Climate Tipping Points

James Price

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1 Introduction

The concept of a “tipping point” refers to a critical threshold at which even a small perturbation triggers a significant response causing a system to move from one equilibrium state to another. A glass on the verge of tipping over provides a good visualisation of the issue at hand. When held at any angle below the critical value and released, the glass tends to move back to a stable upright position. However, when balanced at the required threshold it is extremely sensitive to even the lightest touch and a small force applied in the correction direction results in a rapid transition to a new stable state. This concept can also be applied to the climate system, where a small change in a driving parameter such as mean global temperature can lead to a disproportionate response from a number of components within the system. The new equilibrium state, or even the transition to that state, of such components tends to result in large scale impacts on the earth system which may manifest, for instance, as significant sea level rise, an increased incidence of extreme weather events and widespread forest dieback. Therefore the assessment (Lenton et al., 2008, hereafter L08), or reassessment (Smith et al., 2009), of likely so-called tipping elements, their respective thresholds and probabilities (Kriegler et al., 2009), and possible prediction or early warning of their arrival (Scheffer et al., 2009; Lenton, 2011) have recently received much discussion in the literature.

In the following sections this work shall first review which tipping elements have been identified in the climate system and then go on to discuss the current state of our ability to provide early warnings of their onset as global temperatures continue to rise and we approach their respective predicted thresholds. The tipping elements focused on in this work are those discussed in L08 and therefore follow their definition of being potentially policy-relevant. That is, each element meets the inequality given in their point 1, is at least subcontinental in scale (∼ 1000 km) and is considered as policy-relevant if:

1. Decisions taken within a political time scale, ∼ 100 years, can control whether a given element reaches its critical tipping threshold. L08 note that the Inter-governmental Panel on Climate Change (IPCC) also focus on this timescale.
2. The time scale to observe a noticeable change is not too long term, $\lesssim 1000$ years. As such the potential impact of a given element passing its tipping point is not too disconnected from today's civilisation.

3. The likely outcome of passing a tipping point has a substantial impact on the earth system and particularly on the quality of human life.

Thus, this work covers a range of tipping elements from the abrupt, those that show observable impacts well before 2100, to intermediate, those on time scales of $\sim 100$ years, to slow, only detectable if observed over $\sim 1000$ years.

It is important to note that the critical thresholds of tipping elements given in the following section are derived from some combination of predictive models, paleo-climatic and historical data and as such will always be subject to some degree of uncertainty. Threshold ranges are given with respect to the 1980-1999 global mean temperature, as in L08, unless otherwise stated. Those from Kriegler et al. (2009) are relative to the year 2000 and have been shifted by $\sim +0.2^\circ C$\footnote{http://www.cru.uea.ac.uk/cru/data/temperature/} to be baselined to 1980-1999 levels.

## 2 Tipping Elements

### 2.1 Arctic Sea Ice

**Mechanism:** Measurements of Arctic surface air temperature from a variety of data sources, buoys, Arctic stations and satellite imagery, document an upward trend (Rigor et al., 2000; Comiso, 2003; Serreze et al., 2007) which has been accompanied by a sustained decline in sea ice extent and thickness (Stroeve et al., 2007; Kwok and Rothrock, 2009), particularly in the summer. Sea ice melt is accelerated by increased exposure of the darker ocean surface which absorbs more of the incoming solar radiation, in turn triggering further warming and establishing a positive feedback loop. Observed thinning/retreat occurs in theory due to some combination of surface melt from warmer air temperatures (Comiso, 2003), increase in the length of the melt season (Laxon et al., 2003), less multi-year ice (Kwok et al., 2009) and bottom melt, which itself may be further exacerbated by increased ocean heat transport to the Arctic (Holland and Bitz, 2003).

**Current Observations:** Up-to-date observations of Arctic sea ice continue to show a decline in extent of the annual September minimum, currently computed to be $-12.8 \pm 1.6\%$ per decade in the period 1979-2010 (Perovich, 2011). Peak winter sea ice is also seen to be decreasing in extent (Serreze et al., 2007), although less rapidly than the summer coverage. According to the National Oceanic and Atmospheric Administration 2011 report card for the Arctic, sea ice reached a minimum of 4.33 million km$^2$ on September 9th 2011\footnote{http://www.arctic.noaa.gov/reportcard/sea_ice.html}, the second lowest minimum on record. Furthermore, average September ice thickness has seen a factor of two
decrease, 3m to 1.4m, from 1957-76 to 2003-2007 (Perovich, 2011, based on data from Kwok and Rothrock, 2009).

**Threshold:** According to L08 Summer sea ice loss has a critical threshold in the range +0.5-2°C above current, 1980-1999, global mean temperature. Measurements show that September sea ice extent is decreasing faster than forecast (Stroeve et al., 2007) by the majority of climate models used in Solomon et al. (2007), hereafter IPCC AR 4. Recent modeling work predicts a seasonal ice-free state starting from ~2030-2060 (Holland et al., 2006; Wang and Overland, 2009; Tietsche et al., 2011), although this could occur as early as 2016 (Lenton, 2012). Wang and Overland (2009) is of particular note as they used the September 2007/2008 sea ice extent as a starting point and assessed the time taken there after to reach an ice-free state, i.e. coverage < 1 million km². Indeed, the study of Wang and Overland (2009) prompted Wadhams (2012) to conclude that the Arctic will be ice-free in summer within the next 30 years.

Winton (2006) show that for the Arctic to become sea ice free all year requires a 13°C temperature rise at the North Pole or ~ +3 to 5.5°C globally, depending on model choice. Out of a large number of IPCC AR4 model runs only two warm sufficiently to be completely sea ice free in the Arctic.

**Impact:** The primary impacts from a seasonal lack of sea ice are increased warming due to the ice-albedo feedback and ecosystem change, e.g. lack of hunting grounds for polar bears. This increased warming also has a knock on effect resulting in Europe warming more than the global average (Levermann et al., 2012). Furthermore, Levermann et al. (2012) comment that sea ice retreat can influence storm tracks in the North Atlantic and may be linked to anomalously cold Eurasian winters. A secondary impact of a significantly reduced sea ice cover is that it allows easier access to the Arctic for fossil fuel exploration.

### 2.2 Greenland Ice sheet

**Mechanism:** The key diagnostic parameters for ice sheets, both those in Greenland and Antarctica, is their surface mass balance and total mass balance which are essentially a sum of inputs minus outputs (Rignot and Thomas, 2002). For the former these are accumulation, snow fall, minus melt water run off while the latter also factors in iceberg calving. In response to a warming climate, at some temperature threshold the total mass balance will become negative and result in the ice sheet contracting. As the sheet recedes, ice discharge to the ocean ceases and the total mass balance may stabilise. If warming continues eventually the surface mass balance becomes negative and significant ice sheet collapse is then likely. This occurs primarily because as sheet thickness diminishes so does surface altitude which results in an increased surface temperature and further melting, a positive feedback.

**Current Observations:** Early studies showed that the ice sheet appears to be thin-
ning preferentially around its edges (Krabill et al., 2000; Krabill et al., 2004). More recently, data obtained with both the gravity satellite GRACE (Velicogna, 2009) and from satellite based Radar (Rignot et al., 2008) show that the Greenland ice sheet is losing mass, 286 Gt/yr during 2007-2009 from the former study and 267±38 Gt/yr in 2007 in the latter work. Moreover, both studies find evidence that the mass loss has accelerated in recent years, with Velicogna (2009) reporting -30±11 Gt/yr$^2$. Even more recently, Rignot et al. (2011) found the mass loss of the Greenland ice sheet to be accelerating at a rate of 21.9±1 Gt/yr$^2$ from 1992-2009. This acceleration in mass loss has been attributed to increased ice discharge through faster flowing outlet glaciers and a sustained increase in melt water run off (van den Broeke et al., 2009).

Rignot et al. (2008) still observe the surface mass balance to be positive, however it has been shown to have been decreasing at 12.9±1 Gt/yr$^2$ since 1992 (Rignot et al., 2011).

**Threshold:** Kriegler et al. (2009) give a high probability critical value for global temperatures to be +4.2°C in order to trigger a consistent negative surface mass balance. Furthermore, they give a significant probability, i.e. ~ 50%, for the threshold to be +2.2-4.2°C. L08 predict a lower limit of 300 years for total ice sheet collapse once triggered. IPCC AR4 gives a +1.4-4.6°C global temperature change to trigger collapse, converted from relative to pre-industrial levels to the 1980-1999 mean temperature using a -0.5°C shift as suggested by the IPCC. However, ice sheet mass loss is seen to be accelerating in advance of model forecasts. Indeed, new modeling work conducted by Robinson et al. (2012) has shown that the threshold global temperature could be +0.3-2.7°C with a best estimate of +1.1°C, relative to 1980-1999, for an essentially ice free Greenland. Robinson et al. (2012) also found that the timescale for total ice sheet melt is strongly dependent on the degree of overshoot of regional temperatures above the critical threshold. They report that a 2°C sustained summer overshoot is predicted to melt the ice sheet in ~ 50,000 years while a 8°C regional summer increase is expected to result in a complete melt within 2,000 years.

It is worth noting that current warming in the region poleward of 59°N, based on data from land stations, is seen to be ~ 1.4°C relative to the 1980-1999 mean (Bekryaev et al., 2010, determined by eye from their Figure 2).

**Impact:** The primary impacts here are a 2-7m sea level rise (L08) and of course loss of ice albedo resulting in further warming. On the former, Levermann et al. (2012) note that the maximum height to which dykes can be elevated rarely exceeds 1 m and most coastlines can not be protected against a sea level rise of several metres.

Quantitatively speaking, a sea level rise of 1 metre globally is likely to have a direct impact on the lives of 108 million people while 6 metres will affect 431 million people (Rowley et al., 2007; Li et al., 2009). Dasgupta et al. (2009) studied the impact of sea level rise on 84 developing countries across the globe and found that a rise of 5 metres would affect 2.1%, some 377,930 km$^2$, of their entire agri-
cultural extent. Some regions in their sample suffered much worse than the average figure, with, for instance, countries in the Middle East and North Africa and those in East Asia and the Pacific having $\sim 12\%$ of their agricultural land impacted when subjected to a 5 metre sea level rise. Moreover, certain countries are predicted to be affected even more severely, such as Egypt which may lose $13\%$ of its agricultural land in response to just 1 metre of sea level rise.

For reference, current predictions for sea level rise by 2100 stand at 0.6-1.9m \cite{Vermeer2009, Jevrejeva2010}, with the former study giving a best estimate of $\sim 1.2$m. By 2200 global mean sea level may have risen by some 1.5-3.5m \cite{Katsman2011}, a prediction based on expert assessment rather than modeling work.

2.3 West Antarctic Ice sheet

**Mechanism:** The West Antarctic ice sheet is notably different from that found in Greenland, or indeed the East Antarctic, because it is a marine ice sheet, i.e. it largely rests on land well below sea level, with bedrock that typically slopes down in the inland direction away from its grounding line, the point at which the ice goes a float. It is this facet that led early work by Mercer (1978) to discuss a possible collapse mechanism. The idea focuses on the premise that a stable ice sheet with the features discussed above can only exist so long as its grounded portion, that resting on land, is buttressed by confined floating ice, i.e. ice shelves in an embayment. As reviewed by Joughin and Alley (2011), a strong positive feedback occurs should the confined ice shelf become diminished or removed where the grounding line rapidly retreats in the regime where the bedrock slopes down in the inland direction. This also allows the ocean to further undercut the ice causing increased bottom melt and results in ice sheet contraction, thinning and further separation from the bedrock.

**Current Observations:** While not directly inducing sea level rise, Antarctic Peninsula ice shelves have undergone rapid decline recently perhaps exemplified by that of Larsen B in 2002 following more than 10,000 years of stability \cite{Domack2005}. Following the collapse, speed up was detected in four glaciers feeding into the now collapsed section of Larsen B \cite{Scambos2004}, providing clear evidence of the buttressing role played by ice shelves \cite{Rignot2006}.

Overall, the total mass balance for the Antarctic is $\sim -250$ Gt/yr as of 2010 and is seen to be accelerating at $14.5 \pm 2$ Gt/yr\textsuperscript{2} \cite{Rignot2011}. For the West Antarctic ice sheet, mass losses are estimated to be in the range 100-200 Gt/yr \cite{Joughin2011}.

**Threshold:** L08 give a critical value for global air temperature of $+3.5\,^\circ$C, or $+5.8\,^\circ$C locally, with a smaller increase required for ocean temperatures owing to its ability to intrude under ice shelves and speed up thinning and collapse \cite{Oppenheimer2004}. Again, it is worth noting that the current annual warming level across the West Antarctic ice sheet is seen to be $\sim 0.18\,^\circ$C per decade since
the late 1950s based on the average of four data sets (Schneider et al., 2012). As such current warming above 1980-1999 levels is \( \sim 0.4^\circ C \). Transition times are predicted to be of a similar order to the Greenland ice sheet (\( \sim 300 \) years at lowest) and L08 believe that melt of the West Antarctic ice sheet is more likely to result in rapid sea level rise (\( > 1 \text{m per century} \)) than melt water from Greenland.

**Impact:** Complete collapse of the ice sheet results in a +5m sea level rise (L08). For further information on the impacts of sea level rise see the discussion in section 2.2.

### 2.4 Atlantic Thermohaline Circulation

**Mechanism:** As part of the global ocean circulation system the Atlantic thermohaline circulation moves heat and salt from the tropics to the North Atlantic, relying on temperature and salinity gradients to induce flow. It has been shown that if sufficient freshwater, from rivers or melting ice, and/or heat, from warmer ocean surface waters, are able to enter the North Atlantic then the circulatory system can become altered, with North Atlantic Deep Water formation being shut off because the required density for the water to sink is not reached (Stocker and Wright, 1991; Hofmann and Rahmstorf, 2009). The large amount of heat, \( \sim 10^{15} \text{W} \), transported by this system is a key feature of the northern hemisphere’s climate with the North Atlantic sea surface temperatures being kept some 5\(^\circ\)C warmer than comparable latitude Pacific ocean temperatures (Rahmstorf, 2002).

**Current Observations:** A study by Bryden et al. (2005) reported a slowing of the thermohaline circulation from 22.9 to 14.8 Sv (where 1 Sv = 10\(^6\) m\(^3\) per sec). However, more recent work has shown that the circulation has a year long average of 18.7\( \pm 5.6 \) Sv and therefore more data is required to determine if a real trend is present (Cunningham et al., 2007).

**Threshold:** The experts consulted during the workshop and literature review reported in L08 and Kriegler et al. (2009) give a threshold of +3-5\(^\circ\)C globally to shutdown Atlantic thermohaline circulation with a timescale of \( \sim 100 \) years. IPCC AR4 consider abrupt shutdown to be very unlikely before 2100 but did not account for growing fresh water run off from a melting Greenland ice sheet.

**Impact:** As mentioned earlier, a shut off of the Atlantic thermohaline circulation would lead to a cooling of the Northern Hemisphere, estimated to be -1.7\(^\circ\)C by recent modeling work (Vellinga and Wood, 2008), and result in substantial climate alteration such as significant drying and reduced precipitation in Europe (Levermann et al., 2012). It would also cause sea level rise and a shift in the position of the Intertropical Convergence Zone or Doldrums (L08). Furthermore, ocean uptake of heat and carbon dioxide could strongly reduce leading to accelerated global warming (Levermann et al., 2012).
2.5 El Niño-Southern Oscillation

Mechanism: The El Niño Southern Oscillation (ENSO) refers to a large scale coupling between the ocean and atmosphere in the Pacific basin and is associated with strong fluctuations in ocean currents, surface temperatures and precipitation. Typically the equatorial Pacific is characterised by cool waters in the east, leading to little rainfall, and warm waters in the west, leading to significant rainfall. El Niño events involve warming of the western Pacific waters and are closely linked to surface pressure changes, the Southern Oscillation, which drive variations in rainfall and winds such that the east becomes wetter and the west becomes drier than average. They occur every 3 to 7 years (IPCC AR4). The other extreme of the cycle, La Niña events, effectively result in opposite conditions, i.e. warm water being restricted to the west with more rainfall there and less rainfall over the east. The Southern Oscillation Index (SOI) is used to quantify the strength of ENSO events and is formulated by comparing the sea surface pressure at Tahiti and Darwin. ENSO is known to have global climatic effects (IPCC AR4).

Some climate models indicates that global warming could result in an increased frequency and/or amplitude of El Niño events (Timmermann et al., 1999; Guilyardi, 2006). However, the review article of Collins et al. (2010) concludes that the current generation of models provide no consistent evidence for changes in ENSO in a warming global climate.

Current Observations: Tracking ENSO variability requires sufficiently long and continuous time series and as such it is typically necessary to turn to reconstructions based on geochemical data. Wara et al. (2005) studied the Pliocene warm period, some 4.5 to 3 million years ago, which was globally 3°C warmer than at present and found permanent El Niño like conditions, although others disagree (Watanabe et al., 2011).

Threshold: L08 concluded that a global warming of +3-6°C could trigger an alteration in ENSO with a transition timescale of ~ 100 years. Kriegler et al. (2009) found that experts considered a shift toward more permanent El Niño like conditions fairly likely, 20-45% probability, should global temperatures increase by ≥ 4°C.

Impact: El Niño events result in significant impacts including drought in southeast Asia and elsewhere and flooding in western South America. It has also been linked with global climatic anomalies (Kiladis and Díaz, 1989).

2.6 Indian Summer Monsoon

Mechanism: Indian’s summer monsoon is driven by a land to ocean pressure gradient which brings rain bearing winds in from the Indian Ocean. As such, any perturbation that tends to weaken this pressure gradient may act to destabilise the monsoon (Zickfeld et al., 2005). Indeed, Zickfeld et al. (2005) found that aerosol emissions/land-use change, which have the effect of increasing planetary albedo
and inducing cooling, weaken the monsoon while global warming acts to strength it.

**Current Observations:** Anderson et al. (2002) reconstructed monsoon wind strength over the last four centuries and found that it has increased as the Northern Hemisphere warmed in agreement with theory.

**Threshold:** Zickfeld et al. (2005) predicts a collapse of the monsoon if planetary albedo in the region exceeds $\sim 0.5$. IPCC AR4 projections do not show obvious tipping point behaviour but do agree that aerosols act to reduce precipitation. In light of the fact that brown haze and land-use changes are poorly represented in the models, L08 do predict a transition to a much reduced or collapsed state in a time scale of $\sim 1$ yr based on Zickfeld et al. (2005)’s study and past apparent threshold behaviour (Gupta et al., 2003). However, it is noted that recent modeling work conducted by May (2011), which tested two possible scenarios of global temperature rise and sulphate aerosol concentrations, found monsoon precipitation to increase in both experiments. Thus, the picture regarding how the Indian monsoon will evolve in the coming decades seems somewhat uncertain.

**Impact:** Monsoon failure or transition to a much reduced state would result in drought in regions that currently rely on the monsoon (L08).

### 2.7 Sahara/Sahel and West African Monsoon

**Mechanism:** The West African monsoon, in a similar vein to that found in Indian, is powered by a land to ocean pressure gradient which manifests itself as winds blowing warm moist air over the land in summer. The region has been simulated in terms of a vegetation-atmosphere model which exhibits two stable states: a desert equilibrium with low precipitation and no vegetation and a green equilibrium with moderate precipitation and permanent vegetation cover (Brovkin et al., 1998). The present day climate favours the former conditions.

The amount of rainfall generated by the monsoon has been shown to be correlated with, and significantly controlled by, sea surface temperatures, particularly temperature gradients between hemispheres (Folland et al., 1986). That is, warmer temperatures in the Northern Hemisphere and colder temperatures in the Southern Hemisphere are positively correlated with Sahel rainfall. Therefore, perturbations that alter this gradient result in changes in the monsoon.

**Current Observations:** The Sahel rainfall index is used to track precipitation in the region and is provided online[^1]. Figure 1 shows precipitation data for the Sahel from 1900-2011 and clearly highlights the severe drought in the region from the late 1960s until relatively recently. Shanahan et al. (2009) reconstructed the variability of the West African monsoon over the past three thousand years and found that periods of severe drought lasting decades to centuries are not unusual. Therefore they conclude that the monsoon is capable of longer and more severe droughts.

Threshold: Current predictions regarding monsoon evolution over the next century differ somewhat. Held et al. (2005) forecast an anthropogenic drying trend in the region caused in part by increases in aerosol concentrations and in part by greenhouse gases. IPCC AR4 comments that whether the Sahel will become more or less wet in the future is unclear based on models at that time while Patricola and Cook (2008) predict a possible greening of the region under certain circumstances. L08 give a threshold for increased vegetation in the region of +3-5°C with a 10 year transition period.

Impact: This tipping element may result in a positive outcome with increased vegetation in the Sahara/Sahel region, although this appears to be critically model dependent and still quite uncertain.

2.8 Amazon Rainforest

Mechanism: As discussed in Zeng et al. (1996), a large fraction, ~30% although perhaps up to 50%, of precipitation in the Amazon basin is recycled and so any change in the regional climate system such that evapotranspiration is reduced will likely have a serious impact on the forest. Indeed, increased CO₂ levels tend to cause stomatal closure suppressing transpiration and amplifying warming over tropical rainforests (Cox et al., 1999). Furthermore, land-use changes alone could
potentially bring the forest to a threshold coverage level (L08).

**Current Observations:** In recent decades, Amazonia has been seen to be warming at a rate of 0.26°C per decade (Malhi and Wright, 2004). An increased frequency of dry events in southern Amazonia from 1970-1999 was reported by Li et al. (2008), although no long term trends toward drier or wetter conditions have been observed in the region since the 1920s (Marengo, 2009).

**Threshold:** L08 conclude that dieback of the rainforest would occur following +3-4°C of global warming in a timescale of ~ 50 years based on work by Cox et al. (2004). This occurs because of a more permanent El Niño-like state in the equatorial Pacific which in turn acts to reduce rainfall over the Amazon. Malhi et al. (2009) compared a set of 19 models and found that, after accounting for their significant variability with respect to observed rainfall in the 20th century, eastern Amazonia tends toward a climate more appropriate for seasonal forests. However, they find that Western Amazonia is likely to remain suitable for rainforest, except perhaps at the periphery, during the 21st century.

In the presence of both significant land-use change and climatic forcing, the review of Nepstad et al. (2008) conclude that replacement or severe degradation of more than half of the close-canopy forests in the Amazon could occur by 2030, even without further global warming.

**Impact:** L08 highlight the key impacts associated with this tipping element would be a loss of biodiversity and reduced rainfall. The model of Cox et al. (2004) also shows a reduction in carbon storage capacity in the region with Malhi et al. (2009) commenting that large scale degradation of the rainforest would leave an “enduring legacy on the functioning and diversity of the biosphere”. Nepstad et al. (2008) state that if the droughts of the last decade in the Amazon continue into the future, approximately 55% of the forests will be cleared, logged, damaged by drought or burned over the next 20 years. This would release some 15-26 Pg of carbon (PgC) to the atmosphere. Taking human induced carbon emissions for 2009 to be ~ 8.4 PgC from fossil fuels and cement and ~ 0.9 PgC from land use change (Friedlingstein et al., 2010), this is equivalent to 1.6-2.8 years of total world wide human emissions.

### 2.9 Boreal Forest

**Mechanism:** Boreal forest is a biome in the Northern Hemisphere, characterised by coniferous forests, that separates the tundra in the north from temperate forest in the south and is the worlds largest land biome. A mechanism for its collapse has been discussed by Lenton (2012) and details that increasingly warm summers become too hot for dominate tree species resulting in increased vulnerability to disease, decreased reproduction rates and more frequent fires. This in turn stimulates significantly increased mortality. The forest would be replaced by grassland or open woodlands which act to amplify summer warming and drying and so further increase fire risk, potentially a strong positive feedback.
Current Observations: A recent study by Kurz et al. (2008) showed one example of the effects a warming climate has on the boreal forest. They reported how a pine beetle outbreak, an order of magnitude larger than anything previously recorded, is decimating forest in western Canada. The outbreaks extent and severity have been associated to global warming because of an expansion of suitable habitat, allowing the beetle to penetrate further north and to higher elevations. The outbreak has converted the forest from a small net carbon sink to a large net carbon source both during and immediately after its occurrence.

Threshold: Lenton (2012) state that boreal forest dieback becomes more likely than not at +4.2°C of global warming based on probabilities from Kriegler et al. (2009), see their Supplementary Information.

Impact: L08 give biome switch, leading to a substantial alteration of the current ecosystem, as the key impact from this tipping element. Kurz et al. (2008) also mention significant economic, e.g. disrupted timber mills, and social effects, e.g. unemployment. Furthermore, the forest is likely to become a net carbon source (Kurz et al., 2008) with diminishing sink capacity.

2.10 Other Tipping Elements

L08 identified a number of other candidates tipping elements but did not include them in the bulk of their paper because each element did not meet one or more of the conditions laid out in the introduction. These elements were:

Antarctic Bottom Water: Formation of Antarctic bottom water may be shut off following a tripling of CO₂. L08 conclude that more modeling studies need to be carried out to assess the possibility, and if so, threshold of such a feature.

Tundra: Encroachment of boreal forest into the tundra may setup a positive feedback such that surface albedo is lowered leading to increased warming. However, L08 conclude that it is likely to be a continuous transition and without significant threshold behaviour and hence not a tipping element.

Permafrost: L08 concluded that while melting permafrost accompanied by significant methane release could set up a positive feedback, no studies demonstrated that it is a tipping element by their definition. However, more recent work has found evidence for threshold behaviour in at least one large area of permafrost in northern Siberia. In this area, also called Yedoma, heat released during biochemical decomposition of the carbon acts to reinforce the melting process and establish a positive feedback which could release 2-2.8 PgC per year (Khvorostyanov et al., 2008a,b). Triggering of this tipping element would require a warming of > +9°C in the region, which may be accessible under a high emissions scenario (Lenton, 2012).

Marine Methane Hydrates: Despite large methane hydrate reserves under marine continental shelf and slope sediment, L08 did not classify it a tipping element because of long transition timescales, ~ 1000 to 100,000 years. Indeed,
Lenton (2012) agrees and states that the most likely scenario is a long term chronic methane source made up of many small events.

**Ocean Anoxia:** L08 discuss how increased phosphorus input into the oceans could trigger anoxia, or in other words the extreme depletion of oxygen in the oceans. However, a long transition timescale, $\sim 10,000$ years, rules this candidate out.

**Arctic Ozone:** While the Antarctic ozone hole is a tipping element that has subsequently been stabilised, Europe could face a climate change fueled ozone hole (Lenton, 2012). However, both L08 and Lenton (2012) conclude that whether there is a tipping point is unclear. Furthermore, beyond 2060 such a hole should become impossible due to a reduction in ozone depleting gases (Levermann et al., 2012).

**Mountain Glaciers:** In addition, one further candidate tipping element not included or reviewed by L08 but discussed by Levermann et al. (2012) is the melting/retreat of mountain glaciers, with particular attention given in their study to European glaciers. They cite several mechanisms that serve to amplify glacier retreat such as ice-albedo feedback, increased dust accumulation decreasing ice albedo and a longer melt season experienced by glaciers in the Alps. Recent measurements show the volume of glaciers in the European Alps has decreased from $\sim 200$ km$^3$ in 1850 to $\sim 65$ km$^3$ in 2005 (Haeberli et al., 2007). Globally glacier tongues are seen to be in retreat (see IPCC AR4, Chapter 4, page 357). Levermann et al. (2012) give an increase of $+2^\circ$C in global mean temperature for a complete loss of glacier ice volume in the Alps. Ramanathan and Feng (2008) propose that Hindu-Kush-Himalaya-Tibetan (HKHT) glaciers be considered a tipping element with a threshold in the range of $+1-3^\circ$C, relative to pre-industrial levels. The impacts of melting glaciers include sea-level rise, ground destabilisation due to melting permafrost and a reduction in tourism.

However, it is still unclear as to whether HKHT glaciers will respond in a strongly non-linear way to global warming (Henderson-Sellers and McGuffie, 2012, chapter 17, page 495) and as such further research needs to be conducted to confirmed whether they are, in fact, a tipping element and if so what their threshold is. Given this uncertainty, mountain glaciers are not included in the list of policy relevant tipping elements.

### 3 Prediction and Early Warnings

As discussed by Lenton (2011), the current generation of climate models are somewhat inadequate when it comes to diagnosing our proximity to the threshold of any given tipping element. While these issues are being worked on, interest has recently grown in using statistical techniques for the task at hand (Livina and Lenton, 2007; Dakos et al., 2008; Scheffer et al., 2009; Ditlevsen and Johnsen, 2010; Lenton et al., 2012).

The trigger that forces a tipping element through its threshold is likely to be composed of some combination of inherent, short-term natural variability (or
noise) and a long term trend, in this case induced by human activity. The former, being fundamentally random, is typically more unpredictable, although not completely (Lenton, 2011), while prediction of threshold behaviour caused by the latter has seen significant recent progress. Identified signatures of an approaching transition are as follows:

**Critical Slowdown:** As a system approaches a threshold its response to small perturbations becomes more sluggish and thus it takes longer to return to its initial equilibrium state (Lenton, 2011, see their Figure 2 for a good visual representation of this idea). As a result the system in question possesses an increased “memory” of its previous state at any given moment. Techniques have been developed to take time series data of a relevant system parameter, i.e. flow rate in the case of the Atlantic thermohaline circulation, and look for this increase in sluggishness or in other words a longer decay time in returning to its average value (e.g. Dakos et al., 2008).

**Increased Variability:** In a similar fashion to critical slowdown, as a system nears its threshold a given small perturbation is able to push it further away from its current equilibrium state. In effect, its current state becomes more unstable (see again Figure 2 of Lenton, 2011). This feature manifests itself in time series data as an increased scatter (Ditlevsen and Johnsen, 2010), i.e. variance, about the usual average position of a given parameter, i.e. flow rate of the Atlantic thermohaline circulation.

**Increased Skewness:** Thirdly, when a system close to its threshold is perturbed it may spend more time, and move to a greater amplitude, away from its current average value in the direction of the impending transition (Guttal and Jayaprakash, 2008, see their Figure 1 and 5). As such, when time series data are viewed in frequency terms, i.e. the particular parameter’s value plotted against the number of times that that value occurs, the distribution becomes asymmetric and its tail is elongated, or skewed, toward the value the parameter is destined to have in the new equilibrium state.

Detecting transitions using the signatures discussed above has been tested using model and paleo data and returned encouraging but somewhat mixed results (Lenton et al., 2012). Ultimately, it is recommended that multiple techniques are applied together to improve prediction robustness.

### 4 Conclusions

Table 1 provides a summary of the policy relevant tipping elements, including Yedoma permafrost, reviewed in this work, their respective thresholds and impacts. With current global warming relative to pre-industrial levels standing at \( \sim 0.8°C \), or \( \sim 0.3°C \) with respect to the 1980-1999 mean, ice-free summers in the Arctic and, depending on the model of choice, sustained decline of the Greenland
ice sheet begin to become accessible. L08 concluded that it was these two tipping elements that represent the most sensitive to global warming with the smallest uncertainty in physical mechanism of those discussed previously. Of course it is noted that the precise threshold for triggering ice sheet decline in Greenland has become somewhat less well constrained since L08 went to press. Nevertheless, sustained ice sheet melt, both in Greenland and West Antarctica, would probably have the most severe repercussions of all the identified tipping elements and so the possibility that their thresholds are becoming accessible is of particular concern. Any number of the other tipping points reviewed previously may also surprise us by having an as yet unknown nearby tipping point.

One further point to note is the possibility of interactions between tipping elements, i.e. if one element is tipped how does this alter the probability for other tipping elements to be triggered. This topic has been discussed and schematised by recent studies such as Kriegler et al. (2009), their Figure 2, and Levermann et al. (2012), their Figure 13. One possible example of this theory would be melting of the Greenland ice sheet introducing significant amounts of freshwater into the North Atlantic which would in turn likely trigger collapse of the ocean’s thermohaline circulation, i.e. tipping of the former results in an increased probability of tipping the latter. Interestingly, triggering the collapse of the Atlantic thermohaline circulation may then cause the polar region to cool and act to stabilise ice sheet/sea-ice retreat, reducing the probability of these elements totally collapsing. Ultimately, while many links are theorised, significant uncertainties still exist and Levermann et al. (2012) point out that no model or paleo data has provided evidence that the tipping of one element can lead to the tipping of another, at least for the elements they consider.

To summarise, Figure 5 of Lenton (2011) provides a good review of the likelihood and relative impact of the main tipping elements discussed in this work. This diagram, and its impacts scale which is reproduced in Table 1, shows that the most likely are, as mentioned above, summer Arctic sea-ice loss and Greenland ice sheet melt while those with the highest impact are the collapse of the West Antarctic ice sheet, West African monsoon shift, Greenland ice sheet melt and ENSO amplitude increase. Thus it seems clear that global temperature rise must be constrained rapidly to avoid continued gambling with the climate system and its many tipping points.
<table>
<thead>
<tr>
<th>Relative Impact</th>
<th>Tipping Element</th>
<th>Control Parameter</th>
<th>Global Warming</th>
<th>Transition Timescale</th>
<th>Primary Impacts</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td></td>
<td>Pre-industrial</td>
<td>1980-1999</td>
<td></td>
</tr>
<tr>
<td>High</td>
<td>West Antarctic ice sheet collapse</td>
<td>Local air temp.</td>
<td>+3.5-5.5°C</td>
<td>+3-5°C</td>
<td>&gt; 300 yr</td>
</tr>
<tr>
<td></td>
<td>West African monsoon shift</td>
<td>Planetary albedo/Temperature gradient</td>
<td>+3.5-5.5°C</td>
<td>+3-5°C</td>
<td>10 yr</td>
</tr>
<tr>
<td>Medium-high</td>
<td>Greenland ice sheet collapse</td>
<td>Local air temp.</td>
<td>+0.8-5.1°C</td>
<td>+0.3-4.6°C</td>
<td>&gt; 300 yr</td>
</tr>
<tr>
<td></td>
<td>ENSO amplitude increase</td>
<td>Thermocline depth/sharpness</td>
<td>+3.5-6.5°C</td>
<td>+3-6°C</td>
<td>∼ 100 yr</td>
</tr>
<tr>
<td>Medium</td>
<td>Amazon Rainforest die back</td>
<td>Rainfall</td>
<td>+3.5-4.5°C</td>
<td>+3-4°C</td>
<td>∼ 50 yr</td>
</tr>
<tr>
<td></td>
<td>Atlantic thermohaline circulation collapse</td>
<td>Freshwater input</td>
<td>+3.5-5.5°C</td>
<td>+3-5°C</td>
<td>∼ 100 yr</td>
</tr>
<tr>
<td>Low-medium</td>
<td>Boreal Forest die back</td>
<td>Local air temp.</td>
<td>+3.5-5.5°C</td>
<td>+3-5°C</td>
<td>∼ 50 yr</td>
</tr>
<tr>
<td>Low</td>
<td>Arctic summer sea-ice loss</td>
<td>Local air temp.</td>
<td>+1-2.5°C</td>
<td>+0.5-2°C</td>
<td>∼ 10 yr</td>
</tr>
<tr>
<td>-</td>
<td>Indian summer monsoon collapse</td>
<td>Planetary albedo/local warming</td>
<td>Unknown</td>
<td>Unknown</td>
<td>∼ 1 yr</td>
</tr>
<tr>
<td>-</td>
<td>Yedoma Permafrost</td>
<td>Local air temp.</td>
<td>(\gtrsim +5°C)</td>
<td>(\gtrsim +4.5°C)</td>
<td>∼ 100 yr</td>
</tr>
</tbody>
</table>

* Relative impacts taken from Lenton (2011) and relative to Atlantic thermohaline circulation collapse. The relative impact of Indian summer monsoon collapse and Yedoma permafrost melt was not stated in that study and so is left blank.
† Warming figures from L08 and Lenton et al. (2012), which were given with respect to 1980-1999 mean, have been converted to pre-industrial using a +0.5°C shift where appropriate.
‡ Range updated from L08 to include new modeling work and reflect current uncertainty regarding the threshold of this element.
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