Late Cretaceous ocean: Coupled simulations with the National Center for Atmospheric Research Climate System Model

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1. Introduction
[2] The Campanian (80 Ma), during the Late Cretaceous, had an oceanic environment very different from today. The North Atlantic Ocean was still closed but with significant continental flooding resulting in shallow marginal seas. The South Atlantic Ocean was open but narrow. The paleo-Pacific Ocean extended over 180° of longitude with an open but shallow Isthmus of Panama connecting it to the Atlantic Ocean. The Tethys Ocean, precursor of the Indian Ocean, was bounded by Eurasia and Africa/Australasia/Antarctica. The attachment of Australia to Antarctica and the presence of India and the Antarctic Peninsula at 60°S latitude precluded the existence of an Antarctic Circumpolar Current.

[3] The climate was much warmer than today. Evidence suggests that mean annual temperatures were 7°–14°C warmer than at present [DeConto, 1996] and atmospheric CO₂ concentrations were 2–10 times greater than at present [Berner, 1997]. No significant ice at high latitudes is documented in the proxy record [Frakes et al., 1992], and fossil evidence suggests that the interiors of the midlatitude continents had mean annual temperatures as high as 23°C [Upchurch and Wolfe, 1987]. On the basis of the poleward expansion of the habitats of such organisms as coral reefs, belemnites (ancient relatives of the nautilus), and planktonic foraminifera preserved in ocean sediment cores, sea surface temperatures for the Cretaceous have been estimated to be similar to or slightly warmer than today at tropical latitudes, while ranging from 5° to 15°C warmer at high latitudes [Barron, 1983]. Oxygen isotope analyses of preserved specimens of benthic fauna such as inoceramids (extinct relatives of clams) and benthic foraminifera in deep-sea sediment cores suggest that the deep ocean may have been as warm as 5°–16°C during this period [Douglas and Savin, 1975; Arthur et al., 1979; Boersma and Shackleton, 1981; Saltzman and Barron, 1982; Barron et al., 1984; Barrera et al., 1987; Huber et al., 1995]. For this warm water to be dense enough to sink, it has been speculated that the deep waters were also quite saline.

[4] The first models to explain warm deep water were conceptual. In the early twentieth century, Chamberlin [1906] proposed subtropical and midlatitude marginal seas, with their high evaporation rates, as the formation sites of the Cretaceous bottom waters. Saltzman and Barron [1982] argue for both a low-latitude South Atlantic source and a high-latitude Pacific source based on highly variable deepwater temperatures inferred from isotopic analyses of Deep Sea Drilling Project (DSDP) cores.

[5] Ocean model simulations forced with atmospheric global climate model (GCM) results provided a first test of these conceptual models. Barron and Peterson [1990] used a primitive equation ocean model with coarse horizontal resolution (5°) and simplified convective parameterization and bathymetry. Atmospheric forcing was for mean annual conditions with atmospheric CO₂ at 1 and 4 times the present. Their simulations argue for formation sites both at high latitudes of the Pacific and in the subtropical Tethys, with the former being prominent at lower CO₂ levels and the latter becoming dominant at higher CO₂ levels. Bice et al. [1997] suggested that the subtropical Tethys source would be absent in a simulation in which runoff is included. Brady et al. [1998], using a finer resolution ocean model and bathymetry and seasonal atmospheric forcing, concluded that warm salty water, consistent with proxy data, could be formed by cooling at high southern latitudes. Using an idealized coupled ocean-atmosphere model, Schmidt and Mysak [1996] show that the polar and tropical source regimes are solutions depending on the polar temperature-salinity value.
[6] In this study, we employ the National Center for Atmospheric Research (NCAR) Climate System Model (CSM), a fully coupled atmosphere-ocean-sea ice model, with detailed parameterizations of physical processes. The CSM is a state-of-the-art climate model that includes detailed models of both the atmosphere and ocean. Long, coupled simulations without flux adjustments are possible with these simulations that were started with the full ocean spun up. The seasonal cycle is included. The atmosphere and ocean models have improved horizontal and vertical resolutions and detailed parameterizations compared to previous models. In addition, more detailed reconstructions of the bathymetry and topology are now available.

[7] Two questions are addressed: (1) How were the thermohaline circulations and regions of deepwater formation different than the present to explain proxy inferences of warm water at depth? (2) Do atmospheric CO₂ levels within the range proposed from proxy records and geochemical models affect formation sites and resulting circulations?

2. Model Description

[8] Simulations for this study are conducted with a version of the NCAR CSM, a global coupled atmosphere-ocean-sea ice-land surface model [Otto-Bliesner and Brady, 2001]. The atmospheric model is Community Climate Model 3 (CCM3), the latest version of the NCAR CCM. CCM3 is a spectral model with 18 levels in the vertical. For these experiments CCM3 is run at T31 resolution (an equivalent grid spacing of roughly 3.75° × 3.75°). Details of CCM3 are given by Kiehl et al. [1998]. The land surface model has specified vegetation types and a comprehensive treatment of surface processes [Bonacci and Gent, 1998]. Surface and subsurface runoff is computed in the land model but is not routed back to the oceans in the version of the CSM used for these simulations. Instead, freshwater balance is maintained with the precipitation scaling scheme described by Boville and Gent [1998]. The ocean model is the NCAR CSM Ocean Model (NCOM), with 25 levels, 3.6° longitudinal grid spacing, and latitudinal spacing of 1.8° poleward of 30° smoothly decreasing to 0.9° within 10° of the equator. NCOM uses the Gent-McWilliams eddy-mixing parameterization and a nonlocal K profile boundary layer parameterization. Details of NCOM are given by Gent et al. [1998]. Our version also uses a spatially varying horizontal viscosity scheme [Large et al., 2001] and an ocean background vertical diffusivity of 0.15 cm² s⁻¹ [Otto-Bliesner and Brady, 2001]. The sea ice model [Weaverly et al., 1998] includes ice thermodynamics based on the three-layer model of Semtner and ice dynamics based on the cavitating fluid rheology of Flato and Hibler. The grid spacing is the same as in the ocean model.

3. Forcing and Boundary Conditions

[9] Prescribed conditions for the model include land-sea distribution, elevation and bathymetry, solar luminosity, and atmospheric chemistry. The continental configuration is based on a new reconstruction for the Campanian (80 Ma) [Hay et al., 1999], and paleoshorelines and paleotopography are taken from DeConto et al. [2000]. The continents consist of three continental blocks, North America/Eurasia, South America/Antarctica/India/Madagascar/Australia, and Africa. The highstand of sea level results in shallow epicontinental seas flooding these blocks. Global land area is ~20% less than at present. Bathymetry is based on a reconstruction of Campanian crustal ages, an age-depth relationship, and a correction for sediment accumulation [DeConto et al., 2000]. The oceanic configuration consists of a large open Pacific basin, a wide eastern Tethys, and a circum-African seaway extending from the western Tethys through the subtropical North Atlantic and the South Atlantic Oceans. A high-latitude North Atlantic Ocean is not present during the Campanian. The Isthmus of Panama is open with depths of 400–700 m.

[11] Trace gas concentrations, except for atmospheric CO₂, are set to their preindustrial concentrations as estimated from ice core records [Ice Core Working Group, 1998; Fluckiger et al., 1999; Ingersoll et al., 1999]. Thus methane (CH₄) is set to 700 ppbv, and nitrous oxide (N₂O) is set to 275 ppbv. Chlorofluorocarbons included in present-day simulations did not exist before 1930 and are omitted in the Cretaceous simulation. Atmospheric CO₂ levels for the Campanian have been estimated from both proxy data and models, both with relatively large error bars. Berner [1994, 1997], using a model of the geochemical cycle applied at 1-myrr intervals, calculated levels to be 2 to 5 times the present. Yapp and Poths [1996], using carbon isotope data, estimate levels as high as 9 times the present. For the baseline Cretaceous simulation in this study, an atmospheric CO₂ concentration of 1680 ppmv; 6 times the present (preindustrial), is used. A sensitivity Cretaceous simulation with an atmospheric CO₂ concentration of 1120 ppmv, 4 times the preindustrial, is also presented.

[12] The solar constant is adjusted to be 0.8% less than at present based on theoretical studies of solar physics and models of solar evolution [Endal, 1981; Crowley and Baum, 1992]. Milankovitch cycles of the Earth’s orbital dynamics have been documented in records for the Pleistocene, but the periods and magnitude of this forcing before the Pleistocene are uncertain [Berger et al., 1989]. A circular solar orbit (eccentricity equal to zero) with present-day obliquity (23.4°) is therefore assumed.

[13] The specified vegetation distribution is based on a simulation of Campanian climate using an interactive vegetation model, EVE, with life forms [DeConto et al., 2000] adjusted to be compatible with fossil assemblages (R. DeConto, unpublished data, 1999). Tropical broadleaf evergreen forests cover a broad tropical latitudinal band. Deserts occur over the subtropical land areas of South America and Africa. Needleleaf evergreen and broadleaf deciduous trees populate the high latitudes. No tundra vegetation or land ice is prescribed for this time period.

[14] The simulation uses a spinup procedure modified from Otto-Bliesner and Brady [2001]. First, the atmospheric and land ocean models were integrated with 20 years, which coupled to a slab ocean model with diffusive heat transport [Thompson and Pollard, 1997] and to a thermodynamic sea ice model. Initial conditions for this run were meridionally varying ocean sea surface temperatures (SSTs) of 28°C at the equator and 10°C at the poles and no sea ice. This simulation provided both the initial conditions and forcing to the ocean/sea ice spinup. The ocean/sea ice spinup used SSTs initialized from the slab ocean predictions at the end of year 20. Subsurface ocean temperatures were initialized assuming a bottom-water temperature of 10°C and linear interpolation from the surface distribution. Initial surface salinities were set globally to 34.8 ppt. Daily values of the last 5 years of the atmospheric simulation provided the upper boundary forcing for the ocean/sea ice spinup. The ocean spinup simulation was run for 100 years with the deep ocean accelerated by a factor of 50 (i.e., 5000 deepwater years). Global ocean temperature drift over the last 40 years of the ocean spinup was on the order of 0.1°C per century. The fully coupled simulations were initialized from the atmosphere-land-ocean-sea ice spinups and were integrated for 130 years. No flux corrections are applied. The last 50 years are analyzed in this paper.

[15] Comparisons are made to a present-day simulation with present-day geography and bathymetry and solar constant. Trace gas concentrations for 1990 are used in the present-day simulations. That is, atmospheric CO₂ is set to 354.4 ppmv, CH₄ is set to 1722...
ppbv, N2O is set at 308.4 ppbv, and chlorofluorocarbons are included.

4. Proxy Estimates of Deepwater Temperatures

[16] Although Cretaceous sediments have been recovered from cores at DSDP and Ocean Drilling Project (ODP) sites in the Pacific and Atlantic Oceans since the 1970s, estimates of paleotemperatures at depth are still quite sparse and exhibit a wide range of values (Figure 1). To obtain paleotemperature estimates, analyses of the limited, suitably preserved, benthic foraminifera have been augmented with analyses of specimens of *Inoceramus*, an epibenthic bivalve that is now extinct but was abundant at bathyal depths [Douglas and Savin, 1975; Arthur et al., 1979; Boersma and Shackleton, 1981; Saltzman and Barron, 1982; Barron et al., 1984; Barrera et al., 1987; Huber et al., 1995]. These benthic fauna provide estimates of the ocean temperatures at depth through assumptions about the relationship of temperature and d18O. Many uncertainties limit the accuracy of these paleotemperature estimates, making them suggestive rather than indicative. The isotopic composition of seawater in the Late Cretaceous is unknown and is estimated from present-day tropical Pacific Ocean waters. The method assumes that the carbonate was deposited in isotopic equilibrium with the surrounding waters and that there was no alteration after burial. In addition, the depth habitat of the benthic fauna is difficult to estimate, as is exact dating.

[17] Published Late Cretaceous cores [Douglas and Savin, 1975; Arthur et al., 1979; Boersma and Shackleton, 1981; Saltzman and Barron, 1982; Barron et al., 1984; Barrera et al., 1987; Huber et al., 1995] give deep-ocean temperatures considerably warmer than at present (Figure 1). The two DSDP cores analyzed in the tropical Pacific suggest ocean temperatures at bathyal depths of 12.9°C – 15°C. The difference between the temperatures of surface waters and bottom waters in the tropical Pacific is less than one half of the present-day value [Douglas and Savin, 1975]. Deepwater temperatures estimated for the Atlantic region show a larger range, from 6.2°C – 15.2°C. In addition, benthic foraminifera preserved in exposed marine sequences on Seymour Island, near the coast of Antarctica, indicate bathyal ocean temperatures ranging from 4°C to 8.5°C [Barrera et al., 1987].

5. Atmosphere Results

[18] The annual mean surface temperatures simulated by the CSM for the Cretaceous (Figure 2a) are significantly warmer than at present, with the most pronounced warming over the high-latitude land areas. Isolated areas of subfreezing surface temperatures occur over the high-latitude interiors of the Cretaceous continents. The coldest simulated winter temperatures occur over northeast Eurasia, where December – February surface temperatures reach –20°C. Summer temperatures in the interiors of the Cretaceous high-latitude continents warm sufficiently to support trees in these regions. Only isolated strips along the Arctic Ocean remain cold enough throughout the year to suggest tundra vegetation. The amplitudes of the simulated seasonal cycles of middle- and high-latitude continental areas for the Cretaceous are reduced by 50% compared to the present.

[19] The simulated annual mean precipitation for the Cretaceous (Figure 2b) shows a tropical pattern and amounts similar to the present, with maxima of precipitation straddling the equator associated with the seasonal migration of the Intertropical Convergence Zone. Precipitation is enhanced at middle and high latitudes along the wintertime storm tracks as a result of warmer temperatures and thus greater atmospheric moisture capacity.

6. Ocean Results

6.1. Sea Surface Temperatures

[20] The CSM is able to reproduce the observed present-day pattern of SSTs (Figure 3a) with errors of <2°C except at sea ice
margins and in upwelling regions along the west coasts of South America, Africa, and North America, where the model is too warm by 2°–6°C [Boville and Gent, 1998; Otto-Bliesner and Brady, 2001]. This discrepancy is related to underestimates of upwelling by the ocean model and of marine stratus clouds by the atmospheric model [Boville and Gent, 1998]. The observed pattern of SSTs in the tropical Pacific of a warm pool in the west and a cold tongue in the east are reproduced by the CSM [Otto-Bliesner and Brady, 2001], although the latter is somewhat too intense and extends marginally too far westward, typical of many non-flux-corrected coupled GCMs. Sea ice fraction is overestimated in the Nordic seas.

[21] Enhanced atmospheric CO2 levels for the Campanian result in significantly warmer SSTs (Figure 3b). Zonal and annual average tropical SSTs are 3°–4°C warmer compared to the present day (Figure 4). Habicht [1979] documents a poleward shift in reef deposits suggesting a 10°–15° poleward expansion of warm SSTs to support this fauna [Barron et al., 1981]. Corals are presently found within 30° of the equator with the 21°C water temperature being the approximate limit for extensive carbonate deposition and reef development. The 21°C ocean isotherm is found at ~30° latitude in the present-day simulation and at 45° latitude in the baseline Cretaceous simulation.

[22] Similar to the present-day simulation, Cretaceous trade winds induce equatorial upwelling with SSTs of 30°C in the eastern Pacific and with warmer SSTs, >32°C, in the western Pacific and eastern Tethys regions. Note that these results suggest that cores in the eastern Pacific and western Tethys would be expected to preserve cooler SSTs than the zonal average (Figure 4). The simulated equatorial Pacific east-west SST difference

Figure 2. Annual mean atmospheric fields simulated by the CSM for the Cretaceous. (a) Surface temperatures (°C). Contour interval is 5°C. Temperatures below freezing are shaded. (b) Precipitation (mm/day). Contour interval is 2 mm/day. Light shading denotes precipitation amounts of >2 mm/day, and dark shading denotes amounts of >4 mm/day.
(160°E–100°W) is 4°C for both the Cretaceous and the present. This gradient is important for explaining El Niño-Southern Oscillation (ENSO) dynamics of the coupled atmosphere-ocean system.

SSTs decrease poleward but with reduced meridional SST differences between 30° and 60° latitude averaging 16°C in the Campanian simulation compared to 23°C in the present-day simulation. Cretaceous SST warming at high latitudes is 6°–14°C (Figure 4). The Arctic Ocean with its much restricted access is significantly colder (but above freezing) than corresponding southern latitude waters. No perennial sea ice is simulated for either the Northern or Southern Hemispheres. Small amounts of thin ice form in shallow areas along the polar coastlines of the northern continents and in one small bay off Antarctica during their respective winters, melting completely back during the summer months. Sea ice areas during the winter months are <10% of present-day values. Recall that no permanent land ice exists over the continent of Antarctica in the Campanian simulation.

6.2. Sea Surface Salinities

The CSM captures the observed present-day patterns of sea surface salinities, except in the equatorial regions of the Pacific and Indian Oceans (Figure 5). Salinities in excess of 35 ppt are simulated for the subtropical oceans in regions of net evaporation between 15° and 30° latitude. In the North Atlantic Ocean, salty water extends poleward into the Labrador, Greenland, and Norwegian Seas. At corresponding latitudes in the North Pacific Ocean, salinities are fresher in excess of 2 ppt, in agreement with observations. The major deficiencies in the CSM simulation of present-day salinities include erosion of the subtropical high-salinity regions in the Pacific and Indian Oceans, with the equa-
The geographical distribution of sea surface salinities during the Cretaceous is significantly different than that simulated for the present (Figure 5). Salinities greater than 35 ppt occur in the subtropical and middle latitude Cretaceous oceans from 20°–45° latitude. Two ocean regions have very salty waters in excess of 38 ppt, the Tethys Ocean just west of Asia and the narrow South Atlantic Ocean. Extensive salt deposits off the west coast of Africa and the east coast of South America dated to the Cretaceous [Haq, 1984] agree with the very saline waters (>38 ppt off Africa) simulated by the CSM. Compared to the present day, the North Pacific basin also experiences higher salinities, especially on its western periphery. The narrow North Atlantic basin is quite fresh, as is the Arctic basin.

6.3. Thermohaline Circulations

The meridional overturning stream function in the model is calculated from the Eulerian mean velocity and the parameterized eddy-induced transport velocity. The CSM simulates a present-day, annual mean global meridional overturning (Figure 6a) that captures the observed structure with a large overturning circulation in the Northern Hemisphere underlain by bottom water with a southwestern ocean source region. The Northern Hemisphere overturning circulation reaches a maximum of 30 Sv at 35°N–40°N and a depth of 1 km, with this overturning primarily in the North Atlantic. This cell reaches only 3 km, a common problem in coordinate ocean models, and is ~50% too strong compared to the observed, a problem in the CSM related to the absence of high-latitude river inflow and excessive sea ice formation [Doney et al., 1998]. In the Southern Hemisphere a narrow overturning circulation is found at 50°S near the surface. This circulation reverses at depth, with Antarctica bottom water (~5 Sv) predominating in both hemispheres at depth.

The thermohaline circulations simulated by the CSM for the Campanian are significantly different than at present (Figure 6b). A large overturning cell extending from the surface to depths of 5 km dominates each hemisphere. Sinking in the Northern Hemisphere occurs at 65°N and is in excess of 30 Sv. The Southern Hemisphere cell is the stronger of the two, in excess of 50 Sv with sinking off the coast of Antarctica at 65°S. Both cells extend further poleward than the simulated present-day North Atlantic cell.

Vertical mixing due to buoyancy and mechanical forcing in the ocean model is parameterized by the K profile parameterization (KPP) boundary layer scheme of Large et al. [1994]. Sites of deepwater formation are noted by wintertime boundary layer depths in excess of 500 m, as shown in Figure 7. The boundary layer depth is the depth to which eddies formed by turbulent stresses and buoyancy forcing can reach in the presence of vertical shear and stratification and is diagnosed as the depth where the bulk Richardson number reaches a critical value. Because deepwater formation is a sporadic process with variability on interannual and longer timescales, Figure 7 shows a maximum boundary layer depths found in either the January monthly mean for the Northern Hemisphere or the July monthly mean for the Southern Hemisphere over the 50-year time period examined here.

For the present day in the Northern Hemisphere, the deepest boundary layer depths are found in three locations within the subpolar gyre of the North Atlantic Ocean. Killworth [1983] found that deep open-ocean convection occurs within cyclonic flow because within cyclonic gyres, vertical stability of the water column is reduced. The deepest mixing occurs in the Norwegian Sea at depths exceeding 2500 m. The other two locations, the Labrador Sea and south of the Denmark Strait, exceed depths of 1500 m. These locations agree well with observations. The only other notable area of deep mixing occurs in the Mediterranean Sea. In the Southern Hemisphere the deepest boundary layer depths occurring in July are found within the cyclonically flowing Weddell Sea gyre (as shown in Figure 7a) and exceed 1500 m. Deep mixing is interannually sporadic in this location and occurs in association with openings in the wintertime ice pack similar to observed polynas. The other locations of lesser importance are in the Ross Sea and west of the Antarctic Peninsula.

In the Cretaceous the wintertime maximum boundary layer depths are deeper than what is noted for the present day, but all occur within cyclonic subpolar gyres. In the Northern Hemisphere the deepest mixing occurs in a rather large region along the northwest boundary of the North Pacific within the large subpolar gyre and reaches depths greater than 4000 m. In the Southern Ocean the deepest mixing exceeds 3000 m and occurs within the cyclonically flowing gyre located south of Africa and India and west of Australia.

The annual average barotropic stream functions for the equilibrium present-day and Cretaceous states are shown in Figures 7a and 7c. Mass transport by an Antarctic Circumpolar Current, a dominant feature in the simulation of the present climate, is precluded in the Cretaceous by the attachment of Australia to Antarctica. An anticyclonic subtropical gyre in excess of 60 Sv occurs in the Pacific sector between 20° and 40° latitude in the Cretaceous. The Cretaceous North Atlantic is too narrow and shallow to support the formation of gyre-like circulations. The South Atlantic subtropical gyre is found more poleward, centered at 45°S, with transports in excess of 40 Sv. A large subpolar gyre occurs in the North Pacific Ocean extending northeastward from the Asian coast at 45°N and with transports greater than 30 Sv. This gyre has an intensity 3 times greater than the Pacific subpolar gyre simulated for present-day conditions and is similar to the present-day North Atlantic subpolar gyre.
Figure 8 compares latitude-depth cross sections of temperature and salinity at selected longitudes for the present-day and Cretaceous simulations. North Atlantic deep water is formed in the Labrador and the Greenland, Iceland, and Norwegian Seas in the present-day simulation. This overturning is evident in the temperature and salinity cross sections at 30°W longitude. The 6°C isotherm slopes downward from the surface to 2 km from south to north. The steep isotherms and isohalines north of 55°N suggest concentrated sinking of water that is 8°–10°C and 35.8 ppt at the surface. This warmer and saltier deep water flows to the south, filling the upper 3 km of the Atlantic to 50°S. Below 3.5 km, colder, fresher Antarctic bottom water predominates. While there is no evidence in the section at 180° that deep water is made in the North Pacific at the present day, the shallow equatorward penetrating tongues of fresh water suggest that an intermediate water mass is being formed in both hemispheres. Surface waters poleward of 45°N are significantly cooler and fresher than corresponding waters in the North Atlantic. Relatively flat isotherms and isohalines below 500 m depth indicate little convection of heat or salt vertically in the Pacific sector. Bottom temperatures at 180° average 1°C with salinities of 34.6 ppt.

Temperature-salinity cross sections for the Cretaceous show a significantly different structure (Figure 8). A large deep overturning cell in the Cretaceous South Atlantic is evident in the cross section at 30°W by the steep isotherms and isohalines appearing poleward of 60°S. This overturning results in a penetration of the warm, salty surface waters to depths of 2–3 km. Bottom temperatures in the South Atlantic average 9°C. The North Pacific overturning cell is apparent in the cross section at 150°E. The downturn of the 12°C isotherm and 34.6-ppt salinity contour indicate this cell. Bottom temperatures in the Pacific are warmer than present, averaging 11°C.

6.4. Poleward Ocean Heat Transport

The present-day simulation shows an asymmetric pattern of poleward ocean heat transport (Figure 9). Northward transport...
occurs from 5°S to the North Pole with maximum transport of 2.2 PW at 15°N as a result of the strong North Atlantic overturning cell. This concurs with the estimates of Trenberth and Solomon [1994]. The southward transport in the Southern Hemisphere is much weaker, only 0.4–0.5 PW. The double maxima structure matches the observational estimate of Trenberth and Solomon, but the model underestimates the subtropical transport in the Southern Hemisphere.

[35] The Cretaceous simulation has a more symmetric pattern of poleward ocean heat transport associated with negligible cross-equatorial transport of heat by the oceans. Maximum transport in the Northern Hemisphere decreases to 1.2 PW and shifts to 30°N. Maximum transport in the Southern Hemisphere occurs at 30°S and at 1.7 PW is 4 times greater than in the present-day simulation. Note that the meridional SST gradients are only 70% of the present in the Cretaceous simulation. Atmosphere-slab ocean simulations of the Cretaceous have often invoked increased ocean heat transport in an attempt to explain high-latitude warmth during the Cretaceous [Otto-Bliesner and Upchurch, 1997; DeConto et al., 2000]. The CSM coupled atmosphere-ocean results suggest that the climate system is more complex, with high-latitude ocean convection being an integral component of the Cretaceous warmth.

7. Sensitivity to Atmospheric CO₂

[36] Significant uncertainties exist in the estimates of Cretaceous levels of atmospheric CO₂. For this reason an additional sensitivity simulation with atmospheric CO₂ reduced to 1120 ppmv, 4 times the preindustrial values, was run to assess the sensitivity of the ocean structure and circulations has on our ability to prescribe this parameter.

[37] Figure 10 shows the meridional overturning stream function, sea surface salinity and boundary layer depth distribution, and the Atlantic latitude-depth cross section of temperature for this simulation. The salinity and boundary layer depth distributions in the North Pacific Ocean are similar in the two experiments. The resulting meridional overturning stream functions at northern latitudes are also therefore similar. The Southern Hemisphere cell is reduced by 35% compared to the baseline Cretaceous simulation. This is a result of reduced poleward transport of salty waters in the South Atlantic. At 45°S in the South Atlantic, salinities are ~33.5 ppt in the 6xCO₂ simulation compared to ~35.5 ppt in the 4xCO₂ simulation. Boundary layer depths show reduced formation site area in the South Atlantic with reduced CO₂. With less penetration of warm surface water, middepths (1–3 km) are significantly cooler (8°–10°C versus 10°–14°C) at these high latitudes in the 4xCO₂ simulation compared to the 6xCO₂ simulation.

8. Discussion and Conclusions

[38] To summarize, the coupled atmosphere-ocean simulations with the NCAR CSM for the Late Cretaceous exhibit large overturning cells in both hemispheres extending from the surface to the ocean bottom and with intensity comparable to the present-day North Atlantic simulated overturning. In the Northern Hemisphere the sinking takes place in the Pacific due to cooling of the much warmer and saltier waters compared to the present day. In the Southern Hemisphere the sinking occurs primarily in the southern...
Figure 7. Annual mean barotropic stream function (Sv) and winter maximum boundary layer depth (m) simulated by the CSM for the present and the Cretaceous. Contour interval is 10 Sv for barotropic stream function and 500, 1000, 2000, 3000, and 4000 m for boundary layer depths, with light shading indicating depths of >500 m.
Atlantic and Indian Oceans. For a simulation with atmospheric CO₂ reduced from 6 times to 4 times the preindustrial concentrations, this southern branch is reduced by 35% due to less poleward transport of salty waters in the South Atlantic Ocean. 

Simulated deep-ocean temperatures are significantly warmer than at present. 

[39] Oxygen isotopic paleotemperatures estimated from benthic foraminifera and inocerod bivalves in Cretaceous sediments from

![Figure 8. Latitude-depth cross sections of annual mean ocean temperatures (°C) and ocean salinities (ppt) simulated by the CSM. (a) Present day, 30°W longitude. (b) Present day, 180° longitude. (c) Cretaceous, 30°W longitude. (d) Cretaceous, 150°E longitude. Left-hand column is ocean temperature with a contour interval of 2°C. Right-hand column is ocean salinity with a contour interval of 0.2 ppt.](image-url)
Northward Heat Transport - Annual Mean

Figure 9. Global and annual average northward ocean heat transports simulated by the CSM.

the Deep Sea Drilling Project show the Late Cretaceous deep ocean to be significantly warmer than at present (Figure 1). Site 530A at 35°S in the South Atlantic suggests ocean temperatures at a paleodepth of 4 km of 15.3°C for the late Campanian. This temperature estimate is comparable with deepwater temperature reconstructions at site 511 farther poleward but is significantly warmer than temperatures of 6.2°–9.5°C from inocerods from site 355 located at 25°S. The corresponding temperatures simulated in the baseline Cretaceous are 9°–10°C. Tropical Pacific sites at paleodepths of 2–3 km indicate warm deep-ocean temperatures of 12.9°–15°C, which are reasonably comparable to those predicted in the model (10°–12°C).

[40] Results from the simulation with 4xCO2 may explain the cooling implicated for deep waters in the South Atlantic [Barrera and Savin, 1999] from the late Campanian (75 Ma) to the end of the Cretaceous (65 Ma). Estimates of atmospheric CO2 [Berner, 1997] show a decrease in concentrations during this period. With a reduction in the southern overturning, the model southern deep ocean cools.

[41] CSM Cretaceous tropical temperatures are 31°–33°C, a warming of 3°–4°C compared to simulated present-day values. The simulated Cretaceous temperatures are on the warm end of the traditional range estimated from the limited samples of shallow-dwelling foraminifera. Faunal records have been used to suggest that Cretaceous low-latitude SSTs were generally no warmer and possibly cooler than in the present day [Crowley and Zachos, 2000]. Considerable uncertainty (2°–3°C) in these estimates exists due to diagenesis, metabolic effects, analytical errors, and terms in the oxygen isotope transfer function [Crowley and Zachos, 2000]. Poulsen et al. [1999] reinterpret Cretaceous shallow water temperatures inferred from core data. When they account for spatial variability of isotopic composition of seawater and the paleohabits of the marine organisms, the inferred tropical SSTs are warmer and in better agreement with Cretaceous model simulations with elevated atmospheric CO2. New analyses of ODP core 1052 in the western tropical Atlantic show a maximum SST that is 3°–5°C warmer than today and with pronounced temporal variability [Wilson and Norris, 2001]. Similar to the present day, Cretaceous trade winds would be expected to induce equatorial upwelling with cooler SSTs in the eastern Pacific and warmer SSTs in the western Pacific and eastern Tethys regions. The larger number of ODP cores in the eastern Pacific and western Tethys would thus be expected to preserve cooler SSTs than the zonal average.

[42] The CSM Cretaceous simulations may overestimate the strength of the high-latitude thermosteric circulations as a result of the simulated atmospheric state. The river runoff not being routed back to the oceans affects the salinity distributions simulated by the model. The precipitation scaling scheme achieves global freshwater conservation by fractionally increasing the global precipitation rate and thus redistributing the river discharge to high-rainfall regions. Thus, in the model, regions of high precipitation are freshened compared to regions of low precipitation. The absence of river inflow to the high-latitude oceans results in the simulated present-day North Atlantic being saltier and the North Atlantic overturning being greater than the observed. A present-day simulation observed a newer version of the CSM, which gives reduced North Atlantic overturning and values similar to those observed. If the simulated high-latitude Cretaceous precipitation were allowed to run off into the high-latitude Pacific and southern oceans, the Cretaceous overturning cells would be weaker but still present. This is consistent with the results of Bice and Marotzke [2001], who show that even when subtropical evaporation and high-latitude precipitation are increased in a series of ocean sensitivity simulations, bottom water is formed at high latitudes and not at subtropical latitudes. The cold, winter, high-latitude land surface temperatures simulated for the Cretaceous also enhance the oceanic convection through advection of this colder air over the warmer adjacent oceans.

[43] Earlier ocean-only model simulations gave mixed results in terms of deepwater formation sites. Barron and Peterson [1990] found a North Pacific bottom-water formation site in a Cretaceous ocean simulation forced with present-day atmospheric CO2. The eastern Tethys became a site for deepwater formation for a warmer climate produced by 4xCO2. The models that were available limited their results at the time. The ocean model had a coarse 5° × 5° horizontal resolution and only four vertical levels. Mesoscale eddy effects were accounted for by diffusion processes, and vertical mixing was parameterized simply as a function of static stability. In addition, the atmospheric forcing was for annual means. Using an ocean model with improved horizontal and vertical resolution and seasonal atmospheric forcing from Global Environmental and Ecological Simulation of Interactive Systems (GENESIS), Brady et al. [1998] found only a high-latitude Southern Hemisphere site for deepwater formation. Both these results were for fixed atmospheric forcing and did not allow for ocean-atmosphere feedbacks in determining the climate.

[44] Only one other coupled atmosphere-ocean simulation has been published to date for the Cretaceous: Bush and Philander [1997]. Their Cretaceous simulation was started from the present-day observed ocean temperatures and salinities of Levitus [1982] and ran for only 32 years. This is too short of a simulation to draw any conclusions on the meridional overturning, as the deep ocean takes in excess of 1000 years to equilibrate. Surface ocean conditions did equilibrate in their simulation. Equatorial SSTs in our simulation for 4xCO2 warm by ~2.5°C compared to 3.5°C in their simulation, and the extratropical oceans warm by 3°–9°C in our simulation compared to an average of 5.2°C in theirs. Highest salinities in both simulations are found in the Gulf of Mexico, off the west coast of South America, and in the narrow South Atlantic. Salinities are also greater in the North Pacific than in the present day. These comparable results suggest that the Geophysical Fluid Dynamics Laboratory coupled atmosphere-ocean model would give similar meridional overturning circulations to ours (except
Figure 10. Sensitivity of ocean simulation to atmospheric CO₂ concentration reduced to 1120 ppmv. (a) Annual mean meridional overturning stream function (Sv). Contour interval and light shading are as in Figure 6. (b) Winter maximum boundary layer depth (m). Contour interval and light shading are as in Figure 7. (c) Annual sea surface salinity (ppt). Contour interval and shading are as in Figure 5. (d) Latitude-depth cross section of annual ocean temperature at 30°W longitude. Contour interval is as in Figure 8.
in the North Atlantic, which is much wider in their configuration) if their simulation was extended.

[45] The Cretaceous simulation has reduced meridional gradients compared to the present. Although increased poleward oceanic transport has been proposed as a mechanism to explain warmer SSTs at high latitudes [Schmidt and Mysak, 1996], reduced poleward SST gradients make this difficult to accomplish [Sloan et al., 1995]. Ocean-only simulations [Bradly et al., 1998; Haupt and Seidov, 2001] suggest that the proxy record can be explained without resorting to increased ocean heat transport. These simulations lack the interactions with the atmosphere. In this study, fully coupled Cretaceous simulations have maximum Northern Hemisphere poleward oceanic transport reduced by a factor of 2 with Southern Hemisphere transport enhanced fourfold. Cyclonic wind-driven circulation and moderately high salinities at high latitudes of the North Pacific and southern oceans, combined with cool advection over these waters from adjacent continental areas, drive deepwater formation and two hemispheric overturning cells. The CSM is thus able to create a self-consistent ocean system: zonal SST gradients, ocean heat transport, and deep-ocean temperatures that agree with available proxy data.

[46] Our results suggest that low-latitude sites of deep convection are not required to maintain a general Cretaceous climate state of warm deep and polar water. Instead, warm saline deep water can be formed by deep convection at high latitudes in a process quite similar to deepwater formation in the present-day North Atlantic. The warmth of the deep waters comes from the warm polar surface waters under conditions of high atmospheric CO2. The strength of the overturning is dependent on the advection of high-salinity water from midlatitudes. Surface circulation features (and evaporation) that depend on the amount of atmospheric warming influence this advection.

[47] Other climate models have different sensitivities, so it will be desirable to compare our simulations with future long coupled atmosphere-ocean simulations of warm Cretaceous climates, as they become available. Uncertainties in the boundary conditions and forcing, paleogeography, bathymetry, and trace gas concentrations need to be further evaluated. Despite these uncertainties, our results suggest that higher levels of atmospheric CO2 and the altered paleogeography of the late Cretaceous result in a surface ocean state, temperature, salinity, and circulation, significantly different than the present. This, in turn, results in deepwater features, although formed by mechanisms the same as at present, that are quite different than the present. Warm waters inferred from proxy data in deep-sea cores can be explained by high-latitude sites of overturning. Future work will examine the presence, intensity, and frequency of present climatic modes such as ENSO for a world with an extreme warm background state.

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References


Ice Core Working Group, Ice core contributions to global change research:


