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A review of terrestrial and marine climates in the Cretaceous with implications for modelling the 'Greenhouse Earth'

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Abstract – From the unique perspective of the geological record, it appears that the 'Greenhouse Earth' was a feature of climate for up to 80% of the last 500 Ma, and that therefore our present glacially dominated climate is an anomaly. The Cretaceous in particular was a time of global warmth, an extreme greenhouse world apparently warmer than our current Earth. The geological record provides perspective and constraints against which the success of climate models can be evaluated. At present there are no ways of evaluating model predictions for the future of our 'Greenhouse Earth' until after the event. Retrodicting the past is therefore a very useful way of testing model sensitivity and robustness. The geological record tells us that the characteristics of the Cretaceous greenhouse world were a shallower equator-to-pole temperature gradient, shallow, well-stratified epicontinental seas with a tendency towards periodic dysaerobism, and a well-developed terrestrial flora extending to the high latitudes. Both marine and non-marine data show a global cooling trend throughout Late Cretaceous time, a trend that seems to correlate with declining atmospheric carbon dioxide.

1. Introduction

The Cretaceous (140 Ma to 65 Ma) was a time of global warmth when the Earth's climate was in an extreme 'greenhouse' mode. Fossil evidence for this is varied but the most obvious manifestation is extensive forests at very high (75–85°) palaeolatitudes. The utility of the fossil record in this respect is obvious but the climate signal obtained from palaeontology goes far beyond qualitative observations. Because plants are relatively simple organisms that are not mobile after germination, their morphology and community structure reflects strongly the physical environment to which they are exposed. Those plants not well suited to a particular environment are selected against and patterns of morphological specializations with respect to climate developed early in land plant evolution (Spicer, 1989*a*). These patterns subsequently evolved repeatedly in unrelated groups (Meyen, 1973) and thus the plant fossil record provides reliable data for palaeoclimatic reconstructions. Moreover, because the plants concerned grow on land they reflect the atmosphere directly and careful analysis of the physiognomy (structure and appearance) of leaves and vegetation can yield quantitative data on air temperatures and wetness (Spicer, 1989; Wolfe, 1990).

By contrast marine organisms are buffered from the atmosphere by ocean waters and therefore do not give a direct record of conditions in the air masses above. Nevertheless, marine deposits are extensive, isotopic signatures are often preserved in body parts and can be used as a proxy for temperature, and atmospheric and ocean systems together determine global climate. This means that to understand palaeoclimates in any realistic sense both marine and terrestrial data have to be integrated.

There is no evidence for extensive Cretaceous polar glaciation (although montane glaciation at the poles seems probable; Spicer & Parrish, 1990*a*) and geological evidence also points to high sea levels. Both marine and non-marine rocks of Cretaceous age are widespread globally and there are sufficient data for us to reconstruct the positions of the continents with considerable accuracy. To some extent we can also estimate topography, and by using sedimentological and palaeontological data both quantitative and qualitative patterns of climate can be deduced. Figure 1 illustrates the palaeogeography of the world during early Maastrichtian time. Also shown are the areas of land and sea during this time together with the palaeolocation of the sites discussed in this paper (map from Barron, 1987).

The wealth of Cretaceous geological data provides a valuable perspective for predictive studies of possible future global warming, particularly as the Cretaceous appears to have been a time when atmospheric CO_2 concentrations were greater than those at present. Numerical climate models derived from medium-term weather forecasting are useful tools for understanding the relative importance of a variety of factors that affect climate. However, these models are necessarily simplified versions of the real world and if the present only is used as a baseline we cannot assess the reliability of the model predictions. When the boundary conditions (for example, atmospheric CO_2 levels, average global temperature, amount and distribution

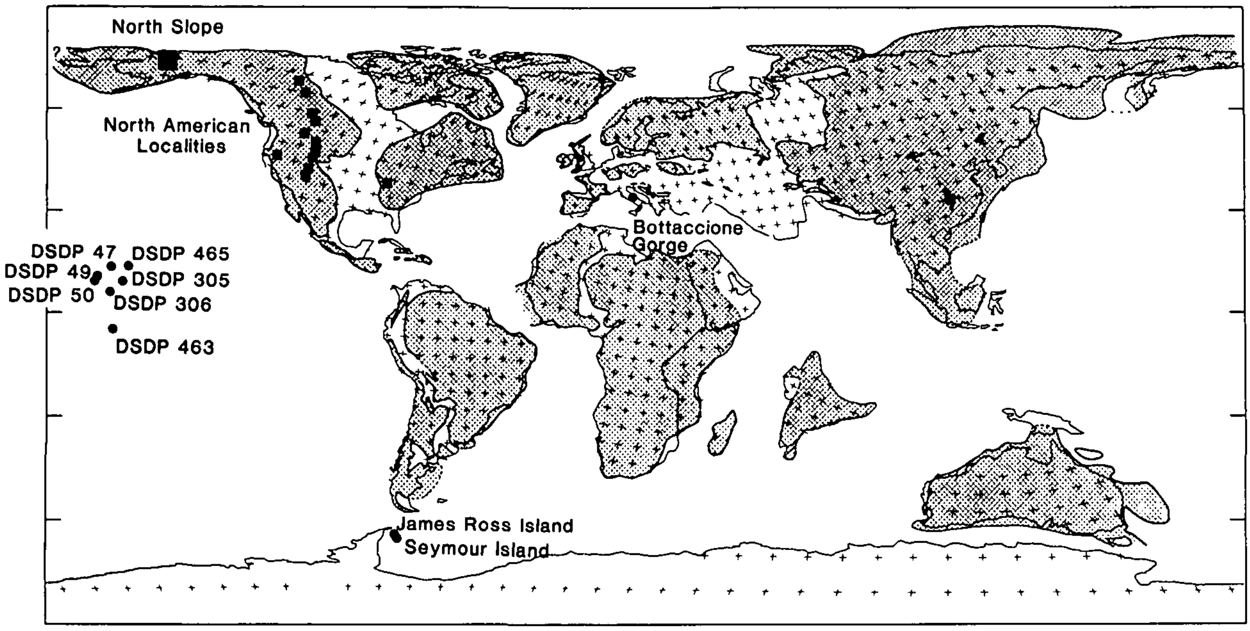

Figure 1. Map of the Cretaceous world during the early Maastrichtian showing distribution of land and sea together with the palaeocoordinates of sites discussed in this review (modified from Barron, 1987).

of vegetation and polar ice) depart from those of the present the model predictions are rendered unreliable due in part to simplification. Only by attempting to model the past and comparing the model results against geological data can model performance and reliability be evaluated. At the present time it is clear that model results (e.g. Sloan & Barron, 1990) fail to match a wealth of independent geological and palaeontological evidence (Wing, 1991). To us this suggests the models are in need of substantial improvement, particularly as palaeontological evidence shows that a wide variety of unrelated organisms, both plants and animals, yield consistent and comparable climatic signals that are at odds with the model predictions. In this paper we consider both marine and non-marine climate data and concentrate on the Cretaceous because of the quality of the dataset and the extreme relative warmth of the Cretaceous Period.

2. Cretaceous terrestrial vegetation and climate

2.a. High northern latitudes

In the simplest of terms, global climate is driven by energy from the sun unevenly distributed over the surface of our globe by virtue of the fact that the Earth's surface is spherical. The poles receive less incoming radiation than the equator and warm air rises at the equator and sinks at the poles. The equator-to-pole temperature gradient is an important factor in driving the system but simple atmospheric heat transfer from equator to poles is complicated by the fact that the Earth is a rotating body. As a consequence of this, atmospheric circulation tends to be zonally organized parallel to latitude (Parrish, 1982). This simple zonal model is further disrupted by land/sea distributions because land and sea have different reflectivities (albedo) and thermal inertia, but it is generally recognized that global climates are defined to a considerable degree by conditions at the poles (Goody, 1980).

Throughout the Cretaceous there is evidence that forest vegetation existed to well within the palaeo-Arctic Circle and in some cases as high as 85° N. Extensive Cretaceous sediments in northern Alaska are highly fossiliferous and allow detailed reconstruction of plant community structure and climate throughout late Cretaceous time (Spicer & Parrish, 1986, 1987, 1990*a*). Prior to the arrival of the angiosperms the Arctic forests were rich in conifers, ginkgophytes, ferns, and even cycads. Previously the presence of cycads was taken to indicate a warm temperate to even subtropical regime (Smiley, 1967; May & Shane, 1985). Furthermore the evergreen nature of living cycads was extrapolated to those of the past and cited as evidence for a more even distribution of light throughout the year rather than the three months or more of continuous winter darkness which is experienced today at latitudes greater than 75° N. This would imply a significantly reduced obliquity (Wolfe, 1978; Douglas & Williams, 1982). Attractive as this idea is for explaining apparently evergreen vegetation, the insolation energy received overall at the poles would fall below present values and would have the effect of cooling, rather than warming, the high latitudes (Barron, 1984). However, many Mesozoic cycads were quite unlike their modern relictual relatives and some of the genera occurring in Alaska were slender, deciduous forms

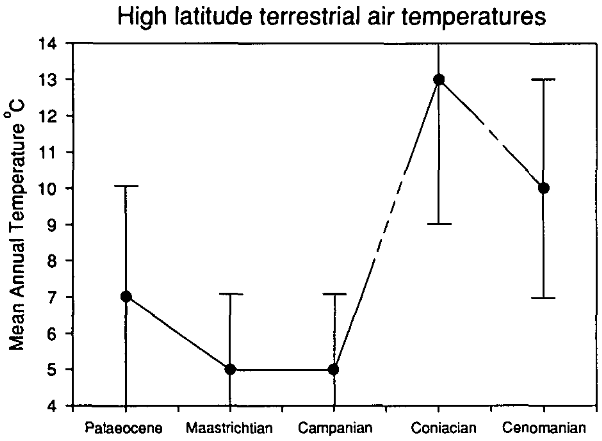

Figure 2. Mean annual air temperatures from the North Slope, Alaska during the mid- and late Cretaceous estimated from leaf physiognomy (from Spicer & Parrish, 1990*a*).

(Kimura & Sekido, 1975; Spicer, 1990). Thus in a simple uniformitarian sense cycads are a poor guide to past climate.

Extensive studies of plant fossils in the context of their depositional environments from mid- to latest Cretaceous times have shown that almost all of the high Arctic taxa were deciduous, and those that were not could enter dormancy or overwinter by dying back to underground organs or in seed form (Spicer & Parrish, 1987, 1990*a*). Thus evergreenness was not a feature of these very high latitude floras even though they were conifer dominated. Angiosperms at high latitudes in the late Cretaceous appear to have been only locally abundant even though they were taxonomically diverse. Although there is no exact modern analogue of this vegetation type the closest is that referred to as low montane mixed coniferous (*sensu* Wolfe, 1979) living under a cool temperate regime. Further evidence predicating against a significant obliquity change is found in tree rings which show a rapid cessation of growth at the onset of winter (Parrish & Spicer, 1988), a feature compatible with the rapid change in daylength seen today at high latitudes (Spicer, 1987).

By using reconstructed vegetation types and angiosperm leaf physiognomic studies, particularly leaf margin characters, it has been possible to construct mean annual air temperature curves for the late Cretaceous and early Tertiary for the near-polar coastal plain of northern Alaska (palaeolatitude 75–85° N); Fig. 2; (Spicer & Parrish, 1990*a*). From a mean annual temperature (MAT) of 10 °C (approximately the same as present-day London) in the late Albian–early Cenomanian, the temperature may possibly have risen slightly before dropping to about 5 °C in the Campanian and early Maastrichtian. This last temperature is an estimate in part constrained by the higher and better understood early Tertiary temperature of 6–7 °C (Spicer & Parrish, 1990*a*). Thus in the northern hemisphere MATs are incompatible with sea-level glaciation (but not seasonal ice) throughout the late Cretaceous. By applying the present globally averaged saturated adiabatic lapse rate of 5.5–6 °C km^{-1} (which may be too high for this situation), it is possible to estimate that MATs of 0 °C (and perhaps permanent ice fields) might not have been encountered below 1000 m above Cretaceous sea level even during the coldest interval (late Campanian/early Maastrichtian) at 85° N (Spicer & Parrish, 1990*a*).

At latitudes greater than 75° and with obliquity close to that of the present (approximately 23°), winter darkness would have lasted more than three months bounded by twilight totalling 1.5 months. Such a long period without insolation would inevitably have led to a drop in temperature while the converse (continuous summer daylight) must have resulted in high summer temperatures. Thus the annual range of temperature tended to be high. In the Cretaceous dark-induced winter dormancy prevented the plants recording winter temperatures (although the mean signal was apparently unaffected (Spicer & Parrish, 1990*a*)) and thus mean annual range of temperature is difficult to estimate. However, the lack of permafrost as evidenced by incomplete freezing of tree root zones, absence of periglacial sedimentary structures (Spicer, 1987), and vegetation physiognomy constraints argue for cold monthly mean temperatures no lower than −11 °C (Parrish *et al.* 1987). A subjective estimate would put the mean annual range of temperature in the region of 25 °C.

In the early late Cretaceous abundant thick coals, wide growth rings in the wood, and a large range of leaf sizes all indicate a generally wet regime with the plants never experiencing any significant water stress. Fossil fungal spores are particularly common (Grant, Spicer & Parrish, 1988). Today polar regions tend to have a low precipitation as the result of a well-developed atmospheric high pressure regime centred close to the pole. Clearly the polar high pressure area must have been weaker in the Cretaceous, allowing penetration to higher latitudes of storm systems carrying both heat and moisture. During the late Cretaceous cooling trend the abundance of Arctic coal-forming mires diminished and the incidence of charcoal increased in the Alaskan rock record (Spicer & Parrish, 1987). This may suggest that drying of the forests was more pronounced and that wildfires were correspondingly more frequent. Cooling seems to lead to drying (the reverse of what might normally be expected due to lowered evaporation rates), suggesting an intensification of the polar high. However, extreme drought was never experienced as evidenced by a lack of preserved mud cracks, and the mostly large cell size in the early wood of fossil logs suggests water stress was never experienced due either to drought or freezing to below the root zone (Spicer & Parrish, 1990*a*, *b*).

2.b. Antarctic

Conditions in the high southern latitudes are less well documented than for the high northern latitudes because the fossil leaf record is not as extensive. However, a number of key observations based on palynomorphs show that physiognomically similar, though taxonomically rather distinct, vegetation existed on the southern polar continents of Antarctica and southern Australia (Douglas & Williams, 1982; Francis, 1986; Dettmann & Thomson, 1987; Askin, 1989; Daniel, Lovis & Reay, 1990; Truswell, 1990; Askin & Spicer, in press). Mid-Cretaceous conifer forests dominated by auraucarians and podocarps grew on the lower coastal plains surrounding Antarctica and are indicative of cool temperate rainforests. Palynological evidence from Livingston and Snow islands suggest a diverse and abundant fern and lycopod understorey flora (Askin, 1983), a feature of the vegetation that appears to last throughout the Cretaceous. The first angiosperm pollen grains appear in the early Albian (Truswell, 1990). Like the northern Alaskan palynoflora, fungal spores and hyphae are abundant.

Late early Cretaceous leaf and wood floras of the Otway and Strzelecki basins, Victoria, Australia (Douglas, 1969, 1973) also indicate the existence of a conifer-dominated forest system with a subordinate cycadophyte and fern component. In a re-evaluation of these floras Parrish *et al.* (1991) note that some leaves commonly occur as mats (e.g. *Phyllopteroides*; Cantrill & Webb, 1987) and have thin cuticles while microphyllous conifers and small-leaved bennetitaleans occur more widely and often have moderately thick to thick cuticles and sunken stomata with overarching papillae for reducing water loss. Forms with thin cuticles are interpreted as being deciduous while the more xeromorphic elements could have retained their leaves throughout the winter provided that temperatures were cold enough to slow metabolic processes significantly.

These floras occupied the rift valley opening up between Australia and Antarctica and were sited at about 75° S. The continental setting for the floras would have produced a climate with a greater mean annual temperature range than that experienced at coastal sites, and the colder winter temperatures would have slowed metabolic processes to such a degree that evergreenness was a viable strategy for overwintering provided that desiccation could be prevented. The retention of leaves throughout the winter conserves the energy that would otherwise be required to produce new leaves each spring. Deciduousness and an enforced dormancy is a more energy-efficient strategy only when winter temperatures are relatively warm and energy-demanding respiration would otherwise continue at a high level. Parrish *et al.* (1991) suggest that the MAT of the rift valley floor was cooler than that implied by Douglas & Williams (1982) but not as cold as that given by Gregory *et al.* (1989) and Rich & Rich (1989), based on oxygen isotopes from non-marine concretions. The MAT was probably about 7 °C. The low temperatures yielded by isotope analysis may be due either to diagenetic factors or to being sourced by meltwater from montane glaciers on the flanks of the rift valley.

By contrast coastal Cenomanian floras from the Clarence Valley, South Island, New Zealand (Daniel, Lovis & Reay, 1990) are physiognomically more comparable with coeval floras of Alaska. Many major groups including angiosperms, conifers, ferns and cycadophytes are broad-leaved, have mostly thin cuticles, and many appear deciduous. It may be that coastal temperatures at high southern latitudes (> 75° S), like those of the north, were high enough to favour wholesale deciduousness.

2.c. Low and middle latitudes

Obtaining quantitative estimates of mean annual air temperature and precipitation from vegetation devoid of angiosperms is currently rather difficult as our ability to calibrate the physiognomy of such vegetation is limited. Research aimed at correcting this deficiency is underway by one us (R.A.S.) but more work is needed. However, from mid-Cretaceous times onward the abundance of angiosperm-rich floras in North America offer insight into the continental temperature gradient and precipitation distribution with respect to latitude.

Comparable physiognomic techniques to those used at high latitudes show that megathermal vegetation (MAT tolerance above 20 °C) extended to palaeolatitude 45–50° N, while mesothermal vegetation (MATs between 13 and 20 °C) with a significant evergreen component thrived up to the palaeo-Arctic Circle (*c.* 66° N). Mesothermal vegetation appears to have been a partly open evergreen woodland with evergreen conifers being dominant (Wolfe & Upchurch, 1987). Some angiosperms were large trees but most were small trees and shrubs. Growth rings in trees suggest precipitation may have been limiting to growth and varied markedly from year to year. Megathermal vegetation was also subhumid, evergreen and conifer-rich, but rainforests were apparently absent or at most isolated in small coastal patches. One explanation for this is that with a weaker polar high, the intertropical convergence zone underwent a greater seasonal excursion about the equator and no low latitude regions were constantly wet (Ziegler *et al.* 1987; Ziegler, 1990). Thus fossil evidence seems to suggest that at times of global warmth tropical rainforests are less extensive than at present. This has important implications because cutting and burning of present-day tropical rainforests may contribute to bringing about a climate that prevents them from becoming re-established and acting as a carbon sink.

Throughout much of the late Cretaceous the North American temperature gradient appears to have been about 0.3 °C per 1° of latitude (Wolfe & Upchurch, 1987; Fig. 3), although fluctuations most evident at higher latitudes modified this from time to time. For the Campanian–Maastrichtian interval the precipitation/evaporation ratio (as expressed by leaf size) shows two peaks: one spanning 45–55° N palaeolatitude and the other above 65° N (Fig. 4).

2.d. Cretaceous polar ice

With a southern polar continent, in contrast to an Arctic ocean, greater annual temperature range might be expected in the Antarctic due to the low thermal inertia of land. In particular the question arises as to the existence and quantity of permanent ice. Superficially short-term fluctuations in sea level (Vail, Mitchum & Thompson, 1977; Haq, Hardenbol & Vail, 1987) would seem to suggest the existence of substantial Antarctic ice. However, extensive coastal cool temperate rainforests indicate a lack of periglacial conditions at sea level, although calving of montane glaciers to the coastal plain or into the sea could not be discounted (as work by Pirrie & Marshall (1990) and Rich *et al.* (1988) suggests). However, if the

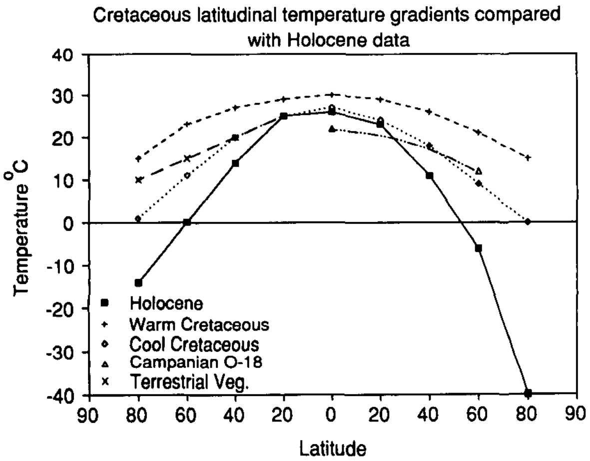

Figure 3. Latitudinal temperature gradients for the mid-Cretaceous and the Holocene compared. Data for the Holocene and warm and cold Cretaceous replotted from Barron, 1983. The 'warm Cretaceous' estimate is based on temperatures suggested by oxygen isotope data, and the 'cool Cretaceous' estimate is based upon the assumption that polar temperatures reached, but did not exceed, freezing in the winter. Also shown is the latitudinal thermal gradient estimate (0.3 °C/1° latitude) of Wolfe & Upchurch (1987) for Cenomanian time ('terrestrial veg.') These estimates fall within the boundary conditions defined by Barron, 1983. Also shown is an estimate of the marine latitudinal thermal gradient for the Campanian (data from Boersma & Shackleton (1981) and Barrera *et al.* (1987)). This plot suggests that although late Cretaceous low-latitude temperatures may have been lower than during the mid-Cretaceous, the latitudinal temperature gradient was not vastly different.

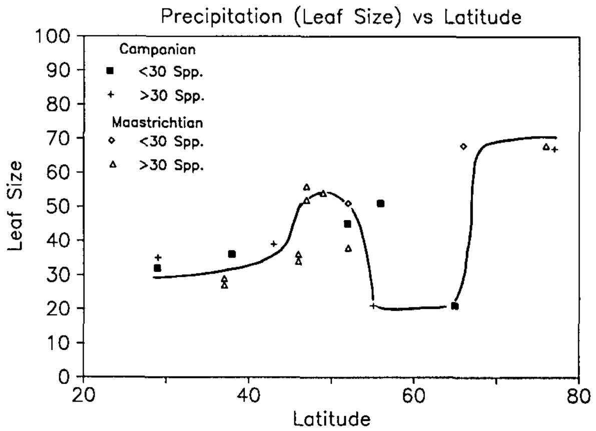

Figure 4. Precipitation estimates with latitude based on leaf size for the Campanian and Maastrichtian of North America (from Wolfe & Upchurch, 1987).

coastal temperatures were similar to those of the higher northern latitudes, MATs of 0 °C would not exist below 1000 m above sea level.

In the northern hemisphere, it is possible that Cenomanian time was characterized by a mean annual temperature of 0 °C or less 1700 m above sea level. In the latest Cretaceous when the North Slope (Alaska) was at 85° N the 0 °C isotherm would have dropped to about 1000 m above sea level (Spicer & Parrish, 1990*b*). From this we conclude that permanent snowfields or even montane glaciation probably occurred in the newly uplifted Brooks Range in the latest Cretaceous although there is no evidence to suggest that glaciers extended down to the coastal plain.

3. Cretaceous marine palaeotemperatures

3.a. Introduction

One of the most widely used methods of palaeotemperature determination in the marine realm has been the oxygen isotope method pioneered by Urey (1947) and his colleagues (Epstein *et al.* 1953; Emiliani, 1954). The method is based on the observation that the ratio of two stable isotopes of oxygen (^{16}O and ^{18}O), when precipitated in carbonate from a surrounding solution, is temperature dependent. In warmer waters there is a greater proportion of the light isotope of oxygen, while in cooler waters there is relatively more of the heavy isotope. This relationship holds even when the calcite is precipitated organically by foraminifera, molluscs, brachiopods or corals, although in some cases within these groups the ratio is altered due to vital effects and absolute temperatures cannot be calculated.

Oxygen isotope ratios are expressed as deviations from an internationally recognized standard: the Pee Dee belemnite from a rock formation in Carolina. The calcite derived from this organism has long since been consumed, and today results are calibrated to PDB by the use of other, more widely available laboratory

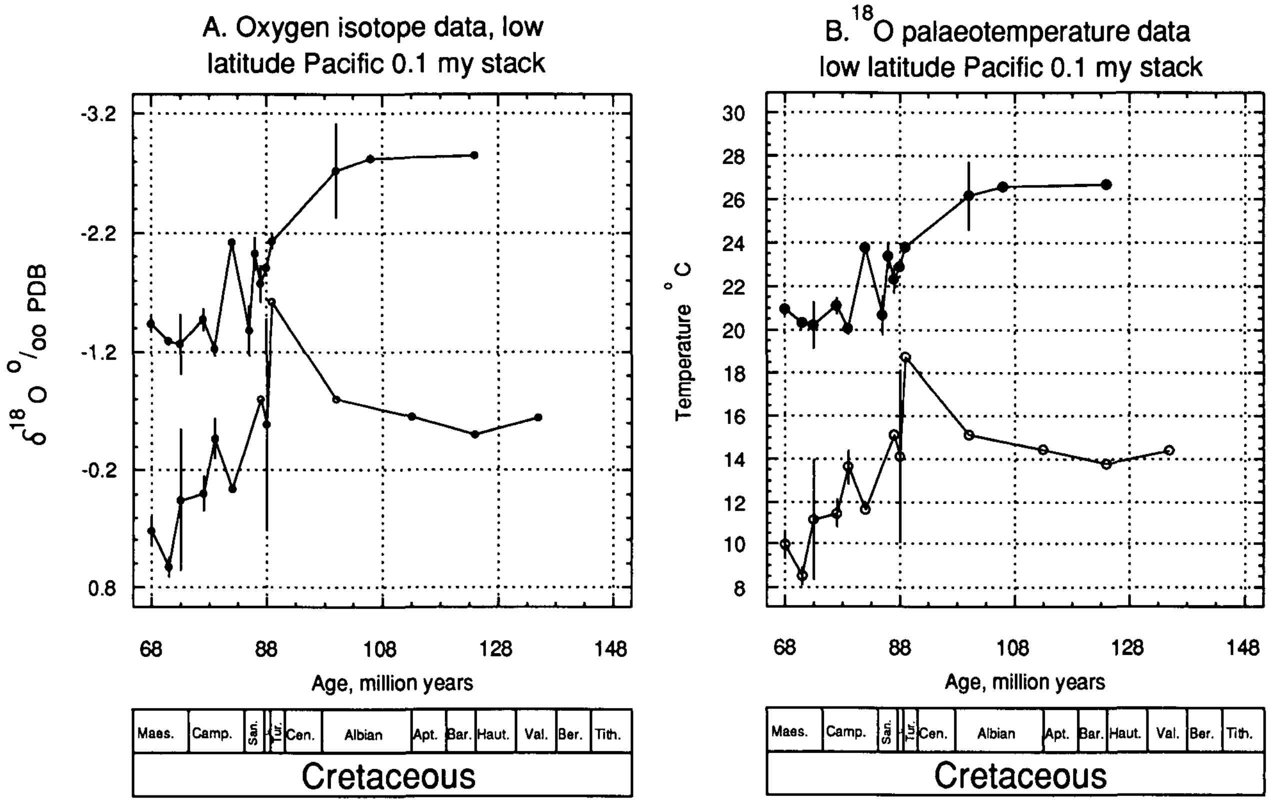

Figure 5. (a) Foraminiferal oxygen isotope data from the low-latitude Pacific Ocean during the Cretaceous. Data from Douglas & Savin (1975); Boersma & Shackleton (1981). The late Cretaceous cooling can be clearly seen in both planktonic and benthonic foraminifera. (b) The oxygen isotope data from Figure 5a are plotted on a temperature scale. Note that for the purposes of this exercise two estimates of the isotopic composition of sea water (δ_w) are employed. Surface waters are considered to have an ^{16}O enrichment of −0.5‰ reflecting both the assumed lack of polar ice-caps (−1.2‰) and a surface water ^{16}O depletion of −0.7‰ due to evaporation (Shackleton, Corfield & Hall, 1985). Closed circles: *planktonics*; open circles: *benthics*.

standards. The deviations are expressed using the δ notation with respect to the heavier isotope such that oxygen isotope ratios are written as δ^{18}O.

Carbon isotope ratios are also becoming widely used in palaeoclimatic work because of their connection with the carbon cycle, and through this with the CO_2 content of ancient atmospheres. These ratios are written as δ^{13}C.

The oxygen isotope technique of palaeotemperature determination has been effectively used for the analysis of Pleistocene climates from deep sea cores (Emiliani, 1954; Shackleton, 1967; Shackleton & Opdyke, 1977; and many others). Because of the advent of more, and better preserved material from the Deep Sea Drilling Project (now the Ocean Drilling Program) these analyses have extended further back into the geological past and we now have estimates of palaeotemperatures from a variety of latitudes from most ocean basins for a good proportion of Cenozoic time.

However, due to the effect of sea-floor spreading, old ocean floor is consumed at the ocean margins and this has the effect of limiting the age of ocean sediments that are available for palaeoclimatic analysis. Hence with increasing age in the Cretaceous it becomes progressively harder to find material to analyse for palaeotemperature estimation.

By way of compensation there are good land sections of Cretaceous sediments such as the deep sea limestones of the Umbrian Apennines in Italy and the softer rocks of the English Chalk that can be analysed. However, the disadvantage here is that because the sections are on land, they have been subjected to weathering and potential isotopic exchange with fluids which pass through the rock. Great care must therefore be taken to control for the effect of this diagenesis when interpreting data from these localities.

3.b. Low latitude marine palaeotemperatures

Where Cretaceous data from the deep sea are available (e.g. Shatsky and Hess rises in the North Pacific; Walvis Ridge in the South Atlantic) it is usually possible to make measurements on individual species of foraminifera. These single-celled organisms secrete a calcium carbonate shell, which when analysed isotopically provide a measure of the temperature of the water in which the organism grew. Planktonic foraminifera therefore provide estimates of the temperature of surface waters and benthic foraminifera that of bottom or deep waters.

Figure 5a illustrates a compilation of available δ^{18}O data from the low latitude Pacific (Douglas & Savin,

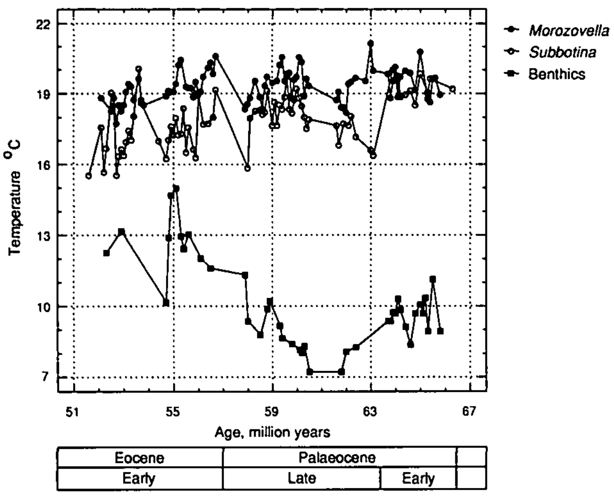

Figure 6. Oxygen isotope palaeotemperature estimates from the Palaeocene of DSDP 577 (equatorial Pacific). Note similarity of values for latest Cretaceous time (Fig. 5) and the early Tertiary.

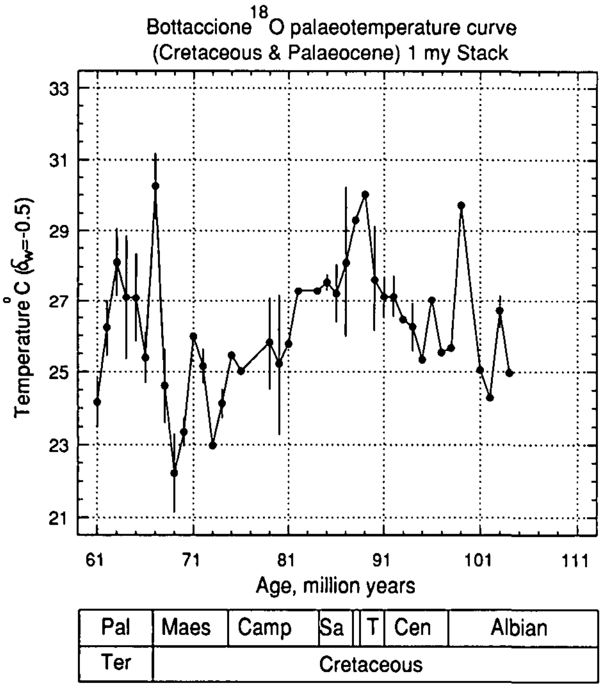

Figure 7. Oxygen isotope palaeotemperatures from the Bottaccione Gorge. Note the post-Turonian temperature decline.

1975; Boersma & Shackleton, 1981). These data have been replotted on a temperature scale in Figure 5b. The data indicate warm surface water temperatures during early Cretaceous time followed by a cooling after the Cenomanian. Deep water temperatures from benthic foraminifera are in broad agreement with this late Cretaceous cooling trend. Data from the Palaeogene of DSDP 577 (Fig. 6) indicate that surface water temperatures were similar in the latest Cretaceous and the early Tertiary of the Central Pacific. Hence the early Cretaceous was clearly a time of anomalous warmth. Although it is only possible to show relative temperature trends from $\delta^{18}O$ analysis of limestone, data from the Bottaccione Gorge (Corfield *et al.* 1991; Fig. 7) also show a cooling from the Turonian to the Maastrichtian. The widespread occurrence of this cooling is confirmed by $\delta^{18}O$ analyses of belemnites and brachiopods from northwest Europe (Lowenstam & Epstein, 1954; Spaeth, Hoefs & Vetter, 1971), as well as by whole-rock analyses of the English Chalk (Jenkyns *et al.* unpub. data).

3.c. Higher latitude marine palaeotemperatures

Barrera *et al.* (1987) and Pirrie & Marshall (1990) have published isotope data from the high southern latitudes of the Antarctic Peninsula on Foraminifera (Barrera *et al.* 1987) and Mollusca (Pirrie & Marshall, 1990) which suggest cool temperature conditions in this region during the Campanian–Maastrichtian interval. In addition, Pirrie & Marshall (1990) have identified a cooling trend between the Santonian–Campanian and the Maastrichtian which may be the local expression of the late Cretaceous cooling trend recognized in the lower latitudes.

Non-marine oxygen isotope records are presently somewhat controversial and rare. Nevertheless Rich *et al.* (1988) report maximum estimated MATs of 5 °C in the late early Cretaceous of southern Australia (palaeolatitude 70–80° S).

Using the isotopic temperature estimates of Barrera *et al.* (1987) and Pirrie & Marshall (1990), we calculate that latitudinal temperature gradients were very much shallower during this interval of the Cretaceous than during the present day (Fig. 3). As observed by other workers (e.g. Barron & Washington, 1985), low-latitude temperatures appear to have been similar to those of today while high-latitude temperatures were very much warmer. We estimate that the marine temperature gradient was in the region of 0.14 °C/1° of latitude, a figure not dissimilar to the estimate of Barron (1983, fig. 3, page 312) who shows a gradient of approximately 0.21 °C/1° latitude. These figures are in broad agreement with the 0.3 °C/1° latitude independently estimated by Wolfe & Upchurch (1987) for the late Cretaceous from North American terrestrial vegetation. The present-day latitudinal gradient is approximately 0.73 °C/1° latitude – at least a factor of two greater than all estimates for the Cretaceous.

3.d. Cretaceous marine dysaerobism

It seems likely that a consequence of the widespread shallow seas of the Cretaceous coupled with the low vertical and latitudinal thermal gradients was a tendency for widespread dysaerobism of the Cretaceous ocean. Such intervals have been termed 'Oceanic Anoxic Events' (Schlanger & Jenkyns, 1976) and occurred during the Aptian–Albian interval (where there appear to have been several Oceanic Anoxic Sub-Events (Arthur *et al.* 1990)), again at the Cenomanian–Turonian boundary (Arthur, Schlanger & Jenkyns, 1987) and possibly again during the Santonian. At present there is debate whether the proximate cause of these OAEs was increased surface water productivity during discrete intervals of Cretaceous time, or whether deep water circulation was periodically reduced leading to deeper water oxygen depletion.

It has been suggested that the absence of sea-level glaciation at the poles would have resulted in a very different pattern of deep water circulation to that which characterizes the present day. The current system, termed thermohaline, operates due to the sinking of cold, dense water formed in the Norwegian Sea and Weddell Sea which then flows along the sea bottom ultimately arriving in the North Pacific Ocean. In the Cretaceous, enhanced evaporation may have caused low-latitude waters to become more saline and therefore sink, a process termed halothermal circulation (Brass, Southam & Peterson, 1982; Bralower & Thierstein, 1984; Kennett & Stott, 1990). The oxygen-carrying capacity of this relatively warm water would be lower than that formed by thermohaline circulation, with the result that deep waters would be considerably less well oxygenated than at present, possibly promoting periodic stagnation of the oceans.

3.e. Carbon dioxide in the Cretaceous atmosphere

It has been suggested (e.g. Arthur, Dean & Schlanger, 1985; Barron & Washington, 1985; Arthur, Allard & Hinga, 1991) that the warm Cretaceous was a product of high CO_2 levels in conjunction with predominantly low-latitude continentality. Estimates of the level of CO_2 in the Cretaceous atmosphere needed to produce the range of temperatures observed range from a factor of 2 to a factor of 10 greater than the present-day (Barron & Washington, 1985). Arthur, Dean & Pratt (1988) and Arthur, Allard & Hinga (1991) have estimated a CO_2 level 8 times higher than present during the mid-Cretaceous.

As noted earlier, one of the most striking features of Cretaceous palaeoclimatology is the marked cooling from the mid-Cretaceous up until the end of the Cretaceous. In this section we investigate the possibility of an inter-relationship between this cooling phase and changes to the CO_2 content of the late Cretaceous atmosphere.

Carbon dioxide concentrations in the Pleistocene atmosphere have been estimated by Shackleton *et al.* (1983) using the ^{13}C difference ($\Delta\delta^{13}C$) between planktonic and benthonic foraminifera. This value reflects atmospheric CO_2 because ^{13}C measurements in this context are a proxy for the rate of organic carbon production and hence the amount of photosynthetic activity in the surface ocean.

Photosynthesis preferentially uses the light isotope of carbon (^{12}C) and therefore leaves the CO_2 of surface waters enriched in ^{13}C. Surface-dwelling planktonic foraminifera therefore have relatively heavy $\delta^{13}C$ values. The $\delta^{13}C$ of benthic foraminifera is lighter because deep waters contain the ^{12}C liberated by the oxidation of organic matter formed at the surface and because photosynthesis does not occur at these depths.

An increase in the level of photosynthesis in surface waters will result in a greater consumption of CO_2, resulting in a lowering of p_{CO_2} in surface waters and a concomitant draw-down of atmospheric CO_2. This increase in photosynthesis will be registered by an increase in the vertical carbon gradient ($\Delta\delta^{13}C$).

To investigate whether the late Cretaceous cooling has been influenced by changes in CO_2 in the atmosphere we have assembled the $\delta^{13}C$ data from the late Cretaceous low-latitude Pacific Ocean and subtracted $^{13}C_{benthic}$ from $^{13}C_{planktonic}$ in order to obtain an estimate of the changing ^{13}C gradient, and by implication the CO_2 content of the Cretaceous atmosphere.

Figure 8 shows a plot of $\delta^{13}C$ *v.* surface water temperature. Also plotted is the relative inferred change in atmospheric CO_2, assuming that the average CO_2 content of the Cretaceous atmosphere was the same as during the Quaternary. This scale may therefore underestimate the gross CO_2 content of the Cretaceous atmosphere (although not the magnitude of CO_2 change) since there is good evidence (e.g. Arthur, 1982; Arthur, Dean & Schlanger, 1985) that Cretaceous atmosphere CO_2 levels may have been higher possibly due to CO_2 outgassing from the mantle via abundant volcanism.

There is a reasonable negative correlation (between -0.6 and -0.69) between surface water temperature in the low latitude Pacific and $\Delta\delta^{13}C$. Therefore, by implication, there is the same positive correlation between surface water temperature and atmospheric CO_2. We conclude, therefore, that the late Cretaceous temperature decline may have been mediated by enhanced photosynthesis in the surface ocean, leading to a decline in the partial pressure of atmospheric CO_2.

We regard these correlations as intriguing but far from definitive. Clearly our conclusion needs to be supported by a considerable body of further work before it attains the authority of the conclusions of Shackleton *et al.* (1983) and Shackleton & Pisias (1985), who demonstrated that fluctuations in p_{CO_2} are instrumental in the control of the Pleistocene ice ages. At present the biggest hindrance to our estimates is the inaccuracy of the timescale. The measurements used to construct Figure 8 were originally placed at best within late Cretaceous planktonic foraminiferal zones (Boersma & Shackleton, 1981), and at worst within the stages of the late Cretaceous (Douglas &

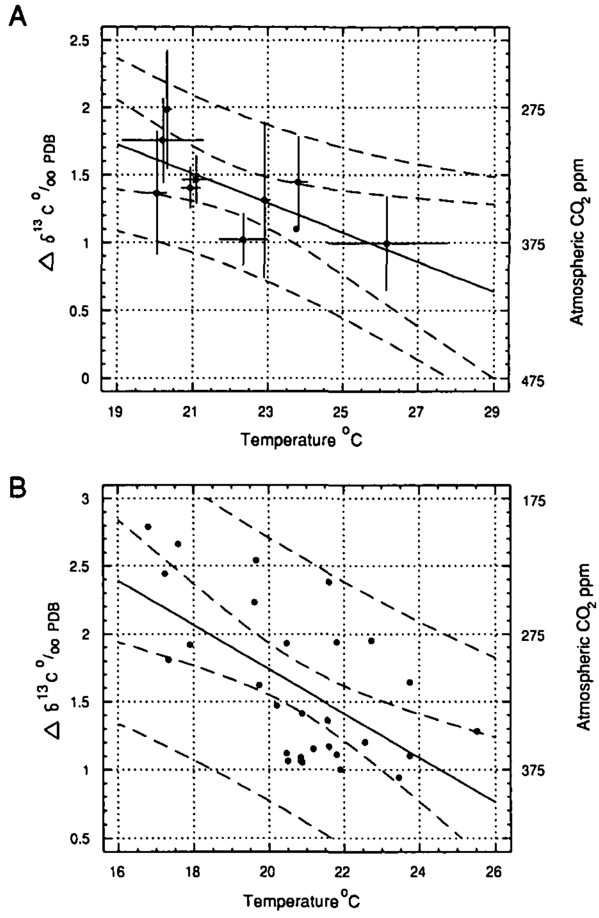

Figure 8. (a) Time and species averaged $\Delta\delta^{13}C$ versus surface water temperature data for the Cretaceous. Despite the limitations of this dataset discussed in the text, a negative correlation of -0.69 between $\Delta\delta^{13}C$ and surface water temperature is observed, suggesting that the late Cretaceous cooling of 6 °C may have been mediated by a decrease of 100 ppm in atmospheric CO_2. (b) Alternative estimate of the correlation between $\Delta\delta^{13}C$ and SWT during the late Cretaceous using raw planktonic and benthonic foraminiferal and nannofossil $\Delta\delta^{13}C$ and SWT data. Using these data the correlation is -0.60. The inner set of dashed lines is the 95% probability level. The outer set of dashed lines is the predictive uncertainty.

Savin, 1975). In addition we have averaged measurements from different species of planktonic foraminifera for surface water $\delta^{13}C$ estimates and different species of benthic foraminifera for deep water $\delta^{13}C$ estimates. This has the advantage that as much data as possible has been incorporated into our database, but the disadvantage is that this necessarily introduces error into the $\delta^{13}C$ estimates of surface and deep waters. These errors are compounded when estimating $\Delta\delta^{13}C$ from averaged $\delta^{13}C_{planktonic}$ and $\delta^{13}C_{benthic}$ measurements (note the large error bars on the $\delta^{13}C$ scale in Fig. 8a).

Future work should concentrate on the retrieval of well-preserved foraminifera from the deep sea through the offices of the Ocean Drilling Program. In addition more work needs to be done to constrain the effects of diagenesis on fossils from land sections so that we can add reliable isotope measurements from this source to our Cretaceous climatic database. Similarly more terrestrial palaeobotanical data are needed.

Clearly, our understanding of the effects of CO_2 change on the warm Earth of the Cretaceous will be vital to our understanding of the effects of anthropogenic CO_2 additions in our own future.

4. Synthesis: implications for modelling the Greenhouse Earth

Both vegetational (Parrish & Spicer, 1988; Spicer & Parrish, 1990*a*) and marine oxygen isotope data (Douglas & Savin, 1975) suggest that the late Cretaceous was a time of declining temperature. The mid- and early Cretaceous appears to have been significantly warmer than the late Cretaceous. Mean annual temperatures estimated from vegetational evidence were in the region of 10 °C near the North Pole in the mid-Cretaceous dropping to 5 °C in the latest Cretaceous. Oxygen isotope palaeotemperatures calculated from the low-latitude Pacific indicate a cooling from equatorial sea-surface temperatures of approximately 27 °C in the Albian to 21 °C in the late Maastrichtian.

Both vegetational evidence and oxygen isotope evidence indicate that latitudinal temperature gradients were very much shallower than present. Estimates from oxygen isotope palaeotemperatures suggest 0.15 °C/1° latitude, those from low to mid-latitude studies of North American vegetation suggest 0.3 °C/1° latitude. These estimates compare with 0.73 °C/1° latitude documented for the present day. Oxygen isotope data from planktonic foraminifera suggest that vertical temperature gradients in the oceans were lower than those today.

Vegetational evidence suggests that there was no significant permanent ice at the poles during the late Cretaceous. This idea is supported by high latitude oxygen isotope data which indicate that the region around James Ross Island in the Antarctic experienced a cool temperate climate.

The likelihood is that the warm, equable climate of the Cretaceous Period was a function of continents clustered mainly in the low latitudes and high CO_2 levels. It appears that the widely recognized late Cretaceous cooling may have its origins in a decline in atmospheric carbon dioxide due to enhanced productivity in surface waters.

The extreme greenhouse climate experienced during the Cretaceous provides an important test for the effectiveness of Global Climate Models. Geological evidence provides the only criterion for evaluating the performance of such models under conditions that depart significantly from the present. The success of such models in accurately predicting the likely course of future climate change depends on their robustness which can only be developed from successful modelling of ancient climates.

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