

The future of ice sheets and sea ice: Between reversible retreat and unstoppable loss

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We discuss the existence of cryospheric “tipping points” in the Earth’s climate system. Such critical thresholds have been suggested to exist for the disappearance of Arctic sea ice and the retreat of ice sheets: Once these ice masses have shrunk below an anticipated critical extent, the ice–albedo feedback might lead to the irreversible and unstoppable loss of the remaining ice. We here give an overview of our current understanding of such threshold behavior. By using conceptual arguments, we review the recent findings that such a tipping point probably does not exist for the loss of Arctic summer sea ice. Hence, in a cooler climate, sea ice could recover rapidly from the loss it has experienced in recent years. In addition, we discuss why this recent rapid retreat of Arctic summer sea ice might largely be a consequence of a slow shift in ice-thickness distribution, which will lead to strongly increased year-to-year variability of the Arctic summer sea-ice extent. This variability will render seasonal forecasts of the Arctic summer sea-ice extent increasingly difficult. We also discuss why, in contrast to Arctic summer sea ice, a tipping point is more likely to exist for the loss of the Greenland ice sheet and the West Antarctic ice sheet.

Greenland | West Antarctic | climate change | tipping point | Arctic

Barely any other component of the Earth’s climate system has received as much public attention with respect to the possible existence of so-called “tipping points” as the ice masses covering the polar oceans, Greenland, and the Antarctic. This attention is probably due to a number of reasons, including (i) the ease with which the ice–albedo feedback and its possible consequence of a tipping point can be explained; (ii) the rapid decrease of Arctic summer sea-ice extent in 2007; and (iii) the consequences for the global sea level if indeed the Greenland ice sheet and/or the Antarctic ice sheet were to melt rapidly.

Although these points may well have led to a tipping point in the public perception of the future melting of the Earth’s ice masses, there still exists a significant lack of scientific understanding of the cryospheric “tipping elements”. In this contribution, we review some of the recent scientific progress regarding the future evolution of the Earth’s cryosphere and discuss conceptually the possible existence of cryospheric tipping points. Following ref. 1, these tipping points are here defined to exist if the ice does not recover from a certain ice loss caused by climatic warming even if the climatic forcing were to return to the colder conditions that existed before the onset of that specific ice loss. This definition is more restrictive than the one recently formulated in ref. 2: For a tipping point to exist according to our definition, the response of a climatic element must show significant hysteresis. We believe that this definition is in accordance with the early parlance of tipping points in the sociological literature (3, 4) and with most of the previous analysis of nonlinear cryospheric behavior that is reviewed in this contribution. Note also that we here do not discuss a tipping of the external atmospheric and oceanic forcing that could lead to the irreversible loss of sea ice or ice sheets. We instead focus on tipping points that are inherent to the cryosphere.

In the following, we first outline in a simple energy-balance model why the existence of the ice–albedo feedback does not necessarily lead to instability of the Earth’s ice masses. We then

briefly review the possible existence of the so-called “small ice-cap instability” before moving on to discuss the stability of the three elements of the Earth’s cryosphere that are probably the most crucial for the future evolution of the Earth’s climate: sea ice, the Greenland ice sheet, and the West Antarctic ice sheet (WAIS). The paper closes with a short discussion.

The Ice–Albedo Feedback

The assumed existence of a tipping point during the loss of the Earth’s ice masses is often motivated by the destabilizing ice–albedo feedback: If a certain ice cover is decreasing in size, the albedo (i.e., reflectivity) of the formerly ice-covered region usually decreases. Hence, more sunlight can be absorbed, the additional heating of which gives rise to further shrinkage of the ice cover until all ice is gone.

As simple as this feedback loop seems to be, it does not necessarily lead to an instability during the loss of ice, as described by the so-called feedback factor. To explain this concept to a reader who is unfamiliar with it, it is instructive to use a very simple zero-dimensional energy-balance model that shows how a very simple negative (i.e. stabilizing) feedback, namely the increase of outgoing longwave radiation, can prevent the accelerating loss of a polar ice cover.

Such a simple energy-balance model is often given by the balance of incoming shortwave radiation and outgoing longwave radiation (e.g., 5); i.e.,

$$c \frac{dT}{dt} = (1 - \alpha(T))F_{Sw} - \varepsilon\sigma T^4. \quad [1]$$

Here, c is heat capacity, T is temperature, t is time, F_{Sw} is the mean incoming shortwave radiation at the ground, α is albedo, ε is the effective emissivity, describing the partial opaqueness of the atmosphere to outgoing longwave radiation, and σ is the Stefan–Boltzmann constant. In a warmer climate, the albedo of the Earth is likely to decrease because of the smaller extent of snow and ice. Following ref. 6, we approximate this temperature dependence of the albedo as linear and set

$$\alpha(T) = \alpha_0 - \beta T.$$

To examine the stability of the temperature to small changes ΔT , we substitute T with $T + \Delta T$ in Eq. 1 and linearize to obtain

$$c \frac{d(T + \Delta T)}{dt} = (1 - \alpha_0 + \beta T + \beta \Delta T)F_{Sw} - \varepsilon\sigma T^4 - 4\varepsilon\sigma T^3 \Delta T. \quad [2]$$

Subtracting Eq. 1 from Eq. 2 leads to

$$c \frac{d(\Delta T)}{dt} = (\beta F_{Sw} - 4\varepsilon\sigma T^3)\Delta T. \quad [3]$$

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If the bracketed expression on the right-hand side is larger than zero, any small temperature change ΔT would grow rapidly, and an unstable feedback would be established. Introducing $\sigma = 5.67 \cdot 10^{-8} \text{ W m}^{-2} \text{ K}^{-4}$ and mean Arctic values of $F_{\text{sw}} = 173 \text{ W/m}^2$, $T = 255 \text{ K}$ (see ref. 7), the bracketed expression in Eq. 3 reduces to the stability criterion

$$\beta < 0.022\text{K}^{-1}\epsilon \quad [4]$$

for an isolated ice cover in the Arctic.

Two important conclusions can be drawn from Eq. 4. First, for too-low β , corresponding to an insufficient strength of the ice-albedo feedback mechanism, an ice cover at the pole might well be stable because any warming not only decreases the size of the ice-covered region but also increases the outgoing longwave radiation, thus stabilizing the ice cover (see ref. 7 for a more-detailed discussion). For a value $\epsilon \approx 0.7$, as estimated from a coupled ocean-atmosphere general circulation model (A. Voigt, personal communication), the stability criterion reduces for today's climate to $\beta < 0.0154\text{K}^{-1}$. Given previous estimates of $\beta = 0.009\text{K}^{-1}$ (see ref. 6) and $\beta = 0.0145\text{K}^{-1}$ (see ref. 8), the polar ice cover were stable in this very simple model. Note that even such "stable" ice cover shrinks if the climate becomes warmer; however, this transition to a smaller ice cover will be smooth without showing the sudden disappearance that is described by the so-called small ice-cap instability (see *The Small Ice-Cap Instability*).

The second important conclusion relates to the role of ϵ in determining the stability in Eq. 4. With increasing greenhouse-gas concentration, the efficiency of longwave emission into space, and hence the value of ϵ , is very likely to decrease in the future. Hence, even if an ice cover were stable in this simple energy-balance model for today's climate, it might well become unstable in a future warmer climate with a smaller ϵ : In such a climate, the transition to a smaller ice cover could be strongly nonlinear. A similar result, but for a different reason as discussed in *Arctic Sea Ice*, is also found in more-complex modeling studies (1, 9).

Note that the simple energy-balance model presented here cannot and should not be used to give a reliable quantitative estimate of the instability of the Earth's ice masses, not least because a number of feedbacks and other relevant factors have been neglected, as discussed recently by Winton (7). His study suggests, for example, that changes in lateral heat exchange might be more important for stabilizing Arctic sea ice than the change in outgoing longwave radiation. Nevertheless, this simple energy-balance model elucidates the fact that any stabilizing feedback, here the increase of outgoing longwave radiation in a warmer climate, might be sufficient to remove the apparently very obvious instability triggered by the ice-albedo feedback. The question of whether or not a finite-sized ice cover will indeed be stable in a warmer climate has received much theoretical attention in the last few decades, often in the context of the so-called small ice-cap instability, which will be reviewed in the following section.

The Small Ice-Cap Instability

In 1924, C. E. P. Brooks discussed at the Royal Meteorological Society in London the stability of a finite-sized ice cover in polar regions (10). He concluded that "only two types of oceanic polar climate are possible, a mild type and a glacial type", with the former referring to an ice-free ocean and the latter to a polar ice cover that extends from the North Pole at least as far south as 78° N . Although his discussion focussed on sea ice, he remarked that his argument also held for ice on land, where an ice cap of less than a certain size would be unstable and disappear rapidly in a warming climate. His argument is based on the self-induced cooling that a large ice cover would cause owing to its high albedo. If the extent of this ice cover were to decrease, this self-induced cooling would diminish more and more. The ice cover would hence become smaller and smaller until it disappears.

Brooks' account is probably the first to describe what is now known as the small ice-cap instability, which in principle claims that the only stable ice cover is one that is large enough to maintain its own climate (9). Whether this instability indeed exists has been debated since the early times of energy-balance models, with many such models showing a similar instability (refs. 11–14, and references therein). In these models, the instability is caused by an effect similar to that initially described by Brooks. Therefore, the threshold for the instability depends crucially on the parameterization of the cooling effect of the ice cover, as exemplified by its albedo and the strength of heat diffusion from the interior of the ice cover. For example, the instability disappears for a nonlinear diffusive heat transport or a smooth albedo transition at the edge of the ice cover (e.g., 14, 15).

The fact that a nonlinear heat transport into and out of the polar regions removes the instability was explained in detail by North (16). Given that more complex models usually employ more realistic representations of heat transport, it was expected that they might not show the small ice-cap instability. Recently, a number of studies found that it is indeed probable that no such instability exists for the loss of Arctic summer sea ice, whereas an instability might well exist for the transition from a seasonally ice-covered Arctic Ocean to an Arctic Ocean that is virtually free of sea ice throughout the entire year (1, 7, 9).

Arctic Sea Ice

The finding that probably no instability exists for the loss of summer sea ice might at first appear to be in contrast to the recent rapid retreat of Arctic summer sea ice. In the course of just one year, from September 2006 until September 2007, the minimum ice extent of Arctic sea ice decreased by more than 1.6 million km^2 , leading to a summer sea-ice extent that was only half as large as during the early 1950s (17). This decrease has given increased momentum to the claim that a tipping point does exist for the loss of Arctic summer sea ice. But was summer 2007 really such an unusual event in the long-term perspective of Arctic sea-ice evolution?

To examine this question, it is instructive to first simply look at the time series of minimum Arctic sea-ice extent from 1953 until 2008 (Fig. 1). The data were derived from a combination of ship observations, aerial reconnaissance flights, and satellite measurements, all collected in the HadISST dataset (18, 19). For consistency with earlier studies, these data were modified similarly to the

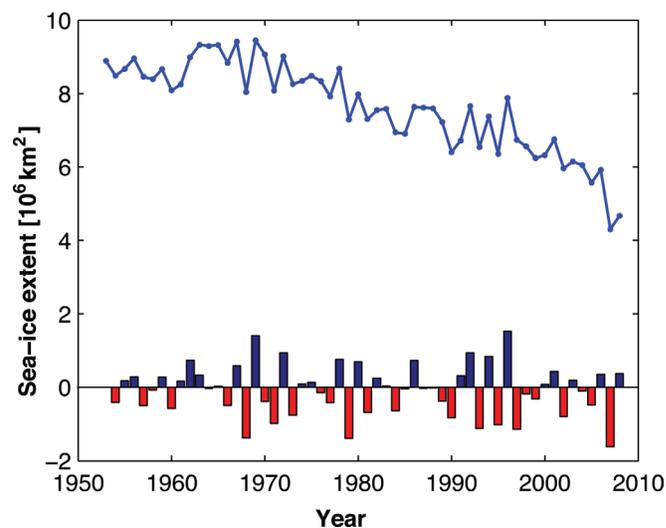


Fig. 1. Evolution of minimum Arctic sea-ice extent from 1953 until 2008. The blue line shows the minimum sea-ice extent, and the bars at the bottom show the change in extent from one year to the next.

procedure described by Meier et al. (20): All data since 1997 were replaced by the updated dataset given by the National Snow and Ice Data Center sea-ice index (http://nsidc.org/data/seaice_index/, ref. 21), and the HadISST data predating 1997 were adjusted by a positive offset of $0.25 \cdot 10^6 \text{ km}^2$. This adjustment results in a somewhat larger recent decrease in ice extent than was described by the original HadISST dataset.

The blue line in Fig. 1 shows the resulting minimum sea-ice extent, and the bars at the bottom show the year-to-year changes. Whereas the extreme minimum of summer 2007 clearly stands out, the year-to-year change from 2006 to 2007 is only somewhat larger than previous extreme changes: From 2006 until 2007, the minimum sea-ice extent decreased by $1.63 \cdot 10^6 \text{ km}^2$. Previous record reductions were observed from 1978 to 1979, then amounting to a decrease in ice extent of $1.39 \cdot 10^6 \text{ km}^2$, and from 1967 to 1968, then with a decrease in ice extent of $1.38 \cdot 10^6 \text{ km}^2$. The rapid recent retreat that was experienced in summer 2007 is hence, per se, not necessarily an indication that a tipping point exists for Arctic summer sea ice.

The observed loss of ice extent in summer 2007 can instead readily be explained by a smooth and gradual change in ice-thickness distribution, with no need to employ the concept of a tipping point. In the Arctic, sea-ice thickness varies greatly on all horizontal length scales from a few meters to the width of the Arctic Ocean basin. While the large-scale differences in ice thickness come about from differences in atmospheric and oceanic forcing, the small-scale differences are usually a result of local sea-ice dynamics. This variety of ice thicknesses can be visualized by their histogram, which then forms the ice-thickness distribution (22). In such ice-thickness distribution, there is usually a typical, modal ice thickness that covers a comparably large area (see sample distributions in the insets of Fig. 2).

Integrating an ice-thickness distribution results in the so-called cumulative ice-thickness distribution, which is shown in the main frames of Fig. 2 for an area covered with thick ice (Fig. 2A is exemplary for Arctic sea ice until the early 1990s) and for an area covered with thinner ice (Fig. 2B is representative of much of today's Arctic sea ice). In each case, the black line indicates which percentage of the area is covered by ice that is thicker than the ice thickness that is given along the x axis.

The black line is assumed to show the cumulative ice-thickness distribution for a certain area at the onset of summer melting, and the three colored lines represent different amounts of thinning during summer. For simplicity, we make the assumption that all ice in this area experiences roughly the same thinning, which implies that the ice-thickness distribution preserves its shape during summer and is simply shifted to the left. The area that becomes ice-free during a particular summer is given by the area initially covered by ice thinner than the assumed total thinning during that summer. Because most ice for the thick-ice case in the top frame is much thicker than 1 m, only a small fraction of the area becomes ice-free during summer, whether the summer is relatively cool with, say, just 0.5 m thinning or relatively warm with 1.0 m of thinning. Hence, with thick overall ice cover, the year-to-year variability of ice extent during summer is relatively small.

Now consider Fig. 2B, which shows the impact of a much thinner ice-thickness distribution: Because here most of the ice has a thickness of around 1 m, the area that becomes ice-free during summer crucially depends on the amount of total melting: In the example shown here, 90% of the area remains ice covered for a total melt of 0.5 m during a relatively cool summer, whereas only 50% of the area remains ice covered for 1 m of total thinning during a warmer summer.

From this qualitative description, we can draw two conclusions, which might prove crucial in the assessment of recent years' melting events: First, with a thinner ice-thickness distribution, the same amount of heat input leads to a much larger ice-free area, as has been noted by several previous studies (23–27). This relationship

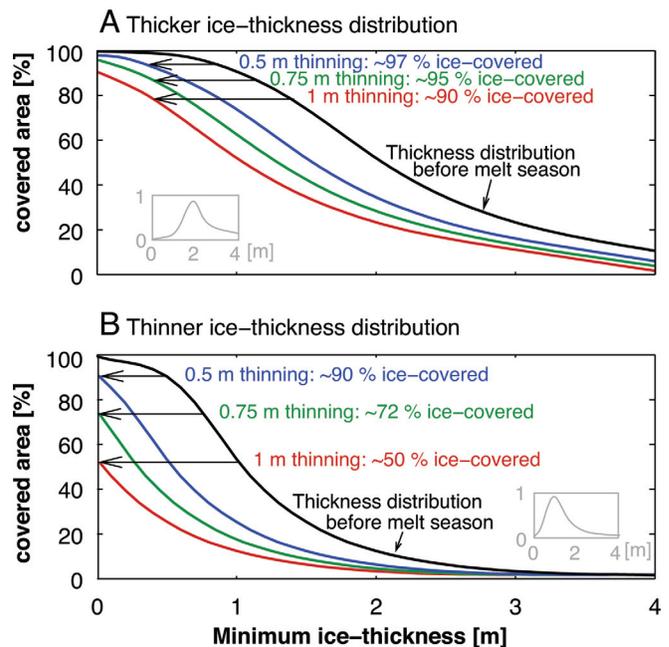


Fig. 2. Comparison of the impact of a certain amount of summer melt on the ice-covered area for two different ice-thickness distributions. The black line shows which percentage of a certain area is covered by ice that is thicker than the ice thickness indicated along the x axis at the beginning of the melt season (cumulative ice-thickness distribution). **A** shows a possible cumulative thickness distribution of thick ice and **B** shows that of thinner ice. The colored lines indicate the shift of the ice-thickness distribution for a uniform melting during summers of 0.5 m (cold summer, blue), 0.75 m (normal summer, green) and 1.0 m (warm summer, red), respectively. Note the much stronger variability of the remaining ice-covered area in the case with thinner ice. The *Inset* shows roughly the noncumulative ice-thickness distribution for each case. Here, the y axis indicates normalized area that is covered by the ice thickness shown along the x axis.

has sometimes been described as an increase in “open-water efficacy” (e.g., 26, 27). Because of the nonlinear ice-thickness distribution, a gradual thinning of the ice cover can initially lead to an acceleration, and, at some point, a very rapid loss of ice-covered area during summer. Once the modal ice thickness is less than a typical summer melt rate, the additional ice retreat will again proceed at a slower rate. This conceptual image might help to rationalize the possibly rather sudden future reduction in summer sea-ice extent, followed by a slower retreat of the remaining ice that has been found in a number of recent modeling studies (26–28).

A second conclusion that can be drawn from considering the shift in the ice-thickness distribution is the fact that with a thinner sea-ice cover, the size of the area that becomes ice-free during summer depends much more on the actual weather during a particular summer than is the case for thicker ice. Therefore, with the ongoing thinning of the ice cover (23, 25), we are likely to experience both large negative and large positive year-to-year changes in Arctic summer sea-ice extent. This variability directly implies a much-reduced predictability of sea-ice extent a season ahead than does thicker ice. The increased variability will continue until most of the ice cover is thinner than the typical summer melting rate, at which point most of the ice will disappear during summer, with only small year-to-year changes. The decreased predictability that goes along with such an increase in variability was also found in a recent more complex modeling study (27).

The transition to a state with a much-reduced summer sea-ice cover will probably show periods of strongly increased rates of ice loss, as has been discussed above. Nevertheless, a number of recent studies find that the transition to a seasonally ice-free Arctic is probably reversible in a climate that is cool once again. (1, 7, 9, 28).

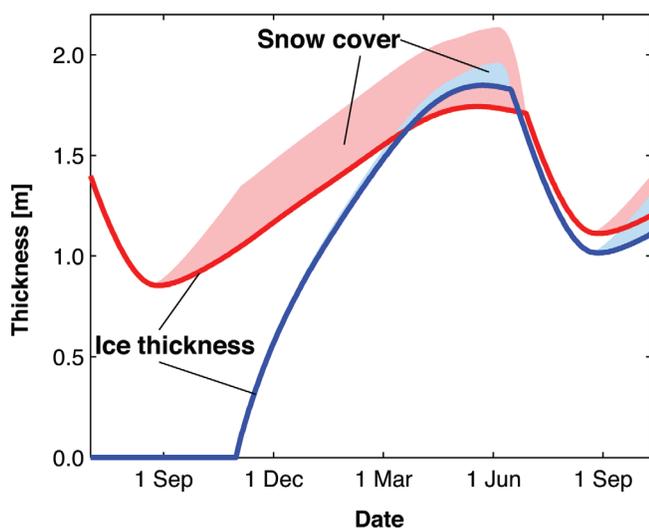


Fig. 3. Schematic comparison of the ice-thickness evolution of first-year ice (blue) and of preexisting ice (red). The snow on top of the ice is represented by the shaded blue and red areas. The atmospheric and oceanic boundary conditions for this simulation, including the incoming radiation, sensible and latent heat fluxes, and the amount of snowfall, are given by Arctic climatological mean values as summarized by Maykut and Untersteiner (66). For simplicity, the ice growth was simulated by a zero-layer model (see ref. 67) and there is no feedback from the ice to the atmospheric or oceanic forcing. Ice formation of the first-year sea ice was prescribed to start on November 1, two months after the end of the melting season of the preexisting ice. Note that, nevertheless, the newly forming ice reaches a larger maximum thickness than the preexisting ice because of the thinner isolating snow-cover on top of the former.

For example, Eisenman and Wettlaufer (1) use a simplified model setup based largely on the setup described by Thorndike (29), which allows them to examine the phase space of the transition from a permanent ice cover to a seasonal ice cover and finally to an ice-free Arctic. These authors find that the ice–albedo feedback, which by itself could trigger an instability during the retreat of sea ice, is compensated by a stabilizing feedback related to the ice-growth rate: Since Stefan’s early works on ice growth in 1889 (30), it has been known that thin ice grows substantially faster than thick ice. This faster growth gives rise to a stabilizing feedback, as has been discussed by a number of studies (e.g., 31, 32). This feedback implies that as long as winter air temperatures remain well below freezing, even after very strong thinning during summer, the ice can recover rapidly during winter—an effect that becomes even stronger if one also considers the impact of snow.

Compare, for example, the evolution of ice that survived a certain summer and the evolution of ice that forms, exposed to the same forcing, in open water some months after the end of the melting season (Fig. 3). In the example shown here, the ice that survived one summer has an ice thickness of about 85 cm at the end of the melting season in early September. During the following winter, it reaches a maximum ice thickness of about 1.75 m (red graph). The ice that forms from open water is assumed to start growing on November 1 (blue graph). Maybe surprisingly, this ice, even though it has two months less to grow during winter, reaches a larger ice thickness than the preexisting ice, which was 85 cm thicker at the onset of its growth. This finding is similar to the early results by Budyko from 1966 (33): In his figure 4c, the maximum ice thickness increases once the Arctic has become seasonally ice-free. The reasons for this increase are twofold: First, because of its low ice thickness the thin ice initially grows very fast, and second, in the example shown here, significant amounts of snow accumulate on the preexisting ice before November 1, leading to a much thicker snow cover on this ice throughout most of the winter. This snow cover very efficiently insulates the existing ice

from the cold atmosphere, significantly slowing down its growth. In effect, the formerly ice-free area is, at the end of just one single growth season, covered with ice of almost the same thickness as the area with preexisting ice.

It is this stabilizing feedback that explains in part why it was possible for the Arctic sea-ice extent to increase again in summer 2008 and 2009 after the minimum extent observed in summer 2007. This stabilizing feedback implies that any measure that is taken to slow down or even reverse global warming will probably allow for a rather immediate corresponding response of the Arctic summer sea ice.

For the loss of winter sea ice in a much warmer climate, some recent studies find the existence of a tipping point (1, 7, 9). This tipping point is in part rationalized by the change in the strength of the ice–albedo feedback: Today, with a widespread sea-ice cover during winter, most of the Arctic remains ice-covered for long periods of the summer. Once a certain area becomes ice-free all year round, the open water will be heated up from spring until autumn, possibly increasing the ice–albedo feedback enough to lead to a rapidly disappearing winter sea-ice cover. In the simple energy-balance model, this is equivalent to an increased value for β in Eq. 4.

By using conceptual arguments, we have rationalized in this section some recent results from more complex studies as to why (i) during the disappearance of Arctic summer sea ice a tipping point according to the definition used here seems unlikely; (ii) the year-to-year variability in sea-ice extent is likely to increase significantly in the future; (iii) the loss of sea ice experienced during the last years is likely to be reversible if the climate were to become cooler again; and (iv) this reversibility exists only if significant amounts of sea ice still form during winter. Although the arguments given here are conceptual, they have all been confirmed in a number of recent studies that employed a large range of model complexity. However, there is still a significant uncertainty regarding the future evolution of sea ice that is also reflected by the spread in model projections for the future evolution of sea ice that were presented in the last Intergovernmental Panel on Climate Change (IPCC) report (34).

Note that the loss of Arctic summer sea ice might well show tipping behavior according to the broader definition of Lenton et al. (2). According to their definition, a climate element shows tipping behavior if it can be “switched into a qualitatively different state by small perturbations.” Given that the observed retreat of Arctic sea ice and the possible complete disappearance of sea ice in the future are triggered by a relatively small change in the temperature forcing, sea ice might well show tipping behavior according to the Lenton et al. definition. This is especially the case given that many models show nonlinearity in the transition to an ice-free Arctic (26).

Antarctic Sea Ice

So far, we have focussed exclusively on the observed and expected changes of Arctic sea ice. In the Antarctic, the situation is rather different: Most of the sea ice that is formed there during winter does not survive the following summer. The loss of this ice is due to the fact that in the Southern Ocean, ocean eddies can transport significant amounts of heat to the underside of sea ice to eventually melt it. For the last few decades, a number of studies suggest no large trends in total Antarctic sea-ice extent (ref. 35 and references therein). Including proxy records based on whaling data, there is some debated evidence for a retreat of Antarctic sea ice since the middle of the twentieth century until the late 1970s (refs. 36 and 37 and references therein). A recent study reports a slight upward trend over the last three decades (38), in contrast to the trend simulated by most of the models used for the last IPCC report (39, 40).

Given that today already less than 15% of Antarctic sea ice survives the summer, it is most likely of greater importance to assess the possible existence of a tipping point for the disappearance of

Antarctic winter sea ice. Such a disappearance would have a profound impact on Southern Ocean circulation (e.g., 41). In light of the so-called Weddell Polynya in the 1970s, and taking the structure of the Southern Ocean into account, there has been some speculation about the possibility of a sudden, sustained disappearance of sea ice over a large area triggered by so-called thermobaric convection that would bring warm water to the surface and prevent winter ice formation (42). However, more-sophisticated modeling studies are needed to assess the likelihood of such an event taking place, not least under a global-warming scenario. Currently, our understanding of the Southern Ocean is too limited to assess realistically the likelihood of a sudden, irreversible decrease in sea-ice extent occurring over the next several decades.

Greenland and WAIS

Most early studies that dealt with instabilities of the Earth's cryosphere did not distinguish between ice on land and ice in the sea. Given that chances are high that there is no tipping point according to our definition for the loss of summer sea ice in the Arctic, the question therefore naturally arises whether the same might hold for the large ice sheets covering Greenland and the West Antarctic. To answer this question, it is instructive to discuss briefly, with no claim to be complete, some of the main differences between the loss of sea ice and the loss of ice sheets, mostly to show why the nonexistence of a tipping point for the former has no consequences for the existence of a tipping point for the latter. These differences can be summarized as follows.

First, as discussed above, the growth rate of sea ice depends crucially on its thickness, with thin ice growing much faster than thick ice. This faster growth allows both for a relatively strong resilience against a sudden acceleration of sea-ice volume loss and for a quick recovery of summer sea ice when the climate becomes cold once again. For ice sheets, there are also a number of stabilizing feedbacks. The growth of ice sheets depends almost exclusively on the accumulation of snow on its surface, and because warmer air can transport more moisture, the amount of snow-fall usually increases in a warmer climate and with decreasing surface elevation. Model studies have also found some stabilizing feedback relating to air-circulation changes at the ice-sheet margin (43). Although these two effects combined provide some negative feedback to keep ice sheets stable over a certain range of climatic conditions, their magnitude is probably not comparable with the very strong stabilizing feedback that is exhibited by the large increase of the growth rate for thin sea ice.

Second, a smaller sea-ice extent basically enhances the original warming only because of the ice-albedo feedback mechanism. Ice sheets, however, will additionally have their surface temperature increase by 0.6–1 °C per 100-m loss of their vertical extent. This increase is due to the adiabatic lapse rate that makes the surface temperature increase by a certain amount with reduced height.

Third, as far as dynamics are concerned, a thinner sea-ice cover is more prone to the ice floes being squeezed together to eventually pile up on top of one another. This tendency provides for two different mechanisms which both can increase ice thickness. First, the piled-up ice floes constitute thicker ice. Second, ice piling up produces more open water (22) (where during summer more solar radiation can be absorbed but where in winter more new sea ice can be formed) than if that area had remained ice-covered. A thinning of sea ice can hence lead to changes in its dynamics that can slow down this thinning—another stabilizing feedback. Such feedback does not exist to the same extent for the loss of ice sheets where, usually, changes in ice dynamics will lead to faster ice loss in a warmer climate. Although the details of the impact of global warming on ice-sheet dynamics remain unclear, in part because of a lack of understanding of the induced changes in basal friction and the seaward margin of the ice sheet (44), it seems likely that the observed acceleration of many of the Greenlandic outlet glaciers is due to a combination of several effects linked to such warming (45–47).

Fourth, there is no paleoclimatic evidence for any tipping-point behavior of Arctic sea ice during the past several million years, during which the Arctic ocean has probably been permanently ice-covered (48). In contrast, there is widespread consensus on the existence of paleoclimatic evidence for ice-sheet instability during that time (49, 50), with at least two sudden events during the last deglaciation that led to a sea-level rise of 20 m in a few centuries (51, 52). Notwithstanding the potential for future instability of the Earth's ice sheets, it should be noted that a sea-level rise even remotely this large seems impossible in the foreseeable future (53), not least because those meltwater pulses contributed the equivalent of between 1.5 and 3 Greenland ice sheets each (51).

Fifth, there is some evidence from general circulation models that the removal of the Greenland ice sheet would be irreversible, even if climate were to return to preindustrial conditions (54), which increases the likelihood of a tipping point to exist. This finding, however, is disputed by others (55), indicating the as-of-yet very large uncertainty in the modeling of ice-sheet behavior in a warmer climate (56).

Sixth, there is also some indication for the existence of a tipping point for the loss of the WAIS in a warming climate. The change of the scientific consensus with time whether this ice sheet is stable is an example for the existence of what has been referred to as “negative learning” (57): soon after the early works of Mercer in the late 1960s, culminating in his famous 1978 publication (58), it was widely agreed that the WAIS was indeed unstable. During global warming, floating ice shelves would disintegrate rapidly, giving way to an accelerating loss of the vast ice sheets covering the western Antarctic. This paradigm lost some momentum during the 1980s and early 1990s when some modeling studies suggested that the ice sheet might be stable even if the floating ice shelves were lost. More-recent satellite observations and laser altimetry have in turn again suggested that the WAIS might indeed be unstable, contributing significantly to sea-level rise in the coming two centuries (for details see reviews in refs. 59 and 60). A recent analysis of the stability of ice-sheet grounding lines has also given renewed evidence that the WAIS might indeed be unstable (61) if global warming exceeds a certain threshold. Recent modeling and paleoproxy studies also suggest that small changes in the forcing can lead to “brief but dramatic” retreats of the WAIS (62, 63). It seems, hence, at the moment well possible that a tipping point exists for a possible collapse of the WAIS.

Note that there are a great number of additional processes that render the behavior of ice sheets inherently different from that of sea ice. Nevertheless, this brief overview already indicates the possible existence of a tipping point for the loss of the Earth's ice sheets. In this context, it should be noted that the local warming needed to slowly melt the Greenland ice sheet is estimated by some authors to be as low as 2.7 °C warming above preindustrial temperatures (64, 65). This amount of local warming is likely to be reached for a global warming of less than 2 °C (39).

Conclusion and Summary

In this contribution, we have examined the existence of so-called tipping points during the loss of sea ice and during the melting of the Greenland and WAIS. We have discussed why sea ice and ice sheets show greatly different behavior, with different consequences for the existence of a tipping point.

For Arctic sea ice, we have explained why the recently observed rapid decrease in ice extent might just be a consequence of a smooth and slow shift in ice-thickness distribution. We have also explained how the ice cover is stabilized in more-complex studies because of the strong increase of ice-growth rate with shrinking ice thickness. These more-complex studies conclude that there is probably no tipping point (according to the definition used here) for the loss of Arctic summer sea ice. Hence, sea ice is probably capable of recovering rapidly once the climate turns cold again, and any measures taken to slow down climate warming can immediately slow down future sea-ice loss. If no measures are taken to

