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Chapter 1

Introduction

1.1 The Present-Day Climate Perspective

We are currently experiencing a rapid man-made change in climate, which confronts us with unprecedented challenges posing risks for humans as well as the natural systems (IPCC 2014; Pachauri et al., 2014). The consequences of our continuous impact on the Earth and its climate are already evident in rising global mean temperatures, altered precipitation patterns and melting ice caps. Different future scenarios, which vary in the amount of emitted greenhouse gases, are used in various kinds of numerical models to achieve an estimate on how the climate system may develop in the future and how this will impact our society and living conditions (Pachauri et al., 2014). This information is needed to develop strategies on how to adapt our society to future climate conditions.

The Earth's climate is a highly complex system and to accurately predict the coming changes, we need to understand how the different climate components (atmosphere, hydrosphere, cryosphere, land surface, biosphere, Fig. 1.1) react to imposed changes such as enhanced greenhouse gas emissions. Another difficulty of correctly predicting the impact of changing climate conditions on humans and nature lies in local characteristics such as orography, cloud cover or surface albedo (the reflecting power of a surface) that regionally alter the prevailing conditions. Numerical models of different complexities (Section 1.2.2) are used to estimate the development of global and regional climate, but the quality of the predictions strongly depends on our understanding of the interactions between the single climate's components and the mechanisms that alter the climate conditions.

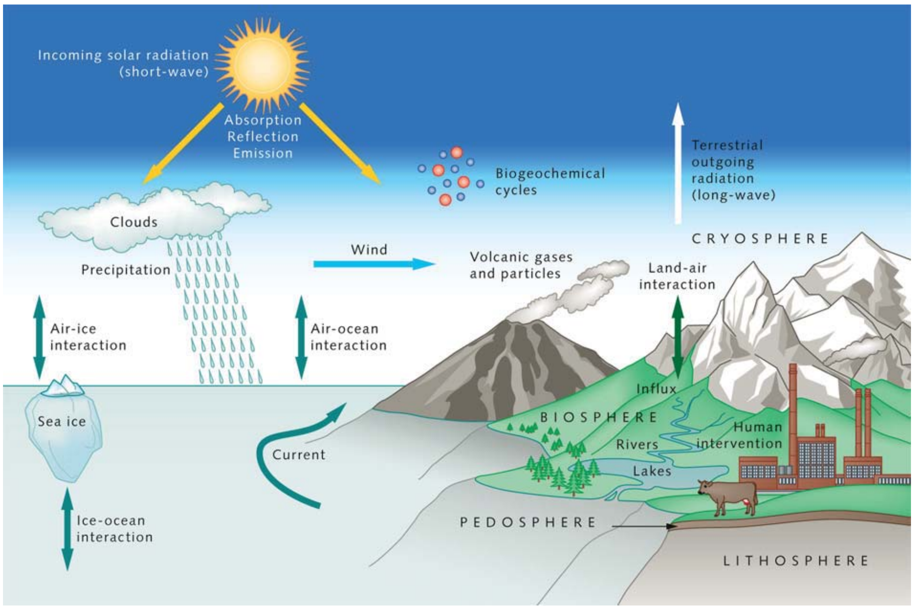

Figure 1.1: The climate system and its sub-systems and relevant processes and interactions (source: `http://worldoceanreview.com/en/wor-1/climate-system/earth-climate-system/`)

1.1.1 Forcing and Feedback Mechanisms of the Climate System

Climate, as we experience it, constantly reflects the effects of the factors that drive it (forcings) and the response of the different climate components to them. Climate forcings impose changes in the planetary energy balance that cause altered global temperatures (McGuffie and Henderson-Sellers, 2014). The most important forcings during the past probably were changes in the incoming solar radiation, changes in the atmospheric composition (e.g. carbon dioxide, methane, nitrous oxide) and volcanic eruptions. The incoming solar radiation varies due to two different reasons. First, the Sun itself experiences changes in its activity. Thus, the incoming solar radiation is not constant over time, instead it varies in the well-known 11-, $\sim$ 88- and $\sim$ 205-year periodicities (Damon and Sonett, 1991) with estimates ranging from $\pm$ 1 W m^{-2} (Steinhilber et al., 2009) to $\pm$ 5 W m^{-2} (Shapiro et al., 2011). Second, the incoming solar variation is altered by the Earth's orbit, which alters the received radiation and consequently affects the prevailing temperatures. Also greenhouse gases directly alter the prevailing temperatures because greenhouse gases absorb the longwave radiation from the Earth

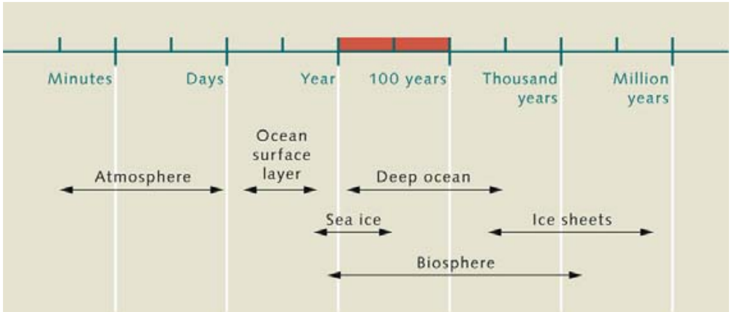

Figure 1.2: Different components of the climate system react to perturbations at different rates. The deep ocean, for example, is an important cause of the slow response of climate. The colored area on the top scale represents the short time span of a human life (source: `http://worldoceanreview.com/en/wor-1/climate-system/earth-climate-system/`)

and re-radiate it. Therefore, an increase in greenhouse gases causes increased warming due to the trapping of longwave energy in the climate system. In contrast, strong volcanic eruptions emit particles into the stratosphere, which alter the atmospheric composition and prevent the incoming solar radiation to reach the Earth's surface, the latter effect causing colder conditions at the surface. It is important to realize that variations in the incoming solar radiation, due to volcanic eruptions or the variations in the Sun's activity, and the astronomical parameters, is forcings that is not influenced by the climate system itself (external forcing).

Any change imposed on the climate system causes it to react, but the response time varies strongly between the different components, from minutes to days in the atmosphere, up to thousands of years in the cryosphere (Fig. 1.2, Bosch et al., 2010). Moreover, the feedback between the components can either enhance (positive feedback) or reduce (negative feedback) the forcing. The feedback mechanisms at work are most likely the processes behind the nonlinearities in the climate (Rial et al., 2004). The fact that the Earth's climate system is a nonlinear system that is characterized by multiple equilibria and internal thresholds (Stocker and Schmittner, 1997) enables the occurrence of rapid changes (Rial et al., 2004) that are evident under different climate conditions such as glacials and interglacials.

In general, the Earth's polar regions display a more pronounced response to climate changes than the global mean due to internal positive feedbacks and the ice sheets' volume and extent was altered substantially

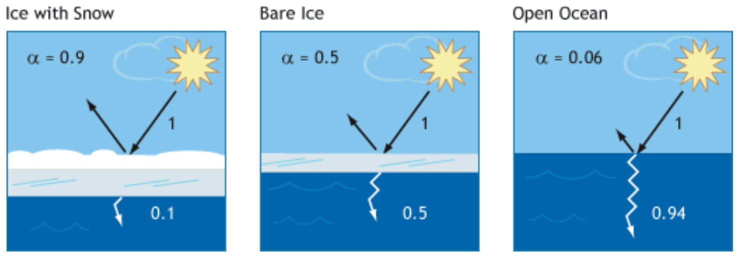

Figure 1.3: The ice-albedo feedback, the numbers correspond to the percentage of incoming radiation that is reflected and that enters and warms the ocean, for example α=0.9 means that 90% of the incoming solar radiation is reflected and 10% (0.1) enter the ocean (The National Snow and Ice Data Center, `https://nsidc.org/cryosphere/seaice/processes/albedo.html`)

from glacial to interglacial cycles. This enhanced sensitivity to changed climate conditions is called Polar Amplification (Manabe and Stouffer, 1980; Holland and Bitz, 2003) and describes the fact that any global change in the net radiation balance results in a stronger temperature change at the polar regions than globally. This is important because the Arctic and Antarctic have a strong impact on global climate, since these regions are dominated by cold conditions and the presence of ice, snow and water (Anisimov et al., 2007).

One major positive feedback mechanism at play in the polar regions is the so-called ice - albedo feedback (Fig. 1.3). It is associated with a decrease in the areal cover of snow and ice due to increased air temperatures. The higher temperatures cause the sea ice to melt, thereby increasing the local albedo. This is because the albedo of sea ice is in the range of 0.5 - 0.9, therefore most of the incoming radiation is reflected, but the albedo of the open ocean is much lower (0.1, Fig. 1.3). Thus a decrease in sea ice causes increasing sea surface temperatures that further enhance the melting of sea ice and decrease the areal cover of snow and ice (Curry et al., 1995). This positive feedback is most likely one of the main reasons for the amplified climate change at the polar regions compared to the rest of the globe (Holland and Bitz, 2003). Apart from sea ice, also the present ice sheets over Antarctica and Greenland strongly react to changing conditions and thereby influencing local and global climate. When an ice sheet is shrinking, its decreasing topography is causing higher air temperatures, altered wind directions and storm tracks (altitude effect, Ridley et al., 2005). Moreover, the positive ice-albedo effect will accelerate the shrinking of the ice sheet (Vizcaíno et al., 2008). The exposed land has a much higher heat uptake than ice because of the lower surface albedo (ice: 0.7 - 0.9, bare soil: $\sim$0.2).

The decreasing ice sheet thickness is directly related to enhanced freshwater fluxes in the form of ice discharge (icebergs) and runoff (surface and basal melting). The freshwater fluxes coming from Greenland currently increase global sea levels by 0,57 $\pm$ 0,1 mm yr^{-1} as measured in 2006 (Rignot and Kanagaratnam, 2006). Both, the runoff and ice discharge alter the ocean's stratification because on the one hand, the fresh melt water causes lower sea surface salinities, thereby increasing stratification, and on the other hand the implementation of cold water decreases the ocean's surface temperature, thereby decreasing stratification. The ice sheet's runoff enters the ocean at the coast. Icebergs, however, transport the freshwater further away from the shore, thereby freshening the ocean over a wider area. Moreover, they take the heat needed to melt from the ocean, thus cooling it at various depths. The resulting oceanic cooling and freshening facilitates the formation of sea ice (Jongma et al., 2009, 2013; Wiersma and Jongma, 2010; Hunke and Comeau, 2011; Bügelmayer et al., 2015a). Further, the icebergs' melt water enhances the ocean's stratification because it decreases the salinity and thereby the density of the water, which prevents it from sinking. Depending on the position and the strength of the applied freshwater flux, this process can cause a reduction of the Atlantic Meridional Overturning Circulation (AMOC, Swingedouw et al., 2009; Roche et al., 2010).

The AMOC is a globally operating ocean circulation system that is characterized by a northward flow of warm, salty water in the upper layers of the Atlantic and a southward flow of colder water in the deep Atlantic (Fig. 1.4). It plays an important role in global climate because it transports water masses, salt and heat over the globe (Broecker, 1991; Johns et al., 2011). At the moment, the AMOC transports 25% of the total global ocean-atmosphere northward heat flux and a water volume of 18.7 Sv (1 Sv = 10^6 $\mathrm{m}^3\,\mathrm{s}^{-1}$) at 26.5° N (Trenberth and Caron, 2001; Ganachaud and Wunsch, 2003; Johns et al., 2011; Rayner et al., 2011). Since the AMOC transports surface waters from the Southern Hemisphere (SH) across the equator into the high latitudes of the Northern Hemisphere (NH), the North Atlantic region experiences much warmer conditions than the North Pacific at similar latitudes (Rahmstorf and Ganopolski, 1999). In the NH and SH these surface waters mix downward at the deep-water formation regions forming the North Atlantic Deep Water (NADW) and Antarctic Bottom Water (AABW), respectively. In the NH the deep water formation sites are currently found in the Nordic Seas and Labrador Sea (Marshall and Schott, 1999; Pickart

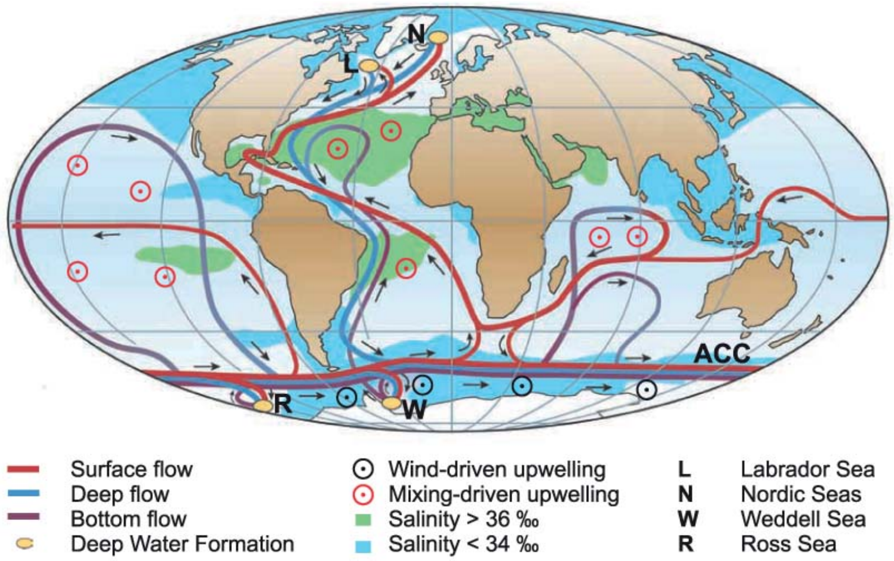

Figure 1.4: Strongly simplified sketch of the global overturning circulation system. In the Atlantic, warm and saline waters flow northward all the way from the Southern Ocean into the Labrador and Nordic Seas. By contrast, there is no deepwater formation in the North Pacific, and its surface waters are fresher. Deep waters formed in the Southern Ocean become denser and thus spread in deeper levels than those from the North Atlantic. Note the small, localized deepwater formation areas in comparison with the widespread zones of mixing-driven upwelling. Wind-driven upwelling occurs along the Antarctic Circumpolar Current (ACC). After Rahmstorf (2002).

et al., 2003) and in the SH they are situated in the Ross and Weddell Sea (Kuhlbrodt et al., 2007, yellow dots, Fig. 1.4). Deep-water formation is a complex process that depends on prevailing wind-related surface stresses, on high-density surface waters and a low vertical density gradient and thus a low buoyancy of the surface waters (Ganopolski and Rahmstorf, 2001; Kuhlbrodt et al., 2007). The density of the surface waters can be increased due to strong cooling whenever the ocean loses heat to the atmosphere or due to increasing salinity. Salinity is enhanced by the inflow of saline waters from lower latitudes, at locations where evaporation exceeds precipitation or by brine rejection due to the formation of sea ice. As mentioned above, freshwater fluxes coming from the ice sheets (calving and runoff) directly impact the AMOC depending on the location and amount of applied fluxes. The added freshwater on the one hand cools the surface waters (increasing its density) and on the other hand decreases its salinity (decreasing its density). Modelling studies indicate that the decreasing salinity causes a more stabilized water column thereby preventing deep mixing and weakening the AMOC (Swingedouw et al., 2009; Roche et al., 2010).

Overall, it is difficult to assess and predict the changes of the polar regions and their impact on global climate in the future because of the different processes involved. However, investigating their evolution over the past millions of years offers the possibility to gain better insight into the mechanisms at work. Ice sheets and the climate in general have undergone strong changes over the past (Gradstein et al., 2004). By studying and improving our understanding of the causes and consequences of these past climate (or paleoclimate) changes, we can make better predictions of the response of the climate system to the present and especially future conditions.

1.2 The Paleoclimate Perspective

The climate over the past 2.6 million years (the Quaternary Period) can be crudely divided into glacial and interglacial cycles. Since the Mid-Quaternary, about 800 ky BP (800,000 years Before Present), glacial cycles occurred periodically. M. Milankovitch (1920) proposed that the Earth's astronomical parameters, with periodicities of 21 ky, 41 ky and 100 ky (Fig. 1.5), are responsible for the occurrence of glacial cycles due to their effect on the seasonal incoming solar radiation. His theory was confirmed by Hays et al. (1976), who showed that the periodicities of the Earth's astronomical parameters could be found in marine sediments. Moreover, Imbrie et al. (1993a,b) displayed a direct relationship between the computed periodicities of the eccentricity (100 ky), obliquity (41 ky) and precession (23 ky, Fig. 1.5) and the occurrence of glacial cycles. Recent studies however have demonstrated that a combination of the non-linear interactions between Earth's astronomical parameters, the ice sheets and oceans is needed for the explanation of glacial-interglacial cycles (Kukla and Gavin, 2004; Wunsch, 2004; Crucifix, 2013).

In the NH extensive ice sheets covered North America, Eurasia and Greenland during glacial cycles. For instance, during the last glacial maximum (LGM, 21 ky BP), sea level decreased by up to 120 m relative to present day (Siddall et al., 2003) and atmospheric greenhouse gas concentrations were much lower than today (Flückiger et al., 1999; Dällenbach et al., 2000; Monnin et al., 2001). In contrast, during warmer

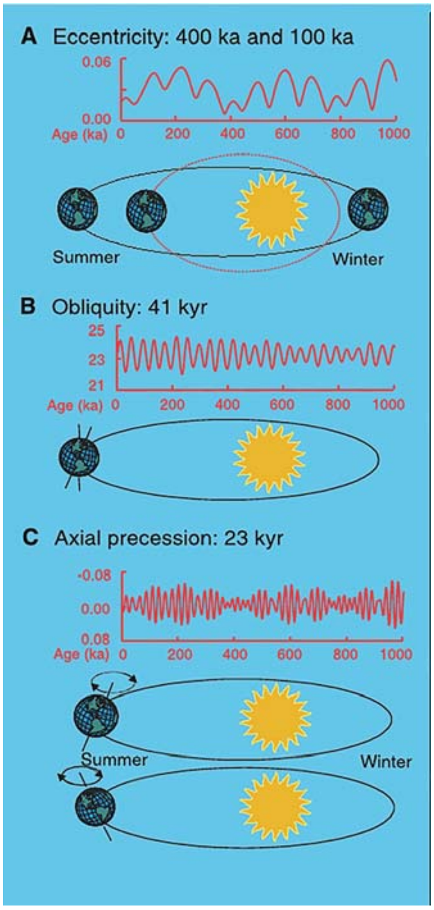

Figure 1.5: Primary orbital components: **(a)** Eccentricity with a periodicity of 400 and 100 ky refers to the shape of the Earth's orbit around the Sun and varies from near circular to elliptical; **(b)** Obliquity with a periodicity of about 41 ky refers to the tilt of the Earth's axis relative to the place of the ecliptic varying between 22.1°and 24.5°. A high angle is related to a stronger contrast between the seasons, especially in the high latitudes; **(c)** Precession with a periodicity of about 23 ky refers to the gradual shift in the orientation of the Earth's axis of rotation. In combination with the orbital eccentricity, it determines where on the orbit around the Sun seasons occur, increasing the seasonal contrast in one and decreasing it in the other hemisphere (Zachos et al., 2001).

interglacial cycles, such as the present one (the Holocene), global temperatures were similar or even higher than present day (Bintanja et al., 2005) as can be seen in paleoclimatic data.

In this thesis, we concentrate on the current interglacial, the Holocene, as well as on the last glacial period.

1.2.1 Proxy Data - Records of Past Climates

Evidence of past climate conditions is stored in natural archives that are influenced by the prevailing climate conditions (Bradley, 1999). Commonly used archives are tree rings, ice cores, and marine sediment cores. The isotopic ratio of oxygen, that is the ratio between the heavy (^{18}O) and the light (^{16}O) oxygen isotopes, measured in paleoclimatic data can be used to gain information about past temperatures because it depends on the temperature at which condensation occurs (Dansgaard, 1964). This method is used in ice cores from the Greenland and Antarctic ice sheets to reconstruct local temperatures. Further, the temperature and salinity of the past ocean can be reconstructed by collecting sediment cores from the ocean bottom. In these sediment cores the skeletons of benthic (bottom dwelling) and planktonic (surface dwelling, Peng and Broecker, 1984) foraminifera, small creatures that live in the ocean at various depths, give indication about the prevailing conditions at their lifetime. This is because the prevailing ocean temperatures and the isotopic composition of seawater during the formation of the foraminifera's shell mainly define the carbonate oxygen isotopic concentration of their shell (Urey, 1947; Shackleton, 1974). Therefore, the isotopic analysis of foraminifera can be used to reconstruct their living conditions and more importantly changes therein (e.g. Waelbroeck et al., 2002).

Moreover, ocean sediment cores recovered near Greenland and Antarctica include ice rafted debris (IRD). IRD come from the surrounding ice sheets and are transported by sea ice and icebergs (Bond et al., 1997; Andrews, 1998; Bianchi and McCave, 1999). When sea ice and icebergs melt, they slowly release the sediments that drop down to the bottom of the sea. Depending on the size distribution of the IRD, it is more probable that they have been transported by sea ice or by icebergs (Molnia, 1972; Ruddiman, 1977; Andrews et al., 2000). Therefore, the occurrence of IRD indicates the passing of icebergs at a specific site. Further, the sediment analysis of IRD gives information about the bedrock from where the icebergs calved, hence the calving locations. Thus, changes in ice sheet topography that alter the calving fronts as well as changes in the prevailing climate conditions that affect the amount of icebergs or their transport paths are displayed in the IRD records of the ocean cores (Andrews, 1998; Andrews et al., 2014).

Apart from the regional information that can be derived from proxies, the combination of archives from all over the world allows to identify common periods of warmer or colder conditions (Wanner et al., 2011). This would ideally enable scientists to reconstruct and, as a further step, understand the climate over the past thousands or even millions of years. Unfortunately, the proxy-based reconstructions are subject to various uncertainties. First, the main theory behind proxy-based reconstructions is that the relationship between a climatic variable (e.g. temperature) and a certain measurable variable (e.g. oxygen isotopic ratio) does not change over time. Thus, the present relationship is also applicable to the past and is not influenced by other components that might have changed over time such as atmospheric greenhouse gases or precipitation. Yet, those factors might play a role, therefore the assumption that the relationship between the climate and the measurable variable is stationary, is unlikely to hold (Birks, 1981). Moreover, the exact timing of the recorded evolution depends on the chosen age-depth model that is needed to transfer the core depths to an accurate time scale (Telford et al., 2004; Huybers and Wunsch, 2010).

1.2.2 Climate Models: Tools to Understand Climate

Proxy data and daily observations offer the possibility to reconstruct climate conditions and also its changes over the past weeks, years and millennia. But it is difficult to investigate the mechanisms that cause the evolution recorded in data. Climate models are therefore a valuable complementary method that allows to test and understand the impact of explicit forcing factors on climate or specific climate components.

Numerical models are based on mathematical equations that represent our understanding of the physical processes involved and atmospheric models have first been used for weather prediction purposes. However, increasing computer power enabled the complexity of the constructed models to grow. Therefore, ocean-, vegetation-, ice sheet- and coupled models emerged, with increasing spatial resolutions. It is important to notice that currently the models are limited by two factors, first, the available computer power that poses a boundary on the complexity of the model. Second, our knowledge of the physical processes involved in climate.

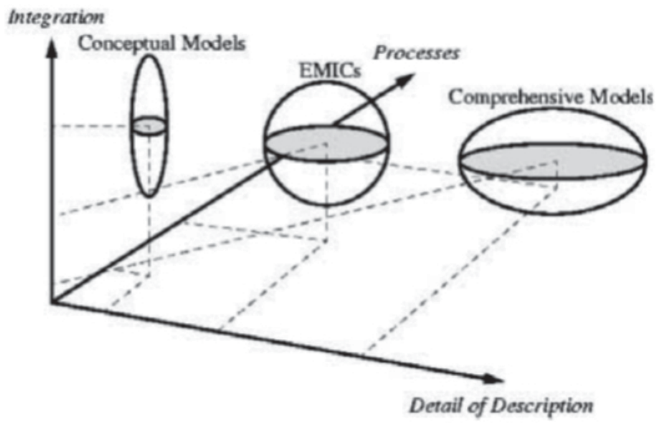

Figure 1.6: Pictoral definition of climate models arranged according to their complexity (incorporated processes), Claussen et al. (2002)

So far, different complexities of coupled models exist. The simplest approach are box models and conceptual models, second are the more complex Earth Models of Intermediate Complexity (EMIC) and finally the most complex are the General Circulation Models (GCM, Fig. 1.6, Claussen et al., 2002). The computation time of one model year directly depends on the resolution and complexity of the various model components (atmosphere - ocean - cryosphere - biosphere). EMICs mostly consist of simplified Navier-Stokes equations and a lot of small-scale processes, such as convection, are parameterized. This introduces biases in the model results compared to the observed present-day climate, but due to the generally coarser resolution experiments covering thousands of years can be easily performed. Therefore, also slow-feedback mechanisms in the ocean and ice sheets can be taken into account. Hence, EMICs are well suited to perform model experiments with long-term changing forcings, such as variations in the Earth's astronomical parameters or ice sheet topographies. Parameterizations are also necessary in GCMs, but small-scale processes can be explicitly represented because of the high resolution of the model components. This fact puts a natural time limit on the lengths of the experiments performed with GCMs and most studies are performed for a couple of hundred years, thereby often neglecting slow feedback mechanisms.

The aim of climate models is to help understand the mechanisms that cause the observed conditions and changes therein as well as to successfully predict future conditions. Proxy-based paleoclimate reconstructions offer the possibility to evaluate the model performance by comparing the simulated climate to these reconstructions. This approach is hindered by the fact that proxy records do not directly give temperature or precipitation. Instead, for example the measured isotopic ratio in the foraminifera's shell depends on both ocean temperatures and the isotopic composition of seawater. Therefore, including the explicit simulation of the isotopic content of the water cycle within a climate model facilitates the comparison between model and proxy output (Roche, 2013). Moreover, isotopic modeling offers the possibility to verify the hydrological cycle in models as errors therein are directly reflected in the simulated δ^{18}O values.

As described in Section 1.1, the ice sheets of the Northern and Southern Hemisphere play an important role in climate and underwent strong changes over the past. Therefore, three-dimensional thermo-mechanical ice sheet models have been developed to conduct longer-time simulations to investigate their evolution and interactions with the other climate components. These models account for changes in thickness and extension of the ice sheet, as well as for the position of the grounding line (e.g. Huybrechts and de Wolde, 1999; Ritz et al., 2001; Pollard and Deconto, 2009). The simplest approach to reconstruct past ice sheet geometries is to force an ice sheet model with proxy or model based time series of precipitation and temperature (e.g. Ritz et al., 2001). This method allows to evaluate how the ice sheet evolves according to the applied climate fields, but without affecting them (one-sided interaction). A two-sided interaction is provided when a climate model (EMIC, GCM) is coupled to an ice sheet model. In this case, the simulated air temperature and precipitation are used as input data for the ice sheet model and the resulting changes in topography, albedo and freshwater fluxes (calving and runoff) from the ice sheet affect the other climate components. In most modeling approaches taken so far, both the runoff and the calving from the ice sheet were applied either at the coastline (e.g. Vizcaíno et al., 2008; Bonelli et al., 2009; Goelzer et al., 2011) or given to the ocean at a pre-defined area (Ridley et al., 2005) to mimic iceberg transport.

Icebergs play an important role in the climate system due to their wide spread cooling and freshening of the ocean. Their impact has been tested with an iceberg model coupled to climate models and using prescribed calving locations and amounts based on climatological data (Bigg et al., 1996, 1997; Death et al., 2006; Levine and Bigg, 2008; Jongma et al.,

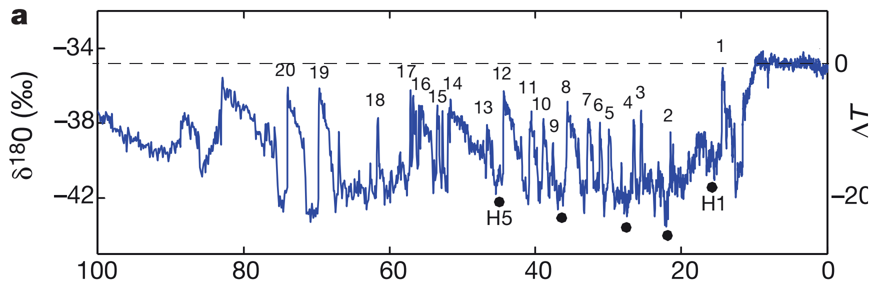

Figure 1.7: Abrupt climate changes in Greenland ice core data: record of δ^{18}O from the GRIP core, approximate relative temperature range is given on the right; the glacial climate is interrupted by Dansgaard - Oeschger warm events (numbers) and the timing of the Heinrich events H1-H5 is indicated by black dots (Ganopolski and Rahmstorf, 2001)

2009, 2013). Those studies showed that implementing icebergs has a different effect on the formation of sea ice, the AMOC's response and consequently on climate than applying direct freshwater fluxes (e.g. Jongma et al., 2009, 2013; Green et al., 2011). So far, iceberg models have been incorporated in EMICs and GCMs, but without being actively coupled to an ice sheet model. Therefore, the calving sites and amounts were either prescribed (e.g. Bigg et al., 1996, 1997; Death et al., 2006; Levine and Bigg, 2008; Jongma et al., 2009, 2013; Roberts et al., 2014) or according to the coastal sites defined by the river routing system of the climate model (Martin and Adcroft, 2010). The latter approach allows the background climate to define the number of icebergs generated, but under the assumption of an equilibrated ice sheet.

Overall, paleoclimatic data and climate models offer the possibility to gain insight into past climate conditions and changes therein.

1.2.3 The Last Glacial Period and the Occurrence of Heinrich Events

The last glacial period started around 115 ky BP and lasted until 11.700 years ago with the maximum ice sheet extent at about 21 ky BP. During these approximately 100,000 years several abrupt climate changes, termed Dansgaard - Oeschger and Heinrich events, occurred (Heinrich, 1988; Bond et al., 1992; Dansgaard et al., 1993; Grootes et al., 1993; Andrews, 1998; Ganopolski and Rahmstorf, 2001,Fig. 1.7).

Dansgaard - Oeschger (D-O) oscillations are marked by a rapid decrease in δ^{18}O ($\sim 10^\circ$ C increase) in Greenland within a few decades and a slow gradual increase until the δ^{18}O rapidly falls back to its stadial value (Fig. 1.7). Evidence of these abrupt warming events is strongest in data for Greenland and the North Atlantic. Nevertheless, synchronous warming has also been found in other locations, such as in the Mediterranean or in North America (Voelker et al., 2002). There are different theories on the cause of D-O cycles, but most involve changes in AMOC strength associated with variations in deep water formation in the North Atlantic, either due to internal oscillations or external freshwater forcing (Claussen et al., 2003 and references therein).

From about 50 - 20 ky BP Heinrich events occurred during some of the coldest intervals between the D-O events (Fig. 1.7). Heinrich events correspond to the discharge of huge armadas of icebergs from the Northern Hemisphere ice sheets as seen in ocean sediment cores as thick layers of ice rafted debris. Most of the IRD was found in the so-called Ruddiman - belt between 40°-55° N and 10° -60° W (Ruddiman, 1977) between Newfoundland and Europe (Grousset et al., 1993).

The exact mechanism behind Heinrich events is not yet clear, but currently three mechanisms are proposed: first, some studies indicated that catastrophic ice-shelf break-up is the driver behind Heinrich events (Hulbe, 1997; Hulbe et al., 2004). This theory indicates that ice shelves would collapse as soon as summer temperatures were high enough to cause surface melting. The formed melt water fills the crevasses of the ice shelf and slowly penetrates the crevasses to the bottom of the floating ice, causing the ice shelf to disintegrate (Hulbe, 1997; Hulbe et al., 2004). Second, the hypothesis of internal ice sheet variability (binge - purge mechanism) proposes that the ice sheet is first growing (binge stage) depending on the prevailing temperatures and the available moisture, until a combination of geothermal heat, advection of heat from the upper surface and internal friction at the base of the ice sheet destabilizes the ice sheet, causing it to rapidly collapse (purge stage; MacAyeal, 1989; Verbitsky and Saltzman, 1995; Hunt and Malin, 1998; Clarke et al., 1999). Third, some studies suggest that warming subsurface ocean temperatures triggered the breakup of ice shelves due to a prior AMOC slowdown (Hall et al., 2006; Jonkers et al., 2010; Alvarez-Solas et al., 2010; Álvarez-Solas et al., 2011; Alvarez-Solas et al., 2013). The first two mechanisms, ice shelf collapse and binge-purge, share that the increased iceberg melt flux strongly weakens the AMOC and thus globally alters the climate conditions (e.g. Broecker, 1994; Rahmstorf, 2002; McManus et al., 2004). The third mechanism however indicates that the slowdown of the AMOC happened prior to Heinrich events.

Heinrich events are clearly indicated in IRD records, but in the recorded calcite signal of planktonic foraminifera their occurrence is not as easy to identify because the $\delta^{18}O_{calcite}$ signal is a combination of the prevailing ocean temperatures and the $\delta^{18}O$ of the seawater. Due to the intrusion of iceberg meltwater, the ocean is cooled and isotopically depleted water is added (mean $\delta^{18}O$ of icebergs $\sim$ -30 ‰). Roche and Paillard (2005) showed in an isotope-enabled climate model that the simulated $\delta^{18}O_{calcite}$ signal near the Iberian margin displays only a small change during Heinrich event 4 ($\sim$40 ky BP) because of the compensating effects of decreasing temperatures and decreasing $\delta^{18}O_{seawater}$ values, which is also seen in ocean sediment core from that area (Shackleton et al., 2000). Moreover, it is difficult to assess the exact amount of freshwater released during Heinrich events, with suggested values ranging from 0.01 - 0.4 Sv (1 Sv= $10^6 m^3 s^{-1}$) over a period of 250 up to 1250 years (Hemming, 2004; Roche et al., 2004, 2014b; Roberts et al., 2014). Roche et al. (2004, 2014b) conducted hosing experiments, thus directly applying freshwater fluxes to the ocean's surface at pre-defined sites, using an isotope enabled global climate model to constrain the estimated duration and freshwater fluxes released during Heinrich event 4 ($\sim$40 ky BP) and 1 ($\sim$17 ky BP). This model set-up allows to directly compare the modeled $\delta^{18}O_{calcite}$ with paleoclimatic data. For Heinrich event 1, Roche et al. (2014b) used a climate model to display that the best fit between the modeled and observed $\delta^{18}O_{calcite}$ is found, if the AMOC is severely reduced, but not completely shut down and for Heinrich event 4 the authors limited the timing to 250 $\pm$150 years and a most probable freshwater flux of 0.29 Sv (Roche et al., 2004). Another approach to constrain the released freshwater fluxes was taken by Roberts et al. (2014), who explicitly computed the sedimentation rates of icebergs using an iceberg model embedded in a global climate model. The authors compared the simulated sediment thickness to IRD paleoclimatic data and found that the most likely scenario of a typical Heinrich event is over a period of 500 years with a constant melt water flux of 0.04Sv.

Other model studies concentrated on the different climate response to hosing experiments compared to explicitly modeling icebergs (Levine and Bigg, 2008; Green et al., 2011; Jongma et al., 2013). All studies found a different AMOC response to icebergs than to direct freshwater fluxes (Fig. 1.8a). Hosing experiments have a stronger impact on the ocean than icebergs, yet, this effect is weakened when implementing

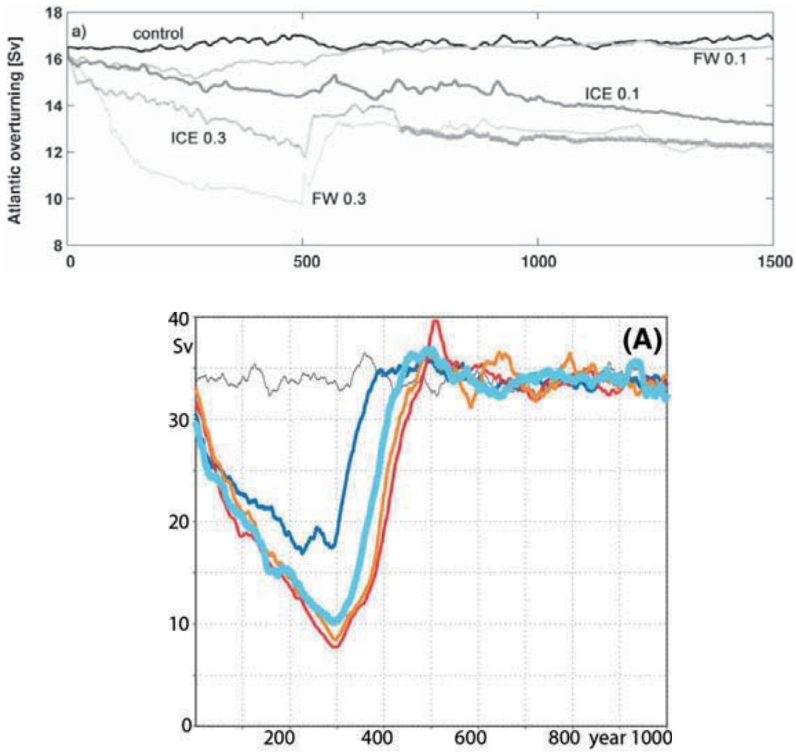

Figure 1.8: **(top)**Response of the Atlantic Overturning circulation to a 500 year freshwater flux of 0.1 and 0.3 Sv, respectively representing the Barents' Sea ice shelf collapse during MIS 6. The black line on top is the control experiment without any freshwater fluxes included, FW corresponds to direct freshwater fluxes (hosing experiment) and ICE to the explicit computation of icebergs (Green et al., 2011); **(bottom)**The yearly maximum meridional overturning circulation in the North Atlantic (Sv). C=Cooling, F=Freshening; dark blue: F-icebergs; light blue: CF-icebergs; red: F-hosing; orange: CF-hosing (Jongma et al., 2013)

icebergs that not only alter the salinity of the ocean (freshen), but also take up the heat needed to melt (cooling) the ocean (Fig. 1.8b).

1.2.4 The Holocene

The present interglacial, the Holocene, started about 11.700 years ago and is characterized by a relatively stable climate. This climatic stability has provided the opportunity to develop our current society (Wanner et al., 2011). It is often divided in three subperiods (Nesje and Dahl, 1993). The Early Holocene is the first period, spanning from 11.7 ky BP to 8.2 ky BP and is characterized by the large remnant ice sheet over North America. During the second phase, the Middle Holocene, from about 8.2 ky to 4.2 ky BP high summer insolation over the Northern Hemisphere caused warm conditions. Moreover, by then the ice sheet over North America was gone and therefore no longer cooled the regional climate. The summer insolation decreased again during the third phase (Late Holocene), starting around 4 ky BP until present (Walker et al., 2012). Besides these long-term orbital induced changes, 6 multi-decadal to century scale relapses to lower temperatures have been identified during the past 11.7 ky (Fig. 1.9; Wanner et al., 2011). These rapid climate change events were characterized by colder and dryer conditions, as well as glacier advances. Moreover, 7 periods of enhanced iceberg melt flux, so-called Bond events, have been detected in the North Atlantic and South-East of Greenland during the past ∼12,000 years (Bond et al., 1997, 2001) that partly coincide with the Holocene cold events (Fig. 1.9).

There are different theories of the mechanisms causing the Bond- and the Holocene cold events. (Bond et al., 2001) suggested that variations in the incoming total solar irradiance are responsible for the occurrence of increased iceberg discharge by causing lower ocean temperatures and a slightly altered atmospheric circulation pattern over Greenland. They found a significant correlation between the reconstructed solar radiation and the events of enhanced IRD recorded in the ocean cores close to the Greenland ice sheet. The impact of TSI on climate was tested in various models. For instance, Renssen et al. (2006) found that in their climate model the probability of a colder ocean state is higher during years of decreased TSI, with the ocean thus further enhancing the relatively weak external TSI forcing. Moreover, the stratospheric ozone formation depends on the incoming solar radiation, consequently lower stratospheric temperatures are present during periods of decreased TSI. These lowered temperatures propagate downward into the troposphere and can result in altered atmospheric circulation patterns (Haigh, 1994, 1996; Shindell

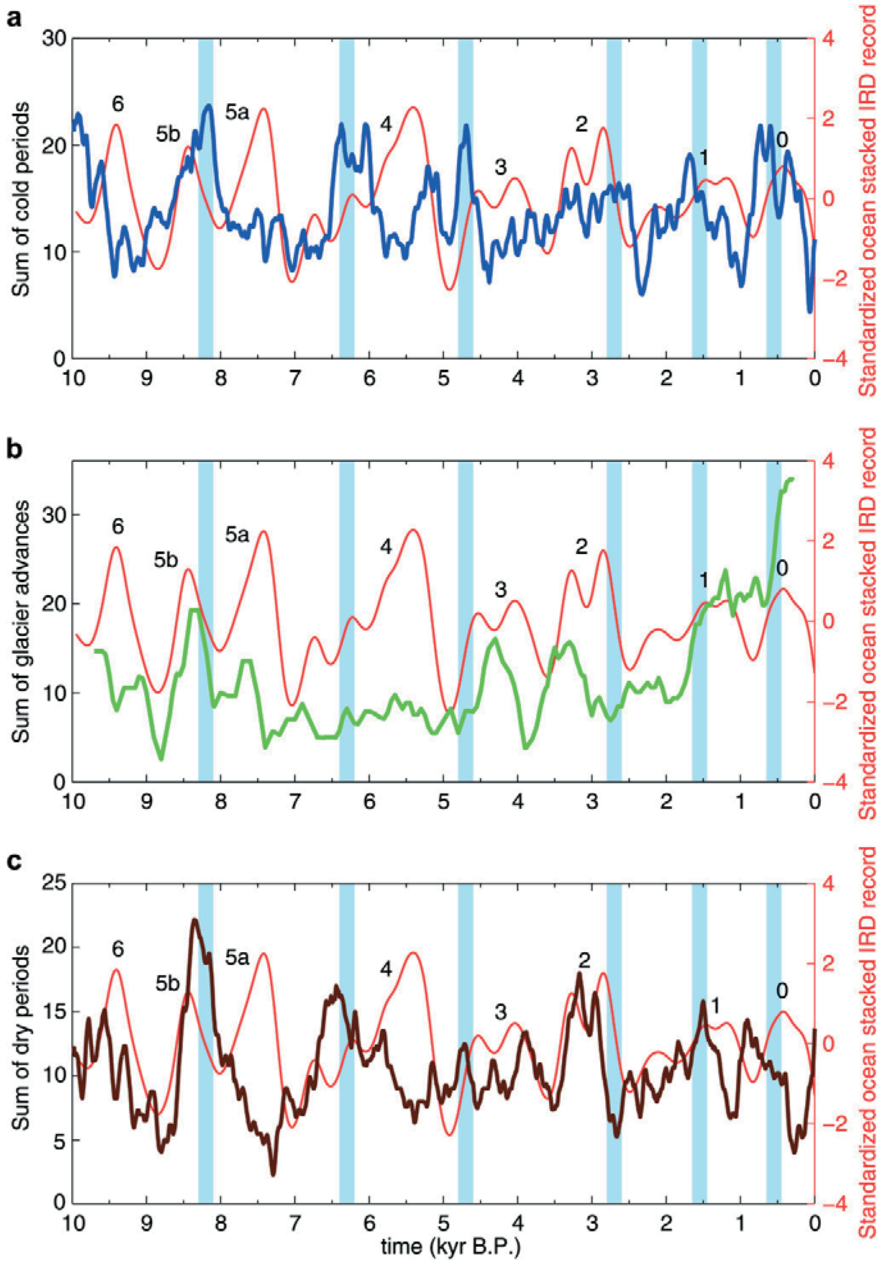

Figure 1.9: smoothed and weighted time series of globally distributed paleoclimatic data: **(a)** sum of cold periods; **(b)** sum of glacier advances; **(c)** sum of dry periods; red line corresponds to ocean stacked ice rafted debris record presented by Bond et al. (2001); blue bars denote the time of the cold events defined within the study of Wanner et al. (2011)

et al., 1999, 2001). Also large volcanic eruptions strongly impact the air temperatures and can cause a cooling of up to 0.2° C because the emitted sulfates decrease the incoming solar radiation (Robock, 2000). Recent modeling studies tested the impact of volcanic eruptions and variations in TSI on the climate of the past 1000 years and concluded that the climate responds to volcanic forcing, but its response to TSI is within internal variability (Jungclaus et al., 2010; Mignot et al., 2011). It has also been suggested that a combination of variations in incoming solar irradiance, volcanic eruptions as well as internal variations in ocean circulation are responsible for the Holocene cold events, rather than one globally active mechanism (Wanner et al., 2011). In contrast, Khider et al. (2014) argue that the climate system itself provokes the Holocene millennial scale variability, independent of external forcings.

1.3 The used climate model and the objectives of this thesis

1.3.1 The *i*LOVECLIM climate model

The climate model used in my thesis is the *i*LOVECLIM model. *i*LOVECLIM includes the same atmosphere, ocean and vegetation modules as LOVECLIM 1.2 (Goosse et al., 2010), but differs in the coupled ice sheet model and the implementation of isotopes, therefore *i*LOVECLIM. The name comes from the combination of the used models: the atmospheric **m**odel **E**CBilt, the oceanic model **CL**IO, the vegetation model **V**ECODE, the carbon cycle model **LO**CH and the ice sheet model GR**I**SLI (in LOVECLIM 1.2: ag **I**sm). *i*LOVECLIM was used because it offers a good trade off between resolution and complexity as it enables us to carry out multi-millennial simulations within reasonable time, but still includes a comprehensive ocean - sea ice model, a state of the art ice-sheet model and a relatively simplified atmospheric model. Moreover, (i)LOVECLIM has already been substantially tested for future (Driesschaert et al., 2007), Holocene (e.g. Renssen et al., 2006), Last Glacial Maximum (Roche et al., 2007) and Eemian (Duplessy et al., 2007).

This thesis was part of the AC^{2}ME project, which aimed at Unravelling the causes of abrupt climate change by means of δ^{18}O proxy forward modelling and model - data integration (NWO project proposal; Roche, 2009). Therefore, the coupling between all model components and the implementation of isotopes in all model components (atmosphere - ocean - ice-sheet - icebergs) was required and was done by Didier M. Roche

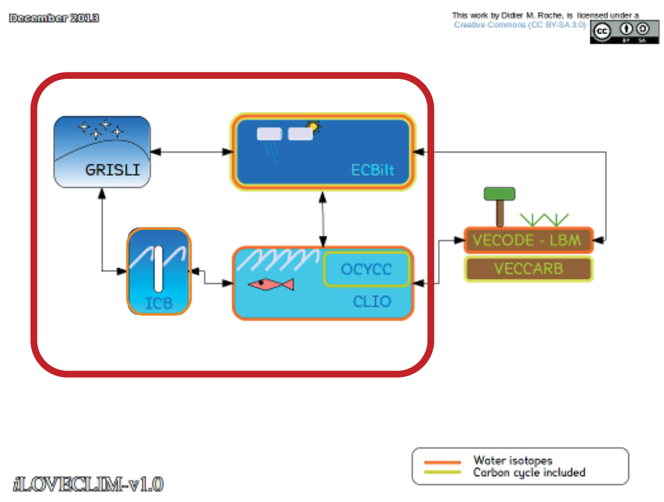

Figure 1.10: iLOVECLIM climate model: the yellow and orange bounding boxes indicate the inclusion of carbon and oxygen isotopes in the various model components. Abbreviations: ICB = iceberg module; OCYCC = ocean carbon cycle model, possible to activate in the ocean model CLIO; VECCARB = vegetation carbon cycle model, possible to activate in the vegetation model VECODE; LBM = Land Bucket Model included in VECODE; red box indicates model components that I actively worked on and (partly) further developed; Figure from Didier M. Roche, December 2013.

(coupling of ice-sheet to atmospheric model, isotopes in atmosphere - ocean model) and me (coupling of iceberg module to ice-sheet model, isotopes in iceberg module).

In all the experiments presented in this thesis, the atmospheric model ECBilt, the oceanic model CLIO and the vegetation model VECODE, as well as the iceberg module were used. Moreover, except for chapter 5, the ice sheet model GRISLI was actively coupled (Fig. 1.10). I will therefore concentrate on those components in the following sections.

1.3.2 The Atmospheric Model ECBilt

ECBilt was developed at the KNMI and is based on the quasi-geostrophic potential vorticity equation (Opsteegh et al., 1998). Since the ageostrophic terms in the vorticity equation and the advection of temperature by the ageostrophic wind are included, also the circulation at low latitudes is relatively well described. Moreover, the temperature is computed from the geopotential height and using the ideal gas law and the hydrostatic equilibrium (Goosse et al., 2010). ECBilt is run on a horizontal T21 truncation, which corresponds to a grid resolution of 5.6° in longitude and latitude. The potential vorticity and streamfunction are computed on three vertical layers, namely 800, 500 and 200 hPa. Temperature is computed at the surface and at 650 and 350 hPa. ECBilt consists of a thermodynamic stratosphere. The cloud cover is prescribed (ISCCP D2 dataset, Rossow et al., 1996) and precipitation is only computed in the lower most layer and is obtained using the available humidity at this level. Therefore, all the water that is transported by atmospheric winds above 500 hPa automatically precipitates. Roche (2013) implemented water isotopes in ECBilt, which are computed according to the precipitable water within the water column. All the processes affecting the isotopic composition (evaporation and precipitation) are taken into account. The surface albedo is computed according to the fractional use of the grid cells as ocean, sea ice, trees, desert and grass. ECBilt has an isotope enabled land-surface model included that computes the surface temperature and snow cover by taking into account the heat budget over a single soil layer with a spatially homogeneous heat capacity. To compute the soil moisture, a simple bucket model is implemented that takes into account evaporation, precipitation and snow melt. If the water content exceeds a defined threshold, the excess water is automatically transported to the corresponding ocean grid point. Moreover, if the snowfall at any grid location exceeds 10 m, it is removed and treated as excess-snow. The excess-snow is used to parameterize the effect of icebergs on the ocean. All variables (e.g. temperature and precipitation) are computed every 4 hours.

1.3.3 The Oceanic Component - CLIO

CLIO (**C**oupled **L**arge-scale **I**ce **O**cean model) is a comprehensive sea-ice model coupled to a free surface ocean general circulation model (Deleersnijder and Campin, 1995; Deleersnijder et al., 1997; Fichefet and Maqueda, 1997, 1999; Campin and Goosse, 1999). Its horizontal resolution is 3° in longitude and latitude and it consists of 20 unevenly spaced layers ranging from 10 m thickness at the surface to about 750 m for the deepest level (Goosse et al., 2010). The free surface allows the direct implementation of freshwater fluxes instead of parameterizing them as negative salt fluxes. The ocean flow is computed based on the Navier - Stokes equations and small-scale processes such as isopycnal mixing, vertical mixing and convection are parameterized. Moreover, a downslope parameterization is implemented to improve the representation of dense water flow (Campin and Goosse, 1999). In CLIO, water isotopes are implemented as passive tracers (Roche, 2013).

The sea-ice component consists of three vertical layers, 1 layer for snow and 2 layers of ice, and has the same horizontal resolution as the ocean model. One CLIO grid cell can consist of partial sea ice cover and open water. Further, the albedo formulation takes into account the state of the ice's surface (frozen or melting). Its time step to calculate the ocean variables (e.g. ocean temperature and salinity) is 1 day.

ECBilt interacts with CLIO by exchanging heat, momentum and freshwater fluxes. Flux corrections are only needed for the freshwater fluxes (precipitation, evaporation, river runoff and sea-ice melt) because ECBilt overestimates the precipitation over the Atlantic and Arctic Oceans and underestimates it over the North Pacific (Opsteegh et al., 1998). Therefore, the overestimated precipitation from the Atlantic and Arctic Oceans is dumped in the North Pacific to avoid a decreased Atlantic Meridional Overturning Circulation (AMOC) compared to present-day. Moreover, the excess snow computed in ECBilt is homogeneously distributed around Greenland and Antarctica, where the according heat needed to melt it is taken up from the ocean's surface layer to mimic the cooling effect due to iceberg discharge.

1.3.4 The Vegetation Model - VECODE

VECODE (**VE**getation **CO**ntinuous **DE**scription model) was explicitly developed to be coupled to a coarse resolution model to perform long-term simulations (Brovkin et al., 2002). It is a dynamic global vegetation model that computes changes in vegetation and terrestrial carbon pools. It takes into account three plant functional types (PFT), namely trees, grass and bare soil. For any given climate there is a unique stable composition of PFTs corresponding to the long-term averaged atmospheric fields. If the climate changes, VECODE computes the transition from the foregoing to the new stable composition (Goosse et al., 2010). The prevailing vegetation depends on the growing degree-days above 0° C and the annual mean precipitation that VECODE receives from EC-Bilt. The atmospheric model in turn is influenced by the vegetation cover that directly affects the surface albedo as well as by the leaf-area-index, which impacts the water storing capacity of the soil.

1.3.5 The Ice Sheet Model GRISLI

The **GR**enoble model for **I**ce **S**helves and **L**and **I**ce is a large-scale three-dimensional thermomechanical ice-sheet model. It was first designed for Antarctica (Ritz et al., 1997, 2001), but was further developed for the Northern Hemisphere (pey, 2007). GRISLI is computed on a horizontal 40 x 40 km grid resolution and distinguishes between three types of ice flow, first, slow flowing inland ice, second, fast flowing ice streams and third, ice shelves. The inland ice follows the 0-order shallow ice approximation (Hutter, 1983; Morland, 1984) whereas the ice streams and ice shelves are computed using the shallow-shelf approximation (MacAyeal, 1989).

Further, the ice sheet's thickness depends on the surface mass balance, ice discharge and basal melting. The surface mass balance is computed according to the accumulation minus the ablation and the latter one is computed using the Positive-Degree-Day method (PDD; Fausto et al., 2009). The isostatic adjustment of the bedrock to the ice load on top is governed by the flow of the asthenosphere with a characteristic time constant of 3,000 years and by the rigidity of the lithosphere. The temperature fields in the ice and the bedrock are calculated using a time-dependent heat equation.

Calving occurs whenever the ice thickness of the grid points at the ice sheet's margin is below a pre-defined value, 150 m in this thesis. If the ice sheet thickness next to an ocean grid is below 150 m and the upstream points cannot maintain the thickness above this threshold, the ice mass from this grid cell is cut off and considered as calved. The runoff from the ice sheet is computed at the end of the GRISLI model year by comparing the thickness at the beginning and at the end of the year and by taking into account the mass lost due to calving.

1.3.6 The Iceberg Model

In iLOVECLIM the optional dynamic - thermodynamic iceberg module has been implemented by Jongma et al. (2009) and Wiersma and Jongma

Table 1.1: Initial size classes

Class	Height (m)	Width (m)	Volume (m^3)	Percentage of total available Volume
1	67	67	$5.16*10^5$	0.15
2	133	133	$4.07*10^6$	0.15
3	200	200	$1.38*10^7$	0.2
4	267	267	$3.28*10^7$	0.15
5	300	333	$5.74*10^7$	0.08
6	300	400	$8.28*10^7$	0.07
7	300	500	$1.29*10^8$	0.05
8	300	600	$1.86*10^8$	0.05
9	300	800	$3.31*10^8$	0.05
10	300	1000	$5.18*10^8$	0.05

(2010) and in all the experiments performed within this thesis it was used with the same parameter set as in Jongma et al. (2009). The iceberg model is based on the iceberg-drift model published by Smith and coworkers (Smith and Banke, 1983; Løset, 1993) and was then further developed by Bigg et al. (1996) Bigg et al. (1997) and Gladstone et al. (2001). It was coupled to CLIO by Jongma et al. (2009) and Wiersma and Jongma (2010). All the icebergs are generated following a prescribed size distribution based on present day observations from one Greenland fjord (Tab. 1.1, Dowdeswell et al., 1992; Bigg et al., 1997).

The icebergs are moved on the CLIO grid and are transported by the Coriolis force, the air-, water- and sea-ice drag, the horizontal pressure gradient force and the wave radiation force. The latter defines how strongly the icebergs are moved by the incident waves. These forces depend on the wind and the ocean currents as computed in ECBilt and CLIO, which are then linearly interpolated from the surrounding grid corners to fit the icebergs location. Icebergs can ground and stay grounded until they have melted enough to be transported again. The icebergs melt due to basal melt, lateral melt and wave erosion. Since melting alters their length to height ratio, they are allowed to roll over. However, breaking up of icebergs is not included in the iceberg module. The heat needed to melt the icebergs is taken from the corresponding ocean layers. The melt water of the icebergs is added to the ocean's surface layer to mimick the complex process of upwelling of the iceberg's basal meltwater.

1.3.7 The model performance

As mentioned above, *i*LOVECLIM is a code fork of LOVECLIM1.2, which has been extensively tested for present-day conditions as well as key past periods. When compared to observations and paleoclimatic data, LOVECLIM satisfyingly captures the main characteristics of present-day, pre-industrial, mid-Holocene and Last Glacial Maximum conditions (Goosse et al., 2010, and references therein).

Nevertheless, there are some biases present in the model, especially at the low latitudes compared to present day observations due to the simplified atmospheric model included. There the temperature is overestimated and the precipitation between the two hemispheres is distributed too symmetrically. Especially in the northern hemisphere the position of the Intertropical Convergence Zone (ITCZ) is not well captured and situated near the equator. Moreover, the precipitation is overestimated in the subtropics causing a too extensive vegetation cover. The general structure of the atmospheric circulation is well captured, but the gradients between high and low pressure regions are too weak, resulting in too weak winds (Goosse et al., 2010).

1.3.8 The model developments concerning this thesis

To address the research question posed in the AC^{2}ME project "How can we combine δ^{18}O-based palaeoclimate records into a modelling framework to quantify the physical mechanisms behind recorded abrupt climate changes?" (project proposal, Roche 2009) certain model developments were needed. During my PhD, Didier M. Roche added the explicit computation of water isotopes in the atmospheric, oceanic and land surface model (Roche, 2013). Moreover, he coupled the ice sheet model GRISLI to the atmospheric model ECBilt (Roche et al., 2014a). As part of my PhD, I was responsible for coupling GRISLI to the iceberg model and for closing the water cycle between ECBilt - GRISLI - CLIO (Bügelmayer et al., 2015a). Further, I added water isotopes as passive tracers in the iceberg model. All the model developments concerning this thesis are described in detail below.

1.3.8.1 ECBilt - GRISLI

[1] A fully coupled climate - ice-sheet model includes the coupling of the atmospheric and the oceanic model to the ice-sheet model. This allows an interactive computation of the surface melting caused by the atmospheric conditions and the basal melting rate of the ice shelves depending on the prevailing ocean conditions. Unfortunately, the question of how to parameterize the melting / refreezing under ice shelves as a function of ocean temperatures, which is a very small-scale process with respect to our model grid, is still an ongoing research topic on its own (Beckmann and Goosse, 2003; Alley et al., 2008). Currently, basal melting is prescribed and depending on the local water depth (2m per year if the water depth is lower than 600 m, 5 m per year otherwise). But since all experiments presented in this thesis using GRISLI are concentrated on the Greenland ice sheet during the Holocene (6 ky BP up to pre-industrial), we do not expect significant ice-shelf areas. Therefore, in the current set-up the coupling is only between ECBilt and GRISLI.

ECBilt provides the monthly surface temperature and annual snow precipitation to GRISLI, to compute the surface mass balance (accumulation minus ablation). Due to the different grid resolutions (ECBilt: 5.6° in longitude and latitude versus 40 x 40 km in GRISLI) downscaling methods are needed. First, the monthly temperature provided by ECBilt is vertically downscaled to account for the differences in steep topographic regions (Fig. 1.11). This is done by taking into account the lowest and highest GRISLI points within one ECBilt grid cell and computing the local vertical temperature gradient between these points. This temperature gradient is then used to compute the altitude dependent GRISLI temperatures based on the ECBilt surface temperature. The downscaled temperature as well as the yearly snow precipitation are used to compute the surface mass balance, which is calculated following the Positive - Degree - Day method (PDD; Fausto et al., 2009). The PDD method accounts for all days with surface temperatures above the melting point to compute snow- and ice melt rates and takes refreezing of ice into account (Braithwaite, 1984; Reeh, 1991; Charbit et al., 2013).

After one GRISLI model year, the updated topography and ice mask are given to ECBilt. The ice mask is needed to compute the surface

[1] This section is based on the paper of Roche, D. M., Dumas, C., Bügelmayer, M., Charbit, S. and Ritz, C.: Adding a dynamical cryosphere to *i*LOVECLIM (version 1.0): coupling with the GRISLI ice-sheet model, Geosci. Model Dev., 7, 1377-1394, doi:10.5194/gmd-7-1377-2014, 2014.

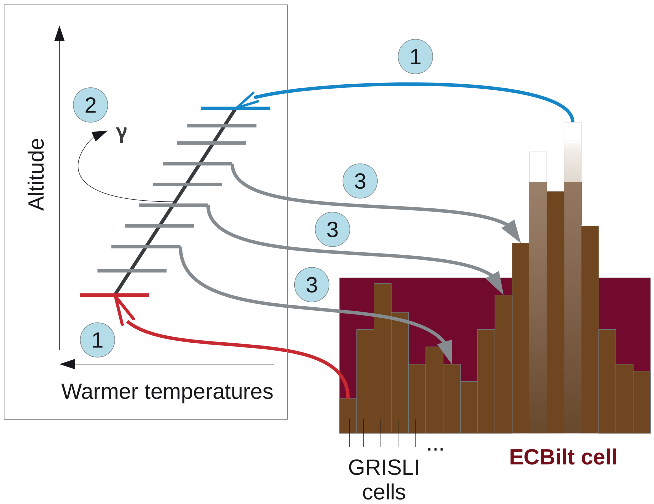

Figure 1.11: Scheme presenting the method used for the vertical downscaling. Numbering indicates the order of processing. (1) The temperature at the highest and lowest GRISLI point (tails of the blue and red arrows respectively) is retrieved for the given ECBilt grid cell (boundary in violet). (2) A vertical lapse rate is computed from these two extreme temperatures, using the line defined by the two temperatures and elevation extrema in addition to the ECBilt cell temperature. (3) Using γ, temperatures are derived for all altitudes in GRISLI (Roche et al., 2014a)

albedo and represents the mean over all GRISLI cells within one ECBilt cell. The ice mask is set to 1 if the ice thickness is above 50 m and to 0 otherwise, this method is used to eliminate small areas of ice that cannot be incorporated in ECBilt correctly.

The main difference between the used coupling method in iLOVECLIM and other climate models is that the absolute temperature and precipitation fields are given to GRISLI, instead of anomaly fields in respect to present day climate (e.g. Vizcaíno et al., 2008; Huybrechts et al., 2011). We chose to use absolute fields because the anomaly method assumes that model biases prevailing in a given climate are of the same order of magnitude as those present-day. The choice of absolute values thus enables the use of the model in different climatic contexts.

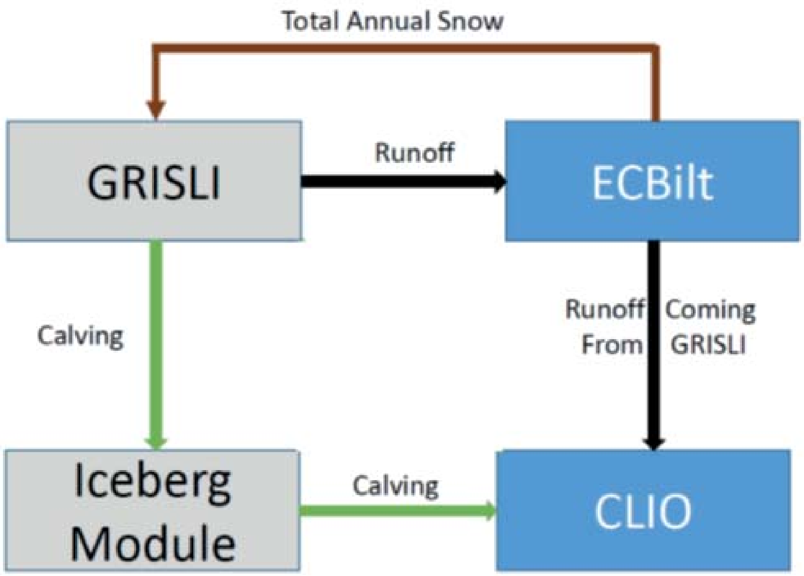

Figure 1.12: sketch of water cycle between ECBilt - GRISLI - iceberg module - CLIO

1.3.8.2 Closing the Water-Cycle

[2] An important issue of climate modeling is a fully closed water cycle. To achieve water conservation between GRISLI, the iceberg module and the climate components, all the water that is exchanged between the various model components has to be accounted for.

First, ECBilt provides the annual snowfall to GRISLI to compute the ice sheet's topography (Fig. 1.12). Second, GRISLI calculates the mass lost due to calving and due to runoff (surface and basal melting). Calving is explicitly computed and occurs whenever the ice thickness at the ice sheet's margin is below 150 m, a pre-defined threshold, and the ice flow from the points upstream does not maintain the height above this value.

The yearly runoff is computed by comparing the ice sheet's thickness at the beginning of the GRISLI model year and at the end of the model year and by taking into account the mass lost due to calving. This method has been chosen due to the initial set-up of GRISLI that was

[2]This paragraph is based on Bügelmayer, M., D. M. Roche, and H. Renssen. "How do icebergs affect the Greenland ice sheet under pre-industrial conditions?-a model study with a fully coupled ice-sheet-climate model." The Cryosphere 9.3 (2015): 821-835.

not designed to give the liquid runoff as output. The runoff is provided to ECBilt where the implemented routing scheme is used to distribute the water to the corresponding ocean grid cells. The calving is given directly to the iceberg module at the CLIO grid point corresponding to the calving position in GRISLI.

The iceberg module then generates icebergs following a monthly seasonal cycle, with the maximum of calving occurring in May and June and the minimum occurring in August. According to the daily available calving mass, icebergs of the ten size classes are generated, following a pre-described percentage distribution (Tab. 1.1). The melt water released by the icebergs alters the ocean's temperature and salinity. Moreover, we implemented isotopes in the iceberg model so that icebergs have a fixed isotopic value of -30 ‰. If the icebergs melt, the released volume of freshwater is added to the ocean's surface layer altering its salinity, temperature and isotopic composition.

After implementing the described developments, *i*LOVECLIM offers the possibility to perform long term experiments to investigate the mechanisms behind rapid climate changes occurring in the past with all the climate components included (atmosphere - ocean - biosphere - cryosphere). Moreover, by adding the water isotopes in the model, a direct data - model comparison is rendered possible.

1.4 Objectives of my thesis

As discussed in Section 1.2.2, climate models present the opportunity to test the impact of different factors on climate conditions as well as the interactions between the climate components. In this thesis the focus is on the Northern Hemisphere ice-sheets, especially on the impact of the icebergs being calved from Greenland during the Holocene and from the Laurentide ice sheet during the Last Glacial.

The effect of icebergs on the Southern Mid -to High latitudes as well as on the Northern Mid- to High latitudes has been investigated by various authors so far (e.g. Bigg et al., 1996, 1997; Gladstone et al., 2001; Jongma et al., 2009, 2013). Some groups forced the iceberg model with climatological data, others coupled it to a climate model and used the atmospheric (wind) and oceanic (ocean currents and sea ice) output as forcing fields. In these studies the authors prescribed the amount and location of the calved icebergs. A more consistent approach is to use the amount of precipitation that would cause the ice sheet to grow under equilibrated conditions as calving mass (Martin and Adcroft, 2010). Yet, none of the studies so far included an ice sheet model, thus the effect of icebergs on the ice sheet itself has never been investigated, nor if directly applied freshwater fluxes (hosing experiments) have a different effect. We therefore pose the following questions in chapter 2:

What is the impact of icebergs coming from the Greenland ice sheet on the climate of the Mid- to High latitudes and the Greenland ice sheet itself under pre-industrial equilibrium conditions? Do directly applied freshwater fluxes (hosing experiments) have a different effect?

The implemented parameterizations in climate models bring uncertainties that influence the obtained results. In the used iceberg model (Section 1.3.6) the size classes of the icebergs are pre-defined (Bigg et al., 1996) following present day observations from one fjord in Greenland (Dowdeswell et al., 1992). Moreover, the modeled wind and ocean currents define the resulting iceberg distribution and inaccuracies in these fields are directly imposed on the spatial pattern of the icebergs' melt flux. Hence, in chapter 3 we want to evaluate:

How sensitive are the modeled climate of the Northern Mid- to High Latitudes and the Greenland ice sheet to the icebergs' size distributions as well as variations in the spatial distribution of the icebergs' freshwater and latent heat fluxes under different climate conditions (pre-industrial, colder and warmer than pre-industrial)?

There are different theories on what caused the Holocene Bond events (periods of increased ice rafted debris). Bond et al. (2001) proposed that variations in the incoming total solar irradiance are responsible for the observed iceberg pulses. Other studies suggest that volcanic eruptions or a combination of the various forcings provoked these events. The impact of variations in TSI, as well as of volcanic eruptions on climate has been tested using climate models, but none of the climate models included an ice sheet or an iceberg model. Therefore, the direct effect of the varying forcings on these climate components has not been analyzed so far. In chapter 4 we consequently address the questions that still remain for us:

Can we reproduce the century to millennial scale iceberg events during the Holocene in our coupled climate - ice-sheet - iceberg model? What are the mechanisms behind these enhanced iceberg events?

During Heinrich events large armadas of icebergs were released from the Northern Hemisphere ice sheets as indicated by a steep increase in IRD in ocean sediment cores. Up to now, the modeling studies conducted on Heinrich events concentrated either on the effect of the freshwater fluxes (as icebergs or hosing experiments) on the climate and especially the AMOC (e.g. Levine and Bigg, 2008; Green et al., 2011; Jongma et al., 2013) or on constraining the estimated duration of Heinrich events and the released freshwater fluxes using isotopic or sediment modeling (e.g. Roche et al., 2004, 2014b; Roberts et al., 2014). But as mentioned above, the recorded $\delta^{18}O_{calcite}$ signal of Heinrich events is difficult to interpret due to the competing effects of lower ocean temperatures and the release of isotopic depleted iceberg melt fluxes. The implementation of isotopes in the ocean, atmosphere, land surface and also the iceberg module in iLOVECLIM offers the unique possibility to investigate (chapter 5):

What is the impact of the duration Heinrich event like iceberg discharges on the North Atlantic Ocean? How do changes in ocean temperatures and $\delta^{18}O_{seawater}$ due to the icebergs melt water impact the $\delta^{18}O_{calcite}$ recorded in proxies at various locations?