CLIMATE-CARBON CYCLE INTERACTIONS ON MILLENNIAL TO GLACIAL TIMESCALES AS SIMULATED BY A MODEL OF INTERMEDIATE COMPLEXITY

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During glacial-interglacial cycles, the atmospheric CO$_2$ content varied by about 100 ppmv. Although atmospheric CO$_2$ variations played a central role in shaping glacial-interglacial cycles, their origin still remains elusive. A recently proposed hypothesis argues for a southern hemispheric wind control of orbital-scale CO$_2$ variability. This hypothesis is tested here using an earth system model of intermediate complexity, LOVECLIM. A series of sensitivity experiments demonstrates that a weakening of the southern hemispheric westerlies leads to reduced upwelling of DIC-rich waters in the Southern Ocean and a reduction of surface ocean pCO$_2$. However, at the same time less nutrients are upwelled in the Southern Ocean and transported to the eastern basin upwelling zones. This leads to a drop of marine export production and an increase in surface ocean pCO$_2$. The net result of these counteracting effects is a small drop of atmospheric CO$_2$ that is insufficient to explain the magnitude of glacial-interglacial CO$_2$ variability.

Another key feature of glacial periods is the occurrence of northern hemispheric meltwater pulses such as Heinrich event 1 (18 ka ago) and the Younger Dryas (11 ka ago). Originating from instabilities of the major ice-sheets, these freshwater pulses led to disruptions of the Atlantic Meridional Overturning Circulation (AMOC) and impacted climate worldwide. As demonstrated by numerous paleo-proxy records, such events were accompanied by a northern hemispheric cooling that lasted hundreds of years and by associated shifts of the Intertropical Convergence Zones. Using LOVECLIM, it is documented that these climate anomalies led to shifts of the major vegetation zones. Under pre-industrial background conditions, the simulated reduced terrestrial primary productivity leads to a carbon release of 120 GtC. A part of this carbon is taken up by the oceans through increased solubility and deep ocean storage. The resulting atmospheric CO$_2$ rise of 20 ppmv leads to a net warming of the Southern Ocean and Antarctica. Comparing paleo-climate model simulations with paleo-proxy data, it is found that at least 25% of the Antarctic warming during Heinrich events can be attributed to the vegetation response to an AMOC shutdown.
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Chapter 1

Introduction

1.1 Variability of the climate system

Climate can be defined as the longterm averaged probability distribution of the states of the atmosphere-ocean-cryosphere-land system. Climate is primarily driven by the energy the Earth receives from the Sun. Due to the angle of Earth’s rotation axis with respect to the ecliptic, tropical latitudes receive more solar energy than polar latitudes. The resulting meridional heating gradient is responsible for establishing the mean atmospheric and oceanic circulation. The atmospheric and oceanic circulation and their synoptic variabilities are key components in transporting moisture and heat to high-latitudes as well as greenhouse gases around the globe.

The cryosphere, which includes the Arctic and Southern Ocean sea ice as well as the land-based ice sheets of Greenland and Antarctica, is an integral part of the climate system through its influence on surface energy and freshwater fluxes. Due to its high albedo, the cryosphere reflects most of the Sun’s incoming radiation, which cools its surface and thus provides an important feedback to the climate system.

The atmosphere also exchanges heat and water with the land surface. Depending on its vegetation coverage, the absorption of solar energy by the land surface will vary, hence affecting the air temperature and hydrology. In addition, the topography of the land surface influences the flow of air.

Finally, Earth’s radiative budget depends on its atmospheric greenhouse gases and aerosol content. An increase in atmospheric greenhouse gases concentration raises Earth’s temperature, while it is the contrary for the direct effect of aerosols. The atmospheric content of one of the most important greenhouse gas, CO\textsubscript{2}, is modulated by the terrestrial and marine biospheres primarily through photosynthesis and respiration. Variations in the Earth’s biosphere can thus have a significant impact on the climate system. The lithosphere is sometimes also mentioned as part of the climate system as greenhouse gases and aerosols are released into the atmosphere by volcanic eruptions. In addition, the uneven interior heating of the Earth can influence atmospheric and oceanic circulations on long timescales.
Due to the variety of the processes involved, climate varies across a wide spectrum of timescales. This study focuses on millennial to glacial-interglacial climate variations (figure 1.1). As postulated by Milankovitch (1930), major external forcings of the climate system at these timescales are astronomically driven variations in insolation. Variations in ice sheet extent, oceanic circulation, and carbon cycle also provide important feedbacks onto the climate system at these timescales. Their possible influence on climate will be briefly discussed below.

### 1.1.1 Orbital parameters and their influence on the climate system

Variations in the Earth’s orbit, which determine the geographical distribution of received solar insolation, are the dominant external forcing of the climate on glacial-interglacial timescales (Milankovitch, 1930). The so-called Milankovitch cycles describe the combined effect of changes in the eccentricity of the Earth’s orbit around the Sun, the obliquity, and the precession of the Earth’s axis (Hays et al., 1976).

Eccentricity is a measure of the deviation of Earth’s orbit around the Sun from a perfectly circular orbit (eccentricity = 0). The orbit of the Earth around the Sun varies from an almost exact circle (eccentricity = 0.0005) to a slightly elongated shape (eccentricity = 0.0607) with periods of 100,000 and 400,000 years (figure 1.2).
1.1. Variability of the climate system

Eccentricity changes influence the amount of solar energy received from perihelion (closest distance from the Sun) to aphelion (farthest distance from the Sun).

Obliquity is the variation of the tilt of the Earth’s spin axis away from the axis perpendicular to the Earth’s orbital plane. The tilt varies between 22.1° and 24.5° on a 41,000-year cycle (figure 1.3). As this tilt increases, the seasons become more pronounced. A strong tilt means warmer summers and colder winters. For an increase of 1° in obliquity, the total energy received by the summer hemisphere at high latitudes increases by approximately 1%.

Finally, precession is the change in the orientation of the Earth’s spin axis, caused by a wobble of the Earth’s spin axis with a period of about 19,000 – 26,000 years (figure 1.4). While obliquity affects the tilt of the Earth’s spin axis, precession of the equinoxes affects the direction of the Earth’s spin axis. Variations in the axis location change the dates of perihelion and aphelion thereby increasing the seasonal contrast in one hemisphere while decreasing it in the other hemisphere. At any given latitude, the precession forcing is eccentricity modulated. Currently, the Earth is closest to the Sun in the northern hemisphere winter, which makes the winters there less severe and the northern hemisphere summers less warm.

1.1.2 The influence of ice sheets on the climate

During glacial times, a large portion of North America and northern Eurasia was covered by ice sheets (Peltier, 1994). In some places the ice sheets were more than 2500m thick. Statistical analyses of paleoclimatic data have shown that northern hemispheric ice sheets have waxed and waned with the same periods (100, 41 and 23 ky) as the orbital parameters (eccentricity, obliquity and precession). Imbrie et al. (1993) suggested that ice sheets might amplify or drive significant variability at millennial as well as orbital timescales. Variations in insolation create large ice sheets
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Figure 1.3: Schematic of the changes in the tilt of the Earth’s axis (obliquity) from 22.1° to 24.5° on a 41,000-year cycle. (Image by Simmon (2008))

Figure 1.4: Illustration of the precession (wobble), change in orientation of the Earth’s spin axis over a 19,000 to 26,000-year cycle. (Image by Simmon (2008))
1.1. Variability of the climate system

through feedbacks between ice accumulation rate, ice sheet extent and albedo. Once in existence, ice sheets influence regional and global climate by altering atmospheric and oceanic circulation (Manabe and Broccoli, 1985).

By decreasing absorption of solar radiation, ice sheets cool the air right above them. Due to their height, ice sheets also block and deflect atmospheric circulation (Clark et al., 1995; Felzer et al., 1996). Modelling studies suggest that large northern hemisphere ice sheets might have induced a southward displacement of the winter jet stream (Manabe and Broccoli, 1985) during glacial times. Justino et al. (2005) further suggest that the northern hemisphere ice sheets also led to a reorganization and strengthening of the storm tracks along the southern margin of the Laurentide ice sheet and across the North Atlantic region. The cooling over the ice sheets is then transmitted through the atmosphere to adjacent regions. Moreover, the enhanced cooling at high northern latitudes due to the presence of the ice sheets induces a greater equator to pole temperature gradient. This in turn leads to stronger winds (deMenocal and Rind, 1993), which cool the tropical regions (Timmermann et al., 2004). However, the influence of northern hemisphere ice sheets is limited to northern hemisphere climate (Manabe and Broccoli, 1985).

Due to the large glacial ice sheets, the amount of water stored on land was much greater during glacial than interglacial times. As a result, the glacial sea level was about 120m lower compared to the modern one and the global ocean salinity was greater by about 1 psu (Yokoyama et al., 2000; Lambeck and Chappell, 2001). Changes in oceanic nutrient reservoir and dynamics have been hypothesized in response to the greater area of continental shelves exposed during glacial times (Broecker, 1982; McElroy, 1983). However, the impact of the lower sea level on glacial nutrient reservoirs is still badly constrained. The changes in coastal areas resulting from the lower sea level also certainly modified coral reef build up, which had implications for the global carbon cycle. The effects of sea level changes on the carbon cycle will be discussed in more detail in Section 1.2. Glacial-interglacial variability/ Role and causes of glacial-interglacial atmospheric CO$_2$ variations.

In addition, ice sheets are the primary driver of millennial scale variability through the intermittent release of freshwater into the North Atlantic or the Southern Ocean. Freshwater inputs in the northern North Atlantic can potentially weaken the North Atlantic Deep Water formation through a lowering of surface water density. As will be discussed below, the oceanic circulation is of outmost importance for the redistribution of heat from the tropics to the poles. Disruptions of this circulation can greatly impact the climate system.

1.1.3 Role of the AMOC in the climate system

About half of the sunlight reaching Earth is absorbed by the oceans, where it is temporarily stored near the surface. Of the heat absorbed by the oceans, a part is released locally to the atmosphere, mostly through evaporation and infrared radia-
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The remainder is transported by currents, generally from low to mid-latitudes. The Atlantic Meridional Overturning Circulation (AMOC), which is a result of wind-driven and thermohaline-driven processes will be briefly discussed here.

The surface part of this circulation is characterized by warm and increasingly salty near-surface waters exported from the Pacific Ocean to the Atlantic Ocean via the Indian Ocean. At present, most of the oceanic poleward heat transport to the North Atlantic is carried by the surface part of the AMOC (Bryden and Imawaki, 2001). Ganachaud and Wunsch (2003) estimated that about 1.8 PW (1 PW=10^{15} W) of heat is transported at 24°N poleward by the AMOC, compared to a maximum of about 5 PW transported at 40°N by the atmosphere (Trenberth and Caron, 2001).

Due to the strong evaporation and cooling occurring in this area, the surface waters become dense enough to sink to the bottom in the Norwegian and Greenland Seas. This newly formed North Atlantic Deep Water (NADW) then flows southward as a deep western boundary current, joining the Antarctic Circumpolar Current (ACC) south of 30°S (figure 1.5). Bottom waters are also formed in the southern hemisphere, in the Weddell Sea and to a lesser extent in the Ross Sea and along the Adélie coast (140-150°E) (Rintoul et al., 2001; Schlosser et al., 2001). The Antarctic Bottom Water (AABW) is formed by ocean-atmosphere and ocean-ice interactions near the Antarctic shelves. Cold dry air blowing off the Antarctic continent cools the surface waters of the Southern Ocean to sub-freezing temperatures. Sea ice forms, rejecting salt that increases the density of the water underneath. These very cold and very saline waters sink over the Antarctic continental shelf into the abyssal ocean. They eventually mix with the NADW in the ACC and flow into the abyssal Atlantic, Indian and Pacific Oceans. Abyssal waters eventually resurface either by large scale upwelling or diapycnal mixing. Remineralization of organic matter causes large quantities of DIC to accumulate in deep waters. In addition, as deep waters are very cold and as CO₂ solubility increases with decreasing temperature, the ocean deep waters are a major carbon reservoir.

During glacial times, large ice sheets covered the high northern and southern latitudes. Due to ice sheet instabilities, intermittent release of freshwater occurred into the northern North Atlantic and the Southern Ocean. The addition of freshwater at the ocean surface lowers the density of the surface waters, thus slowing down the convection process. Meltwater pulses in the northern North Atlantic or in the Southern Ocean weaken the thermohaline-driven part of the AMOC, which considerably reduces the amount of heat transported to high latitudes. Chapter 4 of this thesis describes in more detail the climate and carbon cycle responses to meltwater pulses in the northern North Atlantic.
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Figure 1.5: Sketch of the major water masses in the Atlantic basin. The main water masses are the Antarctic Bottom Water (AABW), the North Atlantic Deep Water (NADW), the Antarctic Intermediate Water (AAIW) and the surface waters.

1.1.4 Role of the carbon cycle in the climate system

On Earth, carbon is stored in different reservoirs: the atmosphere, the terrestrial biosphere and soil, the ocean and the lithosphere (figure 1.6). The atmosphere is the smallest reservoir and it presently holds about 760 GtC (Gigatons of Carbon: $10^{12}$ kgC) in form of CO$_2$. As atmospheric CO$_2$ is a powerful greenhouse gas, changes in its atmospheric content will impact the climate system. The amount of CO$_2$ present in the atmosphere results from complex interactions with the carbon stored in the other reservoirs. For processes taking place on millenial to glacial-interglacial timescales, the carbon exchanges between the atmosphere and the marine and terrestrial carbon reservoir dominate. Indeed, even though the lithosphere is the largest carbon reservoir, the carbon exchanges between the lithosphere and the other carbon reservoirs are not significant on glacial-interglacial timescales. Thus, only the terrestrial and marine carbon cycles are briefly introduced below.

Terrestrial carbon cycle

Terrestrial vegetation plus living and dead carbon in soils make up the terrestrial carbon pool, and account for 2000 GtC. Terrestrial plants perform photosynthesis to convert atmospheric CO$_2$ into carbohydrates, while vegetation respiration and organic matter decay in soils release CO$_2$ to the atmosphere. Terrestrial primary productivity (and therefore CO$_2$ uptake) depends on nutrient availability in soils as well as on climate conditions. About 120 GtC/yr is exchanged in between the atmosphere and the terrestrial reservoir through photosynthesis and respiration.
Carbon transfer from the terrestrial to the marine reservoir occurs by the river transport of inorganic and organic carbon to the coastal zone.

**Marine carbon cycle**

The ocean reservoir contains about 40,000 GtC in the form of dissolved inorganic carbon and organic matter. Dissolved CO$_2$ at the ocean surface is constantly exchanged with the overlying atmospheric CO$_2$ so as to reach equilibrium. The net CO$_2$ flux depends on the pCO$_2$ difference between the ocean (pCO$_{2sw}$) and the atmosphere (pCO$_{2atm}$) as well as on the gas transfer velocity ($k_s$) (equation (1.1)), which is mainly a function of wind speed.

$$\text{FCO}_2 = k_s (p\text{CO}_{2atm} - p\text{CO}_{2sw}) \quad (1.1)$$

The pCO$_{2sw}$ at the ocean surface is given by Henry’s law (equation (1.2)). $K_h$ is the solubility coefficient of CO$_2$ in seawater and is a function of temperature and salinity.

$$[\text{CO}_2(aq)] = K_h p\text{CO}_2 \quad (1.2)$$

In the ocean, dissolved CO$_2$ exists in three different inorganic forms: CO$_2(aq)$, HCO$_3^-$ and CO$_3^{2-}$ (Zeebe and Wolf-Gladrow, 2001). The carbonate species are related by the following equilibria:

$$\text{CO}_2(aq) + H_2O \leftrightarrow K_1^1 H^+ + HCO_3^- \leftrightarrow K_2^2 2H^+ + CO_3^{2-} \quad (1.3)$$

The first and second dissociation constants of carbonic acid are respectively $K_1$ and $K_2$. The dominant inorganic form at the sea surface at a pH of around 8 is HCO$_3^-$. Two quantities are defined to characterize the equilibrium of inorganic carbon in solution: Dissolved Inorganic Carbon (DIC) and total alkalinity (ALK). DIC is the sum of all the inorganic carbon species present in the ocean (equation (1.4)).

$$\text{DIC} = [\text{CO}_2]_{aq} + [\text{HCO}_3^-] + [\text{CO}_3^{2-}] \quad (1.4)$$

Total alkalinity, as defined by Dickson (1981), is a measure of the proton deficit of seawater relative to an arbitrarily defined zero level of protons. Alkalinity is calculated from the equation (1.5).

$$\text{ALK} = [\text{HCO}_3^-] + 2[\text{CO}_3^{2-}] + [\text{B(OH)}_4^-] + [\text{OH}^-] - [H^+] + \text{minor components} \quad (1.5)$$

The partial pressure of CO$_2$ at the ocean surface (pCO$_{2sw}$) is a function of DIC, alkalinity, temperature, salinity and pressure. CO$_2$ solubility at the ocean surface increases with decreasing temperature and salinity (”solubility pump”).
1.1. Variability of the climate system

At constant alkalinity, pCO$_{2sw}$ increases as DIC increases. However, at constant DIC, as alkalinity increases, the inorganic carbon reservoir is shifted away from aqueous CO$_2$ and thus pCO$_{2sw}$ decreases. During CO$_2$ exchange in between the atmosphere and the ocean, only DIC varies.

The marine carbon cycle also involves the production and recycling of two types of carbon-rich materials: organic matter and calcium carbonate (CaCO$_3$). Photosynthesis by marine phytoplankton transforms DIC into organic matter in the euphotic zone and exports it to depth. Photosynthesis can be described by the following idealized chemical equation (Redfield et al., 1963):

$$
106 \text{CO}_2 + 16 \text{NO}_3^- + \text{H}_2\text{PO}_4^- + 17 \text{H}^+ + 106 \text{H}_2\text{O} \rightarrow (\text{CH}_2\text{O})_{106}(\text{NH}_3)_{16}(\text{H}_3\text{PO}_4) + 138 \text{O}_2 
$$

Equation (1.6)

The characteristics C:N:P:O$_2$ ratios of 106:16:1:138 in equation (1.7) are known as Redfield ratios. These ratios can however slightly vary depending on the environmental conditions. Marine export production depends mainly on the light level and on the availability of macronutrients (nitrates, phosphates, silicates). However, in the High Nutrient Low Chlorophyll (HNLC) regions (i.e. the Southern Ocean, the North Pacific and to a lesser extent the Eastern Equatorial Pacific), primary production remains low despite high nutrients content in the euphotic zone. Martin (1990) suggested that the micronutrient iron might be the limiting nutrient in these regions.

A part of the organic matter produced sinks to depth and is then remineralized below the euphotic zone. This so-called “biological pump” lowers DIC and thus pCO$_{2sw}$ at the surface while increasing it at depth.

Marine organisms such as coccolithophorids and foraminifera also secrete calcitic shells, while pteropods produce aragonitic shells. Both calcite and aragonite consist of calcium carbonate (CaCO$_3$) but differ in their mineralogy. The rain ratio, the relative delivery rate of CaCO$_3$ and particulate organic carbon (CaCO$_3$:POC) varies usually between 0.1 and 0.25. CaCO$_3$ precipitation leads to a decrease in DIC and alkalinity in the euphotic zone in a ratio 1:2, and thus increases the concentration of CO$_2$ (“carbonate pump”, equation (1.7)) (Zeebe and Wolf-Gladrow, 2001).

$$
\text{Ca}^{2+} + 2 \text{HCO}_3^- \xrightarrow{\text{precipitation}} \text{CaCO}_3 + \text{CO}_2 + \text{H}_2\text{O} 
$$

Equation (1.7)

$$
\text{CaCO}_3 + \text{CO}_2 + \text{H}_2\text{O} \xrightarrow{\text{dissolution/weathering}} \text{Ca}^{2+} + 2 \text{HCO}_3^- 
$$

Equation (1.8)

Other marine organisms, such as diatom or radiolarians, secrete silica tests (SiO$_2$). CaCO$_3$ and SiO$_2$ shells then sink to depth, where they are partly dissolved (equation (1.8)). The dissolution of CaCO$_3$ shells depends on the calcite (or aragonite) saturation state of the ocean, which is a function of depth. CaCO$_3$ shells are preserved in supersaturated waters and start to dissolve below the saturation
horizon. At present, the calcite lysocline, which is the depth at which significant amounts of calcite start to dissolve on the seabed, lies between 4000m and 5000m in the Atlantic Ocean, whereas it is found at around 4000m in the Pacific Ocean. This is mainly the result of the increasing corrosiveness of the deep waters (Archer, 1996; Zeebe and Wolf-Gladrow, 2001). As aragonite is more soluble than calcite, the aragonite lysocline lies at about 3000m in the Atlantic Ocean whereas it is shallower than 1000m in the Pacific Ocean (Zeebe and Wolf-Gladrow, 2001).

On timescales of 10,000 years or more, the depth of the lysocline is regulated by the carbonate compensation mechanism (Broecker and Peng, 1982). Any imbalance between the influx of dissolved HCO$_3^-$ brought about by weathering on land and CaCO$_3$ burial in the deep sea will act to change ocean pH until the flux balance is restored. The weathering of both carbonates and calcium silicates on land produces dissolved HCO$_3^-$ (equations (1.8) and (1.9)), which is then deposited as CaCO$_3$ in the ocean.

\[
\begin{align*}
\text{CaSiO}_3 + 2 \text{CO}_2 + 3 \text{H}_2\text{O} & \rightarrow \text{Ca}^{2+} + 2 \text{HCO}_3^- + \text{H}_4\text{SiO}_4 \quad (1.9) \\
& \rightarrow \text{CaCO}_3 + \text{SiO}_2 + \text{CO}_2 + 3 \text{H}_2\text{O}
\end{align*}
\]

An increase in the riverine flux of HCO$_3^-$ would increase the ocean inventories of DIC and alkalinity and thus decrease atmospheric CO$_2$. Greater alkalinity would also lead to an increase of CaCO$_3$ burial in the deep sea and a deepening of the CCD. The equilibration timescale for this process is of the order of 10,000 years for calcium carbonate weathering (Archer et al., 2000), whereas it is 100,000 to one million years for calcium silicates weathering.

To summarize, marine primary productivity lowers pCO$_2$ at the ocean surface, whereas carbonate precipitation slightly increases it. As atmospheric CO$_2$ is a powerful greenhouse gas, variations in the marine carbon cycle can impact the climate system through the solubility, biological or carbonate pump.

### 1.2 Glacial-interglacial variability

As recorded in deep-sea sediment cores and ice cores, the climate of the Quaternary (\(\sim 1.6\) million years ago to present-day) has been dominated by glacial-interglacial cycles. During glacial periods large ice sheets covered the northern hemisphere and Antarctica and, as a result, the sea level was lower by about 120m as compared to today (Yokoyama et al., 2000; Lambeck and Chappell, 2001). Analyses of air trapped in Antarctic ice cores also reveal that atmospheric CO$_2$ was about 80 ppmv lower and methane 300 ppbv lower during glacial times compared to interglacials (Petit et al., 1999).

Variations in the Earth’s orbit, which determine the amount and geographical distribution of received solar insolation, are the dominant external forcing of
1.2. Glacial-interglacial variability

the glacial-interglacial cycle. As seen in figure 1.7, during the last 400 kyrs the glacial-interglacial cycles varied on a characteristic timescale of 100-120ka (EPICA community members, 2004; Jouzel et al., 2007). These cycles are often linked to the major eccentricity cycle, which also modulates the amplitude of precessional forcing. However annual mean insolation changes associated with eccentricity cannot explain the full magnitude of glacial-interglacial variations. Additional internal feedbacks within the climate system have to be involved in these cycles.

Role and causes of glacial-interglacial atmospheric CO$_2$ variations

As shown in figure 1.7, atmospheric CO$_2$ and temperature in Antarctica co-varied over glacial-interglacial cycles. As atmospheric CO$_2$ is a powerful greenhouse gas, glacial-interglacial CO$_2$ changes of 80-90 ppmv certainly affected global climate. However, in spite of the clear importance of atmospheric CO$_2$ as an amplifier or a primary driver of the glacial cycles, the causes of its variations are still poorly understood. $\delta^{13}C$ analyses in benthic foraminifers (Crowley, 1991; Bird et al., 1994), pollen reconstructions (Crowley, 1995; Batjes, 1996; Adams and Faure, 1998) and modelling studies (Prentice et al., 1993; Friedlingstein et al., 1995; Francois et al., 1998, 1999; Kaplan et al., 2002; Otto et al., 2002; Köhler and Fischer, 2004) suggest that due to the colder and drier climate conditions during glacial times, the terres-
Figure 1.7: Time series from Vostok ice core (Antarctica) and ice volume for the last 400,000 years; Atmospheric CO$_2$ content (ppmv, thin blue line); deuterium profile (temperature proxy, green line); Ice volume proxy (arbitrary scale, thick blue line) calculated from Bassinot et al. (1994) seawater $\delta^{18}$O; and dust profile ($\mu$g g$^{-1}$, red line)
trial carbon stock was significantly reduced. The terrestrial carbon stock was about 300 to 1100 GtC lower during glacial times as compared to pre-industrial values. Thus, during glacial inceptions, the terrestrial biosphere was certainly a source of carbon to the atmosphere and not a sink. The marine carbon cycle might thus be responsible for a significant part of the atmospheric CO$_2$ draw down observed during glacial inceptions.

Different mechanisms have been suggested to explain the greater sequestration of carbon in the oceanic reservoir during glacial times (figure 1.8, table 1.1). Model results and theoretical considerations (Archer et al., 2003) suggest that any of these mechanism alone is not sufficient to explain the full magnitude of the glacial-interglacial atmospheric CO$_2$ changes and their timing. The mechanisms involving the solubility, the biological, the carbonate or the physical carbon pump are reviewed briefly here.

As ocean waters were globally colder during glacials, the solubility of CO$_2$ in the ocean was certainly enhanced and more CO$_2$ was sequestered in the deep sea (CLIMAP-Project-Members, 1981; Kucera et al., 2005). The estimated effect of lower ocean temperatures during glacial times on atmospheric CO$_2$ is of 20 to 30 ppmv (Heinze et al., 1991; Sigman and Boyle, 2000). But, as the storage of freshwater on land, mainly in ice sheets, was much greater during glacial times, the sea level was about 120m lower (Yokoyama et al., 2000; Lambeck and Chappell, 2001), and thus global ocean salinity was about 1 psu greater compared to today. This salinity increase would reduce CO$_2$ solubility and raise atmospheric CO$_2$ by about 6 ppmv. Increased CO$_2$ solubility can therefore account for one fourth of the glacial-interglacial CO$_2$ amplitude at most.

Biological productivity exports carbon from the surface ocean to the deep ocean in the form of sinking particles. Increased export production could thus lower the partial pressure of CO$_2$ at the ocean surface and therefore consequently decrease atmospheric CO$_2$. Broecker (1982); McElroy (1983) and Broecker (1998) suggested that, due to the lower sea level and the subsequent greater area of exposed shelves, the ocean’s surface nutrient inventory was greater during glacial times. This would have enhanced marine primary productivity. However, during the last deglaciation, between 20ka and 15ka, the sea level rose by only 20m which is not sufficient to flood the exposed shelves. The timing of the events during Termination I is thus not favorable to the shelf hypothesis. Broecker (1982) and Omta et al. (2006) further proposed that a change in Redfield ratio during glacial times could lead to a greater export of carbon to the deep sea. Another proposed mechanism invokes enhanced iron fertilization of the Southern Ocean (Martin, 1990). Indeed, Antarctic ice cores have revealed that the transport of dust to high southern latitudes was greater during glacial times due to the colder and drier conditions (figure 1.7). A higher flux of dust into the Southern Ocean would increase the iron content and potentially lead to greater export production there. But, Röthlisberger et al. (2004) estimate that
iron fertilization during glacial times could have led to a decrease in atmospheric CO₂ of only 20 ppmv. As dust contains also a significant amount of silica, Harrison (2000) proposed that a greater supply of silica to the ocean during glacial times could have shifted the phytoplankton population towards diatoms at the expense of coccoliths. This would have decreased the global production of calcite.

Another way to lower atmospheric CO₂ is to increase ocean pH in order to convert seawater CO₂ into HCO₃⁻ and CO₃²⁻. An increase in ocean pH could be obtained by increasing the silicate weathering on land, which would result in high riverine input of HCO₃⁻ to the ocean. A decrease in CaCO₃ production or a shift of CaCO₃ deposition from shallow to deep waters would also increase the burial of CaCO₃ in the deep sea, which would lower atmospheric CO₂ (Archer and Maier-Reimer, 1994; Sigman et al., 1998). However, a pH increase would lead to a deepening of the lysocline, which seems to be at odds with paleo-proxies (Archer et al., 2000; Anderson and Archer, 2002).

Finally, another set of explanations to lower atmospheric CO₂ involves ocean circulation and stratification, in particular in the Southern Ocean. The upwelling of Circumpolar Deep Waters in the Southern Ocean exposes DIC-rich waters to the surface. Francois et al. (1997); Toggweiler (1999); Stephens and Keeling (2000) and Gildor et al. (2002) suggested that a greater Southern Ocean stratification and sea ice cover during glacial periods would have isolated those DIC-rich waters from the surface. An increase in Southern Ocean stratification during glacial time could be obtained by greater summer sea ice melting. However, until now only experiments using box models have led to significant variations in atmospheric CO₂ through this mechanism (Archer et al., 2003).

Recently, Toggweiler et al. (2006) proposed a new hypothesis by which a weakening and equatorward shift of the southern hemisphere westerlies could have led to a lowering of atmospheric CO₂ content during glacial times. The strong southern hemispheric westerlies induce a northward Ekman transport and the upwelling of Circumpolar Deep Water (CDW) in the Southern Ocean. Those deep waters brought to the surface are CO₂ and nutrient-rich. They return to the subsurface before all the nutrients are utilized, leading to a CO₂ transfer from the ocean to the atmosphere. An equatorward shift of the westerlies would lead to a weaker upwelling of CO₂ rich waters close to Antarctica, which would therefore reduce the CO₂ source to the atmosphere. As CO₂ is a powerful greenhouse gas, a lower atmospheric CO₂ content would induce a cooling of the southern ocean, which would in turn lead to a sea ice advance. Due to the greater sea ice area, less CO₂ can be ventilated from the ocean to the atmosphere, which then causes an even greater equatorward shift of the westerlies. Using an idealized general circulation model, Toggweiler et al. (2006) proposes a self-sustained mechanism that could explain a CO₂ draw down of about 35 ppmv. This hypothesis will be tested in Chapter 6.
Figure 1.8: Schematic of possible mechanisms to explain the glacial to interglacial variations in atmospheric CO$_2$. Rectangles represent the mean change in atmospheric CO$_2$ obtained for each mechanism, while the bars represent the scatter in the changes obtained by different studies (Lenton, 2007).
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Hypothesized changes in the Ocean surface carbon system include:

- **Solubility pump**
  - (-) SST
  - (+) SSS

- **Biological pump**
  - (+) Export Production
  - (+) Surface nutrients
  - (+) Redfield ratio
  - (+) Surface [Fe]

- **Carbonate pump**
  - (-) CaCO$_3$ precipitation
  - (+) Surface [Si(OH)$_4$]
  - (+) Exposed shelves

- **Physical pump**
  - Southern Ocean
  - (+) sea ice area
  - (+) stratification
  - (-) westerly winds strength

<table>
<thead>
<tr>
<th>Hypothesized changes</th>
<th>Ocean surface carbon system</th>
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<tr>
<td>(-) SST</td>
<td>(-) pCO$_{2sw}$</td>
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<tr>
<td>(+) SSS</td>
<td>(+) pCO$_{2sw}$</td>
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<tr>
<td>(+) Export Production</td>
<td>(-) Surface DIC</td>
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<td>(+) Surface nutrients</td>
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<td>(+) Redfield ratio</td>
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<td>(+) Surface [Fe]</td>
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<td>(-) CaCO$_3$ precip</td>
<td>(-) pCO$_{2sw}$</td>
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<tr>
<td>(+) Surface [Si(OH)$_4$]</td>
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<tr>
<td>(+) Exposed shelves</td>
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</tbody>
</table>
| Southern Ocean       | (-) CO$_2$ transfer to atm | (–) Surface DIC
| (+) sea ice area     |                            |
| (+) stratification   | (-) Surface DIC            |
| (-) westerly winds strength | (-) Surface DIC |

Table 1.1: The table shows different mechanisms to explain the lowering of atmospheric CO$_2$ during a glacial inception. Increase in surface nutrients, in the Redfield ratio or in iron availability could have increased the export production. While, increased surface silicic acid [Si(OH)$_4$] and area of exposed shelves could have lowered the carbonate precipitation. The changes (positive (+) or negative (–)) are reported for a glacial inception.
1.3 Millennial-scale variability

Superimposed on the orbitally-driven component of the glacial to interglacial transitions are millennial-scale features such as the Dansgaard/Oeschger events, the Heinrich events (figures 1.9 and 1.10) or the Bølling-Ågerod/Younger Dryas transition.

Dansgaard/Oeschger (D/O) events are rapid climate fluctuations occurring approximately every 1500 years throughout the last glacial period (Alley, 2000). Greenland ice cores suggest that these events were characterized by rapid air warming of ~5 °C at high northern latitudes, followed by a gradual cooling. Some D/O cold spells also coincide with Heinrich events, suggesting a link between the two (Bond and Lotti, 1995).

During the last glacial period, ice sheet instabilities led to abrupt and massive discharges of icebergs into the North Atlantic. The mechanisms proposed to explain ice sheet instabilities include ice sheet internal processes (binge-purge mechanism) (MacAyeal, 1993), ice-load-induced earthquakes (Hunt and Malin, 1998) and ice shelf build up/collapse (Hulbe, 1997). An alternative explanation includes massive flood from glacial lakes (Johnson and Lauritzen, 1995). These so-called Heinrich events, were identified in marine sediment cores of the North Atlantic as pronounced layers of Ice Rafted Detritus (IRD) (Heinrich, 1988). Recent studies suggest that the sea level rose by 10–30m during Heinrich events (Lambeck and Chappell, 2001; Yokoyama et al., 2001; Chappell, 2002; Lambeck et al., 2002; Siddall et al., 2003; Roche et al., 2004), and the duration of these events was estimated at 500 +/- 250 years (Hemming, 2004; Roche et al., 2004). The addition of such a large amount of freshwater into the northern North Atlantic weakened or even shut down the Atlantic Meridional Overturning Circulation (AMOC) temporarily (Vidal et al., 1997; McManus et al., 2004) with impacts on the large-scale climate system.

Ice cores from Greenland (Dansgaard et al., 1993), marine sediment cores from the North Atlantic (Bond et al., 1992; Bond, 1993; Bard et al., 2000), the North Pacific (Harada et al., 2006) and the Mediterranean Sea (Rohling et al., 1998; Cacho et al., 1999) as well as lake and loess records from Europe (Shi et al., 2003; Watts et al., 1996; Thouveny et al., 1994) and Asia (Swann et al., 2005) provide unequivocal evidence for a wide-spread northern hemispheric cooling during Heinrich events, prevailing for several hundreds to thousand of years. As shown in paleo-climate records (Blunier et al., 1998) as well as in some climate model simulations of different complexity (Stocker and Johnsen, 2003; Knutti et al., 2004), a shutdown of the AMOC leads to a temperature increase in the southern hemisphere through the so-called bipolar seesaw effect (Stocker and Johnsen, 2003).

Similar climate anomalies occurred in response to a slow down of the AMOC (McManus et al., 2004) during the Younger Dryas (YD) cold interval around 12.7-11.6 ka BP. Unlike the Heinrich events that were associated with iceberg surges throughout the North Atlantic, the YD event could have been caused by the drainage
Chapter 1. Introduction

Figure 1.9: (Top) Time series for the last deglaciation of $\delta^{18}$O as measured in GISP ice core (top); $\delta$D as measured in EPICA dome C ice core (middle) (Röthlisberger et al., 2003); atmospheric CO$_2$ content as measured in EPICA Dome C ice core (bottom) (Monnin et al., 2004). The green rectangle represents the Younger Dryas, the purple rectangle represents the Bolling-Allerød and the yellow rectangle Heinrich event 1.
1.3. Millennial-scale variability

Figure 1.10: Time series for the last glacial period of $\delta^{18}$O as measured in GISP ice core (top); $\delta$D as measured in Dome C ice core (EPICA community members, 2004) (middle); atmospheric CO$_2$ content (bottom) as measured in Taylor dome ice core (blue) (Indermühle et al., 2000) and in Byrd ice core as measured by Neftel et al. (1988) (green) and Ahn and Brook (2007) (red). The red rectangles represent Heinrich events.
of Lake Agassiz (Teller et al., 2002; Carlson et al., 2007), which added significant amount of freshwater into the North Atlantic (Fairbanks, 1989). Although their triggers were substantially different, it is conceivable to assume that Heinrich events and the YD led to similar climate-carbon cycle responses.

Heinrich event I (∼16–14.5 ka B.P.) and the YD may have helped to amplify the glacial-interglacial transition in the southern hemisphere through the bipolar seesaw effect and a greater atmospheric CO$_2$ content. Indeed, recent analyses of air trapped in Antarctic ice core reveal that millennial-scale cold periods in the northern hemisphere were accompanied by an atmospheric CO$_2$ increase of about 20 ppmv (figure 1.10) (Ahn and Brook, 2007). Reorganizations of terrestrial and marine carbon cycles due to the reduction of North Atlantic Deep Water formation may have led to a greater atmospheric CO$_2$ content. Possible linkages between millennial-scale CO$_2$ changes and Heinrich events were identified and elucidated using 2-dimensional ocean-energy balance-marine carbon cycle models (Marchal et al., 1998, 1999) and vegetation models, such as the Lund-Potsdam Jena Dynamic Global Vegetation Model (LPJ-DGVM). Neglecting vegetation effects, Marchal et al. (1998) and Marchal et al. (1999) found a 10–30 ppmv CO$_2$ change in response to an AMOC shutdown. This response was explained in terms of reduced ocean CO$_2$ solubility, due to higher Southern Ocean temperatures. On the other hand, employing only a crude marine carbon cycle component but a complex vegetation model Scholze et al. (2003) and Köhler et al. (2005) identified the northern hemispheric vegetation as an important source of CO$_2$ during periods of reduced AMOC.

It is timely to revisit the issue of millennial-scale carbon cycle dynamics using a model that captures both, the terrestrial and the marine carbon cycle adequately and that can be run over many centuries.
1.4 Objectives and approach

The proposed research aims at exploring the feedbacks between atmospheric CO$_2$ and global climate that could have played a role in the millennial as well as glacial-interglacial climate variability using a series of transient climate simulations performed with a global coupled atmosphere-sea ice-ocean-vegetation-marine carbon cycle model. The model used in this study is LOVECLIM (Chapter 2), an Earth system Model of Intermediate Complexity (EMIC).

The hypotheses tested in this thesis are:

1. Changes in the southern hemispheric wind strength during glacial-interglacial transitions are a main driver of glacial-interglacial atmospheric CO$_2$ variations.

2. A shut down of the Atlantic Meridional Overturning Circulation during Heinrich events led to an increase in atmospheric CO$_2$. This CO$_2$ increase was primarily due to a release of carbon from the terrestrial biosphere.

3. The increase in atmospheric CO$_2$ during Heinrich events enhances southern hemispheric warming.

To the greatest extent possible, the climate and biogeochemical background states obtained from the modeling simulations will be compared to recent paleo-proxies data representing ocean ventilation, SST, sea ice, marine productivity and terrestrial vegetation.

1.5 Outline of the thesis

The model used for this work is LOVECLIM, a coupled atmosphere-ocean-sea ice-carbon cycle model. Each component of the model is described in Chapter 2.

The experiments were performed from "equilibrium" climate states representing the pre-industrial and the Last Glacial Maximum (LGM) climates. The pre-industrial state is briefly described in Chapter 3. As there is a much greater amount of data describing our present climate than the pre-industrial state, a present-day run was also performed. Its results are compared in detail with present-day observations. Finally, the LGM state is compared to the available paleo-proxies.

The climate and marine carbon cycle responses to changes in the strength of the southern hemispheric westerlies are studied in Chapter 4.

Chapter 5 describes the climate and carbon cycle responses to a shut down of the Atlantic Meridional Overturning Circulation (AMOC) starting from both a pre-industrial and a glacial climate state. Recent analyses of atmospheric CO$_2$ trapped in Antarctic ice cores (Ahn and Brook, 2007) suggested that the atmospheric CO$_2$
content increased by about 20 ppmv during Heinrich events. Chapter 6 investigates the effect of such an atmospheric CO$_2$ increase onto the southern hemisphere climate.

Finally, a discussion and the conclusions of this work are given in Chapter 7.

Publications

Chapters 4 to 6 are manuscripts, which have been submitted and partly published as stand-alone papers to refereed scientific journals.


Chapter 6: Menviel, L., A. Timmermann, A. Mouchet, and O. Timm (2008), Enhanced southern hemispheric warming during Heinrich events due to atmospheric CO$_2$ increase, Paleoceanography, submitted.
Bibliography


Chapter 1. Introduction


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Chapter 2

Model description

The model used in this study is the Earth system model of Intermediate complexity, LOVECLIM, which includes numerical components for the atmosphere, the ocean, sea-ice as well as a marine and terrestrial carbon model. Figure 2.1 shows the different components of LOVECLIM and how they interact with each other. The version of LOVECLIM used here does not include an ice-sheet model. For the experiments performed under glacial state conditions, an ice-sheet mask (ICE4G) is used (Peltier, 1994).

2.1 The atmosphere: ECBilt

The atmospheric component of the coupled model LOVECLIM is ECBilt (Opsteegh et al., 1998), a spectral T21 (∼ 5.625° x 5.625°), three-level model (800hPa, 500hPa and 200hPa), based on quasi-geostrophic equations. At the top of the atmosphere (p=0hPa), the rigid lid condition is applied. Synoptic variability associated with weather patterns is explicitly computed.

Dynamics

The dynamical behavior of the atmospheric model is governed by the quasi-geostrophic equations. These are derived from the primitive equations under the assumption of small Rossby number. The dynamical core is extended by estimates of the neglected ageostrophic terms (Lim et al., 1991; Holton, 1992) in order to close the equations at the equator. These ageostrophic terms are computed from the divergent component of the horizontal wind and the vertical velocity (Justino, 2004). The basic equations in isobaric coordinates are given below.
Chapter 2. Model description

Figure 2.1: Schematic of the interactions between the components of LOVECLIM. For clarity, the ocean-atmosphere interactions are not represented on the graph. The following quantities are exchanged between ECBilt and CLIO: sea surface temperature, radiative, turbulent and freshwater fluxes, wind stress, albedo, thicknesses and fractions of sea ice and snow. The clouds are prescribed in ECBilt-CLIO and the river input of geochemical tracers is prescribed in LOCH.

- The vorticity equation:

\[
\frac{\partial \zeta}{\partial t} + \mathbf{V}_\psi \cdot \nabla (\zeta + f) + f_0 D + k_d \nabla^8 \zeta = -F_\zeta
\]  

(2.1)

\( \zeta \) is the vertical component of the vorticity vector \( (\zeta = \nabla^2 \psi) \), \( \psi \) is the stream-function, \( \mathbf{V}_\psi \) is the rotational component of the horizontal velocity, \( D \) is the divergence of the horizontal wind, \( f \) is the Coriolis parameter \( (f = 2\Omega \sin \varphi) \), \( \Omega \) is the angular velocity of the Earth, \( \varphi \) is the latitude, \( f_0 \) is \( f \) at 45\(^\circ\) North and South, \( k_d \nabla^8 \zeta \) is a highly scale selective diffusion and \( F_\zeta \) contains the ageostrophic terms in the vorticity equation (equation (2.2)).

\[
F_\zeta = \mathbf{V}_\chi \cdot \nabla (\zeta + f) + \zeta D + \omega \frac{\partial \zeta}{\partial p} + \mathbf{k} \cdot \nabla \omega \times \frac{\partial \mathbf{V}_\psi}{\partial p}
\]  

(2.2)

\( \mathbf{V}_\chi \) is the divergent component of the horizontal wind, \( \omega \) is the vertical velocity in isobaric coordinates, \( p \) is the pressure. Estimates of \( F_\zeta \) are used to represent the effect of the ageostrophic circulation on the tendency of the geostrophic vorticity.
2.1. The atmosphere: ECBilt

- The first law of thermodynamics:

\[
\frac{\partial}{\partial t} \left( \frac{\partial \psi}{\partial p} \right) + \mathbf{V}_\psi \cdot \nabla \left( \frac{\partial \psi}{\partial p} \right) + \frac{\alpha}{f_0} \omega + k_d \nabla^s \left( \frac{\partial \psi}{\partial p} \right) + k_R \left( \frac{\partial \psi}{\partial p} \right) = - \frac{RQ}{f_0 p c_p} - F_T
\] (2.3)

\(\sigma\) is the static stability \((\sigma = -\frac{\alpha}{\frac{\partial \theta}{\partial p}})\), \(\alpha\) is the specific volume, \(\theta\) is the potential temperature, \(k_R\) is a Rayleigh damping coefficient, \(R\) is the gas constant, \(c_p\) is the specific heat for constant pressure, \(Q\) is the diabatic heating and \(F_T\) is the advection of temperature by the ageostrophic wind (equation (2.4)).

\[F_T = \mathbf{V}_\chi \cdot \nabla \left( \frac{\partial \psi}{\partial p} \right)\] (2.4)

The hydrostatic equation relating the temperature \(T\) and the geopotential \(\phi\) (equation (2.5)), as well as the ‘geostrophic’ relation relating \(\phi\) and \(\psi\) \((\phi = f_0 \psi)\) have been used for the derivation of equation (2.3).

\[T = -\frac{p}{R} \frac{\partial \phi}{\partial p}\] (2.5)

- The continuity equation:

\[D + \frac{\partial \omega}{\partial p} = 0\] (2.6)

The quasi-geostrophic potential vorticity \((q)\) equation can be obtained by combining equations (2.1), (2.3) and (2.6).

\[
\frac{\partial q}{\partial t} + \mathbf{V}_\psi \cdot \nabla q + k_d \nabla^s (q - f) + k_R \frac{\partial}{\partial p} \left( \frac{f_0^2}{\sigma} \frac{\partial \psi}{\partial p} \right) = - \frac{f_0 R}{c_p} \frac{\partial}{\partial p} \left( \frac{Q}{\sigma p} \right) - f_0 F_T \frac{\partial}{\partial p} \left( \frac{\sigma}{\sigma p} \frac{\partial \psi}{\partial p} \right)
\] (2.7)

where \(q\) is defined as:

\[q = \zeta + f + f_0^2 \frac{\partial}{\partial p} \left( \sigma^{-1} \frac{\partial \psi}{\partial p} \right)\] (2.8)
Chapter 2. Model description

Radiation

Diabatic heating due to radiative fluxes, the release of latent heat and the exchange of sensible heat with the surface are parametrized. Solar radiation, defined as short-wave radiation, is reflected at the top of the atmosphere and at the surface. As the cloud cover climatology is prescribed in the model (Rossow et al., 1996), so is the influence of clouds on the radiation reflection. For our simulations, the time-varying daily averaged incoming insolation is calculated as a function of latitude using the algorithm of Berger (1978).

Absorption of shortwave radiation depends on clouds and on the gases present in the atmosphere. Longwave radiation emitted by the Earth’s surface follow the Stefan-Boltzmann law ($\sigma T^4$). Compared to the standard version of LOVECLIM we enhanced the sensitivity of ECBilt to longwave radiation forcing by a factor of 2 (Timm and Timmermann, 2007). The simulated global mean temperature response to such a perturbation amounts to about 3°C as compared to the 1.5°C for the standard sensitivity. The qualitative structure of the climate change pattern is stable to reasonable changes in the longwave sensitivity.

Moisture budget

The relative humidity is constant from the surface to the 500hPa level. Changes in the specific humidity are described by a single equation (equation (2.9)) for the total precipitable water content between the surface and 500hPa.

$$\frac{\partial q_a}{\partial t} = - \nabla \cdot (V_a q_a) + E - P$$  \hspace{1cm} (2.9)

$q_a$ is the total precipitable water content between the surface and 500hPa, $V_a$ is the transport velocity for $q_a$, $E$ is evaporation and $P$ precipitation. Above 500hPa, the atmosphere is assumed to be completely dry. Below 500hPa, precipitation occurs when $q_a > 0.8 q_{max}$, where $q_{max}$ is the vertically integrated saturation specific humidity. When the surface temperature is below 0, accumulation of snow occurs instead of precipitation. The amount of precipitation is given by:

$$\Delta q_a = \frac{(q_a - 0.8 q_{max})}{1 + \frac{\rho_w L_c g q_{max}}{c_p \Delta p \left. \Delta T_s \right|}}$$  \hspace{1cm} (2.10)

where $r$ is the relative humidity, $\rho_w$ is the density of water, $L_c$ is the latent heat of the condensation, $g$ is the gravity, $c_p$ is the specific heat of dry air at constant pressure and $T_s$ is the land surface temperature.

Evaporation over sea is given by equation (2.11):

$$E = \rho_a C_d \left| V_s \right| \left[ q_s(p_0, T_s) - q(p_0, T_s) \right]$$  \hspace{1cm} (2.11)

$\rho_a$ is the air density, $C_d$ is the drag coefficient, $\left| V_s \right|$ is the absolute value of the wind speed at the surface, $q$ is the specific humidity, $q_s$ is the saturation specific
humidity, $T_s$ is the surface air temperature. The surface wind is assumed to be 0.8 times the wind at 800hPa. The evaporation over land depends on the soil moisture content. The hydrological cycle is closed over land by a bucket model for soil moisture. Runoff occurs when the soil moisture content exceeds a certain maximum value. Each land grid point is connected to an ocean gridpoint to define the river runoff (figure 2.2).

2.2 The ocean and sea-ice: CLIO

The sea ice-ocean component of LOVECLIM, CLIO (Coupled Large-scale Ice-Ocean model) (Campin and Goosse, 1999; Goosse et al., 1999; Goosse and Fichefet, 1999) consists of a free-surface primitive equation model coupled to a thermodynamic-dynamic sea ice model. The resolution is $3^\circ \times 3^\circ$ and 20 unevenly spaced levels. To avoid a singularity at the North Pole, the oceanic component makes use of two subgrids: The first one is based on classic longitude and latitude coordinates and covers the whole ocean except for the North Atlantic and Arctic Ocean. These are covered by the second spherical subgrid, which is rotated and has its poles at the equator in the Pacific ($111^\circ$W) and Indian Ocean ($69^\circ$E) (Campin, 1997).
Chapter 2. Model description

Ocean

The equations governing large-scale geophysical flows are deduced from the Navier-Stokes equations written in a rotating frame of reference with some typical approximations:

- 1. The Boussinesq approximation which states that the variation of the density $\rho$ are small compared to its mean $\rho_0$ except if they are multiplied by a buoyancy term.

- 2. The depth of the ocean is considered negligible compared to the Earth’s radius and the vertical depth scale is supposed much shorter than the horizontal one.

- 3. The hydrostatic approximation is made.

To represent the mixing effects of quasi-horizontal meso-scale eddies that are too small (typical size from 10 to 100km) to be explicitly resolved by the model two parametrizations are applied. The model relies on the isopycnal mixing formulation using the approximation of the small slopes (Redi, 1982) and on the eddy-induced advection terms as proposed by Gent and McWilliams (1990).

The vertical mixing parametrization follows the Kantha and Clayson (1994) version of the 2.5 level turbulence closure scheme of Mellor and Yamada (1982). Moreover, a background diffusivity varying with depth is applied (Campin, 1997).

Due to the coarse resolution of the model, convection on continental shelves is not well represented. However, dense water that flows out of the continental shelves is thought to be a major source of Antarctic Bottom Water (AABW) formation (Gill, 1973). To resolve this problem, Campin and Goosse (1999) included a representation of density driven downsloping flow. Since there is no water cavity below ice shelves in the model, the interactions between ice shelves and ocean are represented through the Beckmann and Goosse (2003) parametrization of heat and freshwater fluxes at the ice shelf base.

Coupling between atmosphere and ocean is done via the exchange of freshwater and heat fluxes, rather than by virtual salt fluxes.

Sea ice

Sea ice evolution is governed by thermodynamic growth and decay as well as by ice dynamics and transport. The ice growth is a function of ice thickness and concentration: the smaller the ice thickness and concentration the faster ice grow and decay (Maykut, 1982), which in turn depends strongly on the advection pattern. Sea ice is considered as a homogeneous slab of ice covered by a snow layer. The evolution of temperature inside the snow-ice system is governed by a one-dimensional
heat-diffusion equation. The surface temperature is determined by considering the budget of a thin layer at the top surface. It is a function of the net shortwave radiation (albedo effect included), the downward longwave radiation, the emitted radiation, the turbulent fluxes of sensible and latent heat and the conductive heat flux from below. If the heat balance requires that the surface temperature is above the melting point, then the surface temperature is held at this point and the excess of energy is balanced by melting snow or ice. At the bottom of the ice slab, any imbalance between the conductive heat flux within the ice and the heat flux from the ocean is balanced by accretion or ablation of ice.

The floes within the ice pack are separated by zones of open water: leads and polynyas. Those areas of open water are very important since they allow direct exchanges between the atmosphere and the ocean. This fraction of open water is represented in the model by the ice concentration variable. The thermodynamic variations of the ice concentration are function of the heat budget of the open water (Fichefet and Maqueda, 1997).

Sea ice is considered as a viscous-plastic continuum (Hibler, 1979). This means that the ice behaves as a rigid plastic medium for high deformation rates and as a viscous fluid for small deformation rates.

**Coupled model**

The time step of the atmospheric model is 8h, whereas the ocean model has a time step of 1 day (3 atmosphere time steps for 1 ocean time step). During these 3 time steps the ocean surface is kept constant. Fluxes of heat, moisture and momentum are exchanged in between the ocean and the atmosphere. The ocean and sea ice cover exchange mass, freshwater, heat and momentum (Goosse, 1997).

### 2.3 VECODE

The terrestrial vegetation module of LOVECLIM, VECODE (VEgetation COntinuous DEscription model), is a two-dimensional dynamical terrestrial vegetation model that uses two plant functional types. It has been described in Brovkin et al. (1997) and has recently been coupled to the LOVECLIM model (Renssen et al., 2005). Based on annual mean values of temperature-dependent and hydrological climatic variables, the VECODE model computes the evolution of the vegetation cover described as a fractional distribution of tree, grass and desert in each land grid cell once a year.

Empirical relationships based on annual precipitation and the positive degree-day index (PPD index) govern the evolution of the vegetation type. The formulation prescribing the trees fraction \( f_t \) is based on the Olson and Watts (1992) classification. The main characteristics are:

- for a fixed temperature, \( f_t \) increases with precipitation increase
for fixed precipitation, \( f_t \) decreases with temperature increase

As trees never occupy a whole cell, the maximum tree fraction is chosen as 0.95.

There are two types of desert: cold and warm ones. The cold deserts are driven by temperature (i.e. Antarctica), while the warm deserts are driven by the lack of precipitation (i.e. Sahara desert). The fraction of grass fills the grid cell after the computation of the tree fraction and desert fraction.

The dynamics of trees and desert depends on climate change, but the change in tree/desert fraction is not adapted instantaneously to the climate change. Indeed, their growth is reduced by an exponential function depending on their respective growing rate.

To better compare the modeling results with climate reconstructions, we developed a module to assign biomes based on VECODE and ECBilt outputs. This module is equivalent to the one described by Roche et al. (2007), except that we grouped the temperate broadleaf forest and the cool mixed forest into a biome called mixed forest. This module is only applied to the outputs to make comparisons between our simulations and observations.

VECODE also computes the Net Primary Production (NPP) of the biomass, as a function of the climate parameters (equation (2.12)). A fertilization effect is also included, so that NPP also depends on the atmospheric CO\(_2\) content (pCO\(_2\), equation (2.13)).

\[
\alpha_1 = [1 - e^{(-0.000664 P_{an})}] ; \quad \alpha_2 = \frac{1}{1 + 3.73 e^{(-0.119 T_{an})}} \quad (2.12)
\]

\[
\text{NPP} = \min(\alpha_1, \alpha_2) \left[ 1 + 0.25 \frac{\log(2)}{\log(280)} \log\left(\frac{p\text{CO}_2}{280}\right) \right] \quad (2.13)
\]

\( P_{an} \) and \( T_{an} \) are respectively the mean annual precipitation (mm/yr) and temperature (°C). NPP is in kgC/m\(^2\)/yr.

Within the LOVECLIM version used here, simulated vegetation changes affect the land-surface albedo and the atmospheric CO\(_2\) content (through LOCH). They however have no influence on other processes such as evapo-transpiration.

### 2.4 LOCH

LOCH (Liége Ocean Carbon Heteronomous model) is a 3-dimensional global model of the oceanic carbon cycle simulating dissolved inorganic carbon (DIC), total alkalinity, phosphate (PO\(_4^{3-}\)), organic products, oxygen and silica (Mouchet and François, 1996; Fichefet et al., 2007). LOCH is fully coupled to CLIO, with the same time step. In addition to their biogeochemical transformations, a 3D tracer advection (and diffusion) scheme is implemented that uses the circulation from the CLIO model.
2.4. LOCH

The uptake of CO\textsubscript{2} by the ocean is governed by the solubility as well as biological pumps. The partial pressure of CO\textsubscript{2} in the surface waters is calculated from the total alkalinity, DIC, temperature and salinity. The difference between the partial pressure of CO\textsubscript{2} in the ocean and in the atmosphere (\(\Delta p\text{CO}_2\)), modulated by a gas transfer velocity (\(k_s\)), determines the CO\textsubscript{2} flux in between the ocean and the atmosphere (\(FCO_2\)) (equation (2.14)). The gas transfer velocity (\(k_s\)) is a function of the square of the wind speed (\(v\)) and of the square root of the Schmidt number (\(Sc\)) (Wanninkhof, 1992) (equation (2.14)). The CO\textsubscript{2} exchange between the ocean and the atmosphere is also dependent on sea-ice fraction. When the surface is totally covered by sea-ice, there is no exchange.

\[
FCO_2 = k_s \Delta p\text{CO}_2 \quad k_s = k \sqrt{\frac{Sc}{660}} v^2
\]  

(2.14)

LOCH computes the export production (XP) from the fate of a phytoplankton pool in the euphotic zone (0–120 m) (equation (2.15)). The phytoplankton growth depends on the availability of nutrients (\(PO_4^{3-}\) is the limiting nutrient) and light (Michaelis-Menten relationship), with a weak temperature dependence.

\[
XP = f(T) f(I) f(B) \frac{[PO_4]}{K_{PO_4} + [PO_4]}
\]  

(2.15)

\([PO_4]\) is the phosphate content, \(K_{PO_4}\) is the half saturation constant for the phosphate and is equivalent to 1\(\mu\)mol/l. \(f(T)\), \(f(I)\) and \(f(B)\) are functions of respectively the temperature, the insolation and the biomass.

\[
f(T) = g_{max} \frac{T}{T + 3}
\]  

(2.16)

\(g_{max}\) is the maximum growth rate (0.7 day\(^{-1}\)), \(T\) is the temperature. The insolation function depends on the incoming radiation at the ocean surface as well as on the sea ice coverage. A grazing process together with natural mortality limit the primary producers biomass and provide the source term for the organic matter sinking to depth. The model contains both Dissolved Organic Matter (DOM) and Particulate Organic Matter (POM).

Remineralization of organic matter (\(R_{POM}\)) occurs only below the euphotic zone. If the oxygen content is greater than 0 \(\mu\)mol/l, then POM is transformed in DOM at a rate defined by equation (2.17). In anoxic zones this rate is equal to 0.9 yr\(^{-1}\). The remineralization rate of DOM (\(R_{DOM}\)) is described by equation (2.18)

\[
R_{POM} = \frac{[O_2]}{[O_2] + 5 \times 10^{-6}} + 0.9 \times 5 \times 10^{-6}
\]  

(2.17)

\[
R_{DOM} = 0.05 \times R_{POM}
\]  

(2.18)
Chapter 2. Model description

Depending on the silicic acid availability, phytoplankton growth is accompanied by the formation of biogenic silica or CaCO$_3$ (calcite and aragonite) shells, which then sink to depth. One fifth of the CaCO$_3$ shells precipitated are formed of aragonite. CaCO$_3$ shells are dissolved depending on the calcite and aragonite saturation states. 94% of the opales produced is dissolved at the bottom of the ocean. The organic matter that is not remineralized and the shells that are not dissolved are permanently preserved in the sediments. This leads to a loss of alkalinity, carbon, phosphate and silica, which is compensated by the river influx.

The atmospheric CO$_2$ content is predicted for each ocean timestep from the air-sea CO$_2$ fluxes calculated by LOCH as well as from annual CO$_2$ fluxes provided by VECODE.

Bibliography


Chapter 3

Reference runs

This chapter describes the simulated pre-industrial and Last Glacial Maximum (LGM) equilibrium states that will be used as reference states throughout this thesis. In order to better compare the model performances with observations, a present-day run is also described.

In the first part, the pre-industrial and modern day runs are compared to the observations. First, the climate representation and its impact on the vegetation distribution are described. Then the simulated climatic variables at the surface of the ocean are compared to observations. The distribution of the phosphate content, the export production and the air-sea CO$_2$ fluxes are also described. Finally, the distribution of the simulated major oceanic water masses are evaluated. For that purpose the zonally averaged latitudinal-depth profiles of temperature, salinity, phosphate, oxygen, apparent oxygen utilization, DIC and alkalinity are compared to observations.

The second part of this chapter describes the LGM reference run and compare it to the available paleoproxies.

The model used in this thesis, LOVECLIM is an Earth system model of intermediate complexity (EMIC). In order to perform experiments on timescales of a few centuries to millenia, the atmospheric model used in LOVECLIM was chosen to be quasi-geostrophic (Opsteegh et al., 1998). This simplified atmospheric model enables LOVECLIM to be about 10 times faster than Coupled General Circulation Models (CGCMs). Even if higher resolution CGCMs represent the climate variables in a more realistic way than LOVECLIM, they oftentimes do not include a marine and terrestrial carbon models. In addition, performing a suite of millennial-scale experiments with CGCMs is hard to achieve due to computing time constraints.

3.1 Pre-industrial and modern day runs

The pre-industrial equilibrium state (PIN) is obtained by forcing LOVECLIM with 278 ppmv of atmospheric CO$_2$ during 500 years, then allowing the atmospheric CO$_2$
to vary freely during 1000 years. After this fully coupled spin-up, the simulated mean atmospheric CO$_2$ concentration is 282 ppmv. The standard deviation of the CO$_2$ concentration due to interannual variability is 1.7 ppmv. The global annual 2m air temperature averaged over the last 400 years of the PIN run is 13.7°C with a standard deviation of about 0.1°C. The global annual sea surface temperature averages to 18.4°C with a standard deviation of about 0.05°C. The partitioning of the pre-industrial carbon-reservoirs seems reasonable. The terrestrial biosphere holds about 2,050 GtC, which is in agreement with the present day estimate of 2,000 GtC (Houghton et al., 2001) and a release of about 120 GtC during the 19th and 20th centuries due to land use changes (Houghton, 1999). The ocean carbon reservoir amounts to 39,100 GtC, which also seems reasonable (Houghton et al., 2001).

The present-day run (MOD) was obtained by forcing the model with the observed atmospheric CO$_2$ variations from 1750 to 2000 year A.D. (Keeling et al., 1996; Etheridge et al., 1996) starting from the PIN state.

### 3.1.1 Climate representation and vegetation distribution

The global annual 2m air temperature at the end of the present-day run amounts to 14.6°C and the global annual sea surface temperature is 18.9°C. Figure 3.1 displays the annual air temperature as well as the annual mean long-term net radiation balance at the top of the atmosphere for the MOD run. The radiation balance is the difference between the amount of solar radiation absorbed by Earth’s surface and atmosphere and the amount of outgoing longwave radiation emitted by the system. The former is governed by the albedo, whereas the latter depends strongly on the atmospheric content of greenhouse gases, such as H$_2$O, CO$_2$, CH$_4$ and N$_2$O. The absorbed solar radiation exceeds the outgoing longwave radiation in the tropical and subtropical regions, resulting in a net radiative heating at these latitudes, while in the middle to polar latitudes there is a net cooling. This equator-to-pole gradient in radiative heating is the primary mechanism that drives the atmospheric and oceanic circulations.

Figure 3.2 displays the mean annual wind velocity at 800 hPa, as simulated during the MOD run and the one averaged from 775 to 825 hPa for the ERA-40 reanalysis. In general, the model and observations exhibit a large degree of similarity. Nevertheless, the simulated westerlies are weaker than in the Uppala et al. (2005) dataset. The simulated maximum of the northern hemispheric westerlies is also slightly shifted to the north. Justino (2004) suggested that the weaker westerlies were the result of weaker simulated transient and stationary eddies. In addition, the simulated tropical trade winds are too zonal and too weak (Opsteegh et al., 1998).

Justino et al. (2005) showed that in an earlier version of LOVECLIM without carbon cycle, the simulated mean storm tracks are well placed in comparison with observations and more complex CGCMs. However, as the amplitude of the storm...
3.1. Pre-industrial and modern day runs

Figure 3.1: Annual averaged air temperature (°C) (left) and net radiation balance at the top of the atmosphere (W/m²) (right) averaged over the final 20 years of the MOD run.

Figure 3.2: Mean annual wind velocity (m/s) at 800 hPa as simulated over the final 50 years of the MOD run (top left) and averaged from 775 to 825 hPa for years 1980 to 2001 for the ERA-40 reanalysis (top right) (Uppala et al., 2005).
Figure 3.3: Annual precipitation (cm/yr) averaged over the last 20 years of the MOD run (left) and present day combined annual precipitation averaged over the years 1979-2006 (right) as described in Janowiak and Xie (1999).

track is dependent on the atmospheric model resolution, the transient eddy kinetic energy at 500 hPa is generally underestimated, a problem that can be found for many T21 AGCMs.

Precipitation

Figure 3.3 shows the annual mean precipitation distribution over the final 20 years of the MOD run as well as for the CAMS-OPI data set averaged over the years 1979 through 2006 (Janowiak and Xie, 1999). The Climate Anomaly Monitoring System ("CAMS") combined with the OLR Precipitation Index ("OPI") provides real-time monthly analyses of global precipitation. Observations from raingauges ("CAMS" data) are merged with precipitation estimates from a satellite algorithm ("OPI").

The simulated precipitation in the MOD run is underestimated (-5%) in the tropical region along the Inter-Tropical Convergence Zone (ITCZ) in comparison with the observational records of Janowiak and Xie (1999) (figure 3.3). Indeed, the model has difficulties to represent the Hadley cell realistically due to its quasi-geostrophic approximation and the relatively weak ageostrophic corrections (Opsteegh et al., 1998). This leads to an underestimation of the vertical velocity in the ITCZ belt. The precipitation in the northern hemispheric stormtrack region is also underestimated by 2%. Due to an underestimation of important orographic features in the T21 resolution, the simulated precipitation in tropical eastern Africa is overestimated. The precipitation distribution is comparable to the results of low-resolution CGCMs. Existing precipitation biases need to be taken into account when discussing AMOC-induced shifts of the ITCZ in Chapter 5.
3.1. Pre-industrial and modern day runs

Figure 3.4: Zonally averaged annual net primary production (gC/m²/yr) for the pre-industrial state (red line). Average annual mean NPP for the NPP model intercomparison project (solid black line) and models standard deviation (dotted black lines) (Cramer et al., 1999).

Representation of the vegetation distribution for pre-industrial times

Figure 3.4 displays the zonally averaged terrestrial long-term mean Net Primary Production (NPP) for the pre-industrial simulation (blue line). The simulated NPP is compared with an average of the NPP simulated by the models used in the NPP model intercomparison project (black line) (Cramer et al., 1999). The vegetation models used in this comparison are more complex than VECODE and were forced by present-day observations. Figure 3.4 documents that the zonally averaged NPP simulated by VECODE is comparable to those simulated by more complex models. However, the NPP simulated by VECODE is in the lower range at the equator as well as at high northern and southern latitudes. In VECODE the NPP is computed annually from the mean annual temperature and precipitation field. In regions where seasonality is important for the terrestrial primary productivity (i.e. northern Eurasia), this annual computation leads to an underestimation of the NPP. As LOVECLIM slightly underestimates the global terrestrial NPP, it might also underestimate the changes in NPP described in the following chapters.

In figure 3.5, the biome distribution obtained for the PIN run is compared to the reconstructions compiled by Crowley (1995). The main desert areas are well represented except for Central Asia. Tropical forests dominate the Amazonian basin, the Congo basin as well as Indonesia and Papua New Guinea. However, rain forest is simulated in southeast Africa and southern India instead of savanna, and forest dominates the southwestern part of the United states instead of grass.
Discussion

LOVECLIM captures the overall structure of the climate variables. However, due to the simplified physics of the atmospheric model, the simulated westerlies and trades are weaker than the observed one. The weaker simulated wind field leads to a misrepresentation of the oceanic circulation in the North and Equatorial Pacific (see following sections).

Due to the quasi-geostrophic approximation of the atmospheric model, some discrepancies in the precipitation field arise along the ITCZ, particularly in the Pacific Ocean. Moreover due to the coarse horizontal and vertical resolution, some topographic features are not well represented (i.e. Western Highlands of Ethiopia), and thus some areas (such as Eastern Africa) exhibit a biased precipitation field (figure 3.3). This can lead to reversed east-west precipitation gradients and associated diabatic forcing over South America and Africa. For example, the simulated precipitation is greater in tropical eastern Africa than in tropical western Africa, whereas it should be the contrary.

As the simulated vegetation type is a direct function of precipitation and temperature, these discrepancies in the precipitation field then lead to erroneous vegetation types in some regions. This is particularly true at low latitudes, where the precipitation amount is the main control on the terrestrial biome.
3.1. Pre-industrial and modern day runs

However, as the main differences between the simulated and observed precipitation are over the ocean, the global vegetation representation is comparable to more realistic vegetation models driven by observed climate forcing. Moreover, if averaged zonally, some of the regional differences are averaged out. These biases in precipitation and vegetation fields have to be taken into account while analyzing the results of the experiments performed in this thesis. In general, specific regional features have to be considered with caution.

Finally, in the experiments performed in this work, the cloud cover was prescribed with the modern day climatology (Rossow et al., 1996), so that cloud-climate feedbacks are not captured here. Although cloud-climate feedbacks can play an important role in the climate system, they are still very poorly constrained by both observations and modelling studies with state of the art models (Bony et al., 2006). Indeed, even state of the art CGCMs exhibit large uncertainties with respect to the sign of the cloud feedback in some areas.

3.1.2 Ocean climatology and marine carbon cycle - Part I: surface processes

Sea surface height

Figure 3.6 shows the sea surface height (SSH) as simulated by the MOD run and compared to the observational values of Maximenko and Niiler (2004). For large scale features, there is reasonable qualitative agreement between the model and the observations. However, the simulated North Pacific subpolar gyre is stronger than in the observations. The simulated North Pacific subtropical gyre is slightly shifted northward, is stronger along the western boundary and extends much less eastwards than in the observations. The simulated Kuroshio extension and Gulf stream penetrate more northward than the observations. This typical low-resolution model feature is associated with enhanced heat transport into the extra-tropics (Justino, 2004). On the other hand the South Pacific subtropical gyre, as well as the North and South Atlantic subtropical gyres are underestimated compared to the observations.

Sverdrup (1947) showed that the location and strength of the gyres was linked to the curl of the surface windstress. The bottom panels of figure 3.6 show the mean annual windstress curl for the MOD run as well as for the ECMWF ocean reanalysis ORA-S3 dataset (Balmaseda et al., 2007). The simulated curl of the windstress in the MOD run is overestimated in the North Pacific. This could be due to the reduced latitudinal extension of the simulated westerlies. The discrepancies in the curl of the windstress can thus explain part of the biases observed in the simulated sea surface height. The coarse resolution of the model might also explain another part of the SSH biases observed.
Figure 3.6: Top panels: Mean annual sea surface height (m) as simulated over the final 20 years of the MOD run (left) and averaged for years 1992 to 2002 for the MDOT (right) (Maximenko and Niiler, 2004); Middle panels: Mean annual windstress curl ($10^{-7}$ m/s) calculated over the final 20 years of the MOD run (left) and averaged for years 1992 to 2002 for the ECMWF ORA-S3 (right) (Balmaseda et al., 2007); Bottom panels: Annually averaged Ekman pumping velocity ($10^{-7}$ m/s) over the final 20 years of the MOD run (left) and for the ECMWF ORA-S3 (right).
3.1. Pre-industrial and modern day runs

Ekman pumping

The bottom panels of figure 3.6 shows the annually averaged Ekman pumping velocity for the MOD run as well as the one calculated from the last 20 years of the ECMWF ocean reanalysis ORA-S3 (Balmaseda et al., 2007). The upwelling and downwelling zones and magnitude are generally represented realistically by LOVECLIM. However, due to the weak simulated trade winds (figure 3.2), the tropical upwelling is severely underestimated. And, as already mentioned, the simulated vertical velocity structure in the North Pacific exhibits significant differences with the ocean reanalysis data.

Ocean temperature and salinity

Figure 3.7 displays the annual mean Sea Surface Temperature (SST) and Sea Surface Salinity (SSS) averaged over the last 20 years of the MOD run (year 1980 to 2000 A.D.) on the left panels and the present day annual SST and SSS observations taken from Locarnini et al. (2006) on the right panels. Globally the model represents the SST distribution reasonably well. However, the simulated zonal mean temperature gradients in the tropical Pacific and Atlantic are not captured realistically. This can be partly attributed to an underestimation of the tropical air-sea coupling strength, owing mostly to the low atmospheric resolution, the weakness of the ageostrophic terms and a diffuse thermocline. Furthermore, we see relatively weak temperature fronts near the western boundary currents - another typical feature of coarse-resolution ocean models.

Due to a misrepresentation of the oceanic gyres, the simulated SSS field exhibits significant differences with the observed data in the subtropical gyres area. More importantly the simulated North Pacific is too salty compared to the observations. This could be due to the overestimation of the northward extension of the Kuroshio current.

Figure 3.8 shows the 15% sea ice concentration contour level computed by the model and averaged over the last 20 years of the MOD run for March and September. The months of March and September were chosen as they represent the months with respectively the minimum and maximum sea ice cover. The model results are in reasonable agreement with observations (Rayner et al., 2003). However, in the northern hemisphere, the model slightly overestimates the sea ice content in the Barents and Norwegian Seas in March. In September, the simulated sea ice extent is underestimated in the Labrador Sea. In the southern hemisphere, the sea ice extent is slightly overestimated in March in the eastern Weddell Sea as well as in the Bellingshausen and Amundsen Seas. In September, sea ice extends slightly too far north in the Bellingshausen and Amundsen Seas. Such biases are not untypical for coupled climate models, even for more comprehensive ones.
Figure 3.7: Top panels: Annual mean SST (°C) averaged over the last 20 years of the MOD run (left) and observed present day annual mean SST as described in Locarnini et al. (2006) (right). Bottom panels: Annual mean SSS (psu) averaged over the last 20 years of the MOD run (left) and observed present day mean annual SSS as described in Locarnini et al. (2006) (right).
3.1. Pre-industrial and modern day runs

Surface phosphate content distribution

The global phosphate concentration averages to 2.11 mmol/m$^3$, which is in good agreement with the climatological observed estimate of 2.17 mmol/m$^3$ as calculated from the Garcia et al. (2006b) dataset. The oceanic phosphate (PO$_4^{3-}$) content distribution in the euphotic zone, obtained during the last year of the MOD run (figure 3.9) is globally in agreement with the data compiled by Garcia et al. (2006b). The model simulates high phosphate content in the upwelling areas (figure 3.6, bottom panels) of the Eastern Equatorial Pacific and southern tropical Atlantic as well as in the Southern Ocean. The model however underestimates the amount of upwelled phosphate in the North Pacific and the Arabian Sea. The phosphate content in the Southern Ocean is also slightly underestimated, but this should not lead to significant bias in export production as phosphate is not the limiting nutrient in this region. Indeed, export production in the Southern Ocean is generally limited by the amount of light in the euphotic zone and the iron content (Martin, 1990), the latter not being simulated in LOVECLIM.

Export production

The simulated export production is shown in figure 3.10, top panel. As there is no global data set representing the global annual export production, the simulated export production is compared to the observed SeaWiFS chlorophyll pattern. The chlorophyll content generally represents the primary production. The bottom panel of figure 3.10 exhibits the satellite-derived chlorophyll concentration (mg/m$^3$) averaged for years 1997 to 2000 (Gregg and Conkright., 2002). The equatorial upwelling areas, such as the Eastern Equatorial Pacific and the Eastern Equatorial Atlantic,
as well as the subpolar Southern Ocean exhibit high export production, in general agreement with the observed primary production. The model also simulates moderate annual export production in the North Atlantic and North Pacific gyres. However, the simulated export production in the Arabian Sea is underestimated due to a weaker simulated upwelling in that region. In addition, due to the coarse resolution of the model, the coastal areas are not well represented. As coastal primary production represents about 25% of the global primary production, a misrepresentation of the coastal area can lead to an underestimation of the global export production. However, even state of the art global ocean biogeochemical models cannot reproduce processes occurring in coastal regions.

The global mean export production amounts to about 6.6 GtC/yr, which is in the lower range of the export production estimates. Observation-based estimates range from 5.3 GtC/yr (Louanchi and Najjar, 2000) to 11.1 GtC/yr (Laws et al., 2000). A large range of export production values is also obtained with global 3D ocean carbon cycle models. For example, the model of Howard et al. (2006) exhibits 4.6 GtC/yr of export production, Matthies et al. (2003) obtain 6.7 GtC/yr, Heinze et al. (2003) get 8.7 GtC/yr, Anderson and Sarmiento (1995) simulate 9.7 GtC/yr, Yamanaka and Tajika (1996) obtain 10 GtC/yr and even larger export production rates are obtained by Najjar et al. (2007). These values are not correlated with the biological model complexity but rather reflect the use of empirical parametrizations. Furthermore, the export production is very sensitive to the level of vertical mixing in the model (Gnanadesikan et al., 2002) as well as to the simulated ocean circulation characteristics (Najjar et al., 2007). By simply varying mixing parametrization in their model, Gnanadesikan et al. (2002) were able to change the export production...
3.1. Pre-industrial and modern day runs

Figure 3.10: Export production (gC/m²/yr) as simulated for MOD (left); Observed annually averaged chlorophyll concentration (mg/m³) for SeaWiFS 1997-2000 (Gregg and Conkright., 2002) (right).

from 9.0 GtC/yr to 22.4 GtC/yr.

**Partial pressure of CO₂ at the ocean surface**

Figure 3.11 displays the difference of partial pressure of CO₂ between the ocean and the atmosphere (ΔpCO₂) over the final years of the MOD run and the observed data compiled by Takahashi et al. (2002). In both, the model and the observations, the Southern ocean, the North Atlantic as well as the North Pacific are net sinks of CO₂, while the equatorial regions, particularly the equatorial Pacific, are net sources. However, in the MOD run, the North Atlantic and North Pacific sinks are underestimated. The model also exhibits a source of CO₂ to the atmosphere in the Atlantic and the Indian oceans at about 40°S. Those discrepancies could be due to differences between the simulated and observed sea surface temperature and salinity patterns (figures 3.15 and 3.9).

**Discussion**

Globally the ocean surface climatology and geochemical tracers distribution are reasonably well represented by LOVECLIM. However, some significant differences with the observations are seen in the North Pacific and in the Equatorial Pacific. Biases in the representation of the oceanic gyres, the Ekman pumping and thus in the geochemical tracer distributions are most likely the result of discrepancies in the simulated wind field and the coarse resolution of the model. Indeed, in the equatorial regions, due to relatively weak trades, the simulated upwelling is reduced compared to observations. Reduced upwelling of phosphate due to the weak wind field could explain why the simulated global export production is in the lower range of the export production estimates. It is important to keep these model biases in
3.1.3 Ocean climatology and marine carbon cycle - Part II: deep water processes

Convection and deep water formation

At high latitudes in winter, due to lower air temperatures the oceanic heat loss is enhanced, resulting in lower SST. Brine rejection, due to the formation of sea ice, also induces an increase in surface salinity. In some regions, the surface waters then become dense enough to sink to depth. The major convection areas in the North Atlantic are the Greenland Iceland and Norwegian (GIN) Sea and the Labrador Sea, which are also the main convection sites simulated by LOVECLIM (figure 3.12). In the GIN Sea the convection depth reaches 1100m in February, whereas in the Labrador Sea the convection depth attains values of about 700m. In our model no convection takes place south of 60°N. The convection depth obtained by LOVECLIM is shallower than the one reported by Killworth (1983). Killworth (1983) reports a convection depth of 1000m in winter in the Labrador Sea. However, a more precise comparison cannot be done as convection depths are poorly constrained from the observations.

Very dense Antarctic Bottom Water (AABW) forms along the Antarctic coast (Rintoul et al., 2001; Schlosser et al., 2001). Cooling of the surface waters of the Southern Ocean lead to sea ice formation. The very cold and very saline water thus formed follow the Antarctic continental shelf and sink into the abyssal ocean. In the southern hemisphere, LOVECLIM simulates coastal convection due to brine rejection in the Ross Sea, the Amundsen Sea as well as in the region 0-60°E close to the Antarctic coast. In reality, convection should also take place in the Weddell
3.1. Pre-industrial and modern day runs

Sea, whereas the model only simulates weak convection on its northeastern flank. In general, the simulated convection in the Southern Ocean does not exceed 450m, which is too shallow.

Figure 3.13 shows the zonal mean vertical stream function (Sv) for the global ocean, the Atlantic and the Pacific basins for the MOD run and for the pre-industrial run of the HadCM model (the UK Met Office HadCM3M2 model, (Gordon et al., 2000)). The HadCM3 model is a coupled model that has been used for numerous studies on North Atlantic climate (Vellinga and Wood, 2002). In agreement with the HadCM model, LOVECLIM simulates a strong northern thermohaline cell (≈ 20 Sv) and more vigorous AABW formation (15–20 Sv). In both models the separation between the two water masses is simulated at about 2500m. The AABW formation and circulation in LOVECLIM are reasonably well reproduced (Campin and Goosse, 1999; Talley et al., 2003) and are comparable to other more complex CGCMs. Because of the weaker simulated trades in LOVECLIM, the wind-driven subtropical cells are also weaker than in HadCM3.

The simulated maximum transport of the Atlantic overturning circulation below 500m amounts to about 26 Sv in the Atlantic, which is somewhat higher than recent observational estimates of 18±5 Sv by Talley et al. (2003). Compared to the HadCM model, the maximum stream function in the North Atlantic is deeper in LOVECLIM by about 500m. The latitudinal location of the maximum value in the North Atlantic is between 40°N and 60°N for both models, in agreement with Talley et al. (2003).

In the Pacific basin, for both models, the AABW fills most of the basin and has a maximum stream function of around 10 Sv. In addition, the models simulate a sluggish intermediate and deep water circulation in the North Pacific.

The simulated stream function in LOVECLIM is thus comparable to the one simulated by a coupled general circulation model of higher resolution.

Latitude-depth temperature and salinity distributions

The latitude-depth ocean temperature and salinity distributions for each basin as simulated over the final 20 years of the MOD run and for present day observations (Antonov et al., 2006; Locarnini et al., 2006) are shown respectively in figures 3.14 and 3.15. The simulated deep waters in the southern parts of the Atlantic and Pacific basins are colder than in the observations, while they are warmer in the Indian basin. Moreover, as can be seen on figure 3.15, the modeled deep waters in the southern part of the Atlantic basin are also slightly fresher than in the observations. The northern part of the Indian basin is poorly represented by the model. The main reason could be the lack of a realistic representation of the monsoon system. The latitude-depth temperature profile thus shows a weaker simulated downward mixing north of the equator in the Indian basin, that could be due to an underestimation of the wind strength. In addition, the simulated salinity in that area is significantly
Figure 3.12: Convection depth (m) at high northern latitudes in February (top) and at high southern latitudes in August (bottom) as simulated over the final 20 years of MOD.
Figure 3.13: Zonal mean vertical stream function (Sv) for the global ocean (top), the Atlantic basin (middle) and the Pacific basin (bottom) as simulated over the final 20 years of the MOD run (left) and the pre-industrial run of the HadCM3M2 model (right).
underestimated.

As was already observed in figure 3.7, the surface salinity in the North Pacific is overestimated. The lack of a halocline in that region leads to an overestimation of the ventilation of the North Pacific waters.

**Phosphate content distribution**

Figure 3.16 displays the simulated zonally averaged phosphate distribution as a function of depth for each basin (left panels) compared to the observed one (right panels) (Garcia et al., 2006b). The phosphate distribution in the ocean depends on marine primary productivity, remineralization of organic matter and the oceanic circulation. Due to restrictions in the model tuning, the total phosphate inventory is lower (4%) than the observed phosphate inventory. However, the simulated phosphate content distribution is globally in good qualitative agreement with the observed data. As phosphate is the limiting nutrient for primary productivity in many areas, the surface phosphate concentration is low in all the basins. The phosphate content then increases with remineralization of organic matter below the euphotic zone. Waters also get enriched in phosphate along the advective path of the conveyor belt. In the Atlantic, the phosphate content displays a maximum centered at about 700m depth near the equator due to local remineralization in a high productivity zone and transport by the intermediate waters. At depth, the phosphate content increases from north to south due to remineralization along the southward path of North Atlantic Deep Water. In the Pacific and Indian Oceans, the phosphate concentration increases with depth as well as from south to north due to remineralization. However, the simulated phosphate content in the intermediate waters of the North Atlantic and North Pacific is respectively 0.2 $\mu$mol/l and 1$\mu$mol/l lower than the observations. The biases observed in the phosphate content of the North Pacific could be the result of a stronger ventilation due to the lack of halocline in that region.

**Dissolved oxygen**

The simulated oxygen distribution, which depends on the gas exchange with the atmosphere as well as productivity and remineralization is displayed in figure 3.17. The oxygen content is generally high at the surface due to air-sea gas exchange with the atmosphere as well as photosynthesis. Oxygen is then consumed during the remineralization of organic matter. The oxygen content decreases from the Atlantic to the Pacific due to the aging of the waters. The areas of minimum oxygen content therefore correspond to areas of maximum phosphate content due to remineralization. However, the same discrepancies with the observations as the ones reported for the phosphate content apply: the oxygen content in the deep Atlantic is underestimated, the oxygen minimum in the intermediate waters of the Pacific is centered at the equator instead of being in the northern part of the basin.
Figure 3.14: Temperature (°C) averaged over the final 20 years of the MOD run (left panels) and for the Locarnini et al. (2006) observations (right panels). The top, middle and lower panels respectively represent the temperature averaged over the Atlantic, the Pacific and the Indian basins.
Figure 3.15: Salinity (psu) averaged over the final 20 years of the MOD run (left panels) and for the Antonov et al. (2006) observations (right panels). The top, middle and lower panels respectively represent the salinity averaged over the Atlantic, the Pacific and the Indian basins.
3.1. Pre-industrial and modern day runs

Figure 3.16: Zonal mean phosphate ($\text{PO}_4^{3-}$, $\mu$mol/l) distribution as a function of depth in the Atlantic (upper panels), Pacific (middle panels) and Indian basins (lower panels) over the final year of the MOD run (left). Observed phosphate content for the Levitus data (right) (Garcia et al., 2006b).
and finally the simulated oxygen content is underestimated in the southern deep part of the Indian basin.

**Apparent Oxygen Utilization, AOU**

Figure 3.18 shows the AOU distribution as a function of depth and latitude in the Atlantic and Pacific basins as compared with the Locarnini et al. (2006) observations. AOU is the difference between the temperature dependent oxygen saturation ($O_{2sat}$) and the in situ $O_2$ concentration ($AOU = O_{2sat} - O_2$). $O_{2sat}$ is calculated from the temperature and salinity data using the Weiss (1970) equation.

At the ocean surface, the oxygen content of the water should be close to $O_{2sat}$. However, as oxygen is consumed during remineralization, the oxygen content generally decreases with the aging of the water. AOU represents the oxygen utilized in the oxidation of organic matter and thus increases along with remineralization.

The general structure of the simulated AOU is in agreement with the observations. However, the simulated AOU is slightly overestimated in the southern and deep Atlantic basin. In the Pacific basin, AOU is also greater than in the observations below 1000m and south of 20°N, while it is underestimated in the northern part of the basin in the first 3000m.

**Dissolved Inorganic Carbon, DIC**

Figure 3.19 shows the DIC distribution as a function of depth in the Atlantic, Pacific and Indian basins for MOD on the left panels and the observed GLODAP data (Key et al., 2004) on the right panels. The DIC content is affected by different processes: air-sea CO$_2$ exchange, productivity, remineralization as well as CaCO$_3$ precipitation and dissolution. Photosynthesis and CaCO$_3$ precipitation lower the DIC content, whereas remineralization of organic matter and CaCO$_3$ dissolution increase the DIC content. The simulated and observed DIC distributions are very similar in all the basins. In the Atlantic, the surface DIC is low due to primary productivity and CaCO$_3$ precipitation. Remineralization of organic matter leads to a subsurface maximum near the equator. At depth, the DIC content increases from north to south due to remineralization of organic matter along the pathway of North Atlantic Deep Water. However, the modeled DIC content is overestimated in the deep North Atlantic. In the Pacific, the surface DIC content is low and increases with depth as well as from south to north. However, the simulated maximum DIC values are near the equator in the model rather than in the northern part of the Pacific.

**Alkalinity**

The alkalinity distribution is strongly dependent on CaCO$_3$ precipitation and dissolution. The range of simulated alkalinity values is comparable to the observational
Figure 3.17: Oxygen (O₂, ml/l) repartition as a function of depth in the Atlantic (upper panels), Pacific (middle panels) and Indian basins (lower panels) over the final year of the MOD run (left) and the observed data (Garcia et al., 2006a) (right).
Figure 3.18: Zonally averaged Apparent Oxygen Utilization (AOU, ml/l) distribution as a function of depth in the Atlantic (upper panels) and Pacific basins (lower panels) over the final year of the MOD run (left) and the observed data (right) (Locarnini et al., 2006).
3.1. Pre-industrial and modern day runs

Figure 3.19: Zonal mean DIC (µmol/l) distribution as a function of depth in the Atlantic (upper panels), Pacific (middle panels) and Indian basins (lower panels) over the final year of the MOD run (left) and for the observed GLODAP data (right) (Key et al., 2004).
data, but the spatial structure is quite different, in particular in the Atlantic. In the
Atlantic basin, the model overestimates the alkalinity content below 1000 m and un-
derestimates it at the surface in the tropics and subtropics (figure 3.20). The larger
alkalinity values in the deep North Atlantic could be the result of greater transport
of high alkaline waters from the Southern Ocean or of greater carbonate dissolution.
In the Pacific and Indian basins, the range and structure of the simulated and ob-
served alkalinity fields are comparable, except for higher values of about 20 to 40
µmol/l in the deep Pacific.

Discussion
Temperature, salinity and biochemical tracers show similar deviations from the ob-
served distributions. A part of the discrepancies between the model and the obser-
vations is certainly the result of differences in the ocean circulation. Specifically, it
seems that the northward extension of the AABW in the Atlantic is overestimated
while the one in the Pacific and Indian basins is underestimated. The northward
extension of the Antarctic Intermediate Water (AAIW) in the Pacific basin is also
underestimated.
The simulated vertical velocity anomalies in the North and tropical Pacific could
be partly responsible for the weaker advection of tracers in the Pacific and Indian
basins. The overestimation of the ventilation in the North Pacific, due to the lack of
a halocline there, is also certainly partly responsible for the observed tracer biases.
In addition, the lower simulated phosphate content in the southern parts of
the basin can simply be the result of the slightly lower total phosphate inventory.
Another explanation for the observed discrepancies in the Southern Ocean could be
an overestimation of the biological pump in that area. This could be due to the lack
of iron limitation in the model or to an underestimation of the light constraint on
export production.
These variations in deep and intermediate nutrient contents arise in order to
obtain a good representation of the surface processes. It has indeed been shown
above that the surface/euphotic nutrient distributions were in good agreement with
the observations. The 3D representation of the nutrient distribution is far from
perfect but should be suitable for global scale experiments. It is important to keep
in mind these caveats when interpreting the paleo-modeling experiments.

3.2 Last Glacial Maximum run, LGM

3.2.1 Experimental set-up
To obtain a quasi-equilibrium state for LGM conditions (simulation LGM), the
coupled ECBilt-CLIO-LOCH was forced for 2200 years with LGM orbital param-
ters, glacial topography (Peltier, 1994), albedo pattern, forest fraction (Crowley and
Figure 3.20: Zonal mean total alkalinity (µmol/l) distribution as a function of depth in the Atlantic (upper panels), Pacific (middle panels) and Indian basins (lower panels) over the final year of the MOD run (left) and the observed GLODAP data (right) (Key et al., 2004).
Baum, 1997) as well as with the LGM atmospheric CO$_2$ concentration (192 ppmv) as recorded in the Taylor Dome ice core (Indermühle et al., 2000). The atmospheric concentrations of CH$_4$ and N$_2$O were prescribed to LGM values, according to ice-core data (Brook et al., 2000; Sowers et al., 2003). The LGM boundary forcing adopted here is described in detail in Timmermann et al. (2005) and Justino et al. (2005). After this 2200-year long forced simulation the VECODE model was activated and the coupled model ECBilt-CLIO-VECODE-LOCH was forced for an additional 1300 years with glacial greenhouse gas concentrations. During this spin-up, the model still used prescribed LGM orbital parameters and glacial topography, as described above. However, albedo and forest fraction were computed directly from VECODE. Thereafter, the atmospheric CO$_2$ was directly computed from the carbon balance between the atmosphere, the vegetation and the ocean. Another fully coupled 2000 years long simulation that was based on the previous forced model run, was conducted using interactive CO$_2$. The river inputs of DIC, alkalinity, organic matter and silicic acid were kept constant at pre-industrial levels. As will be documented below this iterative set-up generated a quasi-steady-state LGM simulation with relatively stable CO$_2$ conditions at 198 ppmv and a weak trend of 3 ppmv/1000 years (figure 3.21). For the short freshwater perturbation experiments analyzed in this study, we anticipate that this weak CO$_2$ trend will not affect our main conclusions.
3.2. Last Glacial Maximum run, LGM

3.2.2 Performance of the climate model

A more detailed review of the LGM climate state simulated in different versions of ECBilt-Clio and ECBilt-Clio-VECODE can be found in Timmermann et al. (2004); Justino et al. (2005) and Roche et al. (2007)\(^1\), respectively.

**Air temperature**

The simulated LGM climate state is characterized by a global surface air cooling of 4.2°C, with respect to the pre-industrial state. The largest glacial temperature anomalies are found near and downstream of the Laurentide and Eurasian ice sheets (figure 3.22). Simulated glacial surface air temperatures drop by 15 to 30°C in the northern extratropics. The equatorward extension of the sea ice in the southern hemisphere leads to a decrease of surface temperatures by about 6 to 15°C. The tropical regions, however, experience only a moderate annual surface cooling of 1-3°C, mostly due to the atmospheric CO\(_2\) reduction and due to remote effects from the glacial ice sheets (Timmermann et al., 2004).

As a result of the greater extent of the glacial ice sheets in the northern hemisphere and the sea ice in both hemispheres, more radiation is emitted back to space at mid-latitudes. The glacial desertification in North Africa and the Middle-East also contributes to further cooling of these areas (figure 3.22, bottom panel.)

\(^1\)In comparison to the LGM simulation conducted by Roche et al. (2007), the model here is forced with the ICE4G palaeotopography, instead of the ICE5G reconstruction. Furthermore, the sensitivity of the longwave radiation to CO\(_2\) changes is enhanced and sea-level changes, changes of the river routing, Antarctic iceberg calving and atmospheric dust loading are neglected.
Precipitation

Both the presence of large orographic barriers and the associated modification of the atmospheric eddy momentum fluxes (Justino et al., 2005), as well as the intensified LGM SST gradient lead to an intensification of the trade-winds (figure 3.24) with implications for the subtropical cell transports and evaporative cooling in the tropical regions. The general cooling of Earth’s atmosphere under glacial conditions leads to a reduced capacity of the atmosphere to hold moisture. This causes a drop of global precipitation by 11 cm/yr (10%) during the LGM (figure 3.24). In particular dry areas such as the southern border of the Sahara, the middle- and far-East, the Tibetan plateau as well as central Greenland experience a drop of annual mean precipitation of about 50%. Other areas such as the southern border of the Laurentide ice sheet experience a regional increase in precipitation, which is suggestive of a positive feedback that may have helped to maintain glacial ice sheets.

Ocean temperature and circulation

Figure 3.25 displays the LGM SST compared to the latest glacial SST reconstructions, from the GLAMAP (Gersonde et al., 2003; Pflaumann et al., 2003) and MARGO projects (Kucera et al., 2005a). SST anomalies in the North Atlantic attain values of up to 8°C (figure 3.22), which is in good agreement with the glacial SST reconstructions (Weinelt et al., 2003; Pflaumann et al., 2003). SST differences around Antarctica compare well with LGM reconstructions of Gersonde et al. (2005). Simulated tropical SSTs are 1 to 3°C lower than for the pre-industrial state. This also compares favorably with paleo-reconstructions of Kucera et al. (2005b), Pflaumann et al. (2003) and Koutavas et al. (2002), but is at odds with other tropical SST estimates (Guilderson et al., 2001; Lea et al., 2003; Nurnberg et al., 2000; Visser et al., 2003; Kienast et al., 2001; Rühlemann et al., 1999; Rosenthal et al., 2003).

It should be noted that some SST paleo-reconstructions, such as the ones based on
Figure 3.24: Annual mean zonally averaged zonal windstress component for the PIN run (solid line) and the LGM run (dashed line) (left); Annual mean precipitation anomalies (cm/yr) for LGM–PIN (right).

Mg/Ca and alkenone, could represent a particular season, whereas the simulated annual mean SST is presented here.

A direct comparison between modeled LGM ocean states and LGM reconstructions (McManus et al., 2004; Gherardi et al., 2005) is not straightforward. This task is rendered difficult by the transient nature of the LGM. As noted in Timmermann et al. (2004), LGM simulations are designed as steady-state experiments, whereas the observed LGM state occurred about 2,000 years after the long-lasting glacial meltwater pulse Heinrich II. It is unclear whether the real AMOC was still recovering from this perturbation, or whether it had already reached a new equilibrium state which could be directly compared with steady-state paleo-model simulations. The simulated vigorous glacial Atlantic overturning with an amplitude of about 26 Sv compares well with the simulation of Roche et al. (2007) and exhibits also a southward displacement of the deep-water formation areas. Roche et al. (2007) argue that this finding is not inconsistent with available proxy data, although a meaningful comparison may be hampered by the fact that the observed LGM state was probably not an equilibrium state.

As shown in figure 3.26, the glacial ocean is globally colder in the Atlantic and Pacific basins compared to the pre-industrial one. Due to intense brine rejection during sea ice formation, the SSS is greater during the LGM compared to present in the Southern Ocean and the North Pacific (figure 3.26). In the North Atlantic, the SSS is however lower due to increased water runoff from land. This greater runoff also leads to negative salinity anomalies in most of the Atlantic basin. Negative salinity anomalies in the Atlantic south of 60°N are also obtained by LGM simulations performed with the HadCM model (the UK Met Office HadCM3M2 model, (Gordon et al., 2000), PMIP2 project) (figure 3.27) as well as with the NCAR-CCSM model (Liu et al., 2005). In LOVECLIM however, the simulated SSS anomaly north of 60°N in the Atlantic basin is negative, which is at odds with the previously cited
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Figure 3.25: SST (°C) for the last 20 years of the LGM run and paleoproxy SST estimates for the LGM (stars) (Kucera et al., 2005a).

LGM simulations. In the Southern Ocean, however, due to the greater sea ice formation, the AABW formed during the LGM is saltier than during pre-industrial times, in agreement with the paleodata obtained by Adkins et al. (2002).

**Performances of the carbon cycle in the fully coupled LGM run**

The simulated LGM atmospheric CO$_2$ is in quasi-equilibrium at 198 ppmv. The mean vegetation and ocean carbon stocks in the fully coupled simulations are 1,370 GtC and 40,480 GtC, respectively. Recent reconstructions of the LGM terrestrial carbon stock were based on $\delta^{13}$C values of calcite shells of benthic foraminifera (Duplessy et al., 1988; Curry et al., 1988; Bird et al., 1994) as well as on pollen analysis (Crowley, 1995; Adams and Faure, 1998). The estimated values for the terrestrial carbon stock during the LGM range from 600 to 1,700 GtC. The simulated terrestrial carbon reservoir is in the range of these storage estimates and is also in agreement with a numerical simulation performed with the LPJ-DGVM forced with simulated LGM climate conditions (Joos et al., 2004).

**Terrestrial biome distribution**

The gross biome distribution obtained for the LGM run (figure 3.28) can be directly compared to the reconstructions compiled by Crowley (1995). Due to the massive ice sheet covering Northern Eurasia and North America during glacial times, ice (desert) has replaced the tundra, taiga or forest present in these areas. In addition,
3.2. Last Glacial Maximum run, LGM

Figure 3.26: Temperature anomalies (°C) for the last 20 years of the LGM run compared to the last 20 years of the PIN run zonally averaged over the Atlantic basin (upper left) and Pacific basin (lower left); Salinity anomalies (psu) for LGM–PIN zonally averaged over the Atlantic basin (upper right) and Pacific basin (lower right).

Figure 3.27: Salinity anomalies (psu) for LGM–PIN zonally averaged over the Atlantic basin for the model HADCM3
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Figure 3.28: Biome distribution (from purple to dark red: Tropical Rain forest, warm forest, mixed forest, conifers, savanna, grass, cold grass and desert) for the last 20 years of the LGM run (left) as well as Crowley (1995) synthesis for the LGM state (right). For comparison, the biome distribution for the PIN run are depicted in figure 3.5.

desert covers most of North Africa, the Middle East and Central Asia during the LGM at the expense of grass. These changes in vegetation cover mainly reflect the drier conditions there, in agreement with Crowley (1995). In Patagonia, due to the colder and drier conditions simulated during the LGM, grass is the dominant biome there, as also suggested by paleoreconstructions. On the other hand forest is still the dominant vegetation type in western Europe whereas Crowley (1995) suggest that grass covered this area. Similarly, in Amazonia, tropical rain forest is still thriving in the LGM simulation, in contradiction with Crowley (1995) reconstructions. Some of the discrepancies between reconstruction and model results can be explained in terms of the mean precipitation biases of our coarse-resolution atmosphere. Due to the uneven distribution of the paleo-data used for the reconstruction, some of the differences observed could be also due to an erroneous paleo-reconstruction.

Phosphate content and export production distributions

Under LGM conditions, the simulated southern hemispheric westerlies are stronger (+10%, figure 3.24), which enhances the upwelling in the Southern Ocean. The phosphate and silicate content are therefore respectively about 0.2 $\mu$mol/l (+16%) and 20 $\mu$mol/l (+44%) higher under LGM than pre-industrial conditions in this area (figures 3.29 and 3.30). Additionally, the glacial increase in sea ice coverage leads to a decrease in marine productivity of about 25%. This result is in agreement with
3.2. Last Glacial Maximum run, LGM

Figure 3.29: Annual mean phosphate content anomalies ($\mu$mol/l) averaged over the euphotic zone (left) and marine export production anomalies (gC/m$^2$/yr) averaged over the euphotic zone for LGM–PIN (right). The solid line represents the annual mean sea ice contour at 0.1m.

LGM nutrient concentrations in the Antarctic surface ocean as deduced from Cd/Ca ratios in planktonic foraminifera (Elderfield and Rickaby, 2000), which shows lower phosphate utilization in the Southern Ocean during glacial times. A reduced glacial export production is also in agreement with the production estimates described by Kohfeld et al. (2005). Similarly, mainly due to sea ice advance, the export production is reduced by 32% in the North Atlantic and 22% in the North Pacific. The lower nutrient utilization leads to a greater phosphate and silicate content in the North Atlantic and in the Southern Ocean.

Due to the intensified trade wind circulation and stronger upwelling in the Eastern Equatorial Pacific and in the Eastern Equatorial Atlantic, more nutrients are brought to the surface, which enhances the marine export production by about 15% (figure 3.29). This is consistent with a compilation of export production as recorded in a large number of sediment cores (Kohfeld et al., 2005), which suggests that during the LGM, the export production may have been equal to or higher than today outside the non polar areas.

Our LGM model results are qualitatively in good agreement with these paleo-reconstructions and with the Peacock et al. (2006) synthesis:

1. Simulated LGM phosphate values are somewhat larger than for the pre-industrial climate, in particular at high latitudes. Anomalies at low latitudes if not weakly positive, are close to zero.

2. The simulated LGM export production is lower in the polar oceans and of the same order or larger in areas comprised between 50°S and 50°N.
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Figure 3.30: Annual mean silicate anomalies ($\mu$mol/l) averaged over the euphotic zone for LGM–PIN

Figure 3.31: Annual mean zonally integrated CO$_2$ flux to the atmosphere (TmolC/yr) for PIN (black) and LGM (red).

Ocean partial pressure of CO$_2$

During the LGM, the high latitude CO$_2$ sinks as well as the low latitude sources are amplified compared to the PIN run. Indeed, the CO$_2$ solubility is enhanced at high latitudes due to colder conditions, while the stronger upwelling at low latitudes lead to a greater CO$_2$ outgassing to the atmosphere (figure 3.31). The globally stronger winds are also partly responsible for the overall enhanced magnitude of the air-sea CO$_2$ flux.

Deep ocean pH and CaCO$_3$ sedimentation

As can be seen in figure 3.32, under LGM conditions, the pH value at a depth of 3000m increases by about 0.1 units in the Atlantic ocean, by about 0.25 units in the
3.2. Last Glacial Maximum run, LGM

Southern Ocean and the South Pacific, and by about 0.1 units in the North Pacific.

The $\text{CO}_3^{2-}$ content anomalies at a depth of 3000m is shown in figure 3.32, right panel. The glacial $\text{CO}_3^{2-}$ concentration is about 30 $\mu$mol/l higher than the pre-industrial one in the Eastern Equatorial Pacific and 40 $\mu$mol/l higher in the Atlantic basin between 20°S and 40°N. In the Southern Ocean, the $\text{CO}_3^{2-}$ content is about 55 $\mu$mol/l higher during glacial than during interglacial times. The pH and $\text{CO}_3^{2-}$ content exhibit the greatest anomalies in the Southern Ocean and in particular on the Pacific side. This could be due to the different deep ocean circulation prevailing in the LGM run compared to the PIN run. Indeed, during the LGM run, most of the AABW flows into the Indo-Pacific, while the Atlantic basin is dominated by the NADW.

In the deep equatorial Pacific, the simulated $\text{CO}_3^{2-}$ content anomalies compare relatively well with the 30 $\mu$mol/l rise in deep $\text{CO}_3^{2-}$ concentration estimated by Marchitto et al. (2005) for the last deglaciation. The simulated glacial deep $\text{CO}_3^{2-}$ content is however slightly higher than the one suggested by recent boron isotope reconstructions (Hoenisch et al., 2008). Indeed, Hoenisch et al. (2008) results suggest a glacial pH increase of about 0.1 units on the Walvis ridge (South Atlantic, 21°S-30°S, 7°E), which corresponds to a deep $\text{CO}_3^{2-}$ content increase of about 20 $\mu$mol/l. In addition, these modelling results are at odds with Broecker and Clark (2003) study, which suggest a lower content of $\text{CO}_3^{2-}$ in the deep glacial tropical Pacific and Atlantic.

Given the uncertainty associated with changes in ocean circulation during glacial times (Wunsch, 2003) and uncertainties in river inputs of alkalinity and DIC during the LGM, the pH and $\text{CO}_3^{2-}$ anomaly patterns presented here are highly uncertain.
3.3 Summary

The analyses above have shown that even with a reduced complexity atmospheric model, LOVECLIM is able to reproduce basic climatic features satisfactorily. It should be kept in mind that LOVECLIM is an EMIC that makes compromises in terms of neglecting certain physical processes. These compromises on the other hand pay off in terms of computing efficiency. In addition, it allowed me to conduct a suite of multi-millennial long simulations, which would not have been possible with a CGCM.

Nevertheless, the simplified atmospheric model leads to biases in the representation of the surface winds, which in turn affect the oceanic circulation and the advection of marine tracers. Significant discrepancies between the model and the observations are thus obtained in the North Pacific and the Northwestern Indian Ocean. The northern part of the Indian Ocean is poorly represented in the simulations due to the lack of a realistic monsoon system. Given the complexity and resolution of the model, the present-day climate, terrestrial biosphere and marine carbon cycle are represented in a satisfying way. In fact, on a zonally averaged basis, some of the regional differences between model and observations cancel each other out. The model is suitable to study large scale climatic and geochemical changes on millennial timescales. Moreover, even if the mean climate and carbon cycle reference states are slightly different from the observations, the changes in variables obtained in the following experiments should be robust.

The LGM equilibrium state simulation ignores the changes in sea level as well as the associated changes in salinity and nutrient concentrations. The LGM state also does not include iron fertilization effects. How mean state biases may affect the realism of the model sensitivities will be discussed in each of the following three chapters.

Bibliography


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Chapter 4

Climate and marine carbon cycle response to changes in the strength of the southern hemispheric westerlies.

It has been previously suggested that changes in the strength and position of the southern hemisphere westerlies could be a key contributor to glacial-interglacial atmospheric CO$_2$ variations. To test this hypothesis, we perform a series of sensitivity experiments using an earth system model of intermediate complexity. A strengthening of the climatological mean surface winds over the Southern Ocean induces stronger upwelling and increases the formation of Antarctic Bottom Water. Enhanced Ekman pumping brings more Dissolved Inorganic Carbon (DIC)-rich waters to the surface. However, the stronger upwelling also supplies more nutrients to the surface, thereby enhancing marine export production in the southern hemisphere and decreasing the DIC content in the euphotic zone. The net response is a small atmospheric CO$_2$ increase (∼5 ppmv) compared to the full glacial-interglacial CO$_2$ amplitude of ∼90 ppmv. Roughly the opposite results are obtained for a weakening of the southern hemisphere westerly winds.

4.1 Introduction

Atmospheric CO$_2$ concentrations recorded in ice cores (Petit et al., 1999) co-varied with temperature and ice-volume changes on orbital timescales. While glacial-interglacial CO$_2$ changes of 80–90 ppmv certainly affected global climate via the greenhouse effect, the origin of these CO$_2$ variations still remain elusive. Colder and drier climate conditions during glacial times were associated with a reduced terrestrial carbon stock. Hence, to account for the glacial CO$_2$ drop, large amounts of carbon must have been sequestered in the ocean. Different mechanisms have
been suggested such as increased CO$_2$ solubility, greater Southern Ocean stratification and sea ice cover (Francois et al., 1997; Stephens and Keeling, 2000; Gildor et al., 2002), deepening of the lysocline (Archer et al., 2000) and increased marine export production due to either greater marine nutrient inventory (Broecker, 1982; McElroy, 1983; Broecker, 1998), or a higher Redfield ratio (Broecker, 1982; Omta et al., 2006), or an increase in iron availability in surface waters (Martin, 1990). So far, however, model results and theoretical considerations (Archer et al., 2003) suggest that there does not seem to be a single mechanism that can explain the full magnitude of the glacial-interglacial atmospheric CO$_2$ changes as well as their timing.

Recently, Toggweiler et al. (2006) (herein referred to as T06) proposed that an overall weakening of the southern hemispheric westerlies during glacial times could have led to a substantial draw-down of atmospheric CO$_2$. The reasoning proposed in T06 is the following: Presently, strong southern hemispheric westerlies generate a northward Ekman transport that leads to upwelling of nutrient and CO$_2$ rich-waters. A potential equatorward shift or weakening of the westerlies during glacial periods could have substantially reduced the supply of carbon-rich deep waters to the surface, leading to an enhanced uptake of carbon in the Southern Ocean. An atmospheric CO$_2$ reduction leads to global cooling, and presumably an extension of the southern hemispheric sea ice belt around Antarctica. Not only could this substantially reduce the air-sea fluxes of CO$_2$ (Stephens and Keeling, 2000), but also move the westerlies farther equatorward. T06 estimated that this positive feedback could explain an atmospheric CO$_2$ reduction of about 35 ppmv. It should however be noted that the model employed in T06 does neither capture sea-ice dynamics properly, nor does it consider a primary production consistent with the modified circulation. Indeed the primary production in T06 is evaluated after restoring the surface phosphate content to the present-day observed values. Firstly, we will show that paleo-proxy data as well as the PMIP2 coupled model simulations are at odds with the notion of a significant equatorward shift of the southern hemispheric westerlies. Secondly, our study will demonstrate that variable phosphate concentrations would alter the T06 result substantially.

As approximately 40% of the anthropogenic CO$_2$ sequestration in the oceans occurs in the Southern Ocean (Sabine et al., 2004; Mikaloff-Fletcher et al., 2006), understanding the impacts of southern hemispheric wind changes on this CO$_2$ sink at present and for future projections is of great importance. Some recent modeling studies suggest that the strengthening and poleward shift of the southern hemispheric westerlies observed over the last 30 years (Thompson and Solomon, 2002) may have weakened the CO$_2$ sink in the Southern Ocean (LeQuéré et al., 2007; Lovenduski et al., 2007). However, these results have been recently challenged by Law et al. (2008) and Zickfeld et al. (2008).

Our study is organized as follows: The intermediate model set-up for the wind-sensitivity experiments is described in section 2. In section 3 several paleo-proxy
records that are sensitive to southern hemispheric westerly winds will be discussed. The section also reviews the glacial wind-response in a series of coupled model simulations for the Last Glacial Maximum (LGM, 21 ka B.P.). The results from the wind-sensitivity experiments will then be described in section 4. The T06 hypothesis will be tested for enhanced and reduced westerlies as well as for a fully prognostic phosphate concentration and a climatologically prescribed one. The paper concludes with a summary of our main results and a discussion.

4.2 Model and experimental setup

The model used in this study is the earth system model of intermediate complexity, LOVECLIM (See Chapter 2), which is based on a somewhat simplified atmosphere model, an ocean general circulation model, a dynamic-thermodynamic sea-ice model, and oceanic as well as terrestrial carbon cycle components. For the sake of simplicity, the model experiments presented here employ a stationary pre-industrial vegetation pattern.

The pre-industrial steady state (PIN) was obtained by forcing LOVECLIM with 278 ppmv of atmospheric CO$_2$ during 500 years, then allowing the atmospheric CO$_2$ to vary freely during 700 years. The model was then run for 500 years with fixed pre-industrial vegetation. All sensitivity experiments described in this paper start from this initial state.

In experiment WINDP (WINDM), the zonal and meridional 10m wind velocities between 60°S-40°S were increased (decreased) by about 15%. The wind modifications were applied in the atmosphere-ocean-sea-ice coupling routine. Not only is the windstress forcing of the ocean modified but also the latent and sensible heat fluxes, in contrast to the model setup of T06. To better understand the effects of changes of marine primary production on the atmospheric CO$_2$, we also repeated WINDM using a climatologically prescribed phosphate field, WINDMCP. WINDP, WINDM and WINDMCP were run for 1,000 years.

4.3 Paleo-proxies and modeling studies of the southern hemisphere westerlies during the LGM

Several recent pollen studies from Chile suggest a wetter climate west of the Andes during the LGM than at present (Moreno et al., 1999; Maldonado et al., 2005; Valero-Garcés et al., 2005). Corroborating evidence comes also from marine sediments cores from the Chilean margin (Lamy et al., 1999; Stuut and Lamy, 2004). The possible mechanism to explain the greater precipitation is an equatorward shift and/or an intensification of the southern hemisphere westerly winds and of the stormtrack (Garreau, 2007). However, these recent studies are in disagreement
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with other paleostudies: Reconstructed drier LGM conditions in Southern America (Markgraf et al., 1992) have been explained in terms of weaker westerlies and a weaker storm track. Trying to reconcile these conflicting paleo-data, Shulmeister et al. (2004) analysis indicates that southern hemispheric westerlies were stronger during the LGM than for interglacial conditions.

Ice cores from the Antarctic Peninsula suggest that the flux of dust originating from Patagonia was much larger during the LGM than at present (Basile et al., 1997; Petit et al., 1999; Delmonte et al., 2002). The main mechanisms which can explain the enhanced dust flux are: (1) Increased westerly winds; (2) Drier conditions in the source region, which would be compatible with the hypothesis of weaker westerly stormtracks. Therefore, dust records from Antarctic ice cores do not provide unambiguous information on the strength of the southern hemisphere westerlies.

It is fair to say that the existing paleo-proxy data do not allow for a very accurate reconstruction, neither of the strength, nor of the position of the mean glacial southern hemispheric westerlies.

To test the conjecture of shifted or reduced glacial westerlies (T06) we analyze pre-industrial and LGM Coupled General Circulation Model (CGCM) simulations that were compiled as part of the Paleoclimate Modeling Intercomparison Project (PMIP2). All PMIP2 LGM CGCMs employ the same forcing: the ICE-5G ice sheet reconstruction for the period 21 ka B.P. (Peltier, 2004), prescribed glacial greenhouse gas concentrations, lowered sea level and insolation anomalies due to variations in the Earth’s orbit (Braconnot et al., 2007).

Our analysis focuses on the pre-industrial and LGM simulations that were conducted using a fixed continental vegetation cover (Braconnot et al., 2007). We analyze the results from four PMIP2 CGCMs: the HadCM model (the UK Met Office HadCM3M2 model, (Gordon et al., 2000)), the CCSM3 model (National Center for Atmospheric Research CCSM3 model, (Otto-Bliesner et al., 2006)), the IPSL model (Institut Pierre Simon Laplace, IPSL-CM4-V1-MR model, (Marti et al., 2005)) and the MIROC model (CCSR/NIES/FRCGC/MIROC3.2.2 medres model, (K-1-Model-Developers, 2004)). The simulated wind changes in these models are also compared with the ones obtained for our LOVECLIM LGM simulation (Men- viel et al., 2008). The latter differs somewhat from the PMIP protocol: we use the ICE-4G orographic reconstruction, a fully interactive vegetation model and the present-day sea-level.

Figure 4.1 shows the zonal component of the annual windstress zonally averaged for the pre-industrial (solid line) and LGM (dashed line) simulations for the HadCM3M2, CCSM3, IPSL-CM4-V1 and MIROC3.2 CGCMs. The HadCM3M2 and the IPSL model exhibit weaker southern hemispheric westerlies (-12%) during the LGM, whereas the CCSM3 simulates stronger westerlies (+8%) in both hemispheres. The southern hemispheric windstress anomalies in the MIROC model are relatively small (-6%). Consistent with the CCSM3 model result, LOVECLIM also simulates stronger westerlies under LGM conditions (not shown). Most strikingly
4.4 Results

Climate

The forced increase in southern hemispheric 10m winds in experiment WINDP leads to a 22% greater zonal windstress in this area (figure 4.2). As a result, the Antarctic Circumpolar Current (ACC) strengthens and the mass transport through Drake
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passage increases by about 40%. The northward anomalous Ekman transport enhances upwelling of Circumpolar Deep Water (CDW) (figure 4.3 and 4.4). Because the upwelled deep water is colder and saltier than the surface waters an overall cooling (figure 4.3) and salinity increase (not shown) are observed near the surface. The stronger 10m winds also lead to increased evaporation (+20%), which further cools the ocean surface and increases surface salinity south of 40°S. On the other hand, the greater northward Ekman transport in the Southern Ocean induces more convergence around 40°S and therefore increased downwelling there (figure 4.3). In the WINDP simulation we obtain an increase in the production of AABW from 16 Sv to 32 Sv. This can be partly explained by the increased surface density. This increase of AABW is likely to affect also the overall strength of the ACC by processes described in Cai and Baines (1996).

As a result of stronger upwelling and evaporation, the Southern Ocean cools initially in the latitudinal band of 40°S–60°S. However, after about 40 years, a positive SST anomaly develops between 100°–150°E that can be related to ocean circulation anomalies. Understanding the mechanism of these advective anomalies is beyond the scope of the paper. A possible mechanism might involve a wind-induced spin-up of the southern hemispheric supergyre (Cai, 2006). Furthermore, as a result of stronger AABW formation and the increased equatorward export of
4.4. Results

Figure 4.3: Same as figure 2, but for Ekman pumping velocity zonally averaged (m/s) for PIN (solid line), WINDP (dashed line) and WINDM (dotted line).
cold water in the bottom layers, an increased poleward heat transport (not shown) occurs that balances the surface heat loss during convection near Antarctica. While the surface density south of 40°S increases, the slope of the isopycnals steepens near the surface leading to an enhanced heat transport via larger bolus velocities (Gent and McWilliams, 1990). This effect has been described in detail in Stocker et al. (2007).

The reduction in 10m winds over the Southern Ocean in WINDM and WINDMCP leads to a 33% decrease of the zonally averaged zonal windstress component (figure 4.2). Due to reduced Ekman pumping near Antarctica (figure 4.3) the upwelling of CDW decreases. Reduced Ekman pumping increase the SST between 40°S and 60°S by up to 2.5°C (figure 4.5). The weaker winds also lead to a 20% decrease in evaporation over the Southern Ocean, which induces a freshening of the surface waters and a reduction of the latent heat flux. The resulting net decrease of surface density leads to a reduction of AABW production from 16 Sv to 10 Sv (~38 %) (figure 4.4). In addition, the poleward heat transport decreases by about 40%. Reduced zonal winds and reduced AABW formation eventually lead to a weakening of the ACC transport through Drake Passage by about 30%.

As will be discussed below, the WINDMCP experiment that mimics the abiotic response to weakened westerlies, simulates a significantly lower CO$_2$ concentration than the WINDM experiment. This leads to an additional global cooling and a reduced warming between 40°S and 60°S compared to WINDM.

4.4.1 Marine biochemical response

As can be seen in figure 4.6, anomalously strong upwelling in experiment WINDP causes an increase in the euphotic phosphate (PO$_4^{3-}$) content (+9%) south of 30°S. The marine export production therefore increases by 7% in the latitudinal band 30°S-60°S.

For experiment WINDM, weaker Southern Ocean upwelling of phosphates into the euphotic zone (~20%) lead to a 14% reduction in export production between 30°S-60°S (figure 4.6).

Under present-day conditions a part of the upwelled waters in the Southern Ocean is exported northward through Ekman transport and is partly incorporated into the Subantarctic Mode Water (SAMW). At low latitudes, this water mixes into the thermocline and can eventually be upwelled in the major upwelling zones (Marinov et al., 2006). The nutrient excess in WINDP or the deficit in WINDM can therefore be transmitted to the low latitudes. As can be seen on figure 4.6, the phosphate content is greater in the euphotic zone from the Southern Ocean to the northern tropics in WINDP, and lower in WINDM. However, for experiment WINDP, there is a negative anomaly in the euphotic phosphate content in the Eastern Equatorial Pacific, which could be the result of a deepening of the thermocline and of the nutricline. The globally averaged phosphate concentration in the euphotic
4.4. Results

Figure 4.4: Global streamfunction (Sv) averaged for the last 50 years of the control run (PIN) (top), averaged for years 450-500 of experiments WINDP (bottom left) and WINDM (bottom right).

Figure 4.5: SST anomalies (°C) for WINDP–PIN (left) and WINDM–PIN (right). The solid line represents the sea ice thickness contour at 0.1m for PIN and the dashed line displays the sea ice thickness contour at 0.1m averaged for the years 600-1000 for WINDP (left) and WINDM (right).
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Figure 4.6: Upper panels: Phosphate anomalies ($\mu$mol/l) averaged over the euphotic zone for the years 600-1000 for WINDP–PIN (left) and WINDM–PIN (right). Lower panels: Same as upper panels but for export production (gC/m²/yr).

...zone is significantly larger in the WINDP experiment than in the WINDM experiment. Hence the global export production increases by +4% in WINDP, whereas it decreases by 10% for WINDM. This has important consequences for the CO$_2$ fluxes in these two experiments, as will be discussed below.

In the control experiment, DIC-enriched waters are upwelled to the surface in the Southern Ocean enhancing the partial pressure of CO$_2$ in the ocean and reducing the net flux of CO$_2$ from the atmosphere to the ocean. In experiment WINDP, the stronger southern hemispheric westerlies enhance the upwelling of DIC-rich waters as documented in figure 4.7. Compared to the control run, the thermocline waters south of 20°N contains about 25 $\mu$mol/l more DIC, which leads to a substantial increase in mean oceanic carbon release in the latitudinal band 35°S-50°S. On the other hand, south of 50°S, even with the greater amount of DIC-rich waters brought to the surface, the ocean still acts as a sink of carbon. Negative temperature anomalies compensate for the DIC increase in surface waters. Stronger winds result in a sink that is even larger in that area in experiment WINDP as compared to PIN.
In the WINDM experiment weaker westerlies reduce the outcropping of DIC-rich waters (figure 4.7). Compared to the control run, the surface waters south of 20°N in WINDM have about 40 µmol/L less DIC, whereas the DIC accumulates in the deeper parts of the ocean. This dramatic DIC decrease in the upper layers of the Southern Ocean more than compensates the warming in that area. As a result, south of 30°S the ocean acts now as a net sink of CO₂ (figure 4.8). The change is clearly visible around 40°S where the net source during PIN reverses into a weak sink during WINDM.

For the final quasi-equilibrium stages of WINDP the oceanic carbon stock has lost about 8 GtC, which induces a net atmospheric CO₂ increase of about 4 ppmv (figure 4.9). For WINDM, the oceanic carbon stock increases by about 10 GtC, which induces a net atmospheric CO₂ decrease of about 5 ppmv. This is much smaller than simulated by T06. In contrast to T06, however, our export production is varying due to the enhanced/weakened upwelling of nutrient-rich waters. In fact, for both sensitivity experiments, WINDP and WINDM, the simulated changes in export production have the tendency to compensate the DIC-effect on southern hemispheric CO₂ fluxes.

To better quantify the effects of anomalous export production on the atmospheric CO₂, we performed an experiment in which the 3-dimensional PO₄³⁻ field is kept at its climatological value, as diagnosed from the pre-industrial control run. Therefore, the export production is basically the same as in PIN. Only minor differences in marine export production may occur as a consequence of temperature variations. The relative change in global export production with respect to PIN amounts to about 1%. As can be seen in figure 4.7, the larger export production in WINDMCP compared to WINDM leads to a DIC deficit at the surface of the Pacific Ocean that is 40 µmol/L greater in WINDMCP than in WINDM. This supports the notion that the lower export production in WINDM compensates for a large part the changes in DIC content brought about by the varying upwelling. In the Southern Ocean (between 40°S and 60°S), the changes in DIC, alkalinity, SST and SSS lead to a local pCO₂ decrease of 32 ppmv, which is twice more than that simulated by WINDM. The local pCO₂ in WINDMCP in the latitudinal band 20°S-40°S is about 20 ppmv lower than in the control experiment, whereas it is only 4 ppmv lower in WINDM. After 1000 simulation years, the ocean carbon stock in WINDMCP has gained about 35 GtC, which leads to a net atmospheric CO₂ decrease of 18 ppmv (figure 4.9). The results of WINDMCP are more comparable to the T06 solution. However, in reality the export production is very likely to vary with the wind-variations, thereby damping the CO₂ effect of wind-induced DIC anomalies substantially.
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Figure 4.7: DIC anomaly (µmol/l) in the Pacific basin (latitude vs depth) for WINDP–PIN (top), WINDM–PIN (middle) and WINDMCP–WINDM (bottom). Please note that the color scale of the top panel is different from the color scale of the middle and lower panels.
Figure 4.8: Zonally averaged flux of CO\textsubscript{2} (TmolC/yr) from the ocean to the atmosphere averaged for the last 400 years of runs (left) PIN (solid line) and WINDP (dashed line) as well as (right) PIN (solid line), WINDM (dotted line) and WINDMCP (dashed line). Please note that a negative value indicates an ocean sink.
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![Graph showing atmospheric CO2 time series (ppmv) for WINDP (dashed line), WINDM (dotted line) and WINDMCP (solid line).](image)

Figure 4.9: Atmospheric CO2 time series (ppmv) for WINDP (dashed line), WINDM (dotted line) and WINDMCP (solid line).

### 4.4.2 Summary and Discussion

T06 suggested that changes in the strength and position of the southern hemispheric westerlies could have an important impact on glacial-interglacial atmospheric CO2 variations. To test this hypothesis we performed a series of experiments using an earth system model of intermediate complexity fully coupled to a marine carbon cycle model. Weaker westerlies in these simulations lead to decreased Ekman pumping near Antarctica. The upwelling of CDW is therefore reduced by about 40%, which causes a reduction in the amount of DIC-rich as well as nutrient rich cold waters brought to the surface. As the partial pressure of CO2 (pCO$_2$$_w$) is a function of the DIC concentration in the surface water, the reduced upwelling therefore induces a lower pCO$_2$$_w$. Carried by the SAMW, nutrient anomalies are transported all the way up to 10°N in the Pacific. The resulting globally averaged negative export production anomaly therefore counteracts the effect of DIC changes on atmospheric CO2. The net effect is a drop of atmospheric CO2 by 5 ppmv. The crucial role of export production was further highlighted in the wind sensitivity experiment with the fixed-phosphate, WINDMCP. Indeed when surface DIC content is only influenced by changes in oceanic transport and solubility, the atmospheric CO2 decreases by 18 ppmv. For an increase in surface wind strength by 15%, roughly the opposite results are obtained.
When increasing the Southern Ocean windstress, T06 obtained oscillations between a state with an active AABW formation and a state without any AABW formation. Using constant enhanced wind forcing, the associated variations in atmospheric CO$_2$ attained magnitudes of up to 35 ppmv. The two different oceanic circulation regimes of T06 are quite similar to the ones obtained in our WINDP and WINDM experiments. However, our 5 ppmv atmospheric CO$_2$ response contrasts with the T06 result. The wind sensitivity experiment with the fixed phosphate (WINDMCP) demonstrated that this difference can be explained by the compensating effect of export production anomalies, mainly in between 30°S and 50°S, that were neglected in T06. While our modeling results clearly document the importance of changes in export production for the glacial wind sensitivity, it should be noted that neglecting that contribution, as in T06, we are able to reproduce the main results described in T06. Figure 4.10 concisely summarizes T06’s hypothesis as well as the counteracting effect due to changes in export production described in this paper.

In our modeling study, the dependence of primary production on iron as well as vegetation changes were not taken into account. In the southern hemisphere, the
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The strength of the westerlies correlates well with the strength of the stormtrack. In fact, transient eddy activity in the stormtrack plays a key role in organizing the Southern Annual Mode (Jin et al., 2006) as well as the mean westerlies. Weaker storms and weaker mean westerlies are likely to result in less Australian and Patagonian dust deposition in the Southern Ocean and hence a reduced iron fertilization. This may result in reduced export production in the Southern Ocean, thereby amplifying the phosphate effect described here. On the other hand, extended source areas and a drier atmosphere could be factors that enhance the dust deposition. To complete our analysis, we also performed a similar experiment to WINDP but with an interactive vegetation and terrestrial carbon component (VECODE module). The globally colder and drier conditions lead to a small release of carbon from the vegetation. The ocean buffers this release, which induces a net atmospheric CO$_2$ increase of only 4 ppmv. Hence the vegetation effects do not alter our main conclusions.

The results presented here are in agreement with the study performed by Winguth et al. (1999), using the carbon cycle HAMOCC3 coupled online with the Hamburg Large Scale geostrophic ocean general circulation model (LSG OGCM). For a 10% enhancement of the southern hemisphere westerlies, they obtained an atmospheric CO$_2$ increase of only 2 ppmv while the global export production increases by 10%. Our results are also supported and complemented by a recent modeling study using a 3D ocean model fully coupled to a marine carbon cycle model (Tschumi et al., 2008). In response to variations in the southern hemisphere westerlies of +/- 50%, Tschumi et al. (2008) obtain +/- 10 ppmv changes in atmospheric CO$_2$, when their model is tuned to a realistic strength of the meridional overturning circulation. However, in their study, the nutrient anomalies generated in the Southern Ocean are not transported to low latitudes.

Based on the fact that neither paleo-proxies, nor the PMIP LGM simulations provide strong support for weaker southern hemisphere westerlies, we conclude that the scenario described in T06 is a very uncertain scenario. Moreover, our model results suggest that weaker or equatorward shifted westerlies would lead to only marginal CO$_2$ variations due to the compensating effects of reduced DIC and nutrient upwelling. We therefore suggest that wind-changes in the Southern Ocean are an unlikely candidate to explain a large fraction of the glacial-interglacial CO$_2$ variations. Further modeling studies using coupled models would be needed to investigate the uncertainties associated with our results.

This study is also relevant to understand the importance of recent changes in the southern hemispheric westerlies on the sequestration of anthropogenic CO$_2$. It has been observed that, since the 1970’s, the winds over the Southern Ocean have shifted poleward and intensified (Thompson and Solomon, 2002) presumably due to ozone depletion in the stratosphere (Sexton, 2001). Our study demonstrates a rather complex air-sea CO$_2$ flux response to intensifying winds (figure 4.8): stronger upwelling of DIC rich waters induces a greater outgassing of CO$_2$ near 40°S, whereas the CO$_2$ sink centered at 60°S is enhanced due to the stronger winds and colder
conditions. In addition, north of 35°S, a simulated increased export production leads to an enhanced sequestration of CO$_2$. In agreement with earlier studies (Lenton and Matear, 2007; LeQuéré et al., 2007; Lovenduski et al., 2007), a strengthening of the southern hemispheric westerlies induces a slight weakening of the CO$_2$ sink in the Southern Ocean. However, for intensifying winds and sustained anthropogenic CO$_2$ emissions, the greater level of atmospheric CO$_2$ might overcompensate the effect of the wind-induced upwelling of DIC as already argued by Zickfeld et al. (2008). Our results also reveal the importance of mid and low latitude export production changes in partly off-setting the effects of high latitude changes in surface DIC.

**Bibliography**


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P.I. Moreno, T.V. Lowell, G.L. Jacobson Jr, and G.H. Denton. Abrupt vegetation and climate changes during the last glacial maximum and last termination in the
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Chapter 4. Climate and marine carbon cycle response to changes in the strength of the southern hemispheric westerlies.
Chapter 5

Meridional reorganizations of marine and terrestrial productivity during Heinrich events

To study the response of the global carbon cycle to a weakening of the Atlantic Meridional Overturning Circulation (AMOC) a series of freshwater perturbation experiments is conducted both, under pre-industrial and glacial conditions using the earth system model of intermediate complexity LOVECLIM. A shutdown of the AMOC leads to substantial cooling of the North Atlantic, a weak warming of the southern hemisphere, intensification of the northeasterly trade winds and a southward shift of the Intertropical Convergence Zone (ITCZ). Trade wind anomalies change upwelling in the tropical oceans and hence marine productivity. Furthermore, hydrological changes associated with a southward displacement of the ITCZ, lead to a reduction of terrestrial carbon stocks mainly in northern Africa and northern South America, in agreement with paleo-proxy data. In the freshwater perturbation experiments, the ocean acts as a sink of CO$_2$, primarily through increased solubility. The net atmospheric CO$_2$ anomaly induced by a shutdown of the AMOC amounts to about +15 ppmv and -10 ppmv for pre-industrial and glacial conditions, respectively. This background state dependence can be explained by the fact that the glacial climate is drier and the terrestrial vegetation therefore releases a smaller amount of carbon to the atmosphere. This study demonstrates that the net CO$_2$ response to large-scale ocean circulation changes has significant contributions both, from the terrestrial and marine carbon cycle.
Chapter 5. Meridional reorganizations of marine and terrestrial productivity during Heinrich events

5.1 Introduction

Paleo-proxys clearly document (Vidal et al., 1997; McManus et al., 2004) that Heinrich events, interpreted as manifestations of major glacial ice-sheet instabilities (Heinrich, 1988), weakened the Atlantic overturning circulation with impacts on the large-scale climate system. Ice cores from Greenland (Dansgaard et al., 1993), marine sediment cores from the North Atlantic (Bond et al., 1992; Bond, 1993; Bard et al., 2000), the North Pacific (Harada et al., 2006) and the Mediterranean Sea (Rohling et al., 1998; Cacho et al., 1999) as well as lake and loess records from Europe (Shi et al., 2003; Watts et al., 1996; Thouveny et al., 1994) and Asia (Swann et al., 2005) provide unequivocal evidence for a wide-spread northern hemispheric cooling during Heinrich events, prevailing for several hundreds to thousand of years. Similar climate anomalies occurred in response to a slow down of the AMOC (McManus et al., 2004) during the Younger Dryas (YD) cold interval around 12.7-11.6 kyr BP. Unlike the Heinrich events that were associated with iceberg surges throughout the North Atlantic, the YD event has been linked to the drainage of Lake Agassiz (Teller et al., 2002; Carlson et al., 2007), which added significant freshwater into the North Atlantic (Fairbanks, 1989). Although their triggers were substantially different, it is conceivable to assume that Heinrich events and the YD exhibited similar climate-carbon cycle responses.

The interhemispheric response pattern of Heinrich events has been successfully reconstructed using high-resolution Antarctic ice cores that were synchronized to well-dated Greenland ice cores via methane concentrations (Blunier and Brook, 2001). The results support the notion of a bipolar seesaw (Broecker, 1998; Stocker, 1998) which is characterized to first order by an out-of-phase relationship between northern and southern hemispheric temperatures. Only recently have modeling studies elucidated the physical mechanisms that are responsible for spreading the northern North Atlantic cooling equatorwards (Stouffer et al., 2006; Timmermann et al., accepted; Krebs and Timmermann, 2007) and for warming the southern hemisphere. While the equatorward spreading of the North Atlantic sea surface temperature (SST) anomaly involves air-sea coupling via the wind-evaporation SST (WES) feedback, the South Atlantic warming can be mostly explained in terms of reduced meridional oceanic heat transports. Heinrich anomalies were also associated with global sea level anomalies in the order of 3–30 meters (Siddall et al., 2003; Lambeck et al., 2002).

Most recent modeling studies of Heinrich events have focused on the large-scale physical effects. However, little is known about the global carbon cycle response to a weakened AMOC. Synthesizing existing proxy evidence of marine and terrestrial productivity changes during Heinrich events has been hampered by the sparseness of these records and the lack of suitable model simulations. Furthermore understanding the apparent linkage between glacial CO\textsubscript{2} changes (Monnin et al., 2001; Stuaffer et al., 1998; Indermühle et al., 1999b) and Heinrich events has been a major
challenge because the relative phasing among the two is still difficult to establish. Possible linkages between millennial-scale CO\textsubscript{2} changes and Heinrich events were identified and elucidated using 2-dimensional ocean-energy balance-marine carbon cycle models (Marchal et al., 1998, 1999) and vegetation models, such as the Lund-Potsdam Jena Dynamic Global Vegetation Model (LPJ-DGVM). Neglecting vegetation effects Marchal et al. (1998) and Marchal et al. (1999) found a 10-30 ppmv CO\textsubscript{2} change in response to an AMOC shutdown. This response was explained in terms of reduced ocean solubility, due to enhanced southern ocean temperatures. On the other hand, employing only a crude marine carbon cycle component but a complex vegetation model Scholze et al. (2003b) and Köhler et al. (2005) identified the northern hemispheric vegetation as an important source for CO\textsubscript{2} during periods of reduced AMOC.

It is timely to revisit the issue of millennial-scale carbon cycle dynamics using a model that captures both, the terrestrial and the marine carbon cycle adequately and that can be run for many centuries. The goal of our study is to quantify the relative contributions of these two components to the CO\textsubscript{2} sensitivity under millennial timescale freshwater forcing, both under present-day and glacial conditions.

5.2 The earth system model LOVECLIM

The model used in this study is the earth system model of intermediate complexity, LOVECLIM (See Chapter 2), which is based on a somewhat simplified atmosphere model, an ocean general circulation model, a dynamic-thermodynamic sea-ice model, and oceanic as well as terrestrial carbon cycle components.

5.3 Experimental setup and model performance

5.3.1 Pre-industrial and Last Glacial Maximum (LGM) control simulations

The pre-industrial steady state (PIN) was obtained by forcing LOVECLIM with 278 ppmv of atmospheric CO\textsubscript{2} during 500 years, then allowing the atmospheric CO\textsubscript{2} to vary freely during 700 years.

To obtain a quasi-equilibrium state for LGM conditions (simulation LGM), ECBilt-CLIO-LOCH were forced for 2200 years using LGM orbital parameters, the glacial topography (Peltier, 1994), the associated albedo pattern, LGM forest fraction (Crowley and Baum, 1997) as well as with the LGM atmospheric CO\textsubscript{2} concentration (191.85 ppmv) as recorded in the Taylor dome ice core (Indermühle et al., 1999a). The atmospheric concentrations of CH\textsubscript{4} and N\textsubscript{2}O were prescribed to LGM values, according to ice-core data (Brook et al., 2000; Sowers et al., 2003). The LGM boundary forcing adopted here is described in detail in Timmermann et al. (2005b)
and Justino et al. (2005). After this 2200-year long forced simulation the VECODE model was activated and the coupled model ECBilt-CLIO-VECODE-LOCH was forced for an additional 1300 years with glacial greenhouse gas concentrations. The model still used prescribed LGM orbital parameters and glacial topography, as described above. However, albedo and forest fraction were computed directly from VECODE. Thereafter, the atmospheric CO$_2$ was directly computed from the carbon balance between the atmosphere, the vegetation and the ocean. Another fully coupled 2000 years long simulation that was based on the previous forced model run, was conducted using interactive CO$_2$. As will be documented below this iterative set-up generated a quasi-steady-state LGM simulation with relatively stable CO$_2$ conditions at 202 ppmv and a weak trend of 4 ppmv/1000 years. For the short freshwater perturbation experiments analyzed in this study, we anticipate that this weak CO$_2$ trend will not affect our main conclusions.

### 5.3.2 Freshwater flux experiments

To study the background-state dependence of the biogeochemical response to freshwater perturbations we conducted waterhosing experiments under pre-industrial (FC) and under LGM conditions (FL). To mimic a typical Heinrich event, anomalous freshwater was injected into the northern North Atlantic ($55^\circ$W–$10^\circ$W, $50^\circ$N–$65^\circ$N). The overall duration of the freshwater perturbation was 200 years. The freshwater flux increased linearly during the first 100 years to 2 Sv [1 Sv=$10^6$ m$^3$/s], and decreased at a similar rate in the following 100 years to return to unperturbed conditions (figure 5.1). The total amount of anomalous freshwater released in these experiments amounts to $6.3 \times 10^6$ km$^3$, which is equivalent to a global sea level rise of around 17.5m. Following the freshwater perturbation the simulations FC and FL continued for another 800 years.

### 5.4 Results

#### 5.4.1 Pre-industrial and LGM control simulations

**Climate mean state**

As demonstrated in Driesschaert (2005) the pre-industrial and the present-day climate states are simulated quite realistically by LOVECLIM, given the reduced complexity of the model. Here only the main characteristics of the carbon-cycle model and model existing biases of the control climate state will be reviewed. Figure 5.2 reveals that the simulated zonal mean temperature gradient in the tropical Pacific is not captured realistically, in comparison with the Reynolds and Smith (1995) observations (not shown). This can be partly attributed to an underestimation of the tropical air-sea coupling strength, owing mostly to the low atmospheric resolution,
5.4. Results

Figure 5.1: Solid line: Maximum of the meridional streamfunction (Sv) in the North Atlantic for FC (left) and FL (right). Dashed line: Anomalous freshwater flux (Sv) into the northern North Atlantic for FC (left) and FL (right).

and a diffuse thermocline. Furthermore, we see relatively weak temperature fronts near the western boundary currents - another typical feature of coarse-resolution ocean models. While the extratropical westerlies are captured qualitatively well, the simulated tropical trade winds are too zonal. Precipitation is underestimated in the tropical and stormtrack regions in comparison with the observational records of Janowiak and Xie (1999). Another bias that might affect our modeling results is a positive precipitation bias in tropical eastern Africa, that can be explained in terms of an underestimation of important orographic features in the T21 resolution.

The simulated strength of the pre-industrial Atlantic meridional overturning circulation (AMOC) below 500m amounts to about 27 Sv in the Atlantic (figure 5.1), which is somewhat higher than recent observational estimates of 18±5 Sv by Talley et al. (2003).

Our simulated LGM climate state is characterized by a global surface air cooling of 4.2°C, with respect to the pre-industrial state (table 5.1). A more detailed review of the LGM climate state simulated in different versions of ECBilt-Clio and ECBilt-Clio-VECODE can be found in Timmermann et al. (2004); Justino et al. (2005) and Roche et al. (2006)\(^1\), respectively.

The largest glacial temperature anomalies are found near the Laurentide and Eurasian icesheets. Simulated glacial surface air temperatures drop by 15 to 30°C in the northern extratropics. The equatorward extension of the sea ice in the southern

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\(^1\)In comparison to the LGM simulation conducted by Roche et al. (2006), we use the ICE4G paleotopography, instead of the ICE5G reconstruction. Furthermore, we employ a higher sensitivity of the longwave radiation to CO\(_2\) changes and neglect sea-level changes, changes of the river routing, Antarctic iceberg calving and atmospheric dust loading.
Figure 5.2: Upper panels: Sea surface temperature (°C) averaged over the last 20 years of the PIN run (left) and difference (right) between the longterm annual mean SST of the LGM experiment and the PIN run (in both simulations SST is averaged over the last 20 years). The thick dashed line represents the sea ice contour (average of the last 20 years of each run) corresponding to an annual mean height of 0.1 m. Lower panels: Same as upper panels, but for annual mean precipitation (cm/yr).
5.4. Results

Hemisphere leads to a decrease of surface temperatures of about 6 to 15°C. The tropical regions, however, experience only a moderate annual surface cooling of 1-3°C, mostly due to the atmospheric CO₂ reduction and due to remote effects from the glacial icesheets (Timmermann et al., 2004).

SST anomalies in the North Atlantic attain values of up to 8°C (figure 5.2) which is in good agreement with latest glacial SST reconstructions (Weinelt et al., 2003; Pflaumann et al., 2003). SST differences around Antarctica compare well with LGM reconstructions of Gersonde et al. (2005). Tropical SSTs are 1 to 3°C lower than for the pre-industrial state. This also compares favorably with paleo-reconstructions of Kucera et al. (2005), Pflaumann et al. (2003) and Koutavas et al. (2002), but is at odds with other tropical SST estimates (Guilderson et al., 2001; Lea et al., 2003; Nurnberg et al., 2000; Visser et al., 2003; Kienast et al., 2001; Rühlemann et al., 1999; Rosenthal et al., 2003).

Both, the presence of large orographic barriers and the associated modification of the atmospheric eddy momentum fluxes, as well as the intensified LGM SST gradient lead to an intensification of the trade-winds (figure 5.3) with implications for the subtropical cell transports and evaporative cooling in the tropical regions. The general cooling of the Earth’s atmosphere under glacial conditions leads to a reduced capacity of the atmosphere to hold moisture. This causes a drop of global precipitation by 11 cm/yr during the LGM (figure 5.2). In particular dry areas such as the southern border of the Sahara, the middle- and far-East, the Tibetan plateau as well as central Greenland experience a drop of annual mean precipitation of about 50%. Other areas such as the southern border of the Laurentide icesheet experience a regional increase in precipitation, which is suggestive of a positive feedback that helps to maintain the glacial icesheets.

A direct comparison between modeled LGM ocean states and LGM reconstructions (McManus et al., 2004; Gherardi et al., 2005) is not straight forward. This task is rendered difficult by the transient nature of the LGM. As noted in Timmermann et al. (2004), LGM simulations are designed as steady-state experiments, whereas the observed LGM state occurred about 2,000 years after the long-lasting glacial meltwater pulse Heinrich II. It is unclear whether the AMOC was still recovering from this perturbation, or whether it had already reached a new equilibrium state which could be directly compared with steady-state paleo-model simulations. Our simulated vigorous glacial Atlantic overturning with an amplitude of about 26 Sv compares well with the simulation of Roche et al. (2006) and exhibits also a southward displacement of the deep-water formation areas. Roche et al. (2006) argue that this finding is not inconsistent with available proxy data, although a meaningful comparison may be hampered by the fact that the observed LGM state was probably not an equilibrium state.
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Figure 5.3: Same as figure 5.2, but for annual mean wind velocity (m/s) at 800 mbar. The reference vector is 10 m/s for PIN conditions (left) and 4 m/s for the LGM anomaly (right).

Carbon cycle

The simulated mean atmospheric CO$_2$ concentration during the fully coupled pre-industrial experiment (PIN) amounts to about 282 ppmv. The standard deviation of the CO$_2$ concentration due to interannual variability attains values of up to 2 ppmv. The partitioning of the pre-industrial carbon-reservoirs compares well with present-day estimates: While the terrestrial biosphere holds about 2,050 GtC in PIN, about 39,100 GtC are stored in the ocean. Present-day estimates of these reservoirs are 2,000 GtC and 38,100 GtC (Houghton et al., 2001), respectively. Houghton (1999) suggest that land use changes during the 19$^{th}$ and 20$^{th}$ centuries induced a release of about 120 GtC from the vegetation. Taking this into account, the pre-industrial terrestrial reservoir would account for about 2,120 GtC, which is close to our estimate.

For the LGM steady state experiment (LGM) we simulate a quasi-equilibrium atmospheric CO$_2$ of 202 ppmv. The mean vegetation and ocean carbon stocks are 1,370 GtC and 40,000 GtC, respectively. Recent reconstructions of the LGM terrestrial carbon stock were based on $\delta^{13}$C values of calcite shells of benthic foraminifera (Duplessy et al., 1988; Curry et al., 1988; Bird et al., 1994) as well as on pollen analysis (Crowley, 1995; Adams and Faure, 1998). The estimated values for the terrestrial carbon stock range from 1300-1700 GtC and 600-1300 GtC, respectively. Our simulated terrestrial carbon reservoir is in the range of these storage estimates and is also in agreement with a vegetation model simulation using the LPJ-DGVM forced with LGM climate conditions (Joos et al., 2004).
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The gross biome distribution that we obtain for the PIN and LGM runs (figure 5.4) can be directly compared to the reconstructions compiled by Crowley (1995) (not shown here). For PIN, the main desert areas are well represented except for Central Asia. Tropical forests dominate the Amazonian basin, the Congo basin as well as Indonesia and Papua New Guinea. However, the dominant biome simulated in southeast Africa is rain forest instead of savanna, and is forest instead of grass in the southwestern part of the United states. During the LGM, the desert fraction increases in north Africa, the Middle East and Central Asia reflecting the drier conditions there, in agreement with paleoreconstructions. On the other hand forest is still the predominant vegetation type in western Europe whereas Crowley (1995) suggest that grass covered this area. Most of the main discrepancies between reconstruction and model results can be explained in terms of the mean precipitation biases of our coarse-resolution atmosphere.

In order to show the performances of the marine carbon cycle model (LOCH) under present day conditions, we compare the difference in partial pressure of CO$_2$ between the ocean and the atmosphere ($\Delta$pCO$_2$) as well as the phosphates (PO$_4^{3-}$) distribution obtained during a present-day run (MOD) with observations. A more thorough comparison of the geochemical tracer distributions can be found in Chapter 2. MOD was obtained by forcing the model with the observed atmospheric CO$_2$ variations from 1750 to 2000 year A.D. (Keeling et al., 1996; Etheridge et al., 2001) starting from the PIN state. Figure 5.5 displays $\Delta$pCO$_2$ for the last years of the MOD experiment and the observed data compiled by Takahashi et al. (2002). In both the model and the observations, the Southern ocean, the North Atlantic as well as the North Pacific are a net sink of CO$_2$ and the equatorial regions, particularly the cold tongue region in the Pacific, are a net source. However, the simulated North Atlantic and North Pacific sinks are a little underestimated. The model also exhibits a source of CO$_2$ to the atmosphere in the Atlantic and the Indian oceans in the southern boundary regions near the fronts of the southern subtropical gyres at around 40°S . Mismatches between model results and the observations can be largely explain in terms of SST, wind and sea surface salinity (SSS) differences between the fully coupled model solution and the observed climatology. During the LGM (not shown here), the high latitude CO$_2$ sinks as well as the low latitude sources have a greater amplitude than in MOD. Indeed, the colder high latitudes induce an increase in CO$_2$ solubility, whereas the stronger upwellings at low latitudes enhance the release of CO$_2$ to the atmosphere.

The PO$_4^{3-}$ distribution averaged over the euphotic zone obtained during MOD (figure 5.5) is qualitatively quite similar to the observed one (Garcia et al., 2006). However, significant quantitative differences can be found in the north Pacific and in the northwestern Indian Ocean, where the simulated PO$_4^{3-}$ is lower by 1 $\mu$mol/L than the observed one. The simulated export production is comparable to the observed one (not shown here), except for the Arabian Sea where it is much lower (figure 5.6). The PO$_4^{3-}$ content is higher (by about 0.3 $\mu$mol/L) during the LGM than under pre-
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Figure 5.4: Biome distribution (from purple to dark red: Tropical Rain forest, warm forest, mixed forest, conifers, savanna, grass, cold grass and desert) for the last 20 years of the PIN run (upper left), the LGM run (upper right) as well as averaged over the years 180-200 for FC (lower left) and FL (lower right). Only the dominant vegetation biome per grid cell is displayed.
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Figure 5.5: Upper panels: Annual mean air-sea differences in CO₂ partial pressure ($\mu$atm) (left) for MOD and (right) for the Takahashi et al. (2002) reconstructions. Lower panels: Simulated longterm annual mean phosphates concentration ($\mu$mol/L) averaged over the euphotic zone (0–120m) (left) for MOD and (right) the observed annual mean phosphates concentration averaged over the euphotic zone (0–120m) from Garcia et al. (2006).
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Figure 5.6: Annual marine export production (gC/m²/yr) averaged over the euphotic zone for (left) the last 20 years of the PIN run and (right) LGM anomalies compared to the last 20 years of the PIN.

industrial conditions (PIN) in the southern parts of the Eastern Equatorial Pacific (EEP) and the Eastern Equatorial Atlantic (EEA) (not shown), as can be expected from the intensified trade wind circulation and stronger upwelling. As a result, the marine export production increases by about 15% in the EEP and by about 30% in the EEA (figure 5.6). This is consistent with a compilation of export production as recorded in a large number of sediment cores (Kohfeld et al., 2005) which suggests that during the LGM, the export production may have been equal or higher than today outside the non polar areas. Principally because of the sea ice advance, the export production is reduced by 32% in the North Atlantic and 22% in the North Pacific. In the glacial southern Ocean, the PO₄³⁻ content is about 0.4 µmol/L higher (not shown) because of decreased nutrient utilization. Indeed, the glacial increase in sea ice coverage leads to about 25% decrease in productivity. This result is in agreement with LGM nutrient concentrations in the Antarctic surface ocean as deduced from Cd/Ca ratios in planktonic foraminifera (Elderfield and Rickaby, 2000), as well as with the lower primary production estimate made by Kohfeld et al. (2005).

Our LGM model results are qualitatively in good agreement with these paleo-reconstructions and with the Peacock et al. (2006) synthesis:

1. Simulated LGM PO₄³⁻ values are somewhat larger than for the pre-industrial climate, in particular at high latitudes. Anomalies at low latitudes if not weakly positive, are close to zero.

2. The simulated LGM export production is lower in the polar oceans and of the same order or larger in areas comprised between 50°S and 50°N.
5.4.2 Freshwater flux experiments

Climate response

In response to the North Atlantic anomalous freshening, the AMOC gradually decreases from a maximum strength of about 26 Sv to a minimum value of around 4 Sv for both the freshwater flux experiment starting from a pre-industrial climate state (FC) and the one using LGM boundary conditions (FL) (figure 5.1). Whereas the pre-industrial AMOC recovers completely after about 600 years, the LGM AMOC recovers already after 400 years. This change in glacial hysteresis behavior has been described in detail in Krebs and Timmermann (2007).

Similar to other recent CGCM waterhosing experiments (Stouffer et al., 2006; Timmermann et al., accepted), the weakening of the AMOC leads to a substantial cooling in the North Atlantic (figure 5.7). While, the North Atlantic cooling is more pronounced in the FC simulation, the southern hemispheric warming of up to 2°C is comparable in both FC and FL. The northern hemispheric sea ice moves equatorwards in both FC and FL by about 5-10°, whereas only small sea ice anomalies are simulated in the southern hemisphere. The interhemispheric cooling asymmetry induced by weakening of the AMOC leads to an intensification of the northeasterly tradewind circulation in the northern hemisphere (figure 5.8) and the establishment of a dipole in the precipitation field. This dipole-like pattern is associated with a southward shift of the Intertropical Convergence Zone (ITCZ) (figure 5.9). The physical mechanisms responsible for this ITCZ shift have been described in detail in Broccoli et al. (2006), Timmermann et al. (2005a), Zhang and Delworth (2005) and Krebs and Timmermann (2007). The tropical Atlantic precipitation anomaly is more pronounced in the FC simulation (figure 5.9) because the pre-industrial atmosphere is warmer and can accumulate more moisture than the glacial atmosphere.

Vegetation response

During the weakened AMOC state, the drier and colder conditions in the northern hemisphere lead to significant changes in the vegetation patterns, and to the dominance of plants that are more adapted to such climatic conditions. In FC, a perennial snow cover leads to a reduction of forest and grass over Scandinavia (figure 5.4). Grass is substituted by forest over some parts of northeastern Europe, Siberia and north America and the desert spreads over northern Europe. In the Sahel, desert replaces savanna and in northern South America savanna replaces forest. The most important vegetation and carbon storage changes occur in the tropical latitudinal band 5°S-10°N, both in FC and FL.

Due to increased precipitation in the southern hemisphere, more carbon-rich vegetation can be maintained, leading to an increase of terrestrial carbon stocks in southern Brazil, Paraguay, northern Australia as well as southwest Africa. For the pre-industrial experiment FC, the decrease in terrestrial carbon storage in the
Figure 5.7: Annual mean sea surface temperatures (SST) anomalies (°C) for FC–PIN (left) and for FL–LGM (right). The SST have been averaged over the years 180-200 for FC and FL, and averaged over the last 20 years for PIN and LGM. The thick dashed line corresponds to the annual mean sea ice height of 0.1m in FC and FL (averaged over the years 180-200).

Figure 5.8: Same as figure 5.7, but for annual mean wind velocity anomalies (m/s) at 800 mbar for FC (left) and FL (right).
northern hemisphere outweighs the increase in the southern hemisphere; the net result being the release of up to 130 GtC during the first 300 years of the experiment (figure 5.10). When the AMOC recovers its initial strength, there is no longer any precipitation anomaly and the vegetation takes up some of the carbon previously released. After the complete recovery of the AMOC in FC, the vegetation carbon stock does not fully recover, leaving a terrestrial deficit of 50 GtC as compared to the control state.

As documented in figure 5.4, similar vegetation changes can be observed in FL. However, due to the cold and dry background climate of the LGM as well as the reduced precipitation anomalies, the amplitude of the vegetation anomalies is smaller in FL than in FC. In FL, the terrestrial carbon reservoir only releases 40 GtC during the glacial Heinrich event (figure 5.10). Furthermore, the original carbon stock in the continental vegetation bounces back within a few centuries to its original value.

**Chemical and biological response of the ocean**

In the first stages of the AMOC weakening, colder and fresher surface waters in the northern hemisphere lead to an enhanced CO$_2$ solubility in the ocean, which can efficiently buffer the excess CO$_2$ released by the vegetation. As estimated from the Takahashi et al. (1993) relationship, the global temperature decrease leads to a pCO$_2$ drop of 5 ppmv for FC and 3.4 ppmv for FL. During the first 200 years, the net flux of carbon is from the atmosphere to the ocean (table 5.1) both in the FC and FL experiments. As can be seen in figure 5.11, the colder climate is associated with an increase in the area of the high latitude CO$_2$ sinks. Moreover, weaker upwelling in the EEP and EEA lead to a reduction of the CO$_2$ release to the atmosphere. The oceanic carbon reservoir increases less in FL than in FC because the smaller
Figure 5.10: Upper panels: Atmospheric CO$_2$ content (ppm) for the FC run (left) and FL run (right). Lower panels: Carbon reservoir anomalies (GtC) for the ocean (solid line) and the vegetation (dashed line) for FC (left) and FL (right).
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Figure 5.11: Anomalies of partial pressure of CO₂ differences (µatm) between the ocean and the atmosphere for FC–PIN (left) and FL–LGM (right) averaged over the years 180-200.

terrestrial carbon release leads to a lower atmospheric CO₂ content.

During the AMOC shutdown stages carbon sequestration in the deep ocean is more efficient because of an enhanced oceanic stratification and an overall reduction of tropical upwelling in the main southern hemispheric upwelling zones. In experiment FC (model years 80–100), the DIC values below 1000m are on average 20 µmol/L higher than in the PIN run (table 5.1).

Due to greater stratification in the North Atlantic, the PO₄³⁻ concentration in surface waters is reduced by about 0.5 µmol/L (0.2 µmol/L when integrated over the euphotic zone, figure 5.13). This leads to 50% decrease in marine export production in that area. In the pre-industrial and LGM control simulations, the strongest mean upwellings occur in both the EEP and EEA due to strong Ekman pumping. The associated supply of nutrients into the euphotic zone leads to large marine primary productivity. Due to the AMOC collapse, the circulation, in both the Atlantic and the Pacific, is dominated by the subtropical cells (STC) and the water upwelled in the equatorial regions originates from the sub-tropics (Haarsma et al., 2007). As a result of the northern North Atlantic cooling during the weakened AMOC, the trade winds intensify in the northern tropical Atlantic. Due to the sign change of the Coriolis parameter the southern tropical Atlantic experiences a weakening of the trade winds. Consequently, as it was previously described by Prange and Schulz (2004) and Haarsma et al. (2007), the upwelling strength is increased north of the equator in the EEA, whereas it is decreased south of the equator. In the EEP, the divergent Ekman transport is enhanced in winter but reduced in summer. The global reduction in the amount of water upwelled and the change in source water lead to a reduction of the nutrient content in the upper 1000m in favor of an accumulation in the deep ocean, especially in the Atlantic. The amount of PO₄³⁻
brought to the upper ocean in the EEP and south of the equator in the EEA is therefore reduced. As can be seen in figure 5.13, in those regions the $PO_4^{3-}$ content of the euphotic zone is reduced by about 0.5 $\mu$mol/L compared to PIN. This results in a decrease of the annual mean export production. However, north of the equator in the EEA, the net effect of the stronger upwelling is a slight increase (8%) in export production. Due to the weaker trade winds in the southern tropics, the upwellings are reduced in the southeastern Atlantic and southeastern Pacific. In addition, the Peru (Humboldt) and the Benguela currents are weaker, leading to a reduction in the amount of nutrient rich waters advected from the Southern Ocean to the southeastern Pacific and southeastern Atlantic, respectively. Those regions therefore experience a decrease in export production. While the export production decreases by about 25% in the tropical Pacific due to the weakened AMOC state, the tropical Atlantic export production drops by about 40% (figure 5.12). Globally, the export production decreases by about 14% in FC (figure 5.12, table 5.1).

Until the beginning of the recovery phase of the AMOC the ocean has taken up about 100 GtC, part of which is released back to the atmosphere once the effect of the freshwater perturbation ceases. At the end of the simulation FC, the ocean has gained 50 GtC (figure 5.10) due to colder conditions in the southern hemisphere. Indeed, at the end of FC, the climate state reaches a slightly different equilibrium than the initial one. The existence of potential multiple equilibria of the carbon-climate system deserves a more thorough investigation in a subsequent study.

Under glacial conditions (exp. FL) the overall biogeochemical response to a weakening of the AMOC differs somewhat from that of the pre-industrial water-hosing experiment FC. The global export production experiences a slight decrease of about 3% (figure 5.12, table 5.1). While the response in the tropical oceans is similar to FC, the negative anomalies in the eastern upwelling regions are smaller in FL compared to FC. On the other hand the positive productivity anomalies in the western tropical areas are somewhat larger than in FC and in particular off the coast of Brazil and in the south China Sea. This pattern can be explained in terms of stronger equatorial currents that advect nutrients eastward across the Pacific and the Atlantic (figure 5.13). The net effect of changes in ocean productivity, solubility and upwelling adds up to a net ocean uptake of 50 GtC during the weakened AMOC state simulated by FL.

**Atmospheric CO$_2$ response**

Here we decompose the net CO$_2$ response into vegetation-atmosphere and ocean-atmosphere fluxes. As described above, these individual fluxes are strongly dependent on the climate background conditions. The vegetation response to the AMOC shut down is quite fast, whereas the ocean biogeochemical conditions adjust on longer timescales to the forcing. This lagged response shapes the time-evolution of the atmospheric CO$_2$ concentration and the carbon stocks, as shown in figure 5.10.
Figure 5.12: Marine export production anomalies (gC/m²/yr) for FC–PIN (left) and FL–LGM (right) compared to marine primary production paleoproxy data (stars). The blue stars represent lower reconstructed productivity and the red stars larger productivity compared to the reference state during which the anomaly occurs. References from 30°E to 25°E and from north to south are as follow: Ivanochko et al. (2005); Lin et al. (1999); Sachs and Anderson (2005); Ortiz et al. (2004); Kienast et al. (2006); Hughen et al. (1996); Vink et al. (2001); Rasmussen et al. (2002); Thomas et al. (1995); Pailler and Bard (2002); Plewa et al. (2006); Haslett and Davies (2006); Little et al. (1997).

Figure 5.13: Same as figure 5.7, but for annual mean phosphates concentration anomalies (µmol/L) averaged over the euphotic zone for FC (left) and FL (right).
In FC a shut down of the AMOC first leads to a fast release of 130 GtC from the terrestrial reservoir followed by an increase of the ocean reservoir of 100 GtC with about 30 years lag. When the AMOC starts to recover, the terrestrial carbon storage increases again and therefore leads to an atmospheric CO$_2$ draw down. For FC, the peak to peak increase in atmospheric CO$_2$ amounts to about 15 to 20 ppmv (figure 5.10). After about 600 years, the atmospheric CO$_2$ content attains pre-industrial values (280 ppmv). The radiative effect of the CO$_2$ increase further amplifies the development of the bipolar seesaw, as already suggested by Rohling et al. (2004). Indeed, the additional terrestrial CO$_2$ release warms the southern polar regions by up to 2–4°C, thereby providing an important element for the bipolar seesaw response.

In the glacial waterhosing simulation FL, the vegetation only releases around 40 GtC and the ocean reservoir increases by 50 GtC (figure 5.10). The ocean lags the vegetation changes by only a few years, which leads to a few ppmv (2-3 ppmv) increase in atmospheric CO$_2$, after which the oceanic sink of CO$_2$ becomes dominant. Overall, the AMOC shut down under glacial conditions leads to a peak to peak decrease of atmospheric CO$_2$ of about 10 ppmv.

Our analysis clearly shows that a rather delicate balance exists between the response of the marine and terrestrial carbon cycle. This also suggests that mean-state model biases might generate substantial errors in the CO$_2$ response to a shutdown of the AMOC. Assessing these errors can be done by a careful qualitative comparison between paleo-proxy data and model simulations as well as by comparing our results with those obtained from model simulations which include e.g. a more sophisticated vegetation component (Scholze et al., 2003b; Köhler et al., 2005).

### 5.5 Discussion and comparison with paleoclimate records

**Climate**

A careful validation of our earth system model of intermediate complexity under pre-industrial and LGM conditions has revealed that despite the simplicity of the atmospheric and the vegetation component, many features of the climate and biogeochemical state are simulated qualitatively well and in agreement with present day observations and paleo-proxy reconstructions. This includes the LGM climate background state (Justino, 2004; Justino et al., 2005; Roche et al., 2006, and references therein), the terrestrial carbon reservoir (Adams and Faure, 1998) as well as the biome partitioning under LGM conditions (Crowley, 1995).

The freshening of the North Atlantic leads to a temporary shut down of the AMOC and a substantial reduction of the meridional heat transport in the North Atlantic of about 0.8 PW at 30°N and an increase of the poleward heat transport in the southern hemisphere of about 0.5 PW. While the recovery takes about 500
years under pre-industrial conditions, it is somewhat faster during glacial conditions (~300 years). The reduction of ocean heat transport in the North Atlantic leads to a substantial cooling of the northern North Atlantic, an equatorward expansion of the perennial sea ice cover and an intensification of the northeasterly trade winds. Furthermore, as a result of stronger poleward heat transport in the southern hemisphere, the Southern Ocean warms significantly. A reduction of the southeastern trade winds amplifies this warming due to the wind-evaporation-SST feedback (Krebs and Timmermann, 2007). A substantial weakening of the meridional overturning circulation as simulated in a series of coupled GCMs lead to a global SST reduction of 0.2 to 1.2°C (Timmermann et al., accepted). We simulate a global SST decrease of about 0.4°C, which is in agreement with the other GCMs.

The meridional reorganization of the trade wind circulation and the tropical SST gradients lead to a southward shift of the ITCZ. Both in FC and FL the northern tropical band gets drier and the southern tropical band experiences much wetter conditions during the collapsed AMOC state. The simulated precipitation anomaly pattern is qualitatively in good agreement with paleoclimate reconstructions, as illustrated in figure 5.9. Lake level studies (Flores-Diaz, 1986; Gasse et al., 1990; Street-Perrot and Perrot, 1990; Hodell et al., 1991; Gasse and Campo, 1994; Bonnefille et al., 1995; VanDerHammen and Hooghiemstra, 1995; Johnson et al., 2002; Stager et al., 2002; Adegbie et al., 2003; Scholz et al., 2003; Turney et al., 2004; Lamb et al., submitted), geochemical precipitation proxies from marine sediment cores (Arz et al., 1998; Maslin and Burns, 2000; Hang et al., 2001; Ivanochko et al., 2005; Haslett and Davies, 2006), speleothems (Wang et al., 2001; Bar-Matthews et al., 2003; Dykoski et al., 2005; Wang et al., 2006), as well as tropical ice cores (Thompson et al., 1995) support the notion that the northern hemisphere was drier during Heinrich events and/or the YD. Our simulated precipitation increase in South America is also in agreement with recent paleoclimatic records (Thompson et al., 1998; Baker et al., 2001a,b; Wang et al., 2004; Cruz et al., 2005). The southward shift of the main precipitation bands during weakened AMOC states appears to be a robust feature, both in climate model experiments as well as in paleo-reconstructions.

**Vegetation response**

The southward shift of the ITCZ induces large-scale vegetation changes in the tropics. In response to an AMOC weakening we simulated changes from savanna to desert in northern Africa as well as a reduction of the forest area in northwestern tropical Africa and the northeastern part of South America. The amplitude of these vegetation anomalies depends strongly on the climate-vegetation background state. Under glacial conditions, vegetation adapted to a drier and colder climate already prevail. The simulated changes of terrestrial carbon stock are therefore smaller under glacial than under pre-industrial conditions. For FC, the terrestrial carbon reservoir releases 130 GtC to the atmosphere, whereas only 45 GtC are released in
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Figure 5.14: Vegetation primary production anomalies (kgC/m²/yr) for FC–PIN (left) and FL–LGM (right) compared to vegetation paleoproxy data (stars). The blue stars represent a reconstructed vegetation type that is adapted to a drier climate. See text for references.

FL. Simulated vegetation changes in FC and FL are in qualitative agreement with some pollen studies from equatorial Africa and northern South America that suggest a change towards drier climate vegetation during the YD period (Gasse and Campo, 1994; VanDerHammen and Hooghiemstra, 1995; Maley and Brenac, 1998; Hughen et al., 2004) (figure 5.14). Indeed, as suggested by Myneni et al. (1995), a precipitation reduction in a certain environment leads to a decrease in vegetation primary productivity. The wetter conditions in the southern hemisphere do not lead to large changes in the terrestrial carbon storage, except for southwestern Africa and western South America.

These results are in qualitative agreement with the ones obtained by Köhler et al. (2005) as well as Scholze et al. (2003b) using a more complex vegetation model. Indeed, the main features of the LPJ-DGVM response to a shut down of the AMOC are a reduction in tree cover north of 55°N, a slight increase in carbon storage due to a southward shift of the tree line around 40°N, a replacement of tropical trees by grass due to drier conditions in the latitudinal band 5°S-15°N and a slight increase in carbon storage between 10°S and 20°S due to greater precipitation. However, for the pre-industrial climate, we obtain a terrestrial carbon release of 130 GtC which is much larger than the 40 GtC obtained by Köhler et al. (2005) but lower than the 180 GtC simulated by Scholze et al. (2003b). Köhler et al. (2005) explained their discrepancy with Scholze et al. (2003b) results by differences in the initial climate state. In our simulation the vegetation cover-albedo feedbacks are included, which could be the reason of the larger terrestrial carbon released obtained compared to the Köhler et al. (2005) study. The different time evolution of terrestrial carbon stock simulated by Köhler et al. (2005) could be partly explained by the difference
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in the freshwater flux forcing applied. The sensitivity of the terrestrial carbon release to the initial climate state is consistent with the studies of Köhler et al. (2005) as well as Scholze et al. (2003a) where less carbon is released when starting from a cold climate state due to the dominance of vegetation adapted to drier conditions. However, those results can not be directly compared to ours as they assumed the same climate anomaly for a freshwater flux experiment starting from a pre-industrial climate state as the one starting from a cold climate state. We have demonstrated earlier in this paper that the climate anomalies have qualitatively the same pattern but not the same amplitude, which further amplifies the differences between FC and FL.

The vegetation response is primarily driven by the changes in precipitation and temperature. The agreement between the climate response to a shut down of the AMOC obtained by different modeling studies (Broccoli et al., 2006; Krebs and Timmermann, 2007; Timmermann et al., 2005a; Zhang and Delworth, 2005; Knutti et al., 2004) and paleoclimate as well as paleovegetation records suggests that the overall vegetation response pattern is robust.

Response of the marine carbon cycle

In FC, the strong release of carbon from the terrestrial reservoir induces an increase in atmospheric CO$_2$, which leads to a higher difference between the atmospheric partial pressure of CO$_2$ and the one in equilibrium with the surface ocean. The ocean therefore acts as a sink of CO$_2$. This effect is further amplified by reduced surface water temperature and salinity and hence greater CO$_2$ solubility. The increase of the oceanic carbon inventory by about 100 GtC is however not sufficient to balance the terrestrial release. In FL, the ocean carbon reservoir gains up to 50 GtC due to increased solubility (figure 5.10). This result is contrary to the Marchal et al. (1999) study where the CO$_2$ solubility was reduced due to greater sea surface temperature and salinity. This difference could be due to the fact that Marchal et al. (1999) do not use a 3 dimensional dynamical atmosphere and only a 2 dimensional ocean model.

Figure 5.12 shows the export production anomalies for FC and FL as well as a compilation of reconstructed changes in primary production for the YD and/or Heinrich events as recorded in marine sediment cores. Export production decreases in the North Atlantic in several sediment cores (Thomas et al., 1995; deMenocal et al., 2000; Pailler and Bard, 2002; Rasmussen et al., 2002), consistent with our modeling results for FC and FL as well as with the results of Schmittner (2005). In our model simulations the lower productivity can be explained in terms of a reduced supply of nutrients to the upper ocean due to increased stratification. The increased productivity due to stronger upwelling off the coast of northern Africa in the Atlantic, in particular in FL is in agreement with Haslett and Davies (2006) paleorecord. Our modeling results further suggest that one of the areas which ex-
hibits very large changes both of temperature and productivity is the eastern south Atlantic near the coast of Namibia. A drastic drop in export production due to a reduction in the advection of nutrient rich waters has also been obtained by Schmittner (2005). This result is also in agreement with the paleo-reconstruction of Little et al. (1997). However, other high resolution marine productivity records from this area are needed to consolidate the results.

Striking discrepancies between the model experiments and the reconstructions occur in the EEP (Kienast et al., 2006) and the Arabian Sea (Ivanochko et al., 2005). However, we can only compare our results with single paleorecords from those sites. Other paleoproductivity records would be needed to validate or falsify our results with confidence. In agreement with our study, Schmittner (2005) simulate a reduction in the nutrient content of the upper Pacific ocean, and therefore a lower primary productivity in the EEP. But as the wind-driven changes in upwelling are not taken into account in Schmittner (2005), the global effect of wind stress and nutrient source variations can not be inferred with certainty.

Our model does not include iron atmospheric transport, but the climate anomalies obtained over the North Pacific and Patagonia (the main dust source of the Southern Ocean) does not support any significant change in dust transport. Moreover, an increase in marine primary productivity due to iron fertilization does not necessarily lead to a significant increase in export production (de Baar et al., 2005). We therefore think that the lack of iron limitation in our model should not significantly impact our results.

**Atmospheric CO\textsubscript{2} response**

Atmospheric CO\textsubscript{2} changes in response to a shutdown of the AMOC are the result of two large opposing carbon fluxes from the vegetation to the atmosphere and from the atmosphere to the ocean. The simulations have shown that the delicate balance between those fluxes can result in quite different atmospheric CO\textsubscript{2} responses. Any qualitative comparison with paleoproxies is, at this stage, associated with large uncertainties.

Variations in atmospheric CO\textsubscript{2} associated with the YD can be reconstructed from ice core records. The analysis of Monnin et al. (2001) suggests that the atmospheric CO\textsubscript{2} concentration increased by more than 20 ppmv during the YD. The immediate effect of Heinrich event I (H1) on atmospheric CO\textsubscript{2} is more difficult to assess. As H1 coincides with the onset of the deglaciation (Marshall and Koutnik, 2006), which was characterized by large orbital forcing anomalies in the northern and southern hemisphere, our idealized experiments FC and FL can not be directly compared with the atmospheric CO\textsubscript{2} record for that time period. Attributing the observed changes of atmospheric CO\textsubscript{2} during 19ka -14ka to variations of the AMOC, orographic icesheet forcing or orbital forcing is difficult and beyond the scope of this paper.
During MIS3 CO$_2$ varied between 185-220 ppmv on a number of different timescales. There is some evidence (Stauffer et al., 1998) that these variations correlated with Heinrich events, although both the exact timing of the Heinrich events as well as of the CO$_2$ variations is quite uncertain. Furthermore, it has to be noted that early glacial meltwater pulses (Heinrich 6 through C25) (Chapman and Shackleton, 1999) occurred on a precessional timescale of 19-23 ka. Orbital forcing might have affected the CO$_2$ history and may have triggered meltwater pulses independently from each other, resulting in a possible correlation between the CO$_2$ and meltwater forcing, but no direct causality.

However, the range of the simulated atmospheric CO$_2$ response [-10:+20 ppmv] obtained in response of a collapse of the AMOC is reasonable compared to the atmospheric CO$_2$ variations as recorded in ice cores.

5.6 Summary

We conducted a series of climate model simulations with an earth system model of intermediate complexity to understand the CO$_2$ response to a shutdown of the AMOC under pre-industrial and glacial background conditions. The AMOC shutdown was induced by a 200-year-long injection of freshwater into the northern North Atlantic, corresponding to a global sea level rise of roughly 20m. Strong cooling in the North Atlantic spreads via changes of the atmospheric circulations (Krebs and Timmermann, 2007) and leads to an intensification of the northeasterly trade winds. These atmospheric circulation changes are accompanied by a southward shift of the ITCZ, generating drier conditions in the north (5°S-15°N) and wetter conditions in the south (8°S-20°S). Following the reduction of northern hemispheric precipitation, the vegetation releases carbon to the atmosphere, a part of which is taken up by the ocean, mainly due to an increased solubility. The atmospheric CO$_2$ therefore increases by about 15 ppmv when starting from a pre-industrial state whereas it decreases by about 10 ppmv when the initial climate state is a glacial one. The atmospheric CO$_2$ response is a delicately balanced sum of the terrestrial and marine inventory changes. This emphasizes the importance of the initial climate state on the CO$_2$ response to large-scale ocean circulation changes. In order to test the robustness of our results, similar simulations should be performed with other earth system models that include both the marine and terrestrial carbon cycle as well as a more comprehensive atmospheric component.
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<table>
<thead>
<tr>
<th>AMOC (Sv)</th>
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Table 5.1: The table shows for each experiment the mean annual value of some climatic and carbon cycle variables averaged over the last 50 years of the PIN and LGM runs and averaged from years 80 to 100 for FC and FL. AMOC represents the maximum strength of the overturning circulation in Sv; T surface and SST are respectively the surface air temperature and the sea surface temperature in degree Celsius averaged over the whole globe; pCO$_2$ a is the atmospheric CO$_2$ content in ppmv; ΔFlux CO$_2$ atm–ocean is the anomalous flux of CO$_2$ from the atmosphere to the ocean in GtC/yr for FC and FL respectively compared to PIN and LGM; Export production averaged over all the basins in gC/m$^2$/yr; PO$_4^{3-}$ is the phosphates content averaged over the euphotic zone (0–120m) in µmol/L and DIC averaged over the deep ocean in µmol/L.
Chapter 6

CO₂ Release from Northern Hemisphere Vegetation Contributes to Millennial-scale Antarctic Warming Events

Simulations with an Earth system model of intermediate complexity demonstrated that strong glacial meltwater pulses into the North Atlantic weaken the Atlantic Meridional Overturning Circulation (AMOC), shift the Intertropical Convergence Zone southward, warm the Southern Hemisphere, and reduce the global terrestrial carbon stock. Generated partly from a collapse of the Northern Hemisphere tropical rainforests, the ∼20 ppmv CO₂ released are superimposed on the classical bipolar seesaw pattern of northern cooling and southern warming due to the weakened AMOC. In Antarctica, this greater atmospheric CO₂ content contributes even further to warming (i.e., to A-events). In the Northern Hemisphere, the radiatively induced warming due to the additional CO₂ is compensated by a cooling due to a reduction of Atlantic meridional heat transport. The results provide a new framework that explains a significant fraction of the observed millennial-scale variance in climate records during Marine Isotope Stage 3 (MIS3) in Antarctica and the Southern Ocean.

6.1 Introduction

Paleoclimate records from the North Atlantic suggest that massive discharges of icebergs into the North Atlantic during the last glacial period (Heinrich, 1988) (so-called Heinrich events) drastically weakened the Atlantic Meridional Overturning Circulation (AMOC) (Vidal et al., 1997; McManus et al., 2004). An associated weakening of the poleward heat transport cooled the North Atlantic substantially (Skinner et al., 2003), shifted the Intertropical Convergence Zone (ITCZ) southward.
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(see Dahl et al. (2005); Krebs and Timmermann (2007) for a review), and warmed Antarctica (Blunier et al., 1998; EPICA and community members, 2006) and the Southern Ocean (Bianchi and Gersonde, 2004). This characteristic temperature response has been referred to as the bipolar seesaw pattern (Stocker, 1998).

A recent comprehensive ocean heat budget analysis of a waterhosing simulation (Stocker et al., 2007) revealed that changes in meridional ocean heat transport play a key role in warming the Southern Hemisphere. In some regions, such as the Benguela upwelling region (Krebs and Timmermann, 2007) and the southeastern tropical Pacific (Timmermann et al., 2007), a substantial weakening of the trade wind circulation can further amplify regional warming by reducing upwelling, evaporation and mixing.

While the majority of recent Coupled General Circulation Model (CGCM) studies support these findings, some model experiments (Rind et al., 2001) have questioned the simple seesaw interpretation and have called for a more active role of the Southern Hemisphere atmospheric circulation. Moreover, the average magnitude of the Southern Hemisphere temperature response simulated by state-of-the-art pre-industrial CGCM waterhosing experiments (Stouffer et al., 2006) is much smaller than that reconstructed from Antarctic ice cores (Jouzel et al., 2007; EPICA and community members, 2006). This finding also suggests that an important amplification mechanism may have been overlooked in our current understanding of the Antarctic warming events (A-events). As paleo-proxy data reveal a high correlation between millennial-scale Antarctic temperature variations and atmospheric CO\textsubscript{2} concentrations (figure 6.1) (Ahn and Brook, 2007), Rohling et al. (2004) recently proposed that these atmospheric CO\textsubscript{2} variations may be contributing to the A-events.

To better understand the origin of the observed atmospheric CO\textsubscript{2} anomalies during Heinrich events, a number of waterhosing experiments have been conducted with coupled climate-carbon cycle models of different complexity. These model simulations support the notion that a weakened AMOC causes an increase in atmospheric CO\textsubscript{2}, either through oceanic (Marchal et al., 1998; Schmittner et al., 2007) or terrestrial (Obata, 2007; Menviel et al., 2008) climate-carbon cycle feedbacks. In accordance with numerous paleo-proxy data, the mechanism proposed by Menviel et al. (2008) and Obata (2007) invokes a southward migration of the ITCZ during an AMOC shutdown and a subsequent release of terrestrial carbon. About 50\% of the CO\textsubscript{2} increase in simulations under pre-industrial conditions can be attributed to the drying and shrinking of tropical rainforests (Menviel et al., 2008).

The goal of this study is to systematically quantify the effect of carbon-cycle climate interactions on the generation and amplification of A-events during Marine Isotope Stage 3 (MIS3). The study is based on a series of intermediate complexity modeling simulations.
Figure 6.1: (top) Atmospheric CO$_2$ content as measured in Taylor dome ice core (blue) (Indermühle et al., 2000) and in Byrd ice core as measured by Neftel et al. (1988) (black) and Ahn and Brook (2007) (red); (bottom) Temperature anomalies over Antarctica derived from δD measurements in EPICA Dome C ice core (red) (Jouzel et al., 2007) plotted on the EDC2 timescale. Temperature anomalies averaged over the latitudinal band 75°S-90°S obtained during the experiment GC (black). Note the different temperature scales for the simulated temperature anomalies and the ice core reconstruction.
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### 6.2 Model and experimental setup

The model used in this study is the Earth system model of intermediate complexity LOVECLIM, which includes components for the atmosphere, the ocean, and sea-ice as well as an oceanic and terrestrial carbon cycle model. LOVECLIM is described in detail in Chapter 2.

The model was spun up to a fully coupled pre-industrial steady state (PIN) by forcing LOVECLIM for 500 years with an atmospheric CO\textsubscript{2} of 278 ppmv and then allowing the atmospheric CO\textsubscript{2} to vary freely for the next 700 years. After the coupled pre-industrial model spin up had been reached, two meltwater experiments were performed. The first was conducted with an interactive atmospheric CO\textsubscript{2} (FC); the second, with a CO\textsubscript{2} content fixed at 280 ppmv (FCC). To mimic a typical Heinrich event, anomalous freshwater was injected into the northern North Atlantic (55\degree W–10\degree W, 50\degree N-65\degree N), as in Menviel et al. (2008). The overall freshwater perturbation lasted for 200 years: freshwater flux increased linearly during the first 100 years to 2 Sv [1 Sv=10\textsuperscript{6} m\textsuperscript{3}/s] and then decreased at a similar rate during the subsequent 100 years, by which time unperturbed conditions had returned (Menviel et al., 2008) (figure 6.2). The anomalous freshwater released in these experiments amounted to 6.3 \times 10\textsuperscript{6} km\textsuperscript{3}, which is equivalent to a global sea level rise of around 17.5m. Following the freshwater perturbation, the simulations continued for another 800 years.

To test the impact of observed atmospheric CO\textsubscript{2} variations on the evolution of southern hemispheric temperatures during the last glacial period, an additional sensitivity experiment (GC) was performed. The model was forced by the time-evolution of the atmospheric CO\textsubscript{2} concentration as measured in the Byrd ice core (Neftel et al., 1988; Ahn and Brook, 2007) and was run from 62 ka B.P. to 34 B.P. ka using an acceleration factor of 5 (Timm and Timmermann, 2007). Topography, albedo, orbital forcing and the atmospheric content of CH\textsubscript{4} and N\textsubscript{2}O were fixed at Last Glacial Maximum values (Timm and Timmermann, 2007; Menviel et al., 2008).

### 6.3 Results

As reported in Menviel et al. (2008), atmospheric CO\textsubscript{2} concentration in FC increased by 20 ppmv in response to the freshwater-induced shutdown of the AMOC (figure 6.2). This response is mainly attributable to the simulated drying of the Northern Hemisphere and the resulting changes in vegetation patterns and terrestrial carbon stock. About 50% of the CO\textsubscript{2} increase originates from a very significant reduction in the size of the tropical rainforests and savanna in Africa and South America during the simulated Heinrich events. This result is consistent with \delta^{13}CO\textsubscript{2} data from the Taylor Dome ice core (Smith et al., 1999), which provide evidence for a source of isotopically light carbon (such as from vegetation) during Heinrich events H1 and H2. Moreover our modeling results are also consistent with the observed \sim 20 ppmv
6.3. Results

Figure 6.2: Upper left panel: Time series of the maximum of the meridional streamfunction (Sv) in the North Atlantic for FC (blue) and the FCC simulation (black), as well as for the anomalous freshwater flux (Sv) into the northern North Atlantic (grey). Upper right panel: Time series of atmospheric CO$_2$ content for FC (blue), FCC (black) and the anomalous freshwater flux (Sv) into the northern North Atlantic (grey). Lower left panel: Time series of 2m air temperature anomaly [°C] averaged over the latitudinal band 60°S–90°S for FC–PIN (blue) and FCC–PIN (black); Lower right panel: Same as lower left panel, but for sea-ice area (x10$^{12}$km$^2$) anomaly in the southern hemisphere. Values have been smoothed using a 50-year running mean filter.
increase in CO$_2$ that occurred at the same time as Heinrich events H4-H6 (Ahn and Brook, 2007) (figure 6.1).

Compared to the FCC experiment, the atmospheric CO$_2$ rise in FC leads to a globally averaged surface temperature increase of 0.3°C, with the greatest increase seen at the poles (figure 6.3). In areas poleward of 60°N and 60°S surface air temperatures rise by as much as 3°C and 1.5°C (e.g. in the Ross Sea), respectively, compared to FCC. In the Northern Hemisphere polar regions, the CO$_2$-induced warming is outweighed by the much stronger cooling during the AMOC shutdown. In the Southern Hemisphere, the increase in atmospheric CO$_2$ is superimposed on the bipolar seesaw temperature response pattern (figure 6.3, left panel). The CO$_2$ contribution to the Antarctic warming during the AMOC shutdown in the present experiments is as large as the one due to the bipolar seesaw effect $^1$. As a result of the CO$_2$ increase in FC (figure 6.2, lower right panel), the Southern Ocean sea-ice extent shrinks significantly, leading to less surface albedo and greater absorption of shortwave radiation by the ocean, and thereby to increased warming.

These modeling results clearly demonstrate that the Northern Hemisphere vegetation response to an AMOC shutdown and its effect on CO$_2$ concentrations can play a significant role in amplifying Antarctic warming during Heinrich events.

To quantify the amount of CO$_2$-induced warming in observed A-events, a transient climate modeling experiment (GC) using glacial maximum boundary conditions (Timm and Timmermann, 2007), except for CO$_2$ forcing, was conducted. The time-varying CO$_2$ concentrations were prescribed using the ice core data of Neftel et al. (1988) for 34 ka B.P. to 47.2 ka B.P. and of Ahn and Brook (2007) for the period from 47.4 ka B.P. to 62 ka B.P. (Figure 6.1), interpolated onto an-equidistant

$^1$Note also that our simulated bipolar seesaw response in Antarctica is significantly larger than the multimodel-ensemble mean response derived from an ensemble of recently conducted CGCM waterhosing experiments (Stouffer et al., 2006).
time-axis. The simulation reproduced the observed time evolution of temperature in Antarctica (Jouzel et al., 2007) quite well (figure 6.1). However, the amplitude of the simulated millennial-scale temperature variations averaged from 90°S-75°S is a factor 3-4 smaller than that derived from the EPICA Dome C temperature reconstruction.

There may be several reasons for this underestimation:

- The CO$_2$-induced warming is only a minor contributing factor to the A-events. The more important factor is the bipolar seesaw effect (Blunier et al., 1998; Stocker, 1998; Blunier and Brook, 2001; Stocker and Johnsen, 2003). However, according to the FC simulation, the bipolar seesaw effect in our model has a similar magnitude than the CO$_2$-induced warming. Even the FCC experiment simulates a stronger Antarctic warming in response to an AMOC shutdown than most of the CGCMs subjected to a similar freshwater forcing (Stouffer et al., 2006).

- The temperature reconstruction obtained by Jouzel et al. (2007) also includes millennial-scale topographic effects that affect the $\delta$D via the lapse-rate effect. If this were the case, this would suggest that the Antarctic ice-sheet is much more unstable on millennial timescales than previously thought.

- The polar amplification in the LOVECLIM climate model version used here is underestimated due to biases in the regional climate sensitivity to CO$_2$ forcing. Such biases may include misrepresentations of polar clouds, water vapor feedbacks and sea-ice sensitivity to external forcing.

- Our model simulation GC uses an acceleration factor of 5, which means that in our case 28,000 years of CO$_2$ forcing history are squeezed together into 5,600 model years. As shown in Timm and Timmermann (2007), such acceleration techniques may lead to under-representations of forced variability, because the model does not have sufficient time to adjust to the accelerated external forcing. This can be particularly problematic in areas such as the Southern Ocean. The GC experiment was repeated with an acceleration factor of 10. The simulated temperature variations in Antarctica were less than those shown in figure 6.1 using an acceleration factor 5. This suggests that an even smaller acceleration than the one used here would further boost the amplitude of the simulated CO$_2$ response in figure 6.1.

- Regional aspects of the EPICA Dome C climate evolution are not well represented by our coarse-resolution model, nor are they captured by the zonal mean temperature (figure 6.2). In fact, as documented in figure 6.3, even our model simulation displays quite significant regional differences in the CO$_2$-induced warming over Antarctica.
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- The CO$_2$-induced warming and the bipolar seesaw-related warming of the Southern Ocean could amplify each other in a nonlinear manner, so that eventually only their combined effect can explain the full magnitude of the A-events.

6.4 Discussion and Conclusion

An AMOC collapse in the present simulation leads to an atmospheric CO$_2$ increase of $\sim$20 ppmv, about half of this CO$_2$ increase resulting from shrinking Northern Hemisphere tropical vegetation. This net CO$_2$ increase mostly raises temperatures in polar regions. In Antarctica, the warming due to this CO$_2$ increase adds up to the bipolar seesaw warming pattern and provides an explanation for the observed correlation between Heinrich events, atmospheric CO$_2$ variations and Southern Hemisphere warming during MIS3. In fact, the simulated CO$_2$-induced warming over Antarctica is comparable to the one generated by ocean heat content anomalies related to the bipolar seesaw mechanism. However, the simulated CO$_2$-induced temperature variations in the present transient model simulations can only explain about 25–30% of the observed temperature range of A-events. It is likely that the acceleration technique used for the GC experiment leads to some underestimation of the simulated variance.

In any case, the mechanism presented here contributes notably to the generation of millennial-scale glacial climate variability around Antarctica. It synchronizes climate events in both hemispheres by the radiative effects of CO$_2$ released mostly in the Northern Hemisphere. According to this mechanism, Heinrich events, A-events and the associated CO$_2$ changes are closely coupled. Given such a tight coupling, it may be possible to identify Heinrich events in other glacial periods simply by looking at joint suborbital temperature and CO$_2$ variations in Antarctica. Furthermore, the proposed mechanism is distinguishable from the bipolar seesaw mechanism in the sense that different timescales are involved. In the former, the typical response depends upon the time for the extratropical SST anomalies to reach the tropical Atlantic (Chiang et al., 2003; Krebs and Timmermann, 2007) and on the time for the vegetation to adjust; in the latter, the seesaw response time depends mostly on ocean dynamical and thermodynamical processes.

Several caveats need to be mentioned. As discussed in Menviel et al. (2008), the 20 ppmv CO$_2$ increase during the idealized pre-industrial AMOC shutdown originates from a delicate balance between terrestrial carbon sources and oceanic carbon sinks. This balance may be model as well as climate-state dependent. For example, a meltwater experiment similar to FC conducted with LOVECLIM under glacial conditions shows no net atmospheric CO$_2$ increase (Menviel et al., 2008), which is at odds with observational records (figure 6.1). This absence of CO$_2$ increase is due to the reduced total amount of vegetation under glacial conditions. Clearly, more modeling work with more comprehensive climate-carbon cycle models needs
to be done to better understand and constrain the mechanisms that led to the suborbital CO₂ variations during glacial periods.

The shorter-term and weaker warming events that occurred between 39-43 ka B.P. (figure 6.1) correspond to the Dansgaard Oeschger events 9-11. According to the most recent CO₂ data (Indermühle et al., 2000), these events were not accompanied by significant CO₂ variations. Whether this can be explained in terms of insufficient temporal resolution of CO₂ records is still an open question. An alternative explanation could be that these warming events are simply manifestations of the bipolar seesaw response to either an internally generated (Timmermann et al., 2003) or an externally forced (Braun et al., 2005) AMOC weakening. The resulting hydrological changes during these events may have been insufficient to release significant amounts of CO₂. Higher resolution CO₂ and δ¹³CO₂ data are needed to further elucidate the mechanisms responsible for millennial-scale Antarctic warming.

Bibliography


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Chapter 6. CO$_2$ Release from Northern Hemisphere Vegetation Contributes to Milennial-scale Antarctic Warming Events
Chapter 7

Summary, discussion and conclusions

The aim of this thesis was to study climate-carbon cycle interactions on millennial and glacial-interglacial timescales. The following hypotheses were tested:

A. Changes in the southern hemispheric wind strength during glacial-interglacial transitions are a main driver of glacial-interglacial atmospheric CO\(_2\) variations.

B. A shut down of the Atlantic Meridional Overturning Circulation during Heinrich events led to an increase in atmospheric CO\(_2\). This CO\(_2\) increase was primarily due to a release of carbon from the terrestrial biosphere.

C. The increase in atmospheric CO\(_2\) during Heinrich events enhances southern hemispheric warming.

7.1 Summary

A. Despite the importance of atmospheric CO\(_2\) variations on glacial-interglacial timescales, the mechanism responsible for these variations is still poorly understood. Toggweiler et al. (2006) (herein referred as T06) recently proposed an interesting new idea, in which weaker southern hemisphere westerly winds during glacial times lead to a weakening of the upwelling of DIC-rich waters and thus to a reduction in the amount of CO\(_2\) released to the atmosphere in this area.

To test T06 hypothesis, the response of the climate and the marine carbon cycle to changes in the wind strength over the Southern Ocean is studied using an EMIC. Weaker southern hemispheric westerlies induce a reduction in the upwelling of Circumpolar Deep Water (CDW), thus reducing the amount of DIC brought to the surface. This leads to an enhanced CO\(_2\) sink in the Southern Ocean. However, due
Chapter 7. Summary, discussion and conclusions

to the reduced upwelling, there is a decrease in the surface waters nutrient content. This negative nutrient anomaly is incorporated into the Sub-Antarctic Mode Water and advected to low latitudes. The export production is thus lower in almost the entire southern hemisphere, which reduces the amount of DIC transformed in organic matter and therefore counteracts the effects of the increased CO$_2$ sink in the Southern Ocean. As a result, a weakening of the southern hemispheric westerlies by about 15% only leads to a small decrease in atmospheric CO$_2$ of only about 5 ppmv. However, when the export production is kept constant, an atmospheric CO$_2$ decrease of 20 ppmv is obtained.

**B.** Millennial-scale experiments, designed to represent idealized Heinrich events and a collapse of the Atlantic Meridional Overturning Circulation (AMOC), were performed by adding freshwater in the northern North Atlantic. A shutdown of the AMOC leads to a cooling in the northern hemisphere and a slight warming in the southern hemisphere. Due to the increased equator to pole temperature gradient in the northern hemisphere the trades strengthen and the Inter Tropical Convergence Zone (ITCZ) is pushed southwards. Due to the colder and drier conditions in the northern hemisphere, there is a vegetation shift to biomes more adapted to the climatic conditions. This leads to a decrease of the terrestrial primary production and a CO$_2$ release to the atmosphere. A part of this CO$_2$ is taken up by the ocean. Due to the globally colder and fresher waters, CO$_2$ solubility increases and the reduced circulation leads to a greater DIC storage in the deep sea. The net effect on atmospheric CO$_2$ is an increase of about 20 ppmv under pre-industrial conditions, while under glacial conditions atmospheric CO$_2$ decreases by 10 ppmv.

**C.** Greater atmospheric CO$_2$ during Heinrich events enhances the warming in the southern hemisphere. A warming of the southern hemisphere in response to a shut down of the AMOC has been previously described by different studies but the mechanism to generate such a warming is still a matter of debate. By comparing simulations with different atmospheric CO$_2$ content, it is suggested that an atmospheric CO$_2$ increase of 20 ppmv can explain at least 30% of the observed high southern latitude warming during the so-called A events.

### 7.2 Discussion

**Changes in the strength of the southern hemisphere westerlies**

Chapter 4 of this thesis showed that changes in the strength of the southern hemisphere westerlies did not produce large variations in atmospheric CO$_2$ content. These results are in agreement with the studies of Winguth et al. (1999) and Tschumi et al. (2008), who obtained a weak atmospheric CO$_2$ increase in response to an en-
enhancement of the southern hemispheric westerlies. In both studies changes in export production play an important role in counteracting the changes in DIC transport to the surface.

It was demonstrated in Chapter 4 that a 15% weakening of the southern hemispheric westerlies could lead to an atmospheric CO$_2$ decrease of about 20 ppmv when export production was kept constant. Martin (1990) suggested that during glacial times, export production was enhanced in the Southern Ocean due to iron fertilization. As export production in the Southern Ocean is thought to be limited by the iron content, export production in this region could have been greater during glacial than during interglacial times even under weaker southern hemispheric westerlies conditions. However, during the Last Glacial Maximum (LGM), the summer sea ice coverage extended to about 55$^\circ$S and at present the High Nutrient Low Chlorophyll (HNLC) region in the Southern Ocean is located south of 40$^\circ$S. It is thus mainly in between 55$^\circ$S and 40$^\circ$S that iron fertilization could have compensated for the reduced amount of nutrients brought to the surface. Further experiments, using a model that includes iron limitation/fertilization, should be performed to study the impact onto the carbon cycle of changes in the southern hemispheric westerlies.

**North Atlantic meltwater experiments**

Chapter 5 described a new mechanism to generate an atmospheric CO$_2$ increase of about 20 ppmv following a shut down of the AMOC.

Using an AOGCM (MRI-CGCM2) coupled to a carbon cycle model, Obata (2007) (herein referred as OB07) obtained an atmospheric CO$_2$ increase of about 7 ppmv due to a shut down of the AMOC. This increase was due to a release of carbon from the terrestrial biosphere of about 25 GtC, while the oceanic reservoir took up 10 GtC. Qualitatively, the results presented in this thesis are in very good agreement with OB07. OB07 obtained a decrease of the terrestrial NPP in the northern tropics, while simulating a NPP increase in the southern tropics (figure 7.1). The NPP is also reduced at high northern latitudes, in agreement with the results presented in Chapter 5.

At the ocean surface, OB07 simulated a greater CO$_2$ solubility due to colder and fresher conditions. And, in agreement with the results presented here, the export production decreased globally (−19%), particularly in the Eastern Equatorial Pacific and the Eastern Equatorial Atlantic due to weaker upwellings (figure 7.2).

The integrated carbon reservoir anomalies obtained in OB07 were however smaller compared to the ones described in Chapter 5. This could be due to differences in the length of the freshwater input. The maximum length of the freshwater input in OB07 is 62 years compared to 200 years for this study. As a result, the AMOC shut down for ∼ 70 years in OB07 compared to ∼ 300 years in this study. 70 years might be too short to generate anomalies similar to the ones obtained in Chapter 5.
Chapter 7. Summary, discussion and conclusions

Figure 7.1: Terrestrial net primary production anomalies (gC/m²/yr) for a collapsed AMOC state compared to a pre-industrial control state for the experiments described in Chapter 5 (top) and for Obata (2007) study (bottom).
Figure 7.2: Export production anomalies (molC/m$^2$/yr) for a collapsed AMOC state compared to a pre-industrial control state for the experiments described in Chapter 5 (top) and for Obata (2007) study (bottom).
Using the UVic model, which comprises an Ocean General Circulation Model (OGCM) coupled to an Energy Balance Model (EBM) of the atmosphere, Schmitz et al. (2007) (herein referred as SC07) obtained an atmospheric CO$_2$ increase of about 20 ppmv due to a shut down of the AMOC. However, in contrast to the results presented in Chapter 5, this CO$_2$ increase was due to a carbon release from the oceanic reservoir, while the terrestrial carbon reservoir took up a part of that CO$_2$. As the atmospheric component of the UVic model is an EBM, the winds are prescribed, so that no changes in the position of the ITCZ were simulated. SC07 simulated a carbon release from the vegetation in northern Europe, while the terrestrial carbon storage increased everywhere south of 20°N and along the western coast of North America. This result is clearly at odds with all the other Heinrich events simulations which include a terrestrial carbon cycle (Scholze et al., 2003; Köhler et al., 2005a; Obata, 2007). The lack of a realistic atmospheric response to an AMOC shut down may explain the different vegetation response in the SC07 study.

The oceanic carbon release in SC07 is due to:

1. A decrease in CO$_2$ solubility caused by a surface ocean warming
2. A redistribution of Alkalinity and DIC at the surface.
3. A weakening of the stratification in the Southern Ocean which leads to a greater surface DIC content.

Similar freshwater experiments performed with CGCMs exhibit a global decrease in SST ranging from 0.2° to 1.2°C. A CO$_2$ solubility decrease due to warmer ocean surface is thus inconsistent with most of the other modeling studies.

A detailed comparison of the simulated alkalinity and DIC distributions in SC07 with our results, is rendered difficult by a lack of information provided in SC07. SC07 states that the global export production is reduced during the AMOC shut down but does not give exact numbers. Using LOVECLIM, a 14% decrease in global export production is obtained under pre-industrial conditions. However, in the experiments performed in Chapter 5, I find that, while the production of opal shells is strongly reduced, the carbonate precipitation remains constant. SC07 model only simulates carbonate precipitation and not opal production. A decrease in export production might thus induce a decrease in carbonate precipitation and thus a relative increase in DIC and alkalinity.

SC07 further suggested that weaker stratification in the Southern Ocean induced a surface increase in DIC and thus a greater atmospheric CO$_2$ content. On the contrary, in the results presented in Chapter 5, the simulated southern hemispheric westerlies were weakened during a shut down of the AMOC (figure 7.3). This induced a negative DIC anomaly at the surface of the Southern Ocean (figure 7.4) and thus an increase of the CO$_2$ sink. In agreement with the results of Chapter 5, OB07 simulated an intensified CO$_2$ sink in the Southern Ocean during a shut down of the AMOC.
7.3 Conclusions

To summarize, the climate response to a shut down of the AMOC obtained with LOVECLIM is comparable to the one obtained with CGCMs (Stouffer et al., 2006). The associated changes in the terrestrial carbon cycle are also in agreement with studies performed with a more comprehensive terrestrial carbon model (Köhler et al., 2005b). Finally, the results of Chapter 5 are in good agreement with a similar study performed with a coupled general circulation model including a terrestrial and marine carbon cycle model (Obata, 2007). A comparison of the results obtained in this study and the one obtained with the UVic model showed that the inclusion of wind changes may be critical to simulate the responses of the terrestrial and marine carbon cycles to freshwater perturbations. Even though LOVECLIM, is a model of reduced complexity, its sensitivity to an AMOC weakening is comparable to those obtained with more complex models.

7.3 Conclusions

The experiments performed in Chapter 5 of this thesis helped to propose a new mechanism to generate an atmospheric CO$_2$ increase of about 20 ppmv during Heinrich events. The meltwater experiments highlighted the important role of terrestrial biosphere dynamics in the carbon system and the necessity to include a terrestrial carbon cycle when studying climate-carbon cycle interactions. This new mechanism could be tested in future studies by performing meltwater experiments using a new version of LOVECLIM which includes terrestrial, marine and atmospheric...
Chapter 7. Summary, discussion and conclusions

Figure 7.4: Zonally averaged DIC anomaly in the Atlantic Ocean for the pre-industrial state without NADW formation compared to the pre-industrial control state.

δ¹³C. Comparing the atmospheric δ¹³CO₂ simulated with ice core records of δ¹³CO₂ would provide a strong test for this new hypothesis.

The work conducted in Chapter 6 further suggested that an atmospheric CO₂ increase during Heinrich events could explain a substantial part of the warming observed over Antarctica during A-events.

The experiments performed in Chapter 4 of this study showed the importance of compensating effects within the marine carbon cycle in response to climatic changes. These compensating effects prevent the generation of significant anomalies in the atmospheric CO₂ content on millennial timescales. For example, changes in upwelling strength involve two opposing mechanisms. Reduced upwelling of DIC-rich deep waters leads to negative DIC anomalies in surface waters. However, this also leads to a decrease in the amount of nutrients supplied to the euphotic zone and thus in export production. The experiments performed in this thesis suggested that this compensating effect plays an important role when varying the strength of the upwelling of the CDW in the Southern Ocean. It can also be of similar importance in the Eastern Equatorial Pacific and Atlantic.

However, the cause of the glacial-interglacial 80 ppmv changes in atmospheric CO₂ still remain unresolved. The work performed in this thesis added some constraints on possible mechanisms suggested to explain these atmospheric CO₂ variations:
7.3. Conclusions

1. McManus et al. (2004) suggested that the AMOC was reduced by about 30% during glacial times. These observational data combined with the modelling results presented in this thesis would suggest that potential changes in North Atlantic Deep Water production during glacial times did not have a significant impact on atmospheric CO$_2$ content.

In addition, Marchitto et al. (2007) suggested that during the last deglaciation, a deep-ocean carbon reservoir well isolated from the atmosphere, was suddenly injected into the intermediate waters of the Pacific. Experiments from Chapter 5, showed that this could not be achieved during the resumption of the AMOC. Indeed, even if the DIC content in the deep-ocean reservoir increases when the production of NADW is reduced, this positive DIC anomaly is then slowly advected away. In LOVECLIM, diffusion and advection do not allow for strong DIC gradients to persist, so that no significant DIC anomalies are observed in the intermediate or surface waters during the resumption of the AMOC. So, unless there is a major misrepresentation of the mixing scheme in LOVECLIM, the hypothesis of a release of very old carbon rich water mass seems difficult to reconcile with the modelling results presented here.

2. Other recent hypotheses to explain glacial-interglacial CO$_2$ changes involved the Southern Ocean (Toggweiler et al., 2006; Watson and Garabato, 2006). It has been shown in Chapter 4 that changes in southern hemisphere wind strength only, cannot account for significant atmospheric CO$_2$ variations. However, the impact on the carbon cycle of a strong cooling and sea ice advance in the Southern Ocean has not been studied in detail. In addition, iron fertilization is not included in LOVECLIM. The glacial southern hemisphere sea ice advance coupled to an increase in export production at around 40$^\circ$S due to iron fertilization could potentially have a significant impact on the atmospheric CO$_2$ content. Future experiments should include more detailed studies of the impact of strong changes in southern hemispheric climate onto the carbon cycle.

This study has demonstrated the importance of climate-carbon cycle interactions in generating millennial-scale variability. While this thesis focused on understanding climate-carbon cycle interactions for past events, it is also relevant for present and future assessments. In the 2007 Fourth Assessment Report, the IPCC analyzed results from a number of state-of-the-art coupled ocean and atmosphere climate models and concluded that the AMOC is very likely (90% possibility) to slow down (0-50%) during the 21st century. The results presented in Chapter 5 of this study addresses this issue by studying the impact on the carbon cycle of a potential slow down of the AMOC.

In addition, the potential marine carbon cycle response to changes in the strength of the southern hemispheric westerlies was investigated. It may help to understand
the implications on the carbon cycle of the poleward shift and intensification of the winds over the Southern Ocean, that has been observed since the 1970’s (Thompson and Solomon, 2002).

Bibliography


