Climate-Biosphere Feedbacks

Victor Brovkin

Potsdam Institute for Climate Impact Research (PIK)
P.O.Box 601203
14412 Potsdam
Germany
Tel.: +49 331 288 2592
Fax: +49 331 288 2570
e-mail: victor@pik-potsdam.de

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Synopsis

Geographical distribution of dominant land plant forms is mainly controlled by climate. If climate alters, vegetation cover closely follows the change in the climatic patterns. In turn, changes in spatial distribution and composition of terrestrial vegetation (as well as marine biology) alter the climate through modifications of heat and water fluxes, atmospheric gas and aerosol composition. Biogeochemical and biophysical mechanisms form numerous feedbacks between biosphere and climate that were active during the Earth's geological past and continue to operate at present. Analysis of climate-biosphere feedbacks and their role in the climate system dynamics during Quaternary is a new, rapidly developing scientific field that relies on climate system modeling and proxy data analysis. A picture emerging from paleo-reconstructions and modeling is that forest cover was greatly reduced at the maxima of Pleistocene glaciations. These dramatic changes in forest cover had several important implications for climate dynamics. Coldness and aridity of glacial climates were amplified by land cover changes through increased land surface albedo, decreased transpiration, and enhanced dust formation on exposed bare ground. At the same time, a role of biological mechanisms in glacial-interglacial CO_2 changes is not yet clarified.

Life is not an external or accidental phenomenon of the Earth's crust. It is closely bound to the structure of the crust, forms part of its mechanism, and fulfils functions of prime importance to the existence of this mechanism. Without life, the crustal mechanism would not exist.

Vladimir Vernadsky, The Biosphere, 1926.

Introduction

A current paradigm in plant ecology is that geographical distribution of dominant land plant forms is mainly controlled by climate (Woodward, 1987). If climate alters, vegetation cover closely follows the change in the climatic patterns. In turn, changes in spatial distribution and composition of terrestrial vegetation (as well as marine biology) alter the climate through modifications of heat and water fluxes, atmospheric gas and aerosol composition. Biogeochemical and biophysical mechanisms form numerous feedbacks between biosphere and climate that were active during the Earth's geological past and continue to operate at present.

Quaternary changes in terrestrial and marine biology are recorded in sediments around the planet. A picture emerging from pollen reconstructions is that forest cover was greatly reduced at the maxima of Pleistocene glaciations, while during warmest phases of interglacials expanded forests covered almost a half of the ice-free land surface. These dramatic changes in forest cover had several important implications for climate dynamics. Coldness and aridity of glacial climates were amplified by land cover changes through increase in land surface albedo, decrease in transpiration, and enhanced dust formation on exposed bare ground.

Less forests and lower productivity during glacials should have led to reduced terrestrial biomass and soil carbon storages and consequent release of CO_2 to the atmosphere that slightly offset the radiative cooling. This stabilizing effect is invisible in ice core CO_2 records due to counteracting amplifying feedbacks between temperature and greenhouse gases that led to lowering of atmospheric CO_2 level during glacials. One of hypothesized biological mechanisms that get some support from proxy data is an iron fertilization of glacial marine productivity due to increased level of dust depositions. A role of slow biogeochemical processes like coral reefs growth and terrestrial weathering in the atmospheric CO_2 dynamics is not yet well constrained.

Main Mechanisms of Climate-Biosphere Interaction

Photosynthesis

Process of photosynthesis that exists on Earth for more then three billion years is a focus point of climate-biosphere interaction. Terrestrial plants capture CO_2 from the atmosphere while marine phytoplankton takes up carbon from seawater, which exchanges CO_2 with the atmosphere (House et al., 2005). Respiration of living organisms returns carbon to the atmosphere. In the ocean, a part of dead organic matter (so-called export production) escapes the ocean surface layer and sinks towards the ocean floor. Because of long turnover time of deep ocean waters, amount of carbon stored in the ocean due to this process (called biological pump) is comparable to the organic carbon storage on land. Changes in terrestrial biomass and soil organic matter, as well as in volume and composition of marine export production affect atmospheric CO_2 concentration directly on annual to millennial time scale. Altered CO_2 concentration modifies atmospheric radiative properties and, via climatic change, feeds back on ecosystem production.

Land surface biophysics

In the lowest atmosphere, or planetary boundary layer (about 1-2.5 km above the Earth surface), vertical profiles of temperature and humidity strongly depend on the partitioning of energy between sensible heat and latent heat. This partitioning is, to a large extent, controlled by terrestrial ecosystems through biophysical mechanisms (Pitman et al., 2004). Over bare ground, the energy is transported via sensible heat, resulting in relatively high surface air temperatures. Transpiration, the flux of water from the ground to the atmosphere through plants is controlled by the opening and closing of tiny pores in the leaf's surface called stomata. Transpiration is tightly coupled to photosynthesis. Low atmospheric CO_2 levels during glacials led to decreased stomatal conductance, lower water use efficiency (productivity per water transpired), and reduced photosynthesis levels.

Vegetation canopies transpire water extracted from the root zone, increase latent heat flux upwards, and cool the surface air during the day. At night, increased air humidity due to daily transpiration reduces heat loss in the near-surface atmosphere (greenhouse effect) and smooth diurnal fluctuations in temperature. Vegetation canopies also intercept a part of rainfall and effectively reduce runoff, especially during heavy rain events. Moisture that is transported into continental interior has been transpired and precipitated several times on its way from the ocean, therefore reduction in vegetation cover during glacials enhanced the intra-continental aridity.

Vegetation affects surface albedo, the fraction of solar radiation reflected back into the atmosphere. In most cases, vegetation reduces albedo compared to bare ground and snow. Forests are more effective in trapping the solar radiation than grasslands. This effect is particularly strong in snow-covered regions where trees extend above the snow, while herbaceous vegetation is covered (see Figure 1). Snow-masking effect of tree cover was likely more important in glacial climates then snow cover lasted longer in spring.

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Long-term carbon cycle processes

Weathering

Starting from late Devonian, development of high vascular plants with deep roots led to formation of soil layer which stores water and nutrients. The soil buildup has contributed to enhancement of carbonate and silicate weathering which, on geological time scale, is leading to a slow decrease in atmospheric CO_2 concentration.

Carbonate sedimentation

Marine calcifying organisms like corals or coccolithophorids have skeletons or shells of calcium carbonate. After organism death, $CaCO_3$ does not dissolve but forms reefs or rains to sea floor and forms sediments. In the deep sea, carbonates are well preserved above lysocline (about 3-4 km depth) while they get dissolved below it (see Chapter 302: Dissolution of Deep-Sea Carbonates). On a time scale of about 5-10 thousand years, carbonate sedimentation in the ocean is constrained by weathering. Since carbonate deposition in shelf regions (e.g. coral reef production) is changing with sea level during glacial-interglacial cycles, this affects deep sea sedimentation and leads to changes in the lysocline depth. A temporal imbalance between weathering and sedimentation causes changes in total inventory of ocean alkalinity. Acidic waters dissolve less CO_2 than basic seawater, and in this way changes in $CaCO_3$ sedimentation affects atmospheric CO_2. This process is complicated by the fact that export flux contains not only carbonate but also organic matter that gets mineralized at the seafloor. The later process releases CO_2 that makes sediment pore water more acidic, leading to carbonate dissolution. Changes in organic to carbonate ratio (rain ratio) in export flux considerably affects $CaCO_3$ sedimentation and, through lysocline dynamics, the atmospheric CO_2 level.

Anaerobic decomposition

Some dead organic matter is trapped in wetlands or marine sediments in anaerobic conditions. Decomposition of this organics by methane producing bacteria may leads to a release of methane. Most of methane gets oxidized in a water column, but a part escapes to the atmosphere. Atmospheric CH_4 concentration is small but important for climate since methane is 20-30 times more potent greenhouse gas than carbon dioxide. CH_4 sink in the atmosphere is regulated by atmospheric chemistry that in turn is affected by plant emissions of volatile organic compounds (VOCs) like isoprene.

Numerical modeling as a methodology of feedback analysis

A climatic effect of planting a single tree is negligibly small. Even for large-scale land cover changes, their climatic consequences are very difficult to detect in observations because several climatic forcings are operating simultaneously in presence of substantial climate variability. Besides, the regional climate can be influenced not only by local changes but also by changes in remote regions. Carrying out numerical experiments with climate and ecosystem models is a main methodology for testing feedbacks between climate and biosphere. In usual design of sensitivity experiments, the simulations are performed twice, once when feedback is neglected and second time then feedback is included. A difference between these two simulations indicates feedback importance for climate. Biophysical feedbacks studied with global climate models are called hereafter biogeophysical feedbacks to stress their significance for global scale processes.

Hot Spots of Climate-Biosphere Interaction in Quaternary

The Holocene

"Green" Sahara

Sahara is a desert at present, but it was not always desert in the past. Multiple proxy data reveal that during the so-called Holocene optimum, ca. 9,000 – 6,000 years BP (before present), the vegetation cover of the Sahel was greatly extended to the north. At that period, the Sahara's climate was much wetter; many lakes and rivers were present in the region including the greatly extended Lake Chad (Prentice et al., 2000; also see chapter 208: Postglacial Pollen Records of Africa). A conventional explanation of the "green" Sahara phenomenon in the mid-Holocene, supported by many modeling experiments, is based on the long-term changes in the Earth's orbital parameters (see Chapter 35: Holocene GCMs). In the early to middle Holocene, the northern hemisphere received considerably more solar irradiation during the summer. In the northern subtropical regions, this led to stronger warming over the continent than over the ocean, an increased temperature gradient between land and ocean, and, consequently, intensified monsoon-type circulation in summer which led to increased rainfall over the Sahel/Sahara region.

The Sahara desert differs from many other subtropical regions. The radiative balance in the region is negative, in other words, there is a net radiative heat loss over the desert. This loss is particularly due to high albedo of the desert, as up to 40% of incoming radiation is reflected back into space (Figure 1). The radiative mechanism for the Sahara was first explored by Jule Charney in 1970s. He pointed out that the heat loss due to high albedo leads to a horizontal temperature gradient and induces an atmospheric circulation which maintains the sinking motion of dry atmospheric masses and suppresses rainfall over the region. Low precipitation results in little vegetation cover, and the surface albedo is determined by bare ground with a high albedo. This positive feedback supports a desert that is self-sustaining. On the other hand, if there is more precipitation, there is more vegetation. Vegetation canopy is darker than sand so the albedo is lower, the surface temperature is higher, and the gradient in temperature between land and ocean increases, amplifying monsoon circulation and upward motion over the desert. As a result, the summer rainfall in the region increases. This radiative feedback loop gets additional support through hydrological feedback as vegetation recycles water to the atmosphere, reduces runoff, and enhances rainfall in comparison with bare ground (Figure 2).

< Figure 2 near here >

Both radiative and hydrological feedbacks are positive, i.e. they amplify changes induced by external forcing or internal variability. Could these positive feedbacks affect the stability of the regional climate in the past? In the early 1990s, Martin Claussen applied coupled atmospheric-vegetation model ECHAM2-BIOME to investigate an effect of different initial land surface conditions on the regional climate. In these simulations, vegetation cover responded to climate simulated by the model and vice versa. In the first simulation with present-day boundary conditions (Figure 3a), which started from the present-day potential vegetation, the desert in the Sahara remained stable. However, when the Sahara was initially covered by forest in the simulation B (Figure 3b), the system converged to another solution, called the “green” Sahara. The western Sahara is covered by a mixture of shrubs and grass; in addition, vegetation cover in the Sahel region is enhanced. Both desert and green solutions are stable for present-day conditions. At the same time, the coupled model revealed the only steady solution, “green” Sahara, for the mid-Holocene orbital forcing (Figure 3c). In the dynamical system theory, a fact that the system has one steady state for one parameter value (mid-Holocene insolation) while two states for another value (present-day orbital parameters) means that in between these points the system undergone bifurcation that could be associated with abrupt climate changes. Indeed, proxy data suggest abrupt changes vegetation cover in west Sahara between 5,000 and 6,000 yr BP (deMenocal et al., 2000). Later simulations with another coupled model that included interactive ocean, ECBILT-CLIO-VECODE, supported the fact that the system underwent bifurcation around 4-6,000 yr BP as green state lost its stability (Renssen et al., 2003). Some other coupled models do not show multiple states, but they all reveal a high sensitivity of rainfall in Sahel/Sahara region to vegetation cover. In summary, strong positive vegetation-atmosphere feedback in the Northern Africa is important amplifier of externally induced changes in the region in the Quaternary. In particular, this feedback had contributed to abrupt desertification of the region about 5,000-6,000 years ago.

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Northern treeline dynamics

At the mid-Holocene, northern tundra-forest boundary was north of its present position in many regions. In particular, in central Siberia the treeline (the forest limit) was shifted to the north by up to 200 km (Bigelow et al., 2003), presumably due to warmer summers, as treeline position corresponds reasonably well with July temperature isotherms of about 10 to 12°C. In the northern high latitudes, changes in forest cover affect regional climate in several ways. Forests, even deciduous ones, significantly reduce the albedo of snow-covered surfaces. This is the basis for the radiative feedback between forest and surface air temperature. Increased forest cover within large region leads to decreased surface albedo during snow season. This results in elevated air temperature and earlier snow melt, longer and warmer growing season, which favors further increase in forest area. This feedback (sometimes called taiga-tundra feedback) amplifies the climate system response to the original external forcing (Figure 4). How fast forest cover changes, depends on many local factors like disturbances (fire), permafrost, soil parent material, and pH. The radiative feedback is dampened during the growing season, when trees have a denser, more productive canopy, as well as deeper roots than herbaceous plants and moss and transpire more water. This cools the surface air in forests as compared to tundra during summer. The hydrological feedback, although negative, is secondary to the radiative feedback on the annual average. Besides, enhanced transpiration from deeper soil horizons improves soil drainage and aeration of soil mineral layers that increases viability of tree establishment. In Arctic environment, the radiative feedback of forests is tightly linked to sea ice - albedo feedback. Elevated surface air temperature due to increased forest cover leads to higher sea surface temperature (SSTs) and decreased sea ice area, decreased sea surface albedo, and, finally, elevated surface air temperature (Figure 4). Model simulation suggests that taiga-tundra feedback and its amplification by sea ice-albedo feedback led to additional warming of northern high latitudes in response to the changes in orbital forcing at the mid-Holocene (Figure 5).

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Historical land cover changes

At present, about one third of land cover has being modified by agricultural and forestry activities. Human-induced land cover changes began probably as early as the middle Holocene, ca. 8,000 years ago, with development of slash-and-burn agriculture. Magnitude of these changes and their temporal dynamics on large regional scales in pre-industrial period are mostly unknown. Most models have been used to evaluate a climatic effect of reconstructed potential vegetation versus present-day land cover, and some models have been applied for testing climatic effect of the last millennium land-cover changes using simplified assumptions about land cover changes prior to 1700 AD. Most model simulations agree that historical deforestation led to some increase in atmospheric CO_2 and consequent warming, but at the same time an increase in albedo due to deforestation of regions covered with snow in winter like northern Europe led to a radiative cooling (Figure 6). On global scale, biogeophysical cooling was likely compensated by biogeochemical warming, but consequences for regional climate could be substantial. Biogeophysical effect due to increasing deforestation rate could have contributed to an observed stabilization of the temperature during the second half of the 19th century, and has damped the CO_2 warming in the 20th century.

< Figure 6 near here >

Last Glacial Maximum

The last glacial maximum was characterized by cold and dry conditions (see Chapter 34: Last Glacial Maximum GCMs) that led to a great reduction of forest cover in the Northern hemisphere. Boreal evergreen forests and temperate deciduous forest were fragmented, while European and East Asian steppes were greatly extended (Prentice et al., 2000). Tropical forests in America and Africa were reduced as well. These changes towards herbaceous and sparser vegetation cover had considerable implications for climate.

Biogeophysical feedbacks

Model simulations using glacial boundary conditions suggest that reduction in forest cover has amplified LGM coldness and aridity. Southward displacement and fragmentation of forest biomes in North America and Eurasia led to increase in albedo during snow period and a regional decrease in temperature by several degrees (Figure 6). Reduction in forest cover led to less transpiration and increased aridity of the continent interiors. Presumably, physiological effect of low CO_2 level during LGM resulted in increased canopy conductance of C_3 plants and consequent drop in water use efficiency (water used per unit of photosynthetic production) (Harrison and Prentice, 2003).

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Terrestrial CO_2 feedback

Cold and dry climate as well as low CO_2 levels led to decreased plant productivity and biomass storage at LGM. Land areas in the Arctic were more extent due to low sea level stand but ecosystems there were not productive. To some degree, reduction in boreal forests was compensated by establishing of new forests at the exposed shelf regions in tropical regions like South Eastern Asia, but on global scale terrestrial carbon storage was drastically reduced. Vegetation models applied for LGM climate change scenarios reveal a range of decrease in terrestrial storages from several hundred to thousand PgC. A central estimate in about 600-700 PgC is higher than changes suggested by a decline in deep ocean $\delta^{13}C$. The later proxy should be carefully interpreted since oceanic $\delta^{13}C$ is also affected by changes in oceanic circulation and marine productivity. Decrease in northern peat storage due to expansion of ice sheets and increased aridity has also contributed to release of land carbon to the atmosphere but magnitude and timing of peat changes are uncertain. In total, a reduction of terrestrial carbon pool during glacial periods provides a negative (stabilizing) feedback to climate by increasing atmospheric CO_2 concentration. This feedback is rather weak since most of released terrestrial carbon ended up in the ocean. After carbonate compensation on time scale of 5 to 10 thousand years, 500 PgC of land carbon source corresponds to about 10 to 20 ppmv increase in atmospheric CO_2 concentration.

Dust feedback

In semi-arid regions, reduced vegetation cover means more bare ground exposed to wind erosion. During glacial periods, these sparsely vegetated regions were a source of mineral aerosols for the atmosphere. The higher aeolian dust transport during glacial periods is supported by sediment data (Kohfeld et al., 2005). The mineral aerosols diffuse light and reduce amount of solar radiation that reached the surface. Radiative transfer models estimate that increased atmospheric dust concentrations at the LGM led to about 2-3 W/m^2 reduction in radiative forcing in tropics which roughly corresponds to the radiative effect of lower glacial CO_2 level in the region (Claquin et al., 2003). As a result, sparser vegetation cover amplified LGM cooling in tropics through the dust feedback.

Marine biota feedbacks

Several hypotheses aimed at explanation of glacial-interglacial changes in atmospheric CO_2 invoke changes in marine ecosystem structure and export production. Proxy data suggest that export production in subantarctic Atlantic was enhanced at the LGM, presumably due to iron fertilization in response to increased aeolian dust deposition. Some other regions, like Pacific, show little changes, while productivity in Antarctic Ocean was reduced. Iron fertilization could be responsible for some drawdown of atmospheric CO_2 levels at the LGM, but it can not explain the whole range of 80-100 ppmv change (Kohfeld et al., 2005). Increased organic to $CaCO_3$ ratio in export production (for example, due to silica leakage hypothesis) could have substantially lowered the atmospheric CO_2 through the sediment response. However, recent data show that ratio of organic to $CaCO_3$ flux that reached seafloor is almost constant. Coral reefs production was reduced at the LGM in response to sea level drop and surface cooling (Kleypas, 1997). Reduced shallow-water sedimentation could have lead to an increase in oceanic total alkalinity, but this effect might be also offset by lower weathering at LGM. In summary, marine biosphere is likely to provide positive feedback to atmosphere by amplifying LGM cooling through lowering of the atmospheric CO_2, but magnitude of this feedback and precise mechanisms are still not well known.

Weathering feedback

On a time scale longer than several thousand years, atmospheric CO_2 is strongly affected by changes in carbonate and silicate weathering. Increased weathering lowers atmospheric CO_2 concentration through oceanic carbonate compensation. Vascular plants contribute to the weathering as roots respiration and litter mineralization lead to elevated soil CO_2 concentration and formation of carbonic acid that enhances weathering of parent soil material. Model estimates of changes in weathering rate at the LGM are equivocal as several factors influencing global weathering flux like increased ice sheet areas, shelf exposure, and decreased runoff affect almost compensate each other (Munhoven, 2002).

Methane and atmospheric chemistry feedback

At the LGM, atmospheric CH_4 concentration (about 400 ppbv) was much lower than in the Holocene (600-750 ppbv) (see: 338 Methane studies). Decreased concentration of methane that is about 20 times more potent greenhouse gas than carbon dioxide led to small but not negligible additional cooling at the LGM. The plausible explanation of the declined CH_4 concentration is that terrestrial CH_4 sources (wetlands, biomass burning) were lower at the LGM. In particular, wetland extent was reduced in the cooler and drier LGM climate, especially at high latitudes, while this reduction might have been offset by establishing of wetlands on exposed tropical shelfs. Another important factor could be a change in atmospheric chemistry. Model simulation suggest that emissions of volatile organic compounds were lower at LGM due to reduction in forest cover, resulting in increased concentration of OH radical (an atmospheric detergent), and this have led to an increased sink for CH_4 in the atmosphere (Valdes et al., 2005). In summary, biospheric feedbacks amplified LGM cooling through lowering of CH_4 concentration, while CH_4-regulating mechanisms involving biology are not yet fully explored.

The Last Interglacial and Glacial Inception

During the last interglacial, an increase in summer insolation in northern hemisphere was stronger than during the Holocene. Pollen evidence and model simulations suggest that summer climate was substantially warmer in northern regions then at present and that forest has expanded in the temperate and high northern latitudes (see Chapter 33: Modeling the last interglacial). The vegetation feedback amplifies the warming in model simulations (Figure 8a). This temperature anomaly is mainly due to the northward advance of boreal forest for the warmer Eemian climate. Conversely, the later cooling trend triggered by the decreasing boreal summer insolation is facilitated by the vegetation response (Figure 8b). Climate model simulations suggest that southward shift of northern treeline is crucial for initiation of glacial inception: if vegetation cover is prescribed to the values simulated for the last interglacial, then cooling in the north due to declining insolation is not sufficient to initiate ice sheet growth. Treeline retreat amplifies cooling through albedo feedback and allows simulation of permanent snow cover on land during the glacial inception.

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Conclusions

Analysis of climate-biosphere feedbacks and their role in the climate system dynamics during Quaternary is a new, rapidly developing scientific field. Hypotheses about roles of biological mechanisms in past climate dynamics usually emerge from analysis of geological data, but quantification of feedbacks is impossible without modeling. Today, the main methodology of feedback analysis is numerical modeling using climate system models of different complexity, from high-resolution GCMs to simplified conceptual models. Model results are tested against proxy data for biogeochemistry and vegetation cover.

A picture emerging from model simulations and proxy data analysis is that biospheric feedbacks played substantial role in evolution of Quaternary climate, but many details are uncertain. In some cases, like glacial inception, model simulations agree that dynamics of northern treeline is an important mechanism amplifying cooling due to summer insolation decrease at the northern high latitudes. Analysis of climate-vegetation interaction in Sahel-Sahara region suggests that biospheric feedbacks might be strong enough to cause abrupt, irreversible changes in the region. At the same time, a role of biological mechanisms in glacial-interglacial CO_2 changes is yet far from clear.

Understanding of role of biospheric feedbacks in the past has important implication for the future climate projections. The challenge in the Earth System science is to constrain feedback mechanisms within the models based on available - and newly developing – proxies for biological activities in the past.

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Figure captions

Figure 1. Map of Land Surface Albedo Captured by the MODIS Satellite Instrument (Schaaf et al. 2002; Lucht et al. 2000). Regions where there was no data available, e.g., due to clouds, are indicated by black. (a) January 2001. In the northern areas during the winter season, snow albedo is very high (up to 0.8, red). The boreal forest belt can be clearly seen in blue and green since trees mask snow, reduce albedo, and warm the surface air during the snow season. (b) June 2001. In comparison with January 2001, the northern land areas have a much lower albedo due to the absence of snow. In this map, the area with the highest albedo (up to 0.5, green and yellow) is the Sahara desert. High albedo in this region suppresses rainfall during the summer rain season. After House et al. (2005).

Figure 2. Idealized sketch of atmosphere-vegetation interactions in the Sahel/Sahara region. The signs attached to the arrows indicate the effect of the outgoing box on the ingoing box, e.g., increased vegetation cover lead to more radiation absorbed at land surface. Both radiative and hydrological feedbacks are positive, i.e. they amplify external changes and therefore may cause instabilities.

Figure 3. Summary of results obtained with an interactive atmosphere-vegetation ECHAM/BIOME model. For present-day climate, two states are stable in the model: desert Sahara as observed now (a), and “green Sahara” state with enhanced vegetation cover in West Sahara (b) in accordance with Claussen (1994). (c) For 6,000 yr ago, “green Sahara” is the only stable state in the model (Claussen and Gayler, 1997). Lower albedo corresponds to a higher fraction of vegetation cover. After Brovkin et al. (1998).

Figure 4. Role of vegetation feedbacks in the mid-Holocene climate in terms of annually averaged surface air temperature (°C) in CLIMBER-2 model simulations of the mid-Holocene climate (6000 yr BP). (a) Difference between the mid-Holocene and the present-day simulations. (b) Contribution of the biogeophysical feedbacks (difference between mid-Holocene simulations with interactive vegetation and vegetation prescribed from present-day simulation). After Ganopolski et al. (1997).

Figure 5. Simplified scheme of climate-vegetation interactions in the northern high latitudes. The signs attached to the arrows indicate the effect of the outgoing box on the ingoing box, e.g., increased forest (tree) cover leads to more radiation absorbed at land surface during snow-covered period. Changes in surface air temperature due to forest cover changes are amplified through sea ice – albedo feedback in the Arctic.

Figure 6. Effects of historical land cover changes on global mean annual temperature (in °C) simulated by CLIMBER-2 model (Brovkin et al., 2004). H_P - biogeophysical effect only (mainly due to albedo changes); H_C – biogeochemical effect only in response to atmospheric CO_2 increase; H – both biogeophysical and biogeochemical effects. Shown are differences with control simulation without land cover changes.

Figure 7. Role of biogeophysical feedbacks in the LGM climate as simulated by CLIMBER-2 model. Shown are annually averaged surface air temperatures differences (in °C). (a) A difference between the LGM and the present-day simulations. (b) Change in the temperature due to the biogeophysical feedbacks. After Jahn et al. (2005).

Figure 8. Role of biogeophysical feedbacks in the climate of last interglacial and glacial inception. Shown are annual temperature differences (°C) between CLIMBER-2 model transient experiments with interactive vegetation and vegetation cover prescribed from present-day simulation. Ice sheets are interactive and radiative effect of dust is accounted for. (a) Warming due to biospheric feedbacks at 125 kyr BP. (b) Cooling induced by vegetation changes at 115 kyr BP. After Calov et al. (2005). With kind permission of Springer Science and Business Media.

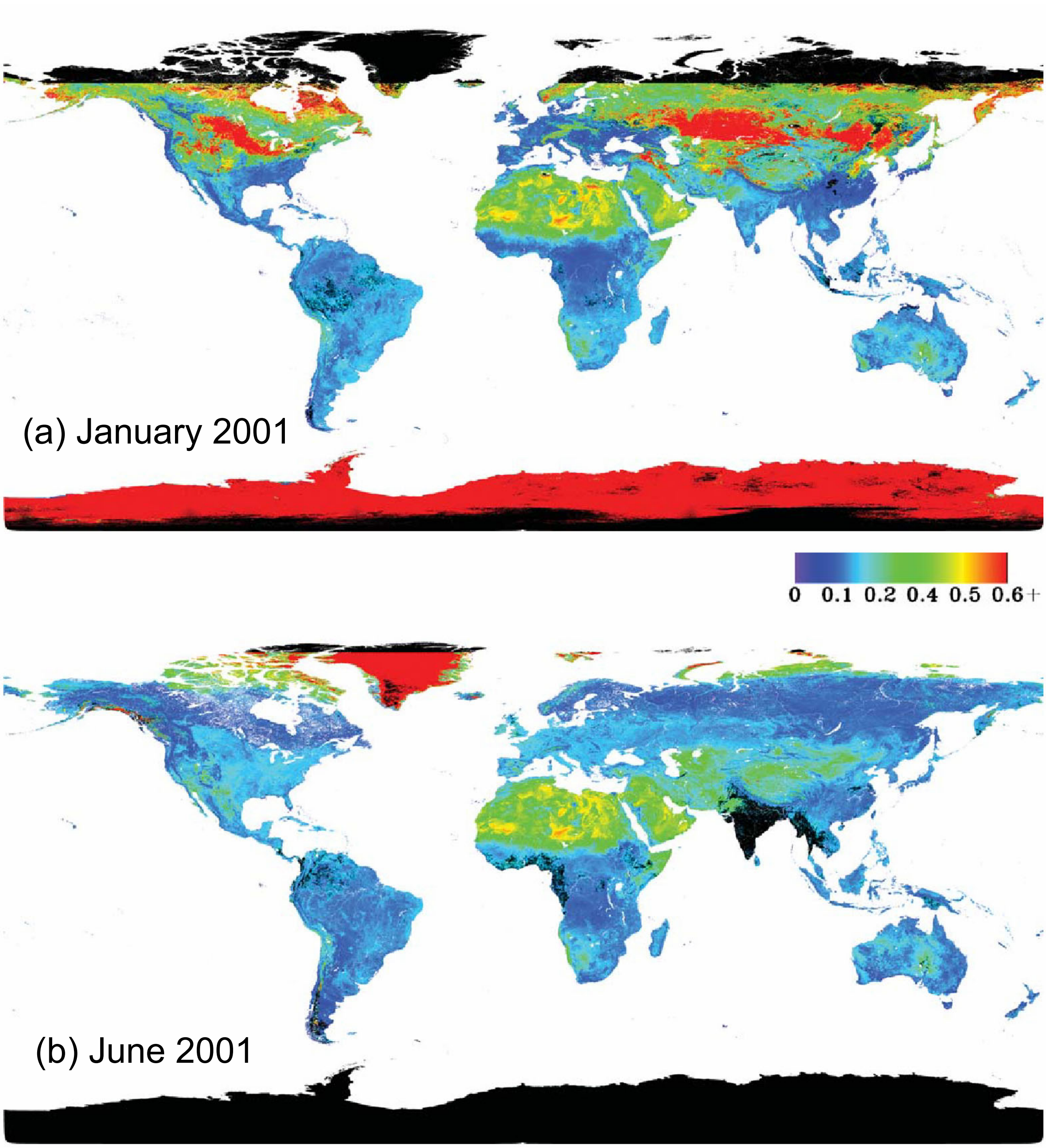


Figure 1

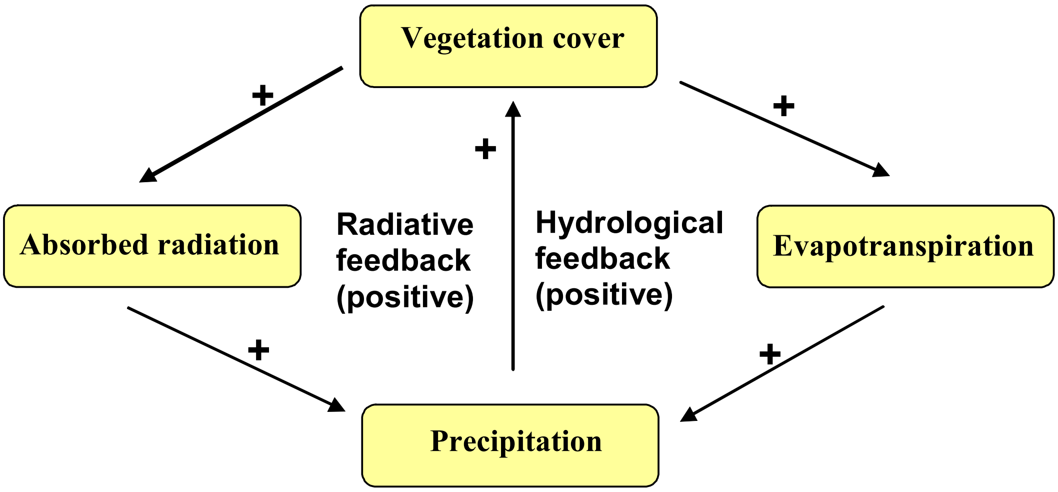

Vegetation cover
+
+
+
Absorbed radiation
Radiative feedback (positive)
Hydrological feedback (positive)
Evapotranspiration
+
+
Precipitation

Figure 2

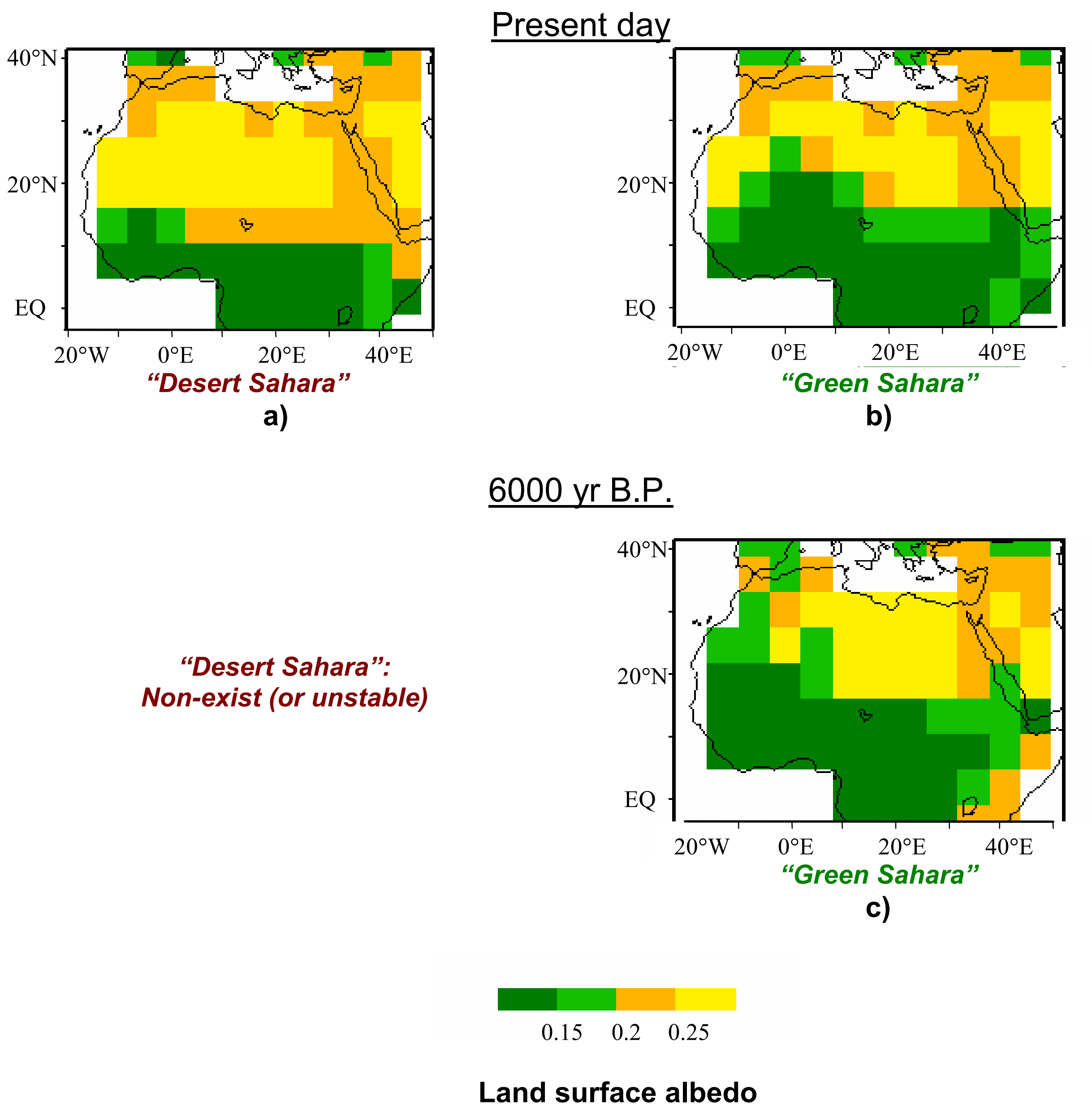

Present day
40°N
20°N
EQ
20°W 0°E 20°E 40°E
"Desert Sahara"
a)
"Green Sahara"
b)
6000 yr B.P.
"Desert Sahara":
Non-exist (or unstable)
"Green Sahara"
c)
0.15 0.2 0.25
Land surface albedo

Figure 3

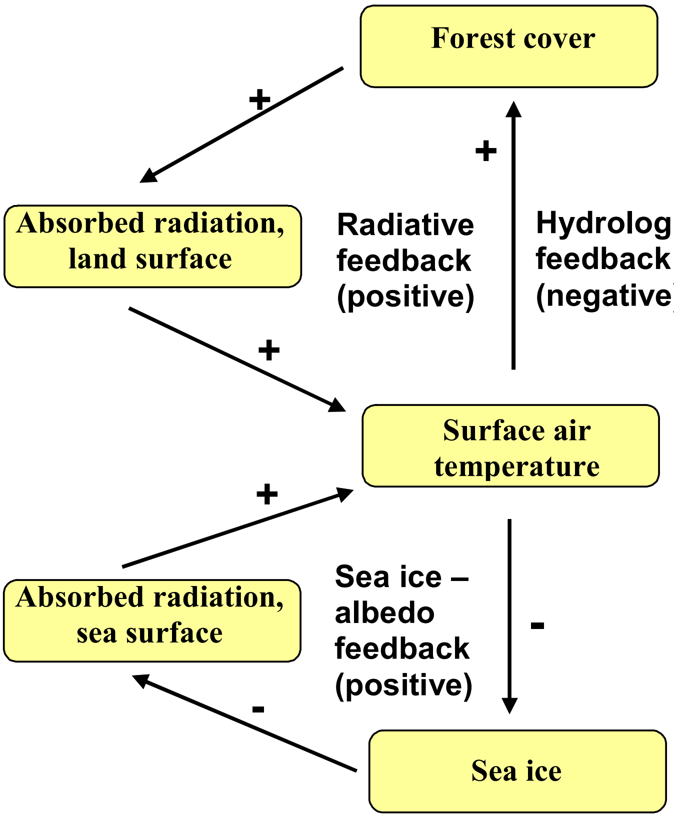

Forest cover
+
+
+
Absorbed radiation, land surface
Radiative feedback (positive)
Hydrological feedback (negative)
Evapotranspiration
+
-
Surface air temperature
+
Absorbed radiation, sea surface
Sea ice – albedo feedback (positive)
-
-
Sea ice

Figure 4

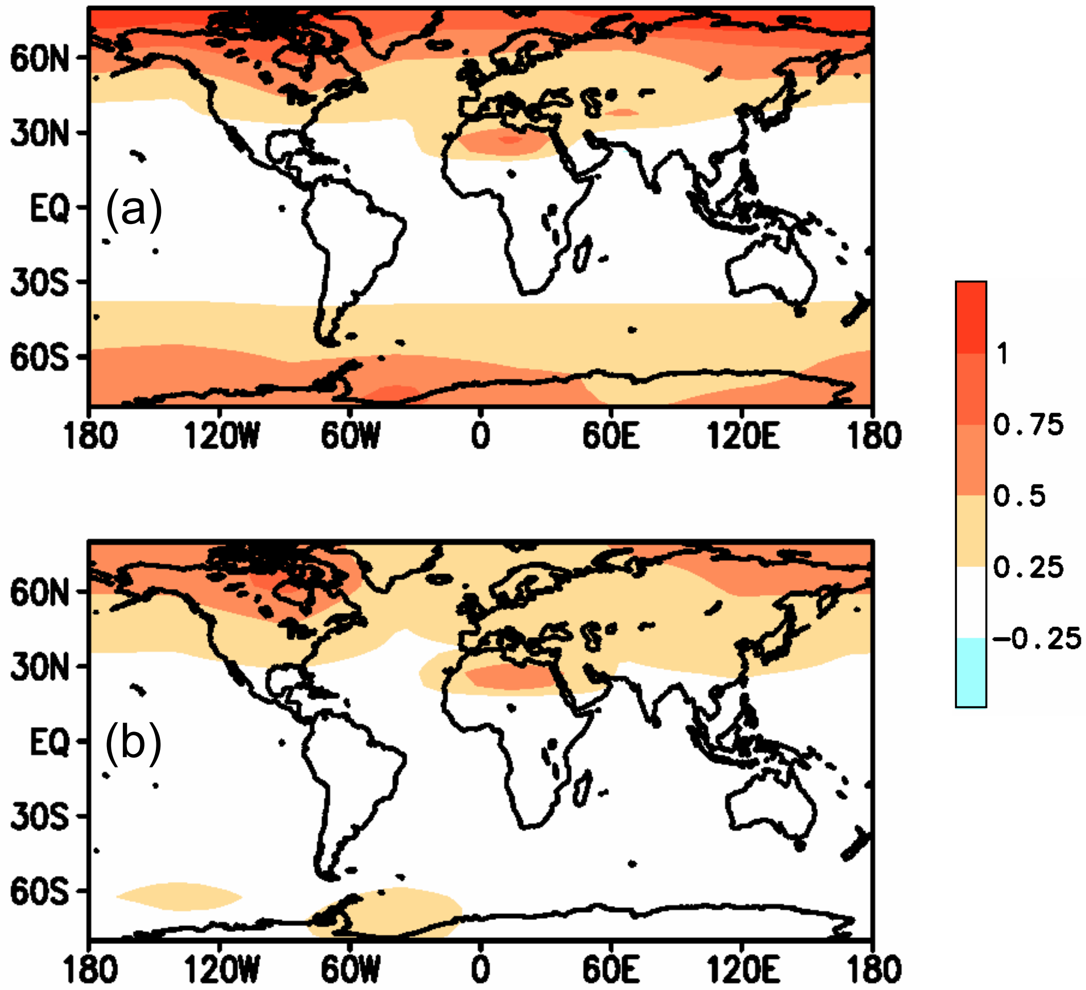
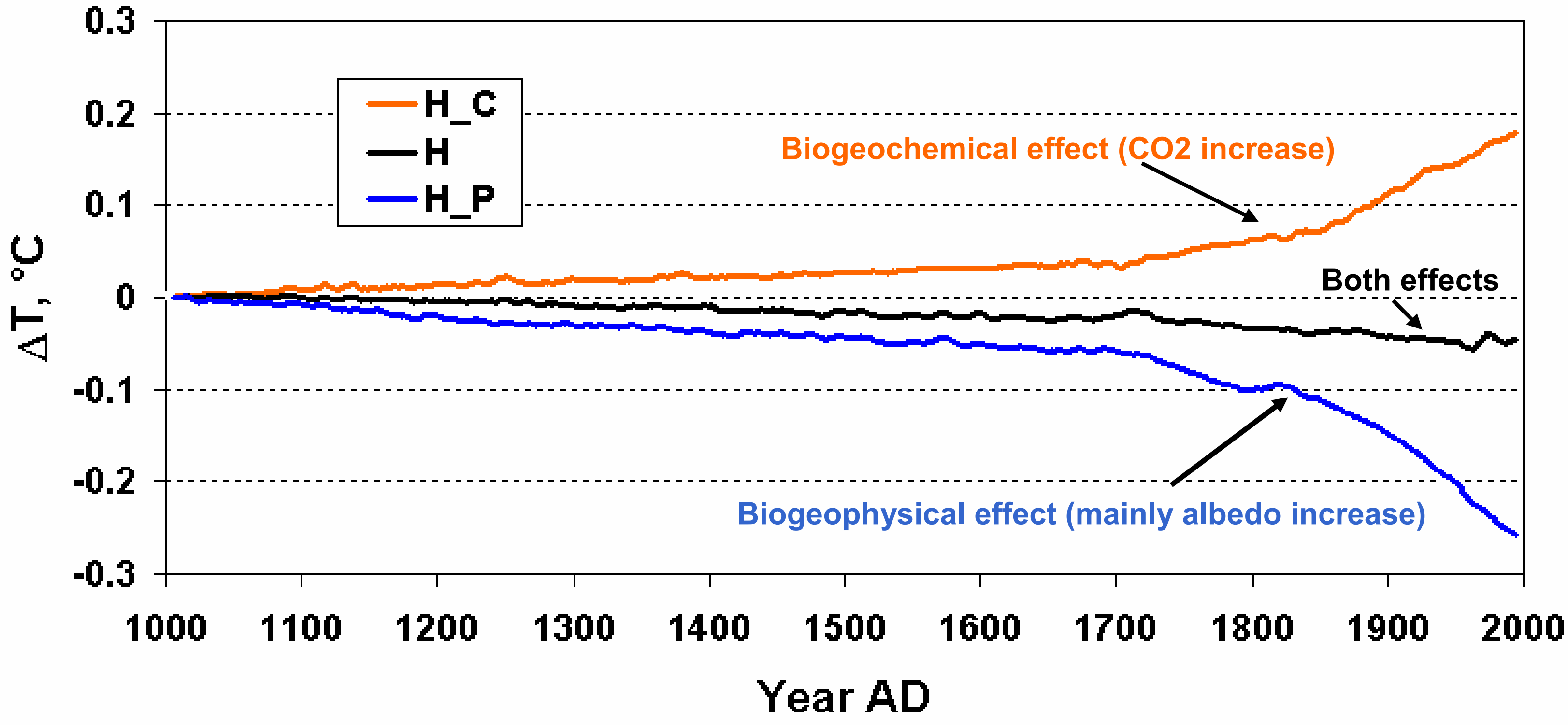

60N
30N
EQ
(a)
30S
60S
180
120W
60W
0
60E
120E
180
60N
30N
EQ
(b)
30S
60S
180
120W
60W
0
60E
120E
180
1
0.75
0.5
0.25
-0.25

Figure 5



0.3
0.2
0.1
0
-0.1
-0.2
-0.3
ΔT, °C
H_C
H
H_P
Biogeochemical effect (CO2 increase)
Both effects
Biogeophysical effect (mainly albedo increase)
1000
1100
1200
1300
1400
1500
1600
1700
1800
1900
2000
Year AD

Figure 6

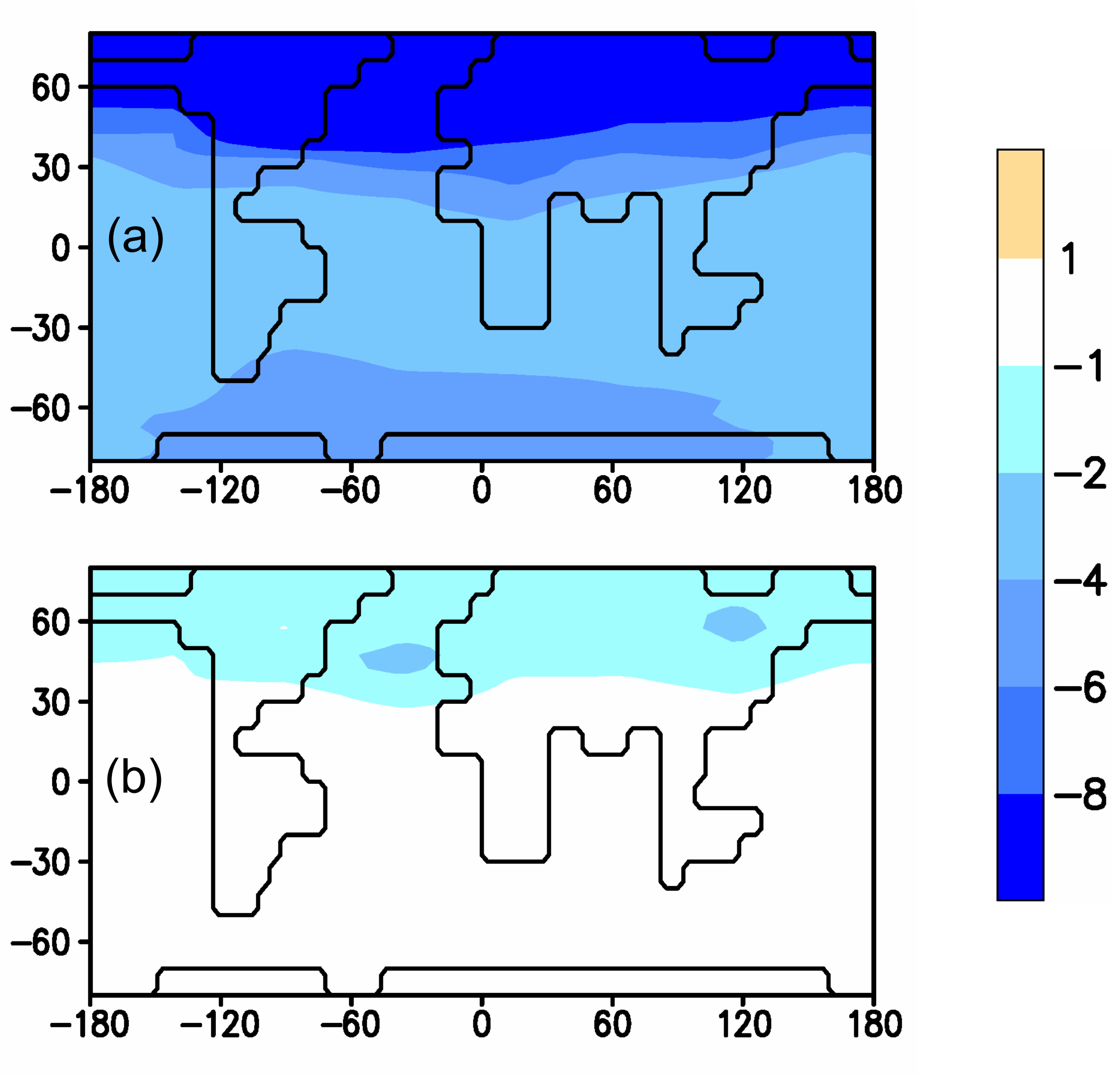
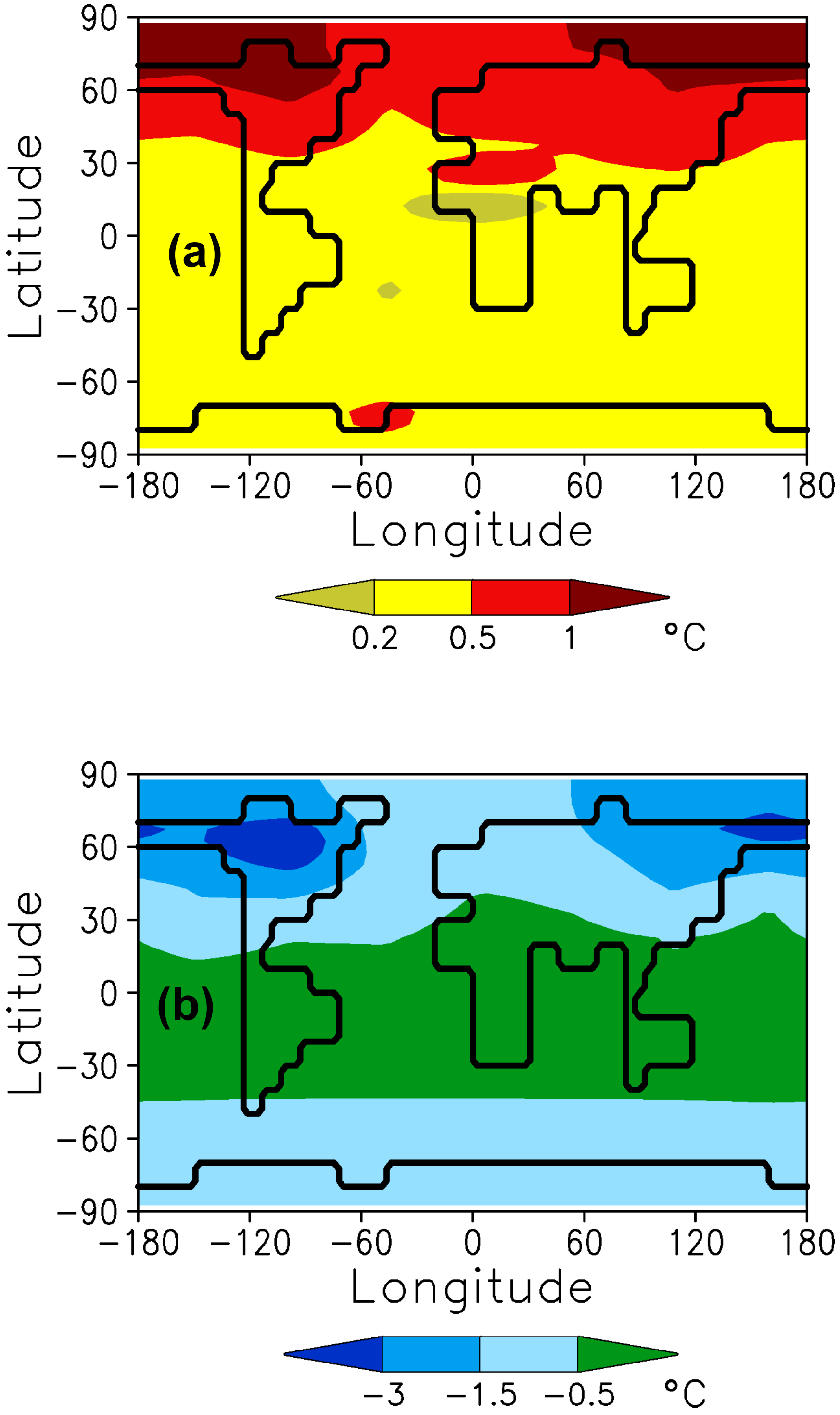


Figure 7



90
60
30
0
−30
−60
−90
Latitude
(a)
−180
−120
−60
0
60
120
180
Longitude
0.2
0.5
1
°C
90
60
30
0
−30
−60
−90
Latitude
(b)
−180
−120
−60
0
60
120
180
Longitude
−3
−1.5
−0.5
°C

Figure 8