1. Introduction

1.1 Framework

1.1.1 Climate change

Traditionally climate is regarded as the constant aspect of our ever-changing daily weather. Stemming from the Greek word "klima", which refers to the inclination of the sun, climate can be defined as the condition of weather in some location averaged over a long period of time (the “climatic normal” being an average over at least 30 years [Henderson-Sellers and Robinson 1992], p.8). But climate itself is not constant and changes at various time-scales. For the past 2.58 million years [Gradstein et al. 2004], Earth has been subject to the Quaternary ice age with cycles of cold glacials and warm interglacials, with variations in our planets orbit as “pacemaker” [Hays et al. 1976]. The glacials themselves are characterized by frequent abrupt climate changes. In contrast, the interglacials are regarded as relatively stable.

Over the past decades we have witnessed a growing public awareness that our modern climate is not as stable as we might have thought, and that we cannot assume that humankind has no influence on the climate. Weather on Earth is chaotic by nature, making it impossible to predict the local weather more than two weeks in advance. However, the averaged weather, or climate, exhibits predictability on longer time-scales. All state-of-the-art computer models predict global warming due to the increased atmospheric concentration of greenhouse gasses [IPCC 2007].

A major concern is the possibility that global warming may lead to abrupt climate changes. Prediction of abrupt climate change and other tipping points [Lenton et al. 2008] in the climate system is one of the biggest challenges in climate modelling. In order to make any meaningful prediction about abrupt state-changes, we need to identify the relevant, possibly cyclical, climatic mechanisms and study their sensitivity to changes. In regards to climatic cyclicity on longer (centennial to millennial) time scales, the characteristic time-scales of the deep-ocean circulation system and the cryosphere make them obvious candidates.

One might say [Seidov et al. 2001b] that ocean dynamics and the ocean-atmosphere-cryosphere coupling during rapid climate changes can be regarded as two of the most fundamental unresolved issues in paleoclimateology. Not only does this explain why this introduction and most of our discussions evolve around these aspects of the climate system, but it also highlights a core motivation behind this study: to further develop the cryospheric coupling in a global climate model that is ideally suited for long-term simulations of deep-ocean circulation.

1.1.2. This thesis

This thesis revolves around further development of the coupled ocean-atmosphere-sea-ice model EcBilt-CLIO (a.k.a. LOVECLIM) by explicit numerical representation of iceberg behaviour. Icebergs are both part of the cryosphere and interact directly with the surface ocean and might therefore have played an important role in abrupt climate change events in the past. Indeed, in glacial times abrupt climate changes were quite common, and many of them (e.g. Heinrich events) are characterized by the release of drift-ice into the North Atlantic Ocean. Theoretically, the fresh melt water from melting icebergs can disturb the ocean circulation [e.g. [Broecker 1994a; Death et al. 2006]], leading to state shifts in the climate system. Furthermore melting icebergs cool their surroundings by absorbing the heat needed to melt the ice. Both thru their freshening and their cooling effects, icebergs can interact with sea-ice, which is a major player in the climate system.

In section 1.2.1.3 of this introduction chapter we put the development of our coupled iceberg-module in an historical perspective, following a brief introduction to climate modelling and to our climate model of choice (LOVECLIM / EcBilt-CLIO).

Then we provide the non-expert reader with some basic information on the climate system (1.2.2), particularly the ocean and the cryosphere. Subsequently, climate variability (1.2.3) is put in the context
of the Quaternary ice-ages, leading to a brief discussion of two climate change events from the geologically recent past, and of the stochastic resonance concept as a tool for a more generalized understanding of cyclical variability. This overview will allow us to pin-point gaps or simplifications in our knowledge, and formulate how the coupled icebergs can help to fill in these gaps and/or assess the justifications for these simplifications. This leads up to the formulation of our research questions in section 1.3.

1.2 General background information

1.2.1.1 Climate Modelling: a spectrum of climate models

Earth’s climate system is complex. The atmosphere, for example, exhibits turbulent and chaotic behaviour, which prevents even the most sophisticated computer models from providing a reliable weather prediction more than two weeks into the future. Interestingly, phenomena that are chaotic on a small scale, can still lead to predictable averages on larger scales (also illustrated in chapter 2). For example, the turbulent convection of warm air is too chaotic for a climate model to solve locally (the computational cost would explode with the number of air particles involved), but on a longer time scale it leads to average air-flows (on larger spatial scales). These air flows can be mimicked quite efficiently with parameterized flux-models. Similarly, on a seasonal or hemispheric scale, simpler models are quite capable of providing reliable estimates.

In fact, one of the simplest climate models is a zero-dimensional energy balance model. This model is based on one of the fundamental laws of physics: the conservation of energy. From this principle the total radiation (heat) emitted by the Earth equals the absorbed radiation from the sun (depending on the effective emissivity and on the albedo respectively). From this single equation model (e.g. formula 3.8 in [McGuffie and Henderson-Sellers 1997]) it can be calculated in a very transparent manner that a 3.3% decrease in Earth’s planetary albedo (~0.3 [Peixoto and Oort 1992], page 103), leads to a 1 degree rise of the steady state average Earth temperature.

This result from a simple global scale model hints at an important role for sea-ice cover in the complex climate system (snow covered sea-ice has an albedo of around 0.8; versus an albedo of less than 0.1 for open water at high latitudes (> 60°) [Henderson-Sellers and Robinson 1992], page 47). Of course, to really assess the importance of the sea-ice albedo for the global temperature, we would have to take into account the latitudinal (and seasonal) dependence of the incoming radiation and the land/water fraction, thus adding dimensions to our model. We might then, for example, obtain a ballpark figure for a steady state global temperature rise if arctic sea-ice were to disappear. But we would not be taking into account the alterations in atmospheric patterns and ocean currents that would undoubtedly accompany such a change, which could have much more dramatic effects locally. For that, we would need a more complex model, with a coupled dynamic-thermodynamic atmosphere and ocean and sea-ice, preferably on a rotating sphere with topology. In such a model the sensitivity of the temperature to changes in sea-ice can be studied and the positive feedback of the sea-ice on the temperature can then be compared to other feedbacks in the climate system. Due to the increased complexity, the resulting quantifications will tend to be more realistic, but more difficult to interpret in terms of cause-and-effect relationships (“less transparent”).

The examples above illustrate that the demands we put on a climate model depend on the nature of the question (and the questions we can ask depend on the model). This has led to the development of a wide range of climate models, ranging from basic conceptual models to complex GCMs [McGuffie and Henderson-Sellers 1997], with EMICs bridging the gap [Claussen et al. 2002]. The term GCM (originally: “gen-
eral circulation model”) refers to full three-dimensional global climate models using a time-step approach to solve primitive equations that represent physical laws like the ideal gas law and conservation of energy, mass and momentum [McGuffie and Henderson-Sellers 1997]. The term GCM (also: “global climate model”) is usually reserved for global models comprising at least the atmosphere and the ocean and operating at the highest spatial and temporal resolution currently feasible; i.e. the state of the art global climate models.

The GCMs can be extended into so-called Earth system models by incorporating additional components such as ice sheets, vegetation dynamics, the carbon cycle etc. However, many of these components operate on longer time scales that are computationally too expensive, so the involved GCMs are usually simplified and/or run at a lower resolution. This has led to the term Earth system model of intermediate complexity (EMIC). In practise the term EMIC refers to the climate models in the middle of the spectrum between simple conceptual models and GCMs. Each model has its own merits in regards to realism, efficiency and transparency. Complex models will generally contain more parameterisations, involve more feedbacks and exhibit multiple solutions, thus they tend to be less transparent. This leads to a need for carefully formulated sensitivity experiments in order to be able to pin-point cause-and-effect relationships.

1.2.1.2 Climate modelling: LOVECLIM / EcBilt-CLIO

For all 4 studies presented in this thesis we have exclusively used climate model EcBilt-CLIO, which is now known as (version 1.0 of) LOVECLIM [Goosse et al. 2010]. EcBilt-CLIO is one of the most elaborate and complex EMICs available [Claussen et al. 2002]. Basically it is a GCM-based EMIC that is suitable for (ensemble) experiments on a multimillennial timescale, due to the simplified atmosphere. With its 2.5 layers and prescribed cloud-cover the quasi-geostrophic relatively simple and fast (T21) atmospheric component (“EcBilt”) is not particularly suited to study atmospheric phenomena, especially at lower latitudes, but it is deemed sufficiently realistic for its purpose of providing coupled feedback to mid-to-high latitude oceanic currents [Goosse et al. 2010].

The ocean component (“CLIO”) on the other hand is a primitive equation, free-surface ocean general circulation model and has a higher resolution (3° by 3° boxes and 20 layers) than the atmosphere. The rationale behind this is that on longer time-scales the ocean is regarded as the major driver of both the atmospheric patterns and the climate variability.

At present, EcBilt-CLIO/LOVECLIM has optional vegetation, ice sheet and carbon modules, which are all kept constant (switched off) in our studies, for the sake of efficiency and transparent sensitivity experiments.

We have coupled an interactive iceberg module to the model, which has allowed for simulation-experiments that have led to our chapters 3, 4 and 5. In the next section we provide an historical perspective for this model-development, referring to past research that has allowed for these developments to take place.

1.2.1.3 Climate Modelling: Iceberg module development

In 1914 the International Ice Patrol (IIP) was established, following the sinking of the RMS Titanic. Their efforts in monitoring iceberg movement in the northern Atlantic Ocean laid the foundations of our iceberg module. In the 1970s, IIP-researchers described iceberg drift and physical properties of icebergs [Robe 1980; Robe et al. 1977; Farmer and Robe 1977, 1976; Wolford and Moynihan 1969; Wolford 1972, 1973 dissertation], which paved the way for the numerical prediction of iceberg drift as a function of water drag, air drag, Coriolis force and the sea surface slope [Mountain 1980].

S.D. Smith [1983; 1993] proceeded to fit model parameters to observed iceberg tracks. By adaptively
adjusting the coefficients for air and water drag, Smith optimized the hind casting of an extensive set of iceberg tracks. From this perspective, imprecise knowledge of ocean state and iceberg properties are compensated for by adjusted drag-coefficients [Smith 1993], which were later adopted by Bigg et al. [Bigg et al. 1996], and subsequently by us.

Laying the foundation for our iceberg module, Grant Bigg et al. [1997] refined this trajectory model by introducing iceberg deterioration due to melt and erosion, based on work by others [Weeks and Campbell 1973; Eltahan et al. 1983; Loset 1993a]. Bigg et al. [1997] demonstrated good general agreement with limited observational data, although not claiming high accuracy for individual iceberg trajectories. They suggested that icebergs may be exploited as previously little-used geophysical tracers.

Bigg et al. proceeded to investigate many aspects of icebergs (e. g. [Bigg and Wadley 2001]) and their relation to the glacial oceans. For example, they compared modelled iceberg trajectories with geological data to gain insight in LGM ocean states and/or the location of iceberg sources both in the North Atlantic [Bigg et al. 1998; Watkins et al. 2007; Death et al. 2006; Bigg et al. 2010] and in the Pacific [Bigg et al. 2008]. Modern Southern Ocean iceberg trajectories and melt rates were studied in an experiment comparable to our chapter 3, albeit without a coupled atmosphere [Gladstone et al. 2001]. From 2008 they started using fully coupled atmosphere-ocean global climate models, similarly to our approach, with which a Heinrich event was simulated [Levine and Bigg 2008].

At the start of our project, in 2003, we extended the approach of Bigg et al. at the time, by implementing a two-way coupling between a melting-iceberg trajectory module and a coupled global climate model. Since 2003, Bigg and his group have partially progressed along the same line. Where they advanced quickly to concrete ocean-model studies aimed in principle at Paleoceanographic reconstruction, we aimed at assessing the feedback of icebergs on the climate. In general we took a sensitivity approach where we compare the effect of dynamic-thermodynamic icebergs on the climate to alternative hosing or flooding set-ups without icebergs. By using this sensitivity approach we aim to pin-point the aspects of dynamic-thermodynamic icebergs that significantly affect the climate. Similarly to the approach of Matsumoto [1996; Matsumoto 1997], when the number of icebergs involved is very high, the modelled iceberg trajectories are regarded as representing ensembles of icebergs of similar size, shape and origin.

By 2008, Bigg et al. had progressed to a similar interactive coupling between icebergs and a global climate model. Simulating a Heinrich event, they concluded that due to the more localized iceberg melt, more fresh water flux was needed to shut down the NADW than classic hosing/flooding experiments would indicate [Levine and Bigg 2008]. However they did not take into account the latent heat that is associated with the phase-transition from solid ice to liquid water, as we did. At this point we note that some other iceberg modellers also took into account the latent heat of melting [Schäfer-Neth and Stattegger 1999], albeit not in a coupled global climate model.

In chapters 3 thru 5 we will illustrate the importance of this latent heat of melting extensively. The role of this latent heat and of other aspects of the dynamical-thermodynamical icebergs is further discussed in the section on research questions (1.3) of this introduction chapter, as well as in the synthesis (chapter 6). For the convenience of the general reader we first provide a brief overview of climate variability in general and Quaternary climate in particular. Any specialist in these fields is encouraged to skip ahead to section 1.3.
1.2.2 Climate System

In this section we briefly introduce the climate system components that are relevant for this thesis. Based on the link between ocean circulation and rapid climate change (e.g. [McManus et al. 2004; Seidov et al. 2001b; Schmittner et al. 2007]) we focus on ocean circulation and the role of the cryosphere therein.

1.2.2.1 Climate System: Global ocean circulation.

The oceans play a major role in the global climate, for one because the heat capacity of water is relatively large and two-thirds of our planet is covered by them. With a total heat capacity that is around one thousand times higher than the atmosphere, the oceans net heat uptake since 1960 is about 20 times greater than the atmospheric uptake [IPCC 2007]. Not only do bodies of water act as a kind of heat capacitor -effectively delaying and smoothing weather variability- oceanic currents are responsible for transporting large quantities of heat from tropical latitudes to the poles (~3.5 x 10^15 W at 25°N [Peixoto and Oort 1992]). Surface-water flowing to the poles must be compensated for by deeper currents in the other direction, to preserve the mass-balance. Schematically, this circulation of ocean currents is known as the "global conveyor belt" [Broecker 1994b, 1991] (see Figure 1.1), or more accurately: meridional overturning circulation (MOC). For example, the Atlantic meridional overturning circulation (AMOC) can be defined as the north-south flow through a transatlantic section at 25°N [Bryden et al. 2005]).

Figure 1.1) Great ocean conveyor logo [Broecker 1987], a cartoon of global circulation of ocean waters. [Illustration by Joe Le Monnier, Natural History Magazine]. Superimposed: a highly simplified thermohaline circulation [Rahmstorf, 2002].
1.2.2.2 Climate System: Thermohaline circulation

Surface currents are primarily driven by wind-stress, Coriolis effects and by density gradients in the surface waters. Water density is determined by both temperature and salinity and the corresponding water circulation is referred to as the “Thermohaline Circulation” (THC, Figure 1.1).

The foundation of our understanding of ocean circulation was laid by Johan Sandström [Rahmstorf 2003]. In 1908 he conducted a series of classic experiments at the Bornö oceanographic station in Sweden. From an adjacent fjord he sampled water of different densities, which he studied under controlled conditions. Blowing air over a tank that was filled with layers of the different water samples he showed the properties of “wind-driven circulation”. By heating and cooling the water layers at different depths he found that thermal forcing can only give rise to a steady circulation if the heating occurs at greater depths than the cooling. This fact is now known as “Sandström’s theorem”.

On Earth thermohaline forcing only occurs at the ocean’s surface (neglecting the small geothermal component). Sandström recognized that the thermal engine of ocean circulation depends on downward penetration of heat at low latitudes due to turbulent mixing (e.g. [Munk and Wunsch 1998; Schulz et al. 2002]). The energy needed for this turbulent mixing is provided by the tides and winds and not by salinity driven convection as was speculated originally by Sandström [Rahmstorf 2003; Wunsch and Ferrari 2004].

1.2.2.3 Climate System: The wind

From the climate modelling perspective, the atmosphere interacts with the ocean in three ways: precipitation/evaporation; radiative/sensible heat transfer; and surface stress or wind. The wind plays two distinct roles in ocean circulation. It causes turbulence, which is crucial for thermohaline circulation, and it partly drives the ocean circulation directly, albeit not in the same direction as the wind. Due to a balance between drag-forces and Coriolis force, surface currents flow at a 45° angle to the wind, leading to perpendicular fluxes and possible up- or down-welling, a phenomenon known as Ekman transport (e.g. [Peixoto and Oort 1992]).

Wind-driven circulation doesn’t require thermohaline surface forcing but it’s closely coupled to the THC because the atmospheric distribution of pressure-fields depends largely on oceanic currents. This atmospheric feedback provides a self-sustaining effect on the prevailing ocean currents. Furthermore, the wind-fields and ocean dynamics strongly affect the sea-ice, especially in the Arctic [McGuffie and Henderson-Sellers 1997]. In other words: The atmosphere provides stabilizing feedback to changes in the ocean circulation pattern, which in turn can lead to multiple stable states and threshold behaviour (e.g. [Manabe and Stouffer 1988]).

1.2.2.4 Climate System: Deepwater formation

When relatively salty surface waters cool off, their density can increase so much that in some areas it can lead to sinking of pockets of dense waters; deep water formation. Global circulation of ocean waters depends partly on deepwater formation in a limited number of key-areas, the most important of which are: the Labrador Sea and the Greenland Iceland Norwegian (GIN) Seas (Figure 1.2) in the North Atlantic and the Antarctic Weddell and Ross Seas (see Figure 3.2 for geographic location).
The global ocean circulation is altered significantly if this deepwater formation is stopped or disturbed. Deepwater formation, or vertical mixing, at high latitudes depends on topography [Wunsch and Ferrari 2004], winds and on the temperature-gradient but also on salinity and on the presence of open water (polynyas) in sea-ice [Peixoto and Oort 1992; Grigg and Holbrook 2001]. During sea-ice formation, salty brine is rejected from the crystal lattice of the ice, which increases the water density and facilitates deep water formation. Furthermore, there is a memory effect; deepwater formation will precondition the situation for the following year [Hakkinen 1995; Jongma et al. 2009].

The ocean circulation has a significant and non-linear effect on the global climate. Small local disturbances in the freshwater and evaporation budget can have a large impact on the circulation pattern. Furthermore, equilibration times of the oceans range from months for surface layers to 3000 years for the deep ocean (Table 1.1). It is therefore very likely that changes in ocean currents have been responsible for, or played a major role in, a variety of climatic changes in the past (e.g. [Broecker 2003, 2006; McManus et al. 2004]).

1.2.2.5 Climate System: Freshwater forcing and the Cryosphere

Affecting the ocean circulation, freshwater forcing can have a variety of causes. Precipitation is the most direct freshwater source, but in principle, icebergs might flow to deep water formation sites and directly influence the deep water formation with their melt-fluxes. River discharge provides a more coastal and slightly delayed distribution of land-based precipitation, with the land acting as a leaky bucket or capacitor. When the precipitation falls in solid form and is not released until spring this capacitor effect becomes stronger and the release of freshwater more concentrated (Figure 1.3). Waxing and waning mountain glaciers can store and delay the water for a couple of hundred years. Fresh water is also stored in frozen ground or permafrost. Even sea-ice must be regarded as a fresh water source. It acts as a dynamic fresh water capacitor since brine rejection leads to a relatively low salinity of the ice.

Ice-sheets can introduce a very long delay and have a huge capacity. This is illustrated by the 120 meters lower sea-level at the last glacial maximum due to storage of water as ice [Rohling et al. 1998]. Icebergs calve from the sheet’s edges. In the glacial climate, episodically icebergs are released on a very large scale. There are several mechanisms that might explain this. Rapid disintegration of stranded ice shelves [Hulbe et al. 2004; Grosfeld et al. 2001] is a likely candidate to explain synchronized iceberg release across the northern North Atlantic. This rapid disintegration could be due to atmospheric climate changes.
("meltwater infilling of surface crevasses"; [Hulbe et al. 2004]) and/or rising sea-levels. Cyclical purging could also be an internal ice-sheet phenomenon. When the sheet’s mass reaches a critical threshold, trapped geothermal heat combined with pressure-melting can produce a bottom layer of melted ice. This might then act as a low-friction film that allows the sheet to collapse rapidly causing a pulse of icebergs [Macayeal 1993]. Basal melting can also result in the build-up of large amounts of dammed-of freshwater under the sheets, which can be released suddenly when the dam is broken (and the dam itself would be a source of icebergs) [Johnson and Lauritzen 1995]. Bottom topography plays a key role in any detailed modelling of glacier and ice sheet processes, making this a challenging subject and a work in progress [Flowers and Clarke 2002].

Given their pulsing character and large capacity, ice sheets have a great potential to affect the fresh water balance, near key deep water formation sites, disturbing directly the oceanic circulation and changing climate abruptly on a hemispheric scale (Figure 1.3).
1.2.3 Climate variability

Climate changes can be characterized in a very general sense by their typical natural time scales [Mitchell 1976], as illustrated in Table 1.1. In Quaternary climate archives three types of changes can be distinguished: Slow variations in the order of tens of thousands of years that are usually related to variations in the Earth’s orbit around the sun [Milankovitch 1941]; Centennial to millennial scale variations that are apparently related to ice-sheet and ocean dynamics [Grootes and Stuiver 1997; Bond et al. 1993]; Smaller decadal scale fluctuations around a mean state, such as the El Nino Southern Oscillation (ENSO) and the North Atlantic Oscillation (NAO) (e.g. [Cassou and Terray 2001]). Of course this rough division is a simplification of the complex climate system, which exhibits fluctuations and feedbacks on most spatial and temporal time-scales. For example, sea-ice provides strong positive feedback to both cooling and warming conditions, and therefore sea-ice fluctuations operate on a seasonal but also on a multidecadal [Yi et al. 1999; Haarsma et al. 2000] and multicentennial to millennial time-scale [Bond et al. 1997].

In this study, we focus mostly on mechanisms that operate on the centennial to millennial time-scale and relate to ocean circulation and the cryosphere. Thanks to their large amplitudes, the mil-
Millennial-scale climate fluctuations over the last glacial cycle are particularly interesting [Seidov et al. 2001b]. They are well documented in high-resolution polar ice-cores, and they can also be studied quite effectively in marine-sediment cores [Bond et al. 1997]. These abrupt climate changes have been related to pattern changes in oceanic currents as well as to pulsing behaviour of ice-sheets (e.g. [Bond et al. 1993]).

1.2.3.1 Climate Variability: Glacial-Interglacial cycles and Milankovitch forcing

Some 2.5 million years ago Earth entered what is now referred to as a “glacial epoch” [Gradstein et al. 2004]. Following the initiation of an Antarctic ice-sheet some 20 Ma ago, a Northern ice-cap started to grow. The ice then grew and decayed (while the sea-level fell and rose) in cycles of roughly 40 thousand, and later (from ~1.2 Ma) 100 thousand, years. A period of extreme climate changes was the result.

For ice sheets to grow, they need both low temperatures and enough precipitation. North-bound ocean currents could provide the necessary moisture to high northern latitudes. Supposedly, closing of the Strait of Panama set the stage some 4.2 Ma ago by initiating the warm gulf stream, which brings precipitation towards Scandinavia [Haug and Tiedemann 1998; Seidov and B.J. Haupt 2003; Sarthein et al. 2009]. Tectonic uplift of the Indonesian seaway preluded an intensification of the North Pacific Gyre, following the initiation of the “modern” Kuroshio Current by 3 Ma [Gallagher et al. 2009]. A Pacific link with the North American ice cap is further supported by increased subarctic Pacific stratification at 2.7 Ma [Haug et al. 2005; Haug et al. 1999]. Meanwhile, the Northward displacement of New Guinea some 5 Ma ago could have cooled the Indian Ocean and reduced poleward heat-transport, by switching the flow from warmer South to colder North Pacific origin [Cane and Molnar 2001]. And the Bering Straight could also play a role since closing of the Bering Straight intensifies Atlantic overturning circulation [Wadley and Bigg 2002].

By capturing water moisture with predominantly lower molecular mass, the global ice mass affects the ratio of oxygen isotopes (\(^{18}O / ^{16}O\)) in the ocean. In Figure 1.4A the marine \(\delta^{18}O\) record is given for the past 3.5 Ma [Clark et al. 1999]. \(\delta^{18}O = \left(\frac{^{18}O}{^{16}O}\right) / \left(\frac{^{18}O}{^{16}O}_{\text{std}}\right) - 1\) x 1000, with std the mean ocean water reference [Mix et al. 1995]. High \(\delta^{18}O\) values reflect increased planetary ice mass and cooler conditions (glacials) and low levels correspond to high sea level (interglacials). In general we see a gradual buildup of the ice over the past 3.5 Ma. Superimposed there are cycles of cooling during cold glacials, followed by a rapid return to warmer climate during interglacials. Sea surface temperature also affect the oxygen isotope ratio in precipitation and hence in the ice sheets themselves, as does the local air temperature above the ice sheet, and their height [Figure 1.4B] [Johnsen et al. 1972; Alley et al. 1997a; Severinghaus et al. 1998], but most of the change is due to the growth and decay of major ice sheets reconstructed in Figure 1.4C. This reconstruction was the basis for the Peltier 2004 reconstruction that was used in chapter 4 of this thesis.

Milankovitch [1941] was one of the first to convincingly link the semi-periodic waxing and waning of ice-sheets (Figure 1.4A&B) to small periodic changes in the Earth’s orbit around the sun (Figure 1.4B, bottom). These changes in tilt, obliquity and precession are due to gravitational interaction with the planetary system [House 1995]. The associated changes in insolation are commonly referred to as Milankovitch forcing and can account for a substantial part of Pleistocene climate variability, specifically at near 20 and 41 ka cyclicly.

At small ~100 ky- and larger ~10 ky - frequencies the waxing and waning of the ice sheets shows more complex non-linear behaviour. These non-linearities find their origin in feedback mechanisms in Earth’s climate system and in delays introduced by the slower components such as the ice-caps [Huybers and...
and reviewed extensively by Saltzman [Saltzman 2002]. Here we take this opportunity to introduce an amplification mechanism, the conception of which was inspired by the dramatic glacial-interglacial cycles at roughly 100 ky.

1.2.3.2 Climate Variability: Introducing Stochastic Resonance

Intriguingly, the Milankovitch forcing mentioned above is relatively small compared to the dramatic climate changes that it correlates with, especially at 100 ka. This indicates that some kind of amplification mechanism is at work in the climate system that can amplify small periodic signals. In chapter 2 of this thesis we investigate a similar (multicentennial scale) amplification mechanism by simulating the effect of a small periodic fresh water flux in the Labrador Sea on the North Atlantic overturning circulation. We show that this weak signal is amplified with the aid of the internal variability of the system. Attempts to generalize such amplifications of small periodic signals have led to the concept of Stochastic Resonance (SR).

In [1981] Benzi et al. showed that some dynamical systems, “subject to both periodic forcing and random perturbation may show a resonance (peak in the power spectrum) which is absent when either the forcing or the perturbation is absent”. They named this cooperative effect between a small periodic external forcing and the internal noise: Stochastic Resonance (SR). They suggested that this phenomenon might explain the 100 ka peak in the power spectra of Quaternary paleoclimatic records, as did Nicolis who was also studying stochastic responses to weak periodic forcing and additive fluctuations at that time [Nicolis 1982]. Since then the SR-mechanism has been identified in a wide range of physical systems [Gammaitoni et al. 1998].

The term stochastic resonance is now used for a very broad class of phenomena, but has been...
shown to lack mathematical robustness [Imkeller and Pavlyukevich 2002]. A more robust definition [Herrmann and Imkeller 2002] of an SR-point, which draws from Freidlin’s theory of large deviations, evolves around transit-times between potential wells, and redefines the SR-point as an optimal timescale on a resonance interval [Herrmann et al. 2005]. Previous studies on SR in the North Atlantic circulation are discussed below in section 1.2.3.6.

1.2.3.3 Climate Variability: Millennial-scale; Dansgaard-Oeschger events and Bond cycles

On a centennial to millennial time-scale the glacial periods themselves show strong and cyclical fluctuations in temperature and ice-mass. The Last Glacial exhibits approximately 23 climate change cycles, known as Dansgaard-Oeschger (D/O) events, (Figure 1.5 b and c).

Figure 1.5) Millennial scale climate variability during the last glacial cycle. Heinrich events 2 thru 5 are marked with yellow verticals. From PAGES [Labeyrie 2002] A) Greenland air temperature. Vertical bars indicate Heinrich events. The series of Dansgaard-Oeschger events from one Heinrich event to the next are known as Bond-cycles. B) Atlantic Ocean 37 °N sea surface temperatures (SST), reflecting N/S shifts of the atmospheric polar front. Reconstructed from stable oxygen isotope fractions of planktic foraminifera. (ocean bottom core MD95-2042). C) Deepwater ventilation, represented by °O content of benthic foraminifera. D) °O content of benthic foraminifera reflecting the state of the thermohaline circulation. E) West-Antarctic air temperature deduced form δ18O in the Bird core.
c). D/O-events typically consist of an abrupt, decadal scale, warming (interstadial); followed by a gradual cooling over a few centuries, and a distinct cold phase (stadial) lasting hundreds of years to a millennium [Stocker 1999]. A typical D/O cycle has a period of 1470 ± 500 years [Grootes and Stuiver 1997; Schulz et al. 1999]. A series of progressively colder D/O cycles followed by a major cold phase (a Heinrich event) is called a Bond cycle [Bond et al. 1993; Hughes 1996].

D/O cycles were found in the Greenland ice cores [Dansgaard et al. 1993; Grootes et al. 1993], where the amplitude of the oxygen isotopic variation over a D/O cycle is about 50-75% of the full glacial-interglacial range [Stuiver and Grootes 2000]. Their equivalents have been shown in North Atlantic marine sediment cores [Bond et al. 1992; Bond et al. 1993; Bond and Lotti 1995] but also in, for example, Mediterranean [Cacho et al. 1999] and Santa Barbara Basin [Kennett et al. 2000] records. D/O-events are most pronounced in the northern hemisphere, but associated rapid climate variability in Antarctica illustrates their global extent (Figure 1.5e) [Blunier and Brook 2001].

An important signature of the cold phase of these millennial scale variations is the ice rafted detritus (IRD) in North Atlantic sediments, indicating episodes of large scale iceberg (and/or sea-ice) rafting. These sediments consist of normal carbonate shells alternating with detritus that originated in basal Canadian and European terrains [Heinrich 1988; Grousset et al. 1993]. Bond et al. have investigated the 1-2 ka cycle in many deep-sea subpolar North Atlantic sediment cores [1995; 1999; 1997; 2001; 1992; 1993]. Robust indicators of ice rafting episodes are two petrologic tracers: Icelandic glass and hematite-stained grains [Bond et al. 1999]. The most pronounced glacial IRD layers are known as Heinrich events, which mark the end of a Bond-cycle.

Figure 1.5e. Cartoon of the Ruddiman belt and inferred circulation (grey arrows) compiled by several authors [Chapman and Maslin 1999]. Black arrows indicate likely sources of icebergs. Adopted from [Bard et al. 2000].
1.2.3.4 Climate variability: Abrupt climate change; Heinrich events

Heinrich events are characterized by horizons of ice rafted detritus (IRD), marking massive release of icebergs from the ice-sheets [Heinrich 1988; Bond et al. 1992; Grousset et al. 1993]. These icebergs melted preferentially in the "Ruddiman-belt" between 40°-55° and 10°-60°W [Ruddiman 1977] (Figure 1.6), between Newfoundland and Europe [Grousset et al. 1993], matching a δ¹⁸O meltwater tongue [Cortijo et al. 1997]. Heinrich events thus are clearly associated with icebergs and their origin: the ice sheets. According to Alley & MacAyeal [1994] the Laurentide ice sheet could have contained enough detritus to account for all IRD of a Heinrich event. Although some of the IRD has an Eurasian origin, Hemming [Hemming 2004] defined a subset of Heinrich events (H1, H2, H4 and H5) that apparently originate in the Hudson Strait. Other paleoclimatic records show that Heinrich events are associated with major abrupt cooling events, and coincide with a global or at least hemisphere-wide climatic footprint [Bond et al. 1993; Grimm et al. 1993].

This strong cooling in the North Atlantic is hypothesized to originate from a disturbed ocean circulation, which is weakened by the meltwater associated with the iceberg armadas [Bond et al. 1993; Broecker 1994a; McManus et al. 2004; Sakai and Peltier 1995]. This is corroborated by numerous coupled atmosphere-ocean model experiments, showing that the freshwater melting from these icebergs can trigger (partial) "shut down" of the meridional overturning circulation (e.g. [Sakai and Peltier 1997; Kageyama et al. 2010; Roche et al. 2010]). In these so-called hosing studies, the freshening effect of the melting icebergs is typically simplified to homogeneous and instantaneous dumping of freshwater on a designated ocean area that broadly corresponds to the Ruddiman belt.

1.2.3.5 Climate variability: Abrupt climate change; the 8.2 ka event

The 8.2 ka climate event is the most pronounced Holocene climate event in the Greenland ice core records [Alley et al. 1997b]. It is a 160-year [Thomas et al. 2007] cold and dry period centred around 8200 years ago. The 8.2 ka climate event is recorded in a large number of proxy archives in the North Atlantic area as well as in areas influenced by monsoons [Alley and AGustsdottir 2005; Rohling and Palike 2005; Wiremsa and Renssen 2006]. The proposed cause of the event is the sudden drainage of the Laurentide lakes (Lake Agassiz and Ojibway) through the Hudson Strait, following the catastrophic break-up of an ice-dam. Early Holocene decay of the remnants of the Laurentide ice sheet triggered this drainage around 8470 yrs BP [Barber et al. 1999]. A mechanism of fresher surface waters inhibiting deep-water formation in the North Atlantic Ocean, leading to a decreased northward oceanic heat transport and cooler conditions has been successfully simulated (e.g. [Bauer et al. 2004; Renssen et al. 2002; Renssen et al. 2001; Wiersma and Renssen 2006]).

However, the break-up of the ice-dam would be expected to involve a large number of icebergs. The possible role of these icebergs in the climatic response has been ignored so far.

1.2.3.6 Climate Variability: D/O cyclicity and Stochastic Resonance

Occurring at intervals of about 1.5 kyr, the cyclical D/O events have been correlated to variations in cosmogenic nuclides, implicating small changes in solar output [Bond et al. 2001]. To explain this correlation as a causal relationship, some kind of amplification mechanism must be at work in the climate system. The amplification of small periodic signals in complex systems falls in the domain of stochastic resonance (SR).
We regard SR as a threshold phenomenon where a periodic forcing is so small that the threshold is practically out of reach, but the internal noise of the system allows for this threshold to be crossed occasionally, which then triggers a state-transition in the system.

The small periodic forcing leads to a preferential threshold-crossing when the forcing is at its maximum, resulting in a tendency for state-switching events to occur at integer multiples of the forcing’s period $T$. In other words: the events tend to follow the beat of the forcing but due to the stochastic nature of the system one or more beats can be skipped. Therefore a recurrence-histogram of waiting times between events typically shows peaks at $T, 2T, ..., nT, ...$ with the peak-height exponentially decreasing with increasing $n$, as long as the threshold-crossing-probability can be considered constant (e. g. [Gammaitoni et al. 1998]).

After analyzing such waiting times between events in a number of filtered paleoclimatic records, Alley et al. argued in [2001] that SR in the North Atlantic could be responsible for the millennial scale regularity of Dansgaard/Oeschger events. He hypothesized that the “noise” might come from changes in the freshwater fluxes in the North Atlantic. The threshold-variable in this case is the relative density of surface waters and the deeper ocean, which does or does not allow for deep water convection.

Ganopolski and Rahmstorf [Ganopolski and Rahmstorf 2002; Ganopolski 2003] proceeded to illustrate such a mechanism by obtaining similar recurrence histograms using an intermediate complexity climate model. In chapter 2 of this thesis we show a similar mechanism at work in a more complex 3-dimensional model.

### 1.3 Research Questions

#### 1.3.1 Research questions on the sensitivity of the Atlantic overturning circulation

Ganopolski and Rahmstorf [Ganopolski and Rahmstorf 2002] demonstrated stochastic resonance by introducing a sinusoidal fresh water-forcing plus “white noise” in the (2-dimensional) North Atlantic of an intermediate complexity 3-basin model (CLIMBER-2), which they had previously shown to exhibit multiple overturning modes for the Atlantic circulation [Ganopolski and Rahmstorf 2001].

It is by no means trivial that these SR results are reproducible in a more realistic 3-dimensional coupled climate model without externally added noise, since the occurrence of stochastic resonance is very sensitive to the ratio between forcing and noise and to the distance of the thresholds relative to the steady states (e. g. Fig 2.4). Furthermore, with such one or two-dimensional models it is feasible to have run-times in the order of 100,000s of years and observe hundreds of switches, which are the ingredients for statistically convincing histograms of recurrence-times between events. However, more complex three-dimensional models have practical runtimes in the order of thousands of years, and the smaller the number of observed switches, the harder to show that the characteristic peaks (at $nT$) in the recurrence histograms are significant. It is therefore not surprising that three-dimensional climate models have not yet been shown to exhibit SR behaviour on a centennial to millennial time-scale.

- **Can we demonstrate the stochastic resonance phenomenon in a more realistic 3-dimensional climate model?**

By showing the SR mechanism at work under modern (preindustrial) boundary conditions (Chapter 2), we illustrate an amplifying mechanism that allows rela-
tively small forcings to have climatic consequences on a hemispheric scale, thus underpinning the notion of oscillations in North-Atlantic overturning circulation as an explanation for D/O cyclicity. This enables us to answer a more specific question:

- **Is there a plausible amplification mechanism to link small cyclical changes in solar activity to (Holocene) D/O cyclicity?** In chapter 2 we will refer to this as Bond’s Hypothesis. [Bond et al. 2001]

If we are to predict abrupt climate change in the future, we need to understand the mechanisms behind abrupt climate changes. Both glacial- and Holocene-D/O cyclicity are marked by millennial-scale variations in proxies of drift ice. This indicates that icebergs are involved in these climate changes, be it as passive tracers of changing ocean currents or actively affecting the ocean and climate. Given the demonstrated (Chapter 2) sensitivity of the Atlantic meridional overturning circulation (AMOC) to small local fresh water fluxes this raises the question:

- **Can icebergs affect the climate?**

In climate modelling, it has been popular practice [Kageyama et al. 2010 and references therein] to simplify drifting icebergs as a hosing or flooding of fresh water. But in doing so, two aspects of iceberg-melt are ignored that might very well prove critical: Firstly, the melting icebergs follow wind- and current influenced trajectories, spreading out from a coastal source. Therefore the distribution of melt water cannot be expected to be homogeneous, as in a traditional “hosing” experiment. Secondly no previous study has taken into account the latent heat that is needed to melt the ice, thus ignoring a possibly crucial cooling aspect of melting icebergs.

Here we have arrived at the core motivation behind our study. The remaining 3 chapters of this thesis are dedicated to the implementation of a coupled iceberg module in the climate model EcBilt-CLIO / LOVECLIM, a sensitivity-investigation of the resulting Earth system model of intermediate complexity (Chapter 3), and case studies of past abrupt climate events involving icebergs (Chapters 4 and 5).

### 1.3.2 Research Questions regarding interactive icebergs

In order to investigate the role of icebergs in the climate system in general and in abrupt climate change events in particular, we implemented an interactive coupling between an iceberg trajectory model and a coupled global climate model (see section 1.2.1.3). In doing so we also took into account an aspect of iceberg melt that has been overlooked in other studies: the latent heat of melting. We assessed the implications from this model-development with sensitivity experiments. From this approach we are led to the following **systematic questions**, or recurring issues:

- **The distribution effect:** What is the effect of the dynamic distribution of melt water by moving icebergs as compared to homogeneous or parameterized “hosing” of fresh water fluxes?
- **The cooling effect:** What is the effect of the latent heat of melting?

Chapters 3, 4 and 5 have all evolved around these two questions regarding the fluxes of melt water and melting heat. To answer these questions we have designed sensitivity experiments aimed at isolating the relevance of the various iceberg characteristics (distribution, melt-water and –heat) in the response of the complex global climate system. Chapter 3 is focussed on the modern (pre-industrial) Southern Ocean. This choice for the Southern Hemisphere was made for the sake of reduced topographic complexity and in the light of the sensitivity of NH overturning circulation which is demonstrated in chapter 2. As we will see in chapter 3 thru 5, the results indicated an important
role for sea-ice in the climatic response to the iceberg fluxes. This has led to an emphasis on a phenomenon we have dubbed “sea-ice facilitation”:

- **Sea-ice facilitation**: What is the role of sea-ice facilitation in these results? Melting icebergs lead to cooler and fresher surface waters, which facilitates the formation of sea-ice. In turn, sea-ice plays an important role in deep water formation.

### 1.3.3 Research questions on abrupt climate change: Dynamic icebergs vs. fresh water hosing in HE1 and the 8.2 ka event

Ultimately, we want to shed light on abrupt climate change involving iceberg armadas so we put forward the key question:

- **What role did icebergs play in abrupt climate change?**

This question involves more specific questions such as: What is the effect of iceberg armadas on the Atlantic overturning circulation? With an interactive iceberg module, can we improve our simulations of major past climate events and our understanding thereof? These questions lie at the base of chapters 4 and 5, which centre on two major climate events; the 8.2 event, the largest climatic event in our Holocene climate; and Heinrich event 1, the most recent glacial climate event with large-scale release of icebergs into the North Atlantic Ocean.

When simulating events such as Heinrich event 1, it is common practice to parameterize the iceberg release with a constant homogeneous freshwater flux over a designated area. Such a hosing approach neglects the dynamic nature of icebergs, floating under influence of the currents and the winds, and it neglects the cooling effect due to the latent heat of melting. From this perspective, a more to-the-point formulation of the key-question on the role of icebergs in abrupt climate change becomes:

- **Is fresh water hosing a reasonable simplification of iceberg armadas?**

Since the bulk of the icebergs will melt near the release site, homogeneous freshwater hosing might be expected to be unrealistically efficient at inhibiting convection, when compared to a more plausible iceberg-melt distribution [Jongma et al. 2009]. This relates to the systematic question on the distribution effect: What is the effect of the dynamic spatial distribution of iceberg melting fluxes, compared to a homogeneous hosing approach?

Indeed, a Heinrich event has been simulated recently with dynamic icebergs [Levine and Bigg 2008], suggesting that with a more realistic, localized freshwater flux of icebergs, the “MOC-shutdown may be harder to induce than previously suggested”. However, neither the classical hosing approach nor the above iceberg study take into account the significant amount of latent heat that is needed to melt the icebergs. Does the latent heat of melting associated with melting icebergs significantly affect the results?

To evaluate the extent to which the hosing approach is an acceptable representation of massive iceberg discharge, we perform a sensitivity experiment (Chapter 5), focusing on qualitative aspects such as: MOC shut down efficiency; MOC shut-down and re-start timing; salinity anomalies; temperature anomalies; spatial patterns and involvement of sea-ice.

With regard to the 8.2 ka event, previous studies have progressed beyond homogeneous hosing, specifying local influx of the fresh lake water, and evolving towards a more detailed scenario of the event. In this context, incorporating all the previously mentioned aspects of the dynamic icebergs will serve to simulate a more refined scenario with iceberg release (Chapter 4).