Earth-Science Reviews, 19 (1983) 305–338
Elsevier Science Publishers B.V., Amsterdam—Printed in The Netherlands

A Warm, Equable Cretaceous: the Nature of the Problem

Eric J. Barron

ABSTRACT

Barron, E.J., 1983. A warm equable Cretaceous: the nature of the problem. Earth-Sci. Rev., 19: 305–338.

The nature of the problem of warm equable paleoclimates is defined by investigating the mid-Cretaceous period. The problem consists of three components: (1) defining precisely the climatic state during any specified interval of geologic time; (2) specifying the external climatic forcing factors which may have been important; and (3) understanding the climatic response to any specific modifying influence. These components are characterized by limitations of critical importance in understanding paleoclimates. The nature of the problem of a warm Cretaceous is defined both qualitatively and quantitatively within these limits.

INTRODUCTION

Maintenance of a globally ice-free climate, which may be typical of much of the geologic record over the last 570 m.y., is a fundamental problem in paleoclimatology. This problem is investigated by examining the mid-Cretaceous, which is the best documented example of a warm, equable paleoclimate. Any attempt to understand the behavior of the atmosphere, hydrosphere, cryosphere system over geologic time encompasses three fundamental problems: (1) determination of the climatic state over a specified interval of geologic time; (2) determination of the controlling factors which influenced the climate; and (3) determination of the climatic response to any modifying influence.

The first problem depends both on the ability to stratigraphically subdivide the geologic record and on proxy indicators of paleoclimatic conditions. Because of the lack of any means of achieving finer resolution, many time stratigraphic units represent intervals of a few million years. Within these units areal synchroneity cannot be demonstrated. The climatic state is also not recorded directly (e.g., as field of temperature or pressure), but rather by a variety of proxy indicators which may not have a clear quantitative

0012-8252/83/$10.20

relationship to climate. Further, the paleoclimatic indicators must be accurately reconstructed with respect to the spin axis of the earth. These factors, which characterize many of the problems inherent in geologic interpretation, can result in ambiguity.

The second major problem is to identify the principal causes of climatic change which occur on geologic time scales. Most hypotheses involve three basic climatic forcing factors: (1) the solar radiation received at the top of the atmosphere; (2) the composition of the atmosphere; and (3) the nature of the surface of the earth. These three factors include a multitude of possible mechanisms, which jointly or independently may have influenced paleoclimates. Each forcing factor is evaluated largely on the basis of evidence of significant variation over geologic time and the extent to which these variations are associated with geologic phenomena. For instance there is a good correlation between the separation of Australia and Antarctica and the opening of the Drake Passage with development of the circumpolar current and the glaciation of Antarctica (Kennett, 1977). Unfortunately, for most of geologic time the nature of the various climatic forcing factors are poorly specified. During the Cretaceous only the nature of the surface of the earth can be well specified and this does not preclude the importance of other forcing factors.

Even in cases where strong arguments can be presented for a specific forcing factor, the climatic response to this factor may be only partially understood. Perhaps the best defined correlation between the geologic record and a specific climatic forcing factor is the correlation between Pleistocene glacial cycles and the earth's orbital elements (Hays et al., 1976). Yet,

---

Eric Barron received his initial training in geology at Florida State University. He continued his education at the Rosenstiel School of Marine and Atmospheric Sciences of the University of Miami, receiving Master of Science and Ph.D. degrees in marine geology and geophysics. After a one-year postdoctoral fellowship at the National Center for Atmospheric Research (NCAR), he joined the NCAR scientific staff in the Climate Section of the Atmospheric Analysis and Prediction Division. He is the first NCAR scientist whose primary interests are in the field of paleoclimatology, paleoceanography and plate tectonics. Present address: National Center for Atmospheric Research, P.O. Box 3000, Boulder, Colorado 80307 (USA).

climate models have so far been unable to reproduce glacial cycles with orbital forcing (e.g., one of the best models to date is by Pollard et al., 1980) without arbitrary tuning of some model parameter. Despite the fact that the geologic record is better defined and the climatic forcing factor is more clearly stated for the Pleistocene than for any other time period, the problem is still only two-thirds solved.

The third component is to demonstrate that climate is sensitive to the change in a specific forcing factor by determining the spatial and temporal response to the modifying influence. This aspect of the problem requires both a qualitative and quantitative understanding of the various interactions and feedback mechanisms within the climate system. This is the goal of climate models which incorporate the basic physics or parameterizations of physical processes governing the climate. However, this approach is constrained by the limited capabilities of present models, whose simulation ability can largely be verified only against observations of the present-day climate.

The focus of paleoclimatic research has been largely on the reconstruction of the climatic history through observations and interpretation of data. Understanding the behavior of the climate system through geologic time requires an investigation which encompasses all three of the basic problems outlined above. In particular there is considerable potential in studying various mechanisms quantitatively by performing simulations with climate models and then using the paleoclimatic data as observations against which to compare model simulations.

The nature of the problem of a warm, equable Cretaceous climate is defined by examining each of the three fundamental problems described above. First, the paleoclimatic evidence for a warm, equable, ice-free mid-Cretaceous climate is reviewed, thus defining the climatic state within the limits of the data. This will, in turn, define the extent to which the paleoclimatic data can be used as verification for model simulations. Second, the major external forcing factors which have been suggested as explanations for the contrast between warm, equable climates and glacial climates are evaluated. Third, the characteristics of a zonal energy balance model, which is the basis for a quantitative investigation of the Cretaceous climate, are reviewed in detail. Finally, a series of *sensitivity* experiments are described, where model parameters, such as land-sea distribution, have been modified one at a time, so that cause-and-effect relationships can be clearly determined.

PALEOCLIMATIC DATA AND INTERPRETATION

The concept of a warm, equable ice-free Cretaceous is based primarily on the extensive latitudinal distribution of floras, faunas and sedimentary indicators presently restricted to lower latitudes. These data must be translated into quantitative estimates of specific climatic parameters, reconstructed with respect to paleogeography, and segregated into the smallest feasible time increments while maintaining a reasonable data set. This review will be restricted to estimates of mean annual surface temperature, reconstructed with respect to paleolatitude, for the Albian–Cenomanian (112–93 m.y. ago). Although surface temperature is probably the best defined paleoclimatic variable, sufficient ambiguity exists that the interpretation must be limited to "maximum" (warmest) and "minimum" (coolest) estimates of mid-Cretaceous paleotemperatures. Each paleoclimatic indicator which contributes to these estimates is examined in detail.

Equatorial temperatures

Isotopic data. Although 137 sites from Deep Sea Drilling Project legs 1–75 penetrated Cretaceous sediments, the amount of isotopic data from planktonic Foraminifera is limited. For the Albian–Cenomanian only sediments from the Shatsky Rise and Hess Rise were suitable for analysis (Douglas and Savin, 1975; Savin, 1977). The Cretaceous paleolatitude of these sites was probably in the tropics (Firstbrook et al., 1979). Assuming a mean δO^{18} of $-1‰$ for ocean water on an ice-free earth, the planktonic Foraminifera record temperatures of 25–27°C (298–300°K). It is reasonable to assume that the isotopically lightest measurement (27°C) reflects the shallowest dwelling Foraminifera. Present-day shallow dwelling forms give isotopic temperatures which are about 3–5°C cooler than the surface and selective dissolution tends to bias isotopic measurements in the cold direction (Savin et al., 1975). If the Foraminifera are unaltered, surface temperatures in the Cretaceous tropics were probably in the range of 27–32°C (300–305°K), similar to or warmer than the present day. However, uncertainties in the mean δO^{18} of the Cretaceous oceans may result in as much as a 2.5°C additional uncertainty in all paleotemperatures.

Midlatitude temperatures

Coral reefs and carbonates. At present, extensive carbonate deposition and reef development are limited to the warm, clear waters of the tropics and subtropics, presently within 30° of the Equator (e.g., Wilson, 1975). The optimum temperatures for modern shallow reef corals are between 25° and 30°C with minimum and maximum temperature tolerances approximately 5–7°C from this range (Vaughn and Wells, 1943). Schwarzbach (1963) has suggested that extensive carbonate deposition and reef development only occurs where the water temperature exceeds 21°C. If this limit can be applied to the Cretaceous, the maximum latitudinal extent of Cretaceous carbonates and reefs can be used as an indicator of the 21°C (294°K) isotherm. The mid-Cretaceous was an optimum time of reef development and Cretaceous carbonate sediments (e.g. Habicht, 1979) plotted on a mid-Cretaceous paleographic map (Barron et al., 1981b) extend 5° to 15° poleward of the present-day distribution. A poleward shift of the 21°C isotherm of approximately 5° in latitude is a conservative interpretation of this data.

Isotopic data from epicontinental seas. Extensive isotopic data have been derived from the calcareous internal skeleton, or rostrum, of belemnites. Belemnites occupied a shelf habitat and were restricted to the mid-latitudes during much of the Late Cretaceous (Bowen, 1964). Several factors have hampered the interpretation of these data: (1) in a shelf habitat salinity may substantially affect the isotopic paleotemperature (Lowenstam and Epstein, 1954); (2) considerable isotopic variation may result from diagenetic alteration (e.g. Spaeth et al., 1971); and (3) the recorded isotopic temperature may be the result of seasonal shell formation or represent seasonal migration (e.g., Lowenstam and Epstein, 1954; Berlin et al., 1967) or even a preference for a specific thermal or depth habitat. Because both freshwater dilution of sea water and diagenesis typically result in isotopically lighter values, there is a tendency to rely only on the isotopically heaviest measurements (e.g., Lowenstam, 1964). Using only minimum temperatures does not insure reliability as this may result in a selection for the depth habitat of the organism, reflecting a cooler temperature than that at the sea surface. If the measurements were completed based on preservation of the specimen, mean isotopic measurements from different regions are probably equally reliable.

Mean isotopic temperatures measured on Albian–Cenomanian belemnites by Lowenstam and Epstein (1954), Bowen (1961a, b), Bowen and Fontes (1963), Teis et al. (1957), Spaeth et al. (1971), and Stevens and Clayton (1971) are plotted with respect to paleolatitude in Fig. 1. Each temperature has been calculated using the paleotemperature equation of Craig (1965) assuming a δO^{18} of $-1‰$ for mean ocean water on an ice-free earth. The paleolatitudes were determined using the paleogeographic reconstruction of Barron et al. (1981b). Mean values were derived by averaging data from specific authors grouped into specific regions. A regression line is plotted through the data as an estimate of the midlatitude surface temperature gradient. A similar line drawn through only *minimum* isotopic values would be approximately 1.5°C below the line through the mean values.

The Northern Hemisphere data seems to imply a reduced midlatitude temperature gradient compared to the present day. If this line can be extrapolated, a polar temperature of 5°C (278°K) would be indicated. Conversely, the data could be interpreted as suggesting little or no change in

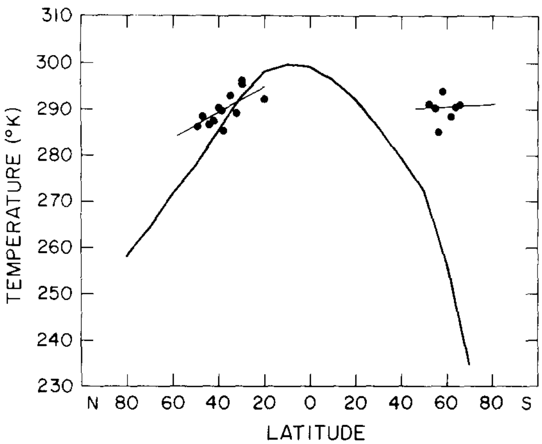

Fig. 1. Isotopic paleotemperature data from belemnites from the Albian–Cenomanian plotted with respect to paleolatitude. See text for an explanation of the data sources and interpretations. The present-day mean annual surface temperature is plotted for comparison.

midlatitude temperatures compared to the present day. The majority of these data points fall within the range of present-day observed January and July sea surface temperatures (Fig. 2). If the isotopic temperatures reflect summer shell formation, they may not be indicative of any significant Cretaceous midlatitude warming. However, the isotopic data would then be indicative of a mean *summer* temperature of 18°C at 35°N. This temperature, which is below the limits described for extensive carbonate deposition and near the temperature tolerances of modern corals, conflicts with the *minimum* temper-

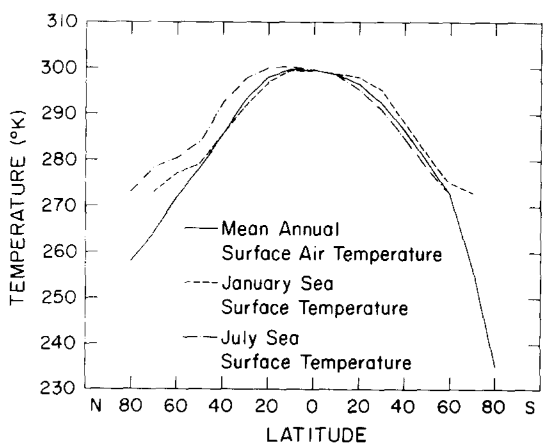

Fig. 2. A comparison of present-day mean annual surface air temperature and Janurary and July sea surface temperature plotted with respect to latitude.

ature at 35°N indicated by corals and carbonate deposition.

The Southern Hemisphere isotopic data appears to reflect increasing temperature toward the poles (as does the minimum values). To a large degree this may reflect a habitat preference for a limited temperature range. Perhaps more important, all the measurements are substantially above present-day temperatures. The coldest Southern Hemisphere determination for the Albian–Cenomanian is 12°C (56°S). This is 6–7°C warmer than present-day summer sea-surface temperatures at 56°S.

Flora and fauna. There exists substantial paleobotanical data which indicate that the boundary between seasonal and nonseasonal floras was displaced as much as 15° poleward compared to the present day (Hughes, 1964; Barnard, 1973; Vakhrameev, 1975). This is indicative of substantially warmer winter temperatures and also suggests that portions of the continents did experience subfreezing winter temperatures. Although it is tempting to speculate that midlatitude mean annual isotherms were displaced 15°, warmer winter temperatures do not necessarily imply warmer mean temperatures.

Perhaps the most persuasive faunal indicator is the record of ectotherms, such as alligators and crocodiles. Terrestrial tetrapod faunas (Colbert, 1964) extended above 60° in paleolatitude during the Cretaceous. Modern alligators become inert below 20°C and have optimum conditions near 34°C (Colbert, 1964). At present, even in midsummer (June–August mean) the 20°C isotherm is displaced poleward of 40° in latitude only in continental interiors (e.g. Newell et al., 1972) and a viable population could only exist equatorward of the summer 20°C mean. These data may suggest a poleward displacement of midlatitude isotherms of near 20° and are additional evidence of a warmer and more equable midlatitude climate during the Cretaceous. However, these data encompass a time span greater than the Albian–Cenomanian, which may explain the extensive distribution.

Polar temperatures

Isotopic data. Isotopic measurements for Albian benthic Foraminifera are as warm as 17°C (e.g. Savin, 1977). These warm deep ocean temperatures have been interpreted as strong evidence for considerably warmer polar temperatures based on a uniformitarian concept of bottom water formation. Although the mid-Cretaceous data are meager, they are substantiated by isotopic data indicative of a progressive decline in deep water temperatures from 15°C to 2–3°C throughout the Tertiary (Emiliani, 1954; 1961; Douglas and Savin, 1971; 1973; 1975; Savin, 1977; Shackleton and Kennett, 1975a; b). A uniformitarian concept of bottom water formation would imply polar temperatures near 15°C (288°K) during the Albian–Cenomanian (Fig. 3). However, temperatures may have been considerably less than 15°C. Brass et

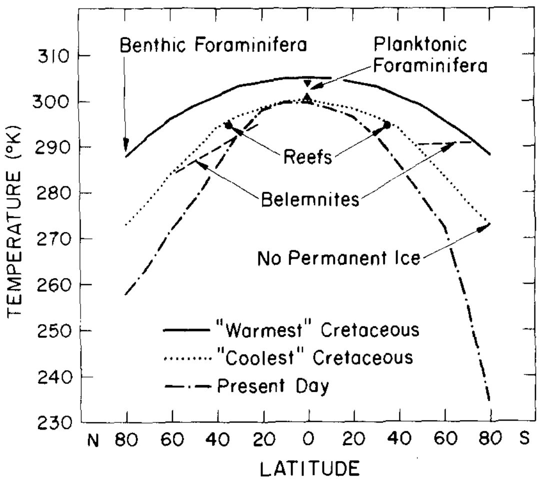

Fig. 3. A summary of data used to define "warmest" and "coolest" mean annual surface temperature estimates for the mid-Cretaceous. The references to Benthic Foraminifera, and Planktonic Foraminifera refer to isotopic paleotemperatures. The two Cretaceous estimates are compared with present-day values.

al. (1982) have shown that during periods of extensive inland seas within the subtropics, deep water temperatures may not reflect a high latitude source of bottom water. Warm salty bottom water forming in the subtropics has the potential to become the global deep water if the buoyancy flux is greater than at high latitudes. The value of 15°C then serves as an upper limit to the temperature of polar seas.

Flora and fauna. Warm polar temperatures receive additional support from the lack of any cold water faunas (Sohl, 1969; Kauffman, 1973) and from the abundance of thermophilous floras at very high latitudes (Smiley, 1967; Taylor, 1972; Krassilov, 1973; 1981; Vakhrameev, 1975). For example, the diverse paleofloras (420 species) from the northern coast of Alaska, near 80°N in paleolatitude, support the interpretation of warm coastal conditions.

Lack of permanent ice. The lower limit to polar temperature is provided by the lack of evidence for any permanent ice. Craddock et al. (1964) have postulated a glacial origin for the erosion of an Antarctic basement complex (73°S 94°W), but the unconformity can only be dated in the range from 10–100 m.y. ago. The first clear evidence for permanent ice on Antarctica is from the Eocene (LeMasurier, 1970; Geitzenaur et al., 1968; Margolis and Kennett, 1971). The lack of permanent ice can be interpreted as a minimum mean annual polar temperature of 0°C (Fig. 3), but allows for the possibility of seasonal subfreezing conditions. Knowledge of the geology of Antarctica is limited and the possibility of some permanent ice, even in the Cretaceous, remains.

Summary

Almost every type of paleoclimatic data has significant limitations. The data give a coherent interpretation of warmer conditions, particularly at higher latitudes, however, for the most part, these data are difficult to quantify. The data allow only estimates which bracket the Cretaceous climate by "warmest" and "coolest" estimates of mean annual, surface paleotemperatures.

The "warmest" estimate is based on a 32°C (305°K) tropical sea surface temperature and a polar ocean temperature of 15°C (288°K). This estimate encompasses all of the different types of paleoclimatic data including all the isotopic data from belemnites in the Southern Hemisphere. It exceeds the warming required by much of the data from the midlatitude Cretaceous.

The "coolest" estimate of Cretaceous temperatures is based on the 27°C (300°K) tropical sea-surface temperature measured from planktonic Foraminifera, the minimum temperature required to explain the distribution of Cretaceous reefs, and the minimum polar temperature required to insure seasonally ice-free poles (0°C). Seasonal indicators, such as paleofloral characteristics, and some of the data from Northern Hemisphere belemnites, may fall below this estimate. This information is summarized in Fig. 3.

EXTERNAL CLIMATIC FORCING FACTORS

Most hypotheses of climatic change on time scales of 10^6–10^8 years (e.g., Wigley, 1981) involve some aspect of three factors: (1) the solar radiation received at the top of the atmosphere; (2) the composition of the atmosphere; and (3) the nature of the surface of the earth. The purpose of this review is to examine these factors in order to evaluate whether they might explain the contrast between the warm, equable Mesozoic and the present-day glacial climate.

The criteria by which each of the proposed hypotheses can be evaluated include: (1) theoretical and observational evidence for variability of a proposed forcing factor (e.g., for variations in solar luminosity); (2) the time scale of the variability; and (3) the magnitude of the variation. In order to explain the transition from a warm, ice-free Cretaceous to the present glacial climate, the forcing factor must result in a global cooling on a time scale of 40–60 m.y.

Variations in solar input

The amount and distribution of incoming solar energy varies with changes in solar luminosity, the orbital elements of the earth and through encounter with dense clouds of interstellar matter.

An increase in solar luminosity of 30–40% since the formation of the sun $4.5 \cdot 10^9$ years ago seems an unavoidable consequence of the increase in mean atomic weight of stellar material (e.g., Schwarzchild et al., 1957; Hasselgrove and Hoyle, 1959; Bahcall and Shaviv, 1968). Gough (1977) has also speculated that thermal oscillations may mix the solar core and destroy nuclear equilibrium resulting in decreased luminosity ($\approx 5\%$) lasting approximately $4 \cdot 10^6$ years followed by an above normal increase of approximately the same magnitude and duration. These theoretical considerations suggest that variations in solar luminosity of relatively large magnitude may occur over geologic time scales.

The least speculative conclusion from solar models is that the solar output was slightly less, 1%, 100 million years ago (from equation given by Gough, 1977) than at present. This is in the opposite sense of the climatic trend. The more speculative hypothesis of thermal mixing is inconsistent with the time scale of the global cooling trend from the Cretaceous to the present. We may conclude that, *as yet*, no hypothesis involving solar luminosity variations can account for the climatic record of the last 100 million years.

The flux of solar radiation is also dependent on the earth–sun distance and the solar angle, thus orbital perturbations modify the spatial and temporal distribution of insolation. Shorter term perturbations in the orbital elements ($< 10^5$ years) are well established (e.g. Berger, 1976); however, there appears to be no theoretical basis for proposing large scale orbital variations over tens of millions of years. Variations in orbital elements may be required to explain the distribution of floras at high latitudes. Wolfe (1978) has suggested that the distribution of light-limited Eocene floras requires an obliquity of 10°. Pronounced growth rings in Eocene floras may be inconsistent with this interpretation (McKenna, 1980) and the decrease in obliquity should result in a greater equator-to-pole thermal contrast based on the distribution of insolation (Van Woerkom, 1953). This would contradict the paleoclimatic data. The possibility of light-limited floras at high latitudes remains a significant problem.

Three galactic mechanisms of climatic change have been proposed to explain glaciations: (1) passage of the solar system through dust lanes bordering the spiral arms of the galaxy (McCrea, 1975); (2) the eccentricity of the orbit of the solar system about the galactic center (Steiner and Grillmair, 1973); and (3) the tidal flexure of the galaxy (Williams, 1975). Each of these hypotheses attempts to explain a glacial periodicity of 200–400 m.y. Although the record of glaciations may not be complete, as yet the Phanerozoic glaciations (e.g., Tarling, 1978) do not lend themselves to simple periodic interpretation.

Of the three galactic mechanisms only the hypothesis of McCrea (1975) can be evaluated in terms of its climatic significance. McCrea (1975) analyzed typical densities of interstellar clouds and noted that luminosity variations on the order of 1% or greater would typically result from the passage of the solar system through a dust lane. The present glacial age would have resulted from the solar system's recent passage through the Orion Arm. Both dust loading of the atmosphere (Talbot et al., 1976) and modifications of the solar wind (Begelman and Rees, 1976) may contribute to the climatic effects caused by this mechanism. Certainly there is evidence that this mechanism operates over geologic time scales but in terms of explaining the climatic record over the last 100 m.y. there are significant problems. First, the Orion clouds may be of insufficient density to explain the observed climatic changes (Dennison and Mansfield, 1976). Second, the duration of the passage of the solar system through the dust lanes (one or several million years) does not explain the global cooling over the last 40–60 m.y.

Atmospheric composition

A change in the atmospheric transparency to incoming solar radiation or to outgoing infrared radiation may also be an important climatic forcing factor on geologic time scales. Two hypotheses have been frequently cited concerning the climatic record of the last 100 m.y.: (1) global cooling as the result of the injection of particulate matter into the atmosphere by intensive volcanism; and (2) global cooling as the result of a long-term reduction in carbon dioxide concentration in the atmosphere.

Kennett and Thunell (1975) proposed that an episode of intense explosive volcanism was the cause of the Quaternary glaciation, based on an analysis of deep sea cores. Whereas individual eruptions are unlikely to have a long-term effect, and episode of intensive volcanism has potential to become an important climatic factor. However, Ninkovitch and Donn (1976) have suggested that the deep sea record only reflects an increase in volcanism because the sea floor is approaching the source of the volcanism (subduction zones). A more important factor is that volcanic ash alters to bentonite over a period of approximately $4\text{–}5 \cdot 10^6$ years (Hein et al., 1978), and therefore visual analyses of core material is not adequate to decipher the volcanic record. Hein et al. (1978) demonstrated, by X-ray analyses, that there has been an increase in explosive volcanism over the last 10^6 years but that this episode is not unusual in quantity or duration for Tertiary sediments. Therefore, volcanism becomes a less likely primary mechanism to explain the Quaternary glaciation.

Geologic history of carbon dioxide in the atmosphere is poorly constrained. The long-term fluxes between the ocean–atmosphere–lithosphere reservoirs are controlled by two major coupled cycles, the cycle of volcanic input, rock weathering, and deposition of sediments and the cycle of photosynthesis, respiration and decay and storage of organic matter. Because the atmospheric reservoir is small compared to the oceans or the lithosphere, and because the amount of fossil carbon is very large compared to the amount stored in living organic matter (Baes et al., 1977), there is potential for significant modification of the atmospheric pCO_2. The size of the reservoirs of carbon has varied significantly through time. For instance the amount of organic matter stored in the lithosphere, as fossil fuels, appears to have varied significantly over the last 100 m.y. (e.g., Irving et al., 1974). Based on mass balance calculations of volcanogenic rocks, total CO_2 in sediments and organic carbon in sediments, Budyko and Ronov (1979) have suggested that the concentration of CO_2 was seven times that at present, 100 m.y. ago.

However, arguments arise over how well buffered the CO_2 system is from changes in any reservoir. Broecker and Takahasi (1977) have determined that the amount of calcium carbonate in the ocean sediment reservoir, available to balance atmospheric CO_2 increases is sufficient to neutralize all known reserves of organic carbon should they be oxidized. However, this may not be an important geologic constraint because the total reserves are only a small fraction of the total sedimentary mass of organic carbon. There are also several additional competing factors. For instance, warming of the planet should act as a feedback mechanism to reduce CO_2 in the atmosphere, because continental weathering reactions are temperature dependent (Walker et al., 1981). During the Cretaceous, as much as 20% of the continents were covered by epicontinental seas. Therefore the area of continent available for weathering would decrease, tending to increase CO_2 in the atmosphere. In addition, in the case of warmer oceans, the solubility of CO_2 in sea water decreases (Eriksson, 1963) and hence an increase in atmospheric CO_2 could occur during warm time periods. It is not known if the changes in atmospheric CO_2 will be balanced by changes in the amount of vegetation or in the storage of organic matter. Unfortunately, not even the relationship between CO_2 concentration and photosynthesis rates are well known (Baes et al., 1977). Variations in atmospheric CO_2 concentrations over the last 100 m.y. are plausible but the estimates are speculative.

Surface of the earth

The hypotheses which have been proposed to explain glaciations include the entire spectrum of tectonic processes from changes in geographic barriers (e.g., Tanner, 1968) to global factors such as episodes of orogenesis (Umbgrove, 1947) or epeirogeny (Damon, 1968), to changes in continental paleopositions and the latitudinal distribution of land masses (Crowell and

Frakes, 1970; Frakes and Kemp, 1972; 1973; Beaty, 1978; Tarling, 1978; Frakes, 1979).

Umbgrove (1947) presented a classical hypothesis which related glaciation with periods of mountain building, because of the resulting increase in the area of land above the snow line. More advanced stratigraphic correlations do not give a particularly well-defined correlation between mountain building and glaciation, although orogenesis is probably a contributing factor (e.g., Hamilton, 1968; Crowell and Frakes, 1970; Fairbridge, 1973).

Epeirogeny and changes in the distribution of land masses have a much better correlation with paleoclimatic trends. Fig. 4 is a comparison of several geographic trends from 100 m.y. ago to the present, including the area of land above 60° in paleolatitude, the area of land in the subtropics, the area of epicontinental seas in the subtropics, and the total global land area, as measured from paleogeographic maps (Barron et al., 1981b). These geographic trends compare favorably with a generalized curve of isotopic paleotemperature data from benthic Foraminifera (e.g. see Savin, 1977). The total land area and the land area in subtropics both slowly increase from 100 m.y. ago to the present. Although the area of land greater than 60° North

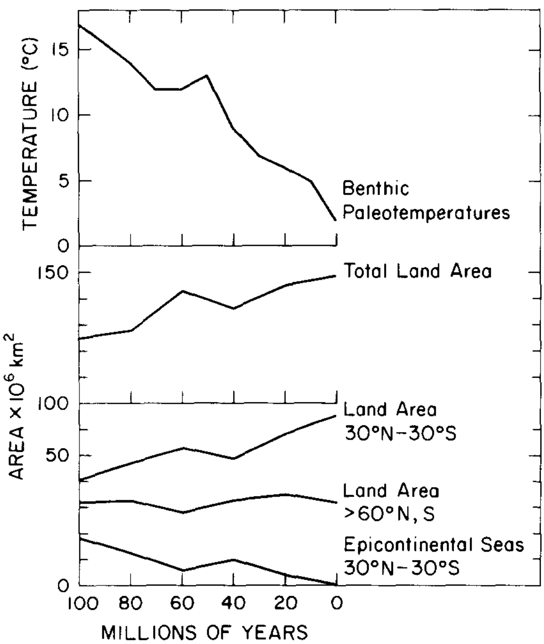

Fig. 4. A comparison of paleotemperature and paleogeographic trends for the period 100 m.y. ago to the present. The paleotemperature trend is the curve from benthic Foraminifera given by Savin (1977). The paleogeographic trends were measured from paleogeographic maps at 20 m.y. increments (Barron et al., 1981).

and South appears to exhibit little trend, the area of land greater than 60° North does increase from 100 m.y. ago to 20 m.y. ago. The area of epicontinental seas in the subtropics exhibits a trend of decreasing area from 100 m.y. ago to the present.

Plate tectonic processes which result in variations in the size and distribution of continents have potential to explain the paleoclimatic record. These processes operate on the correct time scale and they result in relatively large changes in the surface of the earth.

Much more controversy is associated with the specific geographic mechanisms which have been proposed to explain the climatic record. Numerous authors (Frakes and Kemp, 1972; 1973; Beaty, 1978; Frakes, 1979; Tarling, 1978) cite the importance of high latitude land area available for the accumulation of high albedo snow or as a heat transport inhibitor. Barron et al. (1980) suggested that the area of land in the subtropics and changes in total land area due to eustatic sea level variations (Hays and Pitman, 1973; Pitman, 1978) were more important mechanisms in modifying the planetary albedo. Brass et al. (1982) have suggested that the area of epicontinental seas in the subtropics may control the formation of bottom water. Warm saline bottom water formed by evaporative processes in the subtropics would then explain the paleotemperatures determined from benthic Foraminifera. The relative importance of these specific mechanisms cannot be determined simply by correlation because the trends are all in concert.

Summary

Two conclusions can be drawn from this review. First, there are a large number of possible forcing factors; however, most of these are poorly specified. Second, the only factor which can be well specified and which follows from the criteria stated earlier, are the changes in land–sea distribution. However, unless it can be demonstrated that the changes in land–sea distribution do, in fact, explain the climatic record, then other potential forcing factors cannot be eliminated.

The third major problem in paleoclimatology is to determine whether the climate is sensitive to these changes in geography. This aspect of the problem requires a quantitative study of specific mechanisms by performing simulations with climate models. As yet it is uncertain whether changes in climate may be explained primarily by the changes in geography or whether other forcing factors may be required.

CLIMATE MODELS AND CRETACEOUS SIMULATIONS

The primary goal of investigations using climate models is to determine the extent to which any mechanism or series of mechanisms may explain the paleoclimatic data. The insight we gain into the climate system is dependent on the parameterization of physical processes and the extent to which various feedback processes are included in the models. In other words the simulations will reflect model sensitivities. The capabilities of present models can only be verified against observations of present-day climate. In addition the forcing factors which may be required to explain the Cretaceous climate are uncertain and the nature of the Cretaceous climate, as defined by mean annual surface temperature, can be described only within two broad limits. Given these limitations, climate models can be used to examine the importance of changes in geography, and the importance of a number of assumed physical processes, by comparing the simulations with the paleoclimatic data. Our goal is not a comprehensive simulation of Cretaceous climate, but rather a quantitative definition of the nature of the problem of warm paleoclimates.

Previous studies

A number of simple climate models have been utilized to determine if climate is sensitive to a change in the distribution of the continents. A pioneering study in this area was completed by Luyendyk et al. (1972) using the laboratory model of Von Arx (1957) to attempt to simulate the ocean circulation using Middle Cretaceous geography. With this geography there was a significant transport of water into the Arctic Ocean. Luyendyk et al. (1972) suggested that the Arctic circulation may have inhibited glaciation at the pole. Since this study there have been considerable improvements in the positioning of the continents, modeling bathymetry, and determining the history of aseismic ridges, such as the Iceland–Faroes ridge, which may act as oceanographic barriers (e.g., Sclater et al., 1977). The conclusions from Lyuendyk et al. (1972) probably no longer apply to an actual paleogeographic situation.

The pioneering effort in using a mathematical atmospheric model to investigate the importance of paleocontinental positions was by Donn and Shaw (1977) using a simple thermodynamic model. Donn and Shaw completed simulations with four different paleocontinental reconstructions and appear to have demonstrated that the Tertiary global cooling was the direct result of an increase in high latitude land in the Northern Hemisphere. There are, however, several important limitations to these simulations. First, the continental positions Donn and Shaw (1977) used differ significantly from most compilations (e.g. Smith and Briden, 1977). The movement of Eurasia was determined only by paleomagnetic data and this results in a very large change in the area of land at high latitudes compared with reconstructions which are based on sea floor spreading data and paleomagnetic data. More important, the global cooling result was directly dependent on one of the model assumptions. Donn and Shaw (1977) specified a present-day albedo distribution for factors such as snow cover. Apparently the global cooling trend resulted because snow cover was assumed for any land surface greater than 60° in latitude, rather than calculated as a function of temperature. Donn and Shaw (1977) did prescribe equator-to-pole heat transport in a manner that should have reduced this global cooling. The energy transport was defined by an eddy diffusion coefficient which was arbitrarily increased from 1/2 of its present value for the Triassic case to its present value for the modern case. If Donn and Shaw (1977) had performed sensitivity experiments changing one parameter at a time these results might have been easier to interpret. As yet the conclusion that the increase in high latitude land resulted in the global cooling trend is not yet warranted.

There are two additional studies (Sellers and Meadows, 1975; Cogley, 1979) which examine the changes in albedo as a function of the distribution of the continents. Each of these studies is based on assigning a surface albedo for land and sea and calculating the total absorbed solar energy for a fixed value of incident solar energy. These two studies use different continental distributions. The Sellers and Meadows (1975) study is based on paleomagnetic data (McElhinny, 1973), and they conclude that glaciation occurs when land masses are associated with polar regions. Cogley (1979) considers the land area in the subtropics to be more important, but using reconstructions of Briden et al. (1974) he concludes that continental drift is only a modest mechanism of climatic change. Neither study is realistic because they assume that the total land area is a constant (i.e., sea level is fixed). Paleogeographic reconstructions (rather than paleocontinental) clearly show that the changes in land area as a result of eustatic sea level variations are large (Barron et al., 1981b). Barron et al. (1980), using a similar simple albedo model, show that sea level is more important than the positions of the continents over the last 200 m.y.

Neither of the above three albedo studies used comprehensive models. The only variable considered was land fraction in each latitude and the surface albedo was specified. The planetary albedo is a product of a combination of factors, including the relationships between surface temperature and snow cover, atmospheric moisture content, and cloud cover, as well as land-sea areas. Thompson and Barron (1981) constructed a more rigorous planetary albedo model for the mid-Cretaceous. By performing a series of sensitivity experiments, modifying one variable at a time, Thompson and Barron examine the importance of a number of these factors. These experiments suggested that an increase in total absorbed solar energy of between 2–4% during the Cretaceous Period is reasonable. Only about a 1% increase occurred because of the change in geography. The remainder of the increase resulted from an assumption of warm polar temperatures and hence no snow and ice, and an assumption of plausible cloud-climate feedbacks.

A 2–4% increase in total absorbed solar energy could explain a large fraction of the global temperature difference between the Cretaceous and the present day. However, the planetary albedo is only one component of the climate system and any increase in absorbed solar energy must be balanced by outgoing infrared radiation and heat transport across latitudes. In addition, the planetary albedo model requires that some climatic parameters, such as surface temperatures, be specified rather than calculated. The sensitivity experiments with planetary albedo models are an indication that climate is sensitive to a change in geography. However, this does not indicate whether the changes in geography will result in an ice-free climate when a more comprehensive climate model is used to investigate this problem.

Energy balance climate model

A hierarchy of climate models exists, from simple one-dimensional models of the vertical temperature structure of the atmosphere which are globally averaged horizontally, to very complex three-dimensional, time-dependent models intended to "simulate" the atmosphere and oceans. Regardless of the level of complexity, these models are mathematical representations of physical processes thought to be important in determining climate. Each type of model has specific uses and specific deficiencies (see review by Schneider and Dickinson, 1974). Because of the uncertainties in defining the mechanisms of climatic change on geologic time scales and the uncertainties in characterizing the climatic state during any geologic time period, initially we wish to take advantage of a comparatively simple model. Simple models are relatively inexpensive in terms of amount of computational time and are relatively easy to interpret physically. In other words, cause-and-effect relationships can be quantified. This follows the goal to define the nature of a warm, equable climate rather than attempt to perform a comprehensive simulation. A zonal energy balance climate model is best suited for this type of study.

The archetypal energy balance model was developed by Budyko (1969) based on the constraint that over long time scales incoming solar energy and outgoing infrared radiation should balance. Budyko wrote a set of simple equations, one for each latitude, where:

$$F_{\downarrow} - F_{\uparrow} = \mathrm{F}_{\leftrightarrow} \tag{1}$$

$F_{\downarrow}$ is the incoming solar energy which is approximately proportional to cos(latitude)^2. The amount of energy available to the atmosphere is dependent on the amount of incoming energy, Q, and the planetary albedo, α. The planetary albedo is the ratio of the amount of solar radiation reflected by the earth to the amount incident upon it. $F_{\uparrow}$ is the outgoing radiation. The thermal energy radiated by a surface or substance is a function of the temperature of the surface. Thus outgoing thermal radiation, $F_{\mathrm{IR}}^{\uparrow}$, tends to limit any change in the temperature of the atmosphere caused by a change in energy input. $F_{\leftrightarrow}$ is the net energy transport by the ocean and the atmosphere. In most simple models the net energy flux transport, div(F) for latitude, ϕ, is considered to be proportional to the temperature gradient (e.g., Schneider and Gal-Chen, 1973).

The simple balance of Budyko can be rewritten more specifically as a function of latitude and the above variables:

$$Q(\phi)[1-\alpha(\phi)] - F_{\mathrm{IR}}^{\uparrow} = \operatorname{div} F(\phi) \tag{2}$$

This equation is only valid for an annual equilibrium climate because of the lack of any time-dependency and the assumption of no net energy storage. Thompson and Schneider (1979) have developed a version of the zonally averaged energy balance model which is time-dependent and includes seasonal heat storage

$$\frac{\partial}{\partial t}[R(\phi)T_{\mathrm{s}}(\phi,t)] = Q(\phi,t)[1-\alpha(\phi,t)] - F_{\mathrm{IR}}^{\uparrow}(\phi,t) - \operatorname{div} F(\phi,t) \tag{3}$$

On the right-hand side are the three energy balance components as a function of time, t. On the left-hand side, R is the thermal inertia (heat capacity per unit area) and T_{s} is the surface temperature. This equation will be a good approximation of the climate system if each of the terms can be approximated adequately. Obviously even in the most comprehensive model not all physical processes can be included. It is necessary instead to approximate (parameterize) a number of factors. For instance, factors which operate on a scale less than that under consideration, such as small scale eddies or molecular processes, might be approximated as diffusion processes. The parameterization of various energy balance components consitutes the bulk of the problem of low resolution climate models. Each of these components is described in greater detail below.

Planetary albedo. The factors which must be taken into account in order to calculate the planetary albedo are the duration and intensity of the solar insolation, the angle of the solar beam from the local zenith and the mechanisms of modifying the incident solar beam. The beam may be depleted by absorption, in which case there is a conversion of shortwave radiation into other forms of energy. The beam may be depleted by scattering resulting in a change in direction of the radiant energy. A percentage of the shortwave radiation may be reflected by a body or surface. The planetary albedo parameterization is a means of energy bookkeeping, an accounting of insolation, solar zenith angle, absorption, scattering and surface reflections.

The planetary albedo component of the energy balance model is identical to that described by Thompson and Barron (1981). This model accounts for atmospheric absorption by ozone, water vapor and dust; scattering by air molecules (Rayleigh scattering) by dry and wet aerosols; cloud and surface solar zenith angle dependence; multiple cloud–surface reflections; and albedo–temperature feedback for land surfaces. The albedo–temperature feedback occurs because the albedo of the surface begins to increase to take into account the empirical relationship of higher surface albedo due to snow when the surface temperature decreases below 10°C. Water vapor–temperature feedback is also included, with the amount of water vapor increasing with increasing surface temperature.

Outgoing infrared radiation. The outgoing infrared radiation can be approximated, in good agreement with observations (Cess, 1976), as a linear function of the surface temperature.

$$F_{\mathrm{IR}}^{\uparrow} = A + B_s T + C F_c (T_s - T_c) \tag{4}$$

where F_c is the fraction of latitude covered by clouds, T_c is the effective temperature of the cloud tops and A, B and C are constants equal to -248.1 W m^{-2}, 1.8 W m^{-2}°K and -1.73 W m^{-2}°K, respectively (Warren and Schneider, 1979; Ramanathan, 1977).

Clouds have two competing effects on surface temperature (Schneider and Thompson, 1980). Clouds reflect solar radiation thus cooling the surface. Clouds also usually radiate at a lower temperature than the surface, and therefore they act to decrease $F_{\mathrm{IR}}^{\uparrow}$ relative to a cloudless sky. In the energy balance climate model the global mean effect of changes in cloud amount on albedo is larger than on $F_{\mathrm{IR}}{}^{\uparrow}$, in agreement with observational studies (Ellis, 1978; Hartman and Short, 1980). This issue is still in dispute. Cess (1976), employing zonally averaged climatological data, concluded that changes in global total cloud amount resulting from a climate change would have almost no effect on the globally averaged energy balance.

The outgoing infrared radiation also depends on the difference between the surface temperature and the cloud top temperature. The larger the difference, the more the presence of cloud will affect the outgoing infrared radiation.

Both cloud amount and $T_s - T_c$ will be important factors in modeling paleoclimates if there are substantial cloud–climate feedbacks. Unfortunately, these feedbacks are poorly understood (Schneider et al., 1978) and there exists no data from the geologic record.

Meridional heat transport. The energy transported poleward by the atmosphere may be divided into two components, sensible plus potential energy, and latent energy. Since the transport processes occur on space scales similar to or smaller than that of the model (10° latitude zones), then the transport is approximated as a diffusion process proportional to a gradient. Sensible plus potential energy transport is proportional to the meridional gradient of surface temperature, T_s. Latent heat transport is proportional to the meridional gradient of atmospheric water vapor concentration. The diffusion coefficients are latitude dependent and are tuned so that the model matches the presently observed annual mean distribution of atmospheric heat transports (Oort and Vonder Haar, 1976). Outside the tropics the diffusion coefficients are assumed to be proportional to the absolute value of the temperature gradient, and thus the diffusion is nonlinear.

The transport of energy by the oceans is specified as the presently observed annual mean values from Oort and Vonder Haar (1976) for the Northern Hemisphere and from Jacobowitz et al. (1979) for the Southern Hemisphere. The transport of energy by the oceans is specified because, to our knowledge, there are no verified parameterizations for this component for today's conditions, much less for a time of substantially different continental configurations.

Gal-Chen and Schneider (1976), Thompson and Schneider (1979) and Warren and Schneider (1979) discuss the forms and effects of parameterizations of meridional heat transport in energy balance climate models.

Thermal inertia. The thermal inertia, R, is a strong determinant of the amplitude of the annual cycle of temperature. Thompson and Schneider (1979) developed a weighted average of the thermal inertias of land and ocean in order to calculate a zonal average. The thermal inertia of land is taken to be equivalent to that of a 3 m water depth. The effective thermal inertia of the ocean for seasonal time scales is determined by the depth of the mixed layer. The latitude-dependent mixed layer depths for present day simulations are 80 m from 50°N to 30°S, 220 m from 80°N–70°N and 50°S–80°S, and 150 m at 60°N and 40°S. Using these values the amplitude of the annual cycle of surface air temperature matches presently observed values at each latitude within a few percent. Because the effective thermal inertia of the ocean is much greater than land, latitudes with large ocean fractions will have smaller seasonal temperature amplitudes than at other latitudes.

With each term in the energy balance equation specified, eq. 3 is solved numerically for surface temperature by integrating stepwise in time until an equilibrium annual climatic cycle is achieved. A present-day simulation, with specified external conditions, will be a control simulation. A series of sensitivity experiments, with one change in external or internal conditions made for each simulation, can be used to investigate the model response to specific forcing factors or assumed physical processes.

Cretaceous sensitivity experiments

A series of sensitivity experiments were designed (Barron et al., 1981a) to determine the extent to which any external or internal mechanisms or combination of mechanisms may explain the warm, equable mid-Cretaceous climate. An explanation of the mid-Cretaceous climate involves two types of problems.

The first problem involves mechanisms which will increase the global temperature above its present-day value to its Cretaceous value. The present-day globally averaged surface temperature is 287.5°K. The "minimum" and "maximum" Cretaceous globally averaged surface temperature, derived from Fig. 3, are 6° and 14° greater than the present day value. Obviously the mechanisms required to explain the warmest planetary temperature will be of greater magnitude than for the cooler estimate. The major question is whether these global changes can be achieved with mechanisms of plausible magnitude.

The second problem involves the mechanisms which will result in the equator-to-pole distribution of temperature. From Fig. 3 the present-day equator-to-pole temperature contrast is 41°, the "minimum" Cretaceous temperature distribution has a contrast of 26°, and the "maximum"Cretaceous temperature distribution has a contrast of only 17°. The problem is that compared to the present-day Cretaceous polar temperatures apparently increased much more than the equatorial temperatures. An important effect of decreased meridional temperature gradient arises if the heat transport to the poles is proportional to the equator-to-pole surface temperature gradient. The poleward heat transport during the Cretaceous will decrease compared to the present. This is one reason why many hypotheses which have been presented to explain warm climates involve modifications of the high latitude radiation balance (e.g., less high latitude land available for snow accumulation). Again the major question is whether the changes in the distribution of temperatures can be achieved with mechanisms of plausible magnitude.

Boundary conditions. Five boundary conditions must be prescribed for the present-day and for the Cretaceous model simulations. These include (1) fraction of a zone covered by ocean, (2) fraction of a zone covered by sea ice (annual mean), (3) fraction of a zone covered by clouds and the value of $T_s - T_c$ (surface temperature minus cloud top temperature), (4) a "base" land albedo to take into account surface characteristics such as vegetation, and (5) mixed-layer depths for the ocean. The present-day values, which are based on climatological means (Table I), define a control simulation. This simulation compares favorably with observations of mean annual temperature with respect to latitude and with the seasonal temperature amplitude.

TABLE I

Present-day averaged values of fraction of zone covered by ocean, fraction of ocean covered by sea ice (annual mean), fraction of zone covered by clouds and $T_s - T_c$ (surface temperature minus cloud top temperature), base land albedo and mixed-layer depths for the present-day control simulation

Latitude	Ocean Fraction	Sea ice	Cloud fraction	$T_s - T_c$ (°C)	Base land albedo	Mixed-layer depths (m)
85°–75°N	0.81	0.90	0.64	14.0	0.16	220
75°–65°N	0.45	0.50	0.66	16.4	0.16	220
65°–55°N	0.38	0.14	0.69	19.5	0.16	150
55°–45°N	0.42	0.00	0.67	21.6	0.16	80
45°–35°N	0.54	0.00	0.60	24.6	0.16	80
35°–25°N	0.59	0.00	0.52	27.4	0.16	80
25°–15°N	0.68	0.00	0.41	29.4	0.16	80
15°– 5°N	0.76	0.00	0.47	30.0	0.16	80
5°N–5°S	0.77	0.00	0.49	30.0	0.16	80
5°–15°S	0.78	0.00	0.48	29.6	0.16	80
15°–25°C	0.76	0.00	0.46	28.8	0.16	80
25°–35°S	0.81	0.00	0.49	27.2	0.16	80
35°–45°S	0.95	0.00	0.58	25.0	0.16	150
45°–55°S	0.98	0.00	0.73	22.2	0.16	220
55°–65°S	1.00	0.26	0.81	19.5	0.16	220
65°–75°S	0.62	0.69	0.67	13.2	0.16	220
75°–85°S	0.07	1.00	0.47	4.8	0.16	220

For the mid-Cretaceous the land–sea fraction is based on the paleogeographic map at 100 million years ago by Barron et al. (1981b). The mean annual sea ice fraction is assumed to be zero because of the lack of any evidence for significant sea ice. Initially, present-day cloud values are specified because cloud–climate feedbacks are not well understood. A time independent value of 100 m is specified as the mixed-layer depth at all latitudes. (This assumption will modify the seasonal temperature amplitude but will not affect the two problems being considered in this review.) The "base" land albedos for each latitude are the same as those for the present day. The assumptions for each successive Cretaceous simulation are given in Table II for comparison with the present-day control simulation.

The importance of the Cretaceous assumptions are investigated by a series of sensitivity experiments. For each experiment we will examine the change in the globally averaged surface temperature and the distribution of the change in surface temperature with respect to latitude.

TABLE II

A summary of average values used in Cretaceous experiments for comparison with the present-day control simulation (Table I) *[1]

Latitude	Ocean fraction	Sea ice	Cloud fraction	$T_s - T_c$ (°C)	Base land albedo	Mixed-layer depths (m)
85°-75°N	0.67	0.00	0.50	19.8	0.16	100
75°-65°N	0.58	0.00	0.50	21.6	0.16	100
65°-55°N	0.55	0.00	0.50	24.0	0.16	100
55°-45°N	0.55	0.00	0.50	25.8	0.16	100
45°-35°N	0.62	0.00	0.50	28.0	0.12	100
35°-25°N	0.79	0.00	0.50	28.8	0.12	100
25°-15°N	0.90	0.00	0.50	29.4	0.12	100
15°- 5°N	0.89	0.00	0.50	30.0	0.12	100
5°N-5°S	0.81	0.00	0.50	30.0	0.12	100
5°-15°S	0.78	0.00	0.50	30.0	0.12	100
15°-25°S	0.81	0.00	0.50	29.4	0.12	100
25°-35°S	0.83	0.00	0.50	28.8	0.12	100
35°-45°S	0.80	0.00	0.50	28.0	0.12	100
45°-55°S	0.82	0.00	0.50	25.8	0.16	100
55°-65°S	0.83	0.00	0.50	24.0	0.16	100
65°-75°S	0.56	0.00	0.50	21.6	0.16	100
75°-85°S	0.76	0.00	0.50	19.8	0.16	100

*[1] Note that the change in the specification from present day to these Cretaceous values, in the model simulations, is completed by a series of sensitivity experiments.

The Cretaceous simulation. The Cretaceous simulation results in a 1.2% increase in total absorbed solar energy (from 239.0 W m^{-2} to 241.9 W m^{-2}) and a global surface temperature increase of 1.62°K compared to the present-day control. This is a significant global warming but is only a fraction of the 6–14°K warming required by the paleoclimatic data. Fig. 5 compares the increase in surface temperature with respect to latitude required to achieve the minimum temperature estimate for the Cretaceous, with the increase generated by the model. The largest increase in surface temperature is at high latitudes, however, the poles remain too cold.

Geography is apparently an important mechanism of climatic change, but *if* the physical processes in the model are adequately formulated and *if* the paleoclimatic data are interpreted correctly, then geography is inadequate to explain the hypothesized paleotemperatures. Herein lies the problem in quantitatively defining the nature of warm, equable paleoclimates. The discrepancy between the model calculations may be explained by any one or

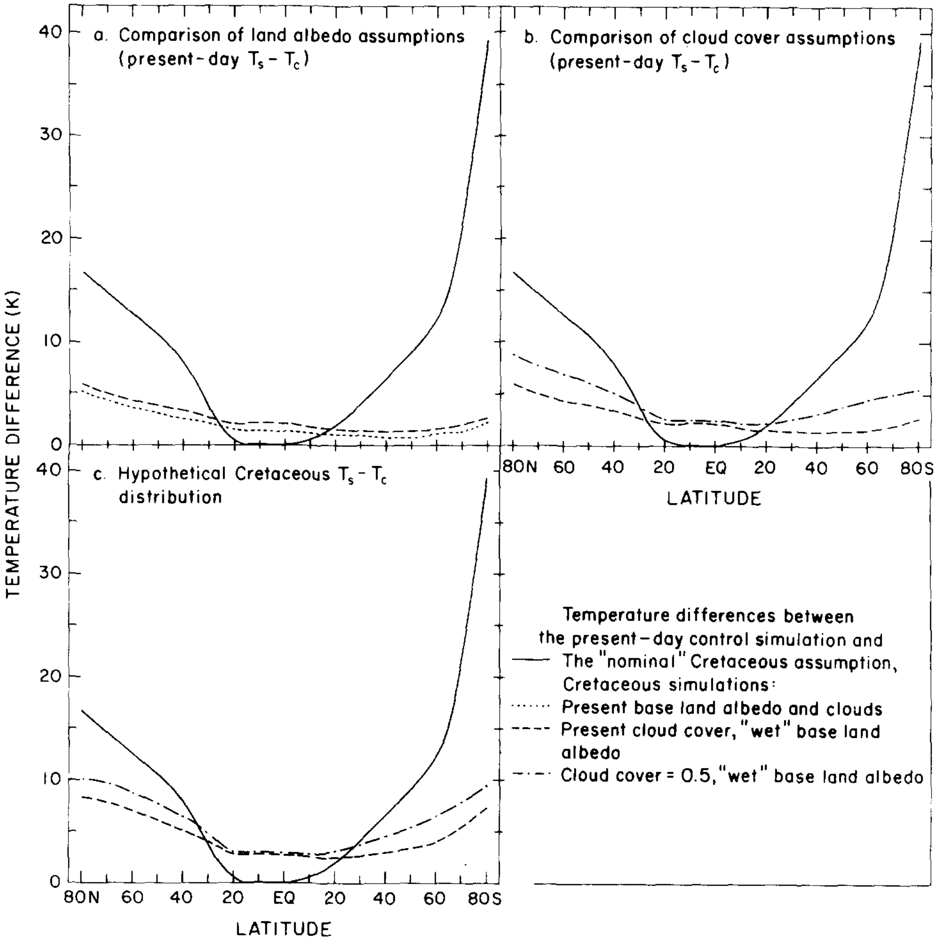

Fig. 5. Temperature differences for the present-day control simulation. The temperature difference between the present-day control simulation and the hypothetical "coolest" Cretaceous distribution (Fig. 3) is indicated in (A), (B), and (C) by a solid line. The temperature difference between the present-day 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 distributions of $T_s - T_c$ (surface temperature minus cloud top temperature) given in Table II. The closer the dotted or dashed curves are to the solid curve, the "better" the simulation.

a combination of three factors: (1) the parameterization of physical processes in the model may be incomplete or incorrect; (2) the paleoclimatic data may be interpreted incorrectly; or (3) other external climatic forcing factors, in addition to paleogeography, may have been important.

The Cretaceous simulation achieves less than a third of the global warming suggested by the "minimum" interpretation of the data. In addition, the warmest estimate, a 14°K increase in globally averaged surface temperature, is much closer to the generally accepted interpretation of the data. It seems unlikely that a cooler Cretaceous is an explanation of the model calculation. There exist several potential climatic forcing factors which might explain warm climates but each of these is speculative. For these two reasons, the remaining sensitivity experiments will consider possible feedback mechanisms which are not incorporated in the model. These experiments will be performed by prescribing a change in an *internal* condition which may be plausible.

Vegetation–climate feedbacks. A feedback between climatic warmth and humidity and increased surface vegetation with decreased surface albedo is plausible. The extensive distribution of Cretaceous floras with a "tropical" character may imply a lower land albedo than at present. A sensitivity experiment is performed by specifying a lower land albedo of 0.12 at latitudes equatorward of 45° in paleolatitude. Compared to the control the total absorbed solar energy increased by 1.6% and the global surface temperature increased by 2.2°K. The temperature increase with respect to latitude was relatively uniform (Fig. 5) despite the fact that the decrease in land albedo was not uniform with respect to latitude. This important point will be discussed later.

Cloud–climate feedbacks. In this model the global mean effect of changes in cloud amount on albedo is larger than the effect on outgoing infrared radiation. Thus, uniformly decreasing global cloud cover tends to increase the global surface temperature.

Table I 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 mid-to-high latitude surface temperature gradient. Since the temperatures and continental distribution during the Cretaceous were apparently much more uniform in latitude than the present day, it is plausible that f_c was also more uniform. Recent studies (Schneider et al., 1978; Roads, 1978) have also indicated that cloud cover amount tends to decrease, at least in midlatitudes as temperature increases.

Although this feedback process between climate and clouds cannot as yet be verified, we can decrease the cloud fraction in our model to test the sensitivity of the model to a plausible change. The present-day globally averaged cloud cover from Table I is 0.55. By setting f_c to 0.50 at all latitudes the global average of f_c is decreased slightly while simultaneously making its distribution uniform with latitude.

Fig. 5 (dash–dot curve) 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 0.5 or less in the tropics. The increased extratropical warming improves the agreement with the "minimum" Cretaceous temperature distribution, although the high latitudes are still much too cold, and the tropics are too warm. The globally averaged surface temperature is 3.6° warmer than the present day, more than double that resulting from changes in geography alone.

The temperature difference between the cloud tops and the surface, as well as the cloud fraction, determines the outgoing infrared radiation. Since Cretaceous temperatures were more globally uniform, it may also be plausible that the Cretaceous distribution of $T_s - T_c$ was more uniform than at present. A hypothetical Cretaceous distribution of $T_s - T_c$ can be derived by assuming that $T_s - T_c$ has a functional dependence on surface temperature as defined by the present-day values in Table I. Cretaceous values are assigned based on the "minimal" estimate of Cretaceous paleotemperatures. Again the validity of the assumption is unknown. Results of experiments using uniform and present-day cloud cover distributions are shown in Fig. 5. The change results in a significant extratropical warming compared to the other simulations. The globally averaged surface temperature increase for the case of uniform cloud fraction and modified $T_s - T_c$ is 5° warmer than the present-day control simulation.

From the cloudiness sensitivity experiments it can be concluded that if changes of this order did occur they would have contributed significantly to a Cretaceous-like temperature distribution. The increase in globally averaged surface temperature in these experiments is close to that required for the "minimal" temperature estimate. The problem is that the poles are too cold and the tropics too warm. This implies that *if* the model heat transports were larger, at least the "minimal" Cretaceous temperatures might be achieved with changes in internal or external conditions on the order of the geographic and cloudiness modifications.

Meridional heat transports. Despite the fact that land albedo, the fraction of ocean, and cloud parameters have been modified non-uniformly with latitude in the Cretaceous experiments, the diffusive heat transport parameterization of the model 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. In addition, the total poleward heat transport decreases as the surface temperature gradient decreases. 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). However, latent heat transport does not fully compensate for the decrease in sensible and potential energy transport. The distribution of temperatures, estimated from the paleoclimatic data, will therefore be very difficult to simulate using the diffusive heat transport parameterization. Another problem is that the oceanic heat transport has been fixed at present-day values.

Rather than attempt to justify specific modifications to the heat transport parameterization, Barron et al. (1981a) calculated the total heat transport *required* to achieve the minimum and maximum estimate of Cretaceous paleotemperatures. It is then possible to consider whether the required changes are reasonable in terms of known paleogeography and physical processes of atmospheric and oceanic circulation.

Starting from any Cretaceous experiment (i.e., those on Fig. 5) we wish to determine the incremental energy input (in Wm^{-2}) at each latitude that will result in either the minimum or maximum estimates of the Cretaceous meridional temperature distribution. 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.

The incremental energy input consists of two components. First, if the amount of energy available from changes in geography or cloud cover is insufficient to achieve a desired increase in global temperature, then the model adds additional energy which is spurious. This additional energy is an estimate of the amount of energy required to achieve the planetary temperature for a specified temperature distribution. It may be considered an estimate of the magnitude of a necessary external forcing factor, a necessary climatic feedback mechanism, or as the magnitude of the error in hypothesized paleotemperatures. Second, a portion of the incremental energy input arises solely because of changes in heat transport. These two components can be easily separated, because the meridional heat transport must be zero when integrated over the surface of the earth.

Barron et al. (1981a) derived several important conclusions by performing the above calculations. First, in the Cretaceous case with modified cloud fraction and modified $T_s - T_c$ no excess (spurious) energy is required to achieve the "minimal" Cretaceous temperature distribution, only a change in heat transport. Remarkably, the total heat transport convergence required is very similar to the present-day control heat transport convergence. This empirically derived result suggests that if the present-day heat transport can be maintained, despite the decrease in meridional temperature gradient, then the "minimum" estimate of Cretaceous temperatures can be achieved with the energy balance climate model. This suggests that non-diffusive heat transport mechanisms may have been important.

However, the energy required to achieve the "warmest" estimate is two to three times greater than the amount generated by changes in geography and clouds. If the Cretaceous temperatures were warmer than the minimal estimate, or if cloud changes are not included, then the model is incapable of simulating the hypothesized Cretaceous planetary temperatures. In addition 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.

CONCLUSIONS

The purpose of this review is to define the nature of the problem of warm, equable climates. The Cretaceous Period was chosen specifically for examination because of the availability of constaining data and because it is a large contrast from the present day. Defining the nature of the Cretaceous climate encompasses three fundamental problems: (1) determination of the climatic state during the Cretaceous; (2) determination of the mechanisms of climatic change on geologic time scales; and (3) determination of the climatic response to any modifying influence. Each of these components of the problem may be defined only within certain limits.

(1) The Cretaceous climatic state can be defined within two limits: (a) a "minimally" ice-free Cretaceous which has equatorial temperatures similar to the present day and a mean annual surface temperature at the pole near 0°C; (b) a "maximum" Cretaceous with warmer equatorial temperatures (3–5°C warmer than at present) and polar temperatures of approximately 15°C. The globally averaged surface temperature for these two cases is 6° and 14°C warmer than at present, respectively. The "minimum" estimate has an equator-to-pole temperature contrast of 26° while the "maximum" estimate has a contrast of only 17°. The present-day equator-to-pole surface temperature contrast is 41°. This definition of the Cretaceous climatic state considers only one climatic variable, the distribution of mean annual surface temperature with respect to latitude. A more comprehensive definition would require the use of proxy data to estimate a variety of climatic variables.

(2) A large number of external climatic forcing factors may have been important in explaining the warm equable Cretaceous and the climatic record over the last 100–200 m.y. However, only the surface of the earth can be well specified. The significance of potentially important factors such as CO_2 are speculative. It is uncertain whether changes in climate may be explained primarily by the changes in geography or whether other forcing factors may be required.

(3) The insight which can be gained from climate models is dependent on the parameterization of physical processes and the extent to which various

feedback processes are included in the models. The capabilities of present models can only be verified against observations of present-day climate. If Cretaceous paleotemperatures could be precisely determined, and if the climatic forcing factors could be well specified, then a quantitative understanding of the various interactions and feedback mechanisms might be achieved through model simulations. Because of the limitations described in points one and two and the limitations of climate models, a comprehensive Cretaceous simulation cannot be achieved. Each of the limitations is of critical importance in understanding warm, equable climates. We cannot eliminate any one factor in explaining the discrepancy between model calculations and Cretaceous temperatures. The discrepancy may be the result of errors in (a) the structure of the model or its parameterizations, (b) the interpretation of paleoclimatic data, or (c) the specification of external climatic forcing factors.

(4) Despite these limitations two problems can be addressed with model sensitivity studies: (a) investigation of the mechanisms which will increase the global temperature above its present-day value, and (b) investigation of the mechanisms which might explain the decrease in equator-to-pole surface temperature gradient compared to the present day. The "minimum" estimate of Cretaceous surface temperatures can be explained with known mechanisms of plausible magnitude. Geography is an important external climatic forcing factor, but other internal mechanisms such as cloud–climate feedbacks are required to explain even the minimum estimate of Cretaceous globally averaged surface temperature. The warmest estimate of Cretaceous temperatures presents us with a more difficult problem. This estimate requires between two and three times the increase in absorbed energy which can be achieved by changes in geography and plausible changes in clouds. This would require substantial changes in other forcing factors, for instance, a 600–800% increase in atmospheric CO_2. Finally, in order to achieve the distribution of Cretaceous temperatures with respect to latitude, we must seek mechanisms which can maintain the heat transport close to present-day values despite the decrease in equator-to-pole surface temperature gradient.

ACKNOWLEDGEMENTS

This research benefited greatly from the comments and discussions with Eric Saltzman, Stephen Schneider, Starley Thompson, William Hay and C.G.A. Harrison. The National Center for Atmospheric Research is sponsored by the National Science Foundation.

REFERENCES

Baes, C., Goeller, H., Olson, J. and Rotty, R., 1977. Carbon dioxide and climate: the uncontrolled experiment. Am. Sci., 65: 310–320.
Bahcall, J. and Shaviv, G., 1968. Solar models and neutrino fluxes. Astrophys. J., 153: 113–125.
Barnard, P., 1973. Mesozoic floras. In: N. Hughes (Editor), Organisms and Continents Through Time. Spec. Pap. Paleontol., 12: 175–188.
Barron, E.J., Sloan, J.L. and Harrison, C.G.A., 1980. Potential significance of land–sea distribution and surface albedo variations as a climatic forcing factor; 180 M.Y. to the present. Palaeogeogr. Palaeoclimatol. Palaeoecol., 30: 17–40.
Barron, E.J., Thompson, S.L. and Schneider, S.H., 1981a. An ice-free Cretaceous? Results from climate model simulations. Science, 212: 501–508.
Barron, E.J., Harrison, C.G.A., Sloan, J.L. and Hay, W.W., 1981b. Paleogeography, 180 million years ago to the present. Ecologae Geol. Helv., 74: 443–470.
Beaty, C., 1978. The causes of glaciation. Am. Sci., 66: 452–459.
Begelman, M. and Rees, M., 1976. Can cosmic clouds cause climatic catastrophes. Nature, 261: 298–299.
Berger, A., 1976. Obliquity and precession for the last 5,000,000 years. Astron. Astrophys., 51: 127–135.
Berlin, T., Naydin, D., Saks, V., Teis, R. and Khabadov, A., 1967. Jurassic and Cretaceous climate in northern USSR, from paleotemperature determinations. Int. Geol. Rev., 9: 1080–1092.
Bowen, R., 1961a. Oxygen isotope paleotemperature analyses of Mesozoic Belemnoidea from Europe, India and Japan. J. Paleontol., 36: 1077–1084.
Bowen, R., 1961b. Paleotemperature analyses of Mesozoic Belemnoidea from Australia and New Guinea. Geol. Soc. Am. Bull., 72: 769–774.
Bowen, R., 1964. Faunal realms in Jurassic and Cretaceous Belemnites. Geol. Mag., 101: 374–376.
Bowen, R. and Fontes, J.C., 1963. Paléotempératures indiquées par l'analyse isotopique de fossiles du crétacé inférieur des Hautes-Alpes (FRANCE). Experientia, 19: 268–275.
Brass, G., Saltzman, E., Sloan, J., Southam, J., Hay, W., Holzer, W. and Peterson, W., 1981. Ocean Circulation, Plate Tectonics and Climate, Climate in Earth History. National Academy Press, Washington, D.C., pp. 83–89.
Briden, J.C., Drewry, D.E. and Smith, A.G., 1974. Phanerozoic equal area world maps. J. Geol., 82: 531–554.
Broecker, W. and Takahashi, T., 1977. Neutralization of fossil fuel CO_2 by marine calcium carbonate. In: N. Anderson and A. Malahoff (Editors), The Fate of Fossil Fuel CO_2 in the Oceans. Plenum Press, New York, N.Y., pp. 213–242.
Budyko, M., 1969. The effect of solar radiation variations on the climate of the earth. Tellus, 21: 611–619.
Budyko, M. and Ronov, A., 1979. Chemical evolution of the atmosphere in the Phanerozoic. Geokhimiya, 5: 643–653.
Cess, R., 1976. Climatic change: an appraisal of atmospheric feedback mechanisms employing zonal climatology. J. Atmos. Sci., 33: 1831–1843.
Cogley, J., 1979. Albedo contrast of glaciation due to continental drift. Nature, 279: 712–713.
Colbert, E., 1964. Climatic zonation and terrestrial faunas. In: A. Nairn (Editor), Problems in Palaeoclimatology. Interscience, New York, N.Y., pp. 617–638.

Craddock, C., Bastien, T. and Rutford, R., 1964. Geology of the Jones Mountain areas. In: R. Adie (Editor), Antarctic Geology. North Holland Publ., Amsterdam, pp. 171–187.

Craig, H., 1965. The measurement of oxygen isotope paleotemperatures. In: Stable Isotopes in Oceanographic Studies and Paleotemperatures. Consiglio Naxionale delle Ricerche, Laboratorio di Geologica Nucleare, Pisa, pp. 1–24.

Crowell, J. and Frakes, L.A., 1970. Phanerozoic glaciation and the causes of the ice ages. Am. J. Sci., 268: 193–224.

Damon, P., 1968. The relationship between terrestrial factors and climate. Meteorol. Monogr., 8: 106–111.

Dennison, B. and Mansfield, V., 1976. Glaciations and dense interstellar clouds. Nature, 261: 32–34.

Donn, W. and Shaw, D., 1977. Model of climate evolution based on continental drift and polar wandering. Geol. Soc. Am. Bull., 88: 390–396.

Douglas, R.G. and Savin, S.M., 1971. Isotopic analyses of planktonic Foraminifera from the Cenozoic of the Northwestern Pacific, Leg. 6. In: A.G. Fischer et al., Initial Reports of the Deep Sea Drilling Project, 6. U.S. Government Printing Office, Washington, D.C., pp. 1123–1127.

Douglas, R.G. and Savin, S.M., 1973. Oxygen and carbon isotope analyses of Cretaceous and Tertiary Foraminifera from the Central North Pacific. In: E.L. Winterer, W.R. Riedel et al., Intial Reports of the Deep Sea Drilling Project, 17. U.S. Government Printing Office, Washington, D.C., pp. 591–605.

Douglas, R.G. and Savin, S.M., 1975. Oxygen and carbon isotope analyses of Tertiary and Cretaceous microfossils from Shatsky Rise and other sites in the North Pacific Ocean. In: R.L. Larson, R. Moberly et al., Initial Reports of the Deep Sea Drilling Project, 32, U.S. Government Printing Office, Washington, D.C., pp. 509–520.

Ellis, J., 1978. Cloudiness, the Planetary Radiation Budget and Climate. Ph.D. thesis, Colorado State University, Fort Collins, Colo.

Emiliani, C., 1954. Temperatures of Pacific bottom waters and polar superficial waters during the Tertiary. Science, 112: 853–855.

Emiliani, C., 1961. The temperature decrease of surface water in high latitude abyssal-nadal water in open ocean basins during the past 75 million years. Deep-Sea Res., 8: 144–147.

Eriksson, E., 1963. Possible fluctuations in atmospheric carbon dioxide due to changes in the properties of the sea. J. Geophys. Res., 68: 3871–3876.

Fairbridge, R., 1973. Glaciation and plate migration. In: D. Tarling and S. Runcorn (Editors), Implications of Continental Drift to the Earth Sciences. Academic Press, New York, N.Y., pp. 503–515.

Firstbrook, P.L., Funell, B.M., Hurley, A.M. and Smith, A.G., 1979. Paleoceanic Reconstructions 160-0 Ma, Deep Sea Drilling Project. La Jolla, Calif. pp. 1–41.

Frakes, L.A., 1979. Climates Throughout Geologic Time. Elsevier, Amsterdam, pp. 1–310.

Frakes, L.A. and Kemp, E., 1972. Influence of continental positions on Early Tertiary climates. Nature, 240: 97–100.

Frakes, L.A. and Kemp, E., 1973. Paleogene continental positions and evolution of climate. In: D. Tarling and S. Runcorn (Editors), Implications of Continental Drift to the Earth Sciences. Academic Press, New York, N.Y., pp. 535–558.

Gal-Chen, T. and Schneider, S.H., 1976. Energy balance climate modeling: comparison of radiative and dynamic feedback mechanisms. Tellus, 28: 108–121.

Geitzenaur, K., Margolis, S. and Edwards, D., 1968. Evidence consistent with Eocene glaciation in a South Pacific deep sea sedimentary core. Earth Planet. Sci. Letters, 4: 173–177.

Gough, D., 1977. Theoretical predictions of variations in solar output. In: O. White (Editor), The Solar Output and Its Variation. Colorado Assoc. Univ. Press, Boulder, Colo., pp. 451–474.

Habicht, J.K.A., 1979. Paleoclimate, paleomagnetism and continental drift. Amer. Petrol. Geol., Stud. Geol., 9: 1–31.

Hamilton, W., 1968. Cenozoic climatic change and its cause. Meteorol. Monograph, 8: 128–133.

Hartmann, D. and Short, D., 1980. On the use of earth radiation budget statistics for studies of clouds and climate. J. Atmos. Sci., 6: 1233–1250.

Haselgrove, C. and Hoyle, F., 1959. Main sequence stars. Mon. Not. R. Astron. Soc., 119: 112–123.

Hays, J. and Pitman, W., 1973. Lithospheric plate motion, sea level changes, and climatic and ecological consequences. Nature, 246: 18–22.

Hays, J., Imbrie, J. and Shackleton, N., 1976. Variations in the Earth's orbit: pacemaker of the ice ages. Science, 194: 1121–1132.

Hein, J., Scholl, D. and Miller, J., 1978. Episodes of Aleutian Ridge explosive volcanism. Science, 199: 137–141.

Hughes, N., 1964. Cretaceous floras and the assessment of past climates. In: A. Nairn (Editor), Problems in Palaeoclimatology. Interscience, New York, N.Y., pp. 44–45.

Hughes, N., 1973. Mesozoic and Tertiary distributions and problems of land plant evolution. In: N. Hughes (Editor), Organisms and Continents through Time. Spec. Pap. Palaeontol., 12: 189–198.

Irving, E., North, F.K. and Covillard, R., 1974. Oil, climate and tectonics. Can. J. Earth Sci., 11: 1–17.

Jacobowitz, H., Smith, W., Howell, H., Nagle, F. and Hickey, J., 1979. The first 18 months of planetary radiation budget measurements from Nimbus 6 ERB experiment. J. Atmos. Sci., 36: 501–507.

Kauffman, E.G., 1973. Cretaceous Bivalvia. In: A. Hallam (Editor), Atlas of Palaeobiogeography. Elsevier, Amsterdam, pp. 353–383.

Kennett, J.P., 1977. Cenozoic evolution of Antarctic glaciation, the circum-Antarctic Ocean, and their impact on global paleoceanography. J. Geophys. Res., 82: 3843–3860.

Kennett, J.P. and Thunell, R., 1975. Global increase in Quaternary explosive volcanism. Science, 187: 497–503.

Krassilov, V.A., 1973. Climatic changes in eastern Asia as indicated by fossil floras, I. Early Cretaceous. Palaeogeogr., Palaeoclimatol. Palaeoecol., 13: 261–273.

Krassilov, V.A., 1981. Changes of Mesozoic vegetation and the extinction of dinosaurs. Palaeogeogr., Palaeoclimatol. Palaeoecol., 34: 207–224.

LeMasurier, W., 1970. Volcanic evidence for Early Tertiary glaciation in Marie Byrd Land. Antarct. J., 5: 154–155.

Lowenstam, H., 1964. Palaeotemperatures of the Permian and Cretaceous Periods. In: A. Nairn (Editor), Problems in Palaeoclimatology. Interscience, New York, N.Y., pp. 227–248.

Lowenstam, H. and Epstein, S., 1954. Paleotemperatures of the post-Aptian Cretaceous as determined by the oxygen isotope method. J. Geol., 62: 207–248.

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

Margolis, S. and Kennett, J.P., 1971. Cenozoic paleoglacial history of Antarctica recorded in subantarctic deep-sea cores. Am. J. Sci., 271: 1–36.

McCrea, W., 1975. Ice ages and the galaxy. Nature, 255: 607–609.

McElhinny, M., 1973. Palaeomagnetism and Plate Tectonics. Cambridge University Press, Cambridge, pp. 1–357.

McKenna, M.C., 1980. Eocene palaeolatitude, climate, and mammals of Ellesmere Island. Palaeogeogr., Palaeoclimatol. Palaeoecol., 30: 349–362.

Newell, R.E., Kidson, J.W., Vincent, D.G. and Boer, G.J., 1972. The General Circulation of the Tropical Atmosphere and Interactions with Extratropical Latitudes, Vol. 1. M.I.T. Press, Cambridge, Mass., pp. 1–258.

Ninkovitch, D. and Donn, W., 1976. Explosive Cenozoic volcanism and climatic implications. Science, 194: 899–906.

Oort, A. and Vonder Haar, T., 1976. On the observed annual cycle in the ocean–atmosphere heat balance over the Northern Hemisphere. J. Phys. Oceanogr., 6: 781–800.

Pitman, W., 1978. Relationship between eustasy and stratigraphic sequences of passive margins. Geol. Soc. Am. Bull., 89: 1389–1403.

Pollard, D., Ingersoll, A.P. and Lockwood, J.G., 1980. Response of a zonal climate–ice sheet model to the orbital perturbations during the Quaternary ice ages. Tellus, 32: 301–319.

Ramanathan, V., 1977. Interactions between ice-albedo, lapse rate, and cloud-top feedbacks: an analysis of the nonlinear response of a GCM climate model. J. Atmos. Sci., 34: 1885–1897.

Roads, J., 1978. Numerical experiments on the climate sensitivity of an atmospheric hydrologic cycle. J. Atmos. Sci., 35: 753–773.

Savin, S., 1977. The history of the earth's surface temperature during the past 100 million years. Ann. Rev. Earth Planet. Sci., 5: 319–355.

Savin, S., Douglas, R. and Stehli, F., 1975. Tertiary marine paleotemperatures. Geol. Soc. Am. Bull., 86: 1499–1510.

Schneider, S.H. and Dickinson, R.E., 1974. Climate modeling. Rev. Geophys. Space Phys., 12: 447–493.

Schneider, S.H. and Gal-Chen, T., 1973. Numerical experiments in climate stability. J. Geophys. Res., 78: 6182–6194.

Schneider, S.H., and Thompson, S.L., 1980. Cosmic conclusions from climate models: can they be justified? Icarus, 41: 456–469.

Schneider, S.H., Washington, W. and Chervin, R., 1978. Cloudiness as a climatic feedback mechanism: effects on cloud amounts of prescribed global and regional surface changes in the NCAR GMC. J. Atmos. Sci., 35: 2207–2221.

Schwarzbach, M., 1963. Climates of the Past. Van Nostrand, Princeton, N.J., pp. 1–328.

Schwarzchild, M., Howard, R. and Harm, R., 1957. Inhomogeneous stellar models, V. A solar model with convective envelop and inhomogeneous interior. Astrophys. J., 125: 133–241.

Sclater, J., Hellinger, S. and Tapscott, C., 1977. The paleobathymetry of the Atlantic Ocean from the Jurassic to the present. J. Geol., 85: 509–552.

Sellers, A. and Meadows, A., 1975. Long term variations in the albedo and surface temperature of the Earth. Nature, 254: 44.

Shackleton, N. and Kennett, J., 1975a. Paleotemperature history of the Cenozoic and the initiation of Antarctic glaciation: oxygen and carbon isotope analyses in DSDP sites 227, 279, and 281. In: J.P. Kennett, R.E. Houtz et al., Initial Reports of the Deep Sea Drilling Project, 29. U.S. Government Printing Office, Washington, D.C., pp. 743–755.

Shackleton, N. and Kennett, J., 1975b. Late Cenozoic oxygen and carbon isotopic changes at DSDP site 284: implications for glacial history of the Northern Hemisphere. In: J.P. Kennett, R.E. Houtz et al. Initial Reports of the Deep Sea Drilling Project, 29. U.S. Government Printing Office, Washington, D.C., pp. 801–807.

Smiley, C., 1967. Paleoclimatic interpretations of some Mesozoic floral sequences. Am. Assoc. Petrol. Geol. Bull., 51: 849–863.

Smith, A.G. and Briden, J.C., 1977. Mesozoic and Cenozoic Paleocontinental Maps. Cambridge Univ. Press, Cambridge, pp. 1–63.

Sohl, N.F., 1969. North American Cretaceous biotic provinces delineated by gastropods. In: Proceedings North American Paleontologic Convention II. Allen Press, Lawrence, Kansas, pp. 1610–637.
Spaeth, C., Hoefs, J. and Vetter, U., 1971. Some aspects of isotopic composition of Belemnites and related paleotemperatures. Geol. Soc. Am. Bull., 82: 3139–3150.
Steiner, J. and Grillmair, E., 1973. Possible galactic causes for periodic and episodic glaciations. Geol. Soc. Am. Bull., 84: 1003–1018.
Stevens, G. and Clayton, R., 1971. Oxygen isotope studies of Jurassic and Cretaceous Belemnites from New Zealand and their biogeographic significance. N.Z.J. Geol. Geophys., 14: 829–897.
Talbot, R., Butler, D. and Newman, M., 1976. Climatic effects during the passage of the solar system through interstellar clouds. Nature, 262: 561–563.
Tanner, W., 1968. Cause and development of an ice age. Meteorol. Monogr., 8: 126–127.
Tarling, D.H., 1978. The geological-geophysical framework of ice ages. In: J. Gribben (Editor), Climatic Change. Cambridge Univ. Press, Cambridge, pp. 3–24.
Taylor, B.J., 1972. Stratigraphical correlation in southeast Alexander Island. In: R. Adie (Editor), Antarctic Geology and Geophysics. Universitetsforlaget, Oslo, pp. 149–153.
Teis, R., Chupakhin, M. and Naidin, D., 1957. Determination of paleotemperatures from the isotopic composition of oxygen in calcite of certain fossil shells from Crimea. Geochemistry, 4: 323–329.
Thompson, S.L. and Barron, E.J., 1981. Comparison of Cretaceous and present earth albedos: implications for the causes of paleoclimates. J. Geol., 89: 143–167.
Thompson, S.L. and Schneider, S.H., 1979. A seasonal zonal energy balance climate model with an interactive lower layer. J. Geophys. Res., 84: 2401–2414.
Umbgrove, J., 1947. The Pulse of the Earth. Martinus Nijhoff, The Hague, pp. 1–358.
Vakhrameev, V., 1975. Main features of phytogeography of the globe in Jurassic and Early Cretaceous time. Paleontol. J., 2: 123–133.
Van Woerkom, A., 1953. The astronomical theory of climatic changes. In: H. Shapley (Editor), Climatic Change. Harvard Univ. Press, pp. 147–157.
Vaughn, T. and Wells, J., 1943. Revision of the suborders, families and genera of the Scleractinia. Geol. Soc. Am. Spec. Pap., 44: 1–363.
Von Arx, W., 1957. An experimental approach to problems in physical oceanography. In: Progress in Physics and Chemistry of the Earth. Pergamon Press, New York, N.Y., 2: 1–29.
Walker, J.C.G., Hays, P.B. and Kasting, J.F., 1981. A negative feedback mechanism for the long-term stabilization of Earth's surface temperature. J. Geophys. Res., 86: 9776–9782.
Warren, S.G. and Schneider, S.H., 1979. Seasonal simulation as a test for uncertainties in the parameterizations of a Budyko-Sellers zonal climate model. J. Atmos. Sci., 36: 1377–1391.
Wigley, T.,M.L., 1981. Climate and paleoclimate: what can we learn about solar luminosity variations? Solar Phys., 74: 435–471.
Williams, G., 1975. Possible relation between periodic glaciation and flexure of the galaxy. Earth Planet. Sci. Lett., 26: 361–369.
Wilson, J., 1975. Carbonate Facies in Geologic History. Springer-Verlag, New York, N.Y., pp. 1–471.
Wolfe, J., 1978. A paleobotanical interpretation of Tertiary climates in the Northern Hemisphere. Am. Sci., 66: 691–703.

[Received June 8, 1982; accepted November 5, 1982]