Simulated warm polar currents during the middle Permian

A. M. E. Winguth,1 C. Heinze,2 J. E. Kutzbach,1 E. Maier-Reimer,2 U. Mikolajewicz,2 D. Rowley,3 A. Rees,3 and A. M. Ziegler3

Received 9 April 2001; revised 12 April 2002; accepted 14 May 2002; published XX Month 2002.

[1] During Permian stage 6 (Wordian, Kazanian) the Pangaean supercontinent was surrounded by a superocean: Panthalassa. An ocean general circulation model has been coupled to an atmospheric energy balance model to simulate the sensitivity of the Wordian climate (~265 million years ago) to changes in greenhouse gas concentrations, high-latitude geography, and Earth orbital configurations. The model simulates significantly different circulation features with different levels of greenhouse gas forcing, ranging from a strong meridional overturning circulation in the Southern Hemisphere at low CO2 concentration (present level) to more symmetric overturning circulation cells with deep water formation in polar latitudes of both hemispheres at high CO2 concentration (8 times present level). The simulated climate with 4 times present level CO2 concentration agrees generally well with climate-sensitive sediments and phytogeographic patterns. The model simulates strong subtropical gyres with similarities to the modern South Pacific circulation and moderate surface temperatures on the southern continent Gondwana, resulting from a strong poleward heat transport in the ocean. An even more moderate climate is generated if high-latitude land is removed from the southern continent so that ocean currents can penetrate into the polar regions of Gondwana.


1. Introduction

[2] The long-term climate trend during the late Paleozoic era is documented in biogeographic provinces, climate-sensitive sediments and organic productivity (e.g., coral reefs, evaporites, oil source rocks, coals, eolian sands, IRD, and glacial tillsites) [Crowley and North, 1991; Ziegler et al., 1998]. These climate proxies indicate a significant climate transition from an “icehouse world” with extensive continental ice sheets during the Carboniferous (~330 Ma) to a generally ice-free state beginning in the Wordian (267 to 264 Ma). The greatest mass extinction in the Phanerozoic occurred at the Permo-Triassic boundary around 252 Ma, to 264 Ma). The greatest mass extinction in the Phanerozoic occurred at the Permo-Triassic boundary around 252 Ma, and several hypotheses have focused on the possible role of changes in the ocean, including a possible stratified ocean with an anoxic deep sea [Sepkoski, 1995; Isozaki, 1997]. The large Late Paleozoic-Early Mesozoic climate changes [Parish, 1998] were accompanied by large tectonic reorganization of the continents. The land-sea distribution during the Permian (~290 to 252 Ma years ago) was characterized by a supercontinent Pangaea with an almost closed meridional barrier [Seyfert and Sirkin, 1979; Ziegler et al., 1979], preventing a circumpolar current in contrast to the modern [Mikolajewicz and Crowley, 1997]. In addition, incoming solar radiation was decreased compared to modern. Simulations with energy-balance climate models have demonstrated the importance of reduced solar radiation (~3% lower than the modern value) and low atmospheric CO2 concentration for the buildup of continental ice sheets during the Carboniferous and Early Permian [Crowley and Baur, 1992]. Several studies indicate that the retreat of the ice sheets during the Middle to Late Permian might be partially explained by an increased atmospheric CO2 concentration, from a value comparable to the modern level (1 × CO2) ~300 Ma to a quadrupled value (4 × CO2) around 265 Ma [Crowley and Baum, 1992; Berner, 1994, 1997; Fawcett, 1994; Barron and Fawcett, 1995]. Comparison of the Wordian climate derived from floral and lithological data with simulations from the GENESIS 2 climate model indicate the fit is improved if the model is forced with 4 × CO2 concentrations or higher rather than 1 × CO2 [Gibbs et al., 2002; Rees et al., 2002]. Variations of the Earth’s orbital parameters could also significantly change the seasonal range in temperature at orbital timescales. Gibbs et al. [2002] showed that a combination of low CO2 levels (such as present-day levels) and a cold summer orbital configuration produces expanded areas of permanent snow cover.

[3] Ziegler [1998] suggested that the polar warming and a melting of the southern ice sheet in the Mid to Late Permian could have been triggered by a northward migration of Pangaea. Increased poleward heat transport of warm polar currents might have an important effect on the warming, and this has been documented by an ocean model simulation with idealized land-sea distribution [Kutzbach et al., 1990]. The importance of oceanic heat transport for a climate of a large supercontinent has also been discussed in previous climate studies for Paleozoic times. For example, ocean...
model simulations for the geographic configuration of the Late Ordovician (~460–440 Ma) indicate an up to ~42% increase in the poleward ocean heat transport in the Southern Hemisphere relative to present day [Poussard et al., 1999].

[4] The purpose of this study is to use climate models to explore the role of changing radiative forcing and changed land-ocean distribution on Wordian climate and ocean circulation. In particular we address the following questions: (1) What is the ocean circulation and oceanic heat transport for a middle Permian (Wordian) land/ocean distribution and idealized bathymetry in a dynamic ocean model? (2) How sensitive is the Wordian climate and ocean circulation to changes in the atmospheric CO₂ concentration? (3) How does a change in land configuration near the southern pole influence the character of polar ocean currents, and how do these ocean currents affect the continental temperatures over Pangaea? (4) How well do the simulated continental and oceanic conditions agree with the geologic evidence? For this purpose, model simulations are compared with climate-sensitive sediments [Ziegler et al., 1998] and with phytogeographic patterns [Rees et al., 2002].

2. Model Description

2.1. Coupled Model

[5] The coupled model consists of two components, an atmospheric energy balance model and an ocean general circulation model (Figure 1). The use of an energy balance model for the atmosphere reduces significantly the computational costs and complexity in comparison to a fully coupled model for the atmosphere reduces significantly the computational costs and complexity in comparison to a fully coupled model. The coupled atmospheric energy balance model (EBM) and large-scale geostrophic (LSG) ocean general circulation model, described in detail by Mikolajewicz [1996], has been applied to paleoclimatic studies [Mikolajewicz and Crowley, 1993; Mikolajewicz and Crowley, 1997]. The nonlinear EBM is two-dimensional (longitude-latitude), having a single (well-mixed) vertical dimension, and is comparable to the model of North et al. [1983]. The atmosphere is dry and the only prognostic quantity is air temperature. The energy budget equation is solved with a time step of one day for surface air temperature T_{air} and incorporates incoming solar radiation F_{sw}, outgoing long-wave radiation F_{lw}, and CO₂ radiative forcing F_{CO₂}:

\[
\frac{\partial T_{air}}{\partial t} + \nabla \cdot \mathbf{U} T_{air} + A_h \nabla^2 T_{air} = F_{sw} - F_{lw} + F_{CO₂} + \frac{(T_{sw} - T_{air})}{
\rho_{air} c_p H} \frac{H}{
\rho_{air} c_p H}
\]

The incoming solar radiation is either absorbed at the surface, reflected (surface albedo is a function of the ground temperature), or scattered/absorbed by prescribed clouds, ozone, and dust. The horizontal atmospheric transport in the EBM is parameterized as advection by the zonally averaged zonal component of the near-surface wind speed (U) described by the second term on the left side of (1) and as diffusion described by the third term on the left side of (1). A_h is the horizontal diffusion coefficient (2.4 \times 10^6 m s⁻¹), and \nabla^2 is the two-dimensional horizontal Laplace operator. The surface wind speed is computed for the Wordian from a GENESIS 2 climate model [Thompson and Pollard, 1997] simulation (Figure 1) and for modern from the European Centre for Medium-Range Weather Forecasts (ECMWF) reanalysis 1979–1993. The heat fluxes between the EBM and ocean model are used to determine the distribution of temperature and sea ice in the mixed layer of the ocean, which is also integrated with a time step of one day. The resulting surface conditions serve as the lower boundary conditions for the atmosphere. The atmosphere-ocean heat exchange uses a coupling coefficient \kappa of 40 W m⁻² K⁻¹ to simulate a strong damping for small-scale temperature changes and is reduced over areas of sea ice coverage. The conductive heat flux through sea ice is inversely proportional to the ice thickness and is proportional to the difference between the air temperature and the sea surface temperature (T_{ss}) at air temperatures below freezing point. For the specific heat of dry air at constant pressure \rho_{air} and the atmospheric scale height of 7.6 km. Over land (not shown in equation (1)) a surface energy balance equation predicts surface temperature with the assumption of zero heat capacity in soil.

[6] The Hamburg large-scale geostrophic ocean general circulation model (hereafter LSG) [Matier-Reimer et al., 1993; Mikolajewicz and Crowley, 1997] is coupled to the EBM via the surface heat exchange (Q_{H}), the last term on
the right hand side of equation (1), and is also forced by
wind stress (τw, τs) and freshwater flux (Q_W) (precipitation
minus evaporation plus river runoff; Figure 1) from 4 ×
CO₂ and 8 × CO₂ Wordian simulations of Gibbs et al.
[2002], which used the GENESIS 2 climate model [Thom-
pson and Pollard, 1997]. In addition, two supplementary
experiments, 1 × CO₂ and 2 × CO₂, with the same Wordian
land configuration in GENESIS 2 have been used to provide
additional prescribed values of τs, τw, U, and Q_W to explore
the sensitivity to low atmospheric CO₂ concentrations.

The LSG uses a time step of 15 days, a spatial
resolution of 3.5° × 3.5° horizontally and 22 layers verti-
cally, and allows computationally efficient long-term model
integrations of several thousand years. The LSG uses two-
dimensional fields of heat fluxes (Q_H), freshwater fluxes
(Q_w), and wind stress (τ_w, τ_s; Figure 1) to predict three-
dimensional fields of the ocean currents, potential tempera-
ture, salinity, tracers, and two-dimensional fields of sea
level, and sea ice thickness. The LSG with 22 vertical layers
includes an explicit formulation of vertical diffusion for
temperature and salinity (diffusivity ranging from 5 ×
10⁻³ m² s⁻¹ at the surface to 2 × 10⁻³ m² s⁻¹ for the
intermediate and deep ocean), which is reduced in the case of
very stable vertical stratification by a factor a/[a + (6p/εz)²]²
with a = 4 × 10⁻⁷ kg² m⁻⁴. The selection of the diffusion
scheme has been tested by various sensitivity experiments to
reproduce the observed modern radiocarbon distribution in
the ocean [Maier-Reimer et al., 1993; Mikolajewicz, 1996]
and the deepening of the mid to high-latitude thermocline
due to winter mixing. Results from these experiments
indicated that the mixing of the deep sea significantly
increases with an increase in the diffusion coefficients.

The thermodynamic sea ice model from Maier-Reimer
et al. [1993] has been modified to allow not only ocean
driven transport of sea ice due to advection by the mean
surface ocean currents predicted by the LSG, but also a
wind-driven transport with an assumed equilibrium velocity
proportional to the surface wind speed.

A simple river runoff model with a bucket depth of 30
cm is used to transport the net precipitation from the land
surface to the coastal ocean areas. The model is initialized
for all Wordian experiments with a mean salinity of 34.3
and a mean temperature of about 4°C (world ocean’s
average mean temperature today). The reduction in salinity
by 0.5 relative to the present-day value is based on the
assumption that water stored in the form of ice, at modern
times essentially in Antarctica and Greenland, was probably
not present during the Wordian.

Because the experimental design allows for forcing
LSG with two-dimensional wind stress and freshwater flux
from the companion GENESIS 2 Wordian simulations, this
study is a considerable improvement over studies that use
modern surface forcing for paleo experiments. However,
there is still the important limitation that the wind stress and
freshwater flux forcing is one-way, i.e. it does not allow for
coupled interaction. Only the sensible heat exchange (Q_H)
is a two-way coupling (Figure 1). Fully coupled ocean-atmos-
phere paleoclimate simulations show important two-way
interactions in polar regions [Schiller et al., 1997] and in
the tropics [Shin, 2002]. We plan to test the effect of these
limitations by studying feedback mechanisms between heat,
fresh water, and momentum fluxes in future fully-coupled
Permian climate simulations with a refined spatial and
temporal resolution.

2.2. Topography and Land-Sea Distribution

The paleogeographic basemaps for our simulations,
including land-sea distribution, bathymetry and land surface
elevation are taken from the Paleogeographic Atlas Project
of the University of Chicago [Ziegler et al., 1997] (see http://
gpap.uchicago.edu). Continental positioning is a crucial
boundary condition for the atmospheric and oceanic circu-
lation. The ocean circulation also depends on the mean depth
of the ocean [Bice, 1997] and on the bathymetry, which
is poorly known for the middle Permian. We assume therefore
a flat bottom bathymetry with a depth of 4000 m based on
the mean depth of the modern ocean. A smoothed transition
from the continental shelves to the deep ocean is derived
from the modern bathymetry (Figure 2a). The deep Tethys is
cut off from the Panthalassa Ocean.

The Permian paleogeography is reasonably well
known, except for the position of islands in the Tethys
Sea which belong nowadays to Southeast Asia. Rees et al.
[1999, 2002] discussed the uncertainties of the positions of
landmasses for these regions. The altitude of Pangaea
mountain chains has a relevant impact on the climate
simulations [Otto-Bliesner, 1993, 1998] and has been speci-
fied for the atmospheric model simulations [Gibbs et al.,
2002]. The estimates of the orography are based on litera-
ture on geochronology, tectonics, sedimentary provinces,
paleo-volcanism, and paleogeology. The Wordian continent
in our baseline experiment is extended to the pole because of
the coarse resolution model grid. Sensitivity experiments
explore the effect of circumpolar currents in the Northern
Hemisphere and in the Southern Hemisphere (4 × CO₂N
and 4 × CO₂S) when the land is removed between 70° to 90°
northern (N) and southern (S) latitude (see section 3.3).

2.3. Solar Luminosity and Orbital Parameters

The solar luminosity is thought to have been steadily
increasing since formation, a consequence of conversion of
hydrogen to helium. According to calculations by Boot-
hroyd [see Caldeira and Kasting, 1992], we chose for the
Wordian a reduction of 2.1% relative to the present-day
solar constant of 1365 W m⁻². Decreases of the Earth’s
rotation rate with time have been estimated from paleonto-
logical analysis [Scrutton, 1978] and tidal analysis [Wil-
liams et al., 1997]. However, changes from the Permian to
present are smaller than 3% and are neglected in the
atmospheric and ocean model simulations.

Previous studies [Crowley et al., 1992] indicate that
orbital variations could be significant on a supercontinent.
To study the sensitivity of the coupled model to orbital
parameters we use the same three configurations as Gibbs et al.
[2002]: For our reference run we assume the Earth’s orbit
about the sun was circular (eccentricity = 0) and that the
Earth’s obliquity (axial tilt) was 23.5°. This setting causes
an equal receipt of solar insolation for both hemispheres.
Two additional extreme experiments with a hot and cold
summer orbit were carried out. Similar to Crowley and
Figure 2. (a) Model bathymetry and (b) annual average of modeled sea surface temperature, (c) surface air temperature in °C, and (d) air-to-sea heat flux in W m⁻² (stripple indicates flux to the atmosphere) for the Wordian climate with 4 × CO₂ (contour interval for Figures 4b–4d is 2°C, 10°C, and 25 W m⁻², respectively).
the Wordian climate to different orbital configurations (cold summer orbit CSO and hot summer orbit HSO) is studied by using experiments. Present-day 

<table>
<thead>
<tr>
<th>Experiments</th>
<th>$T_{oa}$, °C</th>
<th>SST, °C</th>
<th>SAT, °C</th>
<th>$A_{aice(H)}$, %</th>
<th>$A_{aice(S)}$, %</th>
<th>$V_{ice}$, $10^{12}$ m$^3$</th>
<th>$Q_{TA}$, W m$^{-2}$</th>
<th>$T_{oa}$ at 30°S, PW</th>
<th>$T_{oa}$ at 30°N, PW</th>
</tr>
</thead>
<tbody>
<tr>
<td>Present-day</td>
<td>5.0</td>
<td>19.5</td>
<td>12.5</td>
<td>8.5</td>
<td>3.8</td>
<td>6.4</td>
<td>0.75</td>
<td>$-1.45$</td>
<td>1.38</td>
</tr>
<tr>
<td>Wordian 4 × CO$_2$</td>
<td>7.0</td>
<td>21.0</td>
<td>15.3</td>
<td>0.0</td>
<td>0.0</td>
<td>0.0</td>
<td>0.09</td>
<td>$-1.78$</td>
<td>1.10</td>
</tr>
<tr>
<td>Wordian 1 × CO$_2$</td>
<td>4.1</td>
<td>15.6</td>
<td>9.1</td>
<td>13.8</td>
<td>0.3</td>
<td>0.8</td>
<td>0.5</td>
<td>$-2.56$</td>
<td>0.39</td>
</tr>
<tr>
<td>Wordian 2 × CO$_2$</td>
<td>5.4</td>
<td>18.4</td>
<td>12.8</td>
<td>2.6</td>
<td>0.1</td>
<td>0.1</td>
<td>0.01</td>
<td>$-1.79$</td>
<td>1.04</td>
</tr>
<tr>
<td>Wordian 8 × CO$_2$</td>
<td>8.4</td>
<td>22.7</td>
<td>17.5</td>
<td>0.0</td>
<td>0.0</td>
<td>0.0</td>
<td>0.63</td>
<td>$-1.74$</td>
<td>1.07</td>
</tr>
<tr>
<td>Wordian 4 × CO$_2$N</td>
<td>6.7</td>
<td>20.7</td>
<td>15.2</td>
<td>0.0</td>
<td>0.0</td>
<td>0.0</td>
<td>0.33</td>
<td>$-1.90$</td>
<td>0.89</td>
</tr>
<tr>
<td>Wordian 4 × CO$_2$S</td>
<td>6.4</td>
<td>20.7</td>
<td>15.7</td>
<td>0.0</td>
<td>0.0</td>
<td>0.0</td>
<td>0.44</td>
<td>$-1.74$</td>
<td>1.06</td>
</tr>
<tr>
<td>Wordian 4 × CO$_2$CSO</td>
<td>6.9</td>
<td>21.1</td>
<td>15.3</td>
<td>0.2</td>
<td>0.0</td>
<td>0.0</td>
<td>0.25</td>
<td>$-1.84$</td>
<td>1.02</td>
</tr>
<tr>
<td>Wordian 4 × CO$_2$HSO</td>
<td>7.3</td>
<td>21.2</td>
<td>15.5</td>
<td>0.0</td>
<td>0.0</td>
<td>0.0</td>
<td>0.28</td>
<td>$-1.74$</td>
<td>1.05</td>
</tr>
</tbody>
</table>

Table 1. Results From the Wordian Ocean Simulations

Baum [1995], we use Berger’s [1978] extreme values for the Pleistocene with an eccentricity of 0.06 for both experiments and an obliquity of 24.5° for the warm summer orbit (WSO) and 22° for the cold summer orbit (CSO).

2.4. Atmospheric CO$_2$ Level

[15] The assumption of atmospheric CO$_2$ level is crucial for the radiative forcing of the Earth’s climate system. Here we use the functional form of greenhouse gas forcing from Wigley [1987] and a coefficient from Hansen et al. [1988] (consistent with the IPCC 1990 report) to define the changes in the net radiative flux at the tropopause (in W m$^{-2}$):

$$F_{CO2} = 6.3 \ln(p_{CO2}/p_{CO2}\text{ref})$$

with the preindustrial atmospheric partial pressure ($p_{CO2}\text{ref}$ = 280 ppmv. The radiative forcing due to a doubling of CO$_2$ (2 × CO$_2$) is 4.4 W m$^{-2}$ which is about 10% higher than the most recent IPCC2001 estimates of 3.5 to 4.1 W m$^{-2}$ [Ramaswamy et al., 2001]. We are planning to use the improved estimates for future experiments and expect a slightly reduced sensitivity of the climate simulations due to CO$_2$ radiative forcing. However, a small change in CO$_2$ sensitivity is probably negligible compared to various other uncertainties in the reconstruction of the Permian climate.

Concerning CO$_2$ levels for the Wordian, Berner’s [1994] GEOCARB II geochemical model results suggest a CO$_2$ level near 4 times present (4 × CO$_2$) for the Wordian, with a range of error estimates from about 1.5 to about 6 times present day. We conducted four experiments with 1, 2, 4, and 8 times preindustrial CO$_2$ levels to examine the sensitivity of the model climate to CO$_2$.

2.5. Land Surface

[16] A uniform vegetation (mixed tree and grassland or savanna), Dorman and Sellers [1989], is imposed at every land grid point of the EBM and is consistent with the formulation of the land surface in the experiments carried out by the GENESIS 2 atmospheric model simulations [Gibbs et al., 2002]. This prescribed land surface uniformity is clearly unrealistic; for instance a significant area of central Pangaea was probably desert. However, interpretations of results in comparison with observational evidence would be even more complicated if a nonuniform vegetation is assumed because of vegetation’s significant effect on climate [e.g., Dutton and Baron, 1997; DeConto et al., 1999]. We are planning climate model experiments for the Permian that will include interactive vegetation based on the available data. Surface albedo over land is uniform for areas above the freezing point. In the EBM, snow cover with albedo of 0.85 is assumed for areas with temperatures below the freezing point. In the GENESIS 2 experiments, an explicit snow hydrology scheme predicts changes in surface albedo.

3. Simulation Results

3.1. Wordian 4 × CO$_2$ Baseline Experiment

[17] The choice of CO$_2$ for our baseline experiment was guided by Berner’s [1994] geochemical cycle model which indicates that atmospheric CO$_2$ rose from about present-day CO$_2$ levels at about 300 Ma to as high as about 4 × CO$_2$ by the middle Permian. Our Wordian 4 × CO$_2$ will be defined hereafter as the baseline experiment. We will discuss the model’s response to lower levels of atmospheric CO$_2$ concentration (1 × and 2 × CO$_2$) and to high levels of atmospheric CO$_2$ concentration (8 × CO$_2$) in section 3.2. The EBM/LSG baseline experiment is forced, as described in section 2 and Figure 1, with monthly climatologies of surface wind stress and precipitation minus evaporation of the final 5 yr of the 20-yr simulation of the WORD-4 × CO$_2$ experiment [Gibbs et al., 2002] with the GENESIS 2 climate model.

[18] The simulated mean global average ocean temperature for the Wordian is ~2°C higher than for the modern ocean simulation (Table 1). Ocean and land surface temperature (Figures 2b and 2c) indicate generally warmer conditions in high latitudes compared to modern. The moderate high-latitude sea surface temperatures (Figure 2b) are caused by the generally warmer “greenhouse” conditions.
(4 × CO₂) and by significant heat flux from ocean to atmosphere (Figure 2d) reaching 150–200 W m⁻² in the southern high latitudes. Lateral atmospheric heat transport, described in (1), with prevailing westerlies between 30° and 60° [Gibbs et al., 2002, Figure 9d] helps to warm the continents. The minimum temperatures of about −20°C over South Gondwana (Figure 3) in the winter are about 10°–15°C warmer than the minima simulated by Gibbs et al. [2002], and this warmer climate state can be related to a strong heat flux from ocean to atmosphere in the EBM/LSG model, although the atmospheric models are also different and no doubt contribute to the changes as well.

[19] The pattern of annual salinity (Figure 4) follows closely the prescribed P-E pattern [Gibbs et al., 2002, Figure 9d]. It shows the highest salinity in the eastern subtropical Panthalassa between 10° and 30° latitude, where most of the evaporation occurs. Minima in salinity are found just north of the equator at the mean position of the ITCZ and in high latitudes, where precipitation far exceeds evaporation. Differences in salinity of 1 unit between the southern Panthalassa and the Northern Hemisphere are related to a strong poleward current (Figure 5a) in the Southern Hemisphere and low freshwater input along the east coast of South Gondwana.

[20] Simulated east to west gradients of sea surface temperature in tropical Panthalassa are about 3°C, with coldest water delivered by upwelling in the east and a warm pool in west extending into the Tethys Sea (comparable with the modern Pacific; Figure 5b). We note that the variability of the currents and upwelling is probably underestimated in the LSG because the numerical scheme filters out the fast

Figure 3. (left) January (Figure 3a) and July (Figure 3c) modeled sea surface temperature for the Wordian climate with 4 × CO₂ (contour interval: 2°C). (right) January (Figure 3b) and July (Figure 3d) air surface temperature for the 4 × CO₂ experiment (contour interval: 5°C).

Figure 4. Annual average of surface salinity for the Wordian climate with 4 × CO₂ (contour interval: 0.5).
gravitational waves. In the western Tethys Sea, sea surface temperature shows maximum values of over 30°C and minimum values of 27°C in the lee of the equatorial island (South China) due to upwelling of subsurface water masses (Figure 5b). Strong temperature gradients (up to 20°C along the South Tethys Sea coast) are related to strong cooling during winter by outflow of cold air from the continent. Low salinities in the equatorial region are a consequence of well-developed ITCZ precipitation in the prescribed Qw forcing from the GENESIS 2 Wordian simulation. They are additionally influenced by tropical upwelling.

[21] The general features of the tropical ocean circulation in the eastern Panthalassa are comparable to the modern Pacific and show significant N-S differences, in contrast to an ocean simulation with idealized symmetrical land-sea distribution carried out by Kutzbach et al. [1990]. Typical current speeds are 5–10 cm s\(^{-1}\) in the equatorial region of the Panthalassa (Figure 5a). The simulated speeds of the western boundary currents are probably underestimated by a factor of 2 based on the comparison of the modern simulation with observations. This shortcoming of too sluggish simulated currents is typical for models with a coarse resolution and with no nonlinear advection of momentum.

[22] Strong subtropical and polar gyres in each hemisphere, centered around 30° and 50° latitude, are simulated for Panthalassa (Figure 5c). The strength of the subtropical gyre at 30° latitude is about 60 Sv (1 Sv = 10\(^{6}\) m\(^3\) s\(^{-1}\)) which is twice as strong as the gyre simulated for the modern subtropical South Pacific. The poleward directed currents in both hemispheres contribute to a warm temperate climate along the west coast of Gondwana (Figure 5a). On the east coast of Angara (Laurasia), a surface current streams southward similar to the modern Labrador Current and contribute to a cool climate condition around Mongolia (Figures 2b–2d), which is in agreement with Wordian data-derived biomes [Rees et al., 1999, 2002]. However, the authors suggested, on the basis of floral similarities, that the cool climate could also be explained if the position of Mongolia would be moved toward Angara.

[23] The simulated deep ocean also differs from modern. Below 1500 m depth mean salinity is 34.25 and temperature is 5.4°C (compared to 34.66 and 3.8°C for the modern ocean simulation). Main locations for deep water formation (as inferred from maps of potential energy loss by convection; Figure 5d) are northeastern Panthalassa in the northern winter and southeastern Panthalassa in the southern winter. The deep circulation in high latitudes is opposite to the surface circulation (Figure 5e). A powerful westward flow moves across Panthalassa near 70°S latitude. At the east coast of Pangaea, deep western boundary currents circulate in both hemispheres toward the equator. This feature is in agreement with the idealized model of Stommel and Arons [1960]. They envisioned the ocean as a two-layer model with localized sinking in both hemispheres from the upper to the lower layer, and a wide spread slow return of water masses from the abyssal layer into the wind driven upper layer. In accordance with their theory, low current velocities with minima around 30° latitude are simulated in the interior of Panthalassa, which might have implications for an accumulation of organic material from the surface.

Figure 5. (a) Annual average of horizontal velocity at the ocean surface, (b) vertical velocity between surface layer and second model layer (contour interval: 1.5 × 10\(^{-6}\) m s\(^{-1}\); striple indicates downwelling), (c) barotropic stream function (contour interval: 10\(^6\) m\(^3\) s\(^{-1}\)), (d) potential energy loss by convection, and (e) horizontal velocity in the deep sea (~3200 m) for the Wordian climate with 4 × CO\(_2\).
Remineralization and oxidation of the accumulated organic matter might cause anoxic conditions in these locations.

3.2. Response of the Simulated Wordian Climate to Changes in Atmospheric CO2

[24] To investigate the sensitivity of the ocean heat transport and ocean circulation to atmospheric CO2 levels, three additional experiments were carried out with 1/12 CO2, 2/12 CO2, and 8/12 CO2 levels and corresponding changes in the radiative forcing of 0, 4.4, and 13.1 W m−2 (in contrast to 8.7 W m−2 with 4/12 CO2, our baseline). For these experiments, the EBM/LSG uses specified wind stress forcing and precipitation minus evaporation (P-E) forcing from two simulations, WORD-1/12 CO2 and WORD-2/12 CO2, which are analogous to the WORD-4/12 CO2 and WORD-8 experiment from Gibbs et al. [2002]. Differences in global average of sea surfaces temperatures between the 8/12 CO2 level and 1/12 CO2 are 7.1°C (Table 1) and more than 8°C in the Tethys Sea (Figures 6a and 6c). The 1/12 CO2 simulation fails to generate moderate climate in southern Gondwana as a result of the cool surface temperatures, and hence is at odds with reconstructed biomes [Rees et al., 2002], while the 8/12 CO2 simulation produces the most moderate climate (Figure 6c). However, the 8/12 CO2 simulation produces a very warm midlatitude climate (20°C isoline at a latitude of ~40°) and high-latitude climate, which are in conflict with BIOME reconstruction from Rees et al. [2002]. Pole to equator gradients of air temperature in the 1/12 CO2 experiment (Figure 6b) are estimated to be ~10°C higher than the 4/12 CO2 baseline experiment because of effective sea ice albedo feedback in the Northern Hemisphere. The model simulations indicate a sensitive response of sea ice extent in the Northern Hemisphere to an increase of radiative greenhouse gas forcing (Figures 7a and 7b). Sea ice coverage in the Northern Hemisphere in the 1/12 CO2 experiment would be at odds with estimates of water mass properties derived from the climate-sensitive sediments (see section 4). The area covered by sea ice is decreased by ~1/3 if the atmospheric CO2 concentration is doubled from the 1/12 CO2 to 2/12 CO2 (Figure 7b and Table 1).

[25] Using the zonal average profile of net radiation for annual mean conditions at the top of the atmosphere (QTA), we can estimate the total poleward heat transport in the ocean–atmosphere system (TTOT) as a sum of the poleward heat transport by the atmosphere (TA) and by the ocean (TOC):

\[
Q_{TA} = \int_{\text{top}} (1 - A)Q_{SW} - Q_{LW} \, dt \approx 0
\]

\[
T_A + T_{OC} = -\int_{\phi=0}^{\pi/2} Q_{TA} 2\pi R^2 \cos \phi' d\phi'
\]

where QSW and QLW are short-wave and long-wave radiation, A is albedo and R is Earth radius. Values of
...with the moderate marine climate on the modern Norwegian Coast (e.g., near Bodø) strongly influenced by the warm North Atlantic Current. In experiment $4 \times \text{CO}_2\text{N}$, the warm eastward moving currents of middle north latitudes bend to the north (comparable with the flow of the present North Atlantic current into the Arctic Region) and joins the northern circumpolar current (11e). A similar feature is simulated with a southern circumpolar current (Figure 11f). The existence of the southern polar seaway causes a reduction of the southern meridional overturning circulation by about 10 Sv (not shown) which could be...
related to a northward Ekman transport by westerlies [Gibbs et al., 2002, Figure 9] at the southern tip of Gondwana.

3.4. Response to Changes of the Orbital Parameters

Two extreme orbital configurations, one with hot summer orbit (HSO) and one with cold summer orbit (CSO) are considered (section 2.3). The orbital configurations are selected such that the Southern Hemisphere summer will be warmest with HSO and Northern Hemisphere will be warmest with CSO. The two experiments used the $4 \times \text{CO}_2$ baseline and are referred to as $4 \times \text{CO}_2$ HSO and $4 \times \text{CO}_2$ CSO. The specified wind stress and P-E for these experiments are taken from WORD-8 $\times \text{CO}_2$, HSO and WORD-4 $\times \text{CO}_2$, CSO described by Gibbs et al. [2002]. High-latitude summer temperatures are sensitive to changes in the orbital configuration and play a critical role in snow accumulation and formation of continental ice sheets in these areas. Early
Jurassic EBM simulations result in maximum surface temperature changes of up to \( \sim 15°C \) [Crowley et al., 1992]. Simulations of Gibbs et al. [2002] indicate that significant differences between HSO and CSO configurations result in some changes in south polar snow accumulation and hence affect high-latitude climate reconstructions.

Our simulations show HSO minus CSO differences in summer surface temperatures for the Southern Hemisphere (Figure 12a) of \( \sim 20°C \) centered over Gondwana and for the Northern Hemisphere of \( \sim 10°C \) centered over Euramerica (Figure 12b). These significant temperature changes have implications for the length of the growing season and calculation of modeled biome in a fully coupled ocean-atmosphere climate model. The positive temperature anomaly in the Northern Hemisphere in Figure 12a may be due to a positive ice albedo feedback mechanism in the WSO experiment. Changes of the orbital parameters have only little effect on the global annual mean air surface and ocean temperatures (Table 1), similar to results from the Late Ordovician climate sensitivity experiments of Poussart et al. [1999].

4. Comparison With Climate-Sensitive Sediments

Ziegler et al. [1998] constructed climate maps for the Wordian (Figure 13a) based upon Wordian sediments and their present-day relationship of sediments to characteristic water masses (Table 2). The climate-sensitive sediments have been compared with modern marine climate properties, in particular temperature, salinity and productivity related effects. The water mass boundaries listed in Table 2 are based on a summary of parameters derived from world ocean atlases. The boundaries of characteristic water masses (defined by specific temperature and salinity ranges and strength in upwelling) are chosen somewhat arbitrarily because it is difficult to select particular values that universally represent significant climate changes and transitions on the Paleozoic timescale. The water masses of Ziegler et al. [1998] are classified in eight categories (Table 2). These are briefly described (Figure 13a) and then compared with our model simulations for the Wordian baseline experiment \( 4 \times CO_2 \) unless otherwise noted (Figure 13b).

4.1. Glacial Climate

4.1.1. Data

Occurrence of permanent ice flows (like most of the modern Arctic Ocean, or ice shelf of Antarctica) generated by outflow of cold air. Tills could be expected along mountainous coasts of these regions. Sediments of this category are known in the Early Permian but have not been found in Wordian rocks.

4.1.2. Model

In the ocean, sea ice cover (sea surface temperatures \( \sim -1.8°C \)) is not simulated in the \( 4 \times CO_2 \) Wordian baseline experiment, but occurs with lowered \( CO_2 \).

4.2. Cold Temperate Climate

4.2.1. Data

The water masses are defined by the extent of sea ice in the winter season and correspond closely to the \( 0°C \) isotherm (e.g., similar to conditions in the modern Labrador
Dropstones and rhythmites would be indicative of this environment in the fossil record and have been identified in the geological record of Tasmania along the southeastern coast of Gondwana (Figure 11a).

4.2.2. Model

The model (Figure 13b) simulates moderate conditions along the south-polar east coast from Gondwana where data indicate a cold temperate condition. Warm conditions (sea surface temperatures of \(24^\circ C\) in winter time; Figure 3c) are related to the polarward directed currents modeled for these regions. In the observations, the Tasmanian seas might represent a re-entrant in the coast, isolated from the warming effects of the open ocean currents.

4.3. Wet Temperate Climate

4.3.1. Data

Temperatures between \(0^\circ C\) and \(20^\circ C\) and brackish conditions (salinity below 32) define this climate zone (for example modern Baltic Sea or Canadian West Coast). Peats are characteristically developed around the margins of these characteristic water masses and organic muds are commonly associated in the center of these areas with high precipitation. The swamp deposits depend on consistency of rainfall through the annual cycle, or at least through the warmer months [Lottes and Ziegler, 1994], and for confined seaways or broad shallow shelves this translates into lowered salinities. The associated surface runoff provides for abundant nutrients and a stratified water column which in turn allows for organic preservation. The geological data (Figure 13a) for this water mass are best seen around Angara but also occur around southern Gondwana.

4.3.2. Model

Simulated high precipitation [Gibbs et al., 2002] and low salinities (Figure 4) of the high-latitude west coasts of Angara agree with the geological recognition of the wet temperate climate category. In addition, model results show wet temperate regions in northern India along the shore of the southern Tethys Sea similar to the geological findings. These climate conditions are related to the advection of moist tropical air masses and enhanced precipitation. Interestingly, the model shows a wider extent of this water mass type than had been inferred from the geological data but the model results are nevertheless entirely consistent with the data. Here the original water mass interpretation was simply too conservative.
4.4. Temperate Climate

4.4.1. Data

Temperature in this climate zone ranges 0°C and 20°C and salinity between 32 and 37 (e.g., Kuroshio Current for modern climate). No particular climate indicators have been identified for this zone, but peats are quite common due to low evaporation rates in mid to high latitudes.

4.4.2. Model

The equatorial margin of the water mass in this climate zone is placed around 30° latitude for the Wordian 4 × CO₂ simulation (Figure 13b), similar to the reconstructed equatorial margin from climate-sensitive sediments (Figure 13a). The bias between model results and the reconstructions on the West Coast of Pangaea can be explained by the more poleward location of main upwelling areas, which are centered in the model around the equator as a response to the wind stress and bathymetry.

4.5. Cool Subtropical Upwelling Zones

4.5.1. Data

This category is defined by zones, which are mainly limited to the subtropical belt and equatorial region where Ekman-induced upwelling produces high organic productivity and nutrient content. Formation of organic muds and phosphorites constitutes the recognition of these upwelling
zones. No attempt was made to map the equatorial upwelling zone as the Permian ocean floor has been subducted and the sedimentary evidence destroyed.

4.5.2. Model
Upwelling is mapped as green shaded areas and is defined to have upward current velocities greater than $1 \times 10^{-6}$ m s$^{-1}$ (Figure 13b). Typical regions are the coastal upwelling along the west coast of Gondwana at locations comparable with modern Californian coast or the Coast of Peru. A strong upwelling belt is located along the Equator similar to the modern Pacific and around islands. This equatorial upwelling includes shallow platform seas of the South China microcontinent and is actually supported by the presence of organic rich shales and phosphorites. These were originally interpreted as indicating the wet tropical water mass (Figure 13a) but the upwelling interpretation fits equally well.

4.6. Dry Subtropical Climate
4.6.1. Data
Areas with low precipitation minus evaporation with salinity greater than 37 provide favorable conditions for evaporite deposition in coastal lagoons and broad platform seaways (Figure 13a).

4.6.2. Model
Low precipitation and high salinities are simulated along the southeastern Tethys Sea (Figure 4), South China, or the West Coast of Euramerica in agreement with the geological evidence. In addition, embayments along the northern coast of Euramerica have been identified to be dry from model simulations [Gibbs et al., 2002] and geological data. However, the ocean model does not resolve this area because of the coarse resolution of the model (Figure 13b).

4.7. Tropical Climate
4.7.1. Data
Surface temperatures greater than 20°C and normal salinity, which allows carbonate buildup and bottom productivity, characterize this water mass (Figure 13a). Carbonate buildups are generally not found in the geological past above 35° of latitude, as shown by paleomagnetic data [Hulver et al., 1997]. This latitudinal restriction constitutes the zone of light penetration to the seafloor as controlled by the Sun’s zenith angle [Ziegler et al., 1984].

4.7.2. Model
Model simulations (Figure 13b) are in agreement with climatic reconstructions (Figure 13a), but differences on the west coast of Pangaea are associated with upwelling areas, which cannot be entirely resolved with the coarse model resolution.

4.8. Wet Tropical
4.8.1. Data
Wet tropical areas are defined to have sea surface temperatures above 20°C and relatively brackish conditions (salinities below 32). Organic-rich shales do not seem to be associated with wet tropical climate at modern times, but they almost certainly occurred in the Wordian (e.g., in the South China Sea, Figure 13a).

4.8.2. Model
Wet tropical conditions are simulated in the southern Tethys Sea, for the east coast of North China, and the southern coast of South China (Figure 13b). Such areas are characterized by high river runoff, and/or high precipitation (e.g., along the intertropical convergence zone) and correlate well with tropical coals of these areas.

5. Summary and Conclusions
We simulated the Wordian climate with a coupled atmosphere-ocean model of intermediate complexity (EBM and LSG) using realistic land-ocean geometry. The model results ($4 \times \text{CO}_2$ baseline experiment) produce (1) surface water masses with temperature characteristics which are warmer and fresher than at modern times in the high latitudes, (2) a strong westward equatorial current which is blocked by islands at the eastern Tethys Sea, (3) warm poleward directed currents along the Tethys coast and east coast of Gondwana, (4) cold equatorward currents along the east coast of Angara, (5) large meridional overturning circulation cells with deep water formation at both poles, and (6) a warm deep ocean with low current speed in the
eastern equatorial Panthalassa. The model results were compared with climate-sensitive sediments (section 4) and with other climate simulations. The $4 \times CO_2$ baseline experiment shows the best agreement with the paleoclimatic data (section 4). Our experiments suggest that very low CO$_2$ and very high concentrations would cause significant data-model disagreements. Low CO$_2$ concentrations ($1 \times CO_2$) would more likely be associated with a glacial climate.
whereas high \( CO_2 \) (8 \( \times \) \( CO_2 \) experiment) concentrations simulate too warm conditions in the high latitudes. We showed that existence of circumpolar currents in polar seaways would influence the temperature of adjacent high-latitude coasts. In comparison with the previous study of Gibbs et al. [2002], temperature changes over Gondwana were simulated to be up to 15°C higher in the annual mean. We also infer from our simulations that changes of the astronomical parameters have a significant impact on the climate over Gondwana, in general agreement with previous studies [Crowley et al., 1992]. The baseline simulation (4 \( \times \) \( CO_2 \)) has a significantly warmer polar climate than the modern, owing to increased flux of heat from ocean to atmosphere in high latitudes. The results suggest that a moderate climate over South Gondwana could be reproduced by the following two conditions: (1) an atmospheric \( CO_2 \) concentration of at least 4 times present levels and/or (2) the existence of a south polar seaway that further increases the heat flux to the atmosphere from the ocean. With these prescriptions, the simulated moderate climate over South Gondwana agrees reasonably well with geological data from climate-sensitive sediments and phytogeographic patterns. The simulated Panthalassa deep-sea circulation shows a high sensitivity to the ice-albedo feedback and radiative greenhouse gas forcing with strong N-S asymmetry at low \( CO_2 \) levels and moderate asymmetry at high \( CO_2 \) levels. Regions of slow currents in the deep Panthalassa have implications for carbon accumulation and low oxygen conditions.

Although these results lend additional insight into possible climate processes of the Wordian, there are several simplifications that must be noted. The sensitivity of the simulated deep sea circulation to \( CO_2 \) radiative forcing may well be different for different coupled ocean-atmosphere models as documented for modern and future \( CO_2 \) scenarios in the recent IPCC2001 report [Cubasch et al., 2001]. For example, the magnitudes and locations of deep ocean formation are significantly different in different models. Moreover, important processes (such as an interactive hydrological cycle and interactive atmosphere-ocean dynamics) are not included in our model. However, analyses with a coupled atmosphere-ocean study of the role of air-sea flux components in \( CO_2 \)-induced climate changes [Mikolajewicz and Voss, 2000] have shown one can investigate the effect of individual components of air-sea interactions using partially coupled simulations.

The ocean model used for this study has reproduced many of the qualitative features of the global ocean circulation and marine tracer distributions for modern and past times. However, a limitation arises from the use of a non-eddy coarse resolution model which underestimates the strength of the western boundary currents. A more refined model with eddy-resolving grid size will probably yield a more realistic simulation of the western boundary currents and locations of deep water formation.

There were additional limitations to this study. We used prescribed and spatially uniform vegetation and soil parameters. In the future, we will use a fully coupled atmosphere-ocean-land biosphere model to predict climate/vegetation feedbacks that are known to be important for past climate changes and variability. For example, a replacement of tundra with boreal forest could lead to additional climate warming, i.e., a positive feedback, because of the albedo difference in spring and between snow covered forest and snow covered tundra [Kutzbach et al., 1996; Ganopolski et al., 1998].

Uncertainties remain concerning the inadequate representation of the water-vapor feedback mechanism in energy balance models [Pierrehumbert, 1995; Bates, 1999], which might alter our results as well as those from previous studies significantly. Also, consideration of Permian to modern changes in the aerosol loading might be relevant for the radiative budget, but there are no estimates of these properties available so far. The solutions of the simulations with the coupled atmospheric energy balance ocean general circulation model could result in multiple steady states [Mikolajewicz and Crowley, 1997]. However, we expect that the stability of our simulations is robust because of the asymmetry of the geography of Pangea with respect to the equator. Preliminary results from simulations with a transient \( CO_2 \) radiative forcing 2 times \( CO_2 \) to one times \( CO_2 \) support these findings.

Paleogeographic and topographic reconstruction of islands is still uncertain. For example, we deduce from our experiments and comparison with those of Kutzbach et al. [1990] that a more southward position of South China would imply a higher heat transport into the Tethys Sea and a warming of the adjacent coasts. The bathymetry and the details of polar seaways are also uncertain. The outcome of this study will be helpful for the design of future fully coupled climate studies, in particular with respect to changes in the land-sea distribution, greenhouse gas radiative forcing, and changes in the orbital configuration. In addition, a coupling of these climate models to a marine biogeochemical model would give a more detailed insight concerning possible anoxic conditions in the Panthalassa.

---

### Table 2. Surface Water Mass Classification

<table>
<thead>
<tr>
<th>Climate Zone</th>
<th>Water Mass Essentials</th>
<th>Geological Recognition</th>
</tr>
</thead>
<tbody>
<tr>
<td>Glacial</td>
<td>surface permanently frozen; SST &lt; −1.8°C</td>
<td>marine till</td>
</tr>
<tr>
<td>Cold temperate</td>
<td>winter ice flows; SST −1.8 – 0°C</td>
<td>dropstones, rhythmites</td>
</tr>
<tr>
<td>Wet temperate</td>
<td>brackish surface water; SST 0 − 20°C, salinity &lt;32</td>
<td>temperate peats, organic rich shales</td>
</tr>
<tr>
<td>Temperate</td>
<td>mixed water column; SST 0 − 20°C, salinity 32 – 37</td>
<td>clastics</td>
</tr>
<tr>
<td>Cool subtropical</td>
<td>upwelling currents; vertical velocity w &gt; 1 ( \times 10^{−2} ) m s(^{−1} )</td>
<td>organic-rich shales, phosphorites, cherts</td>
</tr>
<tr>
<td>Dry subtropical</td>
<td>high evaporation; salinity &gt;37</td>
<td>gypsum, halite, sabkha facies</td>
</tr>
<tr>
<td>Tropical</td>
<td>deep light penetration; SST &gt;20°C, salinity 32 – 37</td>
<td>carbonates, oolites, coralgal reefs</td>
</tr>
<tr>
<td>Wet tropical</td>
<td>high precipitation or continental runoff; SST &gt;20°C, salinity &lt;32</td>
<td>tropical peats, muddy sediments</td>
</tr>
</tbody>
</table>

\(^{a}\)Following Ziegler et al. [1998].
ocean that might have implications for the Late Permian mass extinction.

[53] Acknowledgments. We thank Cornelius Winguth for reading this manuscript, David Archer for fruitful discussions and Pat Behling for setting up and running the GENESIS 2 climate model experiments and for providing the forcing fields from the GENESIS run of Gibbs et al. [2002] for the LSG ocean model. We thank Matthew Huber and Lisa Sloan for many detailed comments which improved the quality of the manuscript. This work was done in collaboration with the Max-Planck-Institut für Meteorologie, the Deutsche Forschungsgemeinschaft SFB 1054, and with support of the Packard Foundation. John Kutzbach is supported by NSF grant ATM-9815980, and Arne Winguth by NASA grant NAG5-11245.

References


-----------------------------------------------------------------------------------
C. Heinze, E. Maier-Reimer, and U. Mikolajewicz, Max-Planck-Institut für Meteorologie, 20146 Hamburg, Germany.
J. E. Kutzbach and A. M. E. Winguth, Center for Climatic Research, University of Wisconsin-Madison, Madison, WI 53710, USA. (amwinguth@facstaff.wisc.edu)
A. Rees, D. Rowley, and A. M. Ziegler, Department of Geophysical Sciences, University of Chicago, Chicago, IL 60637, USA.
Figure 13. (a) Wordian (Kazanian, early Maokouan) climate sensitive sediments and reconstructed water masses from Ziegler et al. [1998]. (b) Water mass classification derived from the climate simulation with 4 × CO₂ using the classification scheme of Table 1.