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Potential climatic transitions with profound impact on Europe

Review of the current state of six 'tipping elements of the climate system'

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Abstract We discuss potential transitions of six climatic subsystems with large-scale impact on Europe, sometimes denoted as tipping elements. These are the ice sheets on Greenland and West Antarctica, the Atlantic thermohaline circulation, Arctic sea ice, Alpine glaciers and northern hemisphere stratospheric ozone. Each system is represented by co-authors actively publishing in the corresponding field. For each subsystem we summarize the mechanism of a potential transition in a warmer climate along with its impact on Europe and assess the likelihood for such a transition based on published scientific literature. As a summary, the 'tipping' potential for each system is provided as a function of global mean temperature increase which required some subjective interpretation of scientific facts by the authors and should be considered as a snapshot of our current understanding.

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1 General concept of tipping elements

Definition of tipping elements for the paper In this review we follow the formal definition of tipping elements given by Lenton et al. (2008), which was formulated less rigorously for the Synthesis Report of the IARU Congress on climate change (Richardson et al. 2009). For all practical purposes the following concise formulation, which we will adopt for this paper, is sufficient.

> Tipping elements are regional-scale features of the climate that could exhibit a threshold behaviour in response to climate change—that is, a small shift in background climate can trigger a large-scale shift towards a qualitatively different state of the system (Fig. 1).

It should be noted that this definition includes the possibility of irreversible shifts and multiple stable states of a system for the same background climate (so-called hysteresis behaviour as illustrated in Fig. 2). It is, however, not restricted to these.

Role of self-amplification for tipping elements The word *tipping element* suggests the existence of a self-amplification process at the heart of the tipping dynamics. Once triggered it dominates the dynamics for a certain period of time and thereby induces a qualitative change within the system, e.g. from an ice-covered to an ice-free Arctic. If existent, understanding the self-amplification process is crucial to prevent tipping. A prominent example of such self-amplification is the ice-albedo feedback (Fig. 2) that is discussed to be operational in the Arctic sea-ice region and on mountain glaciers such as the Alps and the Himalayas: An initial warming of snow- or ice-covered area induces regional melting. This uncovers darker ground, either brownish land or blue ocean, beneath the white snow- or ice-cover. Darker surfaces reflect less sunlight inducing increased regional warming,[1] the effect self-amplifies.

[1]The phrase ice-albedo feedback is commonly used and refers to the changing reflectivity or *albedo* of the surface.

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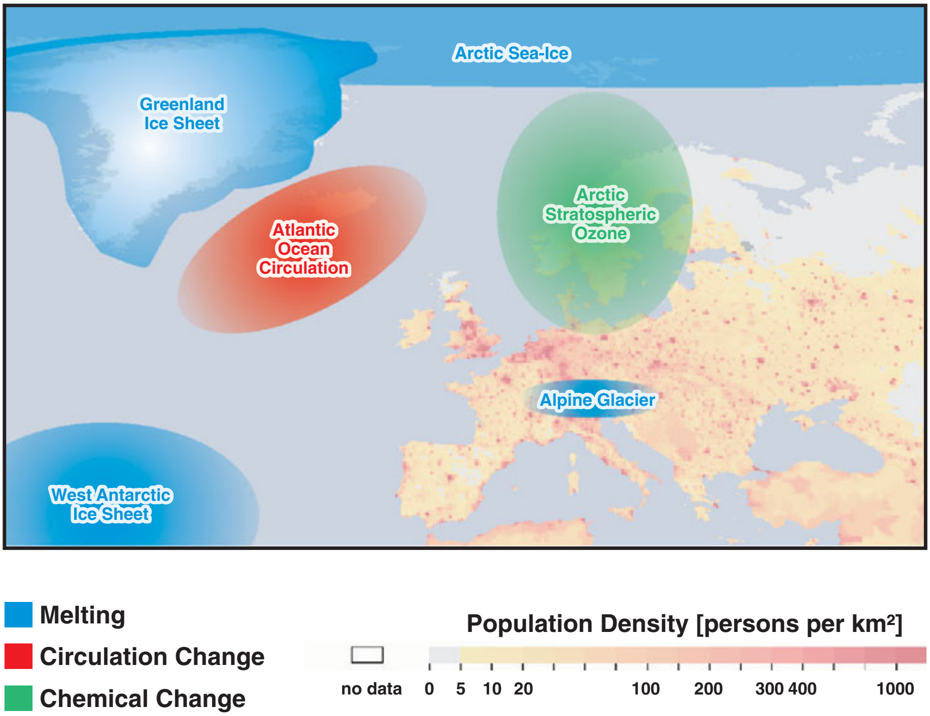

Fig. 1 Potential tipping elements with direct impact on Europe as discussed in this paper

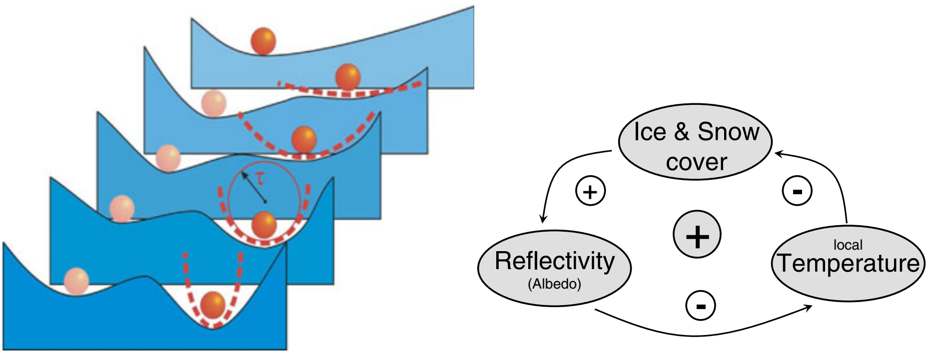

Fig. 2 *Left* Schematic illustrating the tipping of a system (Lenton et al. 2008). Initially (*front*), the system (*dark orange ball*) is stable within its background climate (*blue valley*). Initial changes in background climate do not alter the ball's position (or system's state). At a certain threshold small changes cause the ball to roll over. The system is tipping into a qualitatively different state. *Right* The ice-albedo feedback as an example of self-amplification which is at the heart of most *tipping elements*. A plus between two processes denotes an enhancing influence; a minus denotes reduction. For example, increased temperature reduces ice cover. An even number of minuses yields a self-amplification loop (denoted by '+' in the center of the loop)

Here, following Lenton et al. (2008), the tipping of a system is not defined through such self-amplification, but rather through the ratio of small external perturbation to strong system's response. Such a definition does not comprise any dynamic element. This is justified especially from some stakeholders' perspective (Lenton et al. 2009) which are mainly interested in whether a region will undergo exceptionally strong climate-related changes. For the example of the Arctic summer sea ice, we describe below that it is currently not clear whether the Arctic sea ice decline shows signs of internal acceleration. From the stakeholders' perspective, however, internal self-amplification is of secondary importance as long as the process is abrupt. For local communities as well as Arctic ecosystems it is more important that the sea ice is declining rapidly and that summer sea ice will most likely vanish for a further warming of 1–2°C.

As mentioned above, in this paper, we adopt the stakeholders' perspective and define *tipping elements* through a strong response to small external perturbations (as also applied recently in a Special Issue of PNAS (Schellnhuber 2009)). The authors emphasize however that a dynamical perspective might better reflect the public perception of the word *tipping element* and thus the dynamical perspective will be emphasized whenever it is applicable. In order to minimize any possible misconception, we adopt the term transition as a synonym for the term 'tipping' as defined by Lenton et al. (2008).

Structure of the paper and selection of tipping elements In the following sections, six different 'tipping elements of the climate system' with direct relevance for Europe are discussed (Fig. 1). Even though we can not claim completeness, the tipping elements discussed were selected and sorted according to the severity of their *direct* impact on Europe. It is important to note that a number of global tipping elements might have *indirect* effects on Europe possibly through a major disturbance of the climate system or migration of climate-change-induced refugees. The Himalayan glaciers, for example, store water which is released into the rivers of India, China and neighbouring countries. Current water supply during the dry season in these countries with more than two billion inhabitants depends on this storage mechanism. Comparable to other mountain glaciers the Himalayas are vulnerable to global warming through, for example, the albedo-feedback described in Section 5. Similarly important, monsoon systems in India, Asia and Africa support the livelihood of hundreds of million of people by providing precipitation for regional agriculture. Since monsoon circulations are sustained by a self-amplification process, they might show abrupt cessation (Levermann et al. 2009). Although monsoon rainfall in Asia seems to have undergone abrupt transitions in the past (Wang et al. 2008), their tipping potential has not been evaluated and no robust assessment can be given at this point.

Other processes might further amplify global warming and thereby affect also Europe. An example of a tipping element with such characteristic is thawing of northern hemispheric permafrost (Lashof 1989). The associated biological activity induces the release of methane and carbon dioxide from the ground. These are greenhouse gases and currently represent the two strongest anthropogenic contributions to global warming. The release per degree of global warming depends on a number of regional biological factors and is difficult to assess but poses a potential source of additional warming. Current assessments suggest that the self-amplification is, however, small (Stendel and Christensen 2002; Lawrence and Slater 2005).

In this review we focused on tipping elements with *direct* impact on Europe. It is important to note that we do not seek a comprehensive assessment of the systems but restrict the discussion to a potential transition into a qualitatively different state. Consequently, each section briefly describes the potential tipping element and the impacts of such a transition, followed by an explanation of the associated self-amplification process and a brief assessment of its tipping potential. We conclude with a comparison of tipping potentials and linkages between different systems (Fig. 15).

2 Ice sheets on Greenland (GIS) and West Antarctica (WAIS)

Current sea level contribution and potential future impact on Europe Most European coast line protection was initially built for the last century's sea level conditions and has mainly been readjusted moderately since. Though the situation may strongly differ from region to region, the maximum height to which dykes may be elevated rarely exceeds 1 m. Beyond this region-specific threshold significant rebuilding is necessary to protect land against storm surges and flooding. Most coastlines can not be protected against sea level rise of several meters. Therefore it is important to assess the potential for rapid sea level rise (SLR) within this century and beyond due to accelerated melt of the large ice sheets on Greenland and Antarctica.

Global warming of about $0.7° \pm 0.1°C$ during the last century has increased global sea level by about 0.15–0.2 m (Church and White 2006). Mountain glaciers and ice caps (MGIC) were responsible for about 0.05 m of SLR during 20th century. A similar contribution was due to oceanic thermal expansion. A possible source for the missing 0.05–0.10 m are the large ice sheets on Greenland and Antarctica. Direct observational data (Fig. 4) are, however, extremely limited prior to the 1970s. In the last 10–15 years this has changed. It has now been shown that both the Greenland Ice Sheet (GIS) and the West Antarctic Ice Sheet (WAIS) have been losing mass and this loss has been accelerating (Velicogna 2009). During this period, the much larger East Antarctic Ice Sheet (EAIS) has been approximately in balance (Rignot et al. 2008). These changes in ice sheet behaviour are recent and rapid and were not predicted by any of the current generation of ice sheet models. As a consequence, the Intergovernmental Panel on Climate Change (IPCC) suggested only modest contributions from the large ice sheets in its fourth assessment report in 2007 (Meehl et al. 2007). At the same time, already the observed sea level rise between 1990 and 2006 was underestimated by about 40% (Rahmstorf et al. 2007). It was acknowledged in the report that ice sheet processes were not adequately incorporated into projected sea level rise due to the inadequacy of the current generation of models. As a result, the projected global SLR of 0.20–0.60 m by 2100 underestimates the potential contribution of the ice sheets. A semi-empirical approach that links temperature increase above pre-industrial with the rate of sea level change yields much wider uncertainty for the respective IPCC scenarios in 2100 of 0.50–1.40 m (Rahmstorf 2007).

In recent years (since the mid 1990s) Antarctica exhibits net ice loss and is currently contributing about as much to global SLR as Greenland (Velicogna 2009).

An assessment of the potential contribution of the great ice sheets within this century is the subject of intense research efforts. The water stored in GIS is sufficient to raise global sea level by about 7 m. Although WAIS contains enough ice to increase global sea level by approximately 5 m, only about 3 m SLR equivalent are subject to potential self-amplifying ice discharge because they are grounded below the current sea surface (Bamber et al. 2009). The East Antarctic Ice Sheet could raise sea level by another ~50 m. Even though also in East Antarctica large areas of bedrock are below sea level evidence for the possibility of abrupt discharge there is not established.

During the last glacial period (about 20,000 years ago) large water masses were stored in ice sheets on the Northern Hemisphere. Furthermore colder ocean water was contracted and sea level was about 120–130 m below present levels. About 3 million years ago, global temperatures were higher than presently observed and reconstructions of past sea level show an elevation of 20–30 m above that seen today. Even higher temperatures 40 million years ago were associated with even higher levels of about 60–70 m above present levels. Despite large uncertainty it is clear that, in the past, sea level has responded to temperature changes of a few degrees by sea surface elevations of the order of tens of meters. These changes might have occurred in steps and not gradually and over long periods of time. The most recent period that was warmer than the present was the last interglacial, known as the Eemian, from 130,000 to 115,000 years before present. During this period, sea level was at least 4–6 m higher than today and summer temperatures were 3–6°C warmer (CAPE-Last Interglaicial Project Members 2006; Sime et al. 2009). Thus, it is evident that there is a profound difference between the equilibrium response of sea level to temperature and the transient, centennial to millennial, response that is important here.

Consequently current projections of SLR for the 21st century are one to two orders of magnitude smaller than the expected equilibrium response of SLR for the same temperature derived from paleodata (e.g. Overpeck et al. 2006). This is due to strong inertia in the system which causes sea level response to temperature changes to be relatively slow but also long lasting. The question is: How quickly can sea level rise in response to rapid temperature increase? Due to their potentially self-amplifying ice loss mechanisms, GIS and WAIS are particularly important in a risk assessment of future SLR. Mass loss of an ice sheet is not just associated with more water in the ocean. Loss of big ice masses affects Earth's gravitational field and thereby regional sea level. For example, the loss of the GIS reduces the gravitational pull into the North Atlantic, hence lowering sea levels and offsetting SLR in that region but enhancing SLR in other regions. As a consequence the water distribution within the oceans is changed which alters the sea level pattern. Figure 3 shows the combined effects of additional water and associated gravitational effects for GIS and WAIS. Northern European coastlines will thus be less affected by mass loss in Greenland, while a reduction in WAIS leads to even stronger sea level rise on the European and North American coast compared to the global mean. There are, however, also shorter-term effects related to ocean dynamics that may also lead to large regional variations in sea level rise and which are superimposed on any gravitationally driven changes (Kopp et al. 2010; Yin et al. 2009; Stammer 2008; Levermann et al. 2005).

Mechanism: Self-amplifying ice loss from Greenland GIS covers most of Greenland and reaches a thickness of up to 3,500 m. Since atmospheric temperatures decline

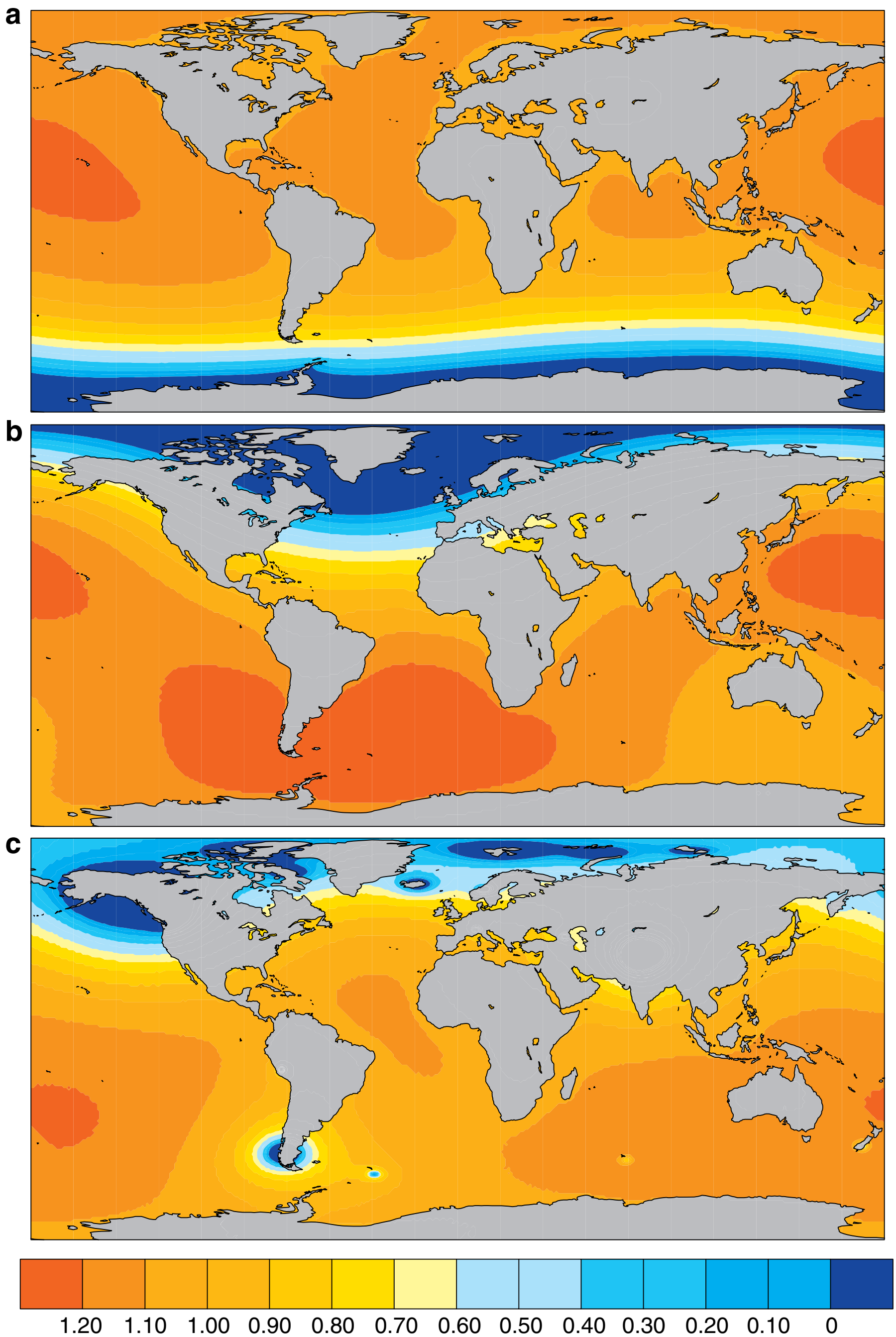

Fig. 3 Regional distribution to sea level rise from **a** West Antarctic Ice Sheet, **b** Greenland Ice Sheet (GIS) and **c** mountain glaciers. Regional heterogeneity arises from gravitational effects and slight changes in Earth rotation. Actual sea level rise (in m) is obtained by multiplication of values in panel a with ~3.5 m (Bamber et al. 2009) and values in panel **b** with ~7 m. Figure from Mitrovica et al. (2001)

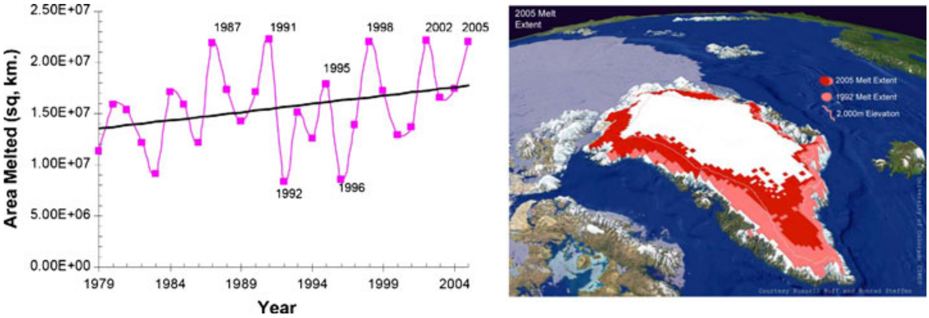

Fig. 4 GIS melting area shows strong variations from year to year with some underlying trend towards larger areas of melting (*left*). Since 1979 with the first available satellite images of the region, the largest melting area was observed during the warmest year on record, 2005, while the smallest melting area was recorded in 1992 after the Mount Pinatubo volcanic eruption (*right*). Figures from K. Steffen, University of Colorado, USA

with altitude,[2] GIS's highly elevated surface is significantly colder than it would be at sea level. This gives rise to a potential self-amplification process: If GIS loses ice, its surface elevation is lowered and its surface temperature increased. This enhances ice loss through melting and possibly the acceleration of iceberg discharge (surface-elevation-feedback).

Assessment of tipping potential for Greenland Ice Sheet (GIS) It is important to note that due to the surface-elevation-feedback, simulations suggest that GIS would not regrow under present climate conditions once it is eliminated and that its present existence is a relict of the last glacial period (Toniazzo et al. 2004; Ridley et al. 2010). From a stakeholder's perspective the relevant question, however, is whether there is a critical threshold temperature at which a complete disintegration of GIS is certain. In 2007, the IPCC-AR4 estimates this threshold to be 4.5 ± 0.9°C of warming over Greenland. Due to enhanced warming in high northern latitudes (Fig. 8) the associated range in global mean temperature is slightly lower (estimated to 3.1 ± 0.8°C by Gregory and Huybrechts 2006) and depends on the rapidity of Arctic sea-ice retreat (Section 4) as well as atmospheric dynamics that contribute to polar amplification of the anthropogenic warming signal. The IPCC-AR4 states that this threshold could be crossed within this century.

This estimate is, however, based on the so-called Positive-Degree-Day (PDD) approach, which employs an empirical relationship between surface melting and surface temperature. This parameterization needs to be calibrated using presently observed climatic conditions and it is questionable whether such calibration is valid for strongly altered boundary conditions as in a markedly warmer climate. More physically based energy-balance models tend to have a reduced sensitivity of the surface mass balance to increasing temperatures which might shift future threshold estimates towards higher values. Nonetheless, it is certain that increased

[2]On average temperatures decline by about 7°C for each kilometer altitude. Locally and temporarily this 'lapse rate' depends on weather conditions, but its order of magnitude is a robust feature which is fundamentally linked to Earth's gravity.

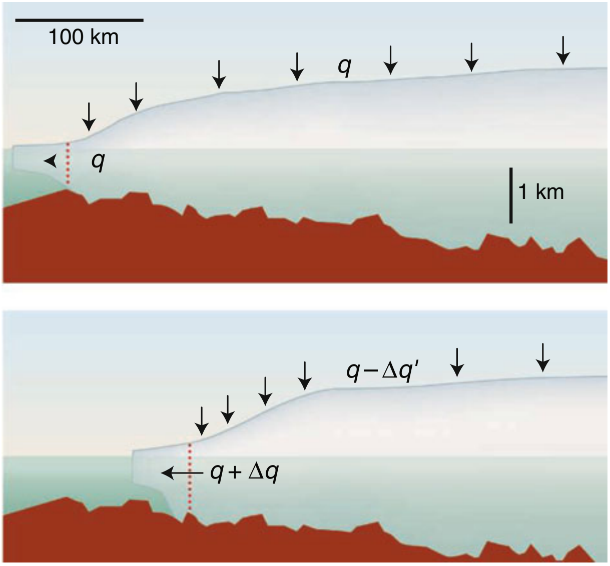

Fig. 5 Tipping of the West Antarctic Ice Sheet (WAIS). Possible self-amplification process of WAIS discharge (schematic from Vaughan and Arthern 2007). For regions in which the ice sheet is grounded below sea level ice flow across the grounding line (*dashed vertical line*) grows with ice thickness. If the bed is sloping down, ice discharge may self-accelerate

temperatures in the Arctic will result in increased mass loss from the GIS. What is less certain is the temperature at which the fate of the ice sheet is sealed. There is currently no evidence from model simulations or observational data that suggest that a near-complete disintegration might occur quicker than on a millennial time scale even for quite extreme warming scenarios (Ridley et al. 2005).

Land ice models are currently not able to capture observed acceleration of ice streams on GIS as for example the doubling in ice speed in the fastest flowing ice stream in Jakobshavn Isbrae (Joughin et al. 2004). Due to difficulties of current state-of-the-art models to simulate fast ice flow processes, such as a sufficient representation of basal ice shelf melting (Holland et al. 2008), models are likely to underestimate GIS sea level contribution of this century. Consequently scientists have employed a different approach to estimating the GIS sea level contribution within this century. Avoiding model simulations, Pfeffer et al. (2008) estimated the maximum contribution of GIS to global SLR as constrained by the maximum ice speed possible and the width of potential ice discharge outlets, to 0.54 m within this century.

Mechanism: Self-amplifying ice loss from West Antarctica Low temperatures in Antarctica inhibit ice-sheet melting and ice loss predominantly occurs (99%) through discharge across the so-called grounding line into ice shelves.[3] Ice shelves are floating ice masses of several hundred meters thickness which are subject to oceanic melting and refreezing, as well as calving into icebergs. Most bedrock beneath the WAIS is below current sea level. For such situations (Fig. 5), theoretical considerations suggest that ice flow through the grounding line increases with ice thickness

[3]The grounding line is the position at which land ice starts to float, i.e. at the grounding line the grounded ice sheet becomes floating ice shelf. Since the melting of floating ice does *not* raise sea level, it is the ice flow across the grounding line that matters for global sea level rise.

(Weertman 1974; Schoof 2007). Since bedrock is sloping down landward from the coastline in most of West Antarctica, this may lead to self-amplification: A retreat of the grounding line shifts its position towards regions of greater ice thickness. This enhances ice flow through the grounding line and yields a thinning of the still grounded ice which causes further retreat of the grounding line.

Assessment of tipping potential for West Antarctic Ice Sheet (WAIS) WAIS has collapsed at least once during the Quaternary, over the last 750,000 years. The most likely period for a collapse is around 400,000 years before present during a particularly warm interglacial (Scherer et al. 1998). Simulations in combination with paleo records suggest that a collapse took place several times during a period of prolonged warming about 3 million years ago (Pollard and Deconto 2009; Naish et al. 2009). During these periods Antarctica, as a whole, contributed to global SLR by about 7 m within a time interval of 1,000–7,000 years. For a complete collapse of the WAIS it would be necessary to largely remove the biggest ice shelves in Antarctica: the Filchner-Ronne and Ross. These buttress much of the vulnerable

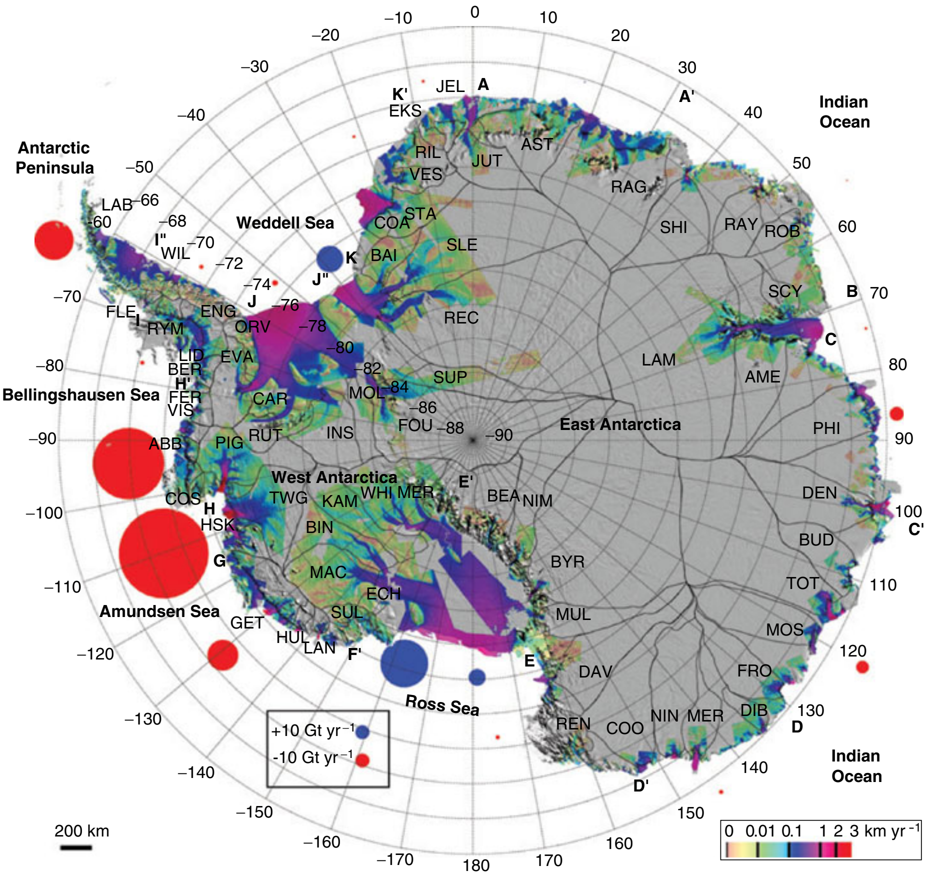

Fig. 6 Ice discharge along the West Antarctic coast has increased by more than 50% in 10 years (Rignot et al. 2008). *Red dots* indicate mass loss, *blue dots* mass gain

inland ice and regional warming of 5°C or more may be required to achieve this (Pollard and Deconto 2009). A partial collapse or retreat of the WAIS is, however, also possible and recent observations from satellites support theoretical analysis of how this might occur (Rignot 1998). Particularly glaciers in the Amundsen Sea sector show strong thinning, a retreat of the grounding line (Rignot 1998) and strong mass loss (Rignot et al. 2008), indicating the possibility that a partial disintegration might have been initiated. The ice volume associated with this region of WAIS is equivalent to a global sea surface elevation of about 1.5 m.

Recent observations in West Antarctica between 1992 and 1998 show a fast grounding-line retreat of the Pine Island Glacier of 1.2 ± 0.3 km (Rignot 1998), and an equally rapid grounding-line retreat (1.4 ± 0.2 km) and mass loss of the Thwaites Glacier (Rignot 2001; Rignot et al. 2002) between 1992 and 1998 (Fig. 6). Dynamic thinning along ice margins has been observed for most of the West Antarctic coast line (Pritchard et al. 2009) that is consistent with what would be expected in the case of grounding line instability. An integrated assessment of the risk of a WAIS collapse is currently not available. An estimate of a maximum contribution to global SLR from WAIS using the same approach as for GIS (Pfeffer et al. 2008) is questionable since outlet glaciers are less constrained by topography in Antarctica compared to Greenland and thus discharge is potentially quicker than on Greenland.

3 Atlantic thermohaline circulation (THC)

Potential impact on Europe The Atlantic thermohaline circulation (THC) is a large-scale ocean conveyor-belt circulation which transports about 1 PW $= 10^{15}$ W of heat towards the Nordic Seas (Ganachaud and Wunsch 2000) and thereby contributes to milder winters in northern Europe compared to regions of similar latitudes in North America and Asia. Climate models in which the THC has been collapsed by addition of artificial surface freshwater flux show a number of consequences of the THC cessation. Without its heat transport (Fig. 7) the Nordic Seas would be about 8°C cooler, and northern Europe, depending on atmospheric conditions and latitude, would be several degrees cooler than at present (Vellinga and Wood 2002). Europe would suffer from significant drying and reduced precipitation. Westerly winds would shift southward with reduced winds in the northern part and increased winds in the southern half of Europe (Laurian et al. 2009). Furthermore, simulations suggest that a THC collapse would increase sea level around European coast lines by up to 1 m (Levermann et al. 2005). This regional contribution would add on to global SLR and could be ten times quicker than presently observed rates, depending on the rapidity of the oceanic circulation changes.

In addition to these regional changes, the global climate system would be significantly perturbed by a THC collapse. Oceanic uptake of heat and carbon dioxide could strongly decrease and thereby accelerate global warming. Atlantic ecosystems are likely to be disrupted (Schmittner 2005; Kuhlbrodt et al. 2009) and the tropical rain belt would shift by several hundred kilometers southward in the Atlantic sector affecting populated areas in West Africa and the Amazon rain forest (Stouffer et al. 2006). Reconstructions of past climate suggest far reaching influences on the Asian monsoon system (Goswami et al. 2006).

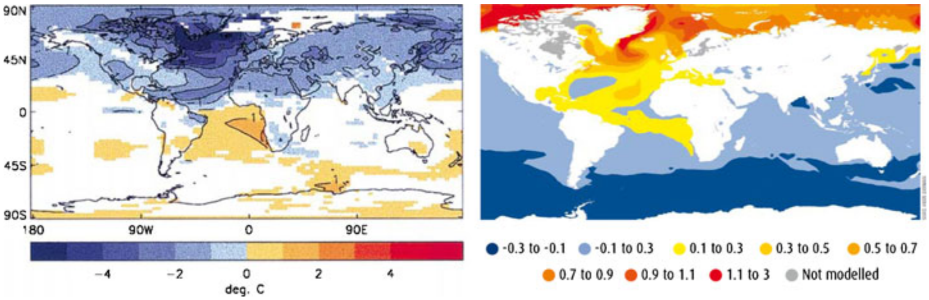

Fig. 7 A collapse of the Atlantic Thermohaline Circulation (THC) would have severe global consequences. Simulations under pre-industrial climate conditions in which a THC collapse has been forced by artificial additional freshwater flux in the North Atlantic show the following patterns: *Left* Temperatures in the Nordic Seas would drop by up to 8°C. Depending on atmospheric transport this yields several degrees of cooling in Europe (figure from Vellinga and Wood 2002) (© Crown Copyright 2002, the Met Office). *Right* In addition to global SLR due to warming, sea level would rise by up to 1 m along the European and North American coast (figure from Levermann et al. 2005)

Though most studies were carried out with climate models under present-day or pre-industrial boundary conditions, most of the effects have been shown to be robust under global warming (Vellinga and Wood 2007). In this situation, the global warming pattern including polar amplification, counters the effect of the reduced heat transport to the North Atlantic.

Mechanism: Self-amplified slow-down of THC The Achilles heel of the THC is deep water formation in the North Atlantic which is an essential component of the circulation. The density of North Atlantic water determines the strength of deep water formation and thereby of the THC. In the North Atlantic densification occurs through heat loss and salinity inflow which is partly provided by the circulation itself through import from the south. An initial reduction of the circulation thus reduces salinity transport to the north and further weakens the circulation (Rahmstorf 1996). Through the release of salt during sea ice formation (brine rejection) there is a strong link of the North Atlantic salinity budget to Arctic winter sea ice extent.

Assessment of THC tipping potential There are three lines of scientific reasoning on which the risk of a THC collapse is based. First, if the THC does indeed transport salt to the North Atlantic, the associated self-amplification process is based on robust large-scale features of the circulation and it is likely to have a significant influence. Observational data suggest that the present-day THC does transport salt into the Atlantic basin (Rahmstorf 1996; Weijer et al. 1999). Secondly, rapid reorganizations of the North Atlantic ocean circulation have occurred during the last glacial period (McManus et al. 2004). These were associated with strong global climatic disruptions (Rahmstorf 2002; Clark et al. 2002) and occurred on decadal to centennial time scales. Freshwater fluxes that caused past circulation changes have been estimated (Ganopolski and Rahmstorf 2001) to be of the order of expected melt water contributions from Greenland (Huybrechts et al. 2004) and potential future changes in net North Atlantic precipitation (Miller and Russell 2000; Winguth et al. 2005). It is, however, possible that stability properties of the Atlantic overturning are different under glacial and interglacial boundary conditions (Ganopolski and

Rahmstorf 2001; Weber and Drijfhout 2007). Thirdly, a variety of coupled climate models at different levels of complexity have shown abrupt THC collapse in response to systematically increased artificial Atlantic freshwater forcing (Rahmstorf et al. 2005). More complex and thus computationally less efficient models which were used for the IPCC-AR4 future projections are not able to perform this kind of systematic analysis. In these models a less systematic approach has been taken in order to assess the stability properties of the THC (Stouffer et al. 2006). Freshwater was externally applied for a period of one hundred years which forces a THC collapse. The cessation of the freshwater flux led to a resumption of the circulation in all of these models. Furthermore none of the IPCC-AR4 models show a THC cessation even for the strongest global warming scenarios during the 21st century (Gregory et al. 2005). These results seems to hold even when taking GIS melt water inflow into account (Hu et al. 2009; Jungclaus et al. 2006).

One needs to keep in mind that this does *not* prove that the models do not have two stable states. Neither is it certain that the models properly represent stability properties of the real ocean. In fact Weber et al. (2007) showed that while in the real ocean the THC transports salt into the Atlantic basin, this is not the case in all of these models. Thus state-of-the-art models seem to have a bias towards mono-stability (Hofmann and Rahmstorf 2009). Under global warming scenarios, all IPCC AR4 models, for which salt and freshwater fluxes are available, show an increased salt import into the Atlantic, i.e. the modeled circulations are moving towards a potential critical threshold (Drijfhout et al. 2010). On the other hand, it needs to be noted that simplified models which are computationally efficient enough to carry out a full stability analysis of the THC, might suffer from relevant short-comings such as unrealistic representation of surface buoyancy forcing which can alter the stability properties of the THC (Yin and Stouffer 2007).

It is also possible that the THC is vulnerable not only to large scale freshening of the North Atlantic as has been observed in recent decades (Dickson et al. 2002) potentially as part of a decadal oscillation (Hátún et al. 2005), but also to small-scale changes in the two main deep water formation regions in the central Greenland Sea and the central Labrador Sea. In both places open ocean convection in winter induces deep water formation in the region. In the Greenland Sea the process takes place in a very limited region of the gyre centre near 75°N 0°W, and was greatly assisted by the fact that the area over the site was covered for several months in winter by a locally-formed ice cover of pancake ice known as the Odden ice tongue, growing in the cold water of the Jan Mayen Polar Current which diverts east from the East Greenland Current. The brine retention by the ocean during pancake ice formation produced a negative buoyancy flux which models showed (Wilkinson and Wadhams 2003) to be the major factor in inducing overturning, which took place by means of convective chimneys extending to 2,500 m (Wadhams et al. 2002, 2004). Since 1998 changes in the atmospheric circulation, and warming of the ocean, have caused the Odden ice tongue to disappear. This is likely to have led to a decrease in the depth and volume of convection and may thereby influence the self-amplifying salt-advection feedback discussed above. (Note that sea ice-AMOC feedbacks may also work in the opposite way if reduction of atmospheric heat loss dominates over brine rejection in the local density budget (e.g. Levermann et al. 2007).)

An elicitation of experts on THC stability provided no clear picture on the risk of a future THC collapse. Subjective probabilities of different experts for triggering a breakdown within this century ranged from 0% to 90% (Zickfeld et al. 2007).

A more recent expert elicitation conducted by Kriegler et al. (2009) suggests less uncertainty and a clear increase of tipping potential with global warming (Fig. 14). The IPCC AR4 assesses the probability of a THC collapse within this century to 10% (Jansen et al. 2007). Generally expert assessments are a snapshot of the current understanding of the problem. In light, for example, of recent findings of an abrupt cooling that has taken place in the northern North Atlantic during 1970s in response to the Great Salinity Anomaly (Thompson et al. 2010), it is possible that the THC may now be viewed more sensitive than during these previous assessments.

Similar to the situation for WAIS also the north Atlantic circulation could exhibit a partial reorganization. While so-called Dansgaard–Oeschger events of the last glacial period might have been associated with abrupt transitions in the meridional circulation (Rahmstorf 2002), also abrupt transitions in the horizontal circulation can not be ruled out (Levermann and Born 2007). A wealth of paleo-records for the so called 8.2 K event at the beginning of the present interglacial can be explained by an abrupt strengthening of the north Atlantic subpolar gyre (Born and Levermann 2010). In model simulations such transitions require significantly less external perturbation than a collapse of the Atlantic overturning circulation. While it is clear that there is a strong link between the meridional and the horizontal circulation, it is not yet established how the respective tipping mechanisms are related.

4 Arctic sea ice

Potential impact on Europe While global mean temperature has risen by about $0.7^\circ \pm 0.1^\circ$C during the last century, Arctic warming has locally been two to four times higher. This polar amplification has a number of causes, one of which is melting Arctic sea ice and associated surface-albedo changes (van Oldenborgh et al. 2009; Winton 2006a). As a consequence, Europe has also warmed more than the global average—an effect that is going to persist under future increases of atmospheric greenhouse gas concentrations (Fig. 8) and would accelerate during accelerated deglaciation of Arctic sea-ice cover. Although models do not provide a uniform picture, sea-ice retreat can influence the North Atlantic atmospheric pressure system and thereby the Atlantic storm track into Europe (Kattsov and Källén 2004). Honda et al. (2009) have shown that strong reduction in Arctic summer sea-ice cover is associated with anomalously cold Eurasian winters and Petoukhov and Semenov (2010) found an up to three-fold increased probability of extreme cold events on northern continents in response to Barents-Kara sea ice reduction in winter. Furthermore, reduced sea-ice cover has profound impact on Arctic ecosystems. This includes marine mammals such as polar bears, seals, walrus and narwhales (Loeng 2004). Strongly reduced sea-ice cover yields improved accessibility to the Arctic including access to potential resources of fossil fuels in the region. The US Geological Survey estimates that about 25% of global oil resources may be found in the Arctic. The estimates are highly uncertain and the error bars range from 0% to 60% (http://www.usgs.gov/). However, potential recovery of these reservoirs will have significant environmental and geo-political implications.

Mechanism: Self-amplification of northern sea-ice melt Possible self-amplification of Arctic sea-ice melt could arise from the aforementioned ice-albedo feedback (Fig. 2), one of four fundamental climatic feedbacks discussed to be responsible for

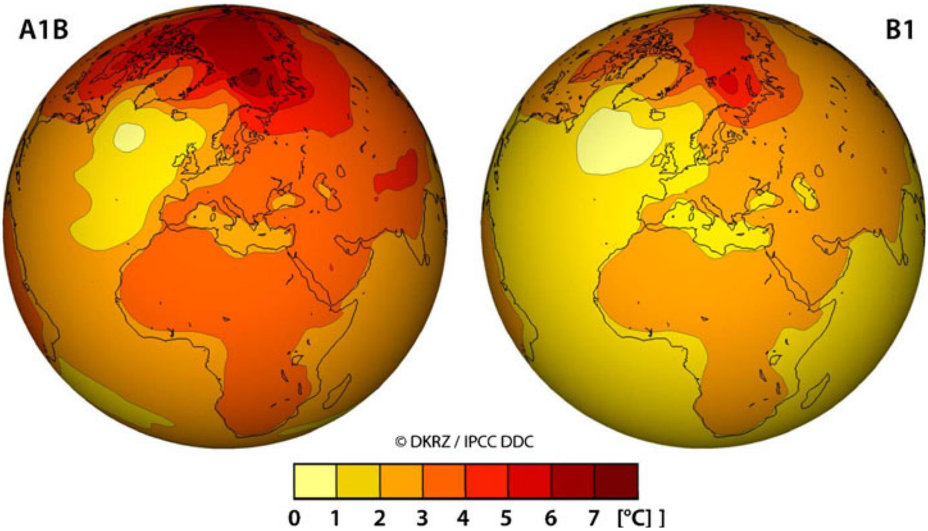

Fig. 8 Polar warming amplification partially caused by sea-ice melting for two scenarios (A1B (*left*) and B1 (*right*)). Temperature anomalies for the time period 2080–2099 compared to the period 1980–1999 were averaged over all models participating in the IPCC AR4 (Solomon et al. 2007). (Visualisation: M. Boettinger, DKRZ, Hamburg, Germany)

enhanced global warming in response to increasing greenhouse gas concentrations (Soden and Held 2006). The mechanism for the ice–albedo feedback is simple to understand: An initial temperature increase in high northern latitudes leads to melting of sea ice. As a consequence, less of the dark ocean is covered by highly reflective ice and snow, which leads to more absorption of sunlight at Earth's surface. This in turn causes more local warming and hence more melting of ice and snow. This self-amplification is mainly relevant for the Arctic summer sea-ice cover, since high-latitudinal solar insolation is strongly reduced in winter and much of the extra ice lost in summer can be regained during winter.

Assessment of tipping potential for Arctic sea-ice cover While all IPCC models agree that Arctic sea ice will decline in a warmer climate, these models do not show an irreversible or self-amplifying meltdown of Arctic summer sea ice (Winton 2006b). Hence, any slow down or even reversal of global warming will have a corresponding effect on Arctic summer sea ice (C.Amstrup et al. 2010; Notz 2009).

There are at least three factors which compensate the self-amplifying ice–albedo feedback and stabilize the Arctic sea ice cover such that its retreat is not self-amplified or irreversible: First, for a reduced summer sea-ice cover more open water is exposed to the atmosphere at the onset of winter. Because during winter ocean water is warmer than the surrounding sea ice, the ocean releases large amounts of heat to the atmosphere. In this way, the heat that has accumulated in the water in summer because of the ice–albedo feedback is released to the atmosphere during winter. Hence, the heat that accumulated in one summer is not carried over to the next summer (Tietsche et al. 2011). Second, thin ice grows much faster than thicker ice also because of the rapid loss of heat. Hence, after an extreme summer minimum the rapid growth of thin ice in winter is a stabilizing feedback that counter-acts the destabilizing ice-albedo feedback. Again, this resets the sea-ice extent each year and thereby reduces the tipping potential for Arctic summer sea ice (Eisenmann and Wettlaufer 2009; Fig. 9, right). Third, in areas that become ice free during summer, the snow that falls at the onset of winter (when snowfall rates are highest) does not accumulate on the ice but simply falls into the water. Hence, snow thickness on the ice that forms late in the season will be greatly reduced. Since snow is a very efficient insulator, such reduced snow cover also allows the ice to recover somewhat during winter.

Model simulations by Tietsche et al. (2011) show that the combination of these stabilizing feedbacks compensates the positive ice-albedo feedback throughout the entire transition to a seasonally ice-free Arctic. They compared a modeled trajectory of Arctic sea-ice evolution in the 21st century with a perturbed trajectory in which all sea ice was artificially removed at the beginning of summer in some years. Even though such artificial full removal maximizes the ice-albedo feedback, the perturbed sea-ice extent recovered to the extent of the original trajectory within two or three

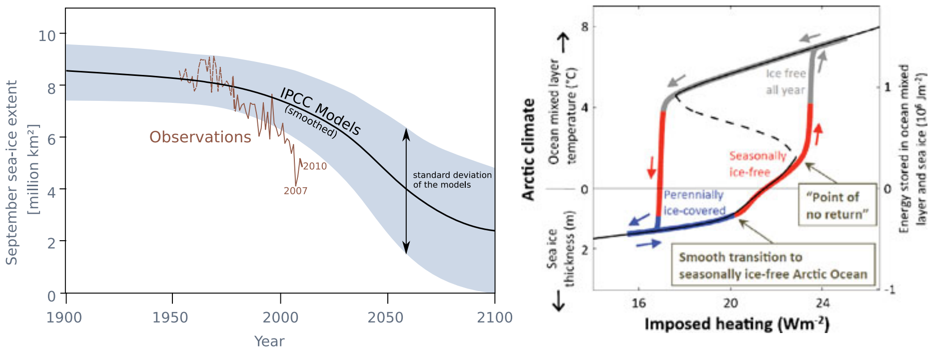

Fig. 9 *Left* Observed decline in September Arctic sea-ice cover (*red line*, in million square kilometers; *solid line* indicates satellite observations). The year 2007 showed an anomalously strong reduction of 23% compared to the previous record in 2005. Sea-ice extent recovered somewhat in 2008 and 2009, but reduced again in 2010. IPCC model simulations of 2007 (shading) strongly underestimated (currently observed) sea-ice decline (after Stroeve et al. 2007). *Right* Evolution of Arctic sea ice in response to warming simulated with an idealized physical model (Eisenmann and Wettlaufer 2009). The vertical axis represents the annual mean state of the upper ocean in terms of how much energy it would take to get to this point from an ice-free ocean that is at the freezing point. Initially (*bottom left*) there is a perennial sea-ice cover (*blue curve*) with an annual mean thickness of about 1.5 m. A transition to seasonally-ice free conditions (*red curve*) occurs in response to warming. At this point, cooling the climate would cause the ice cover to grow back to its original thickness. Further warming, however, causes the system to cross a point of no return and undergo a rapid transition to conditions which are ice-free throughout the year (*gray curve*). This transition represents an "irreversible process": considerable cooling would be required to get the ice to grow again (*arrows to left* along upper branch of the hysteresis loop). The stable and unstable steady-state solutions are indicated by the *solid* and *dashed black curves*, respectively. Note that this figure was derived from a simplified, conceptual model of the Arctic climate system. However, the finding that a tipping point is unlikely to exist for the loss of summer sea ice and that a tipping point might well exist for the loss of winter sea ice is consistent with studies employing state-of-the-art GCMs (Winton 2006a; Tietsche et al. 2011). Such consistency of results for models of largely different complexity is usually considered as a strong indication of their reliability

years, showing that summer sea-ice extent largely responds directly to the smooth rise in global temperature.

However, these stabilizing feedbacks are only functioning as long as there is still significant ice formation during winter. In an even warmer climate with a much reduced sea-ice cover also during winter, a tipping point for the loss of winter sea ice might well exist. In such a climate, Arctic winter sea ice vanishes abruptly and thereby constitutes a qualitatively different tipping element (Fig. 9, right).

Notwithstanding the low probability for a tipping of Arctic summer sea ice in the dynamical sense of a self-acceleration, Arctic sea ice is currently undergoing a significant transition both with respect to its areal extent and to its thickness. Satellite observations show a reduction in ice extent of almost 50% over the last 50 years (Fig. 9). Also ice thickness has reduced significantly in past decades (Haas et al. 2008; Rothrock et al. 1999; Wadhams and Davis 2000). Due to very strong inter-annual variability of sea ice extent especially in response to atmospheric pressure conditions and associated winds (Deser and Teng 2008), it is difficult to assess whether the retreat of Arctic sea ice is currently accelerating. Sea ice in the Arctic Ocean retreated at an annually-averaged rate of 2.8% per decade from 1979 to 1996, as measured by microwave satellites (Parkinson et al. 1999), which sped up to 10.1% per decade from 1996 to the record-low year in 2007 (Comiso et al. 2008). We do not know if this accelerated rate of loss will continue. Some IPCC AR4 models show the most rapid decline in summer sea ice when the summer extent is roughly half of the preindustrial extent while others show a more linear, on average, decline (Wang and Overland 2009).

The situation is further complicated by the fact that the variability of Arctic sea ice extent is probably going to increase in a warming climate and we expect larger negative and positive excursions from the mean downward trend such as that observed during the record sea-ice minimum in 2007 (Goosse et al. 2009; Notz 2009; Lindsay et al. 2009). During that record summer, minimum sea-ice extent dropped by about 23% compared to the previous record in 2005. Though this decline was caused by anomalous atmospheric and ocean conditions which can not directly be attributed to global warming (Kay et al. 2008; Perovich et al. 2008; Zhang et al. 2008; Ogi et al. 2008; Lindsay et al. 2009), the ice in the basin was also preconditioned to be quite thin due to both anomalous wind patterns in previous years and warming winters (Lindsay et al. 2009).

Data from submarines shows that the mean ice thickness in the Arctic Basin declined by 43% between the mid 1970s and the late 1990s (Rothrock et al. 1999; Wadhams and Davis 2000) with a loss of nearly three-quarters of the deep pressure ridges, so that at the beginning of the summer season the ice cover has been thinner and therefore more susceptible to the enhanced summer melt brought about by atmospheric and oceanic factors. The albedo feedback mechanism appears to have acted in this case through the enhanced production of surface melt pools, which preferentially absorb radiation and can melt through to form thaw holes, and the easier break-up of the thinner weaker floes. Both mechanisms create open water which itself absorbs radiation, warms up, and speeds up the melting of existing floe bottoms.

The anomalous wind patterns, particularly in the early 1990's, caused much of the older ice in the basin to be exported through Fram Strait so that the area covered by multiyear ice is now much smaller than in previous years. Thus the average age of the ice is younger and the mean thickness is thinner (Maslanik et al. 2007). While ice that is less than one year old rarely exceeds 2 m thickness, older ice grows to an average of about 3m thickness. Since a number of processes such as ice dynamics and ice transport through winds and ocean currents complicate the picture, current climate models have difficulties in capturing summer sea-ice evolution. The currently observed decline in Arctic sea-ice cover (Fig. 9) is stronger than simulated by any climate model that took part in the latest IPCC intercomparison (Stroeve et al. 2007). This shortcoming of the models is probably caused by a combination of very large internal variability of Arctic sea-ice extent that can lead to extreme minima and a lack of understanding of some underlying processes that are responsible for the recent sea-ice retreat. Since then models have improved and some capture sea-ice decline more satisfactorily (Wang and Overland 2009). Projections are highly dependent on the greenhouse gas emission scenario. Under unmitigated climate change[4] Holland et al. (2006) project an abrupt decline of Arctic summer sea ice starting around 2040 with a complete melting in 2050. This result is supported by Smedsrud et al. (2008) using a different model. While model studies suggest that Arctic summer sea ice will vanish at an additional global warming of 1–2°C, winter sea-ice cover is not likely to be eliminated for a warming of less than 5°C.

5 Alpine glaciers

Potential impact on Europe In concert with mountain glaciers world-wide (Fig. 10), glaciers in Europe have retreated considerably over the last 150 years (Braithwaite and Raper 2002; Oerlemans 2005; Kaser et al. 2006; Zemp et al. 2008; Cogley 2009). According to most recent estimates the ice volume of glaciers in the European Alps has been reduced from about 200–300 km^3 in the year 1850 to 90 ± 30 km^3 at present (Haeberli et al. 2007; Farinotti et al. 2009). Shrinkage of Alpine glaciers and snow cover is reducing surface reflectivity and thus leads to amplified temperature increase in the region. In combination with a generally enhanced continental warming this contributed to the anomalously strong Alpine warming which was about twice as high as the global average with significant acceleration in recent years (Auer et al. 2006). Glaciers are perceived as a symbol for a healthy mountain environment. Their retreat will have strong impact on tourism in Europe (Beniston 2003).

Since mountain glaciers and Alpine snow cover serve as freshwater reservoirs over seasonal to decadal time scales, glacier wastage will affect water availability in the region, in particular during summer. Generally it is observed that seasonality of run-off into rivers has increased. That is, stronger flow has been observed in the peak flow season and reduced flow or even drought in the low-flow season (Arnell 2004). Initially, snow melt and associated glacier retreat is projected to enhance summer flow from the Alps into European rivers. When snow cover and glaciers shrink, however, summer flow is projected to be strongly reduced (Hock et al. 2005; Huss et al. 2008). Through reduced run-off into large rivers such as Rhine and Rhone downstream regions will be affected. A change in hydrological regime is a robust feature of future projections for the European Alps (Eckhardt and Ulbrich 2005;

[4] That is, greenhouse gas emissions follow the so-called business-as-usual scenario, A2, of the IPCC Special Report on Emissions Scenarios (SRES).

Zwierl and Bugmann 2005) and is thereby anticipated in the IPCC 2007 assessment report (Kundzewicz et al. 2007). This will strongly affect hydropower production in Europe (Schaefli et al. 2007).

In addition, thawing of Alpine permafrost will destabilize the ground and result in land slides and debris flows that have been increasingly observed in recent years (Gruber and Haeberli 2007). Although thawing of permafrost is generally a slow process, strong 20th century warming in the Alps has already induced a pronounced thermal anomaly down to about 50–70 m below the surface (Harris et al. 2009; Noetzli and Gruber 2009). During the last century melting of mountain glaciers worldwide contributed to about 25% of the observed global sea level rise (Oerlemans et al. 2007). Over the next decades it is expected to contribute significantly although only about 0.5 m of global SLR equivalent remain in mountain glaciers (Meier et al. 2007). The Alpine contribution is however small compared to other sources like glaciers in Alaska, Patagonia and central Asia.

Self-amplification of Alpine glacier melt Several positive feedback mechanisms amplify the rate of Alpine glacier retreat: The reduction in snow- and ice covered area induces increased regional warming and ice melt through the ice-albedo feedback illustrated in Fig. 2 (Paul et al. 2005). Furthermore, enhanced dust accumulation on the bare ice has significantly decreased surface albedo leading to accelerated ice melt (Oerlemans et al. 2009). Over the last decades a prolongation of the melting season by one month has been inferred for glaciers in the European Alps, and the fraction of precipitation in the form of snow has decreased by more than 10% (Huss et al. 2009). Both processes have significant negative effects on glacier mass balance. The rapid changes in the climate system furthermore induce processes of down wasting of glacier tongues and collapse rather than "active" glacier retreat. This involves the disintegration of glacier systems into small individual parts, subglacial melting out of large cavities and lake formation. The protective effect of increasingly strong debris cover on glacier tongues cannot compensate for the above mentioned positive feedback mechanisms.

Assessment of tipping potential for Alpine glacier melt Over the last century glaciers in the European Alps experienced an average annual ice thickness loss of 0.2 to 0.6 m, the best estimate for the century average mass balance being −0.25 to −0.35 m water equivalent per year (Haeberli and Hoelzle 1995; Vincent 2002; Hoelzle et al. 2003). Strong variability in time and space can be documented (Huss et al. 2010; Paul and Haeberli 2008): Fast glacier mass loss comparable to present-day rates has already taken place in the 1940s and time periods of slightly positive mass balances with intermittent glacier readvance are documented for the 1890s, the 1920s and the 1970/80s (Fig. 10, right). The year 2003 showed exceptional mass loss with a decrease in mean ice thickness of almost 3 meters over the nine measured Alpine glaciers. This rate was four times higher than the mean between 1980 and 2001 and exceeded the previous record of the year 1996 by almost 60% (Zemp et al. 2009).

Glaciers in the Alps probably lost about half their total volume (roughly −0.5% per year) between 1850 and 1975. Roughly another 10% (20–25% of the remaining amount) may have vanished between 1975 and 2000 (updated after Haeberli et al. 2007) and again within the first decade (2000–2009) of our century (corresponding now to about −2% per year of the remaining volume). The melting out of the Oetztal

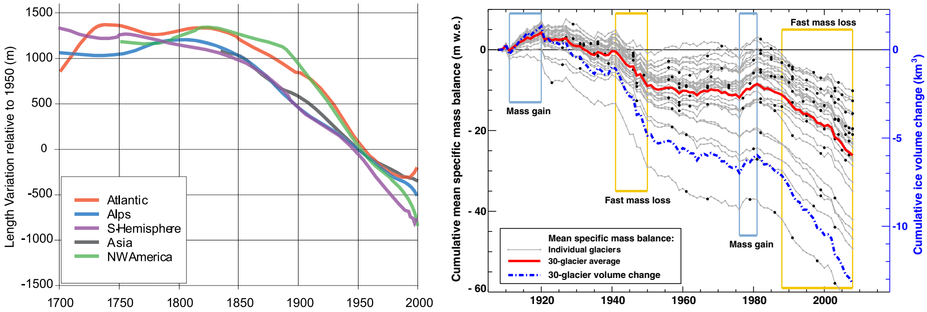

Fig. 10 *Left panel* Mountain glaciers are retreating globally. Large-scale regional mean length variations of glacier tongues (Oerlemans 2005). (Figure from IPCC fourth assessment report (Solomon et al. 2007, Chapter 4, p. 357), data from various sources (reconstructions, long-term observations) extrapolated to large regions.) *Right panel* Cumulative mean specific mass balance of 30 Swiss glaciers and their total cumulative volume change in the 20th century. Series for the individual glaciers are shown in *gray*. The *solid red line* represents the arithmetic average, and the *dash-dotted blue* line the cumulative total volume change of the 30 glaciers. Two short periods with mass gain and two periods with fast mass loss are marked. Figure from Huss et al. (2010). The volume loss, indicated in blue, is calculated from a multiplication of thickness losses and area

iceman in 1991 clearly demonstrated to a worldwide public that conditions in the Alps have reached if not exceeded the "warm" or "energy-rich" limits of glacier and climate variability during many thousands of years before (Solomina et al. 2008).

Simulations of Alpine glacier extent over the 21st century using different model approaches indicate unequivocally that an increase in global mean air temperature of 2°C (corresponding to +3–4°C locally) leads to an almost complete loss of glacier ice volume in the Alps (Zemp et al. 2006; Le Meur et al. 2007; Jouvet et al. 2009). Whereas small glaciers are expected to disappear in the next few decades, considerable amounts of "left-over" ice from large glaciers will persist throughout the 21st century due to thick ice bodies originating from colder conditions.

Mountain glaciers are highly sensitive to small changes in air temperature and precipitation and are thus excellent indicators for climate change. Their response to currently rising air temperatures is strong, and is further reinforced by self-amplification processes. Many Alpine glaciers currently experience accelerated wastage due to increased net forcing. On the basis of available data a clear tipping point in the dynamic sense can not be detected. However, potential re-growth of Alpine glaciers would require decades of cooler and wetter conditions. Near-complete deglaciation of the Alps during this century could only be avoided by strong mitigation efforts. A global limit of two degree temperature increase might not be sufficient to accomplish this goal.

6 Arctic ozone depletion

Potential impact on Europe Stratospheric ozone is absorbing ultra-violet (UV) solar radiation, especially UV-B radiation which is particularly harmful for human skin. The stratospheric ozone layer therefore provides protection against dermatological diseases, corneal and DNA damage. Ozone depletion especially above populated areas may enhance the risk of skin cancer and may cause immune suppression (e.g.

Stick et al. 2006). Due to the generally very low UV-radiation in high latitudes a UV-increase has profound influence on society and ecosystems in the Arctic. The marine food chain is affected through UV-sensitivity of surface layer algae. Due to a very stable stratospheric polar vortex over Antarctica, ozone depletion in response to anthropogenic emissions has been observed in the southern hemisphere stratosphere for several decades. In contrast the Arctic vortex is less stable than over Antarctica, owing to hemispheric circulation patterns. However, for most years since 1992, ozone depletion has been observed also in the Arctic - locally up to 70% below normal [Boreal winter 1999/2000 (Rex et al. 2002)] associated with enhanced UV radiation in northern and central Europe. Substantial reduction in ozone levels can be observed up to mid-latitudes (35°N) of southern Europe.

Mechanism: Self-amplifying northern ozone depletion Low stratospheric temperatures support the formation of polar stratospheric clouds (PSC) which generally enhance ozone depletion due to chemical reactions at their surface (Fig. 11). A strengthening of the polar vortex and associated lower stratospheric temperatures lead to ozone depletion which further cools the stratosphere (e.g. Weatherhead et al. 2004).

Assessment of tipping potential for Arctic ozone depletion The main driver for upper stratospheric ozone loss and for spring losses in the polar stratosphere is the chemistry associated with chlorine and bromine (Solomon 1999). Associated chemical reactions are strongly influenced by human emissions of CFCs which have been banned with the Montreal protocol in 1987. As a consequence northern hemisphere total ozone in mid-latitudes has shown a decline from late 1970s to mid 1990s (Fig. 12). Since then no clear trend is detectable.

Inter-annual variability is particularly strong in the Arctic. This is mainly due to a less stable polar vortex compared to the southern hemisphere and shows the influence of stratospheric dynamics on the ozone layer in the northern hemisphere. Stratospheric dynamics, including stability, strength and temperature of the polar vortex, determine the onset of ozone depletion and also influences the rate and severity of the depletion processes. Global warming of Earth's surface is associated with cooling in the stratosphere which enhances polar ozone depletion. The accumulation

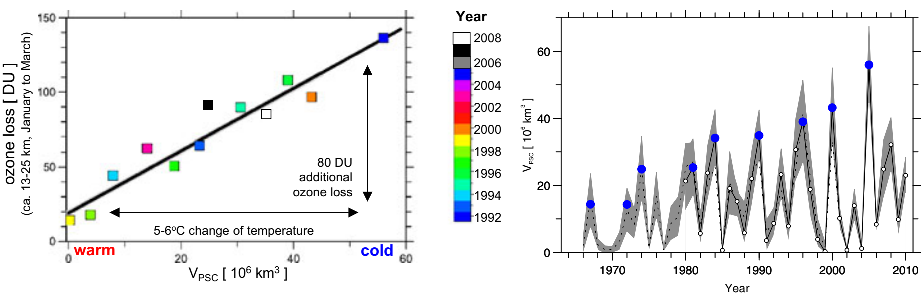

Fig. 11 Polar stratospheric clouds (PSC). *Left* PSC enhance stratospheric ozone loss (Harris et al. 2008). *Right* The volume of PSC estimated from meteorological analyses (*full line* ECMWF; *dashed line* FU Berlin) has increased in the cold stratospheric winters, but not the warm ones (WMO 2011)

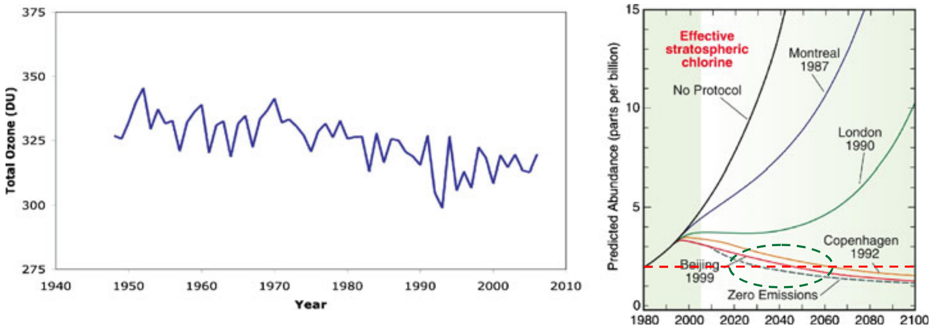

Fig. 12 *Left* Time series of annual mean values of total ozone abundance in Arosa, Switzerland (~45°N, figure from WMO 2007). *Right* Projected effective abundance of stratospheric chlorine in response to international treaties. The observed abundance closely follows the projected one, i.e. the line of zero-emissions will be crossed around 2030 after which Arctic ozone ceases to be a *tipping element*

of greenhouse gases in the troposphere (<~10 km altitude) warms Earth's surface but cools the stratosphere.

In the Arctic, this cooling is likely to lead to increased ozone destruction, as lower temperatures result in the formation and persistence of PSCs which aid in the activation of ozone-depleting compounds and can therefore accelerate polar ozone depletion. The Arctic Climate Impact Assessment of 2004 thus drew the conclusion that such cooling may induce self-amplification through the stabilization of the polar vortex (Weatherhead et al. 2004). Stratospheric cooling resulting from climate change is therefore likely to lead to an increased probability of larger and longer-lasting ozone holes in the Antarctic and extensive, more severe ozone losses over the Arctic (Dameris et al. 1998). In an analysis of approximately 2,000 ozone-sonde measurements, Rex et al. (2004) found that each 1°C cooling of the Arctic stratosphere resulted in an additional 15 DU[5] of chemical ozone loss due to increased PSC volume. Their findings indicate that over the past four decades, the potential for the formation of PSCs increased by a factor of three, resulting in stratospheric conditions that have become significantly more favourable for large Arctic ozone losses. However, a series of warm, disturbed northern hemispheric stratospheric winters occurred since the late 1990s to 2000s due to enhanced planetary wave activity (Manney et al. 2005, 2008, 2009) leading to a low PSC formation and hence ozone depletion potential (Fig. 11, right panel). Thus, a future projection of the Arctic ozone layer is highly uncertain due to the large inter-annual variability observed in boreal winter (WMO 2011).

The situation is further complicated through other radiative effects that influence the ozone budget of the stratosphere. One is a potential increase in stratospheric water vapour due to changes in tropopause temperatures (Evans et al. 1998). Increased water vapour is likely to contribute to increased ozone destruction by affecting the radiation balance of the stratosphere (Forster and Shine 2002; Shindell 2001). Greater

[5] DU = Dobson unit measures atmospheric ozone content. 1 DU corresponds to 0.01 mm ozone layer thickness under standard conditions of 0°C and 1 atm. pressure.

water vapour concentrations in the stratosphere can raise the threshold temperatures for activating heterogeneous chemical reactions on PSCs, and can cause a decrease in the temperature of the polar vortex (Kirk-Davidoff et al. 1999). Few long-term datasets of water vapour concentrations are available, but previous studies of existing observations have suggested that stratospheric water vapour has been increasing up to 1999 (Oltmans and Hofmann 1995). Analyses of 45 years of data (1954–2000) by Rosenlof et al. (2001) found a 1% per year increase in stratospheric water vapour concentrations. Since 1999 there is no evidence for an increasing trend (Jones et al. 2009; Randel et al. 2004) while an overall decrease is observed which feeds back onto the tropospheric temperatures temporarily decelerating the global warming trend (Solomon et al. 2010).

On the other hand, climate change could possibly trigger an increase in planetary waves, enhancing the transport of warm, ozone-rich air to the Arctic (Schnadt et al. 2002). This increased transport would counter the effects of heterogeneous chemistry and possibly accelerate recovery of the ozone layer. Recently Tegtmeier et al. (2008) showed that dynamical transport of ozone into the Arctic polar vortex in the past has contributed equally strong to inter-annual variability as variations in chemical ozone loss. It is currently not possible to make definite statements about the tipping point in the chemical destruction of Arctic ozone. If the emission of ozone-reducing chemicals is reduced in the future following the signed treaties, then the specific risk of a tipping of the Arctic ozone will become insignificant between 2030 and 2060 (Fig. 12). After that, unabated global warming, however, may lead to qualitative changes in atmospheric circulation patterns associated with the polar vortex. Since the lower stratospheric wintertime circulation can strongly influence the probability of extreme surface weather such as minimum daily temperatures in Europe (Scaife et al. 2008), these circulation pattern changes have the potential to exhibit tipping-element-like behaviour in a statistical sense.

7 Conclusions

The necessity to avoiding dangerous climate change (Schellnhuber et al. 2009) had been adopted as a goal by the UNFCC in the year 1992. An assessment of the likelihood of a major transition of different tipping elements is as scientifically challenging as it is crucial for future societal, political and economic decisions. Such assessment needs to be based on a thorough understanding of the systems in question and might evolve while scientific insight deepens. In light of associated risks even incomplete knowledge needs to be exploited to provide 'educated guesses' on the basis of available information. Such assessment will, by definition, always be preliminary and will permanently evolve. In light of natural climate variability, even the detection of the ongoing tipping of a climatic system presents a scientific challenge. Time series of relevant observables rarely exceed several decades in length which might not be sufficient to identify an acceleration in the system beyond any doubt. There are, however, a number of universal precursors such as enhanced variability when approaching a critical threshold which might be used for monitoring systems (Scheffer et al. 2009). This potential has neither been explored nor applied to the largest possible extent.

Linkages between tipping elements Matters are further complicated by the fact that some tipping elements are linked (Fig. 13). For example, increased sea level by GIS melting will elevate ice shelves in Antarctica and might thereby induce a retreat of the grounding line. Most linkages, however, involve the Atlantic thermohaline circulation. A collapse or even only a reduction of its meridional circulation component will cool northern high latitudes which might stabilize melting of the GIS and Arctic sea-ice even though the cooling effect will be strongest in winter while the melting occurs in summer. On the other hand GIS melting will freshen the North Atlantic and might thereby trigger a THC break-down. The resultant changes in tropospheric circulation and weather will be modulated by the lower stratospheric circulation. Strong reduction in Arctic sea ice will change the salinity and heat budget of the Nordic Seas and thereby influence the Atlantic ocean circulation (Levermann et al. 2007). Less sea ice cover will induce enhanced warming in high northern latitudes and increase melting on Greenland and even in the Alps. In the Southern Ocean a THC reduction will lead to a warming around Antartica. Furthermore it might shift the subpolar wind belt (Vellinga and Wood 2002), alter oceanic gyre circulation around Antartica and thereby induce changes in ice shelf melting (Hattermann and Levermann 2010) and influence the stability properties of WAIS. Though these connections exist, so far neither model results nor paleo evidence has clearly shown the tipping of one of these systems due to the tipping of another.

Tipping elements and their transition potential The most recent *comprehensive* assessment of a number of tipping elements and their linkage was presented by Kriegler et al. (2009). They conducted an expert elicitation on subjective probabilities for the tipping under different future warming scenarios (Fig. 14). Results show that experts consider the risk of tipping of major climatic subsystems significant. This holds especially for high warming scenarios but numbers are still far from

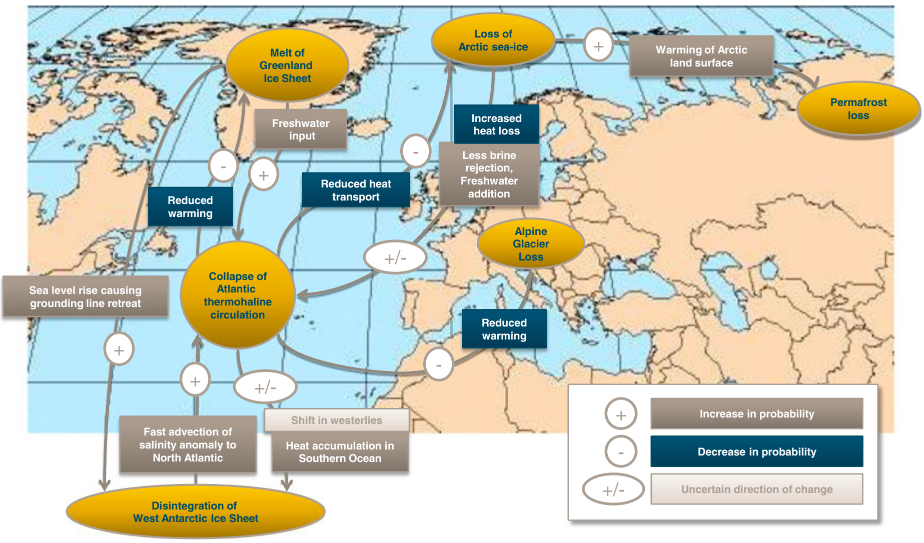

Fig. 13 Potential linkages between tipping elements

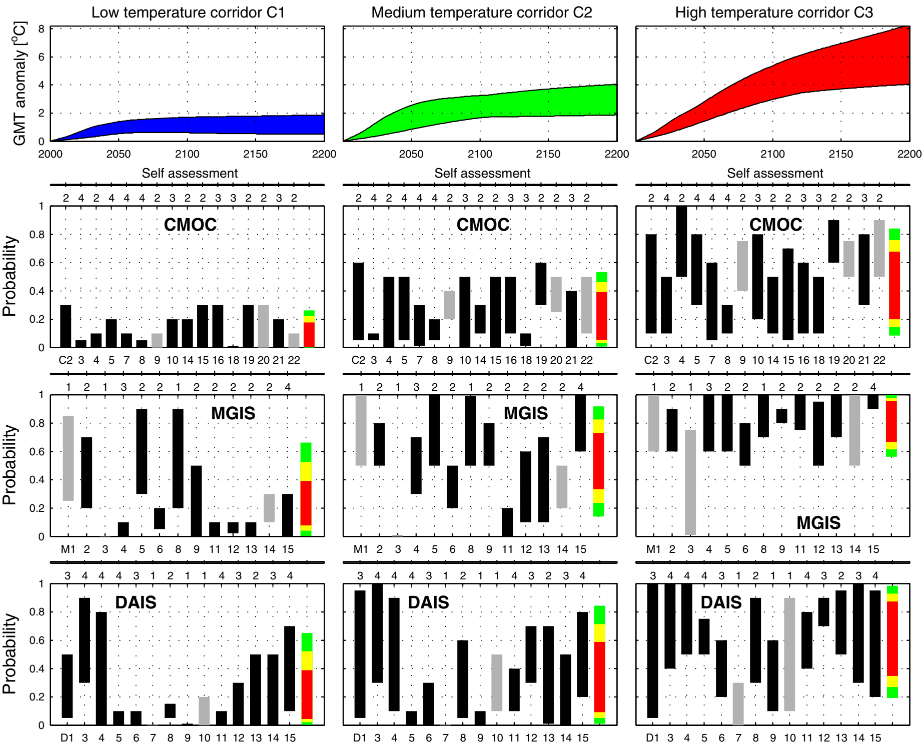

Fig. 14 Subjective probabilities provided by experts for the tipping of THC (denoted CMOC), GIS (denoted MGIS) and WAIS (denoted DAIS). The x-axis provides the number of the expert. *Coloured bars* represent a *core group* of experts for each tipping element which are actively publishing on the subject. The *upper panel row* provides the corresponding climate scenarios as represented by the evolution of the global mean temperature (GMT) during the 21st and 22nd century. High emission scenarios (*right panels*) yield probabilities predominately above 50% for each system and even for low warming scenarios (*left panels*) the elicited tipping potentials are not negligible. The *rightmost bar* in each panel shows the aggregation of probability intervals from core experts based on increasingly restrictive assumptions about expert weights: (i) weights are allowed to vary by ±100% (*green*) or ±50% (*yellow*) around uniform weights, and (ii) unweighed average of lower and upper bounds (*red*). The increasing strength of assumptions leads to nested probability intervals (*Red* < *Yellow* < *Green*). For details confer Kriegler et al. (2009)

small for a moderate temperature increase within this century. Here we provide a condensed assessment of the potential of a transition to occur in each of the subsystems in Fig. 15. Tipping elements are sorted according to the severity of their impact on society. The color coding represents 'tipping potential' for different global mean temperature increase. The width of the columns reflects the confidence that the authors have in their assessment. Naturally confidence is relatively high that no transition has occurred for present-day conditions even though we can not be entirely certain about this. For most systems confidence in the assessment that a transition will occur increases with increasing levels of global warming. A special case is the collapse of the THC. While the THC can be reduced by changes in surface heatflux, a qualitative change in the circulation is induced through changes in the North Atlantic salinity distribution which is only indirectly related to increasing

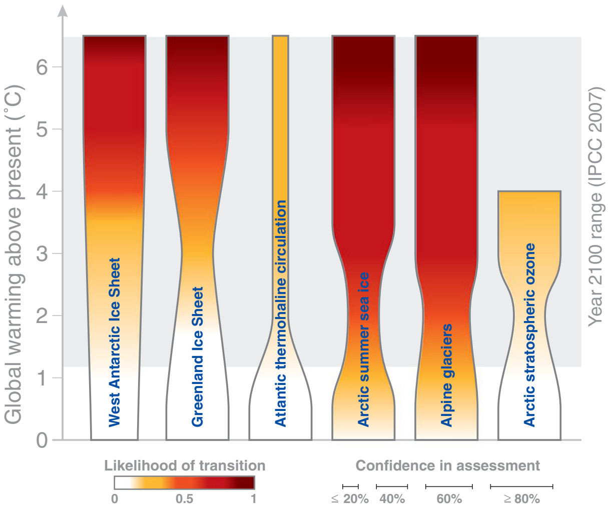

Fig. 15 'Burning rivers' summarizing the authors' general assessment of the potential of a transition of each system into a state that differs qualitatively from their present state. Colour coding represents the authors assessment of the likelihood of a transition for different global temperature increase. The width of the column represents the authors' confidence in their assessment, i.e. the narrower the 'river' the less confident the experts are in their respective assessment. For most systems the risk of tipping increases with temperature along with the confidence in such an assessment. An exception is the potential collapse of the Atlantic overturning circulation. Such a transition depends on the freshwater inflow into the North Atlantic which is only indirectly related to the global mean temperature increase through Greenland melting and precipitation changes. Especially because of uncertainty with respect to future precipitation changes, confidence in the tipping potential for the THC does not increase with temperature. The risk of reaching a tipping point in Arctic ozone depletion will become insignificant when chlorine levels drop below 1980 levels which is projected to occur around 2060 (WMO 2007; SPARC 2010). In the specific case of ozone depletion there exist significant uncertainty on the nature of the state to which the atmospheric circulation might revert to. All other systems are cryospheric and thus the likelihood of a transition increases with temperature. Due to the possibility that a partial disintegration of the WAIS in the Amundson Sea sector might have been already initiated the corresponding confidence that no transition has occurred for zero temperature increase is slightly reduced

temperature through GIS melting and changes in precipitation. Confidence about the likelihood of a collapse thus remains low even for high temperatures.

The WAIS bears the potential of abrupt solid ice discharge in response to oceanic warming, but currently no direct temperature estimates for such tipping is available. Paleo climatic evidence (Naish et al. 2009) in combination with land ice dynamics simulations (Pollard and Deconto 2009) suggest that abrupt discharge has occurred at temperatures 1–2°C above present. It should be noted that also a partial WAIS disintegration is possible. Satellite observations show strong glacier thinning and a retreat of the grounding line in some regions (Pritchard et al. 2009). At present it can not be ruled out that a partial collapse of WAIS within the Amundson Sea sector equivalent to 1.5 m SLR might have been initiated (Joughin et al. 2009; Chen et al. 2009; Pritchard et al. 2009). The stability of the GIS has been investigated more intensely than WAIS stability. Available estimates of the threshold temperature for GIS of 3.1 ± 0.8°C (Section 2) might however not be robust since they are based on simplified parameterizations of the surface mass balance. Our current level of understanding suggests that Arctic sea ice and Alpine mountain glaciers are the most vulnerable to global warming of the presented short list of tipping elements even though currently self-acceleration in a dynamical sense can not be detected. It is possible that even mitigated climate change, which does not exceed 2°C of global warming, is not sufficient to avoid qualitative change of these glacial regions. The risk of a tipping point in Arctic ozone depletion will become insignificant when chlorine levels drop below 1980 levels which will occur by 2060. Since it is very unlikely that global warming will exceed 4°C by 2060 no assessment for higher temperatures is provided.

The schematic of Fig. 15 is based on the scientific evidence presented in this paper. The assessment is, however, necessarily subjective and might change with future studies. Impact associated with the tipping of each of the presented systems are of continental or even global scale.

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