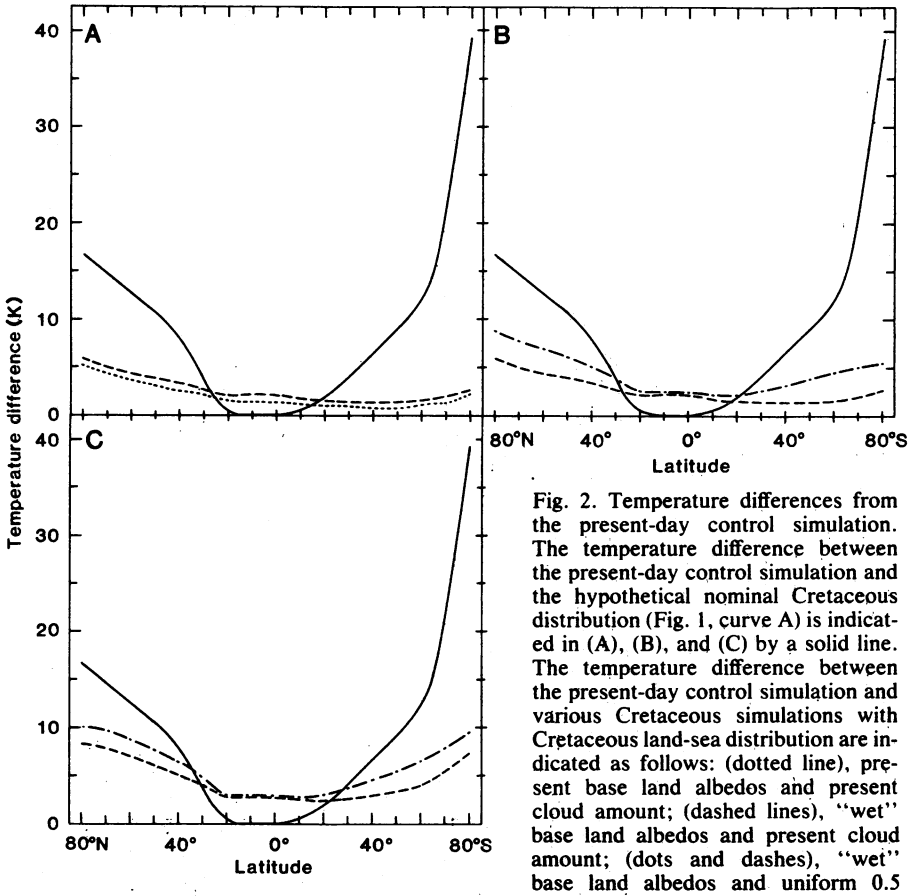


Fig. 2. Temperature differences from the present-day control simulation. The temperature difference between the present-day control simulation and the hypothetical nominal Cretaceous distribution (Fig. 1, curve A) is indicated in (A), (B), and (C) by a solid line. The temperature difference between the present-day control simulation and various Cretaceous simulations with Cretaceous land-sea distribution are indicated as follows: (dotted line), present base land albedos and present cloud amount; (dashed lines), "wet" base land albedos and present cloud amount; (dots and dashes), "wet" base land albedos and uniform 0.5 cloud amount. (A) Comparison of land albedo assumptions. (B) Comparison of present and uniform 0.5 cloud amounts. (C) Comparison of present and uniform 0.5 cloud amounts with a hypothetical Cretaceous distribution of $T_s - T_c$ (surface temperature minus cloud top temperature) given in Table 2. The closer the dotted or dashed curves are to the solid curve, the "better" the simulation.

ertheless be plausible. We concentrate on the internal variable cloudiness.

Clouds have two competing effects on surface temperature (*23*). First, they reflect solar radiation, thus cooling the surface. Second, clouds act to decrease $F_{IR}\uparrow$ relative to a cloudless sky since their tops usually radiate at a lower temperature than the surface. To maintain energy balance, $T_s$ must increase to compensate for the decrease in $F_{IR}\uparrow$ due to clouds. Our model gives a global mean effect of changes in cloud amount on albedo, $\Delta\alpha$, that is larger than the effect on $F_{IR}\uparrow$. Observations suggest that this is the case (*24*). Thus, uniformly decreasing the amount of global cloud cover in our model tends to increase global surface temperature.

Table 2 shows that the present-day fractional cloud cover, $f_c$, increases from the subtropics poleward. This is largely due to the storminess from atmospheric circulation systems associated with the relatively large surface temperature gradient at mid-to-high latitudes. Since the temperatures and continental distribution during the Cretaceous were apparently much more uniform in latitude than they are now, it is plausible that $f_c$ was also more uniform. In addition, some recent studies (*25*) have indicated that global cloud cover tends to decrease, at least in mid-latitudes, as temperature increases. Although this is not proof that there was less cloud cover during the Cretaceous, we can at least decrease $f_c$ in order to test the sensitivity of the model to such a plausible change. The present-day globally averaged cloud cover from Table 2 is 0.55. By setting $f_c$ to 0.50 at all latitudes, we both decrease the global average of $f_c$ slightly and simultaneously make its distribution uniform with latitude.

Figure 2B shows the result of changing from the present-day $f_c$ to a uniform 0.5 $f_c$. Note that the mid-to-high latitudes warm by about 2°C while the tropics show very little change. This distribution of change occurs because the present-day cloud cover is already about 0.5 or less in the tropics. The increased extratropical warming improves the agreement with the nominal Cretaceous temperature distribution, although the high latitudes are still much too cold, and the tropics too warm.

The outgoing infrared radiation to space depends not only on cloud amount, but on the temperature difference $T_s - T_c$ between the cloud tops and the surface (*14*). (For example, if $T_c$ were the same as the surface temperature, the presence of clouds would have minimal effect on $F_{IR}\uparrow$.) Table 2 gives estimates of the mean annual effective $T_s - T_c$, referred to as $\Delta_p T$, for the present day (*26*). (These values have been used in all the previously discussed experiments.)

Since Cretaceous temperatures were more globally uniform than at present, one can speculate that the Cretaceous distribution of $T_s - T_c$ was also more uniform than at present. We have therefore performed experiments using a hypothetical Cretaceous distribution of $T_s - T_c$ (referred to as $\Delta_c T$) shown in Table 2 (*27*). Note that the differences between $\Delta_c T$ and $\Delta_p T$ are the largest in extratropical latitudes. Results of experiments using uniform and present-day cloud cover distributions are shown in Fig. 2C, which is the same as Fig. 2B except we replace $\Delta_p T$ by $\Delta_c T$. The change in $T_s - T_c$ leads to an increase in surface temperature of 1° to 4°C at extratropical latitudes compared to the results in Fig. 2B.

From Fig. 2 we see that there still remains a large difference between our experiments and the nominal Cretaceous temperature distribution. Yet, we can say the prescribed cloud changes, if they occurred, would have contributed significantly to the creation of a Cretaceous-like temperature distribution.

Another principal internal mechanism

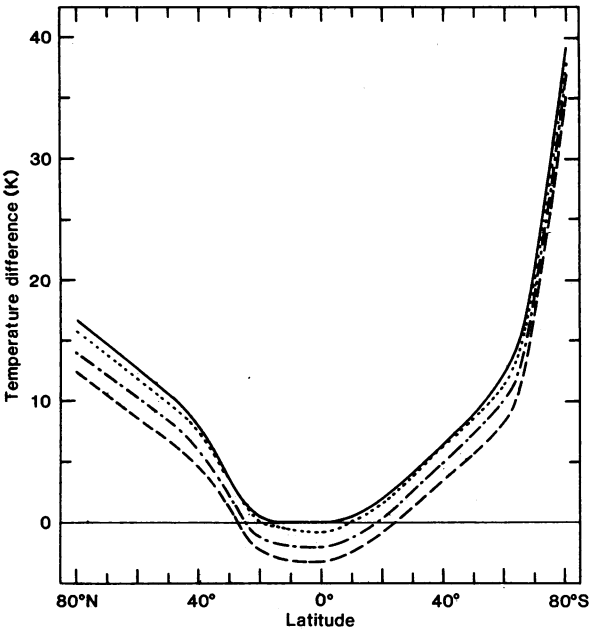


Fig. 3. Temperature differences from the present-day control experiment for three transport-adjusted Cretaceous simulations. The temperature difference between the present-day experiment and the nominal Cretaceous temperature distribution is indicated as a solid line (same as in Fig. 2). Temperature differences between the present-day experiment and the Cretaceous simulations (Cretaceous land-sea distribution and "wet" base land albedos) are indicated as follows: (dashed line), present cloud cover and present $T_s - T_c$; (dots and dashes), uniform 0.5 cloud amount and present $T_s - T_c$; and (dotted line), uniform 0.5 cloud amount and a hypothetical Cretaceous $T_s - T_c$ given in Table 2.

Table 2. Present-day and Cretaceous averaged values for a fraction of a zone covered by ocean, a fraction of ocean covered by sea ice (annual mean), and present-day fraction of a zone covered by clouds (annual mean) (*7*).

| Latitude | Ocean fraction ($f_0$) | | Present sea ice fraction ($f_{si}$) | Present cloud fraction ($f_c$) | Cretaceous base land albedo | $T_s - T_c$ | |
|---|---|---|---|---|---|---|---|
| | Present | Cretaceous | | | | Present ($\Delta_p T$) | Cretaceous ($\Delta_c T$) |
| 85°–75°N | 0.81 | 0.67 | 0.90 | 0.64 | 0.16 | 14.0 | 19.8 |
| 75°–65°N | 0.45 | 0.58 | 0.50 | 0.66 | 0.16 | 16.4 | 21.6 |
| 65°–55°N | 0.38 | 0.55 | 0.14 | 0.69 | 0.16 | 19.5 | 24.0 |
| 55°–45°N | 0.42 | 0.55 | 0.00 | 0.67 | 0.16 | 21.6 | 25.8 |
| 45°–35°N | 0.54 | 0.62 | 0.00 | 0.60 | 0.12 | 24.6 | 28.0 |
| 35°–25°N | 0.59 | 0.79 | 0.00 | 0.52 | 0.12 | 27.4 | 28.8 |
| 25°–15°N | 0.68 | 0.90 | 0.00 | 0.41 | 0.12 | 29.4 | 29.4 |
| 15°–5°N | 0.76 | 0.89 | 0.00 | 0.47 | 0.12 | 30.0 | 30.0 |
| 5°N–5°S | 0.77 | 0.81 | 0.00 | 0.49 | 0.12 | 30.0 | 30.0 |
| 5°–15°S | 0.78 | 0.78 | 0.00 | 0.48 | 0.12 | 29.6 | 30.0 |
| 15°–25°S | 0.76 | 0.81 | 0.00 | 0.46 | 0.12 | 28.8 | 29.4 |
| 25°–35°S | 0.81 | 0.83 | 0.00 | 0.49 | 0.12 | 27.2 | 28.8 |
| 35°–45°S | 0.95 | 0.80 | 0.00 | 0.58 | 0.12 | 25.0 | 28.0 |
| 45°–55°S | 0.98 | 0.82 | 0.00 | 0.73 | 0.16 | 22.2 | 25.8 |
| 55°–65°S | 1.00 | 0.83* | 0.26 | 0.81 | 0.16 | 19.5 | 24.0 |
| 65°–75°S | 0.62 | 0.56* | 0.69 | 0.67 | 0.16 | 13.2 | 21.6 |
| 75°–85°S | 0.07 | 0.17* | 1.00 | 0.47 | 0.16 | 4.8 | 19.8 |
| Global | 0.71 | 0.76 | 0.07 | 0.55 | 0.13 | 25.3 | 27.6 |

*Assumes present-day land area for Antarctica (*1*).

is dynamical interactions between latitudes. In the absence of other external forcing mechanisms, these must have played a large role in maintaining the equable Cretaceous temperatures. We next discuss what would be required of heat transports to bring the model-simulated temperatures into better agreement with our hypothetical "nominal" distribution.

### Effects of Prescribed Meridional Heat Transport Changes

Although factors such as land albedo, fraction of ocean, or cloud parameters have been modified nonuniformly with latitude in the Cretaceous experiments, the diffusive heat transport formulation tends to smooth out temperature differences between latitudes. The diffusive meridional heat transport thus acts as a strong negative feedback mechanism for regional temperature changes. Although the latent heat transport increases in the Cretaceous simulations because it is proportional to the increase in water vapor concentration in the atmosphere (which is itself a function of absolute temperature), latent heat transport does not fully compensate for the decrease in sensible and potential energy transport. We conclude that the distribution of temperatures, estimated from the paleoclimatic data, would be difficult to explain using the diffusive heat transport formulation now in the model [see (*15*)].

Another aspect of the problem is that oceanic heat transport has been fixed in our simulations at present-day values (*16*). On the basis of correlations of continental paleopositions and glaciations, oceanic heat transport to high latitudes has been postulated as a major factor in explaining ice-free climates (*28*). During the Cretaceous there were larger areas of ocean near the poles, although the paleogeography of Antarctica is not well known (*1*).

Rather than attempting to justify specific modifications to the heat transport formulation (itself a major area of research), we determine empirically the total heat transport required to achieve several hypothetical specified Cretaceous temperature distributions (Fig. 1). We can then consider in more detail whether the required heat transport seems reasonable in terms of known paleogeography and physical processes of oceanic and atmospheric circulations.

Starting from the Cretaceous experiments in Fig. 2, we determine the incremental net energy input (in watts per square meter) at each latitude that will result in a meridional temperature distribution closely matching a hypothetical Cretaceous distribution selected from Fig. 1. This incremental energy input is determined through an iterative procedure which ends when the model temperatures match the desired Cretaceous temperatures to within some small tolerance. We call the derived incremental energy input at each latitude $\Delta Q(\phi)$. $\Delta Q(\phi)$ may be conveniently divided into two additive quantities. First, $\Delta Q(\phi)$ may be globally averaged to give $\overline{\Delta Q}$. $\overline{\Delta Q}$ represents, to a close approximation, the change in globally averaged energy input, in excess of that already resulting from changes in geography or clouds, required to reach the global average of the desired temperature distribution. Second, $\overline{\Delta Q}$ may be subtracted from $\Delta Q(\phi)$ to give the component of $\Delta Q(\phi)$ that arises solely from changes in the meridional transport of energy, $\Delta Q_t(\phi)$. $\overline{\Delta Q}$ cannot be due to a meridional transport of energy since the meridional transport term is required to be zero when integrated over a closed surface, such as the earth. The importance of $\overline{\Delta Q}$ is that it is a quantitative estimate of the additional global heating required to achieve a specific Cretaceous global temperature estimate, and it may be interpreted as being a consequence of several possible factors: (i) a change in net energy input (from changes in cloudiness, carbon dioxide, or solar constant, for example) in addition to those already considered, (ii) an incorrect model temperature sensitivity to a change in energy input, thus creating the need for a spurious energy input to reach the desired global temperature, or (iii) inaccurately specified Cretaceous temperatures, to the extent that a spurious energy input is required to achieve them, even though the model sensitivity is accurate.

We may add both $\overline{\Delta Q}$ and $\Delta Q_t(\phi)$ to the model simulations, in which case the desired temperature distribution is almost exactly reproduced. Alternatively, we may add only $\Delta Q_t(\phi)$ to the model, in which case the model temperatures will be nearly uniformly above or below the desired distribution by an amount that is

Table 3. A comparison of total and atmospheric heat transport convergences (watts per square meter) for a present-day control simulation with Cretaceous simulations with indicated cloud parameters, where $f_c$ is the cloud fraction and $\Delta_p T$ and $\Delta_c T$ are, respectively, the present-day and Cretaceous temperature differences between the cloud top and the surface. Columns A are the model experiments with the oceanic heat transport fixed at present-day values, and columns B are the transport-adjusted experiments including the latitude distribution of incremental energy input $\Delta Q_t(\phi)$ required to minimize the difference between the model-generated temperatures and the nominal Cretaceous temperature distribution (see text).

| Latitude | Control | | Present $f_c$, $\Delta_p T$ | | | | $0.5 f_c$, $\Delta_c T$ | | | |
|---|---|---|---|---|---|---|---|---|---|---|
| | | | A | | B | | A | | B | |
| | Total | Atmospheric | Total | Atmospheric | Total | Atmospheric | Total | Atmospheric | Total | Atmospheric |
| 80°N | 137.26 | 126.56 | 103.15 | 92.35 | 116.90 | 69.26 | 101.31 | 90.61 | 107.95 | 71.75 |
| 70°N | 67.12 | 69.37 | 74.74 | 70.99 | 86.77 | 53.72 | 71.36 | 67.61 | 77.52 | 55.68 |
| 60°N | 52.93 | 46.18 | 51.31 | 44.56 | 60.90 | 26.96 | 46.36 | 39.61 | 50.31 | 27.08 |
| 50°N | 46.01 | 11.02 | 47.68 | 12.68 | 59.40 | 0.91 | 41.10 | 6.10 | 45.81 | −1.03 |
| 40°N | 6.19 | −17.31 | 7.61 | −15.88 | 14.40 | −29.77 | 4.71 | −18.78 | 7.13 | −30.49 |
| 30°N | −8.60 | −21.58 | −8.18 | −21.17 | −4.09 | −9.96 | −6.77 | −19.76 | −7.98 | −10.14 |
| 20°N | −29.14 | −18.37 | −30.56 | −19.78 | −32.16 | −1.91 | −23.21 | −12.44 | −29.19 | 0.87 |
| 10°N | −56.26 | −17.42 | −55.76 | −16.93 | −60.89 | −8.69 | −52.52 | −13.69 | −51.26 | −9.30 |
| 0 | −68.44 | −25.70 | −66.54 | −23.80 | −69.47 | −9.35 | −63.94 | −21.20 | −65.91 | −10.79 |
| 10°S | −27.51 | −5.81 | −27.30 | −5.60 | −29.16 | −1.23 | −24.85 | −3.15 | −24.28 | −2.12 |
| 20°S | −28.63 | −29.83 | −28.17 | −29.37 | −25.81 | −14.67 | −23.40 | −24.60 | −30.93 | −13.36 |
| 30°S | −16.06 | −17.36 | −14.11 | −15.41 | −7.98 | −4.76 | −10.30 | −11.60 | −21.18 | −2.70 |
| 40°S | 4.55 | −12.19 | 4.43 | −12.31 | 12.98 | −20.48 | 3.54 | −13.19 | 6.40 | −19.77 |
| 50°S | 36.72 | 10.83 | 42.83 | 16.93 | 55.40 | 15.98 | 34.53 | 8.63 | 41.16 | 13.17 |
| 60°S | 70.90 | 46.15 | 69.49 | 44.75 | 85.60 | 39.00 | 59.96 | 35.22 | 69.10 | 36.68 |
| 70°S | 92.78 | 63.84 | 87.14 | 58.20 | 112.72 | 23.73 | 81.75 | 52.81 | 99.10 | 25.18 |
| 80°S | 111.23 | 111.23 | 99.82 | 99.82 | 126.80 | 10.60 | 94.42 | 94.42 | 114.34 | 10.38 |

proportional to the magnitude of $\overline{\Delta Q}$. In either case we will refer to the resulting experiments as transport-adjusted, since $\Delta Q_t(\phi)$ can be interpreted as the result of an adjustment in net meridional energy transports.

Transport-adjusted experiments have been performed for each of the cloud parameter sensitivity experiments in Fig. 2 with each of the temperature distributions from Fig. 1. First, we examine experiments using the "nominal" Cretaceous temperatures.

The temperature differences from the present-day control experiment are plotted in Fig. 3 for three transport-adjusted simulations using $\Delta Q_t(\phi)$ only. Shown are experiments in which we used (i) present $f_c$ and $T_s - T_c = \Delta_p T$, (ii) $f_c = 0.5$ and $T_s - T_c = \Delta_p T$, and (iii) $f_c = 0.5$ and $T_s - T_c = \Delta_c T$. Also plotted (solid lines) in Fig. 3 is the temperature difference required to achieve the nominal Cretaceous temperatures. The values of $\overline{\Delta Q}$, which were not included in the experiments shown in Fig. 3 are, respectively, 5.2, 3.0, and 1.0 W/m$^2$. In Table 3, both the total and atmospheric-only meridional heat transport convergences for the present-day control simulation are compared with those for the Cretaceous experiments (i) and (iii) above. The table includes results for both the unadjusted model experiments (columns A) and for transport-adjusted experiments (columns B).

Table 3 shows that the atmospheric heat transport convergences for the unadjusted Cretaceous model experiments usually decrease slightly from the present-day control case. This can be explained by noting the small decreases in meridional temperature gradient for these cases, as illustrated in Fig. 2, A and C. However, if the much reduced "nominal" Cretaceous meridional temperature gradient is forced in a transport-adjusted experiment, the atmospheric heat transport convergences decrease dramatically in magnitude. In columns B of Table 3, the large differences between the estimated required total heat transport convergence and the internally calculated (from the diffusive heat transport formulation) atmospheric heat transport convergence illustrate the extent to which nondiffusive heat transport must play a role in explaining the distribution of Cretaceous temperatures. Of course, such a transport could reside in the oceans as well as the atmosphere.

Another interesting point is that the total heat transport convergence in the transport-adjusted experiment in which $f_c = 0.5$ and $T_s - T_c = \Delta_c T$ (which has $\overline{\Delta Q}$ of only 1.0 W/m$^2$) is very similar to the present-day total control heat transport convergence (Fig. 4). This empirically derived result suggests that if the present-day values of heat transport can somehow be maintained, despite the decrease in Cretaceous meridional temperature gradient, then the nominal Cretaceous temperatures can be achieved with an energy balance climate model.

Table 4 shows the model-derived global mean increases in absorbed solar radiation for various Cretaceous simulations (with specified cloud parameters). This table also shows the additional change, $\overline{\Delta Q}$, required in transport-adjusted simulations to achieve each of the global mean Cretaceous temperatures from the profiles in Fig. 1. A comparison of these results gives an indication of how the global heating increase due to the presently included (non-transport–adjusted) processes compares with the remaining increases required to reach the Cretaceous temperatures. Although in some cases the additional $\overline{\Delta Q}$ is small (for example, $\Delta_c T$, 0.5 $f_c$), in other cases (for example, for the "warmest" estimate of Cretaceous temperatures) the additional $\overline{\Delta Q}$ required exceeds that generated internally in the model from the prescribed surface and cloud changes. The larger the $\overline{\Delta Q}$, the larger the unexplained influence of either internal or external mechanisms would have to be to achieve a satisfactory simulation of the hypothesized Cretaceous temperature profile.

## Discussion, Speculations, and Perspective

The results of the Cretaceous climate simulations illustrate two major problems. First, if Cretaceous temperatures were warmer than the "nominal" estimate (Fig. 1) or if cloud changes are not included in the model, then the model is incapable of simulating the hypothesized Cretaceous planetary temperature. Some additional net global heating, $\overline{\Delta Q}$ (Table 4), would be required as well as that resulting from a change in geography. Second, the total heat transport must be maintained at close to present-day values despite the fact that the meridional temperature gradient was apparently considerably reduced.

From Eq. 1, the additional net heating required to explain Cretaceous temperatures may be evaluated in terms of one of three variables: incoming solar energy, the planetary albedo, or the amount of outgoing infrared radiation. Whereas we are fairly confident that a range of possible variations in planetary albedo have been extensively examined (*7*), we note that the relation between temperature and the radiatively active atmospheric constituents, such as $CO_2$ and water vapor, is a major factor in estimating climate sensitivity. An increase in atmospheric $CO_2$ may be related to warmer oceanic temperatures and hence decreased $CO_2$ solubility. A state-of-the-art estimate of the sensitivity of climate to a $CO_2$ doubling is a 2° to 3°C surface temperature increase (*29, 30*). In speculating about this hypothesis, a major drawback is the requirement that a substantially higher atmospheric $CO_2$ concentration must be maintained over millions of years, whereas the relation between volcanic input, weathering, and deposition of sediments and the cycle of photosynthesis, respiration, and decay and storage of organic matter are likely to buffer the $CO_2$ system sufficiently to make long time-scale changes of this type speculative. For the case of warmer oceans this conclusion may depend on the ability of the biosphere to act as a buffer (*31*).

A second factor that may contribute to the additional net heating required to explain Cretaceous temperatures is solar variability. This possibility cannot be eliminated because the past behavior of the sun is not well known. However, it should be noted that most solar models indicate that the past solar input should be less than, rather than more than, the present value (*32*).

Although the value of additional net heating required to achieve the "warmest" estimate of Cretaceous temperatures (~ 12 W/m$^2$) appears excessive, on the basis of present knowledge of climate sensitivity and external forcings even this requirement may not be eliminated as impossible. Estimates of present global temperature sensitivity to a change in surface-tropospheric heating lie in the range of 0.5° to 1.0°C per watt per square meter (*33*). For the Cretaceous, with unknown additional feedbacks, the range must necessarily be more uncertain (*23*). Thus the requirement of an additional surface-tropospheric heating of 12 W/m$^2$ may be reduced to 9 W/m$^2$ or even less. A change in heating of +12 W/m$^2$ is approximately equivalent to a 5 percent increase in solar constant or a 700 percent increase in atmospheric $CO_2$, based on typical model sensitivity estimates.

Even if the increased surface heating was sufficient to explain the hypothesized Cretaceous global mean temperature, the distribution of temperature with respect to latitude remains problematic. The total poleward heat transport must be maintained at close to modern values in order to prevent excessively high tropical temperatures or cold poles despite the fact that the meridional temperature gradient was considerably reduced. This result suggests that a large nondiffusive

heat transport component must have been operating in order to explain the hypothesized Cretaceous temperatures.

There are several possible mechanisms of heat transport that may help maintain the poleward heat transport despite the decreased meridional temperature gradient. It may be expected that a warm atmosphere will transport a greater amount of latent heat than a cool atmosphere, because of the strong dependence of saturation vapor pressure over water on temperature. This idea is supported by experiments with atmospheric general circulation models by Manabe and Wetherald (*29*) showing an increased poleward transport of latent heat in middle to high latitudes when atmospheric temperatures increase. Using a highly idealized ocean-continent configuration, Manabe and Wetherald (*29*) have speculated that a warmer atmosphere, due either to a $CO_2$ increase or a change in solar input, may transport sufficient latent heat to compensate for the decreased meridional temperature gradients. These workers (*29*) further speculate that this enhanced heat transport may be important during the Cretaceous period. But, with the much reduced meridional temperature gradient in Fig. 1, it is questionable whether this enhanced heat transport will be large enough to counterbalance the hypothesized decreases in internal and potential energy transports.

A possible factor in maintaining atmospheric heat transports despite the reduced temperature gradient is the existence of quasi-stationary waves in the atmospheric circulation associated with much of the observed eddy transport of heat (*34*). These planetary waves are induced both by orography and by regional variations in atmospheric heating (*35*). We may speculate that the radically different geography 100 million years ago led to an enhancement of the poleward heat transport by quasi-stationary waves. However, a reliable atmospheric general circulation model with realistic geography is required to test such speculation.

The amount of oceanic heat transport to high latitudes during the Cretaceous is unknown. The distribution of landmasses and the latitudinal extent of coastlines, forming western ocean boundaries such as that of Australia and Antarctica, may be important for high-latitude penetration of western boundary currents. Deepwater connections to polar regions may result in high-latitude upwelling zones, particularly where polar easterlies and mid-latitude westerlies converge. It is interesting that north polar upwelling has been hypothesized from late Cretaceous plankton assemblages (*5*). Unfortunately, the extent or existence of deepwater connections is one of the more speculative aspects of paleogeographic reconstructions.

Table 4. Global heating experiments. (Case 1) Model-generated increases in absorbed solar energy (expressed as watts per square meter) for Cretaceous simulations with indicated cloud parameters with oceanic heat transport fixed. (Case 2) Required additional global heating, $\overline{\Delta Q}$, for each model simulation to achieve a specified Cretaceous distribution of temperatures (curves A to D in Fig. 1). $\Delta_p T$ and $\Delta_c T$ are, respectively, the present-day and Cretaceous temperature differences between the cloud top and the surface (see Table 2); $f_c$ is the fraction of cloud cover, either present-day (Table 2) or uniform 0.5.

| Global heating experiments | $\Delta_p T$ | | $\Delta_c T$ | |
|---|---|---|---|---|
| | Present $f_c$ | $f_c = 0.5$ | Present $f_c$ | $f_c = 0.5$ |
| Case 1 | 4.0 | 8.2 | 4.5 | 8.6 |
| Case 2 | | | | |
| Curve A ("nominal" polar 273 K) | 5.2 | 3.0 | 2.6 | 1.0 |
| Curve B (polar 278 K) | 6.2 | 4.0 | 3.6 | 2.0 |
| Curve C (polar 283 K) | 8.4 | 6.2 | 5.8 | 4.2 |
| Curve D ("warmest" estimate) | 12.9 | 10.7 | 10.3 | 8.7 |

Extensive shallow seas and restricted early ocean basins (for example, the formation of the South Atlantic) in the subtropics may prove to have been major sources of warm, highly saline deep water, which potentially could have resulted in a salt-driven abyssal circulation rather than the essentially thermally driven circulation of the present. This hypothesis, originally suggested by Chamberlin (*36*), is a distinct possibility (*37*). Kraus *et al.* (*38*) suggest that high-latitude upwelling of deep water (~ 10°C) associated with the mid-latitude production of warm, salty bottom waters could have kept polar oceans ice-free.

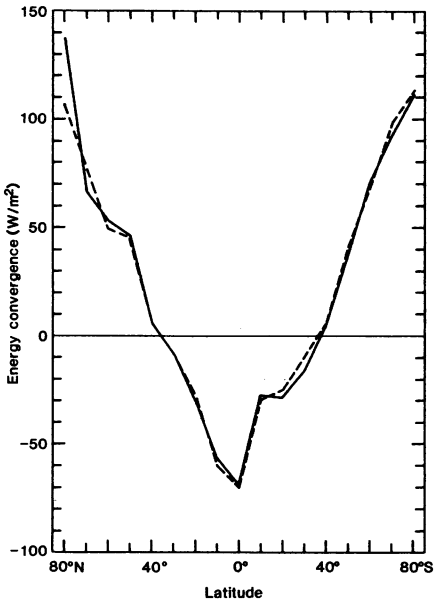


Fig. 4. A comparison of the convergence of total transported energy (in watts per square meter) for the present-day control simulation and the empirically derived convergence of total transported energy (transport-adjusted) required to match the "nominal" Cretaceous temperature distribution (Fig. 1, curve A) for the Cretaceous simulation with uniform 0.5 cloud amount and the hypothetical Cretaceous $T_s - T_c$ given in Table 2.

Whether the veracity of any of the above speculations is required to explain the Cretaceous climate depends on the interpretation of paleoclimatic data.

The oxygen isotopic record for the Cretaceous (which is classically interpreted as a temperature signal) is limited, because Foraminifera collected from deep-sea drilling are typically recrystallized or absent (*39*). The isotopic data are also subject to reinterpretation. For instance, if considerable salinity contrast existed in Cretaceous oceans, then the geographic distribution of oxygen isotope values might be a product of nonuniform evaporation and precipitation as well as temperature. It is conceivable that high-latitude waters, diluted by a freshwater component, give values of isotopic temperature that are too warm. Matthews and Poore (*11*) have noted that the assumption of no ice volume for past climates results in cooler equatorial sea surface isotopic temperatures than present. Similar equatorial temperatures (rather than cooler) may be a plausible assumption, and therefore significant ice volume could have existed during the Tertiary and possibly during the Cretaceous.

Perhaps some of the most important paleoclimatic data are the distribution of vegetation and coal deposits extending into latitudes above 70°N on continental reconstructions (*3, 7, 8*). If much of this record is coastal then it is conceivable that the distribution of "tropical" and "temperate" vegetation is the product of a more local maritime climate reinforced by local warm oceanic currents. The western margin of South Island, New Zealand, in which a coastal rain-forest vegetation is adjacent to the glaciers of the Southern Alps, is a modern example illustrating the potential for overinterpretation of a limited record.

The geologic record on Antarctica is

sufficiently obscured that conclusions concerning the absence of ice may not yet be warranted. Even if a polar ice cap was not present, mean annual temperatures below freezing in the Antarctic interior may have occurred. However, a large body of data supports a very warm, equable temperature. Even our "nominal" estimates of Cretaceous temperatures, with winter temperatures well below freezing, are already considerably cooler than "classical" estimates of Cretaceous temperatures.

We require either other external climatic forcing factors or major departures in internal factors from present, such as cloud cover, to explain even the nominal Cretaceous temperature distribution. The situation is considerably more problematic for the "warmest" Cretaceous temperatures. In particular it is difficult to further reduce the equator to pole temperature gradient with known mechanisms of plausible magnitude. From a state-of-the-art climatic modeling point of view it is much easier to explain a Cretaceous with warmer tropics and cooler poles.

In our Cretaceous simulations the changes in geography are an important external mechanism of climate change but, by themselves, these changes do not alter the net energy balance of the earth sufficiently to allow an ice-free Cretaceous climate. We require much better models to investigate and verify mechanisms that may have operated during situations very different than at present.

We also conclude that paleoclimatologists should reexamine carefully both data and inferences with regard to reconstructions of past climates. More specifically, we believe particular attention should be focused on the commonly held belief that (i) equatorial ocean surface temperatures were cooler than at present, (ii) polar temperatures were above freezing, and (iii) no significant ice volume existed on the earth in Cretaceous times.

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26. The values of $\Delta_p T$ decrease toward the poles. This is due to two effects: (i) cloud heights show a general decrease with latitude, and (ii) the decrease of atmospheric temperature with height itself tends to decrease with latitude; that is, a cloud at a fixed height will generally have a smaller $T_s - T_c$ at high latitudes than at low latitudes. The values of present-day $T_s - T_c$ (that is, $\Delta_p T$) are derived by first assuming an effective cloud top height at the equator of 5 km. Combined with a 6°C per kilometer temperature lapse rate, this gives $\Delta_p T = 30$°C. We then adjust $\Delta_p T$ at 60°N and S so that $F_{IR}\uparrow$ at these two latitudes matches the observed mean annual outgoing infrared radiation. The meridional $\Delta_p T$ distribution is then plotted as a smooth line connecting the three values at 0° ± 60°. The veracity of the $\Delta_p T$ distribution was checked by comparing the model-generated meridional distribution of $F_{IR}\uparrow$ to the data given by H. Jacobowitz, W. L. Smith, H. B. Howell, F. W. Nagle, J. R. Hickey, *J. Atmos. Sci.* **36**, 501 (1979).
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