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# Effect of vegetation on the Late Miocene ocean circulation

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## Abstract

A weak and shallow thermohaline circulation in the North Atlantic Ocean is related to an open Central American gateway and exchange with fresh Pacific waters. We estimate the effect of vegetation on the ocean general circulation using the atmospheric circulation model simulations for the Late Miocene climate. Caused by an increase in net evaporation in the Miocene North Atlantic, the North Atlantic water becomes more saline which enhances the overturning circulation and thus the northward heat transport. This effect reveals a potentially important feedback between the ocean circulation, the hydrological cycle and the land surface cover for Cenozoic climate evolution.

## 1 Introduction

The Eocene-Oligocene and the Mid-Miocene climate transitions are two major cooling steps in the Cenozoic climate evolution (Zachos et al., 2001) from greenhouse to “icehouse” climate conditions. A drastic increase in the oxygen-isotopic composition measured in benthic foraminifer shells points to a combination of Antarctic ice growth and global cooling at 34 Ma and 14 Ma (Ma: million years before present), respectively, which is also indicated by the occurrence of Southern Ocean ice-rafted detritus and eustatic sea-level change (Miller et al., 1987; Kennett and Barker, 1990; Billups and Schrag, 2002). Ocean circulation changes and atmospheric pCO<sub>2</sub> variations are often cited as potential catalysts of these cooling events (DeConto and Pollard, 2003). Large-scale ocean circulation changes, caused by atmospheric circulation changes and/or by tectonic reorganizations of gateway regions, may have altered poleward transports of heat and moisture, which in turn may have resulted in Antarctic ice growth and global cooling (Kennett, 1977; Zachos et al., 2001). Ocean circulation hypotheses are supported by C-13 proxy evidence (e.g. Wright and Miller, 1996; Billups, 2002) and the timing of tectonic events at critical ocean pathways like the Drake Passage, the Tasmanian Seaway, the Indonesian Throughflow (Cane and Molnar, 2001; Lawver and Gaha-

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gan, 2003), the eastern Tethys (Flower and Kennett, 1994), and the Central American Seaway (e.g., Haug and Tiedemann, 1998).

Here, we will examine climate processes in connection with large-scale ocean circulation changes for a selected Cenozoic time slice, namely the Late Miocene or Tortonian (11–7 Ma). The Tortonian was characterized by intensive Antarctic glaciation and the buildup of ice sheets in the North Atlantic realm. Specifically, we focus on the spatial temperature distribution, which is a principal problem in understanding Cenozoic climate change. In the case of the Miocene, elevated global-mean surface temperatures and weak equator-to-pole temperature gradients are proposed (Greenwood and Wing, 1995; Crowley and Zachos, 2000). While numerical simulations exhibit rising global-mean temperatures for increasing greenhouse gas concentrations, they fall far short of attaining the reconstructed reduction in the meridional temperature gradient (Barron, 1987; Huber and Sloan, 2001). Since it seems that atmospheric carbon dioxide concentration hardly varied during the Miocene (Pagani et al., 1999, 2005, Pearson and Palmer, 2000), the mechanism which causes bipolar glaciation in the Tortonian remains even more enigmatic.

Some authors (Schmidt and Mysak, 1996; Hay et al., 1997) have suggested that atmospheric heat transport may have played an important role in resolving this “low gradient paradox”. It is plausible to expect a warmer atmosphere to transport more latent heat poleward, helping to reduce meridional temperature gradients. However, despite the exponential increase of saturation vapor pressure with temperature, this feedback becomes less powerful as temperature rises (Caballero and Langen, 2005). Other possible mechanisms located in the atmosphere involve the atmospheric stationary wave response due to changing paleogeography and sea level.

On the other side, marine proxy data indicate that ocean gateway changes and major reorganizations of the global ocean circulation (e.g. Kennett, 1977; Wright et al., 1992; Zachos et al., 2001) are consistent with a weakening of the ocean heat transport during the Miocene. Concerning the Tortonian, the (then) still opened Central American Seaway (CAS, i.e., the Panama Strait) allowed for the exchange of saline Atlantic water

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with comparatively fresher Pacific water, and it has been shown that this leads to weakening of the thermohaline circulation in the North Atlantic Ocean (e.g., Mikolajewicz et al., 1993; Bice et al., 2000; Butzin et al., 2006<sup>1</sup>). Therefore, the global ocean circulation seems not to be a proper candidate to be responsible for a weaker equator-to-pole temperature gradient.

The question of temperature gradients might be linked to other feedbacks in the climate system, such as changes in the hydrological cycle and vegetation cover. Paleontological and palynological data give evidence for drastic changes in vegetation and therefore climate during the Cenozoic (Retallack, 2001; Willis and Mc Elwain, 2002). For example, during the Eocene/Oligocene glaciation tropical rain forests virtually disappeared poleward of the northern and southern high pressure zones. Grasslands, which had begun to develop under dry conditions during the Eocene, became more and more widespread in the Oligocene. During the Mid-Miocene Climatic Optimum, moist warm forests expanded poleward of the subtropical high pressure zones for a short period. Following the global climatic deterioration after the Mid-Miocene Climatic Optimum, tropical rain forests withdrew again to the equatorial zone. Grasslands and deserts expanded through much of the lower mid-latitudes (Morley, 2000; Bredenkamp et al., 2002). C4 grasslands became widespread during the interval from about 8 to 5 Ma (Cerling et al., 1997; Freeman and Colarusso, 2001). During the Miocene, most of the climatically arranged vegetation belts developed ranging from rain forest along the equator to polar desert at high latitudes. However, to date, little is known about the role of continental vegetation for climate change during the Cenozoic. It is still an open question whether the vegetation just has adapted to hydrological changes or whether it has played an active role as a modifier of major climate transitions. In principle, the vegetation can contribute to a weaker-than-present meridional temperature gradient through modifying the local albedo (e.g., Dutton and Baron, 1997; Otto-Bliesner and

<sup>1</sup>Butzin, M., Lohmann, G., and Bickert, T.: Effect of ocean gateways onto the evolution of the ocean circulation and marine carbon cycle during the Miocene: Sensitivity studies with an ocean-circulation carbon cycle model, *Paleoceanography*, submitted, 2006.

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Upchurch, 1997).

In the light of these findings, we investigate whether such a feedback was effective during the Late Miocene. We examine if the Tortonian vegetation significantly enhanced the hydrological cycle with increased precipitation rates over continental areas providing for a greener land surface. In particular, we are interested in the climate sensitivity of the thermohaline circulation (THC) with the vegetation cover and associated hydrological cycle. For that purpose, we apply an atmospheric circulation model (AGCM) in combination with a coarse resolution model of the ocean. A dynamical vegetation model is used to evaluate the consistency between reconstructed and simulated vegetation cover. The models and experiments are briefly described in the following Section.

## 2 Methods

### 2.1 Atmospheric circulation model

For the Late Miocene climate simulations, we apply the atmosphere general circulation model ECHAM4 (Roeckner et al., 1996). The prognostic variables are calculated in the spectral domain with a triangular truncation at wave number 30 (T30), which corresponds to a Gaussian longitude-latitude grid of approximately  $3.75^\circ$ . The vertical domain is represented by 19 hybrid sigma-pressure (terrain following) levels with the highest level at 10 hPa. The model is coupled to a 50 m slab ocean. This allows a prescription of the Miocene ocean heat transport consistent with proxy data (Steppuhn et al., 2006). Furthermore, the orography is adapted to the Tortonian when the height of mountain ranges was generally reduced. For example, Greenland reaches only about a tenth of its recent elevation. In addition to the above-described boundary conditions, the atmospheric  $\text{CO}_2$  is specified with the present-day level of 353 ppmv for all experiments. This lies within the spectrum of values which are given for the Miocene (Cerling et al., 1997, Pagani et al., 1999; Pearson and Palmer, 2000). For the land surface, sen-

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sitivity experiments were performed which are described below. Each model simulation with the AGCM was run over 20 years. The model reaches an equilibrium state after 5 years, and the last 10 years are taken into account for further analysis. A list of the experiments is given here:

5 **CTRL:** Present day control simulation (Roeckner et al., 1996).

**TGEO:** Tortonian simulation with adapted geography (Steppuhn et al., 2006). The global vegetation represents modern conditions, except that the recent Greenland ice cap is replaced by tundra vegetation.

10 **TVEG:** Tortonian simulation with adapted geography as in TGEO and reconstructed vegetation cover. The Tortonian vegetation was reconstructed on the basis of palaeobotanical data such as fossil pollen and leaf data, and fossil car-poflora (Micheels, 2003). Figure 1 shows the resulting reconstruction of the global Tortonian vegetation. The Tortonian palaeovegetation was generally more lush as compared to today, tropical forests expanded and their margins shifted further poleward. According to the reconstruction of the Tortonian vegetation, land sur-  
15 face parameters are adapted. To consider the changed vegetation in the model, data for the albedo, the leaf area index, the vegetation and forest cover, and the maximum soil water capacity are changed.

## 2.2 Dynamical vegetation model

20 The LPJ dynamical vegetation model (Sitch et al., 2003) combines process-based descriptions of terrestrial ecosystem structure (vegetation composition, biomass and height) and function (energy absorption, carbon cycling). Vegetation composition is described by nine different plant functional types (PFTs), which are distinguished ac-  
25 cording to their physiological (C3, C4 photosynthesis), morphological (tree, grass) and phenological (deciduous, evergreen) attributes. The model is run on a grid cell basis with input of soil texture, monthly fields of temperature, precipitation, as well as short

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and long wave radiation. Each grid cell is divided into fractions covered by the PFTs and bare ground. Both the presence and the covered fraction of PFTs within a grid cell depend on their specific environmental limits and on resource competition among the PFTs. Carbon isotope fractionation is included in the model (Kaplan et al., 2002; Scholze et al., 2003). The model is run on a horizontal  $2^\circ \times 2^\circ$  grid, directly forced with the output of the AGCM experiments.

### 2.3 Ocean circulation model

Our ocean model is an updated version of the LSG circulation model developed by Maier-Reimer et al. (1993). We implemented some significant improvements such as a new advection scheme for tracers (Schäfer-Neth and Paul, 2001; Prange et al., 2003) as well as an overflow parametrization for the bottom boundary layer (Lohmann, 1998; Lohmann and Schulz, 2000). The spatial resolution is  $3.5^\circ \times 3.5^\circ$  in the horizontal and 22 levels in the vertical. We calibrated the model by simulating anthropogenic C-14 (Butzin et al., 2005). The ocean is forced by ten-year averaged monthly fields of wind stress, surface air temperature, and freshwater flux, which serve as background climatology and originate from the simulations with the atmosphere general circulation model ECHAM4 described in Sect. 2.1. A surface heat flux formulation based on atmospheric energy balance model considerations permits that sea surface temperatures (SST) can freely adjust to ocean circulation changes (e.g., see Prange et al., 2003; Knorr and Lohmann, 2003; Butzin et al., 2005). The hydrological cycle is closed by a runoff scheme which considers continental catchment areas and allows for variable land-sea distributions, which permits that sea surface salinities (SSS) can freely evolve. The total integration time of each experiment is 5000 years. For the late Miocene simulations, we assumed a 500 m deep and three gridpoints wide (between  $9^\circ$  N and  $18^\circ$  N) gateway between the Atlantic and Pacific Oceans.

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### 3 Results

#### 3.1 Hydrological cycle and vegetation cover

The control climate simulates the mean hydrological cycle reasonably well as shown by Arpe et al. (2000) and is in agreement with observations (e.g., Peixoto and Oort, 1992; Zaucker and Broecker, 1992). The subtropical highs over the North and South Atlantic and Pacific oceans provide a moisture transport from the subtropics to higher latitudes. In the tropics between 20° S and 20° N, strong easterlies are observed, especially over the Atlantic and Pacific Oceans.

Figure 2 indicates strong changes in the hydrological cycle when comparing TVEG and CTRL. Boreal summer precipitation over the Sahel region is strongly increased for the green Sahara (compare Fig. 1). In accordance with the removal of the inland ice of Greenland, sea ice is drastically reduced caused by a considerably increasing surface temperatures and the ice-albedo feedback, and local precipitation is increased over Northern Greenland (Fig. 2c). The Icelandic Low is slightly shifted to the south-east leading to more precipitation off western Europe and less precipitation between Greenland and Iceland (Fig. 2).

The reduced ocean heat transport causes a southward migration of the thermal equator in both Tortonian simulations TGEO and TVEG. When comparing TVEG with CTRL in Fig. 2, the Intertropical Convergence Zone moves southward resulting in enhanced water vapor export out of the Atlantic catchment area. We evaluate an additional moisture transport from the Atlantic to the Pacific Ocean accounting for an increase of net Atlantic evaporation (0.12 and 0.31 Sv for TGEO and TVEG, respectively). The unit 1 Sv corresponds to a mass transport of  $10^9 \text{ kg s}^{-1}$ , equivalent to a volume transport of  $10^6 \text{ m}^3 \text{ s}^{-1}$  liquid water.

In order to check the consistency of the reconstructed vegetation distribution with the modelled climate in TVEG, we apply the dynamical vegetation model LPJ. We use the monthly output of the last 10 years of the CTRL and TVEG simulations, iterating these simulations 200 times in order to get an equilibrium of the dynamical vegetation model

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after 2000 model years. We build an average over the last 500 years and identify the spatial patterns of the PFTs (in %) for the Tortonian and present-day vegetation cover (Fig. 3). For the late Miocene, tropical trees are spread in the subtropical Africa (North and South) and parts of Australia (Fig. 3a), whereas temperate trees are extended over Asia (Fig. 3b) relative to present conditions. The extension of boreal forests far into the northern high latitudes during the Tortonian (Fig. 3c) is in accordance with proxy data (Boulter and Manum, 1997). Grassland is extended into subtropical areas, over Greenland and over Alaska. The Sahara desert is smaller than today and consists of steppe and open grassland rather than sand desert which is consistent to fossil data (Le Houerou, 1997; Schuster et al., 2006).

### 3.2 Ocean circulation

In the ocean circulation experiments, we employ a hybrid coupled modeling approach, which allows an adjustment of surface temperatures and salinity to changes in the ocean circulation, based on an atmospheric energy balance model (Lohmann and Gerdes, 1998; Prange et al., 2002). No flux correction is applied for present day and other climate conditions. The control experiment for present-day conditions (Fig. 4a) reasonably reflects the modern Atlantic Ocean circulation with a southward water export of 16 Sv at 30° S and a heat transport of 0.96 PW (1 PW=10<sup>15</sup> W) at 30° N, which is in the range of oceanographic observations (Schmitz, 1995; Macdonald and Wunsch, 1996).

A comparison of the control run with the Tortonian experiments (TGEO, TVEG) reveals significant changes in the meridional overturning circulation (Fig. 4bc): The formation of deep water in the North Atlantic is strongly reduced (TGEO) when the Central American Seaway (CAS) is open (Fig. 4). The meridional circulation is only 3 Sv and represents a “mini-conveyor belt” circulation with an ocean heat transport at 30° N of 0.19 PW (Fig. 4b). In experiment TVEG, the circulation strength is similar to the present-day circulation (14 Sv export at 30° S, 0.83 PW at 30° N), but slightly shallower than under present-day conditions. The reason might be the increased flow of bottom

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water from the Antarctic (Fig. 4c).

A detailed analysis of the flow patterns in various depths of the Panamanian gateway shows an export of surface water from the Atlantic to the Pacific Ocean (Fig. 5a). An import of thermocline and intermediate layer water from the Pacific to the Atlantic Ocean is responsible for a reversal of the Northeast Brazil Current (Fig. 5a). The net flux of Pacific water through the CAS into the Atlantic leads to relatively low-salinity thermocline water which hinders deep water formation in the North Atlantic. In T GEO, the surface winds and net freshwater flux in the North Atlantic are not able to overcome this freshening (Fig. 4b), whereas the background conditions in TVEG with stronger northward flow (Fig. 5b) and increase in net evaporation are sufficient to push the ocean circulation into a present-day-like circulation mode (Fig. 4c). Both the increased ocean circulation with a northward shift of the Arctic sea ice, and a local warming associated to the land surface quantities, induce an anomalous warming between TVEG and T GEO of up to 8°C (Fig. 5b).

Caused by the drop in ocean circulation in T GEO relative to CTRL, the sea surface salinity in the North Atlantic is considerably reduced (Fig. 6a). Due to the exchange of surface water close to the CAS, the surface water in the tropical Pacific becomes more saline. In contrast, the stronger ocean circulation for TVEG as compared to T GEO and the increased net evaporation yield considerably higher sea surface salinities in the North Atlantic Ocean (Fig. 6b). The strong increase in North Atlantic upper 500 m salinity is clearly emphasized in the Atlantic zonal-mean salinity distribution (Fig. 7a). The surface and subsurface warming of TVEG relative to T GEO is strongest in the subtropics and polar latitudes (Fig. 7b). At northern polar latitudes, the warming is associated to strong poleward surface currents (Fig. 5b), sea ice retreat, and meridional heat transport.

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## 4 Discussion

The vegetation effect on the ocean circulation may be an important mechanism for the relatively warm late Miocene climate over Europe as reconstructed by terrestrial proxy data (e.g., Mai, 1995; Wolfe, 1994; Fortelius et al., 2003). Sensitivity experiments with atmospheric general circulation models demonstrate that the late Miocene vegetation contributes to a warming of particularly the high latitudes (Dutton and Barron, 1997; Micheels, 2003; Francois et al., 2006; Micheels et al., 2006).

Here, we analyse the hydrological cycle and associated vegetation cover onto the ocean circulation. The atmospheric hydrological cycle has a high mobility and links the THC with the Earth's water budget. We find that the Intertropical Convergence Zone moves southward resulting in enhanced water vapor export out of the Atlantic catchment area. A similar effect has been proposed for tropical water vapor transport during glacials (Lohmann and Lorenz, 2000) and Heinrich events (Lohmann, 2003) which may be responsible for an additional sea surface salinity contrast between the Atlantic and Pacific/Indian Oceans (Broecker, 1992), as well as for El Niño conditions (Schmittner et al., 2000; Soden, 2000; Latif et al., 2000). We find that the water vapor transport out of the Atlantic area is enhanced for the Tortonian climate relative to the control experiment. The net Atlantic freshwater forcing has been recognized as an important external parameter of ocean sensitivity studies (e.g., Birchfield, 1989; Zaucker et al., 1994; Rahmstorf, 1996). The increased export at Central America is caused by an increase in the zonal moisture transport associated to the Atlantic trade winds.

As pointed out by Steppuhn et al. (2006), there is a significant warming of more than 2°C at the eastern margin of the Pacific Ocean associated with a decreased upwelling in this area. This is again linked to the southward shift of the thermal equator, the ITCZ and weaker equatorial Walker circulation. The latter gives rise to a Tortonian permanent El-Niño state. This aspect will be analyzed in a subsequent study using a coupled atmosphere-ocean circulation model for the late Miocene. Fedorov et al. (2006) proposed that a permanent El-Niño state may be important for Pliocene glaciation and

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Cenozoic climate evolution.

We find that the vegetation effect on the ocean circulation can be an important mechanism for the relatively warm late Miocene climate over Europe. Caused by high salinities at northern high latitudes, the sea ice edge is moved poleward which is in general agreement with proxy data (e.g., Wolf and Thiede, 1991). In addition, it is possible that other mechanisms not included in the present generation of GCMs also had an important impact on Tortonian climate, such as high-latitude radiative warming by polar stratospheric clouds (Sloan and Pollard, 1998), increased ocean heat transport driven by tropical cyclone-induced mixing (Emanuel, 2002; Huber, personal communication), or increased levels of methane. Methane can be estimated through stable carbon isotopes (biological processes preferentially incorporate C-12) and areas of wetlands as calculated from the land surface scheme including the vegetation distribution.

## 5 Conclusions

The Cenozoic climate evolution includes significant changes in the oceanic transports which are ultimately linked to the paleotopography and opening/closing of passages. The open Central American Seaway leads to an exchange of fresh Pacific water with saline Atlantic water thereby reducing the density in the North Atlantic Ocean and weakening of the large-scale ocean circulation. For the Late Miocene, we find that the modified vegetation cover can compensate this gateway effect by changes in the subtropical wind system and by more net-evaporation in the Atlantic Ocean. This increases North Atlantic salinity, ocean circulation and poleward heat transport to the north. Due to a “greener” Tortonian land surface and associated atmospheric and oceanic circulation changes, the Tortonian Atlantic meridional overturning and heat transport have almost their present strengths.

The failure of AGCMs to simulate the reduced pole-to-equator gradients of warm climate intervals is a long-standing problem in paleoclimate modeling (e.g., Sloan et al., 1995). The interaction between land surface cover, atmospheric as well as oceanic

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circulation could be the so far unknown mechanism for increasing ocean heat transport at a time when meridional surface temperature and vertical temperature gradients were greatly reduced relative to the modern (Bice et al., 2000).

Based on our finding, it is conceivable that reorganizations of the global ocean circulation, large-scale shifts of vegetation zones, topographical changes and changes in the global carbon chemistry play a dominant role for the major Cenozoic climate transitions. Consequently, it is of utmost importance not only to understand the behaviour of these individual systems in better detail but also to investigate the full dynamics, feedbacks, and synergisms of the coupled system. The results presented show a possible strong connection between the hydrological cycle, vegetation cover, and the ocean circulation. Future work will address the numerous interactions between the climate system components by use of a global atmosphere-ocean-vegetation-carbon cycle model. Dutton and Barron (1997) applied a palaeo-vegetation in a modelling study of the Miocene which led to a significant warming suggesting that vegetation and vegetation-climate feedbacks could be a significant component of the Cenozoic climate evolution. Feedback analysis including synergisms shall be performed to consider the dynamics of the climate system in a similar way as for the Quaternary climate variations (e.g., Ganopolski et al., 1998; Kubatzki et al., 2000). A focus can be on the relative roles of the thermohaline circulation, the atmospheric dynamics including high latitude and monsoon circulation, as well as land surface effects caused by changed vegetation distribution.

In order to further investigate major developments during the Miocene a combined approach between modeling and establishing proxy records from selected key locations is needed. Model results on changing patterns of heat transport can be validated by temperature reconstructions (Mg/Ca, alkenones, TEX86), both from the deep (benthic fauna) and the shallow (planktonic) ocean (Lear et al., 2003; Billups et al., 2002; Sluijs et al., 2006). Major changes in ocean circulation can be traced by using water mass characteristic proxies like Cd/Ca, Nd isotopes, and C-13 (Frank et al., 1999; Frank et al., 2002; Delaney and Boyle, 1987). Combination of temperature reconstruc-

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tions with O-18 gives evidence on changes in salinity and may provide indications on the high salinities in the northern North Atlantic, the position of the ITCZ and associated teleconnections.

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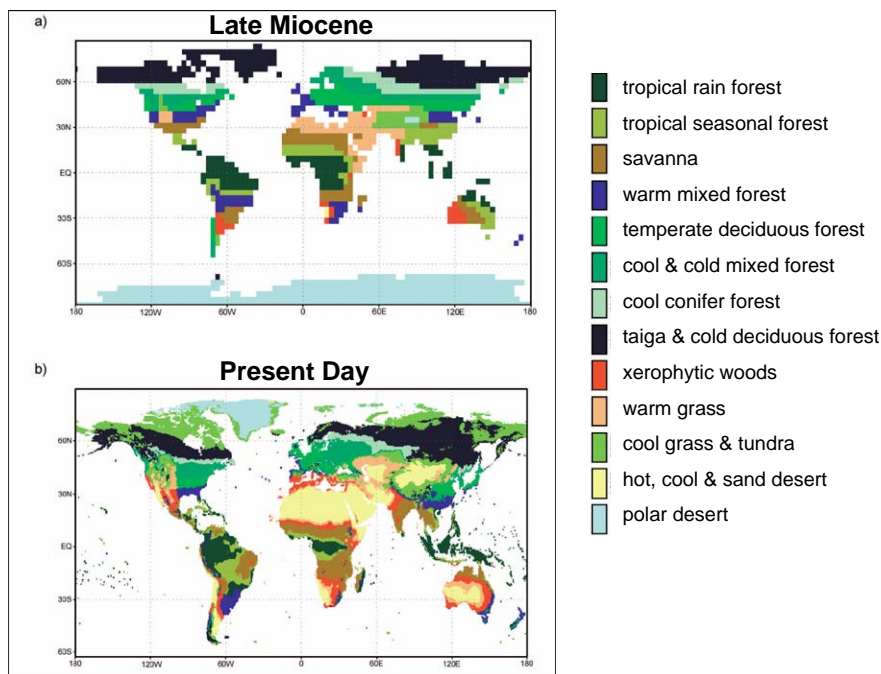
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**Fig. 1.** (a) The proxy-based reconstructed Tortonian vegetation, and (b) the present-day's vegetation (New et al., 1999). These maps serve as an input into the AGCM experiments TVEG, TGEO and CTRL, respectively.

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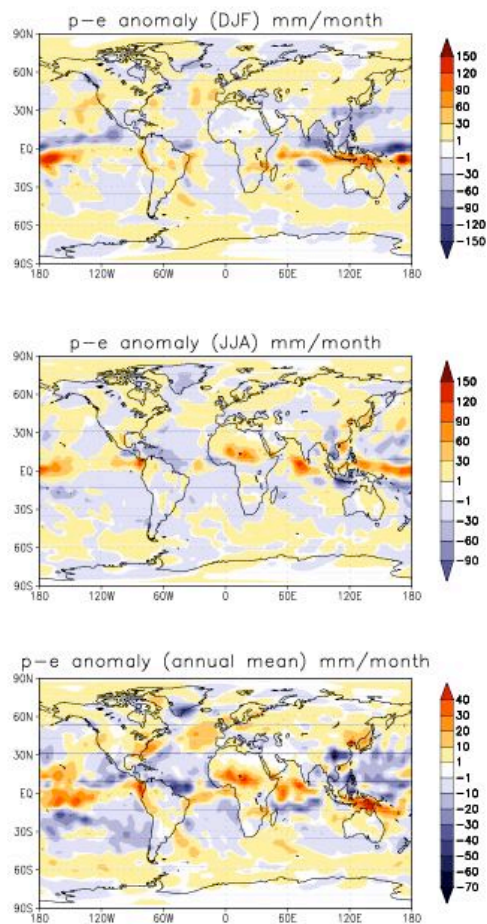
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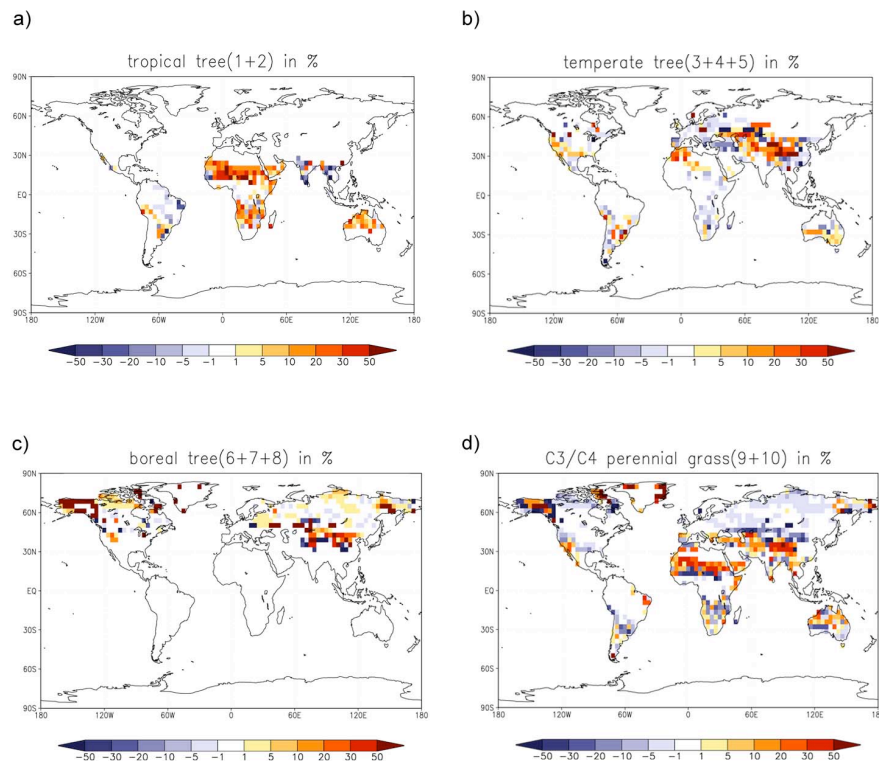


**Fig. 2.** Changes in the net precipitation minus evaporation for TVEG minus CTRL. Units are mm/month.

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**Fig. 3.** Change of the late Miocene relative to the present vegetation (TVEG minus CTRL). Fractional coverage (in percent): **(a)** Tropical tree, **(b)** temperate tree, **(c)** boreal tree, **(d)** grass.

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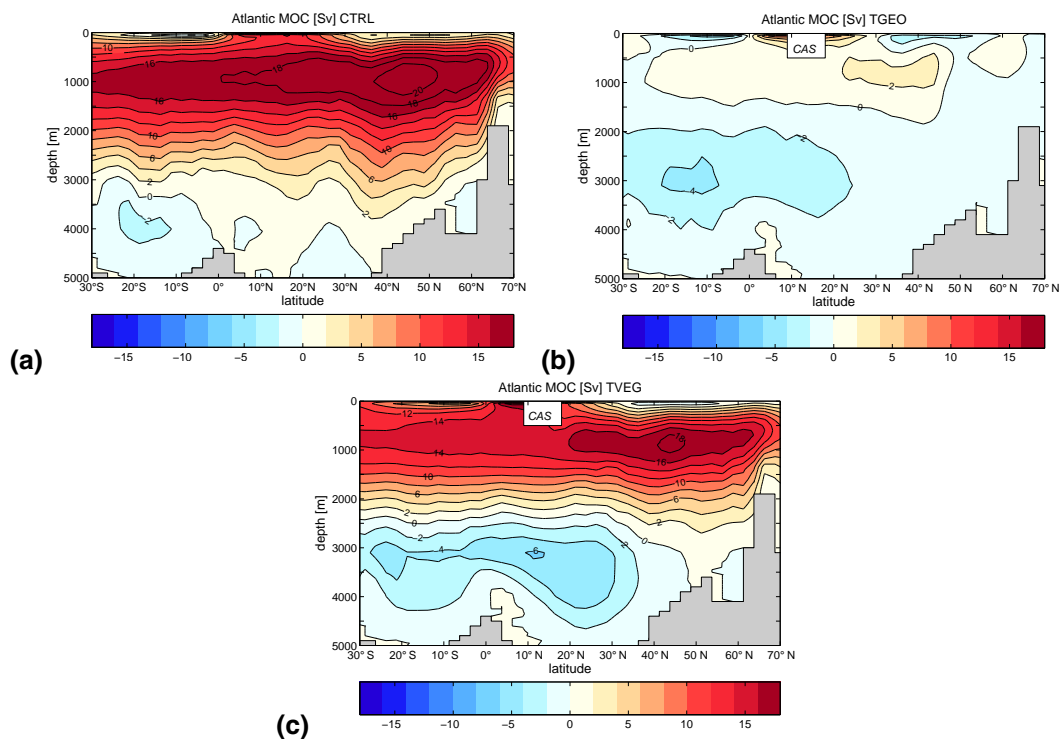
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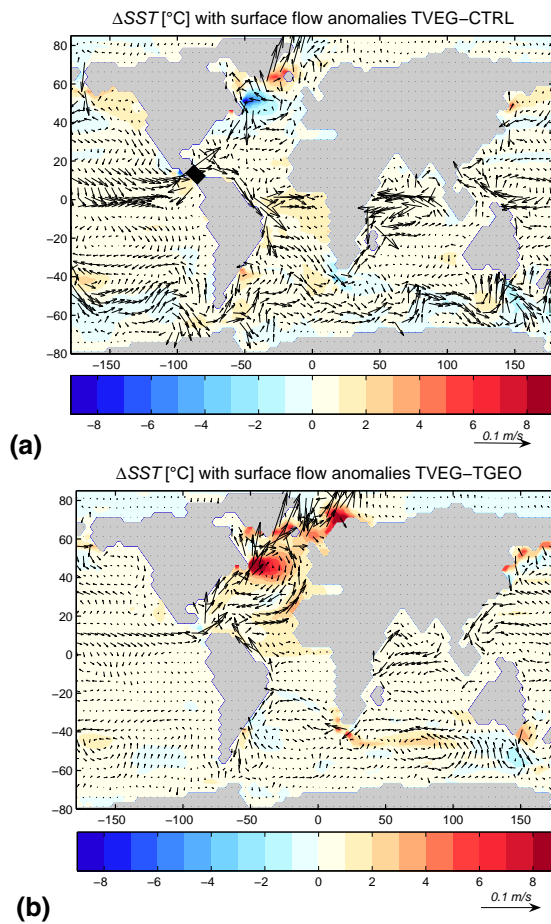


**Fig. 4.** Atlantic meridional overturning circulation ( $Sv=10^6 \text{ m}^3/\text{s}$ ) for present-day **(a)**, and the late Miocene configuration with open Central American Seaway (CAS). **(b)** with present vegetation cover (T GEO), and **(c)** with reconstructed vegetation cover (T VEG).

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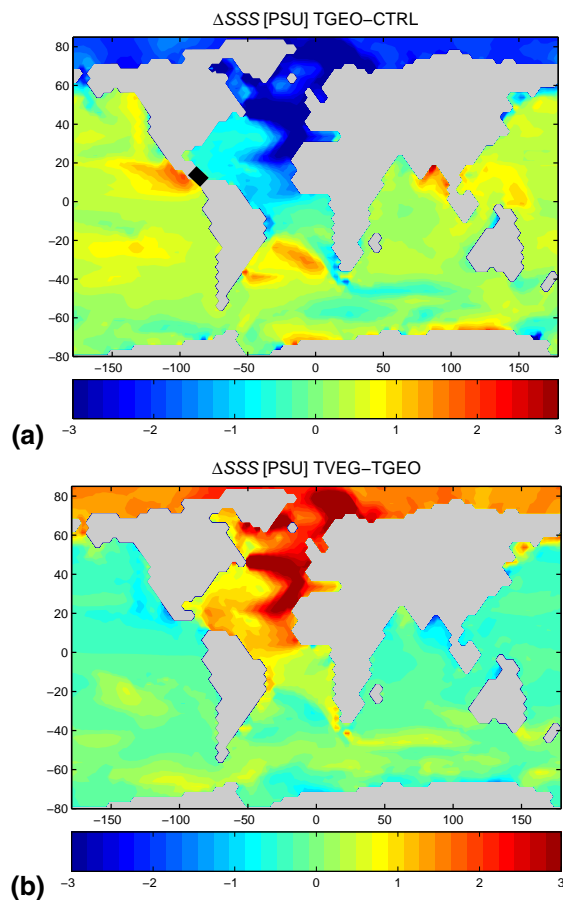


**Fig. 5.** Modelled sea surface temperature anomalies [°C] and surface flow [m/s]. **(a)** difference between TVEG and CTRL, **(b)** difference between TVEG and TGEO.

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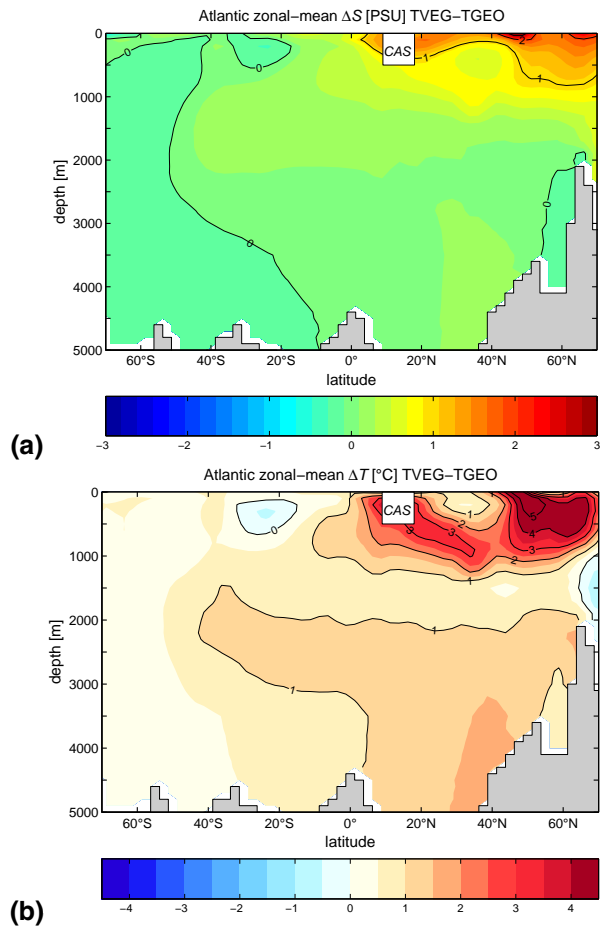


**Fig. 6.** Modelled sea surface salinity anomalies [PSU]. **(a)** difference between TGE0 and CTRL, **(b)** difference between TVEG and TGE0.

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**Fig. 7.** Zonal mean difference between TVEG and TGeo in the Atlantic Ocean: **(a)** salinity, **(b)** temperature.

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