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Arctic Climate Tipping Points

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Abstract There is widespread concern that anthropogenic global warming will trigger Arctic climate tipping points. The Arctic has a long history of natural, abrupt climate changes, which together with current observations and model projections, can help us to identify which parts of the Arctic climate system might pass future tipping points. Here the climate tipping points are defined, noting that not all of them involve bifurcations leading to irreversible change. Past abrupt climate changes in the Arctic are briefly reviewed. Then, the current behaviour of a range of Arctic systems is summarised. Looking ahead, a range of potential tipping phenomena are described. This leads to a revised and expanded list of potential Arctic climate tipping elements, whose likelihood is assessed, in terms of how much warming will be required to tip them. Finally, the available responses are considered, especially the prospects for avoiding Arctic climate tipping points.

Keywords Arctic · Tipping points · Sea-ice · Greenland ice sheet · Atlantic thermohaline circulation · Boreal forest

INTRODUCTION

The Arctic is undergoing striking changes in climate. Regional near-surface air temperatures are rising at two to four times the global average rate (Screen and Simmonds 2010). The record minimum area coverage of Arctic sea-ice in September 2007 was part of an ongoing abrupt decline in ice thickness and total volume (Allison et al. 2009). The Greenland ice sheet (GIS) is losing mass at a rate that has been accelerating (Rignot et al. 2007; Pritchard et al. 2009). Permafrost is thawing rapidly in Northern Alaska and forming thermokast lakes (Jorgenson et al. 2006). A massive insect outbreak has struck the boreal forest in Western Canada (Kurz et al. 2008a). The list goes on.

The unforeseen and abrupt nature of these recent Arctic changes lends support to the view that human-induced climate change is unlikely to involve a smooth and entirely predictable transition into the future. In the past, a variety of abrupt climate changes have occurred in the Arctic region, which are defined by the climate response having been much faster than the factors driving it (Alley et al. 2003). Such 'non-linear' behaviour implies the existence of positive feedbacks in several parts of the Arctic climate system. Recent rapid Arctic warming has been linked to the retreat of the sea-ice, which exposes a darker ocean surface that absorbs more sunlight (Screen and Simmonds 2010), generating a positive feedback that is already amplifying regional climate change.

The Arctic sea-ice, GIS, Atlantic thermohaline circulation (THC), and boreal forest have previously been identified as potential 'tipping elements' in the Earth system—climate subsystems that could exhibit a 'tipping point' where a small change in forcing (in particular, global temperature change) causes a qualitative change in their future state (Lenton et al. 2008). The resulting transition may be either abrupt or irreversible or, in the worst cases, both. In the language of the Intergovernmental Panel on Climate Change (IPCC), these are 'large-scale discontinuities' (Smith et al. 2009), and are arguably the most dangerous type of climate change (Schellnhuber et al. 2006; Lenton 2011a).

This aim of this article is to examine more closely which parts of the Arctic climate system might pass a future tipping point (and whether any have already passed a tipping point). First a climate 'tipping point' is defined. Past abrupt climate changes in the Arctic are briefly reviewed.

Then the current behaviour of a range of Arctic systems is summarised. Looking ahead, a range of potential tipping phenomena are described, and their likelihood is assessed. This leads to an expanded list of potential Arctic climate tipping elements. Some of these involve terrestrial biomes, whereas Arctic marine ecosystem tipping points are dealt with elsewhere (Duarte et al. 2012 [this issue]). Finally, the available responses to approaching Arctic climate tipping points are considered.

DEFINING CLIMATE TIPPING POINTS

In colloquial terms, the phrase 'tipping point' captures the notion that 'little things can make a big difference' (Gladwell 2000). In other words, at a particular moment in time, a small change can have large, long-term consequences for a system. To apply the term usefully to the climate (or in any other scientific context), it is important to be precise about what qualifies as a tipping point, and about the class of systems that can undergo such change. To this end, we introduced the term 'tipping element' (Lenton et al. 2008) to describe large-scale subsystems (or components) of the Earth system that can be switched—under certain circumstances—into a qualitatively different state by small perturbations. In this context, the tipping point (or threshold) is the corresponding critical point—in forcing and a feature of the system—at which the future state of the system is qualitatively altered. For a system to possess a tipping point, there must be strong positive feedback in its internal dynamics, i.e. strong 'self-amplification' of external forcing (Levermann et al. 2011). So, when trying to identify climate tipping elements, we should look for positive feedback processes.

To formalize the notion of a climate tipping element further (Lenton et al. 2008), it is important to define a spatial-scale. As the climate itself has a characteristic length scale of order ~ 1000 km, only components of the Earth system associated with a specific region or collection of regions, which are at least of this sub-continental scale, were considered. Of course tipping points can occur in much smaller-scale systems, and elsewhere several ecosystem examples are discussed (Duarte et al. 2012 [this issue]), but here the focus remains on the sub-continental scale. For a system to qualify as a tipping element, it must be possible to identify a single control parameter (ρ), for which there exists a critical control value (ρ_{crit}), from which a small perturbation ($\delta\rho > 0$) leads to a qualitative change in a crucial feature of the system (ΔF) after some observation time ($T > 0$). In this definition (Lenton et al. 2008), the critical threshold (ρ_{crit}) is the tipping point, beyond which a qualitative change occurs. This change may occur immediately after the cause or not become apparent until much later.

Many scientists take 'tipping point' to be synonymous with a 'bifurcation point' in the equilibrium solutions of a system (Fig. 1a), implying that passing a tipping point necessarily carries some irreversibility (e.g. Tietsche et al. 2011). Others associate a 'tipping point' with a 'point of no return', also implying irreversible change. However, continuous changes without bifurcation, which are therefore reversible (e.g. Fig. 1b), can also meet the tipping point definition (Lenton et al. 2008). In reality, the existence or not of a tipping point should be considered in a time-dependent

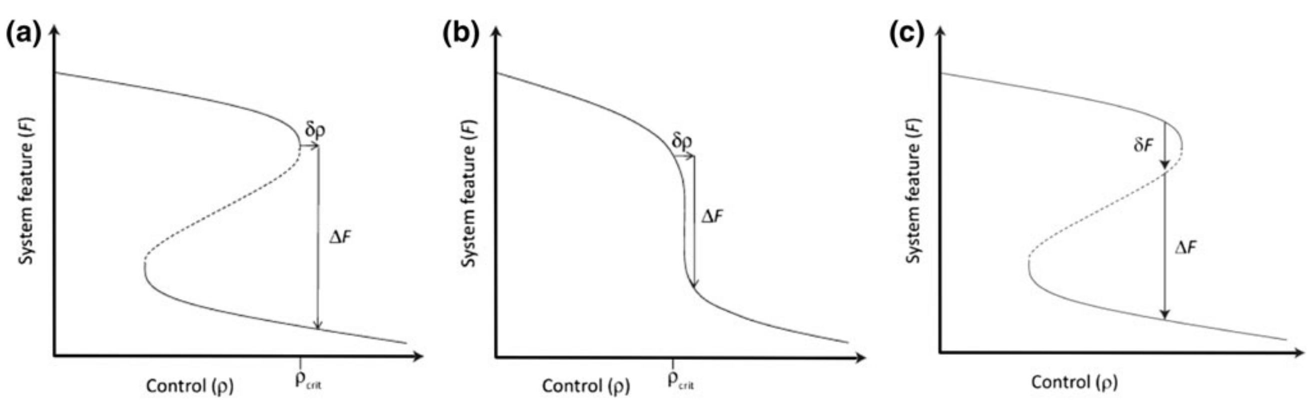

Fig. 1 Different sources of abrupt climate change. Schematics in terms of the time-independent equilibrium solutions of a system, where solid lines are stable steady states and dashed lines are unstable steady states: **a** Bifurcation-type tipping point: a system with bi-stability passing a bifurcation point leading to irreversible change (hysteresis). **b** Reversible tipping point: a mono-stable system exhibiting highly non-linear but reversible change. **c** Noise-induced transition: a bi-stable system transitioning between states due to internal variability. A tipping point (**a**, **b**) occurs when a small change in forcing ($\delta\rho$) results in a qualitative change in system state (ΔF), whereas a noise-induced transition (**c**) happens when internal short-term variability (δF) causes a large change in system state (ΔF). Several recent papers assume that 'tipping point' only applies to (**a**) and use reversibility to rule out the existence of a tipping point. However, a formal definition (Lenton et al. 2008) includes several other types of tipping point, including reversible ones (**b**), all of which could have significant societal impacts

fashion, and there could be several other possible types of tipping element (see the Supplementary information of Lenton et al. 2008). For example, recent work has identified examples of rate-dependent tipping; where a system undergoes a large and rapid change, but only when the rate at which it is forced exceeds a critical value (Levermann and Born 2007; Wieczorek et al. 2011). Unforced internal climate variability, which is relatively large in the Arctic region, can also trigger abrupt changes in Arctic systems. These can be described as noise-induced transitions (e.g. Fig. 1c) between different system states (or attractors) (Lenton 2011c).

PAST ABRUPT CLIMATE CHANGES IN THE ARCTIC

The Arctic has experienced a range of abrupt climate changes on different timescales in the past. The Arctic Ocean became fully ventilated with oxygen with the opening of the Fram Strait around 17.5 Ma (million years ago), coinciding with the middle Miocene climatic optimum (Jakobsson et al. 2007). In the last, Eemian interglacial around 125 ka (thousand years ago), the Arctic climate was warmed by an 11–13% increase in summer insolation, which caused a seasonal loss of Arctic sea-ice, northward advance of tree lines on land, and a substantial shrinkage (though not a total collapse) of the GIS (Brigham-Grette 2009).

During the last ice-age, a series of abrupt climate change events occurred, known after their discoverers as Dansgaard-Oeschger events (or 'DO events' for short). These remarkable events were characterised by abrupt warming of order 5°C within decades in Greenland, into 'inter-stadial' conditions, then some gradual cooling, followed by a more rapid switch back into cold 'stadial' conditions. There were over twenty such events during the last ice age. Some have argued for a periodic ~1500 year recurrence of DO events, suggesting that they represent a stochastic resonance in response to fluctuations in solar activity (Alley et al. 2001; Ganopolski and Rahmstorf 2002). However, the null hypothesis that the DO events are purely noise-induced transitions (Fig. 1c) triggered by stochastic fluctuations in the climate system (with no regular timing) cannot be rejected (Ditlevsen et al. 2005).

At the end of the last ice age, as the ice sheets in the Northern and Southern Hemispheres began to shrink, there was an abrupt warming (14.7 ka) into the Bølling–Allerød inter-stadial period (also known as DO event 1) (Steffensen et al. 2008). This Bølling warming was most likely a large response to a fairly large triggering perturbation, which altered the Atlantic THC, coupled to the atmosphere and sea-ice (Weaver et al. 2000; Liu et al. 2009). The Bølling and Allerød periods themselves were interspersed with cooling events, and ended with a more marked and rapid cooling into the Younger Dryas (12.8 ka). This cool interval persisted for over a thousand years before ending (11.5 ka) in a very abrupt warming event of ~7°C in a few years (Steffensen et al. 2008). This final warming into the Holocene may have involved passing a bifurcation in the climate system (Fig. 1a) in which the 'cold' mode of the North Atlantic ocean–atmosphere–sea-ice system finally lost its stability.

Subsequently, at ~8.2 ka a large pulse of meltwater caused a temporary weakening of the Atlantic THC, but it recovered rapidly. Paradoxically this event coincided with the onset of deep water formation in the Labrador Sea region, and recent work suggests that the freshwater perturbation triggered a switch between stable modes of operation of the subpolar gyre (Born and Levermann 2010). Meanwhile on land, the start of the Holocene was characterised by a switch from steppe grassland to tundra in Siberia with the loss of megafauna, which has been interpreted as a switch between alternative stable states, perhaps triggered by human hunting activity (Zimov et al. 1995; Chapin et al. 2004). Later, black spruce colonised interior Alaska in a transition that appears to have involved self-amplifying increases in fire frequency (despite the climate getting wetter at the time) (Chapin et al. 2004).

The Arctic has seen several subsequent abrupt regional cooling events during the Holocene, which have been linked to shutdown in the inflow of warm Atlantic waters into the shallow Barents Sea (Semenov et al. 2009). These events involved a self-amplifying process in which increased sea-ice cover altered atmospheric wind patterns in a way that reinforced the reduction in inflow and the build up of the sea-ice (Bengtsson et al. 2004; Semenov et al. 2009). Then in the last century, an abrupt Arctic warming occurred in the 1920s, which persisted until the 1940s, with up to 1.7°C warming across 60–90°N at its peak in the 1930s (Alley et al. 2003). This warming has been linked to changes in the opposite direction in the Barents Sea, this time increased inflow of warm Atlantic water, sea-ice retreat, and wind changes that were self-amplifying (Bengtsson et al. 2004). The warming occurred during an interval of strong regional anthropogenic climate forcing due to deposition of black carbon (soot) on Arctic snow (McConnell et al. 2007), so perhaps a role for anthropogenic forcing in triggering this event should be reconsidered.

This past climate evidence provides useful clues as to where to look for potential future Arctic tipping points (Fig. 2). Paleo-data shows the potential for summer Arctic sea-ice loss, large shrinkage of the GIS, and major changes in ecosystems on land, under a warmer Arctic climate than today. It also shows the potential for abrupt transitions in the Atlantic THC, coupled to sea-ice and atmospheric

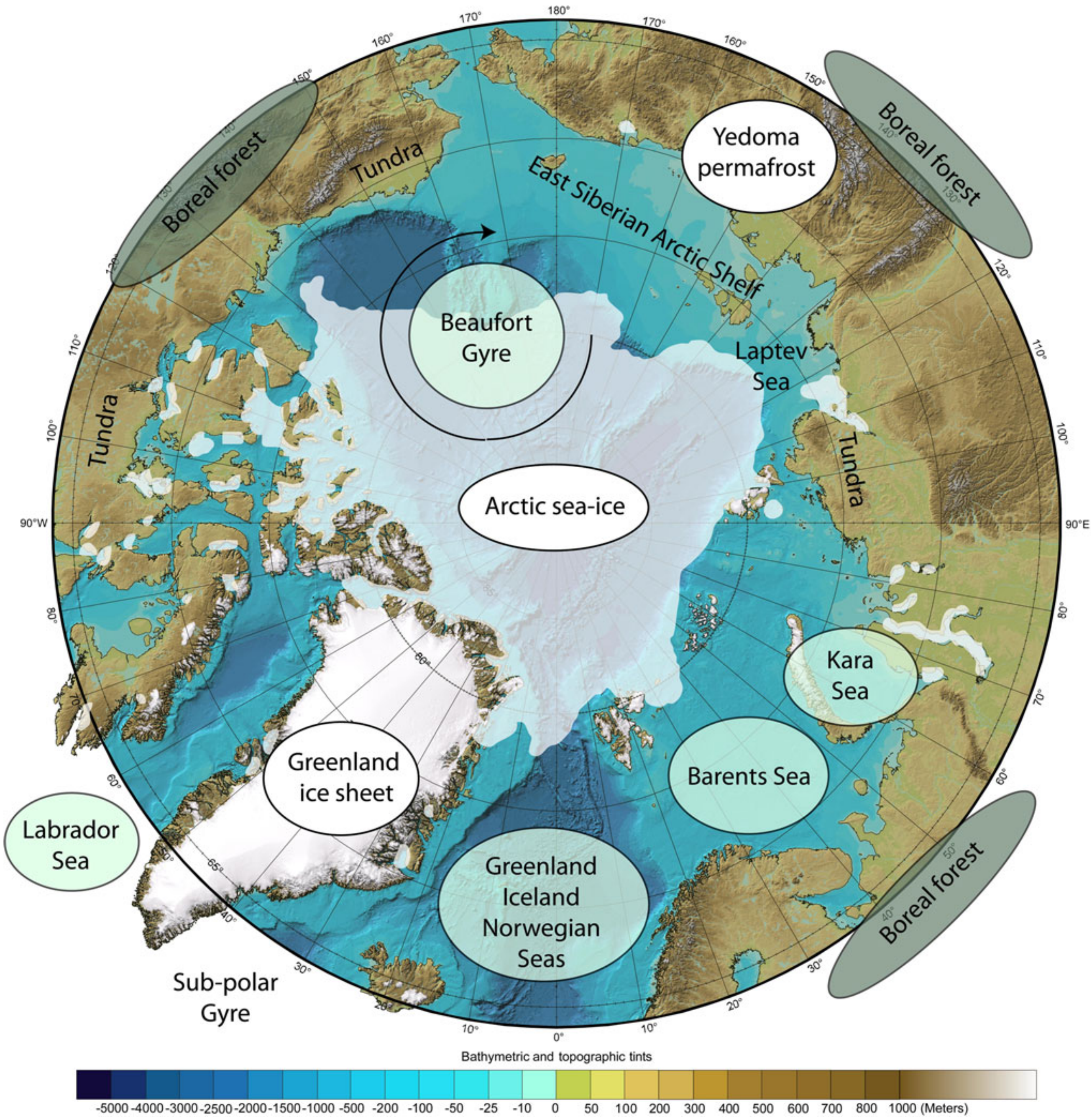

Fig. 2 Map of Arctic climate tipping elements. Based on the International Bathymetric Chart of the Arctic Ocean (IBCAO) with land topography, and the September 2008 minimum sea-ice extent overlain. Systems ringed are tipping elements suggested herein or elsewhere in this special issue, other labels are to help guide the reader (systems discussed herein). Tipping elements are *colour coded*; *white* ice melting, *aqua green* changes in ocean circulation (often coupled to sea-ice/atmospheric circulation), *dark green* involves biome change

circulation, and for smaller scale shifts in the sub-polar gyre (including the Labrador Sea), and in the Barents Sea. However, past abrupt changes generally occurred under different forcing conditions than today (e.g. different orbital configuration or ice sheet volume). Hence, they are not direct analogues for what may happen in future.

RECENT NON-LINEAR ARCTIC CLIMATE CHANGES

During the last 2000 years the Arctic region was cooling thanks to changes in the Earth's orbit, but that trend has now been reversed (Kaufman et al. 2009) and we are in an unprecedented situation where anthropogenic activities are clearly driving Arctic warming (Gillett et al. 2008). Present observations may thus provide some indication of which Arctic climate systems could be vulnerable to future tipping point changes, forced by human activity (Fig. 2). We should look, in particular, for external forcing being amplified by positive feedbacks within parts of the Arctic climate system.

Sea-Ice

The summer minimum area cover of Arctic sea-ice has declined markedly in recent decades, most strikingly in 2007, followed by the second lowest areal coverage in 2008 (Fig. 2), fourth lowest in 2009, and third lowest in 2010. Observations have fallen below all IPCC model projections, despite the models having been in agreement with the observations in the 1970s (Stroeve et al. 2007). Winter sea-ice is also declining in area (though less rapidly), with a loss of 1.5 million km^2 of multi-year ice coverage over 1997–2007 (Nghiem et al. 2007). There is an overall, progressive thinning of the ice cap, with observations showing a decrease of mean winter multi-year ice thickness from 3.6 to 1.9 m over the past three decades (Kwok and Rothrock 2009; Wadhams 2012 [this issue]).

The observed decline in sea-ice is consistent with self-amplification due to the ice–albedo positive feedback, as exposure of the dark ocean surface causes increased absorption of solar radiation. This is warming the upper ocean and contributing significantly to melting on the bottom of the sea-ice. Over 1979–2007, 85% of the Arctic region has received an increase in solar heat input at the surface, with an increase of 5% per year in some regions including the Beaufort Sea (Perovich et al. 2007), where there was a three times greater bottom ice melt in 2007 compared to earlier years (Perovich et al. 2008). Warming of the lower atmosphere in the Arctic is being significantly amplified by sea-ice melt (Screen and Simmonds 2010), and also by a shift from snow to rainfall which lowers the albedo of the remaining ice cover (Screen and Simmonds 2011).

Ice loss has also been encouraged by patterns of atmospheric circulation (Rigor and Wallace 2004; Maslanik et al. 2007) and ocean circulation (Nghiem et al. 2007), that exported multi-year ice out of the Arctic basin through the Fram Strait. Increased input of ocean heat from the Pacific (Shimada et al. 2006; Woodgate et al. 2006) and the Atlantic (Spielhagen et al. 2011) are also contributing to ice melt. Finally, reductions in summertime cloud cover (the only season in which clouds have a net cooling effect in the Arctic) may have contributed to record sea-ice retreat in 2007 (Kay et al. 2008).

Atmospheric Circulation

Recently, radical shifts in atmospheric circulation patterns have occurred in the Arctic (Zhang et al. 2008). The centre of action of the Arctic Oscillation/North Atlantic Oscillation (AO/NAO) that conventionally sits over Iceland shifted northeast into the Barents Sea. Then, the winter AO/NAO pattern of pressure weakened and was replaced in the early 2000s by a dipolar pattern, in which (in its negative phase) a Eurasian Arctic coastal high contrasted with a North Pacific low. This 'Arctic Rapid change Pattern' (ARP), in its negative phase through 2000–2006, strengthened poleward heat transport and brought warm air and warm ocean currents from the Atlantic right into the centre of the Arctic, limiting winter sea-ice growth and thus contributing to summer sea-ice decline (Zhang et al. 2008). The strength of summer storm cyclones in the Arctic basin also increased and contributed to sea-ice decline (Simmonds and Keay 2009). The loss of sea-ice cover in turn began to feed back to change large-scale atmospheric circulation (Overland and Wang 2010). Loss of winter sea-ice cover in the Barents–Kara Seas has been linked to recent anomalously cold winters (e.g. 2005/2006) over northern continents (Petoukhov and Semenov 2010). In 2006–2007 the ARP switched to a positive phase, helping sweep sea-ice towards the warmed Atlantic side of the Arctic and out of the basin, thus contributing to the record decline in summer 2007. In winter 2009/2010, an extreme negative phase of the Arctic Oscillation reasserted itself and should have favoured retention of sea ice through the 2010 melt season, but still the September 2010 sea ice extent fell to a low level (Stroeve et al. 2011).

Ocean Circulation

The Atlantic waters that enter the Arctic Ocean at depth are now unusually warm and are enhancing heat flux to the surface and potentially contributing to sea-ice melt (Spielhagen et al. 2011). The Barents Sea has warmed considerably at depth, tracking a strong phase of the Atlantic Multi-decadal Oscillation (Levitus et al. 2009). Heat input to the Barents Sea has also been enhanced by the radical changes in atmospheric pressure patterns and hence wind forcing (Zhang et al. 2008).

On the other side of the Arctic basin, the Beaufort Gyre is argued to play a key role in Arctic climate variability, storing up freshwater in the dominant anti-cyclonic climate regime (Arctic high), but releasing it to the North Atlantic during shifts to a cyclonic climate regime (Proshutinsky et al. 2002). Recent observations show progressive accumulation of freshwater in the Beaufort Gyre at present, and since the 1950s (Proshutinsky et al. 2009). An interval of freshwater release from the Beaufort Gyre may have caused the Great Salinity Anomaly of the North Atlantic in the 1970s (Proshutinsky et al. 2002), which in turn is associated with a sudden 0.3°C cooling of Northern Hemisphere sea surface temperatures around 1970, centred in the Northernmost North Atlantic (Thompson et al. 2010). Subsequently, in the North Atlantic, the absence of deep convection in the sub-polar gyre for over a decade from the mid-1990s to mid-2000s raised concerns that ocean circulation was already being affected by climate warming. However, there was an abrupt resumption of deep convection in the Labrador and Irminger Seas in winter 2007–2008 (Vage et al. 2009; Yashayaev and Loder 2009).

Greenland Ice Sheet

The GIS is currently losing mass at a rate that has been accelerating (Rignot et al. 2007). In summer 2007, there was an unprecedented increase in surface melt, mostly south of 70°N and also up the west side of Greenland, due to an up to 50-day longer melt season than average with an earlier start (Mote 2007). This is part of a longer-term trend of increasing melt extent since the 1970s. Recent observations show that seasonal surface melt has led to accelerated glacier flow (Joughin et al. 2008; van de Wal et al. 2008). The surface mass balance of the GIS is still positive (there is more incoming snowfall than melt at the surface, on an annual average), but the overall mass balance of the GIS is negative due to an increased loss flux from calving of glaciers that outweighs the positive surface mass balance. The margins of the GIS are thinning at all latitudes (Pritchard et al. 2009), and the rapid retreat of calving glaciers terminating in the ocean, most notably Jakobshavn Isbrae, is probably linked to warming ocean waters (Holland et al. 2008).

Shelf Seas

Unusually warm Atlantic waters are also intruding onto the shallow Laptev Sea (Dmitrenko et al. 2010; Spielhagen et al. 2011). This is part of the East Siberian Arctic Shelf, which is a large area (2.1 million km^2) of submerged permafrost and peatland that was flooded during the early Holocene (Shakhova et al. 2010). Much of the water there is supersaturated with methane coming out of the submerged sediments and the region is venting $\sim$8 TgC-CH_4 year^{-1} to the atmosphere (Shakhova et al. 2010). This is only around 2% of the global methane source to the atmosphere, but if the intrusion of warm Atlantic waters increases this could further destabilise the submerged permafrost (Dmitrenko et al. 2010).

Land Surfaces

Arctic land surfaces are already experiencing strongly amplified warming, partly linked to shrinkage of the Arctic sea-ice (Lawrence et al. 2008), and to lengthening of the snow-free season (Chapin et al. 2005). Around the record minimum of Arctic sea-ice cover in August–October 2007, Arctic land temperatures jumped $\sim$3°C above the mean for the preceding 30 years (Lawrence et al. 2008). There was also record discharge of Eurasian rivers draining into the Arctic in 2007, linked to extreme changes in atmospheric circulation (Rawlins et al. 2009). The overall warming trend is driving thawing of continuous permafrost, which is the perennially frozen soil that currently covers $\sim$10.5 million km^2 of the Arctic land surfaces. This thawing is rapid in some regions, e.g. Northern Alaska where extensive thermokast lakes have formed (Jorgenson et al. 2006). Lengthening of the snow-free season is encouraging shrub growth in the tundra (Chapin et al. 2005), and also greening of the boreal forest further south (Lucht et al. 2002). In western Canada, the boreal forest is suffering from an invasion of mountain pine beetle that is linked to climate warming (Kurz et al. 2008a). This has caused widespread tree mortality and has turned the nation's forests from a carbon sink to a carbon source (Kurz et al. 2008b). Fire frequencies have also been increasing across the boreal forest zone.

POTENTIAL FUTURE ARCTIC CLIMATE TIPPING POINTS

As well as past and present behaviour, model projections can help identify potential future Arctic tipping points (Fig. 2), although one should beware that state-of-the-art climate models have manifestly failed to reproduce past periods of warm Arctic climate (Valdes 2011). Hence, we should also try and identify underlying positive feedback mechanisms that are strong enough to generate tipping point behaviour. Here question marks are used to denote those systems, where the existence of a tipping point is most uncertain. Where a tipping point has been identified, and information is available on its proximity in terms of temperature change, this is summarised in Fig. 3.

Summer Sea-Ice Loss

The Arctic may already be committed to a qualitative change in which the ocean becomes largely ice-free in summer, with projections for when this will happen starting at 2016 ± 3 (W. Maslowski, personal communication) but most model estimates starting around 2050 (Holland et al.

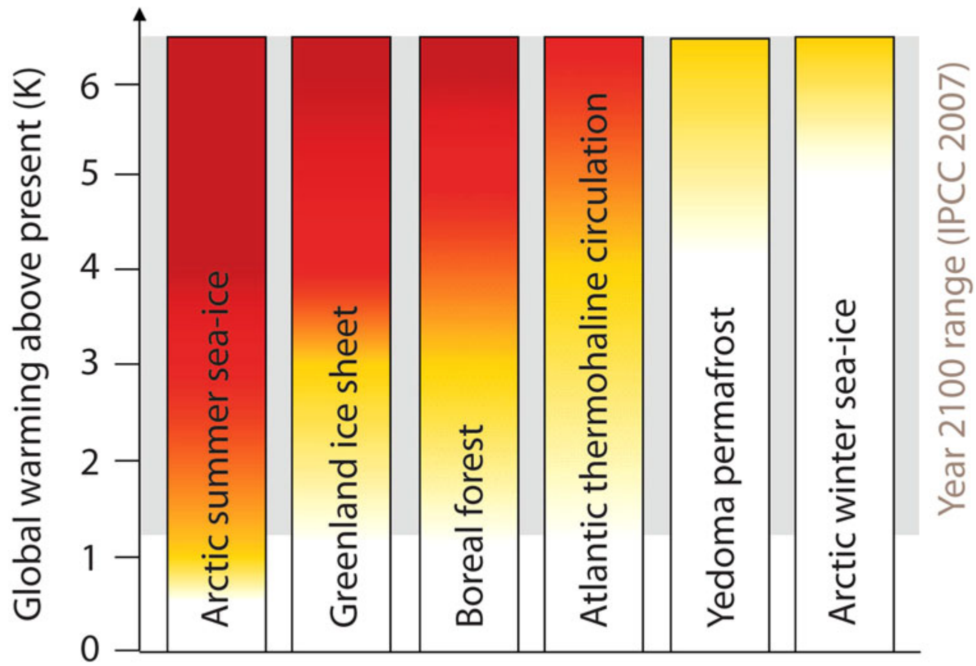

Fig. 3 Proximity of different Arctic climate tipping points. The 'burning embers' capture estimates of the increasing likelihood of passing a tipping point as global temperature increases (above the 1980–1999 mean), and the associated uncertainty. For each tipping point, the transition from *white to yellow* indicates that its likelihood becomes non-zero, the transition from *yellow to red* indicates that it becomes as likely as not (best estimate of threshold location), and the appearance of *dark red* indicates it has become more likely than not. Likelihood information is based on review of the literature, the results of expert elicitation for the GIS, boreal forest and Atlantic THC (Kriegler et al. 2009), and interpretation by the present author. Only Arctic tipping elements for which there is some quantitative information linking a tipping point to global temperature are shown

2006; Boe et al. 2009; Tietsche et al. 2011) when global warming is around 2°C (Fig. 3). The year that the North Pole becomes seasonally ice-free will likely be seen as a 'tipping point' by non-experts. Whilst politically important, several authors argue it is unlikely that such a transition involves an irreversible bifurcation (Eisenman and Wettlaufer 2009; Tietsche et al. 2011) (as in Fig. 1a). Summer sea-ice cover can recover quickly in models when the climate cools into the following winter, because thin ice grows more rapidly (Notz 2009), and with diminished ice-cover, excess heat is more rapidly transferred to the atmosphere and radiated to space (Tietsche et al. 2011) (both are negative feedbacks). However, if cloud cover increases after summer sea-ice loss, then this could act as an insulating blanket in autumn–winter restricting sea-ice recovery (positive feedback) and potentially creating multiple stable states (Abbot et al. 2011). State-of-the-art climate models still struggle to adequately represent e.g. cloud feedbacks, hence using such models to rule out tipping point behaviour (Tietsche et al. 2011), may be placing too much faith in them. More fundamentally, a 'tipping point' need not involve an irreversible bifurcation (Lenton et al. 2008) (e.g. Fig. 1b). Some models show that as the ice cap gets thinner, it becomes prone to larger fluctuations in area, which can be triggered by relatively small changes in forcing (Holland et al. 2006). This could represent 'tipping point' behaviour, even if the changes are readily reversible. Certainly such large ice loss events can have significant impacts, as was seen in 2007.

Year-Round Sea-Ice Loss

Loss of winter (i.e. year-round) ice is more likely to represent a bifurcation (Fig. 1a), where the system can switch rapidly and irreversibly from one state (with seasonal ice) to another (without any) (Eisenman and Wettlaufer 2009). However, the tipping point for year-round ice loss requires around 13°C warming at the North Pole (Winton 2006). This could only occur this century under high anthropogenic emissions scenarios with fairly strong polar amplification of warming (Fig. 3).

Greenland Ice Sheet Irreversible Meltdown

The GIS will be committed to irreversible meltdown if the surface mass balance goes negative, most notably because as the altitude of the surface declines it gets warmer (a positive feedback). Initial assessments put the temperature threshold for this to occur at around 3°C of global warming, based on a positive-degree-days model for the surface mass balance (Huybrechts and De Wolde 1999). Results from an expert elicitation concur that if global warming exceeds 4°C there is a high probability of passing the tipping point (Kriegler et al. 2009). An alternative surface energy balance model predicts a more distant threshold at around 6°C global warming (J. Bamber, personal communication). However, recent work suggests the tipping point could be much closer at 0.7–1.7°C global warming (A. Robinson and A. Ganopolski, personal communication). Figure 3 attempts to capture some of this uncertainty.

Greenland Ice Sheet Retreat onto Land

The actual threshold for massive GIS shrinkage must lie before the surface mass balance goes negative. A more nuanced possibility, which is emerging from some coupled climate–ice sheet model studies, is that there could be multiple stable states for GIS volume, and hence multiple tipping points (Ridley et al. 2010). Passing a first tipping point where the GIS retreats on to land could lead to ~15% loss of the ice sheet and about 1 m of global sea-level rise. Arguably this transition might already be underway, but there is currently insufficient information to link a tipping point to global temperature. As for the rate at which ice loss occur, an upper limit is that the GIS could contribute around 50 cm to global sea-level rise this century (Pfeffer et al. 2008).

Atlantic Thermohaline Circulation Collapse

The archetypal example of a climate surprise is a reorganization of the Atlantic THC, which is prone to collapse when sufficient freshwater enters the North Atlantic to halt density-driven deep water (NADW) formation there (Stommel 1961; Peng 1995). A hysteresis-type response (Fig. 1a) to freshwater perturbations is a characteristic, robust feature of the THC (Hofmann and Rahmstorf 2009). However, the shutdown of the THC may actually be one of the more distant tipping points (Fig. 3). Expert elicitation suggests that THC collapse becomes as likely as not with >4°C warming this century (Kriegler et al. 2009). The IPCC (2007) viewed the threshold as even more remote, but recent analysis suggests the models used were systematically biased towards a stable THC (Drijfhout et al. 2011). Although a collapse of the THC may be one of the more distant tipping points, a weakening of the THC this century is robustly predicted (IPCC 2007), which will have similar, though smaller, effects as a total collapse.

Sub-Polar Gyre Switch

A potential tipping point that occurs in some models with a weakening THC, is a switch of the sub-polar gyre in which deep convection and NADW formation shuts off in the Labrador Sea region (to the west of Greenland) and convection switches to only occurring in the Irminger Sea (near the southern tip of Greenland) and the Greenland–Iceland–Norwegian Seas (to the east of Greenland) (Levermann and Born 2007; Born and Levermann 2010). However, there is currently insufficient information to link this potential tipping point to global temperature.

Barents–Kara Seas Switch

Past and present observations indicate that abrupt shifts can occur in the ocean–atmosphere-sea ice system in the Barents–Kara Seas (Bengtsson et al. 2004; Semenov et al. 2009). A shift to a state like that hypothesised for the 1920s abrupt warming seems feasible, in which much reduced year-round ice cover causes a strong winter ocean–atmosphere heat flux, which in turn supports a locally cyclonic circulation, enhancing westerly winds that support the inflow of warm ocean waters, in a self-amplifying positive feedback loop (Bengtsson et al. 2004). Arguably such a transition might already be underway, but there is currently insufficient information to link a tipping point to global temperature. Such a shift would greatly warm this part of the Arctic, but perversely it could also cause anomalously cold winters in European mid-latitudes, if loss of regional year-round ice supports a larger-scale anti-cyclonic circulation around the pole that extends down to mid-latitude easterlies over the continents (Petoukhov and Semenov 2010).

Ocean Methane Release Events?

Recent model estimates suggest that 1600–2000 PgC are stored globally in methane hydrates and the gas bubbles they trap beneath the ocean floor, of which ~250 PgC is in the Arctic Ocean basin (Archer et al. 2009). As the ocean warms, heat diffuses into the sediment layer and may destabilize this reservoir of frozen methane. Bubbles associated with the melting of methane may trigger submarine landslides (Kayen and Lee 1991), and this raised concern that the destabilization of methane hydrates could result in abrupt and massive releases of methane into the atmosphere. The most vulnerable methane hydrate deposits are shallow ones, such as those found in the Arctic (Reagan and Moridis 2007), including under the East Siberian Arctic Shelf (Shakhova et al. 2010). However, estimates that this region could produce an abrupt release of ~12 times the current atmospheric methane burden (Shakhova et al. 2008), are completely at odds with state-of-the-art models (Archer et al. 2009) and palaeo-climatic evidence (Archer 2007). Instead, the most likely scenario is a long-term chronic methane source made up of many small events (Archer et al. 2009).

Yedoma Permafrost Collapse

Permafrost area could be reduced to as little as 1.0 million km^2 by the year 2100, which would represent a qualitative change in state (Lawrence and Slater 2005). However, permafrost as a whole is not on the shortlist of tipping elements (Lenton et al. 2008) because of a lack of evidence for a large-scale tipping point for permafrost melt. Instead, in future projections the local threshold of freezing temperatures is exceeded at different times in different localities. Yet more recent work has suggested that at least one large area of permafrost could exhibit coherent threshold behaviour. The frozen loess (windblown organic material) of northeastern Siberia (150–168°E and 63–70°N), also known as Yedoma, is deep (up to 25 m) and has an extremely high carbon content (2–5%); thus it may contain ~500 PgC (Zimov et al. 2006). This regional frozen carbon store could undergo self-sustaining collapse, due to an internally generated source of heat released by biochemical decomposition of the carbon, triggering further melting in a runaway positive feedback (Khvorostyanov et al. 2008a; Khvorostyanov et al. 2008b). Once underway, this process could release 2.0–2.8 PgC year^{-1} (mostly as CO_2 but with some methane) over about a century, removing ~75% of the initial carbon stock. The collapse would be irreversible in the sense that removing the forcing would not stop it continuing (Fig. 1a). To pass the tipping point requires an estimated >9°C of regional warming (Khvorostyanov et al. 2008a, b), which may be accessible this century under high emissions scenarios (Fig. 3).

Tundra Loss?

A warmer future climate should enable northward expansion of the boreal forest into tundra regions (Scholze et al. 2006; Sitch et al. 2008). This typically occurs when regions exceed 1000 growing degree days (GDD) above zero, and it initiates a positive feedback whereby the trees obscure snow thus amplifying warming, as happened in the early Holocene (Foley et al. 1994). However, models suggest the transition from tundra to boreal forest will be a continuous process without strong nonlinearity or tipping point behaviour (Joos et al. 2001; Lucht et al. 2006; Schaphoff et al. 2006).

Boreal Forest Dieback

In the future, widespread dieback of the boreal forest has been predicted in at least one model, when regional temperatures reach around 7°C above present, corresponding to around 3°C global warming (Fig. 3). Expert elicitation concurs that above 4°C global warming dieback becomes more likely than not (Supplementary Information of Kriegler et al., 2009). Increasingly warm summers becoming too hot for the currently dominant tree species, increased vulnerability to disease, decreased reproduction rates, and more frequent fires causing significantly higher mortality, all contribute. The forest would be replaced over large areas by open woodlands or grasslands, which would in turn amplify summer warming and drying and increase fire frequency, producing a potentially strong positive feedback.

Arctic Ozone Loss?

Although the Antarctic ozone hole was tipped by human activity, it is widely believed that the stratospheric ozone layer has now been saved by the Montreal protocol. However, the Arctic could face a climate change-induced ozone hole (Shindell et al. 1998; Austin et al. 2003). Global warming implies global cooling of the stratosphere that supports formation of ice clouds, which in turn provide a catalyst for stratospheric ozone destruction. Furthermore, there exists a strong coupling between the troposphere and the stratosphere in the Northern Annular Mode (NAM) and strong synergistic interactions between stratospheric ozone depletion and greenhouse warming are possible (Hartmann et al. 2000). Whether a tipping point exists is unclear, and beyond 2060 it should become impossible thanks to reductions in ozone depleting gases (Levermann et al. 2011).

DISCUSSION: RISKS AND RESPONSES

The revised list of potential Arctic climate tipping elements are; Arctic sea-ice, the GIS, the Atlantic THC, the subpolar gyre (determining the location of North Atlantic Deep Water formation), the Barents–Kara Seas, the Yedoma region of permafrost and the boreal forest (Fig. 2). As noted above, some of these systems could exhibit more than one tipping point (e.g. Arctic sea-ice and the GIS). The list is not intended to be comprehensive, and others have suggested additional Arctic tipping elements, e.g. the Beaufort Gyre (Carmack et al. 2012 [this issue]). Hopefully critical reflection will demote others. The intention is to provide a starting point for an ongoing risk assessment of Arctic tipping points, which could potentially extend well beyond climate to encompass ecosystems (Duarte et al. 2012 [this issue]), and human systems.

Risk in the technical sense is the product of the likelihood of something happening and its impact (the magnitude of the potential loss). There is already some information on the likelihood of crossing Arctic climate tipping points (Fig. 3), although there is still a long way to go in correctly identifying tipping points and assessing their proximity. However, the impacts of crossing different climate tipping points have barely begun to be quantified

(Lenton 2011c). There is some literature on the global impacts of a collapse of the Atlantic THC, and the global impacts of sea level rise, including from GIS melt. In addition, a switch off of Labrador Sea deep convection would have regional, dynamic effects on sea level, increasing it by around 25 cm on the northeast coast of North America (Yin et al. 2009). The impacts of climate change on human access to the Arctic have also recently been assessed (Stephenson et al. 2011).

Despite the shortage of tipping point impact studies, it is clear that abrupt changes in Arctic climate systems are already having a tangible impact on Arctic ecosystems and communities. The question then becomes; how to respond? Broadly there are two options; try to reduce the likelihood of passing a particular tipping point, or try to reduce the impacts of crossing it. Clearly not all potential abrupt future changes in the Arctic may be avoidable, but some certainly are. To address this, it is important to know what is currently driving rapid Arctic climate change. The basic driver can be expressed in terms of anthropogenic radiative forcing (the imbalance of radiative energy fluxes at the tropopause, due to human activities). Whilst there is a great deal of focus on carbon dioxide (CO_2) in the international policy arena, importantly it does not dominate current anthropogenic radiative forcing of the Arctic region. Instead a steep increase in absorbing black carbon aerosols and a decline in reflective sulphate aerosols, together account for up to 70% of Arctic warming since 1976 (1.1 ± 0.8 of 1.5 ± 0.3°C) (Shindell and Faluvegi 2009). Also, the combined contribution of methane (CH_4) and tropospheric ozone (O_3) to Arctic radiative forcing is comparable to that of CO_2 (Hansen et al. 2007).

This mixture of forcing agents opens up avenues for mitigation policy (Lenton 2011a; Lenton 2011b). CO_2 is an extremely long-lived gas, so we can only change its concentration gradually by limiting our CO_2 emissions, and we must act globally. Methane has a shorter lifetime of around a decade, offering a more rapid response of its concentration to reducing emissions. Tropospheric ozone and black carbon have much shorter lifetimes still, such that a reduction in production translates almost instantaneously into a reduction in radiative forcing. Furthermore, particular regions of the world make a disproportionate contribution to Arctic radiative forcing from these agents. Consequently, efforts to restrict black carbon emissions through e.g. national air pollution policies and appropriate technologies, in e.g. China and India, could be a quick way to start limiting Arctic radiative forcing. The incentives (financial or otherwise) needed to help such countries protect the Arctic in this way, merit consideration. Of course CO_2 must also be globally tackled, and we should start reducing CO_2 emissions now to reduce the risk of more distant Arctic tipping points.

The other avoidance strategy is the more controversial deliberate geoengineering of Arctic climate (Zhou and Flynn 2005; Caldeira and Wood 2008). Some commentators have suggested replacing the declining blanket of cooling sulphate aerosols in the Arctic troposphere with a deliberate injection of sulphate (or other cooling aerosol) into the Arctic stratosphere (Caldeira and Wood 2008). The amount of sulphate required would be much less (as the aerosol has a much longer lifetime in the stratosphere), but the side-effects include ozone depletion (Tilmes et al. 2008). Also, returning annual mean Arctic temperatures towards e.g. pre-industrial, would presumably over-cool the summer (with potentially detrimental effects on people and ecosystems), as sunlight-scattering aerosols would have little effect in the dark Arctic winter. Geoengineered enhancement of cloud albedo in the Arctic has also been suggested (S. Salter, personal communication), but this too would be biased to cooling the summer, as for most of the year, clouds are net warming agents in the Arctic. Hence, it is unclear that anyone has yet proposed an Arctic geoengineering strategy that actually reduces overall risk. A constructive way forward would be to engage the people of the Arctic region in the geoengineering debate through the Arctic Council (Egede-Nissen and Venema 2009).

For those Arctic climate tipping points that cannot be avoided, the challenge becomes one of adaptation, or building resilience to reduce their impacts. To help with this, early warning of some approaching climate tipping points may be achievable (Lenton 2011c), which could at least buy some time for those likely to be impacted.

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