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Résumé de l'article

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At present, models are not well utilized to reproduce a global climatology for a particular time period. Rather, the models are used for sensitivity experiments. One geologic factor is varied {e.g., geography or C02 level) while all others are held constant. In this way, the effect on the climate of changing that one factor can be tested, and insights into the mechanisms of global change are gained.The results are then compared with the geologic record. Two case studies are given asexamples.

A Geological Perspective on Climatic Change: Computer Simulation of Ancient Climates

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SUMMARY

The possibility of global warming resulting from the anthropogenic addition of carbon dioxide and other industrial gases into the atmosphere is a topic of much recent concern. Global climate models are used to make predictions for possible future climatic changes, but their ability to simulate climatic states other than the present day is not well constrained. The sedimentary record contains a wealth of information on past global climatic change. From this, we can get a good idea of what past climatic extremes were, and some idea of their underlying causes. If we attempt to simulate the climate of a past time period, we can evaluate the ability of a climate model to represent a climatic state substantially different from the present day.

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INTRODUCTION

The influence of climate on many global systems is enormous, and the possibility of future global climate change has profound implications for agriculture, plant and animal ecosystems, water supplies, weather patterns and sea-level rise. Modification of climate by our industrial society *via* the release of so-called greenhouse gases (*e.g.*, CO_2, CH_4) into the atmosphere is a very real possibility, but the consequences of this modification are not well understood. A general hypothesis is that over the next century, a doubling of atmospheric CO_2 will cause a rise in global average temperature of 2°-4°C (Schlesinger and Mitchell, 1987).

These predictions are based on the results of numerical climate models which attempt to quantify the change in climate as a result of increasing atmospheric CO_2. While these global climate models do a good job of simulating large-scale features of present-day climates, their ability to simulate accurately climatic change is not well constrained. Thus, to gain a perspective on possible future climatic change, we study past climates. Specifically, the task is to determine whether climate models can accurately predict the important elements of ancient climate which were substantially different from present-day climates. This is done by comparing the model output to ancient climates reconstructed from climatic indicators in the sedimentary record, when forcing factors are known. In this way, we can test the ability of climate models to simulate accurately climates other than present day.

MODERN CLIMATOLOGY AND PALEOCLIMATOLOGY

Modern climatology is concerned with the dynamics of, and the interactions among, the atmosphere, the hydrosphere and the cryosphere; yet, with the exception of the seasonal cycle, it lacks any real sense of climate change. Paleoclimatology, however, has as its basis a tremendous sense of climatic change, and thus offers the best opportunity to study its causes and effects. Two themes arise from the study of ancient climates: one is the long-term stability of the Earth's climate system (*i.e.*, continued presence of liquid water on the Earth's surface) in the face of significant perturbations to that system (for example, the "faint young sun paradox", *see* Kasting, 1989). The other is shorter termed (thousands to tens of millions of years), lower-amplitude climatic oscillations, within the bounds of the long-term stability. These oscillations range from warm, equable conditions over most of the globe (*i.e.*, middle Cretaceous) to large glaciations covering significant portions of the globe (*i.e.*, Pleistocene). It is these shorter termed climatic variations that significantly influence a variety of global systems.

A qualitative diagram of climatic variability over the Phanerozoic is given in Figure 1. This diagram shows the inferred global mean temperature for past time periods as departures from the present global mean. Major features of interest include glaciations during the Late Ordovician, Carboniferous/Permian and the Pleistocene, and extremely warm time periods during the Triassic, the Cretaceous and the Eocene. It is important to note that the time scale on the diagram is not linear, but is progressively expanded in younger time units. This reflects the importance of stratigraphic resolution in the definition of paleoclimates. With increasing age, stratigraphic resolution becomes more diffuse and time stratigraphic units can represent periods of a few million years. Within these units, it is not possible to demonstrate areal synchroneity of events. Climate may be represented within these rocks as an average condition, but it may also contain recognizable shorter term changes and even apparently instantaneous events. Thus, the large difference in stratigraphic resolution between the Early Paleozoic and the Pleistocene results in large differences in the meaning of "climate" for these time periods.

SEDIMENTARY PALEOCLIMATIC INDICATORS

The traditional approach to paleoclimatology has been to interpret ancient climates from biologic, sedimentologic and chemical signatures preserved in the sedimentary record. For example, climatic zones can be determined from fossil floral and faunal communities by comparison with present-day analogues. However, this type of analysis becomes less precise for older time periods because of diminishing numbers of extant genera. It is very difficult to assign precise climatic zones to extinct taxa. Certain types of sedimentary deposits can be very diagnostic for climate types. Mixtites, dropstones in varves and striated pavements, when found together, are almost exclusively formed under glacial conditions (*e.g.*, Frakes, 1979) and bauxites and laterites are mainly formed in humid tropical climates. The formation of coals requires high amounts of precipitation under temperate or sub-tropical temperatures, while evaporites form under arid conditions.

Less obvious sedimentary features indicative of climate include sedimentary structures, such as hummocky cross-stratification which is produced by very large storms (hurricanes and severe winter storms; *e.g.*, Duke, 1985). The compositional maturity of fluvial sandstones expressed in terms of quartz, feldspar and rock fragment ratios can be used to determine climate if the effects of tectonics (relief) and source rock are known (*e.g.*, Suttner *et al.*, 1981). The biologic and sedimentologic categories of climatic indicators are complex, and are variable in their reliability. Climate is most often not expressed as specific temperatures or pressures, but in relative terms, such as hotter or colder, or wetter or drier. Chemical signatures, such as the oxygen isotope record in marine sediments, are capable of giving quantitative measures of paleotemperature for the Mesozoic and Cenozoic under certain conditions (Savin, 1977).

PALEOCLIMATIC RECONSTRUCTIONS FROM SEDIMENTARY PALEOCLIMATIC INDICATORS

A qualitative reconstruction of climate for a given time period can be obtained by compiling the above types of sedimentary paleoclimatic indicators onto a paleogeographic reconstruction. An example of this is shown in Figure 2 for the Cretaceous (Albian). Three types of paleoclimatic indicators only are shown for simplicity; bauxite deposits, the localities of fossil crocodiles and the extent of fossil coral reefs (data from summary by Habicht, 1979). The high latitude positions of the fossil crocodiles and coral reefs show that, during this time period, warm temperatures extended significantly further poleward than during the present day. A belt of high precipitation as shown by the presence of bauxite extended across the 40°N latitude line. This would have corresponded to the mid-latitude frontal zone of higher precipitation in the northern hemisphere.

These types of climatic reconstructions are able to show large differences in the global climatic states between different geologic time periods. For example, a distinct contrast exists between the Mesozoic, mostly a period of warmth and climatic equability, and the Neogene, a period characterized by glacial-interglacial fluctuations (*e.g.*, Frakes, 1979). This difference is very apparent when the distributions of climatic zones derived from vegetation patterns of the two time periods are compared (*see* Figure 3). However, these reconstructions do have a number of limitations. They are static in terms of the dynamics of the global

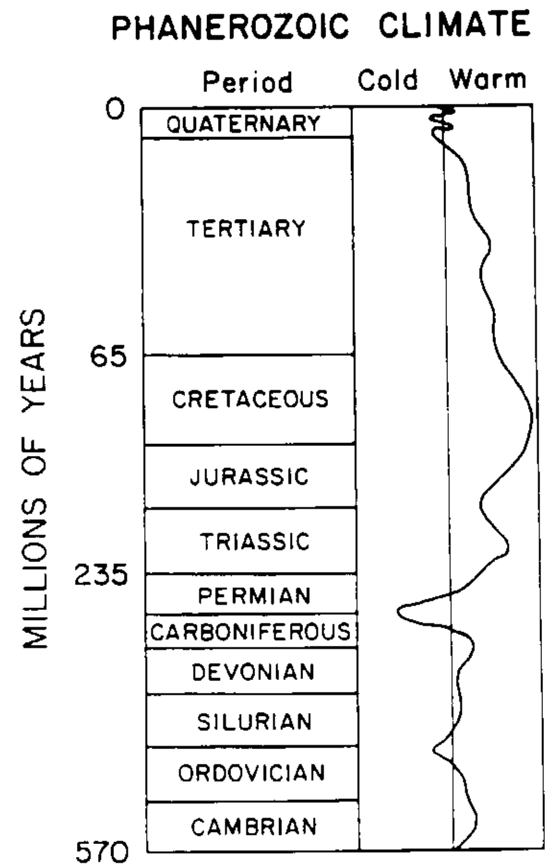

Figure 1 *Climatic variability over the Phanerozoic relative to present-day global mean climate. (Modified from Frakes, 1979).*

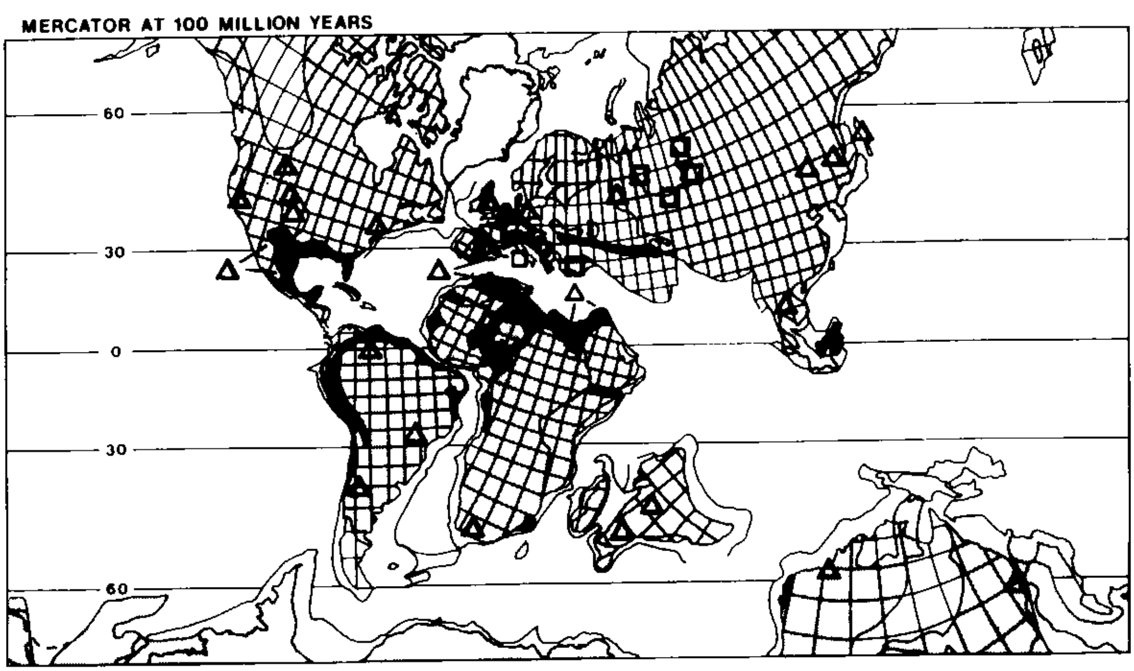

Figure 2 *A paleogeographic reconstruction for the Cretaceous (Albian). Legend: Bauxite deposits (boxes), fossil crocodiles (triangles), fossil coral reefs (black shading). Paleoclimatic data from Habicht (1979), paleogeographic reconstructions from Barron* et al. *(1981). See text for further discussion.*

climate system operating for past geologic time periods and they do not allow assessment of the relative contribution of various climate forcing factors to climatically unique configurations (as existed during past time periods).

Another approach is to assume that a given set of laws has governed the climatic system through geologic time and that these laws can be expressed in a simplified form as a series of mathematical equations — a numerical climate model.

GLOBAL CLIMATE MODELS

There are four broad classes of numerical climate models, ranging from relatively simple energy balance models and radiative-convective models, through more complex statistical dynamical models, to general circulation models (GCMs), which are the most complex and powerful climate models available today. An excellent review of each type of model is given by Meehl (1984). Most detailed paleoclimatic modeling is done with GCMs, because they have the highest spatial resolution of all of the global models and are best able to incorporate the very different conditions of past time periods such as geography and the presence or absence of ice sheets.

The following description of GCMs is condensed from Schneider and Dickinson (1974) and Washington and Parkinson (1984). The fundamental equations contained within general circulation models are derived from the basic laws of physics, particularly the conservation laws for momentum, mass and energy. An equation of state relates the pressure, density and temperature within the model atmosphere, and a moisture equation describes the atmospheric portion of the hydrologic cycle. A radiative transfer equation describes atmospheric heating and other radiative processes. The GCM solves each of these equations simultaneously to describe the evolution of the dynamic and thermodynamic state of the atmosphere. The scale at which the various parameters are calculated typically is a few degrees of latitude and longitude and a few kilometres vertically. (The National Center for Atmospheric Research (NCAR) Community Climate Model calculates these variables on a global grid of 40 latitudes (4.5°) by 48 longitudes (7.5°), with nine levels in the vertical. Case studies using this model will be given below.) All processes occurring at scales larger than this are calculated explicitly, while processes occurring at smaller scales (*e.g.*, cumulus convection, latent heating, cloud formation) must be parameterized.

Constraining the model climatology is a set of non-varying boundary conditions which includes features such as continental positions and land-sea distribution, topography, incoming solar radiation, atmospheric composition (*e.g.*, CO_2 levels), surface albedo and sea-surface temperatures. The model is run until it generates a set of climatic variables which are in a steady state with these boundary conditions. The climatic variables which are most important to the sedimentary record include atmospheric temperatures, precipitation, evaporation, atmospheric circulation including both surface winds and upper level winds, and oceanic boundary currents. The latter must be obtained from ocean circulation models.

An example of GCM output is given in Figure 4, which shows mean annual surface temperatures in degrees Kelvin for a realistic mid-Cretaceous geography and topography (from Barron and Washington, 1984). In this case, sea-surface temperatures were explicitly calculated by the model. Note that the surface temperatures are much warmer at high latitudes than for the present day. Compare this with Figure 2, which shows the distribution of warm temperature paleoindicators (crocodiles and coral reefs).

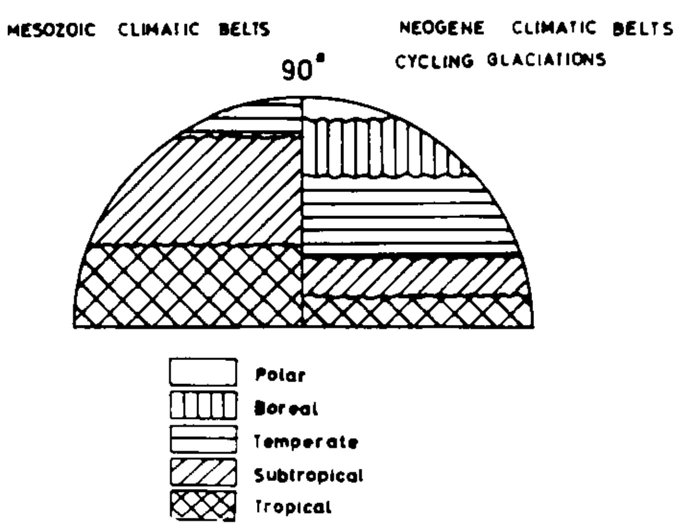

Figure 3 *A general climatic zonation for the Mesozoic and the Neogene from floral patterns. (After Pantic and Stefanovic, 1984).*

Paleoclimatic Modelling

The objective of paleoclimatic modelling is not to produce a full reconstruction of global paleoclimate for a given time period. There are too many uncertainties involved in the simplification of the complex climatic system, and the ability of the models to simulate climatic change is not completely verified. For example, it is not certain that all factors which cause climatic change are correctly specified in the models (*e.g.*, solar variability or atmospheric composition). For these reasons, the main goal of paleoclimatic modelling at present is to use the models to further our understanding of the climate system. When considering paleoclimates, the underlying assumption is that the basic physics of the climate system have not changed and so physically based models derived from the present-day climate system can be applied to past time periods. What have changed over geologic time, however, are many of the boundary conditions or climatic forcing factors (mentioned above) which constrain the climatic state. These act to change the climate system by altering the atmosphere's radiation balance or by altering the nature of the atmospheric or oceanic circulation. Therefore, we can postulate that the Cretaceous climate was very different from the present day because many of the boundary conditions were different, *e.g.*, a different land-sea distribution, no large continental ice sheets, higher sea level, possibly elevated levels of CO_2 and so on. The GCMs are used to determine the most important forcing factors contributing to past climates by means of sensitivity tests. One geologic factor (*e.g.*, geography or CO_2 level) is varied according to paleogeographic/geologic reconstructions for a specific time period while all other boundary conditions are held constant. The resulting model climatology can be compared with the paleoclimatology reconstructed from the sedimentary record for that time period. In this way, the effect of each geologic factor on the model climatology relative to a control case (*i.e.*, present day) can be assessed.

It is important to note that a change in climate can have an impact on its own controlling boundary conditions. A well-known example of this is the so-called ice-albedo feedback where the formation of an ice sheet favours its own continued growth. The presence of ice raises the regional albedo which causes a cooling. This in turn promotes more ice to form with its higher albedo and further cooling and so on. To be successful, climate models must incorporate this type of feedback.

CASE STUDIES

There are many paleoclimatic studies that have been done using climate models that are illustrative of this new approach, but only two are discussed here for space considerations; one for the Cretaceous and one for the very late Pleistocene. These time periods were chosen because they represent end-member climate types for the Phanerozoic, and therefore can illustrate the extent to which GCMs can simulate climatic extremes. As stated earlier, the middle Cretaceous was a time of global warmth and equability, while the Late Pleistocene was a time of glacial-interglacial fluctuations. These studies are summarized from Barron and Washington (1984, 1985) for the Cretaceous, and from COHMAP (Cooperative Holocene Mapping Project) (1988) for the Late Pleistocene.

The Cretaceous

Mid-Cretaceous (Albian) paleotemperatures have been reconstructed from geologic paleoclimatic data within two limits as shown in Figure 5 (Barron and Washington, 1984). Globally averaged surface temperatures apparently were at least 6°C higher than the present day. The greatest contrast in surface temperatures between the Albian and the present day is at mid- to high latitudes, hence the term "equable" for the Cretaceous climate. The mid-Cretaceous was also a large geographic contrast from the present day with very different continental positions and elevations, and a 20% reduction in land area due to high eustatic sea level.

Barron and Washington (1984) conducted a series of sensitivity experiments designed to estimate the role of geographic variables in explaining the warm Cretaceous climate. In particular, they attempted to estimate the surface temperature sensitivity to a change in geography by comparing present-day and mid-Cretaceous simulations. The geographic factors that were varied include high snow albedo, topography, continental positions and sea level. The model used in this study was the NCAR Community Climate Model which was described above.

The net effect of a Cretaceous geography simulation was to produce a globally averaged surface temperature warming of 4.8°C. As shown in Figure 6, the warming predicted in the model, in comparison with a present-day control, increases at higher latitudes and is hemispherically symmetrical. The individual sensitivity experiments indicate that the warming in the northern hemisphere is largely due to changes in continental position and the warming in the southern hemisphere is largely caused by the deglaciation (decreased elevation and removal of the specified high albedo) of Antarctica. They found that changes in topography and ice sheet albedo otherwise had only small regional effects on temperature sensitivity. The model sensitivity to sea-level change was found to be almost negligible — a surprising result given that warm paleoclimates are strongly associated with high eustatic sea level and reduced land area.

When these results were compared with the geologic record, it was obvious that the combined effect of these geographic variables was not enough to explain fully the Cretaceous warmth. A comparison of Figures 2 and 4 shows that the higher latitudes were too cool in the model results to match the geologic data (*i.e.*, the model mean annual temperatures over crocodile fossil localities at higher latitudes are at or just below freezing). Barron and Washington (1985) then considered an additional forcing factor, elevated levels of atmospheric carbon dioxide. The primary motivating factor for considering higher CO_2 levels was the work of Berner *et al.* (1983). Their calculations of atmospheric CO_2 over the Phanerozoic are based on the carbonate-silicate geochemical cycle, where the primary source of CO_2 to the atmosphere is volcanism at sea-floor spreading centres and the primary sink is the weathering of silicate minerals. The mid-Cretaceous was a time of rapid sea-floor spreading which would have resulted in a greater CO_2 flux to the atmosphere. It was also the main reason for higher eustatic sea level because of changes in the ocean basin volume (Pitman, 1978). The rate of weathering of silicate minerals, which is dependent on the area of continents, would have been reduced, leading to an increase in atmospheric CO_2.

Two additional climate model sensitivity studies were performed with the NCAR Community Climate Model by quadrupling the atmospheric CO_2 in both the present-day control and the Cretaceous geography experiments. It was found that a quadrupling was of the right order to place the Cretaceous globally averaged surface temperature within the two limits derived from geologic paleoclimate data (Figure 7).

The Late Pleistocene

Well-dated geologic paleoclimate data assembled by COHMAP (1988) show that substantial changes in the global climate have altered vegetation, ice-sheet volumes and sea-surface conditions. The nature and timing of these fluctuations have long been a

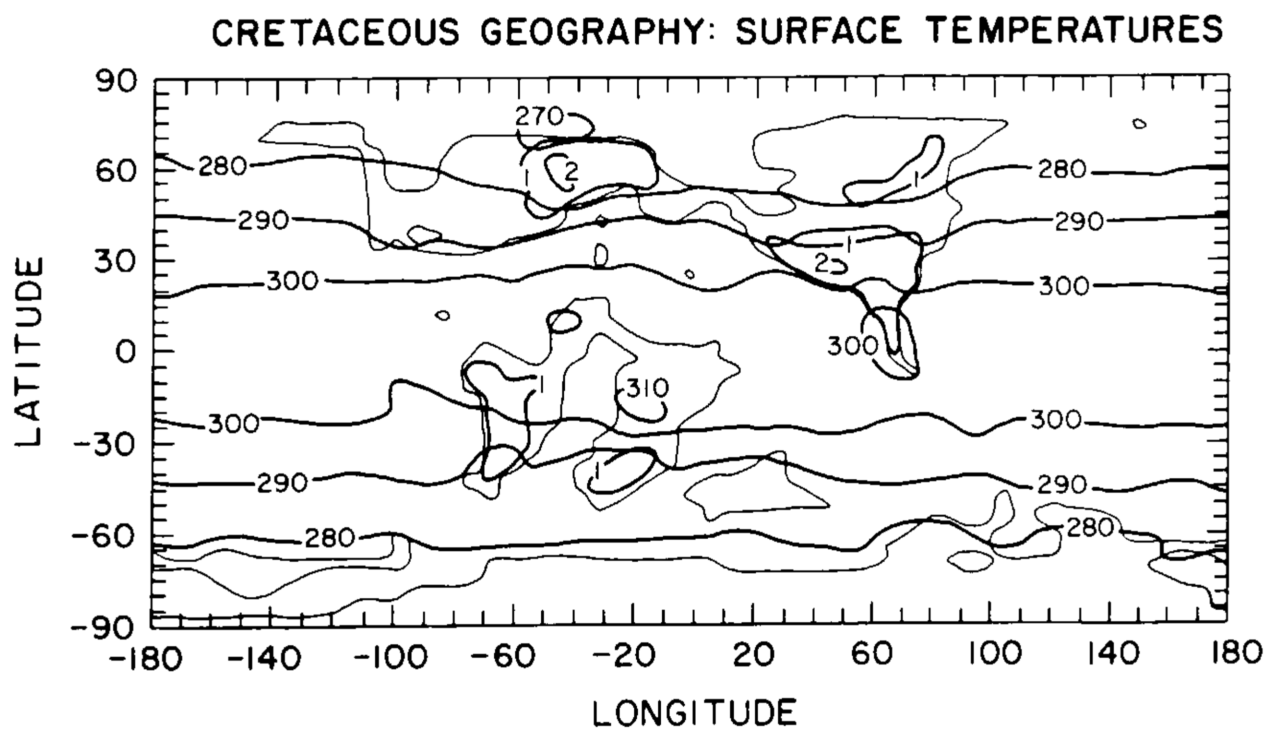

Figure 4 *The distribution of mean annual surface temperatures in degrees Kelvin (0°C = 273.15 K) for the mid-Cretaceous with a "realistic" geography. Topography contours are in kilometres. (From Barron and Washington, 1984).*

problem in Quaternary geology. One possible source for this variability is changes in solar insolation arising from changes in the Earth's orbital parameters (*i.e.*, precession, obliquity and eccentricity), the so-called Milankovitch cycles. A strong correspondence has been found by many workers between fluctuations in marine oxygen isotope records from deep-sea sediments, taken as a proxy for ice volume, and the periodicity of the orbital parameters (*e.g.*, Emiliani, 1955; Hays *et al.*, 1976; Imbrie *et al.*, 1984; Imbrie, 1985). It is thus reasonable that these orbital variations in solar insolation could be responsible for the range of climatic changes experienced since the Last Glacial Maximum.

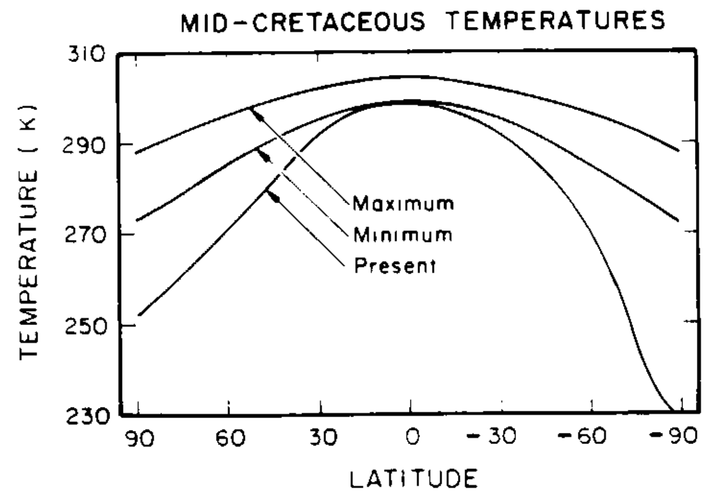

Figure 5 *Zonally averaged mean annual surface temperature limits for the mid-Cretaceous (Albian), derived from geologic paleoclimate data. (From Barron and Washington, 1984).*

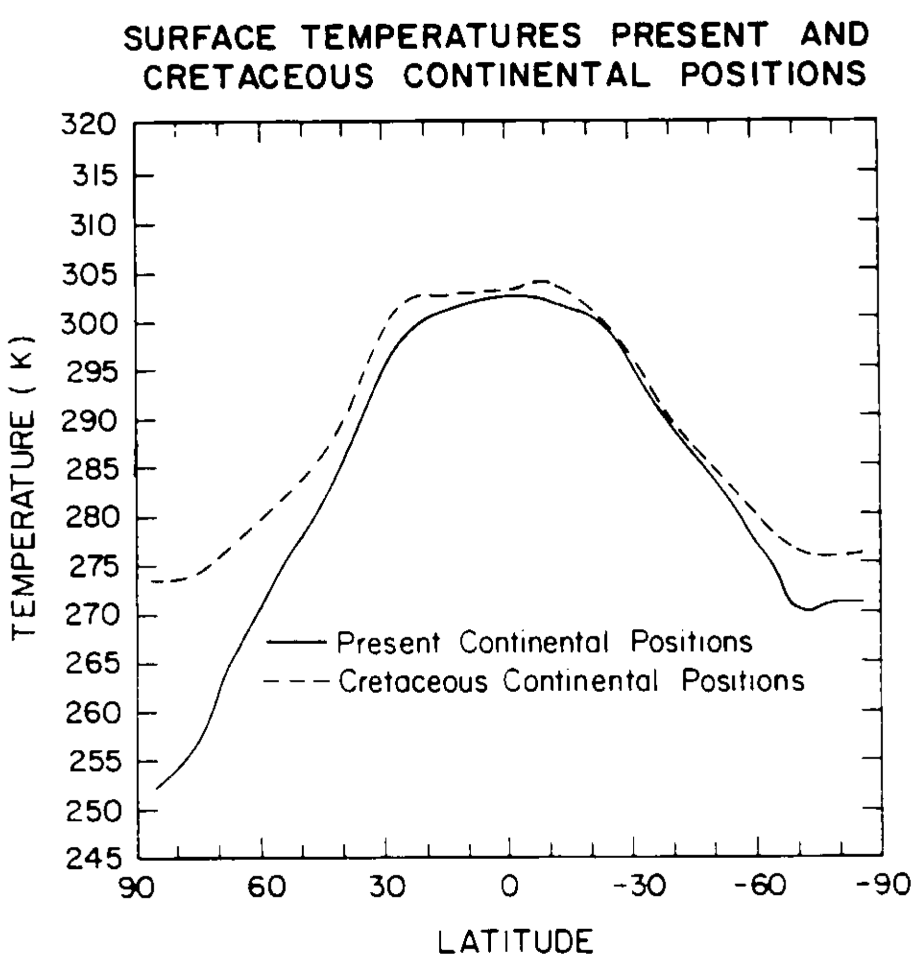

Figure 6 *A comparison of zonally averaged surface temperatures for a present-day control simulation and a Cretaceous geography sensititivy experiment. (From Barron and Washington, 1984).*

Some dispute has been raised, however, about the exact role of the orbital variations in driving climate change over this time period, particularly for the global ice sheets. Broecker and Denton (1989) cite the abrupt terminations of the glacial-interglacial cycles as recorded by the marine oxygen isotope record and the synchroneity of the growth and decay of the ice sheets in both polar hemispheres as being incompatible with a simple orbital variation-ice sheet link. The sinusoidal nature of the orbital forcing should produce a smoother glacial cycle, and it is important to note that this forcing is not uniform over the globe, but is strongly dependent on latitude. At the time of the glacial terminations, insolation was approaching a peak for the northern hemisphere summer, while it was actually declining for the southern hemisphere summer (Broecker and Denton, 1989). The fact that marine oxygen isotope records from all over the globe correlate only with the insolation curve for northern hemisphere high latitudes and not for their own respective latitudes implies that the orbital forcing received at this latitude must be transmitted globally as a climatic signal to the cryosphere.

To explain the rapidity and global synchroneity of glacial terminations and the accompanying changes in other climatic factors, such as atmospheric carbon dioxide and methane and dust (*see* Figure 8), Broecker and Denton (1989) invoke major reorganizations of the ocean-atmosphere system. These constitute jumps between stable modes of operation which cause changes in the atmosphere's greenhouse gas content and albedo. These abrupt reorganizations may have been driven by orbitally induced changes in freshwater transports which impacted the salt structure of the world's oceans. Particularly sensitive to change is the site of North Atlantic Deep Water formation, located around 65°N. The reader is referred to Broecker and Denton (1989) for a much more complete description of this possible mechanism of climate change.

Thus, it appears that while Milankovitch orbital forcing is responsible at a basic level for the glacial cycles, the linkage between the orbital forcing and the climatic response of the ice sheets is quite complex, and is not yet fully understood (Barron, 1984). To adequately model this system, a fully coupled atmosphere-ocean general circulation model linked to a dynamic ice sheet model will be required, as a minimum. There are, however, many other aspects of climate change over the Late Pleistocene that may be more easily explained in terms of known forcing factors. The COHMAP (1988) study used both geologic paleoclimatic data and a GCM in an attempt to better understand the effects of changing boundary conditions over the past 18,000 years on tropical monsoons and mid-latitude climates. The boundary conditions considered were the orbitally induced variations in solar insolation and variations in surface conditions including ice-sheet size, sea-ice extent and sea-surface temperatures. Their approach was to run the NCAR Community Climate Model at 3000 year intervals from 18 ka to 0 ka, and compare the results to paleoclimates reconstructed from geologic data. For each time period, the solar insolation was specified according to the three aspects of earth-sun geometry (*i.e.*, precession, obliquity and eccentricity). At 18 ka, the seasonal and latitudinal distributions of solar radiation were similar to those of today. Between 15 and 9 ka, seasonality increased in the northern hemisphere and decreased in the southern hemisphere. After 9 ka, these values decreased to present-day values. The ice-sheet geometries, sea-ice geometries and sea-surface temperatures were specified according to geologic reconstructions. Because the various aspects of the cryosphere were fixed in this modelling, there was no dynamic interaction between the orbital insolation variations and the ice sheets. This effectively removed the problem of simulating the complexity of this interaction so that the other aspects of Late Pleistocene climatic change could be examined.

The COHMAP (1988) simulations generated a series of climates that did undergo significant changes in the mid-latitude and tropical regions. Many of the important features of these simulations correlate well with the comprehensive global record of climate assembled in this study. Only a few of the major features will be discussed here: the reader is referred to COHMAP (1988) for a complete discussion.

The presence of the large and high (3 km) Laurentian ice sheet at 18 ka, in the first simulation, split the flow of the winter jet stream across North America, with one branch over southern North America and the other along the northern edge of the ice sheet. The geologic data for this time show an expansion of woodlands in the American Southwest along with higher lake levels, indicating higher amounts of precipitation. This is consistent with the splitting of the jet stream, which would bring its associated storm tracks and increased amounts of precipitation to this region. In subsequent simulations, the Laurentian ice sheet had decreased in size and thickness enough so that the winter jet stream was no longer split. By 12 ka, lake levels in the Southwest had decreased. The temperatures simulated through the series of model runs were largely in accord with vegetation patterns reconstructed from the geologic record; however, the model summer temperatures in southern North America were too high from 18 ka to 12 ka. Also, apart from the success at predicting the southwest lake levels, the model had only fair success at reproducing mid-latitude precipitation patterns. This is a common problem in GCMs because most precipitation processes occur at a smaller scale than the model resolution and must be parameterized.

The present-day Asian monsoonal circulation is driven by the land-sea thermal contrast between southern Asia and the Indian Ocean, which is strongly enhanced by the Tibetan Plateau. Between 12 and 6 ka, rising insolation in July and decreasing insolation in January increased this land-sea thermal contrast. The summer monsoons were strengthened in the simulated climates for these time periods and brought greater amounts of precipitation to the Sahara, Arabia and south and east Asia. After 6 ka, July insolation decreased, the simulated temperatures over land decreased and the monsoons weakened. The northern subtropical deserts expanded and present-day climatic patterns developed. Both vegetation pattern reconstructions and lake levels show that from 12 to 6 ka, precipitation was higher across northern Africa, Arabia and into northwest India. After 6 ka, the lake levels in these regions decreased.

DISCUSSION

Apart from the seasonal cycle and historical records, the geologic record is the only proxy of climatic change available for study. The two case studies above illustrate the extent to which general circulation models can be used to examine climatic states substantially different from the present day. The models

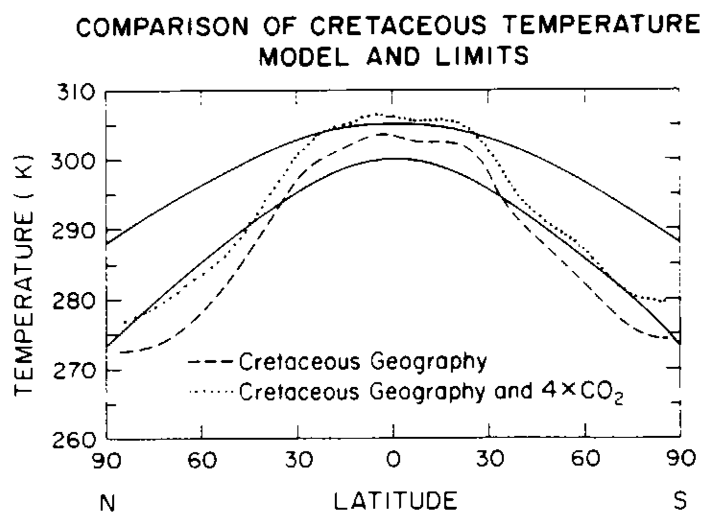

Figure 7 *Cretaceous zonally averaged temperature limits from geologic paleoclimate data in comparison with Cretaceous model-derived surface temperatures for the geography and geography plus CO_2 quadrupling experiments. (From Barron and Washington, 1985).*

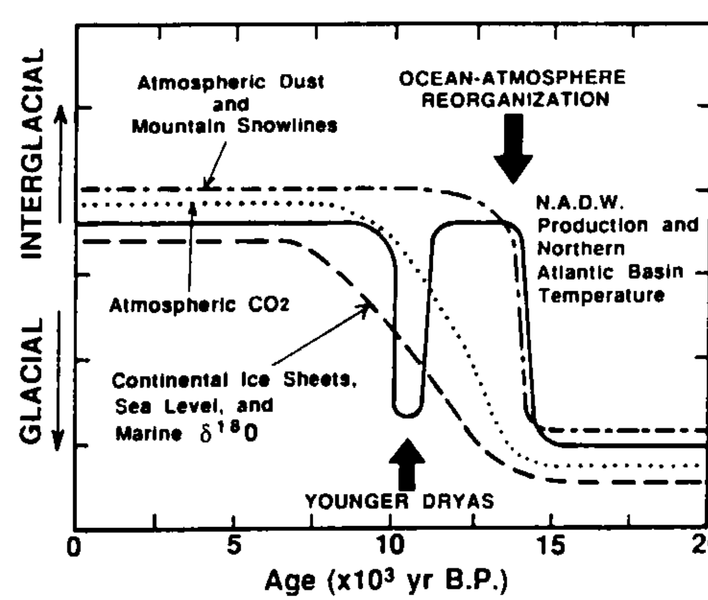

Figure 8 *Summary of the timing of the events associated with the last glacial termination. (From Broecker and Denton, 1989).*

appear to be capable of reproducing many first-order climatic effects found in the sedimentary record when the appropriate boundary conditions are altered. However, this success must be tempered by the fact that there are still many features that are poorly simulated or not simulated at all by the GCMs. The reasons for failures may be threefold: the model sensitivity is inappropriate because a feedback process has not been included or a parameterization is in error, additional climatic forcing factors have not been included, or the paleoclimatic data have not been interpreted correctly.

It is clear that general circulation models are contributing to our understanding of paleoclimates and climatic change, but more work with ancient climates is needed to specify any model's accuracy, and its utility in predicting possible future global climatic change.

The two case studies above are illustrative of two approaches to this task. The first is to produce a better "map view" of one particular climatic variable, which can extend to a broader view of the whole record. A good example of this is a study of the NCAR GCM's ability to simulate accurately the location of wind-driven oceanic upwelling for modern and ancient examples (Kruijs, 1989; Kruijs and Barron, 1991). This is an important climatic variable because oceanic upwelling is associated with high primary productivity and the deposition of organic carbon-rich sedimentary rocks. The emphasis of this study was to assess rigorously the model's predictions as follows: (1) determine the actual relationship between wind-driven upwelling and high primary productivity in order to establish rules which can be applied to Earth history, (2) determine the capability of the GCM to predict the present-day wind stress that drives oceanic upwelling, (3) determine the sensitivity of the model to changes in boundary conditions such as geography (*i.e.*, are the predictions different for past time periods?), and (4) determine the capability of the model to predict past regions of upwelling and productivity by comparison with a suite of upwelling indicators in the sedimentary record. If this type of approach is taken with other climatic variables, the limits to the model's predictions will be much better known.

The second approach is to assess the model's capability to reproduce the global climatic change preserved in the sedimentary record. Because of their complexity and cost to operate, model runs are only executed for a snapshot of geologic time. Thus, to evaluate climatic change over geologic time scales, several model runs from different time periods must be examined together. The model fields of climatic variables are literally stacked on top of each other and "cored" through. This was essentially the approach of COHMAP (1988) for the last 18,000 years, where orbital variations are thought to have controlled the global climatic state. However, the same approach can be taken with longer stretches of geologic history, where changing geography, CO_2 compositions and so on are the important boundary conditions controlling the climatic state.

A combination of these two approaches in future studies will lead to a much better understanding of the capabilities and limitations of general circulation models in representing climatic conditions substantially different than the present day. If the models can simulate a map view of a time period and also the changes through time, then we can have much more confidence in the model results. This will affect attempts to predict future global climatic change and add more insights into the study of ancient climates.

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