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Interim Report

IR-04-055

**Response of European Mountain Forests
to Abrupt Climate Change**

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December 2004

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Abstract

Paleoclimatic data reveals large abrupt climate changes such as the Younger Dryas and the 8.2 ka event. Alpine forests are characterized by distinct ecotonal borderlines and their vulnerability to such events is potentially high, although there are hardly any assessments made up to now. Both paleo-records and climate-model projections are used to conduct a model simulation with the forest ecosystem model PICUS v1.3. Impacts on Norwegian spruce-dominated mountain forests along an altitudinal transect are investigated. Ecosystem-specific thresholds are identified, the potential magnitude of loss is quantified and possible negative feedbacks to the carbon cycle are assessed.

Acknowledgments

I owe gratitude not only to Michael Obersteiner, who made this work possible and gave me visions and perspectives, but also to Manfred Lexer for putting my feet back on the ground with his critical support.

I would like to thank all my fellow 2003 YSSPers and colleagues at IIASA for creating such an amicable and productive spirit at the Institute.

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“...we find no vestige of a beginning, no prospect of an end...”
Hutton (1795)

“...a little taste of what may come, a mere glimpse of what has gone...”
Corgan (2000)

Response of European Mountain Forests to Abrupt Climate Change

Rupert Seidl

1 Introduction

Mankind has experienced peaceful times — climate-wise — since coming into being several thousand years ago. Some even see a connection between this stability and the arrival of the human race. But change is a key element of our Earth; it is the key to development and progress as well as to decline and extinction. Terrestrial ecosystems, being subject to these changes on Earth for a much longer period of time than humans, have adapted to the flickers in the planet's behavior. Moreover, the way we see our environment today is the product of the Earth's convulsions and the ecosystems' reactions. However, the full meaning of “change” is often neglected in modern thinking, where the only direction of “progress” is *upwards*. But, the circle only has one side.

The past 10,000 years are anomalous in the 455 billion year-long history of the Earth. This period, during which civilization developed, was significant for the weather, which was more consistent and equable than any similar time-span in the past 100 millennia (Broecker, 1995). And although this statement, applying the scales of geologists and climatologists, is highly certain there is evidence of human societies being forced into decline by environmental stress, such as the Maya-civilization (DeMenocal, 2001) or the Vikings on Greenland (Alley *et al.*, 2003). Even if modern man sees himself more independent of his environment than ever, the issue of the changing climate is prominent on today's political agenda. By now, there is a whole research community involved in estimating the temper of a beast known as “climate” (Broecker, 1999), where it is almost undeniable that humans and their activities on this planet have been poking this beast substantially. The impacts of recently experienced and the near future predicted gradual climate changes are not yet clear. And adaptive measures and mitigation strategies are not ready for application. But right now, there is another concern being raised — what if the climate is to change rapidly, step-like, over decades or less (Overpeck and Webb, 2000)?

Exploring the paleo-records granted by ice-cores and other proxies of the past makes it increasingly clear that large, abrupt, climate changes have repeatedly affected much or all of the Earth (Alley *et al.*, 2002). Cores drilled through several parts of the Greenland ice cap show a series of cold snaps and warm spells — each lasting 1,000 years or more — that raised or lowered the average winter temperature in northern Europe by as much as 10°C over the course of as little as a decade (Broecker, 1995). It seems that in addition to the orbitally forced, long-term climatic trend associated with glacial-

interglacial cycles, there are major short-term, abrupt changes in the Earth's system that occur on time scales of only a few decades to centuries (Oeschger and Arquit, 1989). Although the ability of simulating past climate events is still limited (Broecker, 1999) some of the assumed mechanisms are well established in state-of-the-art models and confirm the abruptness and amplitude of events recorded in the paleo-data (Ganopolski *et al.*, 1998; Ganopolski and Rahmstorf, 2001b; Knutti and Stocker, 2002; Alley *et al.*, 2002). In this connection, the term abrupt implies not only rapidity but also reaching a breaking point, a threshold — it implies a change that does not smoothly follow the forcing but is rapid in comparison to it (Rahmstorf, 2001). Past abrupt changes were especially prominent when climate has been forced to change from one state to another. Although rapid transitions could occur at any time because of the chaotic elements of the climate system, the existence of forcing greatly increases the number of possible mechanisms. In this context, human activities seem to drive the climate system towards one of these thresholds increasing the likelihood of an abrupt climate change (Alley *et al.*, 2003).

But how can terrestrial ecosystems cope with such an event? Vegetation is a continuously changing set of variables chasing a continuously changing set of variables, namely, climate (BESR, 1995). And although forests, as a class, have proven resilient to past changes in climate, today's fragmented and degraded forests are highly vulnerable (Noss, 2001). Ecosystems are subject to many pressures, such as land-use changes, deposition of nutrients and pollutants, harvesting, grazing by livestock, introduction of exotic species, and natural climate variability (IPCC, 2001b). These human activities add a surplus to the already high vulnerability of ecosystems caused by longevity and immobility. Alpine ecosystems are likely to be particularly vulnerable to these anticipated changes. System characteristics, as steep altitudinal gradients and thus distinct ecotonal borders and a history of intensified land use, contribute to this fact. As for the climate system, ecosystem response shows threshold behavior involving biological processes on all levels of organization (Tobolski and Ammann, 2000). It is likely that plants respond rapidly to a changed environment according to their specific physiological potentials and limits (Tinner and Lotter, 2001). Such ecosystem reorganizations could, on the one hand, lead to major impacts on the natural resource supply in altering the forests' production capacity. On the other hand, large-scale ecosystem changes have the potential to feed back on the climate system via reduced potential for carbon sequestration or changed albedos.

2 Objectives

*“Research programs should be initiated to collect data to improve understanding of thresholds and nonlinearities in geophysical, ecological, and economic systems. [...] Data collection should target sectors where the impacts of abrupt climate change are likely to be largest or where knowledge of ongoing changes will be especially useful in understanding impacts and developing response alternatives” (Alley *et al.*, 2002).*

This statement of the Committee on Abrupt Climate Change recently published in the “standard-work” in this field, shows that we are lacking knowledge of the processes

such an event might trigger. What we definitely know is that forests are among those systems with the largest potential for a highly adverse impact. Lifetimes on geological time scales and slow migration speed as well as human fragmentation and impoverishment contribute to their vulnerability. But there is a great deal of complexity in the response to climate change, involving processes from organism to ecosystem level with the possibility of interaction and mediation between species as well as processes (Hebda, 1996; Birks and Ammann, 2000; Shugart, 1998). There is no unique pattern of reaction. Evidence from the latest period of rapid climate warming, at the end of the last glaciation, shows that in some areas the initial responses involved replacement of species and whole communities over a few decades at most whereas other results show that during the same period the establishment of forest ecosystems lagged behind temperature changes by up to two thousand years (PAGES, 2001). Possible mechanisms of biotic response may involve different processes, with the time consuming process of migration being the most prominent in the pollen data available. Migration rates, compiled in Noss (2001), have been estimated to range from 50 m up to 2000 m per year for different species. But also faster processes, such as changes in biomass production and its allocation to both vegetative and generative production have to be considered (Tobolski and Ammann, 2000). However, the final reaction is extinction, being driven rather by the rate of environmental change than by its magnitude (BESR, 1995).

This study is an attempt to get a better understanding of the reaction of forests to a rapidly changing climate. And because sharp ecotones, as found in alpine regions, seem to be especially sensitive to such changes (Petee, 2000), they are crucial to such an investigation. This work identifies the forest response to abrupt changing climate parameters focusing on mountain forests in the eastern Alps. In a second step, the paper quantifies this reaction to the extent possible and assesses the capability of auto-adaptation.

Ecosystems understanding, as well as previously conducted studies, indicate that reactions are likely to be non-linear showing threshold behavior. Since forests are highly adapted to their environment, climate changes have the ability to push them across a threshold (Alley *et al.*, 2002). The greater the rate and magnitude of temperature and other climatic changes, the greater the likelihood that critical system imminent thresholds are surpassed (IPCC, 2001b). One scope of this work is to gain knowledge for possible adaptive management strategies in approaching these thresholds performing a “sensitivity analysis” to the ecosystem.

This sensitivity is of high interest when it comes to assessing the implications of abrupt climate changes and particularly in developing suggestions to overcome or mitigate adverse effects. Among possible impacts is a strong alteration of the carbon budget, since terrestrial carbon storage primarily occurs in forests (Dean and Gorham (1998) in Falkowski *et al.*, 2000). In the near future forests will have the potential to act as a major sink for atmospheric CO₂, but there is uncertainty concerning the response of terrestrial ecosystems to anticipated climate changes (GCP, 2003). With regard to this fact, an objective is to assess the potential feedback of major ecosystem alterations on the terrestrial carbon budget.

3 Methods and Materials

3.1 Background

3.1.1 The Earth's Changing Climate

This section gives a short summary of the knowledge on the Earth's changing climate system, which is the background to this study.

Climate is and has always been changing on all time and space scales and the Earth's history has been determined by these changes. From the climate archive in ice shields or sediment records that extend back tens of thousands of years, three major types of change have been distinguished (following Stocker, 1999):

- ◆ Natural fluctuations about a mean state such as the Holocene, for which El Nino Southern Oscillation (ENSO) or North Atlantic Oscillation (NAO) serve as examples, varying modern climate on time scales of 3 to 15 years on subcontinents to continents.
- ◆ Sequences of abrupt changes during the last glacial named Dansgaard/Oeschger (D/O) events (Figure 1) and Heinrich events, which occur in 3 to 100 years up to global extent. D/O events typically start with an abrupt warming of Greenland by 5–10°C over a few decades or less, followed by gradual cooling over several hundred to several thousand years. This cooling phase often ends with an abrupt final reduction of temperature back to cold (stadial) conditions. D/O climate change is centered on the North Atlantic and on regions with strong atmospheric responses to changes in that area, and shows only a weak response in the Southern Ocean or Antarctica (Ganopolski and Rahmstorf, 2001a).
- ◆ Slow variations that are probably caused by changes in parameters of the Earth's orbit are often referred to as Milankovich circles. Typical time scales are 20,000 to 400,000 years and the changes are global. Close inspection of these cycles as well as modeling results indicate that we are presently enjoying an unusually quiet period in the climatic effect of these cycles, owing to the present minimum in eccentricity of the Earth's orbit. The next large change in solar radiation that could trigger a new ice age is probably tens of thousands of years away. If this is correct, it makes the Holocene an unusually long interglacial, comparable to the Holstein interglacial that occurred around 400,000 years ago when the Earth's orbit went through a similar pattern (Rahmstorf, 2001).

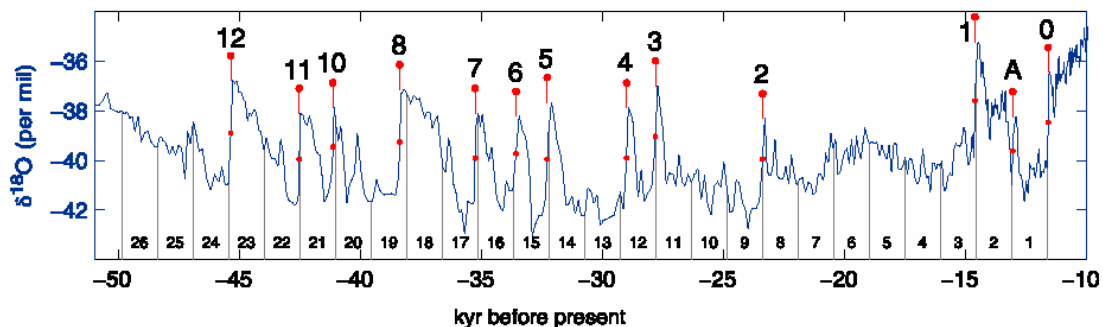


Figure 1: Daansgaard-Oeschger events (numbered) displayed as oxygen-isotope fluctuations in Greenland ice cores (GISP2). Source: Rahmstorf (2003).

But besides these three major ways of climate change identified in the Earth's history the recently added human-induced alteration of the climate system has to be taken into account.

By now, it is scientifically clear that human emissions and land-use practice account for the major part of global warming in the last century (IPCC, 2001a). The global annual surface temperature has increased by 0.6°C since the late 19th century and there is evidence that the shown rate of warming is larger than at any other time during the last 1,000 years. Furthermore, the 1990s are likely to have been the warmest decade of the millennium in the Northern Hemisphere, and 1998 is likely to have been the warmest year in this period (Figure 2). Besides temperature, a global decrease in snow cover, decreasing diurnal temperature ranges and increasing amount of cloud, precipitation and total water vapor has been recorded (IPCC, 2001a).

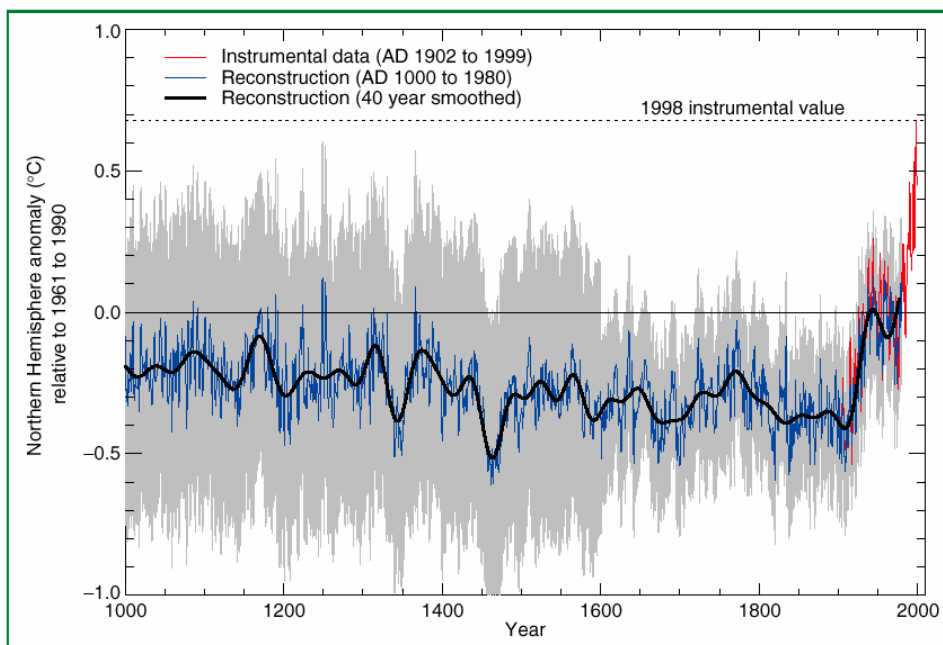


Figure 2: Millennial Northern Hemisphere (NH) temperature reconstruction (blue: tree rings, corals, ice cores and historical records) and instrumental data (red) from AD 1000 to 1999. A smoother version of NH series (black), and two standard error limits (grey shaded) are shown. Source: IPCC (2001a).

Greenhouse gases (GHGs) play a crucial role in the human alteration of the Earth's climate system and are understood, in this context, as a gaseous composition of the atmosphere modifying climate through the alteration of the radiation budget (Shugart, 1998). Over the millennium before the Industrial Era, the atmospheric concentration of GHGs remained relatively constant, whereas since then they show a rapid increase (Figure 3). Besides others, CO_2 is a major contributor to the GHG-induced radiative forcing and is of special interest because of large-scale human emissions and relatively long atmospheric lifetime (depending on the removal process, varying between 5 to 200 years). The atmospheric concentration of CO_2 has increased from 280 ppm in 1750 to 367 ppm in 1999 (31%). Today's CO_2 concentration has not been exceeded during the

past 420,000 years and likely not during the past 20 million years. Furthermore, the rate of increase over the past century is unprecedented, at least during the past 20,000 years (IPCC, 2001a).

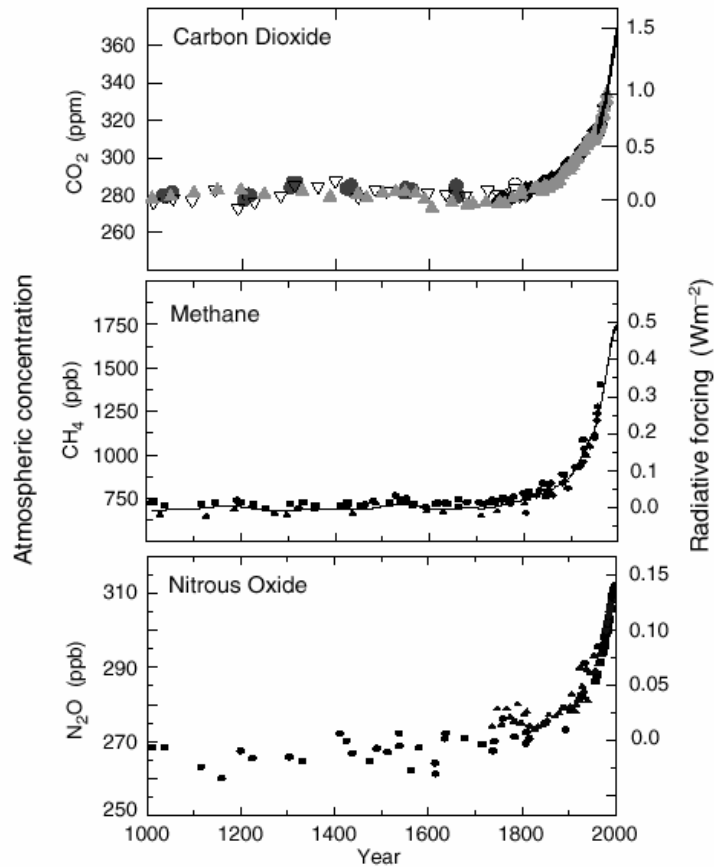


Figure 3: Records of changes in atmospheric GHG concentrations and their respective radiative forcing. Reconstructed from ice core and firm data, supplemented with data from direct samples. Source: IPCC (2001a).

However, the study of paleo-proxies reveals growing evidence for large, rapid and irreversible changes to have occurred in the past without human contribution, supposedly involving many processes and feedbacks interacting in complex non-linear ways. These interactions have the potential to push the climate system across a threshold, if it is perturbed sufficiently (IPCC, 2001a). Chaotic processes in the climatic system may allow the cause of such threshold crossing to be undetectably small, although greenhouse warming and other human alterations of the Earth’s system may increase the possibility of such events entitled abrupt climate changes (Alley *et al.*, 2002). However, since the extent of “abrupt” might differ from its common use when talking about climate time scales, this term needs a closer definition. Following the National Research Council on Abrupt Climate Change (Alley *et al.*, 2002) abrupt climate change is characterized by a threshold being crossed forcing the system into a new state at a rate determined by the climate system itself and faster than the cause. From the point of view of societal and ecological impacts, abrupt climate change is understood as significant change in the climate relative to the accustomed or background climate, having sufficient impacts making adaptation difficult.

At this point, it is important to clarify that rapid coolings, such as being the subject of this study, are not a contradiction to the recent global warming. Abrupt regional cooling and gradual global warming can unfold simultaneously (Gagosian, 2003). Moreover, greenhouse warming has the (paradox sounding) potential to make abrupt climate changes more likely.

3.1.2 Paleoclimatic Evidence for Abrupt Climate Change

Evidence of past rapid changes in the Earth's climate system can be found in different kinds of paleoclimatic records. What is common to all of these records is to rely ultimately on the use of present or recent instrumental records as the key to the past — using them for validating periods that overlap (Mann, 2002). To accomplish this, an assumption of constancy is applied in comparing recent data with paleo-data. Basically one can distinguish between three kinds of paleoclimatic indicators (following Alley *et al.*, 2002):

- ◆ Physical paleoclimatic indicators often rely on the fewest assumptions and have therefore high significance in the reconstruction of past climate conditions.
- ◆ Isotopic indicators are widely used since they have proved sensitive indicators of the paleoclimate.
- ◆ Biological indicators of environmental conditions typically involve the presence or absence of indicator species or assemblages of species.

Any of these different data requires age estimation, which can be done in different ways. One common way is to use the annuality of events, such as plant growth, precipitation or sedimentation patterns to identify layers, but there is also the widely used C14 method that can date exactly.

Ice cores, for instance, give access to paleoclimate series that include local temperature and precipitation rate, moisture conditions, wind strength and aerosol fluxes of marine, volcanic, terrestrial, cosmogenic and anthropogenic origin. Furthermore, they are unique in providing direct records of past atmospheric traces of gas compositions through entrapped air inclusions (Petit *et al.*, 1999). The use of multiple indicators increases the reliability of paleoclimate reconstructions (Alley *et al.*, 2002) as well as provides the possibility of reconstructing a whole “paleo-situation” describing vegetation, climate, atmospheric gas composition, etc.

Although proxy records of past climate show reoccurring rapid climate shifts, from now on I will concentrate on the two most recent ones, because they can be tracked clearly in different kinds of paleoclimatic data and have received extensive study. However, a similar pattern of change (in size, rate and extent) occurred more than 24 times during the 110,000 year-long interval covered by the Greenland ice core record (Alley *et al.*, 2002).

3.1.2.1 The Younger Dryas event

The Younger Dryas cold snap could be considered as the best documented example of abrupt climate change, because of its millennial duration and its extensive geographic

coverage. It is understood to have nearly been a global event with an emphasis on the northern hemisphere and is prominent among all different kinds of paleo-data, dated with 11,000 to 10,000 C14 years (12,500 to 11,500 calendar years) before the present day (Peteeet, 2000).

After a 4,000 year-long transition process out of the last ice age, temperatures were about to reach modern ranges when the Younger Dryas climate reversal took place. Hemispheric to global climate shifted dramatically, in many regions by about one third to one half of the difference between ice age and modern conditions, with much of the change occurring over a few years (Alley (2000) in Alley *et al.*, 2002). The North Atlantic polar front re-advanced far southward to approximately 45°N latitude (only 5 or 10° north of the glacial maximum) (Broecker *et al.*, 1988). The rapidity and extent of this climate shift are in accord with ice core and marine sediment records as well as with paleo-botanic ones (Figures 4 and 5). Actually this paleo-botanic evidence provides the name for this major cooling event. The leaves of the arctic-alpine tundra indicator *Dryas octopetala* (Mountain Avens) reoccur in layers above the remains of trees and shrubs, indicating vegetational shift in response to this dramatic cooling (Peteeet, 2000). Data from the Greenland ice cores (GRIP, 1997) indicate that cooling into the Younger Dryas took place in a few prominent decade-long steps, with a magnitude of up to 6°C in about 20 years (GRIP, 1997), whereas warming at the end of the event occurred primarily in one especially large step of about 8°C in about 10 years (Alley *et al.*, 2002). Regionally, drops in temperature as dramatic as -16° were recorded (Alley, 2001). Major alterations of the precipitation pattern with dry conditions during the millennial duration of the event and a rapid increase at the end of it is significant for the Younger Dryas (Alley *et al.*, 2002 record a doubling of snow accumulation in Greenland in only three years at the termination of the cold phase).

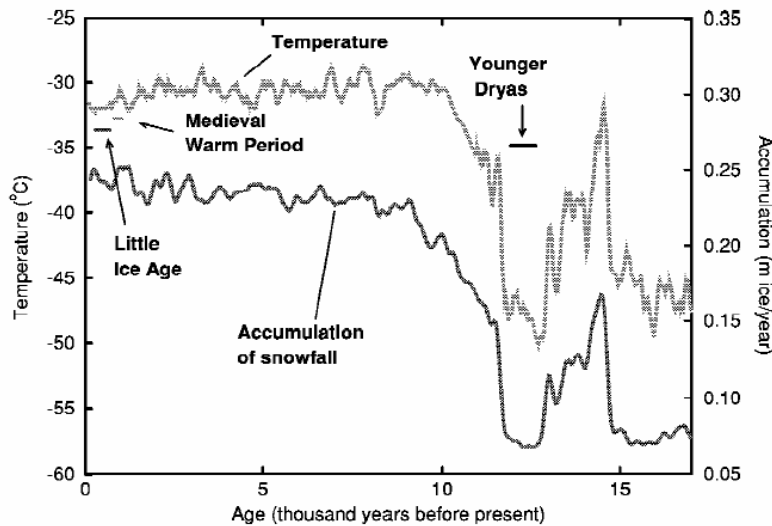


Figure 4: Climate records from Greenland ice-cores indicating the Younger Dryas event. Source: Alley *et al.* (2002).

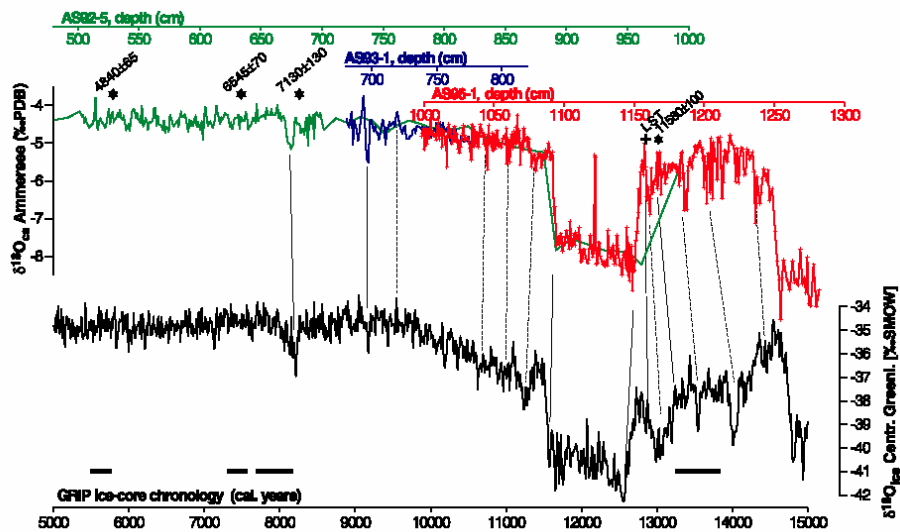


Figure 5: Climate variations including the Younger Dryas event represented in oxygen isotope values of calcite shells from central European sediment cores (top graphs) and ice cores (GRIP) in Greenland (lower graph). Source: von Grafenstein *et al.* (1999).

Generally, the Younger Dryas can be pictured as a cold, dry and windy time with transition periods at the onset and termination during which wind speed, precipitation, temperature, and sea ice changed on sub-decadal time scales (Taylor *et al.*, 1997).

3.1.2.2 The 8.2 ka event

The most prominent Holocene climate event in Greenland ice-core proxies (Alley *et al.*, 1997) as well as in central European high resolution pollen studies (Tinner and Lotter, 2001) occurred 8,000 to 8,400 calendar years ago and is therefore called the 8.2 ka event. This event can also be tracked nearly all over the globe — proxy records from the tropics to the north polar region show a cool, dry and windy event of similar age (Alley *et al.*, 1997, Thompson *et al.*, 2002). The significance of the 8.2 ka cooling is amplified by the fact that the climate before the rapid transition was similar or even slightly warmer than today (Alley *et al.*, 2002). It was estimated that the amplitude was about half of the Younger Dryas with temperature depression locally reaching up to 10°C. However, this amplitude most likely just reached over Greenland — for continental Europe a drop of about 2°C was estimated (Keigwin and Boyle, 2000; Tinner and Lotter, 2001; Alley *et al.*, 2002) (Figure 6).

The 8.2 ka event has received intensive examination because of its singularity in the Holocene and there is a likely explanation of what might have triggered this. Available data indicates that not all transitions were identical, but most abrupt changes seem to have exhibited broadly similar patterns (Alley, 2000). Possible triggers for past and future abrupt climate change are therefore reviewed in the following section.

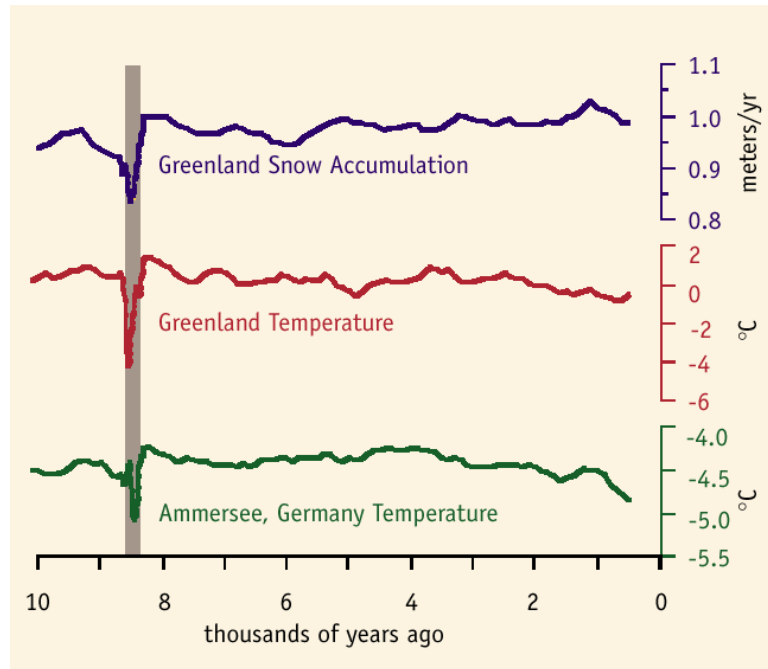


Figure 6: The 8.2 ka event mapped in snow accumulation and isotopically inferred temperature from GISP 2 record (Greenland) and fossil shell isotopes in sediments of Lake Ammersee (Germany). Source: PAGES (2001).

3.1.3 Possible Triggers of Abrupt Climate Change

Two major characteristics of the climate system are crucial when it comes to identifying possible triggers of abrupt climate change, namely non-linearity and multiple equilibria. Both attributes are well recognized in climate science, but make the estimation of probabilities and uncertainties difficult. An undetectably small cause could lead to a huge change as well as to a minor oscillation within the band.

Therefore, general characteristics of possible triggers are the ability to force the system across a threshold, leading to amplification and allowing a considerable persistence of the effects caused (Alley *et al.*, 2002). Key components of the climate system are oceans, ice shields and the atmosphere, where the latter plays a crucial role in connecting the others by serving as mediator or vector. The investigation of these complex relations and interactions is founded on models that are able to reproduce the behavior of and interrelation between all of the crucial components. The knowledge gathered from Stommel's first two-box model (Stommel (1961) in Alley *et al.*, 2002) to modern global circulation models (GCM) suggests three fundamentally different ways that can lead to abrupt climate change (Alley *et al.*, 2002).

First of all a rapidly varying external parameter or forcing can alter the Earth's system, where the external parameters can be a meteorite impact or a major change in the earth-sun liaison. However, these mega events are basically not related to climate itself and are therefore not further investigated here. Considerable crossing of a threshold is another possibility — a sudden disintegration of ice shields or a major freshwater release into the ocean can serve as examples in this category. Such a surge of freshwater

with its consequences on the thermohaline circulation (THC) is believed to have been the trigger of the previously described 8.2 ka event (Alley *et al.*, 1997). And third, there is reason to believe that the climate system contains chaotic elements — regime transitions can occur spontaneously in chaotic systems. Especially near a system-imminent threshold such chaotic processes reduce predictability or contribute to the risk of transition to another equilibrium (Knutti and Stocker, 2002).

Despite the manifold ways abrupt climate change can be triggered, there is only one well-developed theory established through extensive model studies by now (Alley *et al.*, 2002). Sudden changes in the thermohaline circulation are thought to have played a major role in past events such as the Younger Dryas, the 8.2 ka event as well as in examined Dansgaard/Oeschger and Heinrich events. A short review over ocean-climate interactions should be given here in order to comprehensively summarize our current knowledge on possible triggers of abrupt climate change.

The thermohaline ocean circulation is a major heat transport mechanism, which causes the relatively mild climate in the North Atlantic region including Europe in modern times (Rahmstorf, 1999). The driving forces of this current are differences in temperature and salinity in ocean water (Rahmstorf, 1997; Clark *et al.*, 2002; Broecker, 1995; Keigwin and Boyle, 2000). They steer a giant overturning motion where warm surface water flows north throughout the Atlantic (because of the differential heating in high and low latitudes). The Pacific is not showing a similar pattern because there is less salty surface water and lower surface water temperature in the North Pacific (Broecker, 1995; Weaver *et al.* (1999) in Clark *et al.*, 2002). In the Atlantic, however, this mechanism, referred to as Gulf Stream or North Atlantic Current, provides 10^{15} Watt of heat to the northern North Atlantic region (Rahmstorf, 1997). The cooled and therefore denser water sinks into the abyss in high latitude regions of the Atlantic — a mechanism called North Atlantic Deep Water (NADW) formation. The ocean conveyor belt is “closed” by a cold-water current at a depth of about two kilometers flowing south along the Hudson Bay (Figure 7). The surface water flowing north is on average 8°C warmer than the cold deep water flowing south (Broecker, 1995).

The crucial point of the whole system is that the strength of the circulation, and thus the rate of heat transport, depends on relatively small density differences, which in turn depend on a subtle balance between cooling at high latitudes and the input of less dense freshwater from rain, snowfall, melting and river runoff. This sensitive system is believed to have several thresholds for freshwater input or surface warming (which influences water-density as well as the freshwater budget) and acts clearly nonlinear. The concept of multiple equilibria of THC and the transitions between these states is widely accepted and seen as a possible explanation of past abrupt climate changes (Clark *et al.*, 2002). For the above mentioned events at 12 ka and 8.2 ka, respectively, strong evidence points at the importance of the THC as their trigger.

Furthermore, there is indication of at least eight invasions of freshwater into the North Atlantic, either as huge icebergs or floods from melt-water lakes. A well-known example for the latter is Lake Agassiz, a melt-water lake on the North American continent that extends up to 840,000 km² (Teller *et al.*, 2002) in the topographic depression created by the weight of the retreating ice cap during the transition from the last ice age. Changes in the surrounding topography caused by further melting of ice

triggered a catastrophic drop in lake level (estimated at approximately 100 m), releasing a huge amount of freshwater (up to 163,000 km³) to the North Atlantic (Broecker, 1995, 1999; Teller *et al.*, 2002). This outburst of melt-water is understood to have led to a complete shutdown of the THC, which was most likely the trigger of the Younger Dryas event (Alley *et al.*, 2002; Hostetler *et al.*, 2000). A similar input, also linked to a Lake Agassiz freshwater flood, under a close-to-modern climate-constellation marks the onset of the 8.2 ka event (Teller *et al.*, 2002). Simulating the effects of fresh water forcing to the North Atlantic indicates that the ocean convection is most sensitive to high latitude inputs, near the region of present day deep-water formation (Manabe and Stouffer, 2000; Keigwin and Boyle, 2000; Manabe, 2001). In tracking the 8.2 ka event with their three dimensional atmosphere-sea ice-ocean model Renssen *et al.* (2002) find further support for the hypothesis that freshwater-induced THC-weakening caused this event. Several simulation studies add evidence to the hypothesis of THC being a prominent element of abrupt climate change as well as demonstrate that simulated climatic changes associated with freshwater perturbations are realistic (Rahmstorf, 1999; Knutti and Stocker, 2002; Ganopolski and Rahmstorf, 2001b) (Figure 8).



Figure 7: The ocean currents in the North Atlantic. Source: McCartney *et al.* (1996).

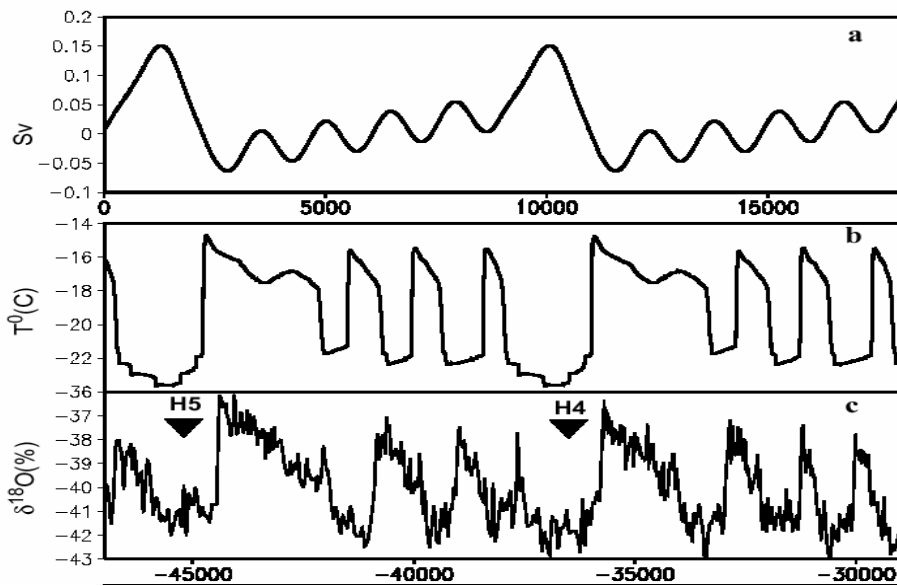


Figure 8: Freshwater-induced temperature fluctuations simulated with a climate system model of intermediate complexity compared to oxygen isotope data from GRIP; (a) magnitude of freshwater forcing; (b) simulated temperature; (c) oxygen-isotope records representing temperature fluctuations. Source: Ganopolski and Rahmstorf (2001b).

However, there is another factor to consider, possibly not being a trigger itself, but having the ability of pushing the system towards a threshold or catalyzing a trigger — human-induced alteration of the Earth’s climate. While the strength of the THC has been relatively stable during the Holocene, GHG warming may push it towards an instability threshold. As discussed, it is the consensus of the scientific community that the changes observed over the last few decades are most likely human-induced, and the trend towards a warmer Earth is understood to continue. This is likely to lead to an enhanced atmospheric water cycle with a higher amount of precipitation in the North Atlantic area as well as increased melt-water runoff from Greenland and other glaciers (Rahmstorf, 2001; Alley *et al.*, 2002). Those two effects, in concert, lead to an increased buoyancy of the North Atlantic surface waters, which in turn weakens the THC. These facts can already be tracked in measurements; Dickson *et al.* (2002) observed salinity data for the northern Atlantic and found a strong decline during the last four decades, referring to it as “the largest changes observed in the modern instrumental oceanographic record” (Figure 9).

Moreover, signs of a possible weakening of the THC are already measured — a study focusing on the Faroe Bank reports a decrease in deep water flow of at least 20% relative to 1950 (Hansen *et al.*, 2001). Prolonging the current development into the future using model simulations, Stocker (1999) reports a total shutdown of the thermohaline circulation between two and four times the pre-industrial CO₂ level (Figure 10) with the climate system moving to a structurally different state. Comparable results are also found by other authors (Ganopolski *et al.*, 1998; Rahmstorf and Ganopolski, 1999; Vellinga and Wood, 2001; Schiller *et al.*, 1997) as well as some intermediate stages are reported (i.e., partly shutdown or weakening; Manabe and Stouffer, 2000; Rahmstorf, 2001; Knutti and Stocker, 2002).

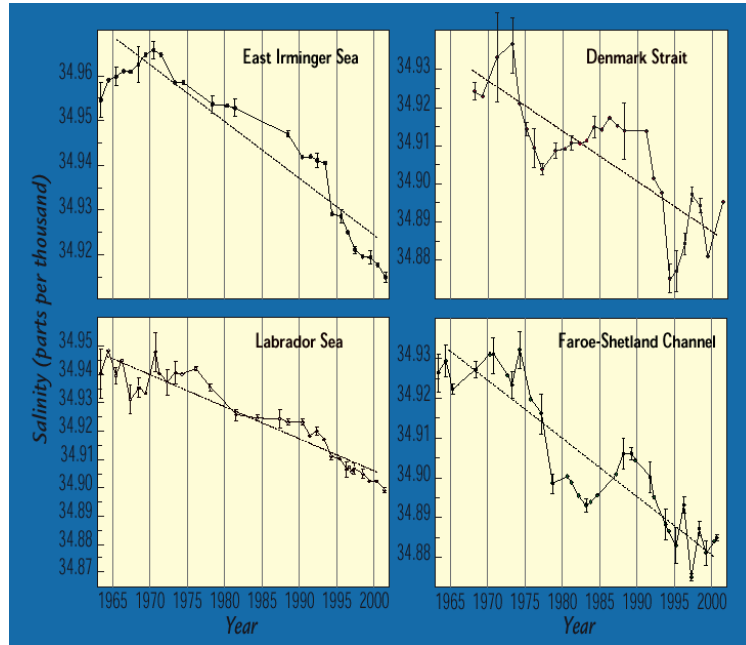


Figure 9: Decrease in salinity in the northern Atlantic. Source: Dickson *et al.* (2002).

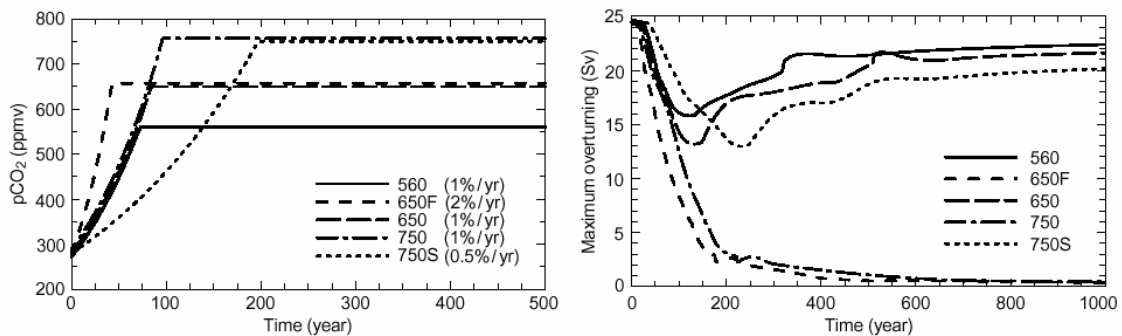


Figure 10: Scenarios of future CO₂ increase (left) and their impact on the overturning volume transport of the thermohaline circulation (right). Source: Stocker (1999).

Further investigation finds a correlation between the rate of CO₂ increase and the stability of the thermohaline circulation and reduced predictability of the THC near the (still unknown) stability threshold. Knutti and Stocker (2002) showed that even after the initial forcing (in their example a doubling of the atmospheric CO₂) has disappeared, rapid transitions may occur with long delays (Figure 11).

One can conclude by saying that threshold behavior, multiple equilibria and non-linearity limit the exact explanation of past events as well as the accurate prediction of possible future abrupt climate change. But the knowledge gathered through paleo-data as well as through distinctive models suggests that abrupt climate changes were prominent in the past and are not only possible but likely in the future (Alley *et al.*, 2002).

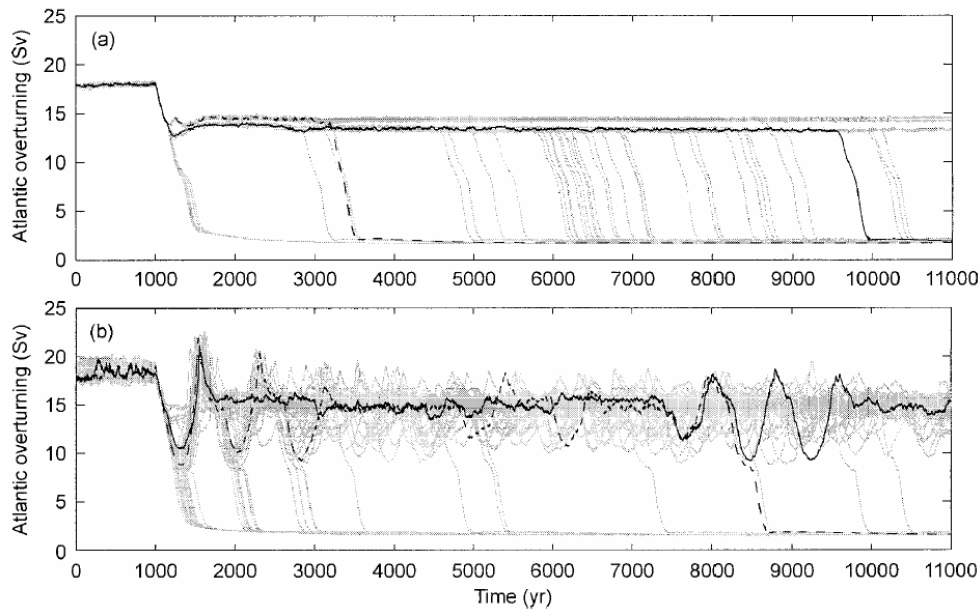


Figure 11: The impacts of two freshwater forcings (a) and (b) on the Atlantic overturning. Spontaneous shutdown of the THC near its stability threshold long after the forcing has stopped is displayed. One hundred simulations (grey) are shown for each forcing where some are highlighted in black for clarity. Source: Knutti and Stocker (2002).

3.1.4 Forest Response to Abrupt Climate Change Recorded in Paleo-data

In general, a close relationship between vegetation and climate can be established beyond the Last Glacial Maximum. A review of pollen proxies leads to the conclusion that when the combination of large-scale climate controls was unique, so too was the vegetation (Whitlock and Bartlein, 1997). The fact that plant history does not repeat itself in an analogue to climate history shows again the advantages of using ecosystem models to provide a comprehensive window into the future. However, paleo-botany provides valuable results with which to compare and reveals the general ways of plant reaction to climatic change — adaptation, migration and extinction. All of these are prominent in paleo-botanic records, but for the Quaternary range the shifts are tracked as the major response of woody vegetation to climate changes (Davis and Shaw, 2001).

A first interesting fact that can be tracked in pollen records around the world are differences in response to similar events. The Younger Dryas cold snap may again serve as an example. Yu (2000) finds no forest transformation related to the Younger Dryas in the Great Lakes region of North America because the location was in the middle of a broad *Picea*-belt at that time. Here, the Younger Dryas only led to changed understory composition and forest structure, indicated by a slight increase in herb-pollen percentages (Yu, 2003). However, about 3,000 years later the study site was near a *Picea*-*Pinus* ecotone and showed a major alteration in vegetation cover (reoccurrence of *Picea*) due to a minor Preboreal Oscillation. This example indicates that distinct ecotones emphasize small changes in climate, whereas reactions to major events such as the Younger Dryas can be found all over the world. Newby *et al.* (2000), for instance, found a drop in pollen accumulation rates of about 70% as an effect of the substantial

cooling during his observations in southeastern Massachusetts. In addition, Shuman *et al.* (2001) recorded a return of spruce (>50%) after a phase of pine-domination related to cool climates during the Younger Dryas for the same region. In the Appalachian, a decrease in deciduous broadleaved tree taxa and *Pinus strobus* pollen is reported, simultaneously with a re-expansion of *Abies* (Kneller and Peteet, 1999). Reports from Russian sites (Andreev *et al.*, 2002) find a decline in tree pollen (*Betula nana*) and an increase of herb taxa. Studies conducted in central Europe are of particular interest for this work, and show similar patterns as described above (Figure 12). For the Swiss mountains, several studies indicate a general decrease in pollen and other macrofossils (Tobolski and Ammann, 2000) and a relative increase in *Pinus* pollen accompanied with rising herbaceous species (Wick, 2000). Pine dominated forests are also reported from other studies focusing on the European Alps (Kral, 1979; Magny *et al.*, 2003) and forests appear to have been less dense during the Younger Dryas period. Furthermore, the study of Wick (2000) supports the theory that vegetational response to abrupt climate changes is more distinct near ecotones, i.e., at medium and high altitudes compared to lowlands.

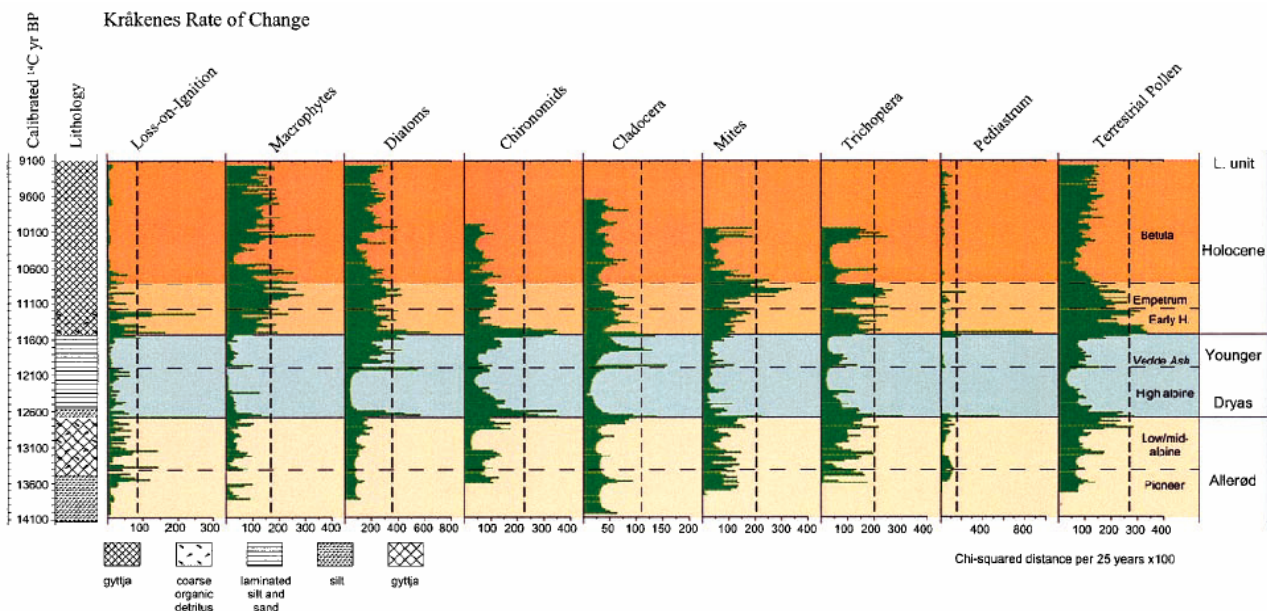


Figure 12: Rates of change at Kråkenes Lake, Norway, over 25 calendar year intervals for the fossil groups and percentage loss-on-ignition at 550°C. Changes in vegetation cover are indicated to the right, revealing decreased terrestrial pollen during the Younger Dryas. Source: Birks and Ammann (2000).

An event with similar rapid transition but less extreme temperature changes the 8.2 ka event interrupts the period of the so-called Boreal thermal optimum (Andreev *et al.*, 2002), which was characterized by the strong abundance of Hazel in central Europe (Kral, 1979). The rapid changes at 8.2 ka led to a sudden collapse of *Coryllus avellana*, within 150 years from 30 to 13% and from 40 to 16%, respectively, at two studied sites in Switzerland and Germany, accompanied by a rapid expansion of the pioneer species *Pinus* and *Betula* (Tinner and Lotter, 2001). As reported for the Younger Dryas, a

decrease in forest density deflected from declining pollen is reported from other studies and regions (Sweden: Hammarlund *et al.*, 2001; Russia: Andreev *et al.*, 2002).

Surprisingly, reported response times are short. Tobolski and Ammann (2000) compared the biotic responses of trees to those of algae, which both reacted within 20 years after the very beginning of the shift in oxygen isotopes at the onset of the Younger Dryas. Time-lags from other studies contribute to this finding — *Betula*, *Pinus* and *Coryllus* responded in less than 40 years, sometimes even within the sample resolution (Wick, 2000; Tinner and Lotter, 2001). The influence of decreased precipitation during rapid coolings is found to be pronounced, but effects differ according to the ecological characteristics of the tree species. Whereas lowland forests under continental conditions show decreased density related to drought (Kral, 1979) in the Alps the drought-stress was relieved because of the cooler conditions, which gave shade-tolerant species (Norway spruce, beech, fir) the chance to outperform *Coryllus* during the 8.2 ka event (Tinner and Lotter, 2001).

Not only was the spatial abundance of trees altered but also major tree level shifts are found in pollen-proxies. Rapid cooling resulted in deforestation of high altitude areas with timberline levels reported at 1000–1200 m asl (Wick, 2000) and 1200–1400 m asl (Kral, 1979), respectively. This drop in tree level is accompanied by increased erosion, marked by a pronounced erosion surface in the according sediment layer (Magny *et al.*, 2003; Yu, 2000).

One can conclude by saying that forest ecosystems reacted quickly to abrupt climate changes in the past — large amplitude changes such as the Younger Dryas are recorded in all altitudes and ecosystems, whereas minor oscillations are detectable in distinct ecotones and at sensitive altitudes (Birks and Ammann, 2000).

3.2. The PICUS v.1.3 Model

3.2.1 Introduction

Global change involves the simultaneous alteration of several key parameters controlling forest dynamics. And since predictions of possible consequences through direct multisite, multifactorial experiments are impossible (Aber *et al.*, 2001), models serve as the best available tools for such assessments.

In the following section the core functions of the forest dynamics model PICUS v1.3 are described with an emphasis on the abiotic environment, which is crucial to this study. PICUS has been developed at the Institute of Silviculture at the University of Natural Resources and Applied Life Sciences (BOKU), Vienna and the model version 1.3 used for this paper is as yet unpublished. The general structure is, however, an analogue to the second version, which is published in Lexer and Hoenninger (2001) and applied in several papers including Lexer (2001) and Lexer *et al.* (2001). Therefore, this section introduces the model structure of version 1.2 as described in the literature but interlaces personal communication with the model author on the updated version 1.3.

PICUS is a spatially explicit three-dimensional gap model developed for site-specific decision support regarding long term vegetation development in alpine landscapes (Lexer and Hoenninger, 2001). It is named after a Roman god of the forest and is the result of a comprehensive review of strengths and weaknesses of existing gap models and modified or new model formulations. The base elements of the simulation are patches of 10×10 m, extended into the third dimension with crown cells of 5 m depth (Figure 13).

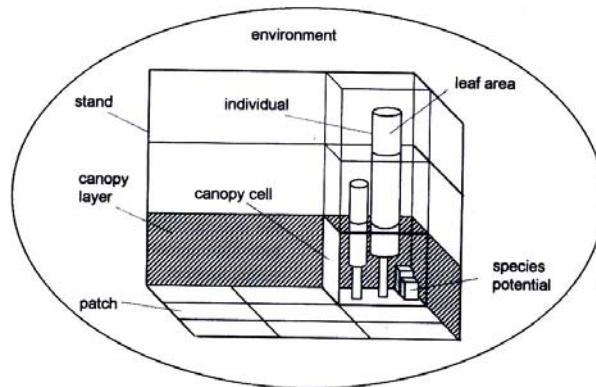


Figure 13: Schematic picture of the spatial structure of PICUS. Source: Lexer and Hoenninger (2001).

In contrast to conventional patch models, PICUS simulates forest changes over time as an interactive unit instead of a series of independent plots. A detailed model of above- and within-canopy light regime, interactions among neighboring patches as well as the consideration of slope, surrounding topography and orientation contribute to this improvement. The core structure of version 1.2 can be seen in Figure 14.

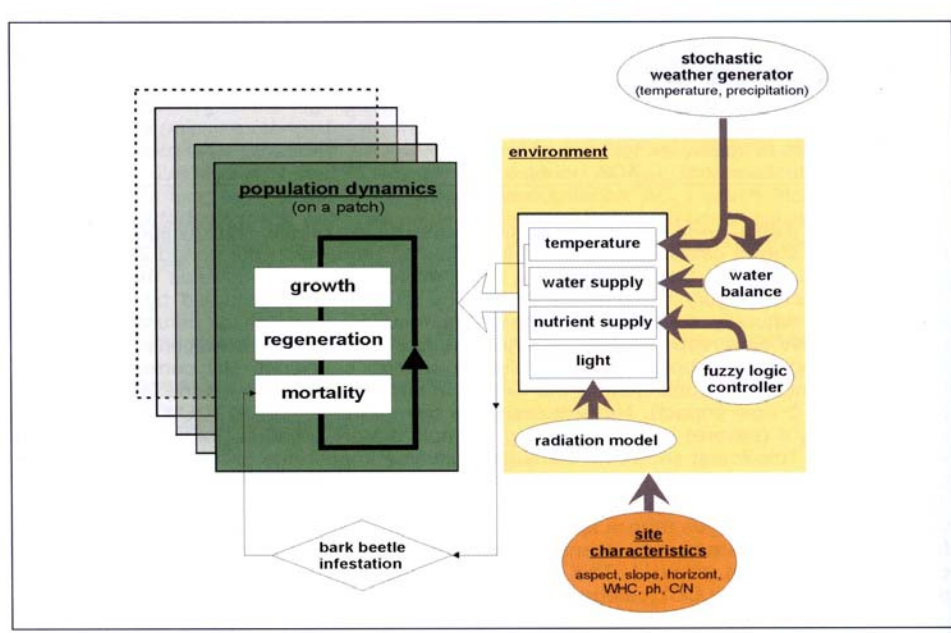


Figure 14: Structure of the PICUS model. Source: Lexer *et al.* (2001).

3.2.2 Population dynamics

3.2.2.1 Tree growth

As in the majority of gap models, actual individual tree growth is derived from a species specific growth potential, which is modified by environmental constraints. Tree dimensions are defined by diameter at breast height (1.3 m), tree height, height to the live crown, leaf weight and leaf area. The potential diameter and height growth are derived from data on open grown trees and adjusted with environmental operators, taking both compensation and intensification among environmental factors into account. While these features of the classical gap model approach are well suited to characterize the inter- and intra-competitive relationships among individual trees, the results of conventional gap models were less reliable with regard to biomass production estimates. Therefore, a main feature of the model version v1.3 is that the “bottom-up” growth estimated for every individual tree in the simulation is corrected with a second, spatially non-explicit physiology based estimate of stand level NPP (net primary productivity), referred to as “top-down” growth. Top-down stand growth is derived from fractioning total incoming radiation into utilizable photosynthetically active radiation determined by environmental constraints, as first introduced by Landsberg and Waring (1997) in their model 3PG. Their study showed that canopy quantum efficiency is stable about the value of 1.8 g C/MJ for a variety of different forest ecosystems and that at stand level it may be possible to use relatively simple environmental modifiers to model the effects of stomatal closure due to high atmospheric vapor pressure deficit, soil drought and sub-freezing temperature as well as temperature on carbon fixation processes. Bottom-up and top-down submodels are linked via the leaf area of the simulated forest. Thus, competition at the tree level is modeled within the bottom-up model component while primary production is estimated by the top-down approach. This advanced growth prediction leads to increased accuracy of the PICUS model and adds value to it as a decision support tool for forest resource managers.

3.2.2.2 Recruitment

New trees are generated by a recruitment submodel and are explicitly modeled beyond a diameter threshold set at 1.0 cm. Seed production and dispersal by overstory trees is considered explicitly, where seed production of an individual tree is calculated as the function of tree size, species seed production characteristics and light consumption. Seed dispersal of each seed-producing tree is modeled as a cone-shaped density function around the center of the tree’s patch. The seed potential of each patch is modified by a species specific germination rate and a number of environmental filters including heat sum, frost occurrence, chemical site properties (pH, C/N-ratio), water supply and available light at the forest floor. A uniform random number decides on the regenerating species for every position open to recruitment with a restriction of a maximum number of 100 individuals per patch.

3.2.2.3 Tree mortality

Mortality is modeled as a stochastic process where trees can die either from an inherent risk of death (chance events) or from growth-related mortality. The hierarchical pattern of carbon allocation shows that diameter increment is of low priority and therefore an indicator of reduced tree vitality. Thus, when a tree fails to realize a specific minimum

diameter increment for a number of successive years, it dies in the simulation. For Norway spruce, bark beetle mortality is modeled explicitly with a two-stage stand risk model integrated into PICUS.

3.2.3 Abiotic environment

The abiotic environment in the bottom-up component of PICUS is represented by the temperature regime, soil moisture supply and nutrient status of the site as well as the light regime.

3.2.3.1 Temperature

The temperature regime is represented by a heat sum (GDD) above a threshold (5.5°C) for net photosynthesis and winter minimum temperature. The response of tree species to GDD was derived from analyzing data of the Austrian Forest Inventory (AFI) according to an index of growth vitality, which led to second-order polynoms for the “cold” left-hand side of species response. Because growth at a species’ southern range limit is not necessarily restricted by temperature conditions per se, there is no constraint at super-optimal thermal conditions. The coldest month of a year is taken as an estimate for winter minimum temperature, which is considered regarding the frost tolerance of a tree species as well as concerning the chilling requirements for species showing dormancy.

3.2.3.2 Soil moisture

In this module, plant water supply is calculated from a site-specific water balance. Soil water content is the state variable derived from integration input and output flows, such as precipitation, snow melt, evapo-transpiration and runoff. The ratio of evaporative demand and supply integrated over the growing season is used to calculate an index indicating drought induced stress. Again, data from the AFI was used to parameterize soil moisture response functions for the main alpine tree species.

3.2.3.3 Radiation regime

The radiation regime of the PICUS model distinguishes between above-canopy radiation and radiation within the canopy. Above the canopy the radiation submodel takes into account shielding effects of the surrounding topography as well as modified light regime on slopes. Diffuse and direct radiation is treated separately. Radiation within the canopy considers the sun position as well as the neighboring patch, which leads to a fairly complex computation. A detailed description is given in Lexer and Hoenninger (2001). The individual tree response to light is calculated as an index relating available light per tree crown to the corresponding value of an open grown tree of the same size.

3.2.3.4 Nutrient supply

In PICUS the variables, water holding capacity (WHC), pH-value, C/N-ratio and plant available nitrogen, characterize the nutrient status of a site. The abundant “soft” expert knowledge on the relationship between tree growth and soil variables was utilized by means of a fuzzy logic control unit. This rule-based approach led to the construction of

seven growth response categories, to which every species was assigned according to statements in ecological literature.

3.2.4 Recent applications of PICUS

Because of the unpublished status of PICUS v1.3 we can not refer to studies carried out with the latest version of the model. However, for the purpose of this work we carried out a sensitivity analysis in combination with preliminary validation runs, which is presented in the next section of this paper. Here, a concluding review of some applications of PICUS v1.2 is given.

In Lexer and Hoenninger (2001) the model was used to simulate the potential natural vegetation spatially explicit along a transect through the Eastern Alps. This study revealed the capability of the model to reproduce large-scale vegetation patterns as well as differentiated species behavior on varying soil types. Furthermore, structural parameters of the simulated equilibrium forests showed good agreement to literature in terms of stem number distribution (reversed J-shape) and maximal tree diameter. Lexer (2001) conducted another study using version 1.2 of PICUS to simulate the potential natural vegetation for Austria. As a result, PICUS was found to be able to sufficiently reconstruct the potential natural vegetation on plots of the Austrian Forest Soil Survey (AFSS) compared to expert assessment on the site. An additional analysis of a warming scenario also showed a plausible behavior of the model with results that were sensitive to a warmer and drier climate. A comprehensive assessment of the sensitivity of Austrian forests to climate change was carried out using PICUS v1.2 in Lexer *et al.* (2001). In an evaluation against the expert assessment of potential natural vegetation of the AFI, the model realistically captured the syn-ecological behavior of the major European tree species. In this work, PICUS was successful in assessing the impacts of climate change and identifying regions and conditions where the impacts of a changing climate might be severe.

Founded on these positive examples of PICUS applications to Austrian forests under stable as well as changing climate conditions, and having an enhanced version readily at hand, makes the PICUS v1.3 model a rational choice for this study.

3.2.5 Validation runs and sensitivity analysis of PICUS v.1.3

3.2.5.1 General

The quantitative output of the model simulation is compared to Austrian yield tables (Marschall, 1992) to crosscheck the model's behavior with regard to site quality and climate. Therefore, three soil categories are defined with respect to water holding capacity (WHC), carbon-nitrogen-ratio (C/N), pH-value and plant available nitrogen (see Table 1). The three altitudinal zones are used to test the model's response to varying climate conditions.

The simulation was carried out using the same initial stand conditions for all runs, namely a nine year old pure Norway spruce stand with 2309 stems/ha.

Table 1: Soil set-up for the validation runs.

	Low	Medium	High
WHC [mm]	100	150	200
pH	3.8	4.5	5.6
C/N	20	20	20
N [kg/ha*a]	50	75	120

3.2.5.2 Total yield and yield classes

In comparing the growth pattern of PICUS at a given total yield (= stand volume at certain age + cumulative volume of death/removed to this age) with the Austrian yield table the table “*Hochgebirge*” was found to be the closest match for both the montane and lower subalpine site. For the submontane site, the table “*Bayern*” fits the simulated stand quite well (Figure 15). Although both *Hochgebirge* and *Bayern* tables have straighter yield curves than the other Norway spruce tables available for Austria, young stands show a higher total yield at all sites and altitudes. However, this fact is possibly linked to the initial stand — from the age of about 60 years the yield table and simulation show virtually no differences.

The differing results for varying altitudes and soil qualities document the sensitivity of the model to these parameters and are first of all in good accordance with experimental observations (decreasing total yield with increasing altitude, decreasing total yield with decreasing soil quality, see Table 2 and Figure 16). A closer view reveals a higher sensitivity to soil parameters (range from 463 m³/ha to 1746 m³/ha at submontane climate) than to climate (represented by the different altitudes in this layout, ranging from 1424 m³/ha to 1746 m³/ha for high soil quality). However, the results are reflecting Herzberger’s (1996) findings regarding a relative Norway spruce optimum in the montane belt, that is, total yield decreases more from montane to lower subalpine than from submontane to montane. Though, these model derived yield classes (mean annual increment (MAI) at age 100) can hardly be compared to “real” conditions because the whole range of ecosystem changes linked with increased altitude are not taken into account in this sensitivity analysis. But a comparison of the relative changes found is of interest to back up the results of this study. Herzberger (1996) found a negative correlation between site class and altitude, but added that only little more than a quarter of the variation of site class can be explained through altitude (Figure 17). The slope of his regression is -0.005 , whereas we find -0.0037 , -0.0021 and -0.0005 for high, medium and low soil quality, respectively.

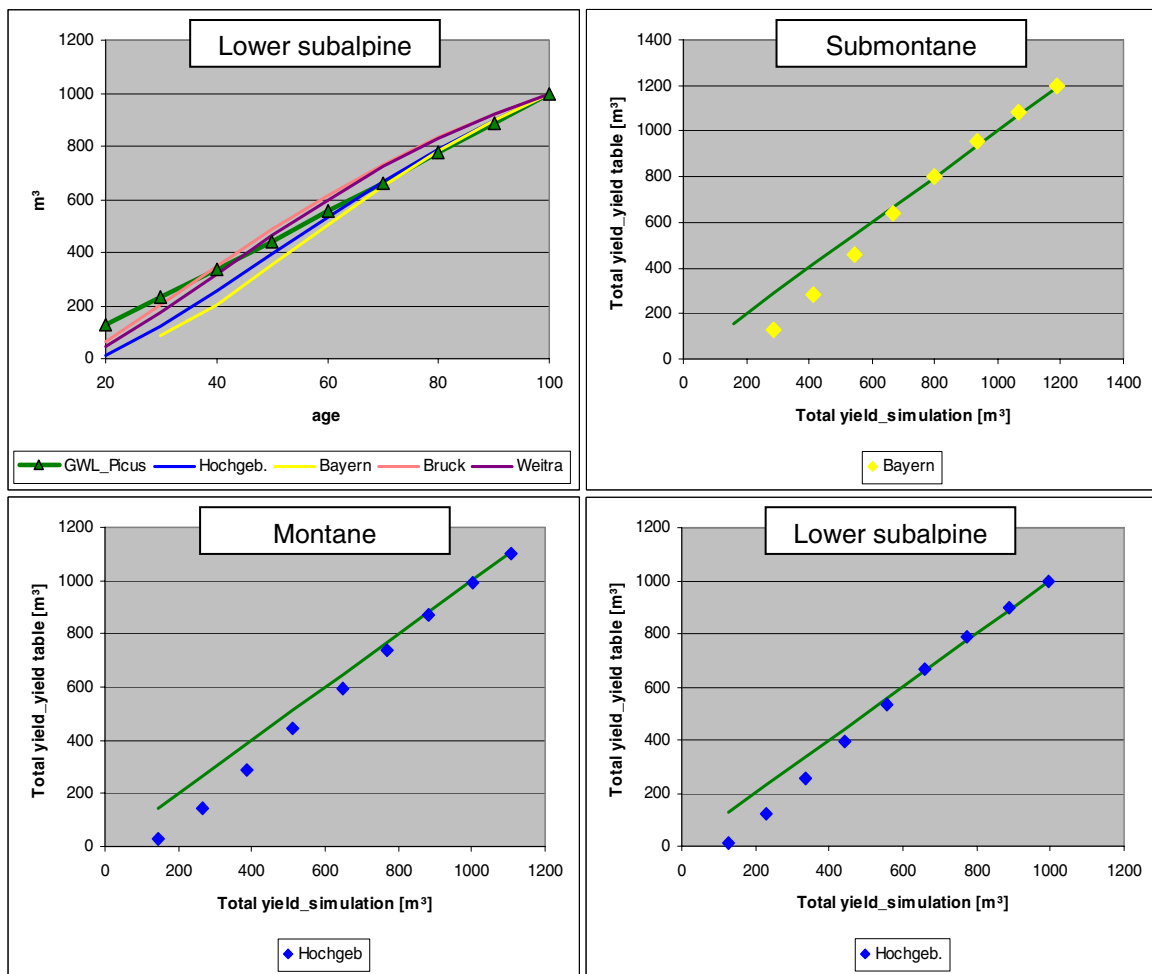


Figure 15: Top left: Simulated total yield at subalpine climate and medium soil quality and the respective four Austrian Norway spruce yield tables. Top right: Total yield cross plot of chosen yield table and simulation (submontane, medium soil quality). Bottom left: Total yield cross plot of chosen yield table and simulation (montane, medium soil quality). Bottom right: Total yield cross plot of chosen yield table and simulation (lower subalpine, medium soil quality).

Table 2: PICUS simulated mean annual increment at age 100 (equals dGz yield classes) in m³/ha*a.

	Low soil quality	Medium soil quality	High soil quality
Submontane	4.6	11.9	17.5
Montane	4.6	11.1	16.7
Lower Subalpine	4.2	10.0	14.2

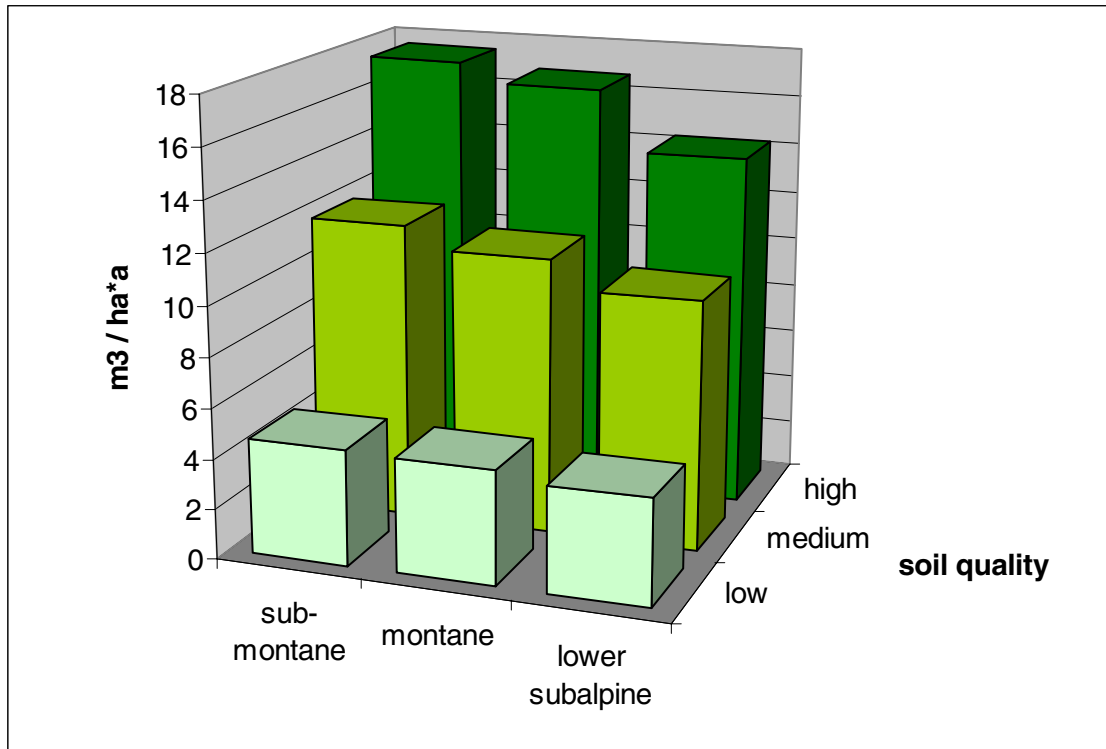


Figure 16: PICUS simulated MAI at age 100.

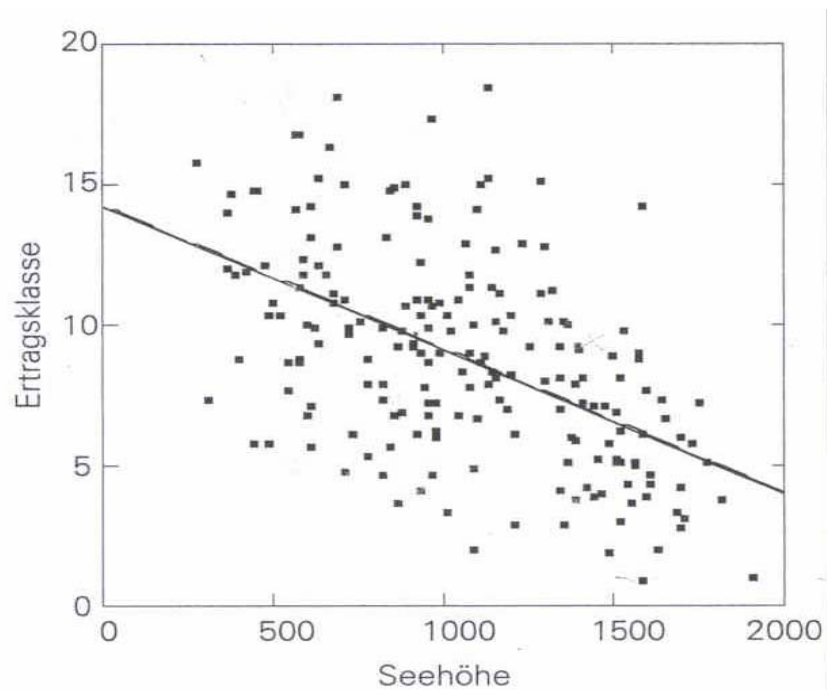


Figure 17: Site class (*Ertragsklasse*; mean annual increment at age 100 in m³/ha*a) over altitude (*Seehöhe*) derived from the Austrian Forest Damage Monitoring System. Source: Herzberger (1996).

3.3 Data and Scenarios

3.3.1 Climate

Because of their potentially high vulnerability to climate change this study focuses on European alpine forest ecosystems. Forests in the Eastern Alps have been subject to human use and alteration for centuries (Glatzel, 1994), which holds true for most parts of the European Alps (Theurillat and Gusian, 2001). An exceptional high ecological exposure to changes in climate and ever-growing human demands on the environment coincide at present and are anticipated to further grow in the near future. This background gives high significance to a first impact assessment of abrupt climate change to mountain forest ecosystems.

In the Alps, mean air temperature decreases regularly as elevation increases, at a rate of about 0.56°C per 100 m difference. Conversely, precipitation tends to increase with elevation, but is strongly influenced by regional weather patterns and oceanicity or continentality of the site. These changes determine vegetation belts, which are altitudinal sections characterized by a given vegetation linked to a given climate (Theurillat and Gusian, 2001).

To capture a broad band of conditions, an altitudinal transect consisting of three sites each representing an elevation belt was chosen. The climate data refers to the period 1960 to 1990 and is de-trended and replicated over 300 years. These three independent climates serve as control climate as well as basis for the paleo-records derived scenarios described below. The characteristics of the transect sites are described in Table 3.

Table 3: Climate characteristics of three altitudes representing the altitudinal transect studied in this paper. Means and standard deviations are calculated from 100 years de-trended monthly values.

	Mean annual temperature [$^{\circ}\text{C}$]	Mean annual precipitation [mm]
Submontane (550 m asl)	7.59 ± 0.53	1313 ± 249
Montane (950 m asl)	5.44 ± 0.67	1339 ± 243
Lower subalpine (1450 m asl)	2.69 ± 0.52	1550 ± 353

3.3.2 Range of Scenarios

The scenarios developed for the simulation experiment correspond to the reviewed literature presented in Section 3.1 and serve the data needs of the PICUS v1.3 model. Each scenario contains monthly averages of temperature, precipitation, radiation and vapor pressure deficit (VPD) for a 300 year simulation period.

Although there is a substantial body of research on the ecological and societal impacts of climate change, virtually all research has relied on scenarios with slow and gradual changes (Alley *et al.*, 2003). And since the instrumental record is becoming more valuable as it is lengthened, but is insufficient to have sampled the full range of climatic behavior, both paleo-data derived as well as model generated projections are

incorporated in this study, each approach serving as the basis to one storyline. For studies of THC-shutdown, Gagosian (2003) suggests contemplation of both an immediate collapse and one a century from now, taking into account prolonged GHG warming, which is realized in the two scenario families.

Storyline 1: Paleo-record derived scenarios

In this storyline the assumption is made that events being reported from the past can serve as a blueprint for possible changes in the future. Since causalities of such events as the Younger Dryas or 8.2 ka are still the center of hot discussions among paleoclimatologists and modelers, we do not intend to estimate the likelihood or cause but apply those former leapfrogs of the climate system on modern climate conditions. Since it was shown that abrupt climate changes were especially common when the climate system was being forced to change (Alley *et al.*, 2002), the urgency of testing the implications of such a catastrophe cannot be denied.

Although the THC has been identified as a major trigger of abrupt climate change, its weakening at a given rate of CO₂ increase, whether it may collapse or may eventually recover, is still clearly model dependent and thus uncertain (Tziperman, 2000). Furthermore, there are still differences in understanding and modeling climatic changes recorded in paleo-proxies. Broecker (1999) states that, to date, there is not a model that is up to the task. And, although there are reservations in applying past analogies on modern conditions, these records can at least serve as a worst case scenario (Broecker, 1999). According to Fuessel and Minnen (2001) the applied “incremental changes approach” — changes in a few climate variables applied uniformly across a study region — is useful for a general assessment of potential climate change impacts. In the following paragraphs the scenarios are characterized according to their major variables and changes. The literature-derived changes are applied on the climate baseline of three altitudinal zones (see Section 3.3.1), with linear modification during a transition period, resulting in the respective climate signal for the scenarios. What all scenarios in this storyline have in common is that they stay within the recent band of inter-annual variability after the succeeded rapid climate transition (i.e., the last two centuries of the simulation).

Temperature

Magnitudes of abrupt climate changes have been estimated as extreme as -16°C (Alley, 2001), but there is reason to believe that the most severe impacts concerned sites outside the study area — the high latitude North Atlantic region. And although we base this storyline of scenarios mainly on Greenland ice core records (GRIP, 1997), which serve the high resolution needed, the relation to the central European region is well established; Schwander *et al.* (2000) demonstrated on high-resolution stable-isotope records of Swiss lake sediments that past climatic changes caused an almost simultaneous signature for Greenland as well as the Central Alps. However, the results and interpretations in the literature are manifold, so we decided to cover a broad range of changes with a whole family of scenarios (Table 4 and Figure 18).

Precipitation

Although temperature is understood to be the key variable of climate change, precipitation is as important as temperature for ecosystems — forest growth is often limited by the amount of water available. But general predictions of precipitation

changes are hard to make since patterns in rainfall are strongly driven by sea and air currents as well as topographic elements. Besides the regional drivers, the general physical assumption of a warmer air bearing more moisture and vice versa holds true for the records of the Younger Dryas, where a major decline in precipitation is reported alongside cooling (Alley *et al.*, 2002). For the scenarios GRIP data was again used to establish a relationship between changes in temperature and snow accumulation serving as an indicator for precipitation. A linear model was used to explain changes in snow accumulation from changes in temperature over a period of 5,500 years, covering the Younger Dryas and the 8.2 ka event. This results in precipitation decreases of -24.0% , -36.2% and -48.5% for the respective temperature drops of -4°C , -6°C and -8°C (Figure 19).

Table 4: Temperature changes, length of transition period and acronyms of the respective scenarios in the paleo-data derived storyline. Departing from control climate, the monthly temperature values decrease linearly during the transition period having shifted the baseline for the respective amount at the end. This procedure is applied at all three altitudes of the study layout. The scenario names consist of a figure representing the full difference in temperature between baseline and the respective scenario. The last letter indicates the speed of transition: (r) rapid, (m) medium, (s) slow.

	Temperature Change			
	Years	-4°C	-6°C	-8°C
Transition Period	20	W4R	W6R	W8R
	50	W4M	W6M	W8M
	100	W4S	W6S	W8S

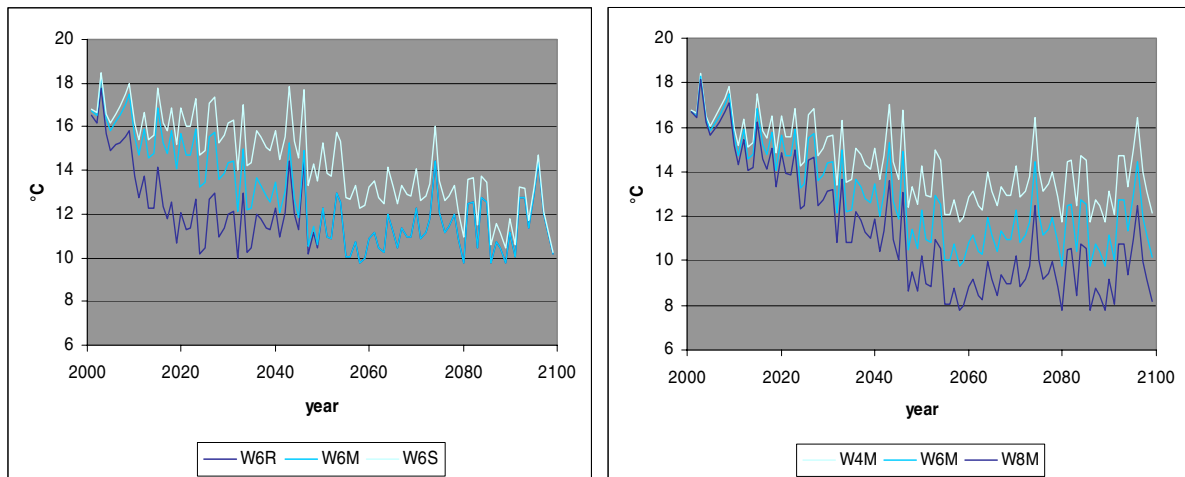


Figure 18: Abrupt change in July temperature according to the scenarios applied on one baseline climate file demonstrating the varying magnitudes and rapidities. Left: Magnitude of temperature change ranging from -4°C to -8°C . Right: Different transition periods being 20, 50 and 100 years.

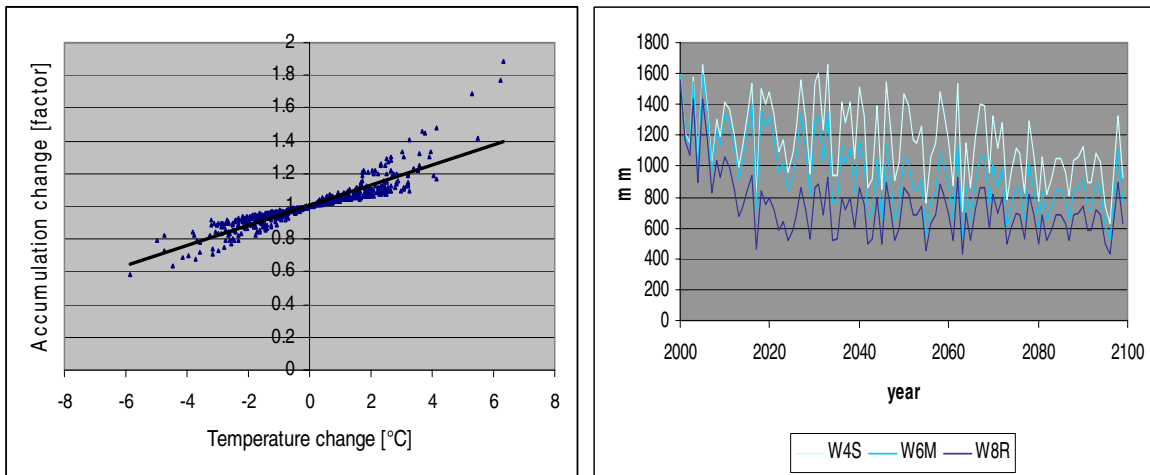


Figure 19: Left: Linear regression between temperature change and snow accumulation change derived from GRIP ($dP = 0.0614(dT) + 1.0063$, $R^2 = 0.824$). Right: Changes in annual precipitation according to the scenarios applied on one climate file to demonstrate magnitude and speed of decrease.

Vapor pressure deficit

A third major climate-induced input variable of the PICUS v1.3 model is vapor pressure deficit (VPD). Since no explicit information concerning the changes of air water content under past abrupt climate changes is available, the VPD is calculated assuming stable relative humidity. However, the changes in temperature resulted in a reduced VPD according to the cooling in the respective scenario.

Radiation

Although radiation is the fourth important climate variable used by PICUS v1.3, we did not modify the radiation budget of the sites studied. Bearing in mind that estimations of a changed radiation regime are hard to derive even under better investigated warming scenarios (atmospheric dust, cloud cover), yet have potentially high importance for plant growth, we applied an assumption of constancy.

The climate characteristics for the first storyline are summarized in Table 5.

Storyline 2: Climate model derived scenarios

Since the conditions that determine the climate of our planet are ever changing on all time scales, it is not possible to find a direct analogue for a future climate in the past. In model simulations the past is used in understanding mechanisms and putting them together in models (Rahmstorf, 1999). Despite the reservations concerning the completeness of our process understanding today (Broecker, 1999), models serve as powerful tools in a comprehensive climate change assessment. One particular advantage is that the state of the modern atmosphere, i.e., the raised and still rising CO₂ content, is taken into account.

Table 5: Mean annual temperature and mean annual precipitation of control climate and the scenarios after the rapid climate transition in the three elevation belts studied in this paper.

	Scenario	Temperature (°C)	Precipitation (mm)
Submontane	Control	7,6	1313
	w4r, w4m, w4s	3,6	998
	w6r, w6m, w6s	1,6	838
	w8r, w8m, w8s	-0,4	676
Montane	Control	5,4	1339
	w4r, w4m, w4s	1,4	1018
	w6r, w6m, w6s	-0,6	854
	w8r, w8m, w8s	-2,6	690
Lower Subalpine	Control	2,7	1550
	w4r, w4m, w4s	-1,3	1178
	w6r, w6m, w6s	-3,3	989
	w8r, w8m, w8s	-5,3	798

The data used in this storyline is provided by the Potsdam Institute of Climate Impact Research and is derived from a series of simulations with the CLIMBER-2 climate model of intermediate complexity (Petoukhov *et al.*, 2000). In model hierarchy, this model is placed between energy balance models and general circulation models (GCMs). It describes a large set of processes and feedbacks in the climate system containing modules for the atmosphere, ocean, sea ice, land surface processes, terrestrial vegetation cover and global carbon cycle and is therefore comparable to GCMs, but has much faster turnaround times due to low spatial resolution and simplified governing equations (Rahmstorf and Ganopolski, 1999). Inter-comparison of CLIMBER-2 results with GCMs show good agreement and extensive validation with paleoclimate simulations suggest the climate sensitivity to be plausible. A detailed description of the model is given in Petoukhov *et al.* (2000).

The two scenarios used are part of a long term global warming study carried out by Rahmstorf and Ganopolski (1999). The simulation starts at pre-industrial equilibrium climate in year 1800 and used observed CO₂ levels until the present. Then the forcing follows the IPCC (1995) scenario with the fastest increase in carbon dioxide — IS92e. However, the review in Figure 20 shows that this former most extreme forcing is now in the midst of the envelope of the recent scenario family of the IPCC (2001a).

In the CLIMBER-2 simulation, carbon dioxide concentration culminates in 2150 at 1200 ppm and declines from then onwards. The maximum global warming is reached several decades after the CO₂ concentration peak (Figure 21). The general assumption is that fossil fuel use will eventually cease as diminishing reserves, increased extraction costs and new energy technologies bring an end to the fossil fuel era.

The two scenarios of this storyline differ in freshwater forcing to the ocean, where the scenario labeled ECT is the standard scenario applying the freshwater response as in the original model and EHS is a scenario with enhanced freshwater forcing, taking into account increased melt-water runoff from the Greenland Ice Sheet.

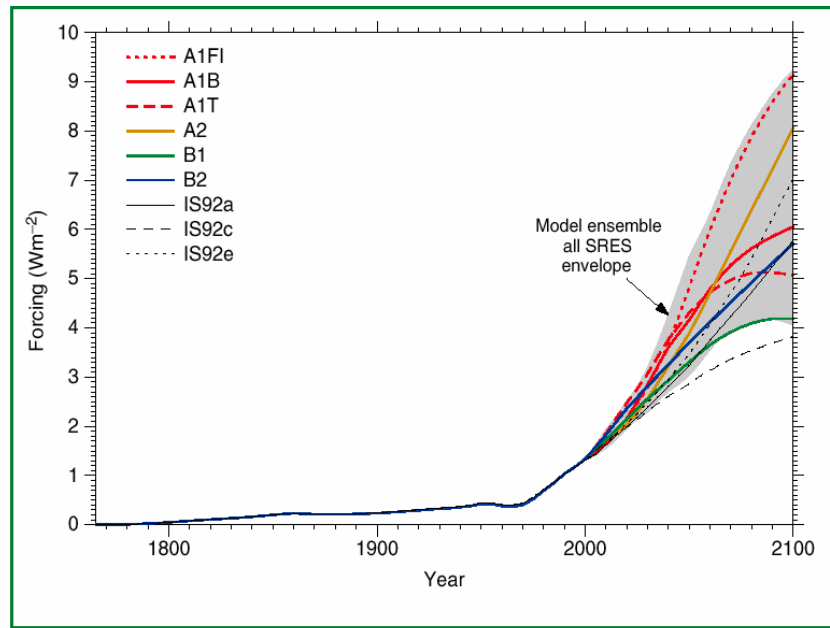


Figure 20: Radiative forcing for six recent (2001) IPCC scenarios (colored) and three from the second assessment of the panel in 1995 (black). Source: IPCC (2001a).

In the simulation experiment Rahmstorf and Ganopolski (1999) found a weakening of the North Atlantic Deep Water (NADW) formation during the first half of the 21st century in all scenarios (Figure 22). However, after 2100 the scenarios diverge, falling into two categories — those in which the NADW formation recovers and those in which it shuts down completely. This indicates the existence of a threshold of the THC as described in Section 3.3.

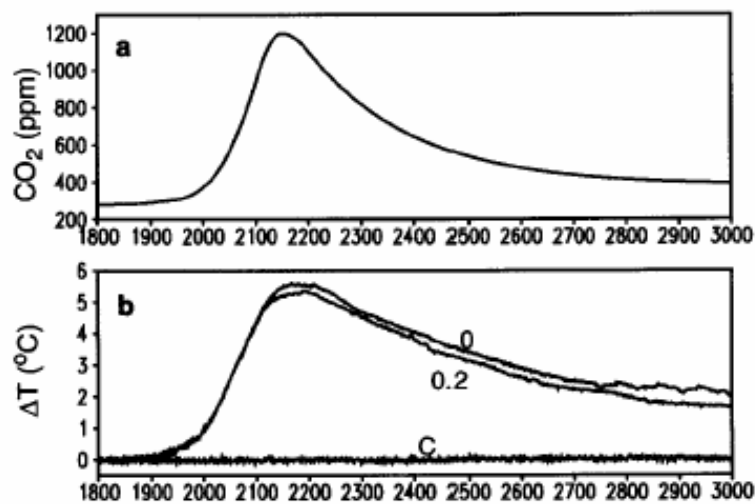


Figure 21: (a) Development of atmospheric CO₂, and (b) Global annual mean surface air temperature in the CLIMBER-2 simulation. The experiment labelled ‘C’ is a pre-industrial control run, ‘0’ refers to the scenario ECT and ‘0.2’ to EHS. Source: Rahmstorf and Ganopolski (1999).

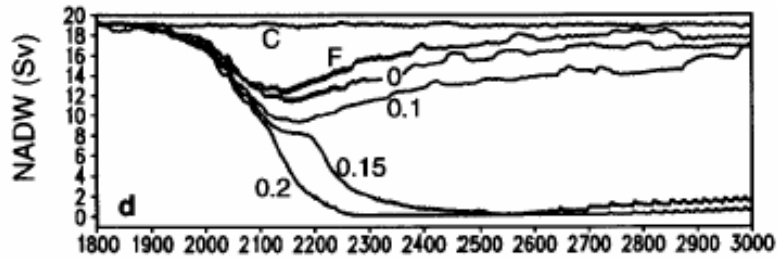


Figure 22: NADW circulation rates (Sverdrup) for the scenarios (0 = ECT, 0.2 = EHS) indicating threshold behavior of the THC. Whereas the THC recovers in scenario ECT, scenario EHS leads to a complete collapse of the thermohaline circulation. Source: Rahmstorf and Ganopolski (1999).

As seen in Figure 21, the global mean temperatures of the two scenarios remain within 0.5°C of each other, even after the ocean circulations have diverged. However, regional differences are large with maximum extent over the North Atlantic in high latitudes. At 55°N latitude the temperature rises by 3°C until 2100 but then drops by almost 3°C within 40 years as an effect of the rapid shutdown phase of the thermohaline circulation. Furthermore, the anticipated cooling does not stop there but continues until the end of the next millennium, leaving the North Atlantic over 6°C colder than at pre-industrial times (Figure 23).

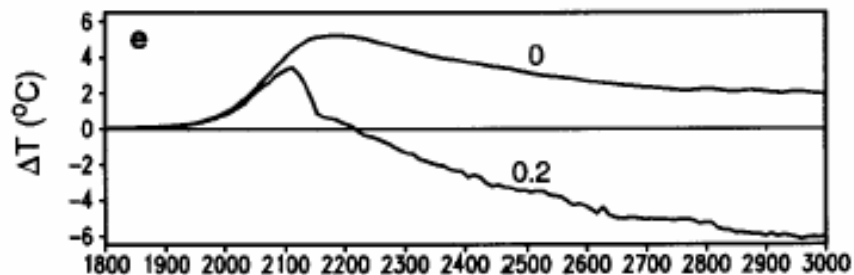


Figure 23: Winter surface air temperature change over the Atlantic at 55° N (0 = ECT, 0.2 = EHS) extended over a millennium. Source: Rahmstorf and Ganopolski (1999).

This study concentrates on a window in time and space, focusing on the Austrian Alps and development of the years 2000 to 2300. Therefore, three grid cells (spatial resolution half-degree grid) were chosen, which are loosely linked to three altitudinal zones of the climate baseline (Figure 24). To grant comparability to the control climate the CLIMBER-2 output is adapted in adjusting all relevant variables. The years 1960 to 1990 served as reference periods, which are the basis of the de-trended control climate. Adjustments are made on the basis of mean monthly values in the reference period.

Vapor pressure deficit was calculated using model-derived air vapor pressure and monthly mean temperature. The radiation regime was again kept the same as for the climate baseline.

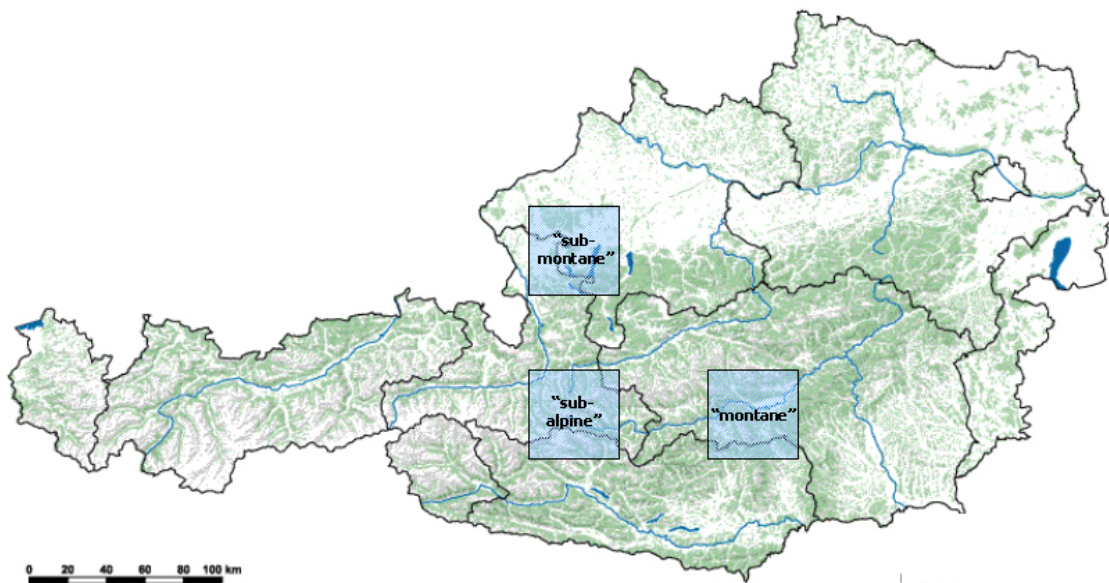


Figure 24: Location of the three grid cells of the CLIMBER-2 simulation chosen for this work. Altitudinal belts loosely associated with these sites are displayed.

Although the changes in this storyline are much smoother than those of the other scenario family, their investigation fits into the abrupt climate change theme since a cooling of almost 3°C within 150 years in the EHS scenario can already be classified as abrupt. Furthermore, beyond the year 2300 CLIMBER-2 suggests ongoing cooling (Figure 25) which would consequently lead to a temperature regime as anticipated in the first storyline.

3.3.3 Soil

Since soil plays a major role in forest ecosystems but is not explicitly modeled in PICUS, assumptions regarding nutrient availability under the investigated cooling are made.

PICUS requires the definition of soil properties in terms of water holding capacity (WHC, in mm), pH-value, carbon-nitrogen-ratio (C/N) and plant available nitrogen (kg/ha*a). As our study mainly aims for climate effects instead of soil response we parameterized “average” soils relying on the Austrian Forest Soil Survey (WBZI) (Killian, 1992). For alpine Norway spruce forests, two soil types are of major importance according to the survey, namely Dystric Cambisol (“*arme Braunerde*”) and Cambic Podzol (“*Semipodsol*”).

According to the relative frequency shown in Figure 26, we chose the characteristics of Dystric Cambisol for the submontane and montane belt and Cambic Podzol for subalpine simulations (Table 6).

Since especially nitrogen is understood to be a major driver of tree growth in forests, a cautious modification of the plant available nitrogen is applied in the simulations,

although literature reveals high uncertainty concerning change of soil carbon and nitrogen pools as well as nitrogen mineralization rate under changing climate.

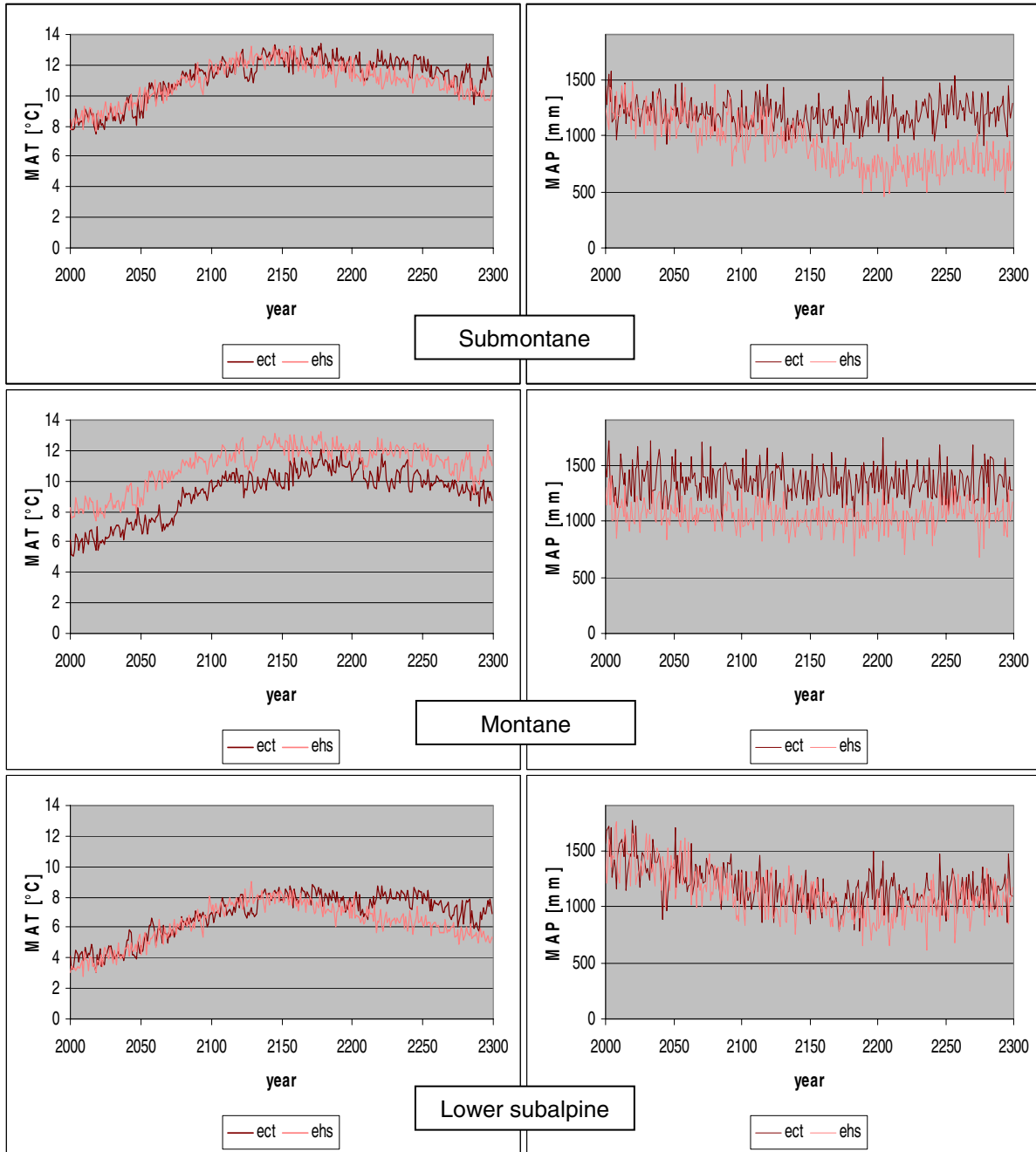


Figure 25: Mean annual temperature (left) and mean annual precipitation (right) derived from the CLIMBER model simulation. Top charts: Submontane site. Center charts: Montane site. Bottom charts: Lower subalpine site. The scenario with a shutdown of the THC (EHS) shows a more pronounced cooling in the second phase of the simulation.

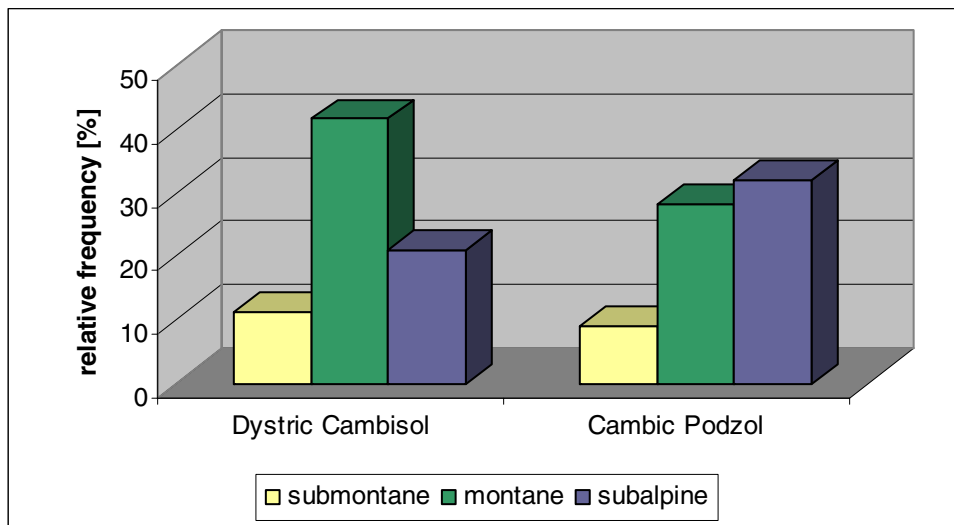


Figure 26: Percentage of soil types, Dystric Cambisol and Cambic Podzol, in different altitudes according to the Austrian Forest Soil Survey. Source: Kilian (1992).

Table 6: Soil specifications as found in the WBZI.

	<i>Dystric Cambisol</i>	<i>Cambic Podzol</i>
PH	4.0	3.9
C/N	20.5	22.4
Total nitrogen (0–30 cm including Litter) t/ha	5.33	5.98

First, climate factors such as temperature and precipitation are likely to have a direct impact on the soil organic matter content (IPCC, 2001b), where soil organic matter is likely to increase with decreasing temperature. But there is strong interference of this relationship with other factors, such as soil type and moisture or vegetation cover. However, to quantify these changes we carried out a spatial analysis reviewing changes in pools along a geographical gradient.

Our findings in reviewing carbon densities in Russian soils along a temperature gradient show an increase in soil carbon with cooler mean temperatures (Figure 27). To get reasonable results, peat influenced soils were excluded in the regression and the data was stratified according to vegetation types. To apply this relationship for our study, we assumed constancy in the C/N-ratio, backed by Figure 28 displaying no significant trend for C/N-ratios of relevant Holdridge life zones.

Another crucial yet uncertain point is the translation of the total nitrogen pool into plant available nitrogen. There is high disagreement among soil scientists on the behavior of soil carbon and nitrogen with regard to temperature changes. Where some studies do not find a relationship between turnover rates and temperature at all (Giardina and Ryan, 2000), others indicate a more or less strong decline of organic matter decomposition with decreasing temperature (Kaetterer *et al.*, 1998; Knoepp and Swank, 2002). Bearing this in mind, we reviewed several relations between temperature and mineralization

rate, derived from both laboratory measurements (Knoepp and Swank, 2002; Francisca et al., 2002) and field observations (Xu and Qi, 2001). After testing several functions implemented in soil models (see Keryn, 2001; Kaetterer *et al.*, 1998) with the cornerstones of Austrian forest soils as given in the WBZI we decided on an exponential relation. We used the decomposition function of the widely applied CENTURY model as found in Keryn (2001; see also <http://www.nrel.colostate.edu/PROGRAMS/MODELING/CENTURY/CENTURY.html>). In comparison to mean mineralization rate of 2% named in the WBZI, the value of 1.75 (for 8°C) is an appropriate match; the underlying assumption for maximal mineralization is 8% (Evers, 1967; Jandl, 1992; both in Englisch, 1992).

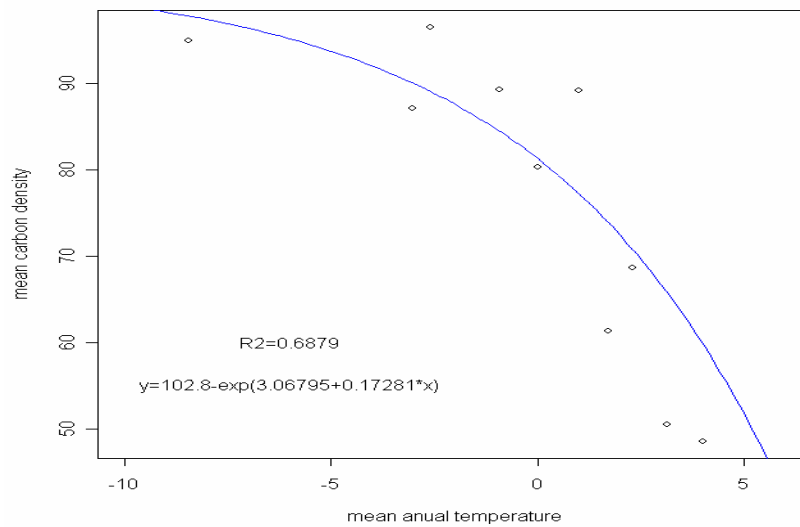


Figure 27: Mean carbon density (t/ha) in Russian forest soils (0–30 cm without litter) over mean annual temperature. Data provided Stolbovoi (2002).

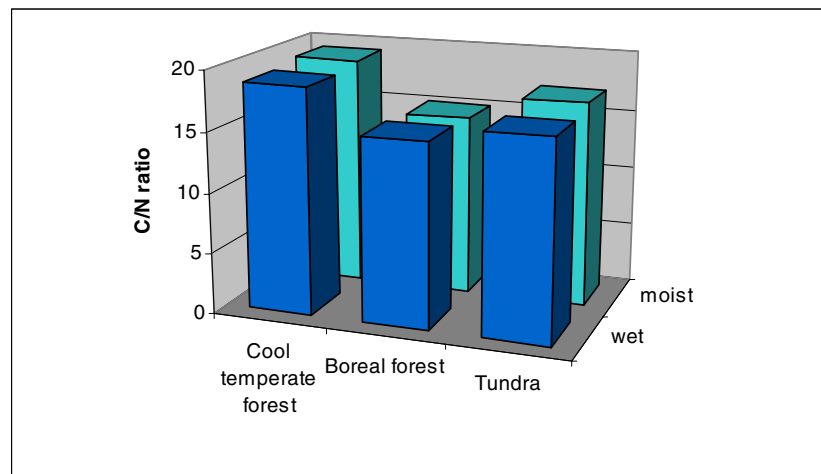


Figure 28: C/N-ratios for different Holdridge life zones showing no significant trend with temperature. Source: www.biology.ualberta.ca/courses/biol366/uploads/w03/public/lectures/Biogeochemical_and_Nitrogen_Cycles.pdf.

The application of the presented ideas concerning soils leads to the following picture for our scenarios: The effects of the increasing nutrient pools and the decreasing decomposition rate (Figure 29) lead to an almost equal amount of nitrogen available to the plants over the course of the simulation in the scenario family anticipating the immediate onset of cooling. The two climate model derived climate scenarios (EHS and ECT) are, in turn, characterized by a significant temperature increase through the first decades, which is directly translated to soil processes in our approach. At this higher temperature level the increased mineralization rate outperforms the effect of lowered nutrient stocks, which leads to a pronounced increase in plant available nitrogen for the first half of the simulation (Figure 30). The findings of Giardina and Ryan (2000) question the correlation of temperature and decomposition in general and highlight the uncertainties in this field — we, however, applied widely accepted assumptions to take into account certain soil processes in our study. Rapid climate change can have potentially severe implications on soils, which needs further investigation of the soil science community.

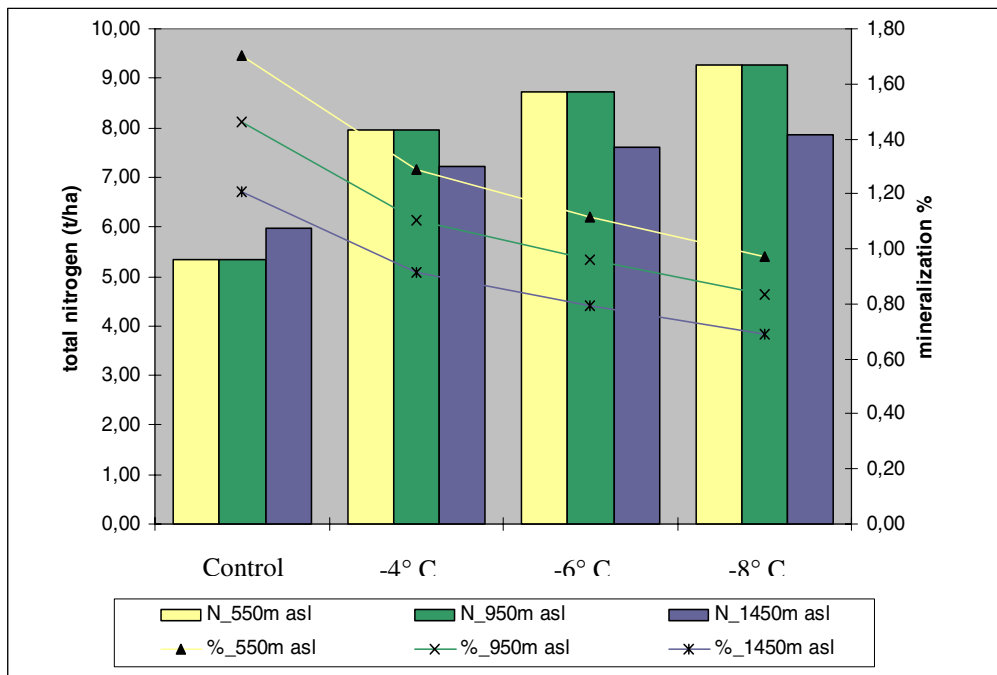


Figure 29: Nitrogen pools (bars) and mineralization rates (lines) and their anticipated change in the respective scenarios. Initial conditions (control) as well as maximal mineralization rate derived from WBZI. Source: Killian (1992).

3.4 Methods

3.4.1 Stand level

The general layout of the study is given by the climate scenarios as described above. To better cover the variety of soil conditions, each of the scenarios was investigated on poor, medium and good soil water holding properties. The simulations were conducted over 300 years, starting at the present day (e.g., 2000 in the simulation) with the same initial conditions for every scenario (9 year old pure Norway spruce stand with 2300

individuals per hectare). An area of 2500 m² was simulated, providing ample reproduction of ecosystem processes in a mono-species experiment as well as model time efficiency. The study focuses on Norway spruce, which is the dominant tree species in Austria and the eastern Alps (see Figure 31). Over the course of the simulation the regeneration submodel of PICUS provides new Norway spruce individuals on the simulated patches.

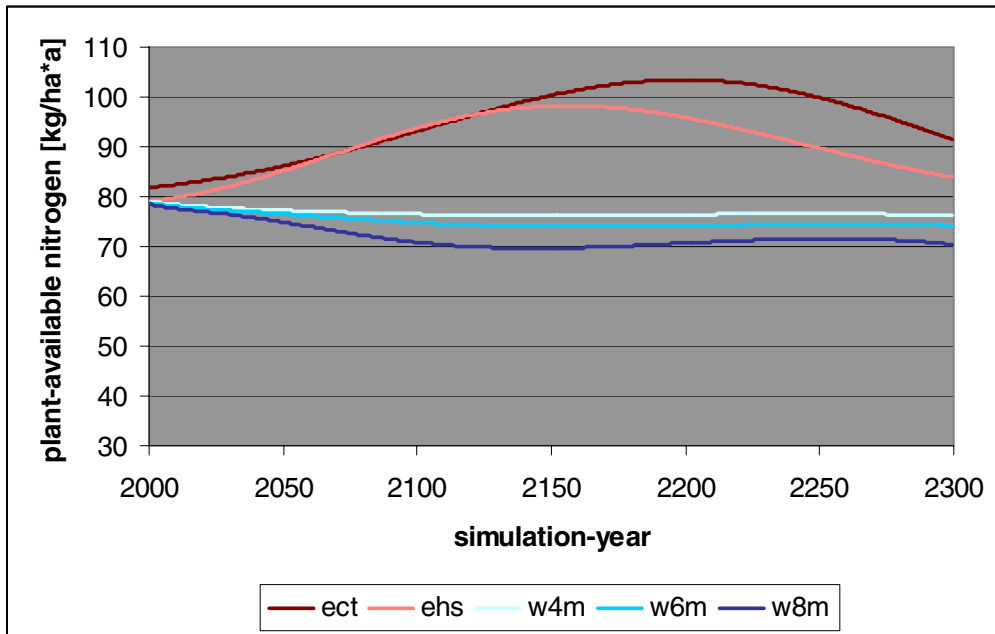


Figure 30: Response of plant available nitrogen to the scenarios of abrupt climatic change over the simulated period (montane site).

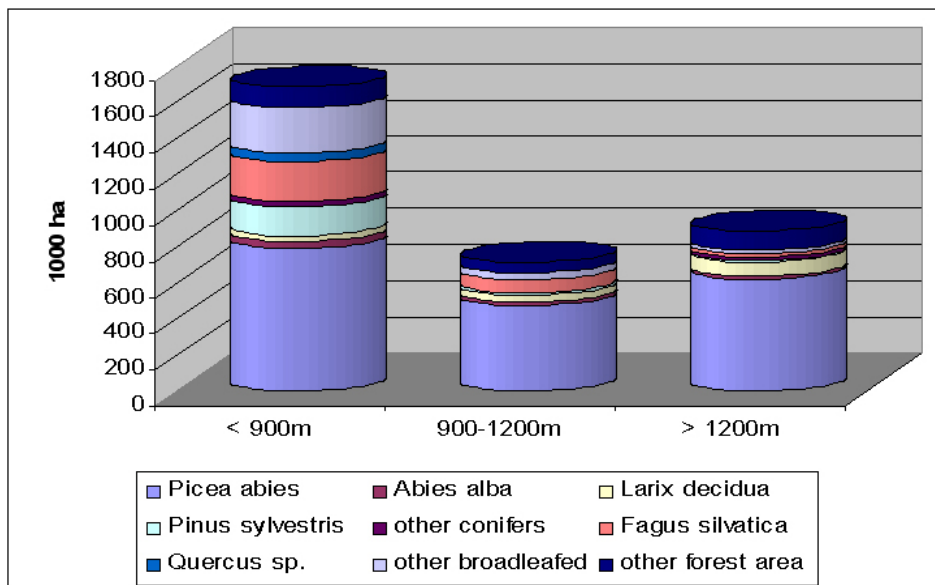


Figure 31: Share of tree species according to the Austrian Forest Inventory 1992/96 in three altitudinal belts. Source: Schieler (1997).

The 300 years of simulation are divided into three 100-year periods to better compare scenarios to baseline and take age dependent growth trends into account. Whereas the first 100 years are dominated by more or less pronounced growth, the following two 100-year periods show biomass stock patterns overlaid by aging induced mortality. Although a “no management” experiment was conducted, the simulated stand does not approach a “real” equilibrium, because of the restriction to one species. The comparison of this climate altered “spruce equilibrium” to the undisturbed control simulation is the major source of the interpretations presented below. The comparability of the two scenario storylines is restricted by differences in the altitudinal specifications — although a loose link between the altitudes of the transect and the three grid cells of the climate model does exist. Due to this fact they are treated separately in the interpretation, but inter-comparisons are made to the extent possible.

3.4.2 Aggregation

The UNFCCC (1992) defines dangerous interference with the climate system by inter alia the condition that “greenhouse gas stabilization should be achieved within a time-frame sufficient to allow ecosystems to adapt naturally to climate change”. Abrupt climate changes, as investigated in this study, have a high potential for exceeding the ecosystems’ adaptive capacity through magnitude and rapidity of change. Our findings at stand level are thus linearly upscaled to the ecosystem level, to tentatively quantify possible adverse effects. Although this approach is limited especially by the small sample size of three transect plots, the study layout aims at reaching the highest possible spatial interpretability in, for instance, choosing “average” soil characteristics of the AFSS. In addition the effect of varying soil water holding capacities is taken into account and the most abundant alpine tree species is studied.

As result of this upscaling, we follow Fuessel and Minnen (2001), using the percentage of an area where the current biome is no longer viable as the main indicator to quantify the impacts of abrupt climate change. The resolution of this assessment is restricted by the AFI providing detailed data on the administration-district level. Also, due to data availability, only forest in yield (“*Ertragswald*”) is considered here and since Norway spruce is an important species of protection forests out of yield mostly situated at high altitudes the presented figures are understood as cautious estimates.

Since there is growing concern that human alteration of the global carbon cycle could act as trigger of abrupt climate change, possible forest related feedbacks are investigated. This assessment is, however, narrowed to the paleo-data derived scenarios because including the second storyline would lead to several shortcomings that can not be overcome within the frame of this work, for instance, whether drought tolerant species would be able to fill the gap at sites too dry for Norway spruce — but, the full species set and species interactions are not considered in this paper. Another shortcoming of the study in this context is that the effect of management in general and adaptation strategies in particular are not considered, which have, however, substantial potential to mitigate adverse effects and are to be developed as a successive step with regard to the severe impacts presented.

The carbon cycle is central to the Earth’s system, being coupled with climate, the water cycle, nutrient cycles and the production of biomass by photosynthesis on land and in the oceans (GCP, 2003). Therefore, the carbon cycle is subject to several international scientific research programs (Falkowski *et al.*, 2000) and substantial work is done to quantify global and local carbon pools and fluxes. As a result, terrestrial ecosystems and namely forests are identified as both major carbon storage and possible sink of atmospheric carbon dioxide. The IPCC assessment for biological mitigation on the global scale is in the order of 100 Gt C (cumulative) by 2050, equivalent to about 10% to 20% of projected fossil fuel emissions during that period (IPCC, 2001c). Furthermore, the importance of forests for the mitigation of fossil fuel emissions is evident in statements, such as the US government planning to meet half of its annual commitment to GHG reduction through land-based carbon sinks (Noss, 2001). Also for Austria, forests can be identified as the major carbon pool, storing the highest absolute amount of carbon in both soils and aboveground biomass (Figure 32). However, threshold events such as abrupt climate change interfere with these projections having the potential to shift the system to another level of operation.

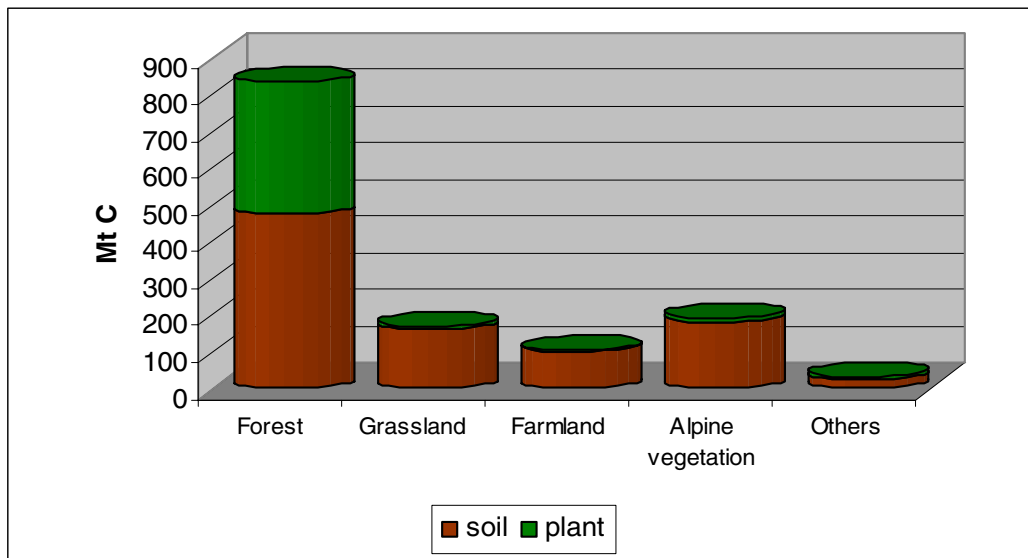


Figure 32: Austria’s carbon stocks as identified by Körner *et al.* (1993).

The assessment regarding changes in the carbon budget is not understood as a comprehensive multi-sectoral carbon cycle study. Rather, it is a first, crude estimation demonstrating the magnitude and severity of carbon feedback through changed forest ecosystems. This is done in conducting a stock change comparison taking the forest sector as quantified by Körner *et al.* (1993) as the base line. The estimated abrupt climate change interfered aboveground forest carbon pool is corrected with both the area where forests can no longer be sustained and changed biomass stocks on the remaining forest area.

This was done in splitting the forest carbon pool into three different altitudes, related to the studied elevation transect, corrected with the respective share of the total stock in the altitudes as reported by the AFI 82/86 (Schieler *et al.*, 1995). Consequently, forest

soils are considered in terms of change of forested area and temperature induced reduced decomposition. In the case of cooling, it is assumed that lost forest area is exposed to soil erosion and a certain percentage is lost to rocks (in accordance with Karl and Danz (1972) we assumed a share of 30% of the deforested area to be affected by substantial erosion and converted into rocky landscape). Other deforested areas are supposed to change to alpine meadows and sparse alpine grasslands — mean soil and plant carbon densities of both are taken from Körner *et al.* (1993).

4 Results

4.1 Stand Level

At stand level, this study focuses on two indicators, namely total biomass (t/ha) and periodical current annual increment ($\text{m}^3/\text{ha}\cdot\text{a}$) derived from total yield. Although they are closely correlated by their very nature both are of interest since abrupt forest response is often first tracked in just one of both (cf. tree mortality, regeneration).

The review in Figure 33 displays that the anticipated cooling in the first storyline leads to a reduction in aboveground biomass stock in all scenarios compared to the control simulation. The magnitude of this reaction is different in the three investigated altitudes: Whereas the submontane ecosystem is able to sustain Norway spruce forests even under an 8° cooler and significantly drier climate, subalpine spruce forests dies back within the first 100 years even with a cooling of 4°. Although the montane stand endures considerable cooler conditions, it is extinct in the most extreme scenario investigated. In the sustained stands, the reduction in biomass stock is substantial, varying between -11% to -28% for the submontane and -8% to -45% for the montane stands. The analysis of different soil water holding capacities led to similar percentages of loss although on another absolute level and so did varying rapidity of the climatic change. For all simulations in this storyline the absolute and relative values of biomass loss stay more or less stable after the transition to a cooler climate is completed.

The response to the CLIMBER derived scenarios does not show such a stable phase, since climate is shifting over the whole course of the simulation. Figure 34 reveals that storyline two also causes major loss in aboveground biomass, but basically related to the warming in first 200 years of the scenario. The decline reaches its maximum at the peak of the warming phase around year 2150 but then a decrease in atmospheric CO₂ and THC shutdown induced cooling helps Norway spruce to recover. Especially the low altitudes suffer under the warmer and drier conditions but biomass losses are substantial in all three elevations. Both climate model scenarios differ only slightly in the first time period, whereas EHS shows a significant advantage at high elevations at the end of the simulation because of the more pronounced cooling compensating the CO₂ induced warming in this scenario. A closer analysis of this storyline leads to the finding that the water regime is crucial to the survival of Norway spruce in a warmer climate. Therefore, the investigation of different physiological soil water properties is of interest.

Whereas different soil water holding capacities lead to different total biomass levels, the percentage of biomass loss as well as the time of dieback varies just slightly for the first storyline (Figure 35). For the CLIMBER derived scenarios the soil water holding

properties decide over life and death — stands on physiologically good soils are less affected than those on poor or mediocre water balanced soils (Figure 36).

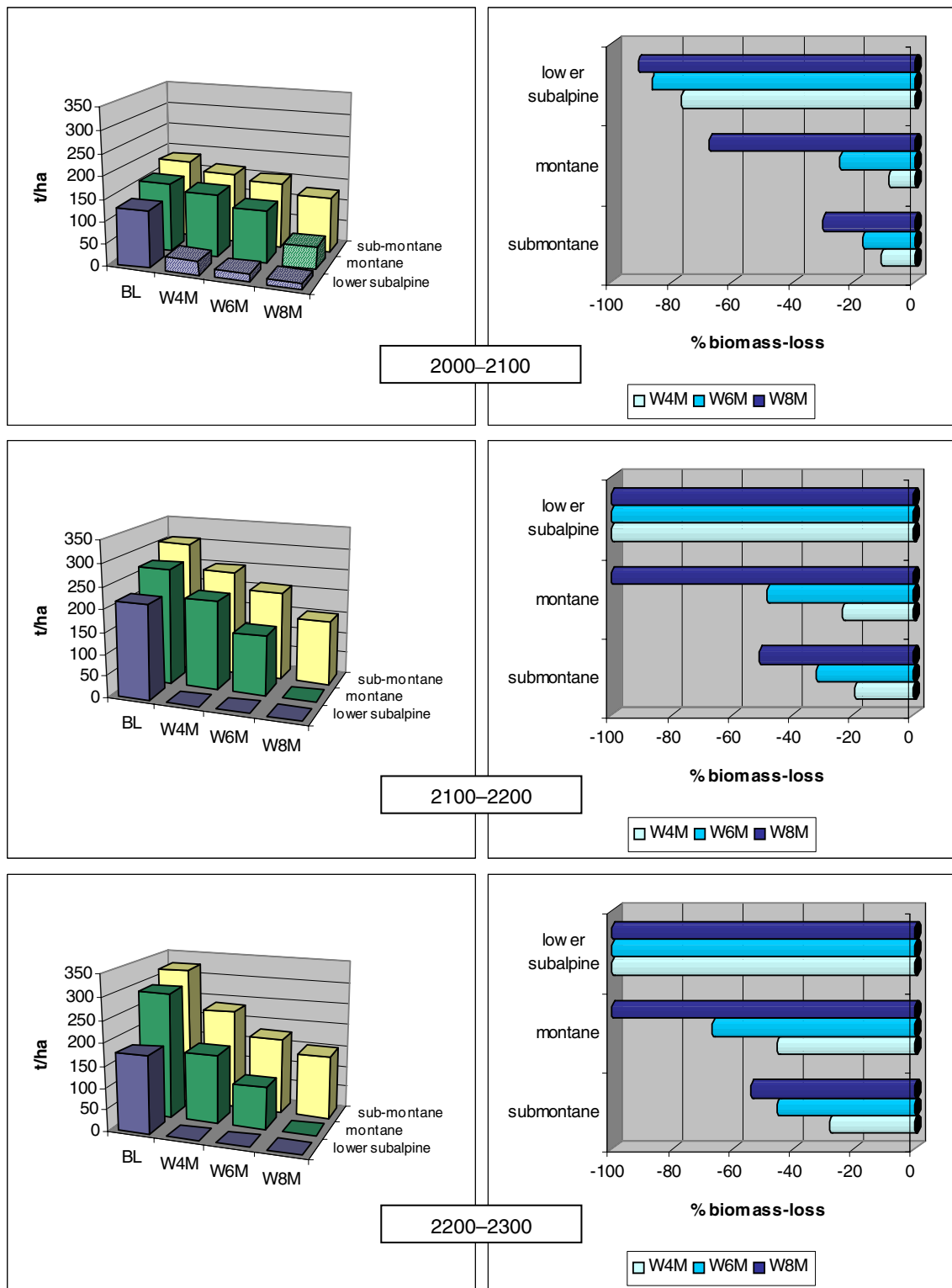


Figure 33: Absolute and relative to baseline (BL) mean stock biomass changes in the three periods: Top charts 2000–2100; Center charts: 2100–2200; Bottom charts: 2200–2300. Hatched stands die during the respective period.

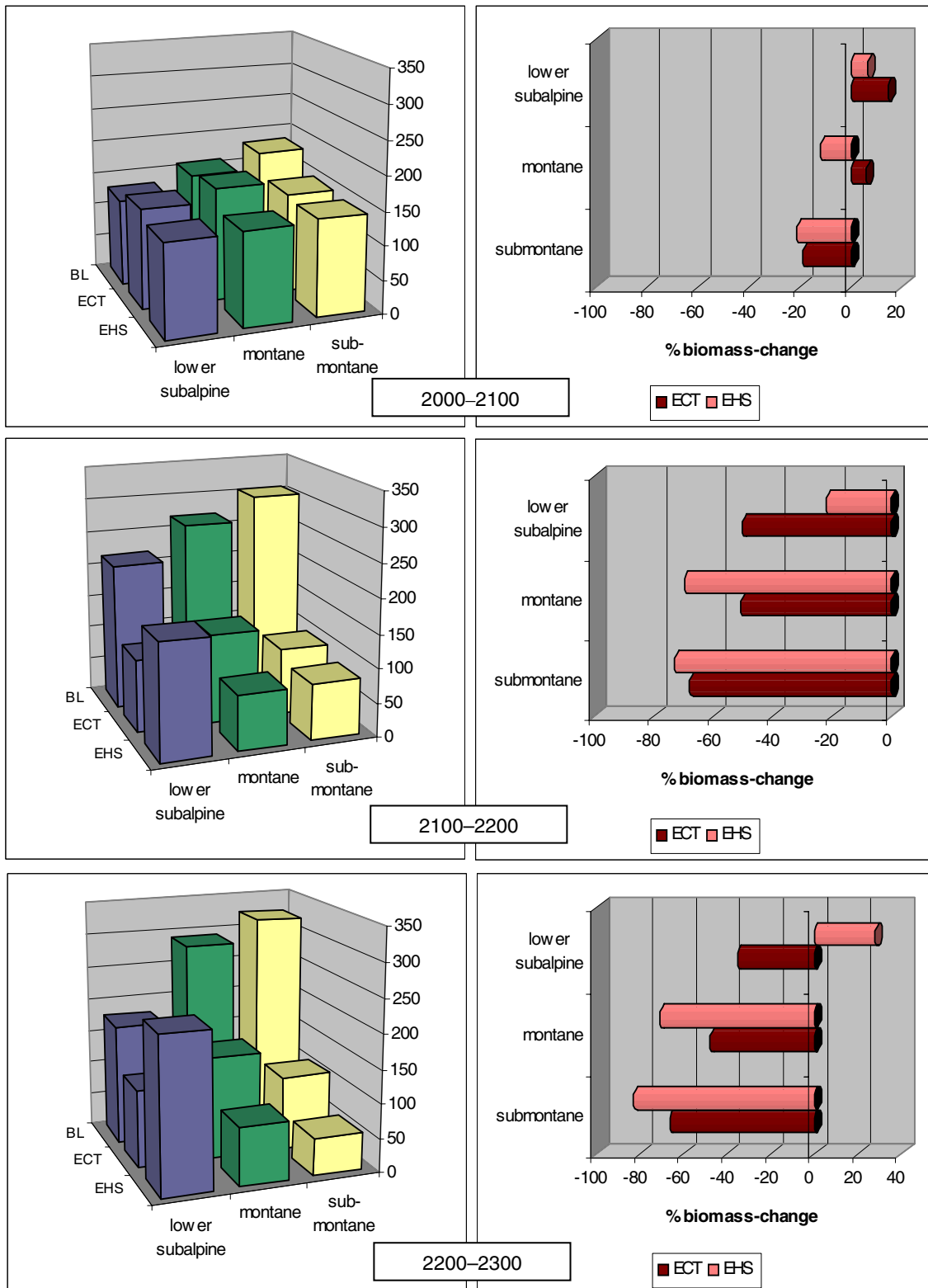


Figure 34: Absolute and relative to baseline (BL) mean stock-biomass changes in the three periods; Top charts 2000-2100; Center charts: 2100-2200; Bottom charts: 2200-2300.

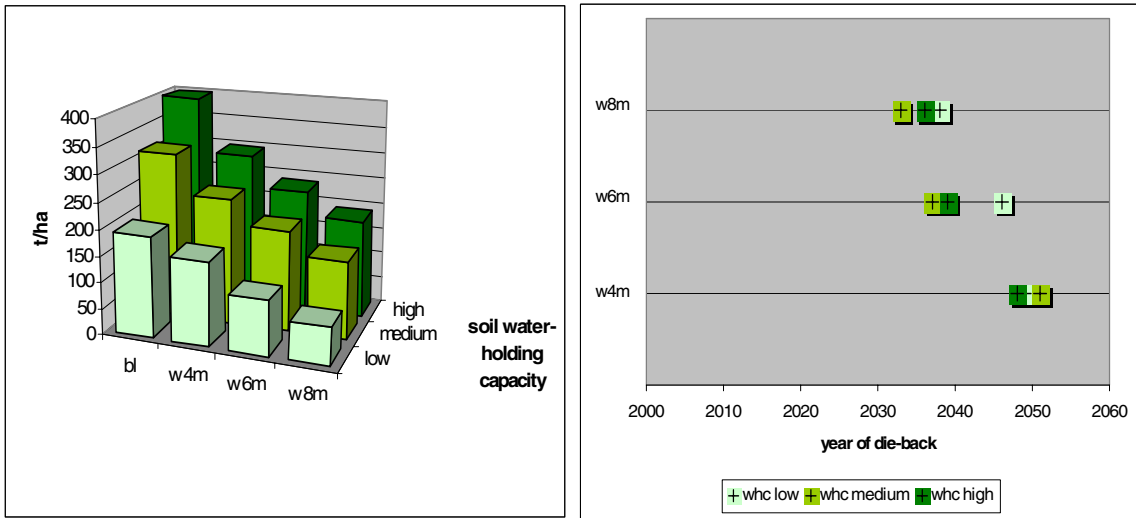


Figure 35: Left: Mean biomass stock for the period 2100–2299 decreasing with soil water holding capacity (submontane site). Right: Time of death for scenarios with immediate cooling (subalpine site); water holding capacity (WHC) has no significant influence.

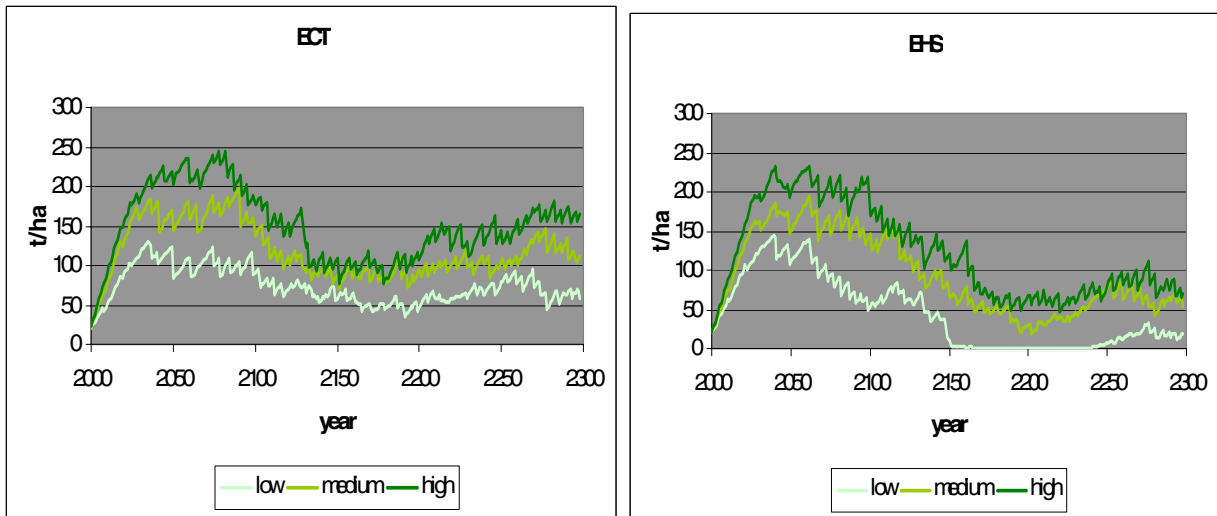


Figure 36: Total biomass at different soil water holding capacities in the climate model derived scenarios. Biomass level and survival are clearly influenced by soil water properties (submontane site).

As for increment, a response similar to biomass stock is found. However, slight divergences can be tracked; especially for the montane and submontane sites under the paleo-data derived scenario, increment corresponds faster to the cooling than biomass stock. Whereas the mid and lower elevation forests are first able to sustain their stock at a relatively higher level compared to productivity losses, these differences are evened out in the last period where losses in stock and productivity are almost equal. In the climate model, derived scenarios productivity is higher than the control for almost all

runs in the first 100 years, reflecting better growing conditions due to higher temperatures and raised plant available nitrogen. However, the second century reflects the heat and drought induced stress, but relative losses in increment are smaller than those in biomass. And whereas biomass levels in the third period are mostly still below the control simulation, productivity responds immediately to the slight cooling.

Furthermore delays in response and implications of different cooling rates were studied for the paleoclimate derived scenarios and their different rates of change. Generally, the experiment showed an immediate response to the changing climate in all simulations. The lower subalpine stand is highly sensitive to both the magnitude and the speed of cooling: Both minus six and minus eight degrees cooler climates lead to an early dieback, even before the climate transition is completed. In contrast, a drop in temperature of 4°C seems to push the ecosystem just slightly over the edge — the dieback follows the rate of cooling and is even delayed for the most rapid scenario. A similar pattern can be tracked at the montane stand for the strongest cooling-scenarios. Although the dieback does not limp behind the cooling it does not occur until the full magnitude of change is reached or even slightly later (Figure 37).

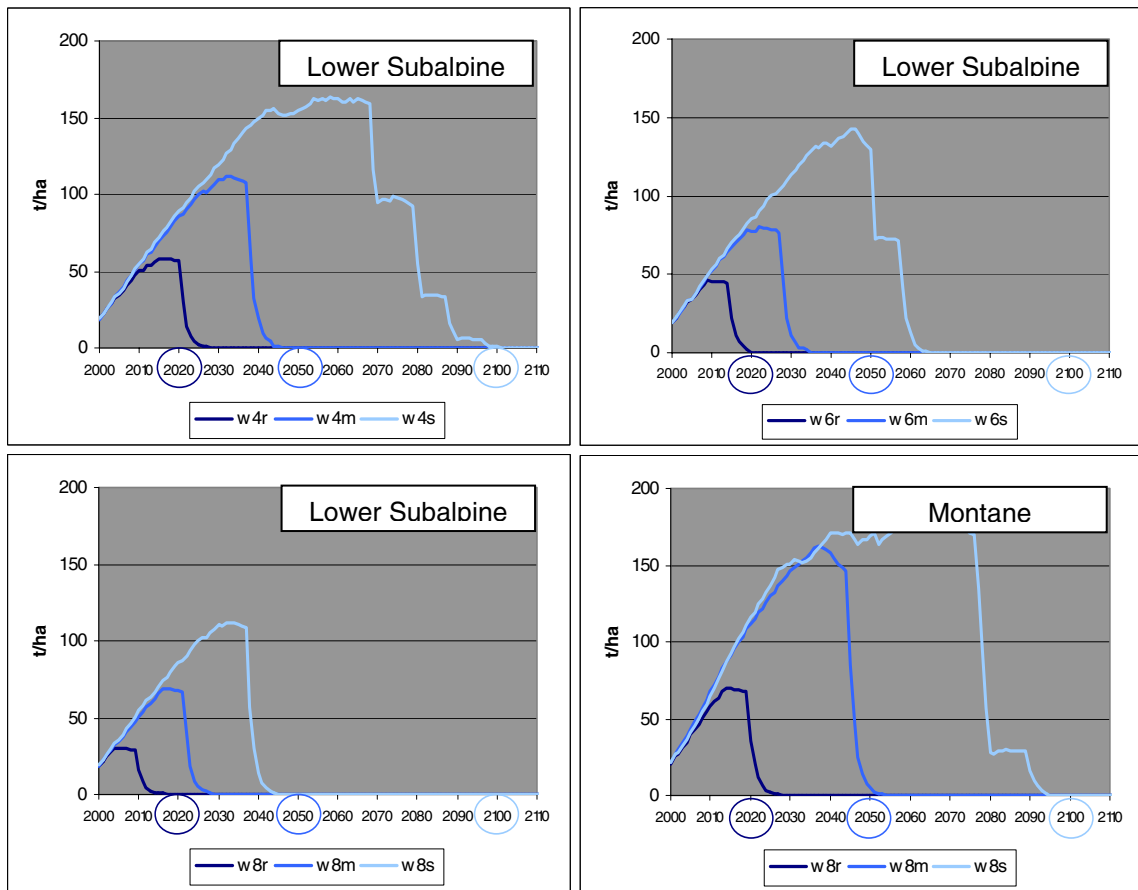


Figure 37: Total aboveground biomass over the simulation year is displayed to identify time lags in the response to climate change. The generally rapid response (end of transition is indicated by colored markings of the respective year) is slower, where ecotonal borderlines are only slightly exceeded (Lower subalpine: cooling -4°C , montane: cooling -8°C).

4.2 Thresholds

Moving from stand level to ecosystem level the results are condensed to isoline charts displaying the aboveground biomass relative to base line simulation (Figures 38 and 39). This method allows the identification of the critical thresholds for the maintenance of spruce forest ecosystems under the investigated abrupt climate change scenarios. For the first storyline the “steady” state after the transition is displayed, whereas for the CLIMBER derived scenarios the whole course of the simulation is covered because of the changing scenario characteristics.

It can be shown that the scenarios derived from paleo-records lead to major loss in aboveground biomass with little difference at varying soil water holding capacities. Interestingly, we find that physiologic dryer sites, where overall growth is generally lower, lose less in comparison to medium and high soil water holding capacities. However, all sites above 1,000 m asl are particularly vulnerable, suffering biomass loss exceeding 50%, even at a cooling of four degrees. Low latitudes are less affected by moderate cooling but also show considerable losses at extreme cooling of minus six to eight degrees Celsius. The threshold for cold induced dieback of Norway spruce forests is found to be somewhere around -4°C for sites above 1,400 m asl and critical altitudes drop significantly with stronger cooling. The lower subalpine stand suffered dieback in all cooling scenarios, whereas the montane stand responded only to an extreme cooling of -8°C with death.

The second storyline of climate change scenarios investigated spruce forests show a different response. For the first time period the high altitudes profit from the CO_2 induced warming and react with higher growth and biomass stock (up to +17%). In turn, Norway spruce stands in low altitudes suffer from hotter and drier conditions and show slightly reduced aboveground biomass stock (max 25% loss). Under the ECT scenario biomass stocks in Norway spruce forests are reduced at all altitude because of drought-stress, and this effect is sustained until the end of the simulation although CO_2 levels go down after the year 2150. As one could expect with regard to the scenario description, we find the biggest impacts in low altitudes on stands with low water holding capacity. The second scenario (EHS), in which climate is further perturbed by a shutdown of the termohaline circulation shows an immediate response of forest ecosystems to this alteration. Although losses in the lower altitudes are even bigger than for the ECT scenarios, which is due to lower growing season precipitation, the THC shutdown induced cooling in the third century of the simulation relieves the high altitudes and leads to increased forest growth above 1,300 m asl (given that soil-water properties are at least mediocre).

4.3 Ecosystem Loss

To assess the spatial significance of the stand level findings, results are related to the distribution of Norway spruce in Austria with respect to topography. As displayed in Figure 40, the adverse effects of paleo-records related rapid climate change concentrates clearly on the alpine districts. Even for a cooling of four degrees the total amount of Norway spruce forest ecosystem lost is estimated to be 450,100 ha. This value rises for -6°C and -8°C to 618,000 ha and 1,084,400 ha, respectively. Thus, at the most extreme scenario, more than 1 million ha forest ecosystem is lost, equaling 58% of all Norway spruce forests or one third of Austria’s total forest area.

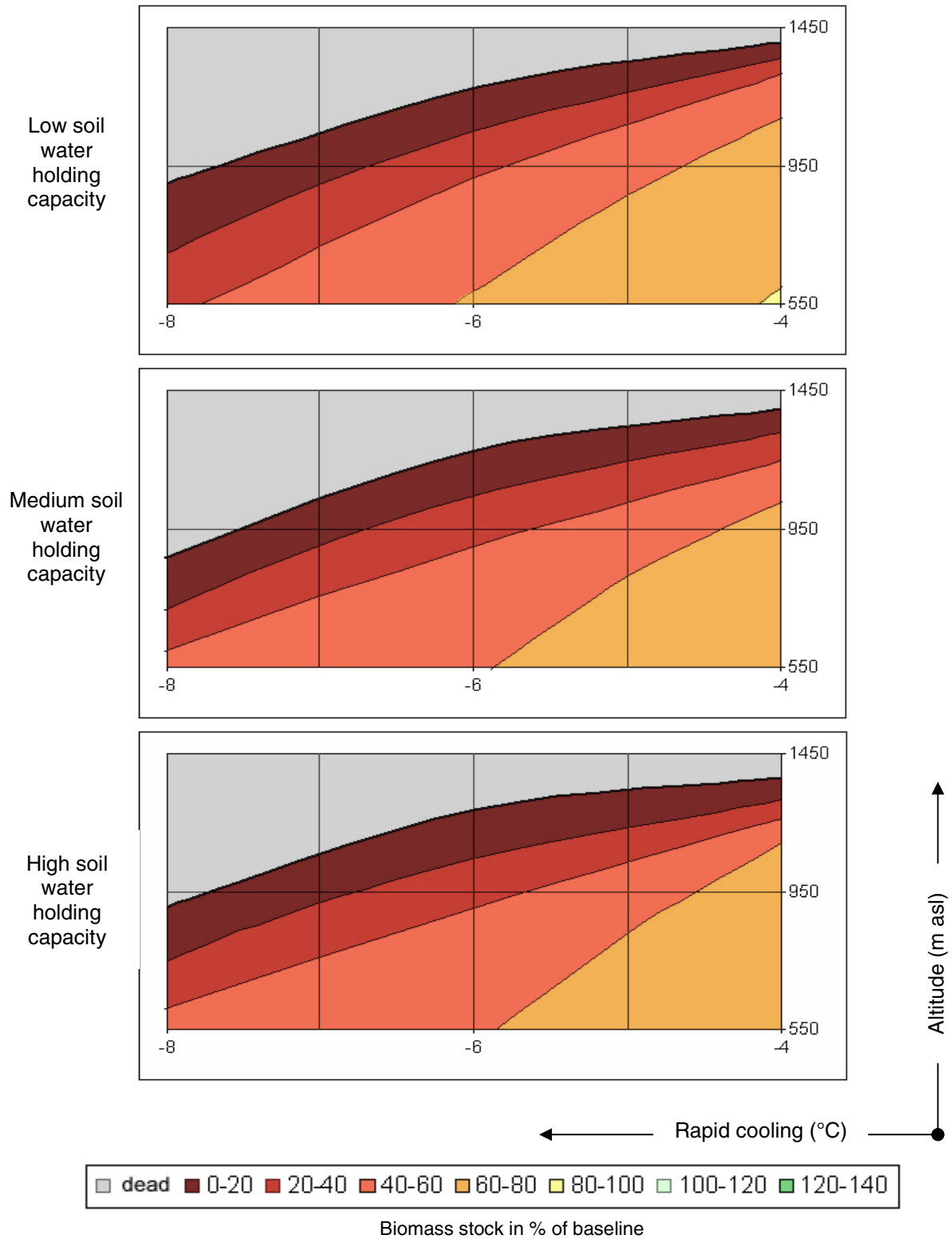


Figure 38: Biomass levels relative to control simulation and ecosystem thresholds beyond which dieback was found for the paleo-data derived scenarios displayed as isolines over respective temperature drops.

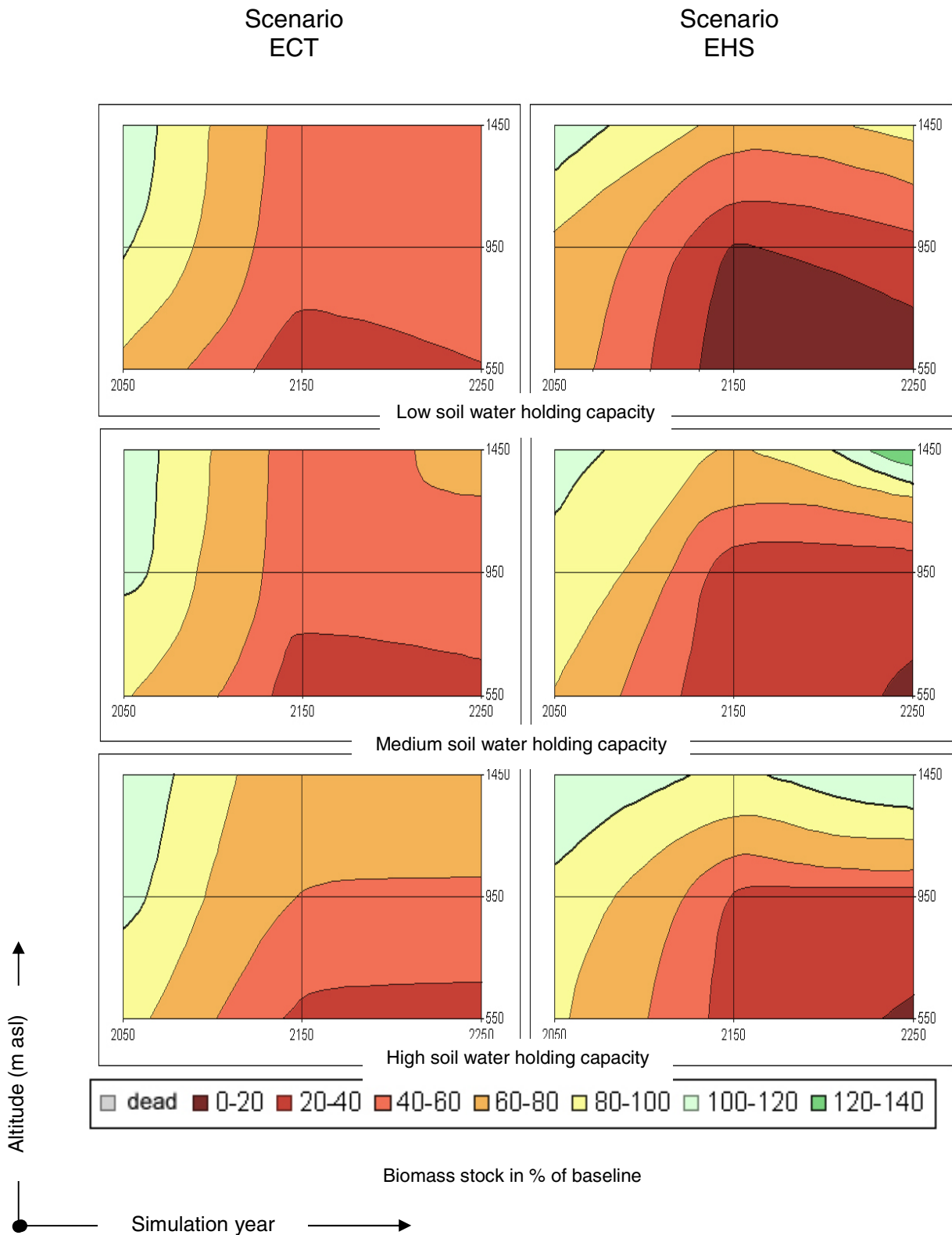


Figure 39: Biomass levels relative to control simulation for the climate model derived scenario -family displayed as isolines over the course of the simulation. Left charts: ECT. Right charts: EHS.

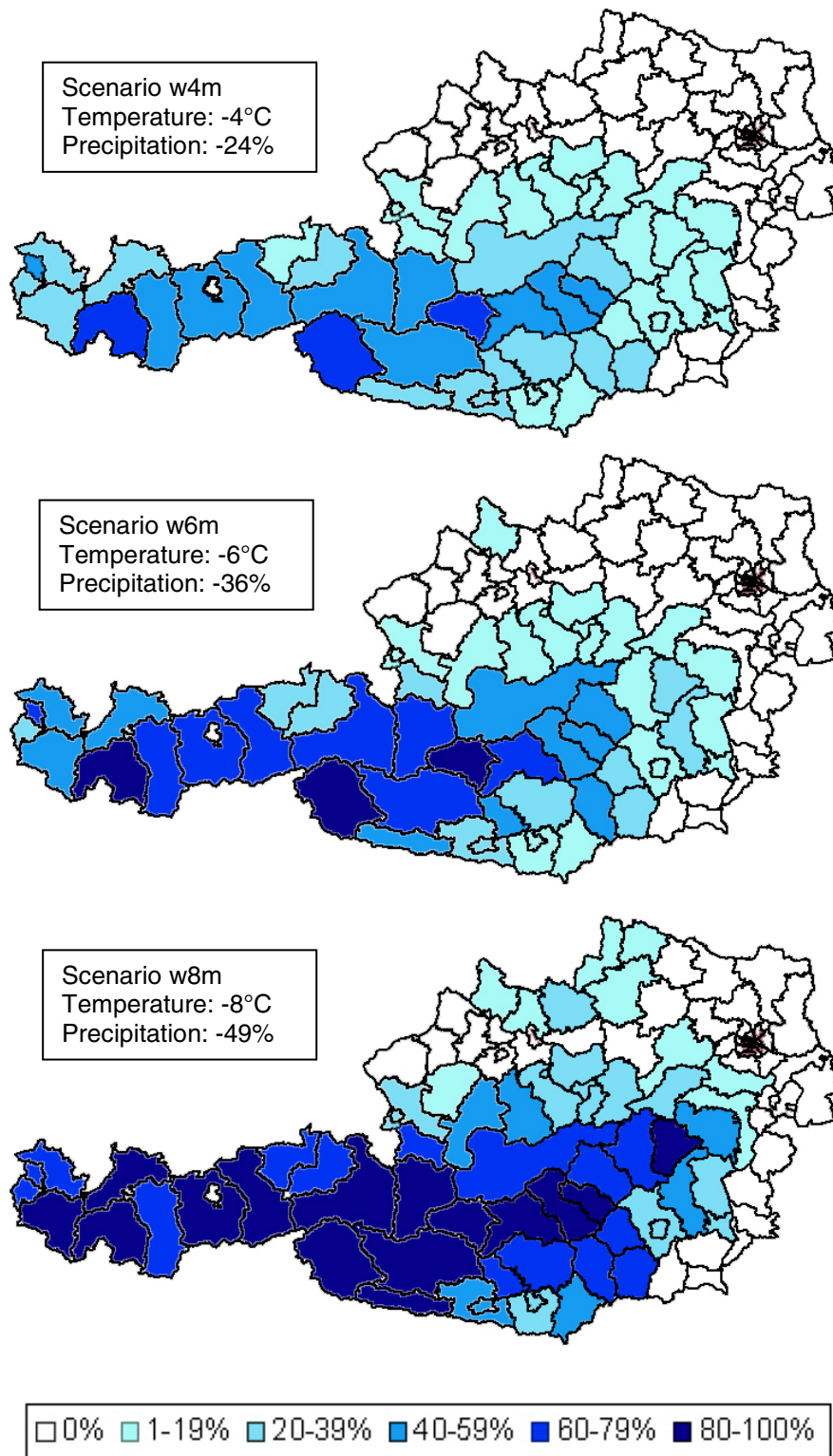


Figure 40: Loss of Norway spruce forest ecosystems in percent of the total Norway spruce covered forest area under paleo-data derived rapid climate change scenarios (Resolution: administration districts, exploitable forest).

For the second scenario storyline a similar quantification is difficult to obtain: Since soil water holding capacity is found to be a crucial element under a warmer climate and soil water properties vary widely according to hillside location, macro- and micro-relief, as well as soil types, simple upscaling is not possible.

4.4 Carbon Cycle

In a subsequent step, efforts are made to quantify these changes with respect to the terrestrial carbon balance. As one could expect from the results presented above, we find a major reduction of the forest carbon pool.

Although soil carbon densities in alpine vegetation are significantly higher than in forest soils (18.0 Kt C/km² compared to 12.3 kt C/km², see Körner *et al.*, 1993) the total amount of carbon stored is considerably reduced (Figure 41). The forest carbon is reduced to 59%, 38% and 25% of the current value in the respective scenarios -4°C , -6°C and -8°C . In total, we account a difference of 79.9 Mt C, 143.9 Mt C and 191.4 Mt C in the respective scenarios.

5 Discussion and Conclusion

A study on the impacts of abrupt climate change on Norway spruce mountain forests is presented. A review of our knowledge on the history of the Earth's climatic system leaves little doubt that the climate has jumped from one mode of operation to another (Broecker, 1995). Abrupt climate change can occur for many reasons, but it is conceivable that human forcing is increasing the probability of large, abrupt events (Alley *et al.*, 2003). It is agreed that future rapid climate changes might have large and unanticipated impacts, and that increased knowledge of impacts and tolerances of plant species is needed to consequently develop effective response strategies (Shugart, 1998; Alley *et al.*, 2003). In our layout, we chose both scenarios derived from paleo-records of past abrupt climate changes such as the Younger Dryas event as well as simulations of a state-of-the-art climate model, following the advice of Crutzen (2003) to not yet overly rely on models to make predictions. Since climate models are still hardly capable to reconstruct past climate events (Broecker, 1999) but past climate constellations are unique and not likely to appear again, a combination of both approaches seemed most suitable for this study. Furthermore, human changes in the atmospheric GHG composition pushed the system outside its "operating range" of the last 420,000 years (GCP, 2003), moving into uncharted territory (Figure 42).

We presented ecosystem response to two scenario storylines along an elevation transect finding major losses in biomass stock and increment for an immediate onset of rapid cooling. According to our results, major reorganizations can be expected in ecosystem distribution even for the lowest magnitude of undelayed cooling, being in good accordance to paleo-botanic findings (Tinner and Lotter, 2001; Wick, 2000). Increased magnitudes make impacts even more severe and bring large areas close to their ecotonal borderlines or even cause dieback of considerable zones and areas. The reaction to abrupt climate change seems to be closely related to the placement of the forest stand in the physiological range of the ecosystem. This study supports the theory established in paleo-botany that response is extensive close to an ecotone. In addition, the experiment leads to the conclusion that the closer a stand is to an ecotonal borderline, the faster is the response.

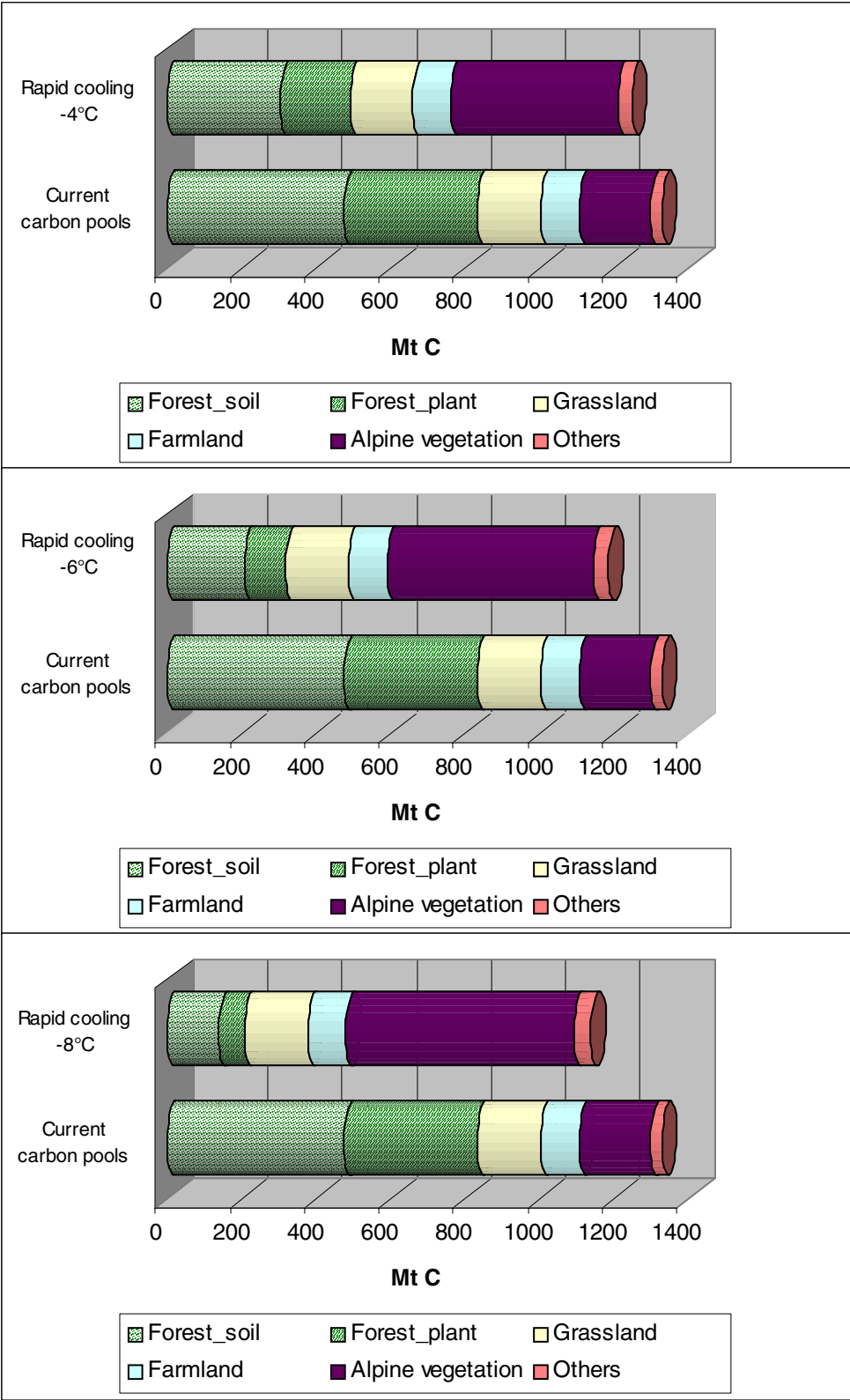


Figure 41: Changes in Austria's carbon budget as anticipated in the respective scenarios. Dieback and reduced biomass stock as well as changed soil carbon densities are taken into account.

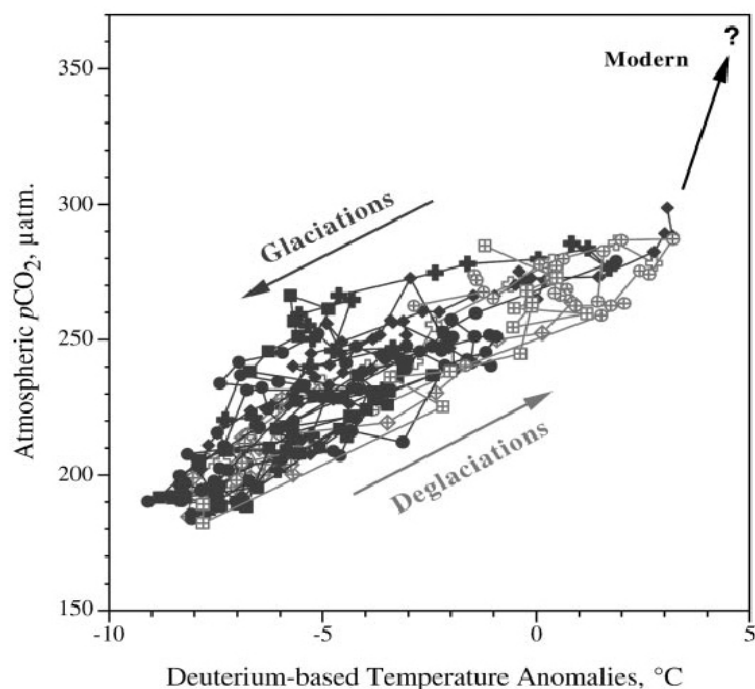


Figure 42: Atmospheric partial pressure of carbon dioxide over isotopic temperature anomalies as recorded in the Vostok ice cores demonstrate the relative constrained domain of climate variability in the last 420,000 years. Source: Falkowski *et al.* (2000).

For an anticipated onset of THC shutdown induced abrupt climate change 200 years from now, as projected with the CLIMBER-2 model of the Potsdam Institute of Climate Impact Research, the response is more complex. The distinct GHG warming of as much as 5°C in the first 150 years of the scenario leads first to enhanced growth of Norway spruce in the simulation. The more favorable growing conditions in terms of temperature are, however, offset in the second part of the warming phase by decreased precipitation, and the system changes from being temperature-limited to water-limited. This finding supports the concern that especially at low altitudes Norway spruce forests will suffer substantially from warming and eventually die back at sites where soil-water is already crucial. In this respect simulation results are in good accordance with a comprehensive study on the sensitivity of the Austrian forests to scenarios of climate warming (Lexer *et al.*, 2001), where especially Norway spruce stands at low elevations suffer severe impacts. During the warming phase of the studied scenarios in this assessment similar results are obtained finding dieback of stands at low elevation and poor soil water holding conditions. For the scenario simulating THC shutdown, forests show immediate response and growth is enhanced due to recovery from drought stress. Thus, as some authors suggest, a possible THC shutdown in the far future has the potential to at least regionally counteract the CO₂ induced warming. It would be of interest to extend the assessment beyond the year 2300 to see if ongoing cooling (see Figure 23) would eventually re-introduce temperature as a controlling factor of alpine Norway spruce forest ecosystems. Furthermore, the investigation of a multi-species set and a broader range of sites could add value to the interpretation of forest response to abrupt climate change.

The latter two aspects comprise major limitations of the interpretation of the results beyond stand level, which is, however, tentatively tackled in this study. In general, findings are in good agreement with pollen and macrofossil records of past abrupt changes, finding major drops in timber line. Simulated especial vulnerability of forest ecosystems above 1,000 m asl is in good accordance with results found for past events of rapid cooling (Wick, 2000; Kral, 1979). Also the higher vulnerability of ecosystems close to their ecotone could be found in the simulation experiment (cf. Yu, 2000; 2003). These broad-scale ecosystem responses are furthermore translated to the carbon budget in a rather crude equilibrium stock-comparison approach. As a result, an immediate onset of cooling would ultimately lead to a distinctly lower carbon stocks in biological storages. Although the transition to this projected future equilibrium does possibly take decades to centuries and the fate of carbon in eroded sediments remains still uncertain (Izaurrealde *et al.* 2001), this quantity is likely to have notable feedbacks on the climate system. This gets even clearer if one compares the projected amount of carbon release to Austria's energy related CO₂ emissions of 17.7 Mt C (65.0 Mt CO₂) as estimated in Jonas and Nilsson (2001) for 1990. Other adverse effects alongside such forest reorganization as increased frequency of mud-rock flows and avalanches as well as a collapse of the timber related economy are not studied here but can be expected in the wake of such an event. However, a comprehensive carbon cycle assessment under abrupt climate change would have to include not only forest changes but also changes in the agricultural sector or sea uptake because of changed ocean currents. Although there are no substantial studies conducted on the impact of abrupt climate change until now, the results of this study reflect the understanding of such events quite well. As an interesting reference Vellinga and Wood (2001) study impacts of a THC collapse and also try to quantify changes in vegetation net primary production (NPP). Their anticipated cooling of 2–3°C for Europe leads to a reduced NPP of 0.9 Gt carbon. Without quantification they state further "...that a fraction of the vegetation cannot be sustained and would die in a fully interactive vegetation scheme" (Vellinga and Wood, 2001, p.14). In addition they see a considerable release of carbon to the atmosphere as the effect of vegetation's response to abrupt climate change, introducing feedback between the THC and carbon cycle. Although they used a top-down approach assessing on the global scale and we applied a bottom-up layout starting at the level of individual trees, both results coincide in general findings as far as comparable.

Concluding, this study supports the assumptions of high vulnerability of alpine regions with steep climatic gradients and distinct vegetational belts and adds to the fact that the Alps would be especially affected by abrupt climate change. To make results more valuable with regard to mitigation of abrupt climate change impacts, it would be of great interest to take into account elements of uncertainty in this assessment. However, although scientists' abilities to predict the future will always have a component of uncertainty, this uncertainty should not be confused with lack of knowledge nor should it be used as an excuse to postpone prudent policy decisions based on the best information available at the time (Falkowski *et al.*, 2000). Thus, it is recommended to take abrupt climate changes into account in future research on issues of global change as well as in the development of response strategies across all sectors to mitigate its potentially dire impacts.

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