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Vegetation-climate interactions in the warm mid-Cretaceous

J. Zhou et al.

Vegetation-climate interactions in the warm mid-Cretaceous

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Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

⏪

⏩

◀

▶

Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion

Abstract

Vegetation-climate interactions are thought to have amplified polar warmth during past warm periods. Here, we explore the vegetation-climate interactions in the mid-Cretaceous using a fully coupled ocean-atmosphere general circulation model with a dynamic vegetation component. We run simulations with 1×, 10× and 16× pre-industrial atmospheric CO₂. Results show that forests expand from mid-latitudes to high latitudes as CO₂ increases from 1× to 10× and 16×, mainly due to the CO₂-induced warming. This expansion of mid-to-high latitude forests are largely supported by the distribution of mid-Cretaceous fossil woods and coal deposits. Globally, the presence of vegetation increases mean annual temperature and precipitation by 0.9 °C and 0.11 mm day⁻¹ relative to bare ground. High-latitude warming induced by the presence of vegetation (~ 1.9 °C) is less than half of that reported in previous studies. The weaker warming here is mainly due to less pronounced albedo feedbacks, and to a less extent, reduced poleward heat transport via weakening of the meridional overturning circulation. Our results suggest that other mechanisms in addition to high atmospheric CO₂ and high-latitude vegetation are required to maintain the polar warmth.

1 Introduction

The mid-Cretaceous (~ 100 Ma) was a period of extreme polar warmth. Paleoclimate proxies indicate that mean annual temperature were as high as 20 °C and 13 °C over polar ocean and land (Bice et al., 2003; Jenkyns et al., 2004; Spicer et al., 2002, 2008). The cause of global warmth is widely accepted to be caused by high atmospheric CO₂ levels (Barron and Washington, 1985) resulting from volcanic outgassing. However, simulations of Cretaceous climate using general circulation models (GCMs) with high atmospheric CO₂ predict polar temperatures that are too cool (Barron et al., 1995; Poulsen et al., 1999, 2004; Spicer et al., 2008) challenging our understanding of greenhouse climate.

Vegetation-climate interactions in the warm mid-Cretaceous

J. Zhou et al.

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures



Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion



Vegetation-climate interactions in the warm mid-Cretaceous

J. Zhou et al.

[Title Page](#)[Abstract](#)[Introduction](#)[Conclusions](#)[References](#)[Tables](#)[Figures](#)[Back](#)[Close](#)[Full Screen / Esc](#)[Printer-friendly Version](#)[Interactive Discussion](#)

Several hypotheses have been proposed to reconcile this high-latitude model-data discrepancy including enhanced poleward heat transport via strengthening of atmospheric or oceanic circulation (e.g., Covey and Barron, 1988; Farrel, 1990; Korty et al., 2008), increased local radiative forcing through high-latitude cloud feedbacks (e.g., Abbot and Tziperman, 2008; Sloan and Pollard, 1998; Kump and Pollard, 2008), high atmospheric methane levels (Bice et al., 2006) and high-latitude forests (Deconto et al., 2000; Otto-Bliesner and Upchurch, 1997; Upchurch et al., 1998). Of these mechanisms only the existence of high-latitude forests is directly supported by observations in the geological records (e.g., Herman and Spicer, 1996; Falcon-Lang et al., 2001; Spicer and Parrish, 1986; Spicer et al., 1993).

Previous studies have reported that the expansion of forests into polar regions in the Late Cretaceous may have warmed high-latitude regions by 4–7 °C (Deconto et al., 2000; Otto-Bliesner and Upchurch, 1997; Upchurch et al., 1998). The warming is mainly attributed to a reduction in surface albedo due to the masking of snow via forests and melting of sea ice via transfer of heat from vegetated land to ocean. The role of vegetation feedbacks on polar warming has not been examined for the mid-Cretaceous when CO₂ levels were thought to have been higher. In addition, previous Late Cretaceous modeling studies coupled atmospheric general circulation models (AGCMs) to mixed-layer ocean models that do not allow for dynamic ocean feedbacks. Paleoclimate modeling of Cenozoic intervals has demonstrated that vegetation-induced climate change can influence the strength of the ocean meridional overturning circulation (MOC) (Brovkin et al., 2002, 2009; Ganopolski et al., 1998; Lohmann et al., 2006). Of possible relevance to Cretaceous polar warmth, Lohmann et al. (2006) report that the prescribed present to warm late-Miocene vegetation changes strengthen the North Atlantic MOC by more than 3 times, leading to a warming over the mid-to-high latitude North Atlantic by up to 8 °C. Furthermore, previous studies investigated the Cretaceous vegetation-climate interactions using prescribed vegetation changes (Otto-Bliesner and Upchurch, 1997; Upchurch et al., 1998) or an equilibrium vegetation model (Deconto et al., 2000), which do not allow realistic transient climate-vegetation

interactions as in dynamic global vegetation models (Peng, 2000).

In view of the limitations of these earlier studies, we re-examine the vegetation-climate interactions in the Cretaceous under high atmospheric CO₂ using a fully coupled ocean-atmosphere GCM with a dynamic vegetation component (CCSM3). This study addresses two main questions: (1) can CCSM3 simulate the mid-Cretaceous vegetation distribution, and if so, at what atmospheric CO₂ levels? (2) How does the simulated vegetation affect mid-Cretaceous climate? We first examine the climate control on vegetation by comparing the vegetation pattern with low (1 ×) and high (10 × and 16 ×) atmospheric CO₂. We then investigate the impact of vegetation on climate by comparing a simulation with the maximum simulated vegetation mass to a simulation without vegetation.

2 Methods

Cretaceous simulations were completed with NCAR's Community Climate System Model version 3 (CCSM3), a fully coupled GCM with dynamic global vegetation (Collins et al., 2006; Levis et al., 2004). CCSM3 was run in a T31 × 3 configuration, with a T31 spectral truncation (3.75° × 3.75°), 26-level atmosphere component model (Community Atmospheric Model) coupled to a nominal 3°, 25-level ocean component model (Parallel Ocean Program). The land surface model (Community Land Model) and dynamic global vegetation model (CLM-DGVM) are set to the same horizontal resolution as atmosphere; the sea ice component model (Community Sea Ice Model) has the same horizontal resolution as the ocean. Vegetation distribution in CLM-DGVM is largely determined by solar radiation, air temperature, soil moisture and atmospheric CO₂ concentration (Bonan and Levis, 2006). For example, the survival and establishment of vegetation are constrained by bioclimatic parameters including monthly air temperature, growing degree days (GDD) and mean annual precipitation (Table 1). CLM-DGVM classifies vegetation as patches of plant function types (PFTs) (Levis et al., 2004). A PFT represents the average individual for a group of species that share

Vegetation-climate interactions in the warm mid-Cretaceous

J. Zhou et al.

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures



Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion



Vegetation-climate interactions in the warm mid-Cretaceous

J. Zhou et al.

[Title Page](#)[Abstract](#)[Introduction](#)[Conclusions](#)[References](#)[Tables](#)[Figures](#)[⏪](#)[⏩](#)[◀](#)[▶](#)[Back](#)[Close](#)[Full Screen / Esc](#)[Printer-friendly Version](#)[Interactive Discussion](#)

similar physiological and morphological characteristics. All PFTs can coexist in a grid cell's soil-covered portion when climate allows. CLM-DGVM calculates vegetation dynamics including allocation, biomass turnover, mortality, aboveground competition, fire and establishment yearly. Levis et al. (2004) provide a detailed description of vegetation dynamics. Vegetation biogeography and structure is also updated yearly. Canopy phenology is updated daily. The communication between CLM-DGVM and the atmosphere model is achieved through exchanges of energy, momentum, and moisture every 20 min (Levis et al., 2004). Our version of CLM-DGVM does not simulate transient changes in atmospheric CO₂.

Eight PFTs are defined in this study, with one herbaceous PFT (C₃ grasses) and seven woody PFTs (trees) (Table 1). We exclude C₄ grasses in this study because C₄ grasses did not become ecologically important until the Early-middle Miocene (Curling et al., 1993; Jacobs et al., 1999). Though the origin of C₃ grassland is no earlier than the Late Cretaceous (Jacobs et al., 1999), we include C₃ grasses for two reasons: (1) CLM-DGVM does not have a PFT representing shrubs, therefore, shrublands are often simulated as desert or grasslands (Bonan et al., 2003); and (2) the fossil record suggested a widespread distribution of herbaceous vegetation during the mid-Cretaceous (Spicer et al., 1993).

To determine the equilibrium vegetation distribution corresponding to the mid-Cretaceous warm climate, we conducted three simulations implementing CLM-DGVM, namely, 1×DGVM, 10×DGVM and 16×DGVM (Table 2). An additional experiment (10×BG) was run with land surface prescribed as bare ground. Atmospheric CO₂ concentration was set to 1×, 10× or 16× pre-industrial levels, and physiological CO₂ concentration was set to 355 ppmv in the experiments with DGVM. 16× pre-industrial levels CO₂ likely falls in the upper limit of the estimated mid-Cretaceous atmospheric CO₂ levels (see summary in Bice and Norris, 2002). All experiments were run with a reduced solar constant (99% of modern) (Gough, 1981) and present day orbital parameters. The model geography and topography were based on mid-Cretaceous (100 Ma) reconstructions from PALEOMAP project (Scotese, 2001). Because sea level was at

highstand, global land area are 18.6% less than present day. Following Poulsen et al. (2003), we applied an age-depth relationship to reconstruct bathymetry for the deep ocean regions with known magnetic lineations (Barron, 1987). Elsewhere, deep-ocean depths were set to 5300 m. River runoff is routed to the ocean via a river transport model (Branstetter, 2001). All experiments were initiated using zonal ocean temperatures and salinity from a warm Late Cretaceous simulation (Otto-Bliesner et al., 2002). Vegetation was initialized from a 300 year integration of a Cretaceous CAM3-DGVM simulation with $10\times$ pre-industrial CO_2 levels. Levis et al. (2004) report that global vegetation distribution approaches equilibrium in a few hundred years in CLM-DGVM. Each of our Cretaceous experiments was run in excess of 1600 yr (Table 2). The linear trend for global vegetation cover is on the order of 10^{-3} /century and is statistically insignificant. The drifts in simulated global mean ocean temperature over the last 200 years are also small (on the order of $0.1^\circ\text{C}/\text{century}$) for all experiments.

3 Results

3.1 Response of vegetation to CO_2 levels

3.1.1 Simulated surface temperature and precipitation

The increase of atmospheric CO_2 leads to an increase in global annual mean surface temperature from 13.6 to 24.0 and 25.6°C in the $1\times\text{DGVM}$, $10\times\text{DGVM}$ and $16\times\text{DGVM}$ experiments, respectively (Table 2). Continental surface temperatures increase by up to $\sim 10^\circ\text{C}$ higher at low latitudes and 29°C at high latitudes as a result of increasing CO_2 from 1 to $16\times\text{PAL}$ (Fig. 1a). Consistent with the increase in surface temperature, global annual mean precipitation increases from 2.90 mm day^{-1} in the $1\times\text{DGVM}$ experiment to 3.47 and 3.56 mm day^{-1} in the $10\times\text{DGVM}$ and $16\times\text{DGVM}$ experiments (Table 2). The increase in continental precipitation between $1\times\text{DGVM}$ and $16\times\text{DGVM}$ experiments is up to 1.8 ($\sim 33\%$) and 1.9 ($\sim 150\%$) mm day^{-1} at

low and mid-to-high latitudes. In contrast, the continental precipitation decreases by $\sim 0.3 \text{ mm day}^{-1}$ ($\sim 20\%$) in the southern subtropical latitudes ($\sim 20\text{--}30^\circ \text{ S}$) (Fig. 1b).

3.1.2 Simulated global vegetation distribution

In the $1\times\text{DGVM}$ experiment, the tropics, subtropics, mid-latitudes and high latitudes are dominated by evergreen trees, grasses, mixture of evergreen and deciduous trees and polar desert, respectively (Fig. 2a–c). In the high CO_2 experiments, trees (especially deciduous) and grasses advance into the northern high latitudes and Antarctica (Fig. 2d–i). In the northern high latitudes ($60\text{--}90^\circ \text{ N}$), average fractional tree cover over the soil-covered land increases from 0.7% in the $1\times\text{DGVM}$ experiment to 61% ($10\times\text{DGVM}$) and 59% ($16\times\text{DGVM}$). The change in tree cover at the southern high latitudes ($60\text{--}90^\circ \text{ S}$) is slightly less than the northern high latitudes, increasing from 4% ($1\times\text{DGVM}$) to 51% ($10\times\text{DGVM}$) and 43% ($16\times\text{DGVM}$). In contrast, tree cover at low latitudes ($30^\circ \text{ S}\text{--}30^\circ \text{ N}$) decreases from 60% ($1\times\text{DGVM}$) to 54% ($10\times\text{DGVM}$) and 50% ($16\times\text{DGVM}$). As a result of the poleward expansion of trees and grasses, area-weighted global vegetation cover increases from 56% ($1\times\text{DGVM}$) to 85% ($10\times\text{DGVM}$ and $16\times\text{DGVM}$). Global net primary production (NPP), which is equal to the net gain of carbon through photosynthesis and respiration, is estimated to be $61.8 \text{ Pg C yr}^{-1}$ in the $1\times\text{DGVM}$ experiment, comparable to observational estimates of present-day NPP (Schlesinger 1997). Due to the increase in global vegetation cover, NPP increases by $13.5 \text{ Pg C yr}^{-1}$ (21.8%) in the $10\times\text{DGVM}$ experiment and $10.9 \text{ Pg C yr}^{-1}$ (17.6%) in the $16\times\text{DGVM}$ experiment relative to the $1\times\text{DGVM}$ experiment.

The large-scale changes in vegetation distribution are linked to changes in surface temperatures and precipitation. The expansion of trees and grasses at mid- and high latitudes in the two high CO_2 experiments is mainly due to CO_2 -induced warming (Fig. 1a). In the $1\times\text{DGVM}$ experiment, GDD (i.e., annual growing degree-days above 5° C), which measures the accumulation of growing season warmth in a year, is largely less than 350° C day at high latitudes inhibiting tree growth (Table 1) and the minimum monthly temperature is lower than -17° C prohibiting grass growth (Table 1).

Vegetation-climate interactions in the warm mid-Cretaceous

J. Zhou et al.

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

⏪

⏩

◀

▶

Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion



(Figs. 4b and 5f). Mid-latitude Northern Hemisphere surface temperatures increase during the cold seasons and decrease during the warm seasons (Fig. 5k). Changes in surface temperatures over ocean are generally less than 1 °C at low latitudes and up to 3 °C at high latitudes (Fig. 4b). The maximum increase in surface temperature (> 5 °C) occurs in the mid-latitude North Pacific, and is caused by the poleward shift of western boundary currents (explained in Sect. 3.2.3).

The changes in land surface temperature can be explained by the net effect of vegetation-induced changes in latent heat fluxes and surface absorbed solar radiation. The former results from changes in evapotranspiration (Fig. 6a); the latter is mainly due to changes in surface albedo (Fig. 6b) and low-cloud cover (Fig. 6c). As a result of canopy evapotranspiration, the total annual mean evapotranspiration over land increases by 0.38 mm day⁻¹ (22.7 %) in the 10×DGVM experiment relative to the 10×BG experiment (Fig. 6a). And the increase (> 1.0 mm day⁻¹) is most significant over the low-to-mid latitude evergreen forests. In contrast, the annual mean land surface albedo decreases by 0.04 (~20 %) in the 10×DGVM experiment (Fig. 6b and Table 3), due to the darkening of land surface via forests, and to a less extent via reduced snow cover. The reduction (up to 0.2) is most pronounced over mid-to-high latitude deciduous forests through snow-masking (Fig. 6b). Seasonally, over high-latitude forests, the net surface radiation (i.e. the difference between surface absorbed solar radiation and net upward longwave radiation) increases by ~ 24 W m⁻² due to the large gain in surface absorbed solar radiation (> 31 W m⁻²) during late spring and early summer (Fig. 5c), which greatly exceeds increases in latent heat fluxes (Fig. 5b), leading to significant warming (Fig. 5a). The increases in surface absorbed solar radiation are due to (1) decreases in surface reflected solar radiation associated with the reduction in surface albedo (Fig. 5d), and (2) increases in incoming solar radiation associated with reduced low cloud cover (Fig. 6c). The decreases in low cloud cover are linked to the reduced boundary layer relative humidity with increased temperatures. In contrast, over low-latitude forests, increases in latent heat fluxes are up to 31 (26) W m⁻² greater than increases in surface absorbed solar radiation in the Northern (Southern) Hemisphere

Vegetation-climate interactions in the warm mid-Cretaceous

J. Zhou et al.

[Title Page](#)[Abstract](#)[Introduction](#)[Conclusions](#)[References](#)[Tables](#)[Figures](#)[⏪](#)[⏩](#)[◀](#)[▶](#)[Back](#)[Close](#)[Full Screen / Esc](#)[Printer-friendly Version](#)[Interactive Discussion](#)

**Vegetation-climate
interactions in the
warm
mid-Cretaceous**

J. Zhou et al.

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

⏪

⏩

◀

▶

Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion



(Fig. 5g ,h) leading to significant cooling (Fig. 5f). The increases in absorbed solar radiation are significantly smaller than the surface albedo feedback (Fig. 5i), because low-latitude forest evapotranspiration leads to denser low cloud cover (Fig. 6c), and thus offsets the surface albedo feedback by reducing incoming solar radiation at the surface (Fig. 5j). Over the northern mid-latitude forests, the increases in latent heat fluxes are up to 35 W m^{-2} greater than increases in absorbed solar radiation during summer (Fig. 5 l,m) leading to summer cooling (Fig. 5k).

Unlike over land, changes in surface albedo are negligible over ocean except the northern Eurasian coastal area (Table 3 and Fig. 6b). The large-scale background warming over ocean results from the increase in downwelling longwave radiation (7.3 W m^{-2}) (Table 3) associated with elevated atmospheric water vapor content (correlation > 0.7). Warming over the southern South Atlantic and Tethys Ocean, the equatorial and northern mid-latitude Pacific Ocean, and the Western Interior Seaway is also enhanced by positive cloud radiative forcing due to reduced low cloud cover (Fig. 6c). In contrast, warming over the subtropical Pacific Ocean is suppressed by negative cloud radiative forcing due to increased low cloud cover.

3.2.2 Precipitation response

From a global perspective, the presence of vegetation slightly intensifies the mid-Cretaceous hydrological cycle (Table 3). As a result of the increases in the surface temperature and evapotranspiration, the global precipitable water (i.e., vertically integrated specific humidity) and precipitation increases by 3.3 W m^{-2} (8.7 %) and 0.11 mm day^{-1} (3.3 %), respectively. In contrast, snowfall decreases by 0.02 mm day^{-1} (17 %) over land. The increases in precipitation are more pronounced over land (0.26 mm day^{-1} , 9.7 %) than over ocean (0.07 mm day^{-1} , 2.0 %). Due to enhanced precipitation, the freshwater flux into the high-latitude North and South Pacific Ocean increases by 0.05 Sv (15.2 %) and 0.03 Sv (11.5 %), respectively.

Vegetation-climate interactions in the warm mid-Cretaceous

J. Zhou et al.

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

◀

▶

◀

▶

Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion

The presence of vegetation leads to significant regional changes in precipitation (Fig. 4d), though the large-scale precipitation pattern remains unchanged. The most significant regional changes ($> 1.0 \text{ mm day}^{-1}$) occur over the low-to-mid latitude forests, intertropical convergence zones (ITCZ) and mid-latitude Western Pacific. The increases in continental precipitation are mainly due to the enhanced recycling of water via canopy evapotranspiration, with a point-to-point correlation larger than 0.6 over most areas. In contrast, the changes in precipitation over ocean mainly reflect changes in large-scale atmospheric circulation, as shown by the changes in zonal-average annual mean meridional streamfunction (Fig. 7a). The Hadley cells weaken by $3.2 \times 10^9 \text{ kg s}^{-1}$ (3%) and $6.3 \times 10^9 \text{ kg s}^{-1}$ (7%) in the Northern and Southern Hemisphere, respectively, due to the reduced equator-to-pole temperature gradient. In addition, the ascending branches of the Hadley cells contract equatorward (from $\sim 16.7^\circ \text{ N/S}$ to 13.0° N/S). The equatorward contraction in each hemisphere mainly occurs during winter and spring seasons when the Hadley cell expands to the other hemisphere (not shown). The weakening and contraction of Hadley cells cause anomalous ascending motion at equatorial latitudes ($\sim 5^\circ \text{ S}$ – 5° N) and ~ 20 – $40^\circ \text{ N}^\circ \text{ S}$ and anomalous subsidence at ~ 10 – $20^\circ \text{ N}^\circ \text{ S}$. As a result, precipitation increases at $\sim 5^\circ \text{ S}$ – 5° N and 20 – $40^\circ \text{ N}^\circ \text{ S}$ and decreases at ~ 10 – $20^\circ \text{ N}^\circ \text{ S}$. The increase in precipitation at mid-latitude North Pacific is associated with warming due to a poleward shift of the western boundary currents.

3.2.3 Ocean circulation response

The reduction in the equator-to-pole temperature gradient in the $10\times\text{DGVM}$ experiment leads to a decrease in zonal wind strength. As shown in Fig. 7b, the maximum zonal-mean wind stress over the Pacific Ocean decreases by 0.02 N m^{-2} ($\sim 13\%$) in the Southern Hemisphere and 0.03 N m^{-2} ($\sim 20\%$) in the Northern Hemisphere. As a result, the maximum intensity of the subtropical gyres decreases by 7.9 Sv (10%) from 77.1 Sv ($10\times\text{BG}$) to 69.2 Sv ($10\times\text{DGVM}$) in the Northern Hemisphere and by 14.0 Sv (13.3%) from 105.4 Sv ($10\times\text{BG}$) to 91.4 Sv ($10\times\text{DGVM}$) in the Southern Hemisphere.

In addition, the mean position of the maximum wind stress shifts poleward by a grid cell (3.75°) from 50.1°N to 53.8°N in the Northern Hemisphere. Consistent with this poleward shift in wind stress, the western boundary current shifts from $\sim 43^\circ\text{N}$ in the $10\times\text{BG}$ experiment to $\sim 46^\circ\text{N}$ in the $10\times\text{DGVM}$ experiment in the western North Pacific. This poleward displacement of warm, saline subtropical water leads to the significant increase in temperature ($> 5^\circ\text{C}$) (Fig. 4b) and salinity ($> 2\text{PSU}$) at $\sim 45^\circ\text{N}$.

The global MOC also decreases in the $10\times\text{DGVM}$ experiment. As shown in Fig. 8, the maximum magnitude of MOC decreases by 1.0Sv ($\sim 11\%$) from 9.3Sv in the $10\times\text{BG}$ experiment to 8.3Sv in the $10\times\text{DGVM}$ experiment in the Southern Hemisphere and decreases by 4.5Sv ($\sim 21\%$) from 21.4Sv in the $10\times\text{BG}$ experiment to 16.9Sv in the $10\times\text{DGVM}$ experiment in the Northern Hemisphere. This decrease is 0.8Sv larger than CO_2 -induced decrease between the $10\times\text{DGVM}$ and $1\times\text{DGVM}$ (20.6Sv) experiments. The slight weakening of MOC in the Southern Hemisphere is caused by warming (Fig. 4b) and freshening through enhanced precipitation (Fig. 4d) of the high-latitude South Pacific. In addition to the freshening via enhanced precipitation, the poleward migration of the low-salinity subpolar front associated with the poleward shift of the western boundary currents, also contributes to the decrease in seawater salinity at the high-latitude North Pacific. As a result, deep water formation along the Northeast coast of Asia vanishes and MOC slows down in the Northern Hemisphere. In response to the weakening of MOC, poleward oceanic heat transport decreases by 0.07PW ($\sim 6\%$) in the Southern Hemisphere and 0.23PW ($\sim 13\%$) in the Northern Hemisphere (Table 3). This decrease in poleward heat transport diminishes the warming over the high-latitude Pacific Ocean ($\sim 70^\circ\text{S}/^\circ\text{N}$) (Fig. 4b).

4 Discussion

4.1 The role of vegetation feedbacks under high CO_2

Our results show that vegetation acts to warm high latitudes and cool low latitudes in a warm climate with high CO_2 , similar to the role of modern vegetation (Bonan, 2008).

Vegetation-climate interactions in the warm mid-Cretaceous

J. Zhou et al.

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

⏪

⏩

◀

▶

Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion



Vegetation-climate interactions in the warm mid-Cretaceous

J. Zhou et al.

[Title Page](#)[Abstract](#)[Introduction](#)[Conclusions](#)[References](#)[Tables](#)[Figures](#)[⏪](#)[⏩](#)[◀](#)[▶](#)[Back](#)[Close](#)[Full Screen / Esc](#)[Printer-friendly Version](#)[Interactive Discussion](#)

The simulated warming effect of high-latitude forests on polar regions is also consistent with previous Cretaceous studies (Deconto et al., 2000; Otto-Bliesner and Upchurch, 1997; Upchurch et al., 1998). However, our simulated reduction in high-latitude surface albedo (~ 0.06) and the associated warming ($< 2^\circ\text{C}$) are more than one-half less than in these studies, mainly because we assess the impact of vegetation in a warm mean climate with little snow (< 0.15) and sea ice (< 0.01) (Table 3). In contrast, using an atmospheric CO_2 value of 580 ppmv, Upchurch et al. (1998) report that snow and sea ice fraction are up to 0.62 and 0.38, allowing a stronger vegetation-snow-sea ice albedo feedback. The weaker warming here is also linked to the decrease in poleward oceanic heat transport resulting from the weakening of MOC, which is not resolved in the earlier studies. The simulated mean annual temperatures over the Eurasian interior and Arctic Ocean only increase $\sim 2^\circ\text{C}$ by vegetation and are up to 10°C lower than those estimated by proxies such as leaf physiognomy (Spicer et al., 2002, 2008) and TEX_{86} method (Jenkyns et al., 2004). Therefore, we conclude that high-latitude forests can only be considered a secondary mechanism for the mid-Cretaceous polar warmth. Other mechanisms in addition to high atmospheric CO_2 and high-latitude vegetation are required to maintain the polar warmth.

We find that $\sim 20\text{--}40\%$ of the increase in surface absorbed solar radiation over high-latitude forests is due to cloud feedbacks. In fact, the increase in cloud radiative forcing surpasses the albedo feedback during late spring and early summer (Fig. 5d, e). This positive cloud feedback contradicts a previous study (Betts, 1999), which reports that the presence of vegetation induced a high-latitude summer cooling of $> 2^\circ\text{C}$ mainly by reducing surface solar radiation via enhanced cloud cover. These contrasting results highlight the uncertainty of cloud feedbacks in determining the overall effect of high-latitude forests.

4.2 Oceanic circulation

Our simulations are consistent with modeling studies with a dynamic ocean component that vegetation can have an important impact on MOC, due to vegetation-induced

changes in hydrological cycle and sea surface temperature (Brovkin et al., 2002, 2009; Ganopolski et al., 1998; Lohmann et al., 2006). Moreover, our simulations suggest that the response of MOC to vegetation could exceed that of a tenfold increase in atmospheric CO₂ concentration. Proxy records and model studies demonstrate that global vegetation pattern is highly correlated with climate changes during the geological past and future (e.g., Spicer and Chapman, 1990; Alo and Wang, 2008; Cramer et al., 2001). Therefore, vegetation-climate interactions should be taken into account for simulating the MOC response to past and future climate change.

5 Caveats

In the absence of global paleovegetation reconstructions, most previous modeling studies of the mid-Cretaceous have assumed a uniform vegetation type on all continents (e.g., Donnadieu et al., 2006; Poulsen et al., 2007; Zhou et al., 2008). In this contribution, we use a dynamic vegetation model to overcome this limitation. We emphasize, though, that the simulation of pre-Quaternary vegetation is subject to uncertainty, mainly due to a paucity of evidence for the construction and physiology of past ecosystems. Due to our lack of understanding of past ecosystems, paleo-vegetation modeling relies on a modern understanding of PFTs and their bioclimatic, physiological, and dynamic relationships that may not be entirely appropriate for past times. For example, grass, which is a dominant PFT at present, was not ecologically dominant in mid-Cretaceous. We keep grass in our simulation because it is the only available herbaceous PFT. Another substantial change between the modern and mid-Cretaceous is angiosperms. The modern angiosperm-dominated tropical forests may not have appeared until the Paleocene (Burnham and Johnson, 2004). Yet our simulations show extensive tropical forests resembling the modern distribution, an overestimation that may be linked to the inclusion of PFTs that represent modern angiosperm trees. Consequently, CLM-DVGM may overestimate the impact of low-latitude vegetation on mid-Cretaceous climate, because angiosperms have a much larger impact on hydrology

Vegetation-climate interactions in the warm mid-Cretaceous

J. Zhou et al.

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures



Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion



relative to other plants due to their high transpiration capacity (Boyce and Lee, 2010). To more accurately simulate the mid-Cretaceous vegetation and its feedbacks, more paleo-vegetation data is needed to appropriately parameterize the mid-Cretaceous dominant PFTs.

6 Conclusions

We investigate the vegetation-climate interaction using a coupled ocean-atmosphere GCM (CCSM3) with a dynamic vegetation component. The model demonstrates a positive feedback between vegetation and polar climate – CO₂-induced warming promotes expansion of high-latitude forests, which leads to additional polar warming via reducing surface albedo and low cloud cover. However, the warming effect of vegetation at high latitudes is smaller than reported in previous studies, mainly due to the less pronounced surface albedo feedback in the absence of large snow and sea ice cover under the simulated warm mean climate. Over low latitude continents, the feedbacks between vegetation and climate are opposite; forest cover decreases with higher CO₂, and the presence of forests leads to year-round cooling. The cooling is associated with enhanced evapotranspiration and low cloud cover.

The simulation of mid-Cretaceous vegetation has unexpected effects on the ocean. First, as a result of the enhanced atmospheric vapor content with vegetation, downwelling longwave radiation increases over ocean, leading to surface warming at sea surface. Second, vegetation alters wind fields, resulting in weakening of wind-driven subtropical and subpolar gyres and poleward displacement of the North Pacific western boundary current. Lastly, the intensified precipitation enhances the freshwater flux into the high-latitude Pacific. This freshening in conjunction with surface warming reduces the meridional overturning circulation in both hemispheres, which reduces meridional heat transport and damps high-latitude warming.

Vegetation-climate interactions in the warm mid-Cretaceous

J. Zhou et al.

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures



Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion



Acknowledgement. We thank Esther C. Brady for her help in setting up the simulations. This study was financially supported by grants (0433440 and 0310032) from the National Science Foundation's Atmospheric Program (to C. Poulsen, D. Pollard, and T. White) and the Barbour Scholarship from University of Michigan (to J. Zhou). The CCSM3 mid-Cretaceous simulations were run at the National Center for Atmospheric Research (NCAR).

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Vegetation-climate interactions in the warm mid-Cretaceous

J. Zhou et al.

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures



Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion



Vegetation-climate interactions in the warm mid-Cretaceous

J. Zhou et al.

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

◀

▶

◀

▶

Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion



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**Vegetation-climate
interactions in the
warm
mid-Cretaceous**

J. Zhou et al.

[Title Page](#)[Abstract](#)[Introduction](#)[Conclusions](#)[References](#)[Tables](#)[Figures](#)[⏪](#)[⏩](#)[◀](#)[▶](#)[Back](#)[Close](#)[Full Screen / Esc](#)[Printer-friendly Version](#)[Interactive Discussion](#)

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Vegetation-climate interactions in the warm mid-Cretaceous

J. Zhou et al.

[Title Page](#)[Abstract](#)[Introduction](#)[Conclusions](#)[References](#)[Tables](#)[Figures](#)[⏪](#)[⏩](#)[◀](#)[▶](#)[Back](#)[Close](#)[Full Screen / Esc](#)[Printer-friendly Version](#)[Interactive Discussion](#)

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Vegetation-climate interactions in the warm mid-Cretaceous

J. Zhou et al.

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

⏪

⏩

◀

▶

Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion



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Vegetation-climate interactions in the warm mid-Cretaceous

J. Zhou et al.

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

⏪

⏩

◀

▶

Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion

Table 1. Bioclimatic parameters for survival and establishment of plant functional type (PFT). Adapted from Table 1 in Bonan et al. (2003).

PFT	$T_{c,\min}$ (°C)	$T_{c,\max}$ (°C)	T_{hs} (°C)	GDD_{\min} (°C day)	P_{\min} (mm yr ⁻¹)
Tropical broadleaf evergreen	15.5	NA	NA	NA	100
Tropical broadleaf deciduous	15.5	NA	NA	NA	100
Temperate needleleaf evergreen	-2.0	22.0	NA	900	100
Temperate broadleaf evergreen	3.0	18.8	NA	1200	100
Temperate broadleaf deciduous	-17.0	15.5	NA	1200	100
Boreal needleleaf evergreen	-32.5	-2.0	23	600	100
Boreal deciduous	NA	-2.0	23	350	100
C ₃ grasses	-17.0	NA	NA	0	100

All values here are based on 20-yr running mean. $T_{c,\min}$ = coldest minimum monthly air temperature for survival, $T_{c,\max}$ = warmest minimum monthly air temperature for establishment, T_{hs} = heat stress air temperature for survival, GDD_{\min} = minimum annual growing degree-days above 5°C, which is the sum of all positive values of difference between average daily temperature and 5°C in one year, P_{\min} = minimum annual precipitation for establishment. The only difference between our defined PFTs and the ones in Bonan et al. (2003) is our treatment of $T_{c,\max}$ for C₃ grasses, in order to allow C₃ grasses to expand over the C₄ grass range, we eliminate the maximum temperature threshold for C₃ grasses.

Vegetation-climate interactions in the warm mid-Cretaceous

J. Zhou et al.

Table 2. Mid-Cretaceous CCSM3 experiments and global annual mean surface temperature (SAT) and precipitation.

Experiment	Atm. CO ₂ (ppmv)	Vegetation	Integration (year)	SAT (°C)	Precipitation (mm day ⁻¹)
1×DGVM	280	DGVM	2125	13.6	2.90
10×BG	2800	Bare ground	1626	23.1	3.36
10×DGVM	2800	DGVM	1637	24.0	3.47
16×DGVM	4480	DGVM	1615	25.6	3.56

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

⏪

⏩

◀

▶

Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion

Table 3. Global and high-latitude annual mean climatic variables in the 10×BG and 10×DGVM experiments.

Variable/experiment	10×BG	10×DGVM	10×DGVM – 10×BG
SAT (land, 60–90° N, °C)	4.4	6.3	1.9
SAT (ocean, 60–90° N, °C)	8.1	10.0	1.9
SAT (land, 60–90° S, °C)	4.7	6.4	1.7
SAT (ocean, 60–90° S, °C)	11.0	12.3	1.3
Snow fraction (land, 60–90° N)	0.12	0.09	–0.03
Snow fraction (land, 60–90° S)	0.14	0.10	–0.04
Polar sea ice fraction	0.005	0.002	–0.003
Surface albedo (land)	0.203	0.163	–0.04
Surface albedo (land, 60–90° N)	0.273	0.216	–0.057
Surface albedo (land, 60–90° S)	0.304	0.242	–0.062
Surface LW↓ (ocean, W m ^{–2})	396.7	404.0	7.3
Latent heat fluxes↑ (W m ^{–2})	97.1	100.5	3.4
Latent heat fluxes↑ (land, W m ^{–2})	48.8	59.9	11.1
Precipitable water (kg m ^{–2})	37.8	41.1	3.3
Precipitation (land, mm day ^{–1})	2.69	2.95	0.26
Snow (land, mm day ^{–1})	0.12	0.10	–0.02
Precipitation (ocean, mm day ^{–1})	3.57	3.64	0.07
Evapotranspiration (land, mm day ^{–1})	1.69	2.07	0.38
MOC (Northern Hemisphere, Sv)	21.4	16.9	–4.5
MOC (Southern Hemisphere, Sv)	9.3	8.3	–1.0
OHT (Northern Hemisphere, PW)	1.84	1.61	–0.23
OHT (Northern Hemisphere, PW)	1.16	1.09	–0.07

Vegetation-climate interactions in the warm mid-Cretaceous

J. Zhou et al.

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

◀

▶

◀

▶

Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion



Vegetation-climate interactions in the warm mid-Cretaceous

J. Zhou et al.

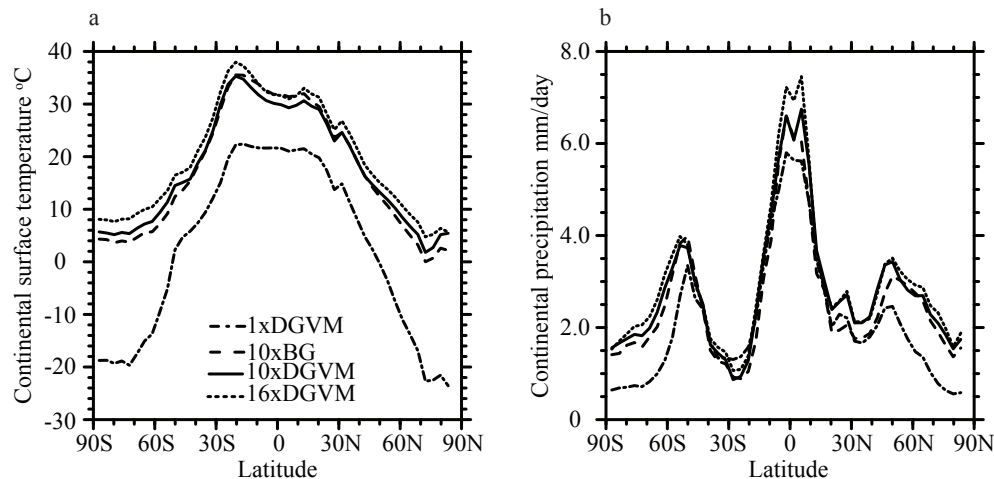


Fig. 1. Zonal-average annual mean continental surface temperature ($^{\circ}\text{C}$) (a); and precipitation (mm d^{-1}) (b) for the 1xDGVM (black dash-dot line), 10xBG (black dashed line), 10xDGVM (black solid line) and 16xDGVM (black dotted line) experiments, respectively.

[Title Page](#)[Abstract](#)[Introduction](#)[Conclusions](#)[References](#)[Tables](#)[Figures](#)[⏪](#)[⏩](#)[◀](#)[▶](#)[Back](#)[Close](#)[Full Screen / Esc](#)[Printer-friendly Version](#)[Interactive Discussion](#)

Vegetation-climate interactions in the warm mid-Cretaceous

J. Zhou et al.

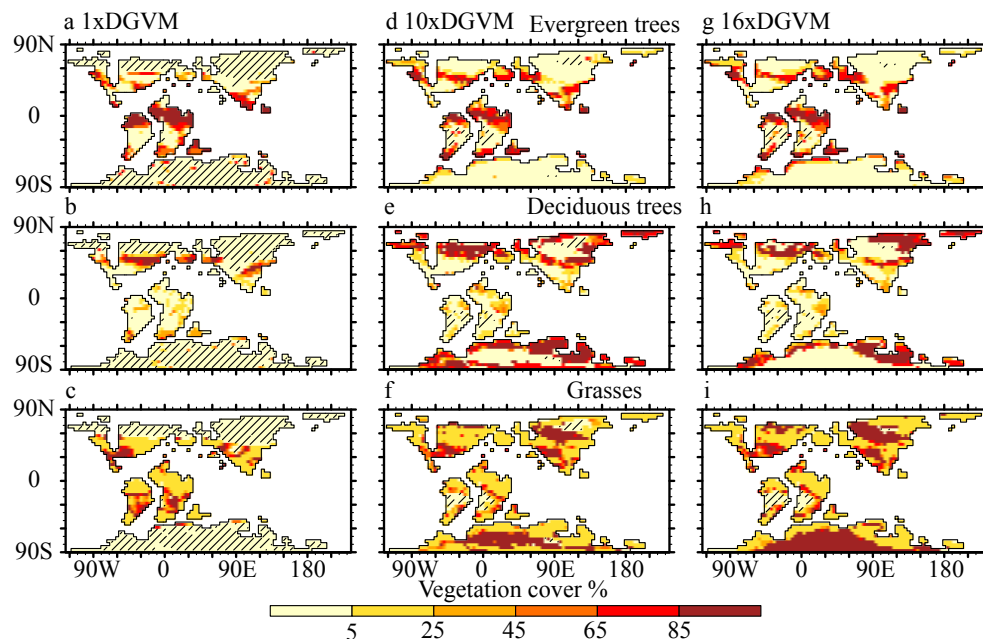


Fig. 2. Global distribution of evergreen trees (a), (d), (g), deciduous trees (b), (e), (h) and grasses (c), (f), (i) (%) for the 1×DGVM, 10×DGVM, and 16×DGVM experiments, respectively. Land area with < 10% vegetation cover, which is defined as desert, is shaded. Note, hereafter, continental outlines delimit area with > 50% land cover. Evergreen trees include tropical and temperate broadleaf evergreen, and temperate and boreal needleleaf evergreen. Deciduous trees include tropical and temperate broadleaf deciduous and boreal deciduous.

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

◀

▶

◀

▶

Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion

Vegetation-climate interactions in the warm mid-Cretaceous

J. Zhou et al.

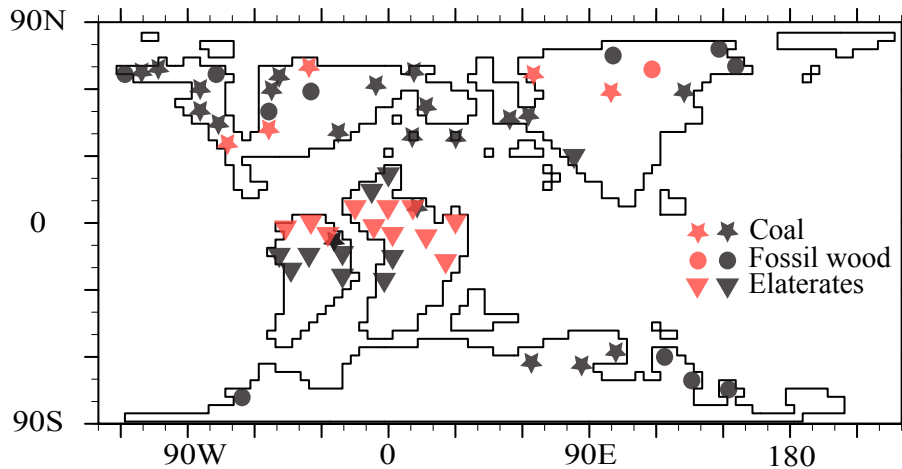


Fig. 3. Distribution of mid-Cretaceous vegetation inferred from fossil wood, pollen of Elaterate assemblages and coal deposits and comparison with simulated vegetation. Fossil wood (solid circle) and coal deposits (solid star) are indicators of high-productivity forests; fossil pollen of Elaterates assemblages (down triangle) indicates dry climate and/or mangrove swamps and savannah (Dino et al., 1999; Ziegler et al., 2003). Paleofloral data is adapted from Beerling et al. (1999), Dino et al. (1999), Falcon-Lang et al. (2001), Herman and Spicer (1996), and Price et al. (1997). The shading of symbols shows agreement between the simulated vegetation in the 10×DGVM experiment and paleofaunal data. Black indicates that the model simulates the observed floral type; red indicates that it does not. The model-data comparison (not shown) is nearly identical for the 16×DGVM experiment.

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

◀

▶

◀

▶

Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion

Vegetation-climate interactions in the warm mid-Cretaceous

J. Zhou et al.

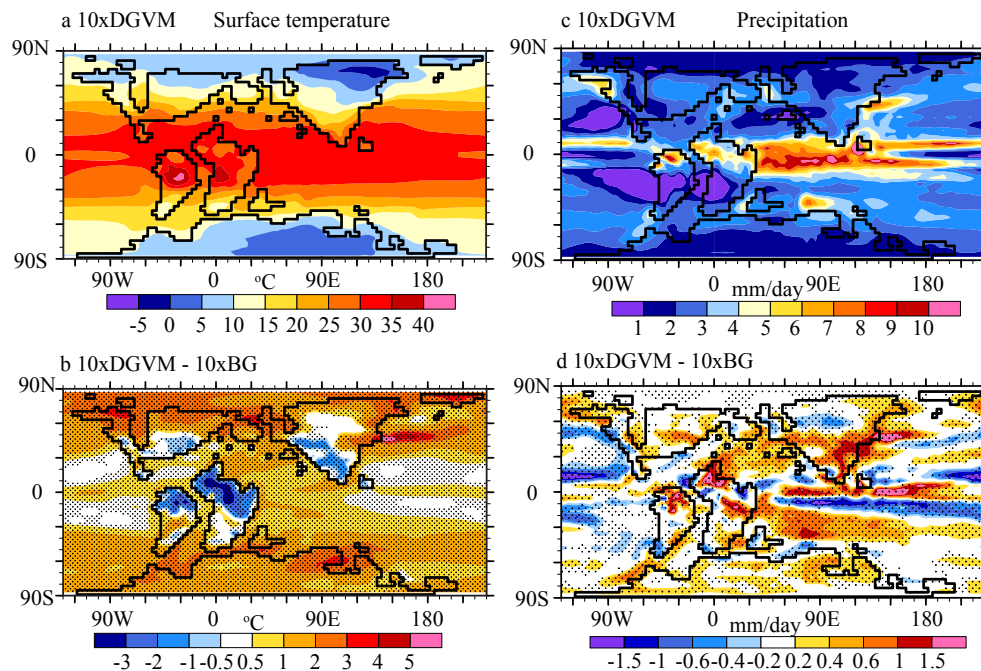


Fig. 4. Mean annual surface temperature ($^{\circ}\text{C}$) and precipitation (mm day^{-1}) for **(a, c)** 10x DGVM experiment; and **(b, d)** 10x DGVM – 10x BG experiments. In **(b, d)**, the differences exceed 95 % confidence level are shaded.

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

◀

▶

◀

▶

Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion

Vegetation-climate interactions in the warm mid-Cretaceous

J. Zhou et al.

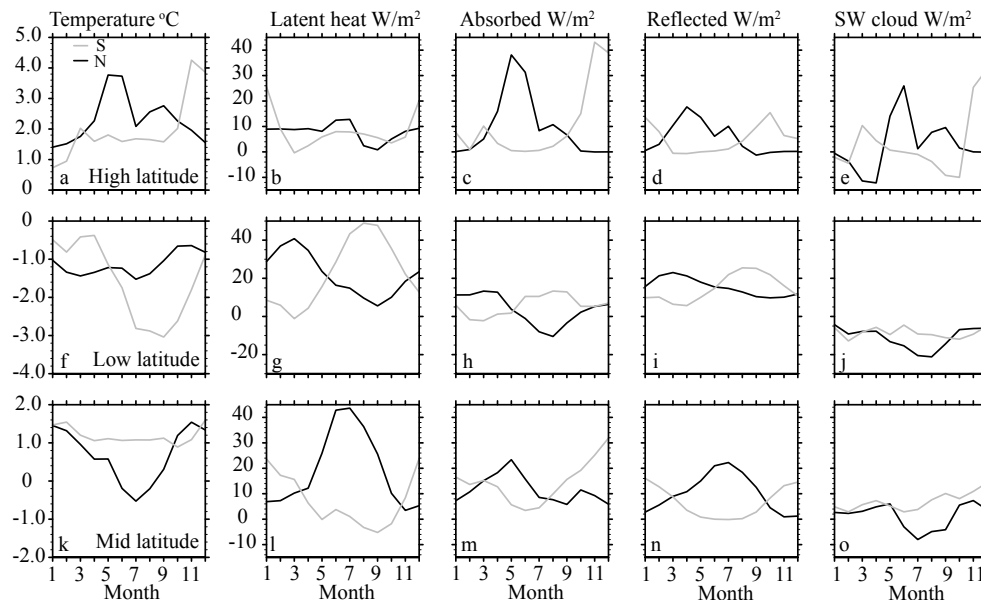


Fig. 5. Annual cycles of surface temperature ($^{\circ}\text{C}$), latent heat fluxes (Wm^{-2}), surface absorbed solar radiation (Wm^{-2}), surface reflected solar radiation (Wm^{-2}), surface shortwave cloud forcing (Wm^{-2}) over the high-latitude forests (**a–e**), low-latitude forests (**f–j**) and mid-latitude forests (**k–o**) for $10\times\text{DGVM} - 10\times\text{BG}$ experiments. Solid black lines denote changes in the Northern Hemisphere and grey lines denote changes in the Southern Hemisphere.

[Title Page](#)
[Abstract](#)
[Introduction](#)
[Conclusions](#)
[References](#)
[Tables](#)
[Figures](#)
[⏪](#)
[⏩](#)
[◀](#)
[▶](#)
[Back](#)
[Close](#)
[Full Screen / Esc](#)
[Printer-friendly Version](#)
[Interactive Discussion](#)

Vegetation-climate interactions in the warm mid-Cretaceous

J. Zhou et al.

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

◀

▶

◀

▶

Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion

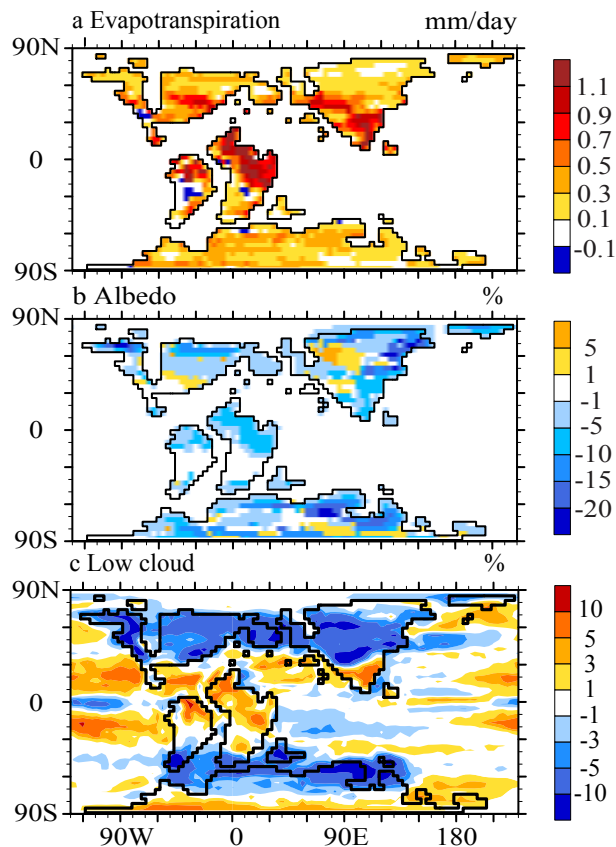


Fig. 6. Annual mean **(a)** total evapotranspiration over land (mm d^{-1}); **(b)** surface albedo (%); and **(c)** low cloud cover (%) for the $10\times\text{DGVM}$ – $10\times\text{BG}$ experiments.

Vegetation-climate interactions in the warm mid-Cretaceous

J. Zhou et al.

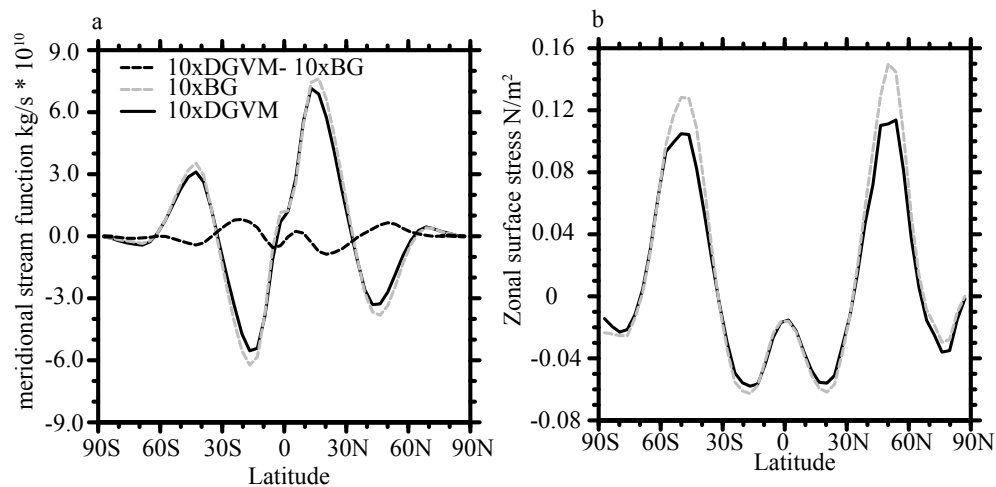


Fig. 7. Zonal-average annual mean meridional streamfunction ($10^{10} \text{ kg s}^{-1}$) at 400 mb **(a)**; and wind stress (N m^{-2}) over the Pacific Ocean **(b)** for the 10xDGVM (black solid line), 10xBG (grey dashed line) and 10xDGVM – 10xBG experiments (black dashed line in **(a)**), respectively.

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

◀

▶

◀

▶

Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion

Vegetation-climate interactions in the warm mid-Cretaceous

J. Zhou et al.

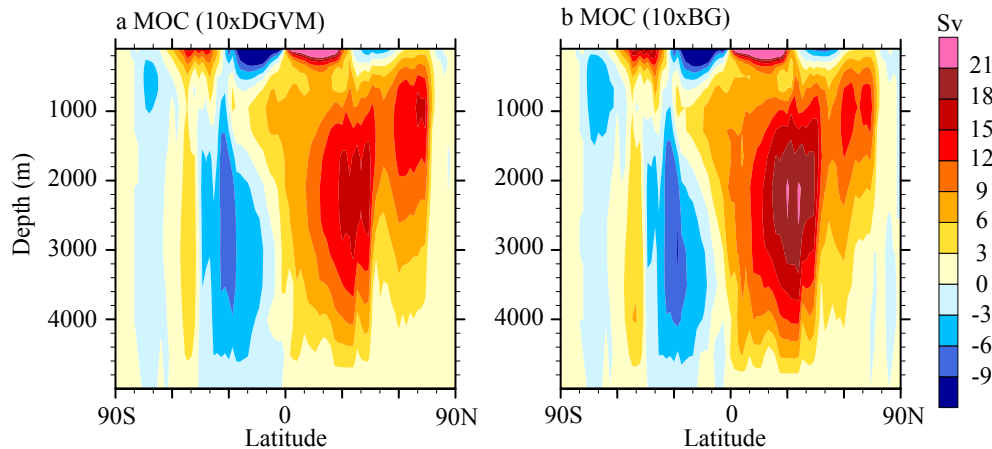


Fig. 8. Annual mean Meridional overturning circulation (Sv) for **(a)** 10×DGVM experiment; and **(b)** 10×BG experiment. Positive (in yellow and red) means clockwise flow direction and negative means counterclockwise flow direction.

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

◀

▶

◀

▶

Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion