Chapter 1.2 Tipping points in the cryosphere

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Summary

Drastic changes in our planet's frozen landscapes have occurred over recent decades, from Arctic sea ice decline and thawing of permafrost soils to polar amplification, the retreat of glaciers and ice loss from the ice sheets. In this chapter, we assess multiple lines of evidence for tipping points in the cryosphere – encompassing the ice sheets on Greenland and Antarctica, sea ice, mountain glaciers and permafrost – based on recent observations, palaeorecords, numerical modelling and theoretical understanding.

With about 1.2°C of global warming compared to pre-industrial levels, we are getting dangerously close to the temperature thresholds of some major tipping points for the ice sheets of Greenland and West Antarctica. Crossing these would lock in unavoidable long-term global sea level rise of up to 10 metres. There is evidence for localised and regional tipping points for glaciers and permafrost and, while evidence for global-scale tipping dynamics in sea ice, glaciers and permafrost is limited, their decline will continue with unabated global warming.

Because of the long response times of these systems, some impacts of crossing potential tipping points will unfold over centuries to millennia. However, with the current trajectory of greenhouse gas (GHG) emissions and subsequent anthropogenic climate change, such largely irreversible changes might already have been triggered. These will cause far-reaching impacts for ecosystems and humans alike, threatening the livelihoods of millions of people, and will become more severe the further global warming progresses.

The scientific content of this chapter is based on the following manuscript in preparation: Winkelmann et al., (in prep)

Key messages

- Large-scale tipping points exist for the Greenland and Antarctic ice sheets, as inferred from multiple lines of evidence. Crossing these tipping points would lead to multi-metre sea level rise over hundreds to thousands of years.
- There is evidence for localised and regional tipping points in glaciers and localised tipping points in permafrost, but evidence for large-scale tipping dynamics in sea ice, glaciers and permafrost is limited.
- Some ice sheet tipping points could be close at current warming levels, with further warming increasing their likelihood. Localised tipping can already be observed for permafrost, and will worsen with further warming, along with non-tipping impacts.

Recommendations

- Protect the cryosphere through urgent and ambitious phase-out of GHG emissions, as well as reducing co-drivers such as black carbon.
- Reduce and/or better understand deep uncertainties, including: 1) instabilities in marine-based ice sheet dynamics; 2) the coupled dynamics of the Southern Ocean, sea ice, and ice shelf system; 3) integrating local glacier feedbacks into glacier modelling; and 4) the impact of abrupt permafrost thaw dynamics on the global permafrost-carbon feedback.
- Invest in observations and improved modelling to constrain projected impacts for the next decades and beyond, and detect early warning signs of cryosphere tipping. Foster data sharing and international collaboration.
- Co-design research, bringing together natural and social sciences and multiple knowledge systems, including Indigenous knowledge, to improve decision making under deep uncertainty, reduce risks and effectively adapt to unavoidable impacts.

1.2.1 Introduction

The Earth's cryosphere, encompassing large expanses of frozen landscapes, is critical to its climate system (Fox-Kemper et al., 2021). From the vast ice sheets on Greenland and Antarctica to mountain glaciers, sea ice and the permanently frozen soils of the Arctic, the cryosphere plays a crucial role in storing freshwater and carbon, regulating global climate patterns and influencing major ecosystems (Figure 1.2.1). However, it is also one of the parts of the Earth system most vulnerable to climate change. As our climate undergoes unprecedented shifts due to human-induced global warming, the cryosphere is at risk of crossing potential tipping points (Lenton et al., 2012; Armstrong McKay et al., 2022; Wang et al., 2023).

Cryospheric tipping dynamics are triggered when changes in part of a system become self-perpetuating beyond some threshold, leading to substantial, widespread, often abrupt and often irreversible impacts (see section 1 Introduction). This definition highlights different characteristics of tipping systems that have been discussed previously – namely the existence of critical thresholds and the potential for abrupt and possibly irreversible change, all of which we assess here for ice sheets, sea ice, glaciers and permafrost.

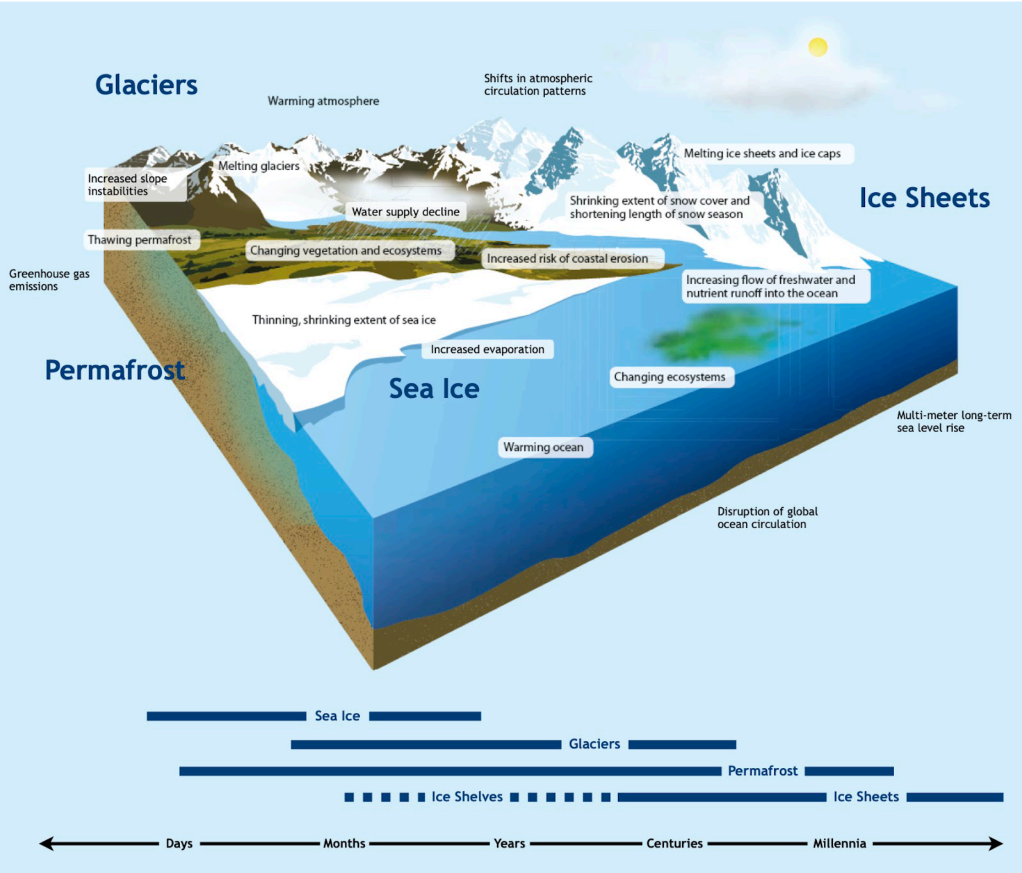

Figure 1.2.1: Key biophysical impacts resulting from crossing tipping points in the cryosphere. Diagram below gives approximate timescales of changes in the respective domain/system. Background graphic from: AMAP (2017).

The consequences of crossing cryospheric tipping points amplify the effects of climate change and have widespread impacts, affecting sea level, ecosystems, wildlife habitats, coastal infrastructure, human livelihoods and regional climate patterns (Fox-Kemper et al., 2021). They could further lead to cascading effects to other climate tipping systems, which would result in far-reaching consequences for the entire Earth system (Steffen et al., 2018; Wunderling et al., 2021; Wunderling and von der Heydt et al., preprint).

1.2.2 Current state of knowledge on cryosphere tipping points

In this section we assess available scientific literature relating to tipping points in the cryosphere, as summarised in Figure 1.2.2 and Table 1.2.1. We focus on the following systems: the ice sheets on Greenland and Antarctica, sea ice (in the Arctic and Antarctic), mountain glaciers, and permafrost.

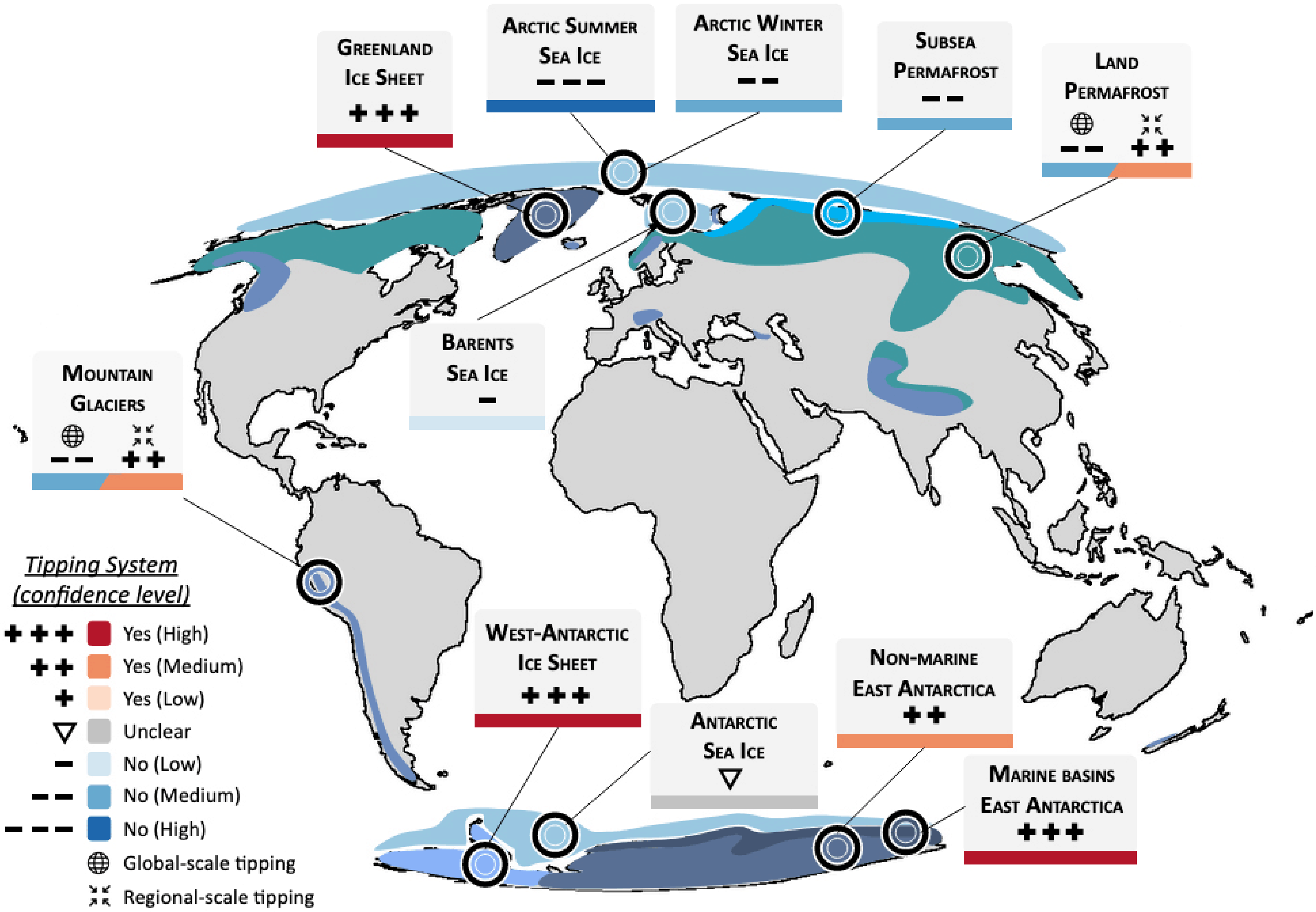

Figure 1.2.2: Map of cryosphere systems considered in this chapter (shading). The markers indicate which of the systems are in this report considered a tipping system (+++ high confidence, ++ medium confidence and + low confidence) and which are not (- - - high confidence, - - medium confidence and - low confidence), ∇ indicates systems for which a clear assessment is not possible based on the current level of understanding.

Table 1.2.1: Summary of evidence for tipping dynamics, key drivers and biophysical impacts in each system considered in this chapter

Key: +++ Yes (high confidence), **++** Yes (medium confidence), **+** Yes (low confidence), **- - -** No (high confidence), **- -** No (medium confidence), **-** No (low confidence)

Primary drivers are bolded, DC: Direct Climate driver; **CA:** Climate-Associated driver (including second-order and related effects of climate change); **NC:** Non-Climate driver**, PF:** positive (amplifying) feedback (FB), **NF:** negative (damping) feedback. Drivers can enhance **(↗)** the tipping process or counter it **(↘)**

System (and potential tipping point)	Key drivers	Key biophysical impacts	Selected key feedbacks	Abrupt / large rate change?	Critical threshold(s) (warming > preindustrial)	Irreversible? (decadal / centennial)	Tipping system?
Ice Sheets							
Greenland Ice Sheet (collapse)	DC: atmospheric warming (↗) DC: precipitation increase (↘) DC: ocean warming and circulation changes (↗/↘) DC: black carbon deposition (↗) CA: sea ice decline (↗) CA: atmospheric circulation changes (↗|↘)	• Sea level rise (up to 7m) over centuries to millennia • Disruption of global ocean circulation • Substantial shifts in atmospheric circulation patterns	• PF: melt-elevation • PF: melt-albedo	+++	0.8-3°C	+++	**+++**
West Antarctic Ice Sheet (collapse)	DC: ocean warming and circulation changes (↗) DC: atmospheric warming (↗) DC: precipitation increase (↘)	• Sea level rise (up to 3m) over centuries to millennia • Disruption of global ocean circulation • Substantial shifts in atmospheric circulation patterns	• PF: marine ice sheet instability • NF: glacial isostatic adjustment • ?: melt-stratification	+++	1-3°C	+++	**+++**
Marine basins East Antarctica (collapse)	DC: ocean warming and circulation changes (↗) DC: atmospheric warming (↗) DC: precipitation increase (↘)	• Sea level rise (up to 19m) over centuries to millennia • Disruption of global ocean circulation • Substantial shifts in atmospheric circulation patterns	• PF: marine ice sheet instability • NF: glacial isostatic adjustment • ?: melt-stratification	+++	2-6°C	+++	**+++**
Non-marine East Antarctic Ice Sheet (collapse)	DC: atmospheric warming (↗) DC: precipitation increase (↘)	• Sea level rise (up to 34m) over centuries to millennia • Disruption of global ocean circulation • Substantial shifts in atmospheric circulation patterns	• PF: melt-elevation	+++	6-10°C	++	**++**

System (and potential tipping point)	Key drivers	Key biophysical impacts	Selected key feedbacks	Abrupt / large rate change?	Critical threshold(s) (warming > preindustrial)	Irreversible? (decadal / centennial)	Tipping system?
Sea Ice							
Arctic summer sea ice (loss)	DC: atmospheric warming (↗) DC: atmospheric circulation shifts (↗/↘) DC: ocean warming (↗) DC: ocean circulation shifts (↗/↘) DC: black carbon deposition (↗) DC: storminess increase (↗) CA: ocean stratification increase (↘)	• Regional warming (polar amplification) • Ecosystem disruption • Impacts on ocean circulation • Impacts on atmospheric circulations • Increased evaporation	• PF: Ice-albedo FB • NF: Snow FB • NF: Growth FB • NF: Radiation FB	- - -	N/A	- - -	- - -
Arctic winter sea ice (loss)				+++	3-6 °C	- -	- - (abrupt loss due to Arctic geometry)
Barents sea ice (loss)				- (linear relationship in most models)	unclear	unclear	-
Antarctic sea ice (loss)				unclear	unclear	+ (reversible over millennia)	unclear
Glaciers							
Glaciers (retreat)	DC: atmospheric warming (↗) DC: deposition of dust, black carbon, etc. (albedo) (↗) DC: reduced snow (input and albedo) (↗) DC: local thermokarst (↗)	• Water supply decline • Ecosystem disruption (e.g. wetlands, water chemistry) • Increase in number and size of glacier lakesIncrease in slope instabilities • Transition from glacial to para-glacial landscapes • Sea level rise	• PF: melt-elevation FB • PF: calving front retreat • PF-: ice-dynamic FBs • NF: retreat to higher altitudes	+ + (regional) - - (global)	Regionally variable	- -	**++ (regional)** - - (global)
Permafrost							
Land permafrost (thaw)	DC: atmospheric warming (↗) CA: vegetation increase (increase albedo ↗, increase summer shading ↘, and vice versa for forest die-back) CA: wildfire intensity increase (↗) CA: precipitation increase (rain extremes, snow cover albedo ↗)	• Greenhouse gas emissions • Landscape disruption • Ecosystem disruption	• PF: carbon-climate FB • PF: thermokarst development • PF: summer soil drying • PF-: vegetation interaction	- - (global) ++ (regional)	N/A	+++ (wrt carbon loss) - - - (wrt frozen soil)	**++ (regional)** - - (global, on 10s-100s year timescale)
Subsea permafrost (thaw)	DC: ocean warming (↗) CA: sea ice loss (↗) CA: water pressure reduction (↗)	• Greenhouse gas emissions	• PF: Carbon-climate FB • NF: sediment sink • NF: water column sink	+	N/A	+ + (w.r.t. gas hydrate dissociation) + + (w.r.t. frozen sediment)	- - (global, on 10s-100s year timescale)

1.2.2.1 Ice sheets

Most of Earth's freshwater is stored in the ice sheets of Greenland and Antarctica (Figure 1.2.3). These represent by far the largest potential sources for sea level rise under ongoing and future warming: if the Greenland Ice Sheet (GrIS) were to melt entirely, global sea levels would rise by about 7 metres (Morlighem et al., 2017), for the Antarctic Ice Sheet, the total sea level rise potential is 58 metres (Fretwell et al., 2013; Morlighem et al., 2020). Even if only part of these masses were to undergo abrupt ice loss or tipping behaviour, this would have far-reaching consequences for coastal communities, infrastructure and ecosystems worldwide (Fox-Kemper et al., 2021).

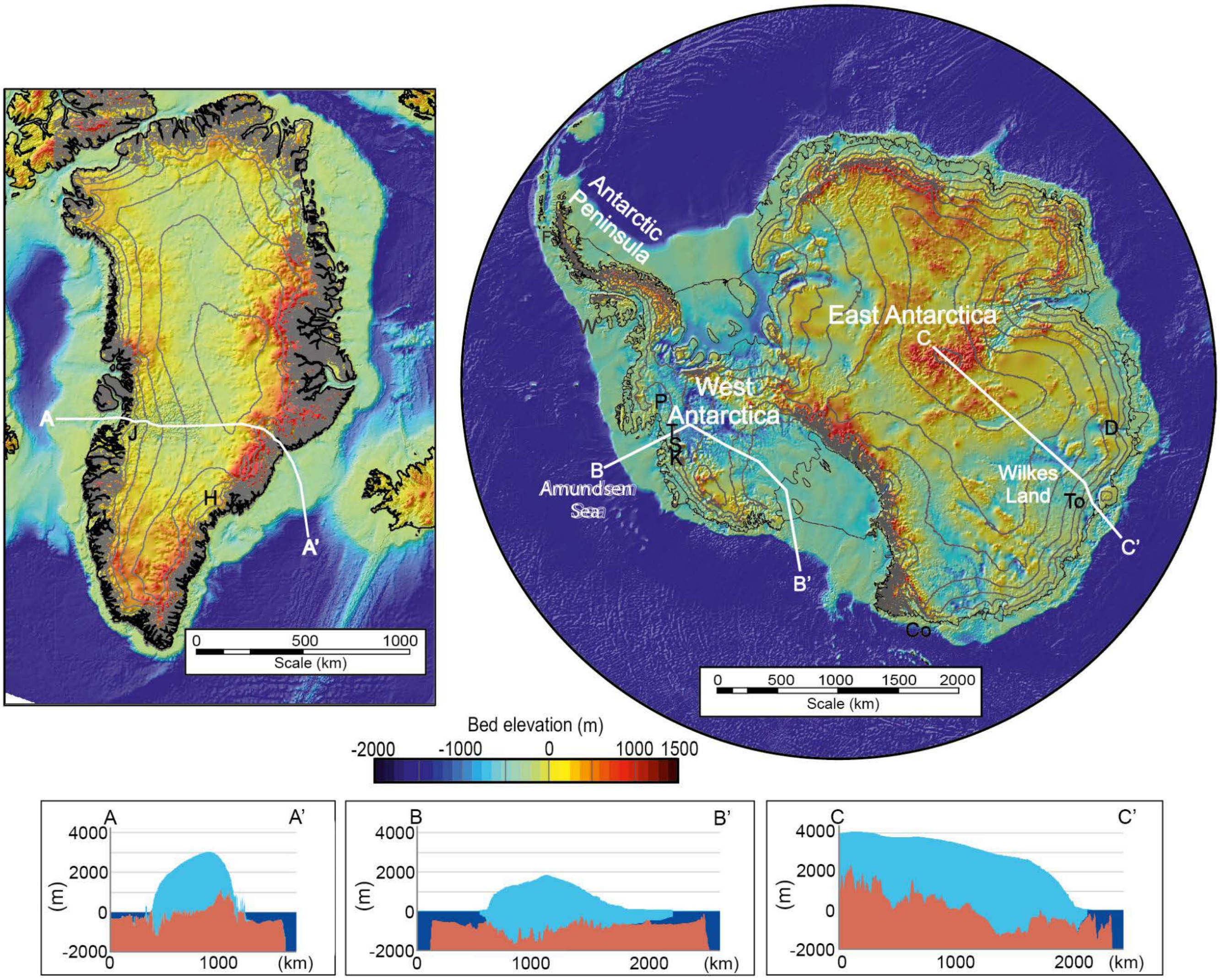

Figure 1.2.3: Greenland and Antarctic ice sheets. Given is the bedrock topography of the GrIS (left, based on Bamber et al., 2013) and the Antarctic Ice Sheet (middle and right, based on Fretwell, 2013) alongside cross sections marked in the maps by white lines. In marine ice sheet sectors (blue-green shading in the maps) the ice sheet rests on a bed submerged below sea level.

The ice sheets have been losing mass at an accelerating rate: from an average of about 105 gigatonnes (Gt – i.e. one billion tonnes) per year between 1992 and 1996 to around 372 Gt per year between 2016 and 2020 (Otosaka et al., 2023) (Figure 1.2.4). The Greenland ice sheet is (still) the major player, with an average mass loss rate of 169±9 Gt per year between 1992 and 2020, similar to the mass lost from glaciers outside of Greenland and Antarctica (Fox-Kemper et al., 2021; Hugonnet et al., 2021). Over the same period, ice losses in Antarctica were predominantly occurring in West Antarctica (The IMBIE team, 2018; Otosaka et al., 2023).

The long-term stability of the ice sheets depends on a complex interplay of amplifying (including self-sustaining) and damping feedbacks (e.g. Fyke et al., 2018). Based on multiple lines of evidence from modelling studies, observations and palaeo evidence, the ice sheets or parts thereof are considered 'global core' climate tipping systems (Armstrong McKay et al., 2022). In the following, we describe the underlying mechanisms, critical thresholds, timescales and potential for (ir)reversibility. Since the ice loss is dominated by different processes, we differentiate between the GrIS, the West Antarctic Ice Sheet (WAIS), the marine basins of East Antarctica and non-marine parts of East Antarctica.

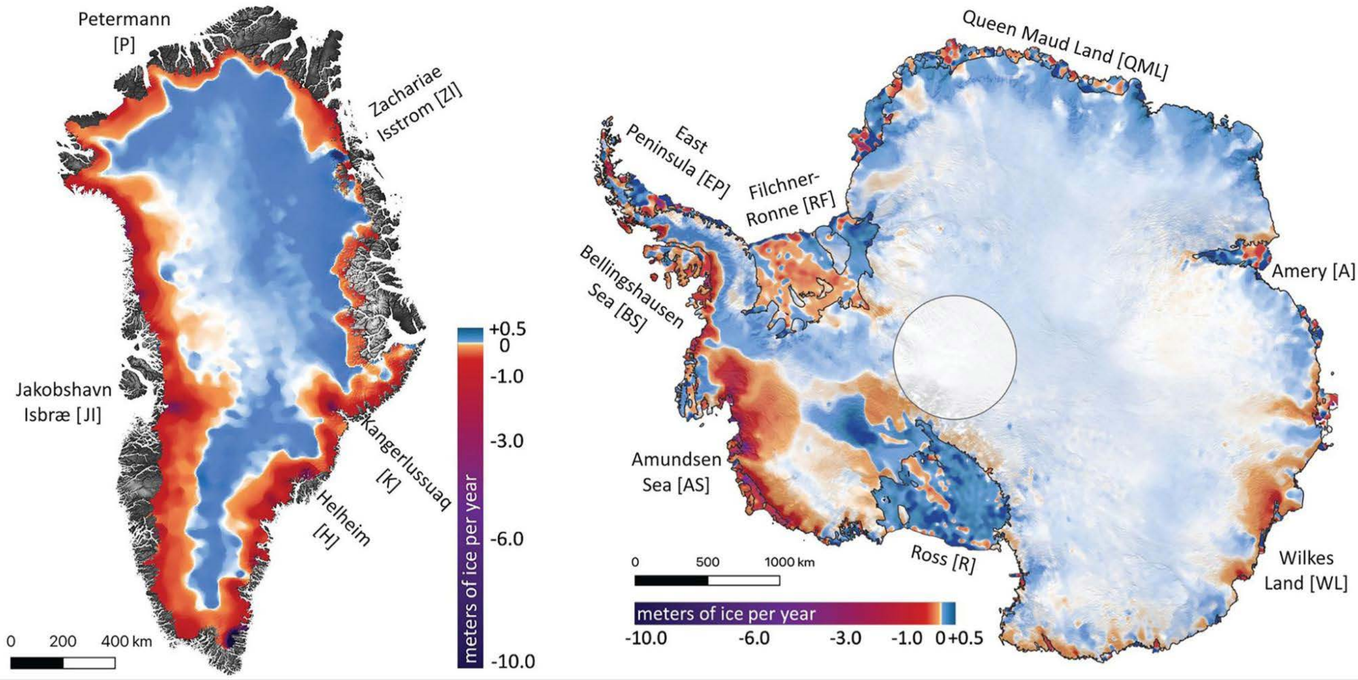

Figure 1.2.4: Observed mass change of Greenland and Antarctic ice sheets. Mass change (mass loss in red, mass gain in blue) between 2003 and 2019 for Greenland and Antarctica, given in metres of ice equivalent per year (from Smith et al., 2020).

Greenland Ice Sheet

The GrIS is a land-based continental ice sheet, with an area of 1.71 million square kilometres. At its margins, ice flows to the sea through marine-terminating outlet glaciers. The currently observed mass loss predominantly occurs through enhanced surface melting and iceberg calving (breaking at the edges) (King et al., 2020; Shepherd et al., 2020). Interactions with the atmosphere play an important role for the overall stability of the ice sheet. Several amplifying and damping feedbacks between the ice sheet and atmosphere are active in a warming climate, and these are associated with different timescales. On short timescales, a warmer climate will, on average, produce more precipitation via the added moisture-carrying capacity of the air. This mitigates some of the mass losses, since it increases accumulation (snow fall) as the climate warms. Atmospheric circulation and wind patterns will also change in response to a changing ice sheet geometry, but the effect on the overall mass balance (i.e. the balance between snow inputs and meltwater/calving outputs) of the ice sheet is not well understood.

Evidence for tipping dynamics

Associated with surface melting is a self-amplifying feedback, the *melt-elevation feedback* (Oerlemans, 1981), in which substantial melt can cause parts of the ice sheet surface to sink to lower elevations, exposing the surface to warmer air masses which in turn can lead to further melt (Figure 1.2.5). This effect is compounded by the *melt-albedo feedback*: as snowpack melts to bare ice, surface albedo (level of reflection) reduces, leading to increased absorption of solar radiation. This in turn leads to further melting and a further albedo reduction (e.g., Box et al., 2012). Glacier algae growing on bare ice can lower albedo further, a process known as the biological albedo feedback (Cook et al., 2020). Both ice sheet modelling and palaeoclimate data indicate that a tipping point can occur when the melt-elevation feedback gets strong enough to support self-accelerating mass loss (Huybrechts, 1994; Robinson et al., 2012; Ridley et al., 2010; Levermann and Winkelmann, 2016).

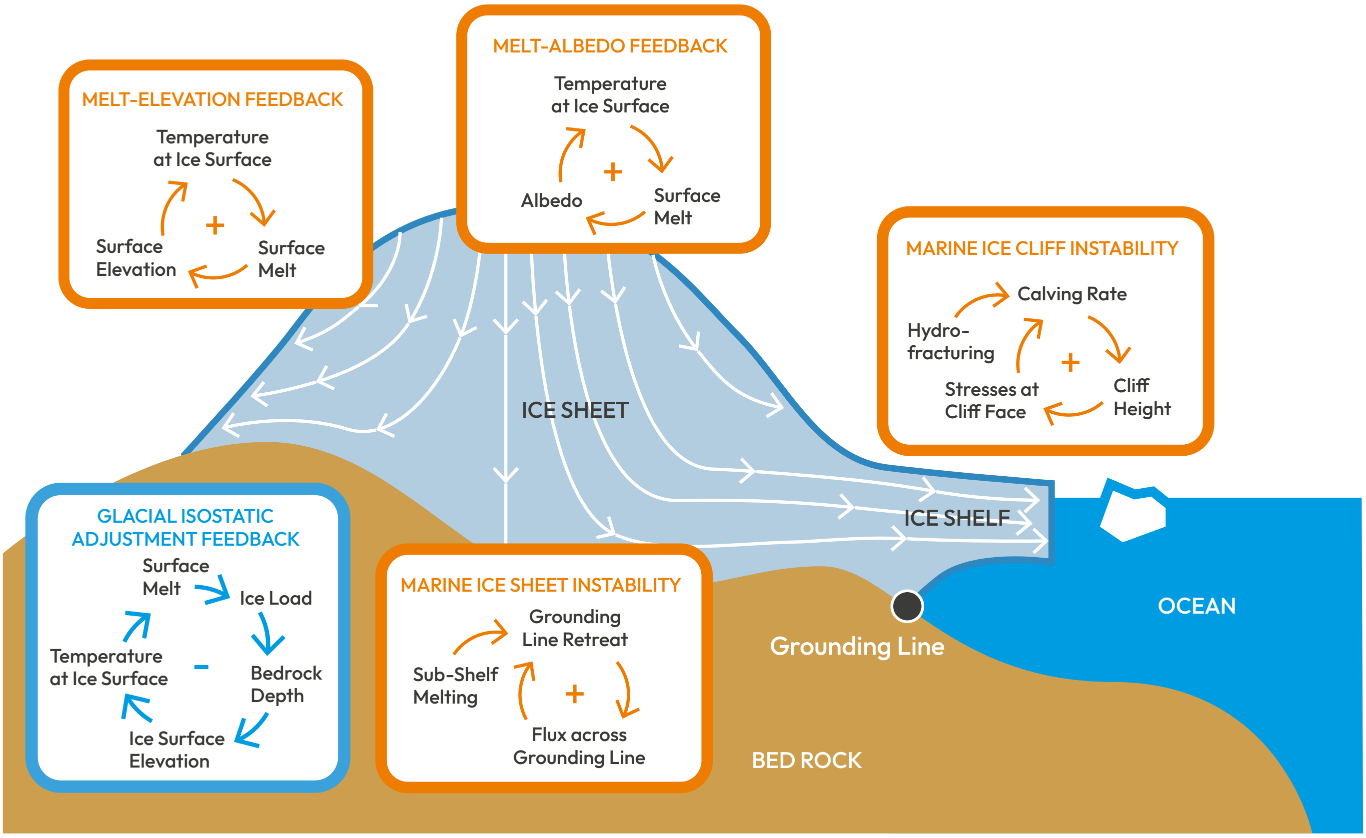

Figure 1.2.5: Schematic illustrating some of the key feedbacks in the ice sheet-climate system. Note that this depiction is limited to the most relevant and widely examined feedbacks, further self-amplifying or damping feedbacks may, however, exist.

On longer timescales (over the course of centuries to millennia), isostatic rebound can also act as a negative/damping feedback on ice sheet retreat (*glacial isostatic adjustment (GIA)*; Whitehouse et al., 2019): a decrease in ice load leads to a slow rebound of the bedrock underneath – as the ice surface is thus lifted to higher elevations with colder surrounding air masses, this can lead to a reduction in surface melt, or even to net accumulation at the surface.

Current estimates for a critical threshold for the GrIS range from 0.8°C to 3°C of warming relative to pre-industrial levels, with a best estimate of about 1.5°C (Robinson et al., 2012; van Breedam et al., 2020; Noël et al., 2021; Höning et al., 2023). This is supported by palaeorecords which indicate that GrIS had at least partially retreated during the MIS-5 interglacial, and likely collapsed during MIS-11, which was 1-2°C warmer than pre-industrial (Christ et al., 2021). At lower warming levels, simulations with a coupled ice sheet atmosphere model indicate that additional atmospheric dynamic changes in precipitation patterns can restabilise the ice sheet, but above 2°C warming, positive/amplifying feedbacks leading to loss of the majority of the ice sheet cannot be overcome (Gregory et al., 2020).

While the respective warming threshold could be reached within the coming decades (Fox-Kemper et al., 2021; Tebaldi et al., 2021), the response times of the ice sheet are such that the ice loss and resulting sea level rise would unfold over several millennia (Robinson et al., 2012; van Breedam et al., 2020). The timescales of ice sheet decline depend on the magnitude of warming beyond this threshold, where stronger warming leads to a faster ice sheet decay (Robinson et al., 2012). Several studies further indicate a strong hysteresis of the GrIS, meaning that substantial ice loss is likely irreversible on multi-millennial timescales (Robinson et al., 2012; Höning et al., 2023).

Slow-onset tipping processes such as ice sheet collapse might also be able to withstand a short period of temperature overshoot if the overshoot time is short compared to the effective timescale of the tipping system (Ritchie et al., 2021). For ice sheets this overshoot time could be in the order of decades to centuries (Ritchie et al., 2021; Bochow et al.,2023), which might for example theoretically allow global warming to overshoot a tipping threshold of 1.5°C and return below it by 2100 without triggering ice sheet collapse (Armstrong McKay et al., 2022). However, such overshoot times are very uncertain, and given the distinct challenges of reducing global temperatures over short time horizons, this possibility should not be relied upon in policy.

Assessment and knowledge gaps

Given the broad evidence base, we have high confidence that the GrIS is a tipping system. This is in line with previous assessments (Fox-Kemper et al., 2021; Armstrong McKay et al., 2022).

West Antarctic Ice Sheet (WAIS)

Since temperatures in Antarctica are generally lower than in Greenland (being centred over the South Pole) and the surface is generally brighter, there is overall less surface melt (Broeke et al., 2023). Recent observations show melt occurrences on ice shelves along the coastline of Antarctica, with most intense melting occuring on the Antarctic Peninsula (Trusel et al., 2013; Jakobs et al., 2020; Lenaerts et al., 2016; Stokes et al., 2019). In contrast to Greenland, however, the currently observed mass loss, especially in the WAIS, is dominated by ocean-induced melting at the underside of the floating ice shelves (e.g., Otosaka et al., 2023; Millilo et al., 2022; Paolo et al., 2015; Adusumili et al., 2020).

Large parts of the WAIS are grounded below sea level (so-called marine basins), surrounded by floating ice shelves, and where these ice shelves are in contact with warmer ocean waters, melting at their base occurs. While the direct contribution to sea level rise of this ice shelf melting is negligible, it plays an important indirect role for the overall mass balance. Due to the thinning of the ice shelves, the buttressing (i.e. the backstress imparted to the grounded ice) is reduced, causing the movement of grounded ice upstream to accelerate, which in turn can lead to substantial sea level rise (Scambos et al., 2004; Rignot et al., 2004; Reese et al., 2018). Substantial ocean warming and ice shelf basal melting is committed in the Amundsen Sea over the 21st Century, which will likely accelerate the retreat of several key WAIS outlet glaciers including the Thwaites and Pine Island glaciers (Naughten et al. 2023).

Evidence for tipping dynamics

Different amplifying feedbacks can lead to self-sustained ice loss from the WAIS once a critical threshold is passed (Figure 1.2.5). One of the key feedbacks is the *marine ice sheet instability* (MISI – Figure 1.2.6, top) (Weertman, 1974; Schoof, 2007; Mengel and Levermann, 2014; Feldmann and Levermann, 2015; Garbe et al., 2020), which can occur where the grounding – the separation line between the grounded ice sheet and floating ice shelves – sits on retrograde bedrock slopes. If the grounding line retreats into regions of greater ice thickness, for instance due to enhanced sub-shelf melting, this increases the flux across the grounding line, leading to further retreat. Such self-sustained retreat may be stabilised by the buttressing effect of ice shelves (Gudmundsson et al., 2012; Pegler, 2018; Haseloff and Sergienko, 2018).

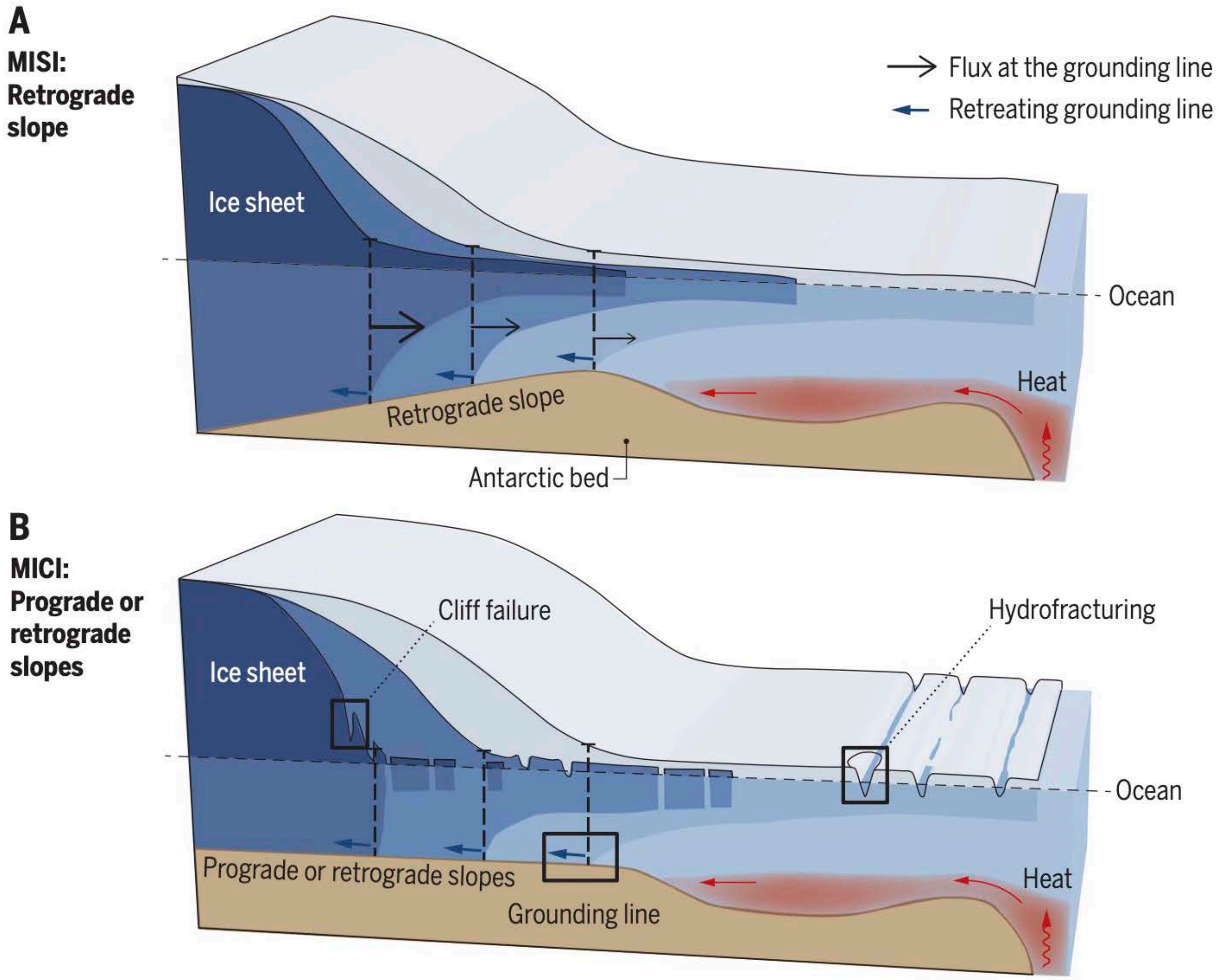

Figure 1.2.6: Schematic illustration of marine ice sheet instability (MISI; top) and marine ice cliff instability (MICI; bottom). From Pattyn and Morlighem (2020) .

The MISI has been suggested to have driven the collapse of WAIS during previous interglacials (Pollard et al., 2015; DeConto and Pollard, 2016; Sutter et al., 2016; Turney et al., 2020; Thomas et al., 2020; Weber et al., 2021). There is also palaeoclimate evidence for a collapse of WAIS and around 20m higher sea level (implying substantial Antarctic Ice Sheet loss) during ~2-3°C warmer periods of the Pliocene, (Naish et al., 2009; Grant et al., 2019; DeConto et al., 2021). It has further been suggested that this instability might already be underway in the Amundsen Sea Embayment, including at the Thwaites and Pine Island glaciers (Rignot et al., 2014; Joughin et al., 2014; Favier et al., 2014; Turner et al., 2017; De Rydt et al., 2021).

While a recent intercomparison study using three different ice sheet models (Hill et al., 2023) concluded that the current observed retreat of grounding lines in West Antarctica is not yet driven by this instability, mounting evidence from modelling studies (e.g., Reese et al., 2023; Seroussi et al., 2017; Arthern and Williams 2017; Golledge et al., 2021; Garbe et al., 2020) suggests that, unless the current warming trend is reversed to colder conditions in the near future, parts of the WAIS such as the Amundsen basin would be committed to long-term irreversible grounding-line retreat driven by MISI. The loss of the Amundsen basin alone would raise global sea levels by roughly 1.2 metres, (Morlighem et al., 2020). Additional large-scale ice sheet changes in West Antarctica could be triggered in the coming decades in response to projected warming. Due to the long response time of the ice sheet, the respective mass loss would unfold and sea level thus keep rising for centuries to millennia (Golledge et al., 2015; Winkelmann et al., 2015).

Another proposed destabilising feedback mechanism is known as *marine ice cliff instability* (MICI – Figure 1.2.6, bottom) (Bassis and Walker, 2012; Bassis and Jacobs, 2013; Pollard et al., 2015; DeConto and Pollard 2016). The MICI hypothesis proposes that tall marine-terminating ice cliffs, which could result from ice shelf collapse, for example, are inherently unstable and could rapidly collapse, potentially associated with a self-reinforcing and irreversible inland ice retreat on both retrograde and prograde sloping marine beds. Such retreat would proceed until water depths shallow or the ice cliff is buttressed (DeConto and Pollard, 2016). The critical height of the ice cliff resulting in its failure depends on the ice properties and the extent of crevassing, but is currently poorly constrained (Bassis and Walker, 2012). In addition, processes potentially mitigating or slowing the self-sustained ice retreat due to MICI such as mélange buttressing or the speed of the preceding ice shelf disintegration introduce additional uncertainties (Clerc et al., 2019; Edwards et al., 2019; Robel and Banwell 2019; Schlemm et al., 2022; Pollard et al., 2018). Low confidence has been assigned to this process in the latest IPCC assessment (IPCC AR6 WG1 Ch9), partially because it has not yet been observed (Needell and Holschuh, 2023).

Assessment and knowledge gaps

Based on these different lines of evidence, there is high confidence that the WAIS is a tipping system, with the potential for widespread, and at least partly irreversible ice loss. Recent estimates of the respective global warming levels at which such tipping dynamics are triggered range from 1°C to 3°C of warming compared to pre-industrial levels (Garbe et al., 2020; Golledge et al., 2017; Reese et al., 2023). This means that the complete decline of the WAIS could be triggered by warming projected under higher-emission scenarios for this century (Chambers et al., 2022; Golledge et al., 2015).

Due to the complexity of interacting processes with the other parts of the climate system and their lack of representation in fully coupled (Earth system) models, it remains a challenging task to reduce the respective uncertainty range and project the resulting ice loss in the near future. For example, the potential effect of ocean stratification or solid-Earth feedbacks on grounding line migration is currently not well-constrained (e.d., Kachuk et al., 2020; Larour et al., 2019; Gomez et al., 2020; Coulon et al., 2021; Golledge et al., 2019). Given the high vulnerability of the WAIS and the far-reaching consequences of its potential collapse, it is important to narrow down the critical thresholds, and in particular the timing of the onset of potential large-scale retreat.

Marine basins East Antarctica

The East Antarctic marine basins include the Wilkes, Aurora and Recovery Basins, and 19.2 metres of sea level equivalent (Fretwell et al., 2013). They have been proposed as 'global core' climate tipping systems, due to the potential for instabilities in the marine ice sheet and ice cliff (Garbe et al., 2020; Armstrong McKay et al., 2022). The processes affecting the marine basins of East Antarctica are thus similar to those described above for the WAIS.

Evidence for tipping dynamics

Outlet glaciers in the Aurora subglacial basin, for instance Totten and Denman glaciers, already experience acceleration, retreat and mass loss at present (e.g., Rignot et al., 2019; Shepherd et al., 2019; Rintoul et al., 2016; Li et al., 2015, 2016; Miles et al., 2021; Shen et al., 2018). There is limited evidence for change in Recovery and Wilkes basins in current observations (e.g., Gardner et al., 2018). However, palaeorecords and models suggest the ice margin may have undergone substantial retreat deep inland of Wilkes subglacial basin during Pleistocene interglacials (Blackburn et al., 2020; Wilson et al., 2018; Iizuka et al., 2023) and in warm periods of the Pliocene (Cook et al., 2013; DeConto et al., 2021; Blasco et al., 2023 [in review]) with global mean atmospheric warming of at least 1-2°C above pre-industrial, as suggested by palaeorecords (Blackburn et al., 2020). Other work has suggested that ice sheet retreat in the Wilkes subglacial basin remained relatively limited during the Last Interglacial, when Southern Ocean sea surface temperatures were about 1-2°C and Antarctic surface air temperatures were at least 2°C above pre-industrial averages (Capron et al., 2017; Hoffman et al., 2017; Chandler and Langebroek, 2021), placing an upper sea-level contribution from the Wilkes basin during that period at 0.4-0.8 m (Sutter et al., 2021).

Recent model simulations show that the risk of substantial sub-shelf melt-induced or calving-induced ice loss and the associated timescales vary strongly for the individual subglacial basins (Garbe et al., 2020): A drainage of the Recovery basin may be driven by oceanic warming of 1-3°C (Golledge et al., 2017), while self-sustained grounding-line retreat in the Wilkes basin is initiated in models when exceeding an atmospheric warming of 2-4°C above present-day levels (Garbe et al., 2020; Golledge et al., 2017). The decay of the drainage basin may occur over a time period of centuries to tens of thousands of years, as indicated in palaeorecords (Bertram et al., 2018) and model experiments (Mengel and Levermann, 2014), depending on the warming trajectory (DeConto and Pollard, 2016). Modelling studies suggest that ice loss from the Aurora subglacial basin is triggered when sustaining stronger warming of about 5-8°C above present-day levels (Garbe et al., 2020; Golledge et al., 2017; Winkelmann et al., 2015; Bulthuis et al., 2020; Van Breedam et al., 2020; Golledge et al., 2015). Palaeo evidence and models suggest that, once triggered, ice loss from these marine basins can only be reversed if the climate were to cool far below pre-industrial levels, leading to hysteresis behaviour (Garbe et al., 2020; Mengel and Levermann, 2014).

Assessment and knowledge gaps

Being characterised by self-sustained dynamics as well as abrupt and irreversible changes beyond a warming threshold in various studies, we identify the marine basins of East Antarctica as parts of the cryosphere exhibiting tipping behaviour with high confidence, in line with previous assessments (Armstrong McKay et al., 2022). Further work is needed to better constrain existing estimates of critical thresholds and timescales from available ice sheet modelling and palaeoclimate data for individual subglacial basins – for example, by improving the treatment of sub-shelf melt and taking into account model and parametric uncertainty.

Non-marine East Antarctica

In East Antarctica, a major part of the ice sheet initially built up at the Eocene-Oligocene transition is grounded above sea level (DeConto and Pollard, 2003; Liu et al., 2009; Morlighem et al., 2020; Hutchinson et al., 2021). At present, observations still indicate mass gain in this terrestrial part of the Antarctic Ice Sheet (for instance, in Dronning-Maud Land) though mass balance estimates are associated with high uncertainties (Otosaka et al., 2023; Schröder et al., 2019). As such, West Antarctic ice loss over the past decades was balanced to some extent by mass accumulation in East Antarctica (Medley and Thomas, 2019).

Evidence for tipping dynamics

Long-term model assessments suggest that large-scale ice loss from terrestrial regions of East Antarctica may be induced for global mean atmospheric warming of 6°C or higher above pre-industrial levels (Garbe et al., 2020) until East Antarctica potentially becomes completely ice-free. Given the wide range of warming projected in the recent sixth phase of the Coupled Model Intercomparison Project (CMIP6), exceedance of respective critical forcing levels cannot be excluded beyond the end of this century under high emissions (e.g. SSP5-8.5 and SSP3-7.0 in the 22nd century; IPCC AR6 WG1 Ch4) in combination with a high climate sensitivity (Tebaldi et al., 2021). The disintegration of the land-based portions of the East Antarctic Ice Sheet may eventually raise global mean sea level by ~34 m (Fretwell et al., 2013), but unfolding over multi-millennial timescales (~10,000 years or longer) according to modelling studies (Winkelmann et al., 2015; Clark et al., 2016).

Here, the *melt-elevation feedback* (similar to the GrIS) propels self-sustained mass loss by enhancing surface melt once the respective tipping point is crossed. It also gives rise to pronounced hysteresis behaviour with distinct stable ice sheet configurations within a range of climatic boundary conditions (Garbe et al., 2020; Pollard and DeConto, 2005; Huybrechts 1994). A strong cooling is consequently required for regrowth of the terrestrial East Antarctic Ice Sheet, and sustained cooling to at least pre-industrial temperature levels to recover its present-day volume and extents (Garbe et al., 2020). Due to this hysteresis, large land-based portions of the East Antarctic Ice Sheet persisted for more than 8 million years (Shakun et al., 2018) through the warm intervals of the early to mid-Miocene, 23-14 million years ago (Gasson et al., 2016; Levy et al., 2016).

Assessment and knowledge gaps

Self-amplifying feedback mechanisms (such as the melt-elevation feedback) can occur in East Antarctica, contributing to abrupt and irreversible ice sheet changes with a substantial impact through sea level rise beyond a critical threshold. There are few modelling studies on multi-millennial timescales covering the warming range that may be relevant for the potential nonlinear response of the terrestrial ice sheet in East Antarctica.

Thus, there is medium confidence in the assessment of the non-marine East Antarctic Ice Sheet as a cryospheric tipping system. Reducing uncertainties in temperature thresholds and timescales of collapse requires multi-model ensembles and better representation of ice surface processes, as well as the inclusion of interaction with the rest of the climate system. Additionally, more research on how climate forcing varies regionally and interacts with regional processes and feedbacks would help better constrain the drivers and timescale of tipping.

1.2.2.2 Sea ice

Sea ice is frozen sea water that floats on the sea surface. It forms in the polar oceans whenever the temperature of the sea water drops below its freezing point of around -1.8°C. The formation and growth of sea ice therefore requires a sufficient heat loss from the ocean to the atmosphere, which in today's climate occurs in both polar regions from autumn to spring. During this period, sea ice is expanding, while during summer it is retreating.

While the formation of sea ice through heat loss to the atmosphere is similar in both polar regions, the dominating process for sea ice decay in summer differs between the two hemispheres. In the North, where the sea ice is largely landlocked by the land masses surrounding the pole, the loss of sea ice is primarily driven by atmospheric heat input that melts the sea ice. In the southern hemisphere, however, the summer loss of sea ice is primarily governed by the export of sea ice through northward winds that move the ice into regions of warmer sea water, which then melts the ice from below. The freeze-melt cycle of sea ice gives rise to substantial seasonal variations in the polar sea ice coverage (Figure 1.2.7 and Figure 1.2.9), whose magnitude is an indicator for the very fast response time of sea ice, in particular relative to other cryospheric systems such as permafrost, glaciers and ice sheets.

Given the different processes that are relevant for the regional and seasonal response of sea ice to global warming, in the following we differentiate our assessment of tipping potential between Arctic summer sea ice, Arctic winter sea ice, Barents Sea ice, and Southern Ocean sea ice.

Arctic summer sea ice

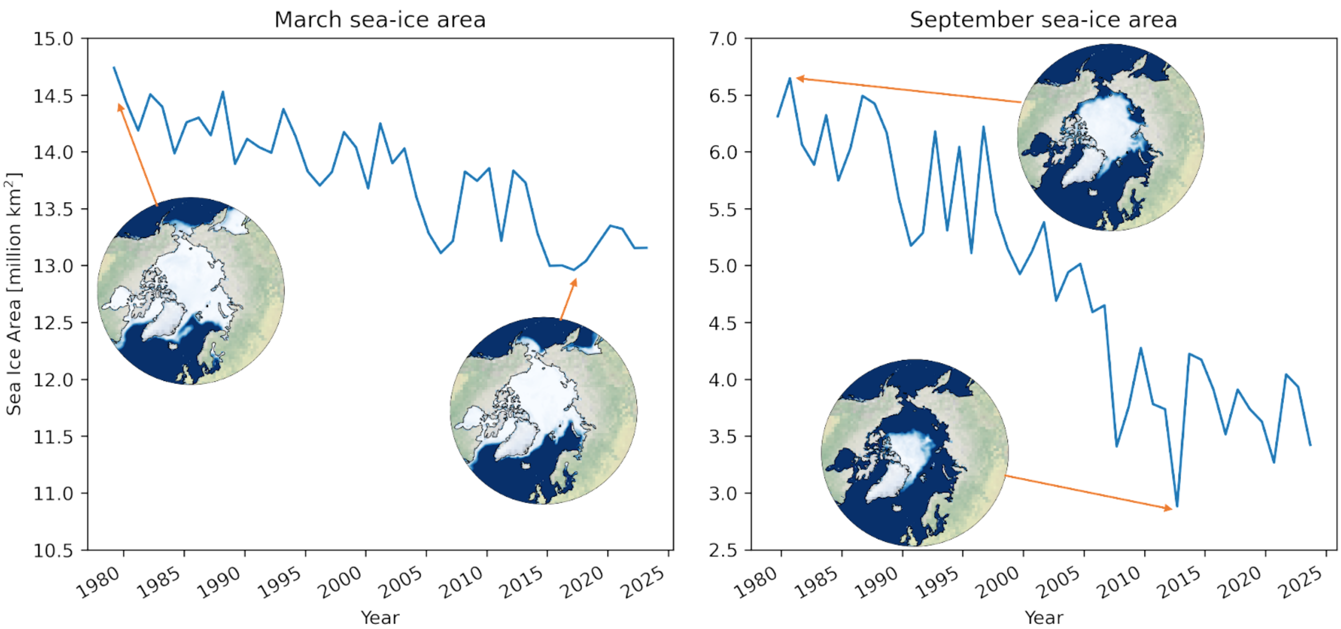

Figure 1.2.7: Arctic sea ice evolution 1979-2023. Time series of Arctic sea ice area, with insets showing sea ice concentration in selected years. March is usually the month of maximum sea ice area ('winter sea ice'), September is usually the month of minimum sea ice area ('summer sea ice'). Data: OSI SAF (Lavergne et al. 2019) [time series: OSI SAF Sea ice index 1978-onwards (v2.2 2023); sea ice concentration before 2020: OSI SAF Global sea ice concentration climate data record 1978-2020 (v3.0, 2022); sea ice concentration after 2020: OSI SAF Global sea ice concentration interim climate data record (v3.0, 2022)].

Evidence for tipping dynamics

In summer, the retreating sea ice cover in the Arctic exposes the much darker ocean surface to the atmosphere, giving rise to the ice-albedo feedback: Less ice implies an additional uptake of heat, implying further ice loss. This mechanism was hypothesised to give rise to a nonlinear tipping point behaviour for the loss of Arctic summer sea ice (e.g., Lenton et al., 2008).

However, a large variety of studies based on both conceptual models and coupled Earth system models have provided convincing evidence that the summer ice-albedo feedback is compensated by damping feedbacks in winter that minimise the long-term memory of the Arctic summer sea ice cover (Figure 1.2.8). This dominance of negative/damping feedbacks gives rise to a linear retreat of the Arctic summer sea ice cover with ongoing global warming (e.g., Gregory et al., 2002; Winton, 2006; Winton, 2008; Notz, 2009; Tietsche et al., 2011; Mahlstein and Knutti, 2012; Wagner and Eisenman, 2015).

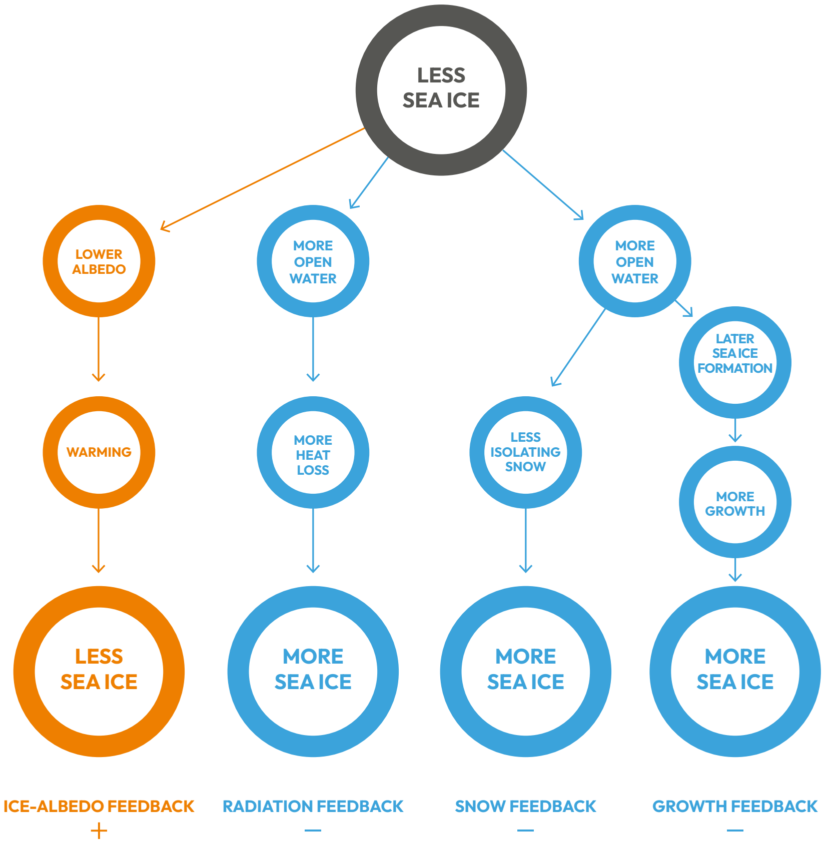

Figure 1.2.8: Schematic illustrating some of the key feedbacks related to Arctic sea ice loss. Note that this depiction is limited to the most relevant and widely examined feedbacks; further self-amplifying or damping feedbacks may, however, exist. Based on Notz and Bits (2016).

Based on this understanding, the response of the sea ice cover to global warming is expected to remain linear as a function of global mean temperature (e.g., Gregory et al., 2002; Winton, 2011; SIMIP 2020) and thus as a function of CO_2 emissions (Zickfeld et al., 2012; Notz and Stroeve, 2016) until the complete loss of the summer sea ice cover that is expected to occur for the first time before 2050 in all future climate scenarios (SIMIP, 2020; Kim et al., 2023). If, in the future, atmospheric CO_2 were to decrease, for example by the technological removal of CO_2, there would be some time lag before global temperature would decrease in response. This hysteresis then carries over to the relationship between CO_2 concentration and sea ice area. The relationship between sea ice area and hemispheric mean temperature, however, has been found to remain linear also for a cooling climate (e.g., Armour et al., 2011; Li et al., 2013; Jahn, 2018).

Assessment and knowledge gaps

The assessment of a linear, threshold-free loss of Arctic summer sea ice is in line with recent assessments (Fox-Kemper et al., 2021; Armstrong McKay et al., 2022). Given the very broad evidence base, we have high confidence in the assessment of Arctic summer sea ice not being a tipping system. This confidence could be increased further if climate models would more reliably capture the observed evolution of the Arctic sea ice cover – for example regarding its linear sensitivity to observed global warming (SIMIP, 2020). A comprehensive assessment of climate model performance is, however, hampered to some degree by the difficulty to obtain reliable, long-term observations of the sea ice thickness distribution (SIMIP, 2020). Some progress in this regard can be expected in the near future, with the recent development of an approach to retrieve sea ice thickness throughout the entire seasonal cycle using remote sensing (Landy et al., 2022).

Arctic winter sea ice

For the loss of summer sea ice, the existing ice cover needs to be melted completely, which is a gradual process. The loss of winter sea ice, however, is governed by a different mechanism: given that the Arctic will already be ice-free in summer, the formation of new ice needs to become impossible to lose the winter sea ice cover. Winter sea ice will form in the Arctic Ocean as long as the water temperature at the ocean surface drops below the freezing point – around -1.8°C for typical saline ocean water – but will no longer form once the water temperature remains above freezing all year round. This binary behaviour of the Arctic Ocean lies at the heart of the analysis of the ongoing loss of the Arctic winter sea ice cover.

Evidence for tipping dynamics

Both in some simple models and in some complex climate models, the loss of Arctic winter sea ice area accelerates drastically once a given warming threshold has been reached (e.g., Winton, 2006; Eisenman and Wettlaufer, 2009; Bathiany et al., 2016). However, this acceleration is simply a consequence of the geometry of the Arctic Ocean: as the climate warms, the winter sea ice edge moves northward. As long as the ice edge is located in the narrow straits that connect the Arctic Ocean to the south, the freely moving ice edge is short and only a little ice is lost by its northward movement. Once the ice edge becomes located in the central Arctic Ocean, more sea ice area is lost for a given retreat of the ice edge, and ice loss accelerates. This acceleration therefore occurs in most models as soon as the winter maximum sea ice area drops below around 8m sq km, which is roughly the area of the Arctic Ocean and its adjacent seas (Goosse et al., 2009; Eisenman, 2010).

Beyond this threshold, the loss of the winter sea ice cover occurs faster than the loss of the summer sea ice in CMIP5 models. This can be explained by the fact that the future formation of winter sea ice from a largely ice-free ocean will lead to a geographically rather homogenous distribution of winter sea ice thickness, such that larger areas can become ice-free simultaneously (Bathiany et al., 2016).

In modelling studies, the faster loss in winter compared to summer has additionally been found to be related to the increased humidity and the related increased downward longwave radiation, for example from convective clouds in areas of open water (Abbot and Tziperman, 2008; Abbot et al., 2009; Li et al., 2013; Hankel and Tziperman, 2021). While this process could potentially imply hysteresis behaviour of the loss of Arctic winter sea ice, the loss of winter sea ice has been shown to be fully reversible in a number of dedicated modelling studies (Armour et al., 2011; Ridley et al., 2012; Li et al., 2013). In particular, for a cooling of the climate induced by the removal of CO_2, studies have found no hysteresis of Arctic winter sea ice area as a function of hemispheric mean temperature, while they found a time lag between the decrease of atmospheric CO2 concentration and the resulting increase of Arctic winter sea ice area. This can be explained by the delayed response of atmospheric temperature to the removal of CO_2, and the potential nonlinear response of oceanic heat transport (Li et al., 2013; Schwinger et al., 2022).

Assessment and knowledge gaps

Based on this assessment, there is currently only very limited support for a dominating role of self-perpetuating processes that would make Arctic winter sea ice a tipping system. Given the difficulty of climate models to realistically simulate the processes that govern the loss of winter sea ice and the related oceanic response, we have medium confidence in the assessment of Arctic winter sea ice not being a tipping system.

Barents Sea ice

Sea ice in the Barents Sea – the sector of the Arctic Ocean north of Scandinavia and Western Russia – is treated as a sub-case of Arctic winter sea ice in Armstrong et al., (2022), who categorised it as a regional impact climate tipping system with medium confidence.

Evidence for tipping dynamics

In the Barents Sea, which is only ice-covered in winter, sea ice loss is primarily driven by an increase in lateral oceanic heat inflow of warm Atlantic water (Docquier et al., 2020; Smedsrud et al., 2021; Muilwijk et al., 2023). Because of this tight coupling, in almost all models the sea ice loss is largely linearly related to changes in oceanic heat transport (Docquier et al., 2020) with only one model showing an abrupt loss of the Barents Sea sea ice cover in winter in a dedicated study (Drijfhout et al., 2015). The loss of the Barents Sea winter sea ice cover might reinforce itself through related changes in atmospheric circulation, but there is no consensus among studies that examined these linkages (e.g., Haarsma et al., 2021; Smith et al., 2022 and references therein). The sea ice loss could also reinforce itself through a related increase in the inflow of warm Atlantic water (Lehner et al., 2013) but very few studies have examined this in detail.

Assessment and knowledge gaps

In summary, there is currently no clear support for the Barents Sea winter sea ice cover being a tipping system. We have low confidence in this assessment, given the very low number of respective studies.

Southern Ocean sea ice

In the Southern Ocean, the amount of sea ice is much more dominated by the combination of oceanic and atmospheric processes than in the Arctic, which gives rise to a much more pronounced seasonal cycle of the Antarctic sea ice area compared to the Arctic (Figure 1.2.9). Generally, the area of sea ice in the Southern Ocean is determined by the balance of ice formation near the continent and ice melt through oceanic heat further away from the coast, where the ice is advected by the prevailing winds and currents. Variations in ice coverage can therefore largely be explained by weaker northward transport of the ice, by increased melting from increased upward oceanic heat transport, and/or by weakened ice formation (e.g., Maksym, 2019). The regional distribution of sea ice growth with its related brine release, and sea ice melt with the release of freshwater, in turn affects the stratification and circulation of the Southern Ocean (see Chapter 1.4 and e.g., Abernathey et al., 2016).

Over the full satellite record from 1979 onwards, there is no significant trend in Antarctic sea ice coverage (e.g., Fox-Kemper et al., 2021). The maximum sea ice coverage of the observational record was recorded in 2014, while the minimum sea ice coverage was recorded in 2022/2023 (Figure 1.2.9). The low ice coverage of the past two years can be linked to changes in the prevailing wind patterns that are caused by changes in the prevailing large-scale atmospheric modes (e.g., Zhang and Li, 2023; Wang et al., 2023), and 2023's historic low has been suggested to represent a new low ice regime resulting from ocean warming (Purich and Dodderidge, 2023). However, given the shortness of the signal, it is currently unclear whether this change in the sea ice forcing will persist, which then could cause a significant, long-term decline of the Antarctic sea ice cover.

Evidence for tipping dynamics

Given the very long response time of the Southern Ocean to climatic changes, and given the potential long-term changes in the Southern Ocean circulation in response to irreversible changes in ice sheet dynamics, hysteresis behaviour can be expected to exist for the long-term loss of Southern Ocean sea ice. Such hysteresis is indeed identified in a number of dedicated studies (Ridley et al., 2012; Li et al., 2013), but is explained by a lagged response of the sea ice cover to the imposed warming and cooling. This dynamic hysteresis behaviour is therefore a consequence of the long response time of the Southern Ocean. Whether or not one considers this behaviour truly hysteretic is a question of the timescales of relevance.

Assessment and knowledge gaps

There is currently limited evidence for a self-amplification of Southern Ocean sea ice loss, and we cannot estimate a related temperature threshold. We have low confidence in the assessment of the future evolution of Antarctic sea ice given the difficulties of large-scale climate models to reproduce its observed evolution. This shortcoming of the models might be related to the dominating impact of small-scale eddies in the ocean which low-resolution climate models cannot explicitly resolve. Another shortcoming is the current absence of reliable satellite retrievals of Southern-Ocean sea ice thickness that would be crucial for a detailed model evaluation. This is expected to be addressed with new satellite technologies including, for example, the Surface Water and Ocean Topography (SWOT) mission (Armitage and Kwok, 2021).

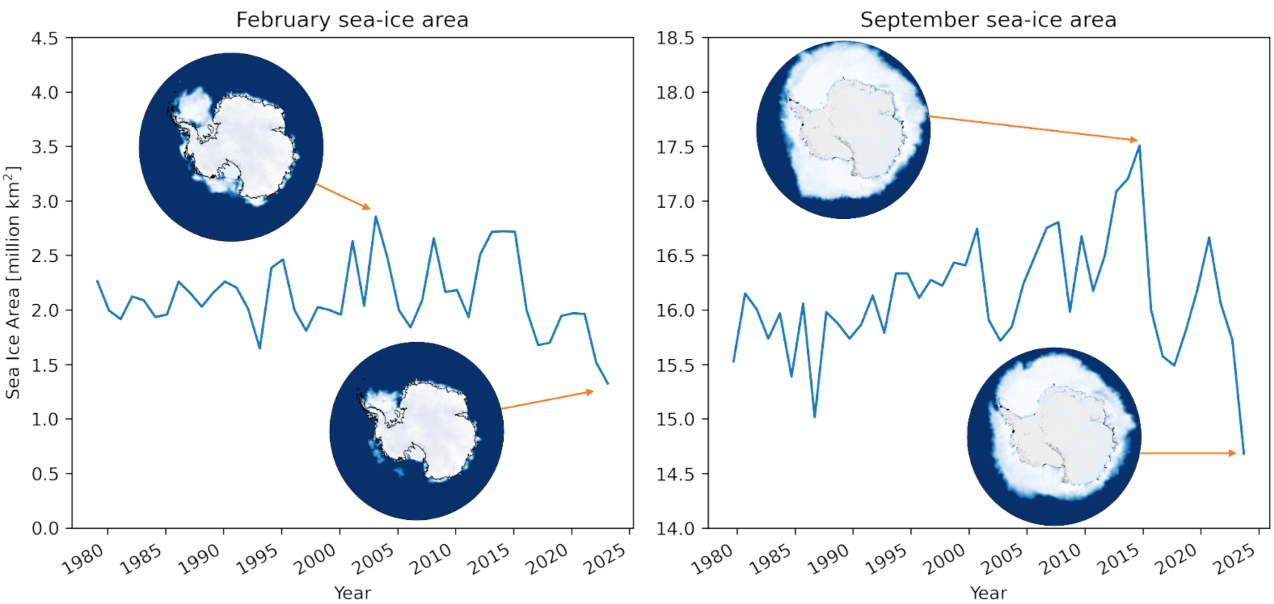

Figure 1.2.9: Antarctic sea ice evolution 1979-2023. Time series of Antarctic sea ice area, with maps showing sea ice concentration in selected years. February is usually the month of minimum sea ice area ('summer sea ice'). September is usually the month of maximum sea ice area ('winter sea ice'). Data: OSI SAF (Lavergne et al. 2019) [time series: OSI SAF Sea ice index 1978-onwards (v2.2 2023); sea ice concentration before 2020: OSI SAF Global sea ice concentration climate data record 1978-2020 (v3.0, 2022); sea ice concentration after 2020: OSI SAF Global sea ice concentration interim climate data record (v3.0, 2022)].

1.2.2.3 Glaciers

Glaciers outside the Greenland and Antarctic ice sheets (here termed mountain glaciers) are spread over high altitudes and high latitudes. A range of processes contribute to their individual mass balances, most notably solid precipitation (mainly snow) and surface melt, but also, among others, calving into lakes or ocean (Hock et al., 2019; Meredith et al., 2019). Mass balance thresholds and feedbacks may impact individual glaciers, but when aggregated to the global scale glacier changes are projected to respond relatively linearly this century (Rounce et al., 2023). At longer timescales and higher warming levels, nonlinear characteristics are projected as glaciers disappear (Marzeion et al., 2018).

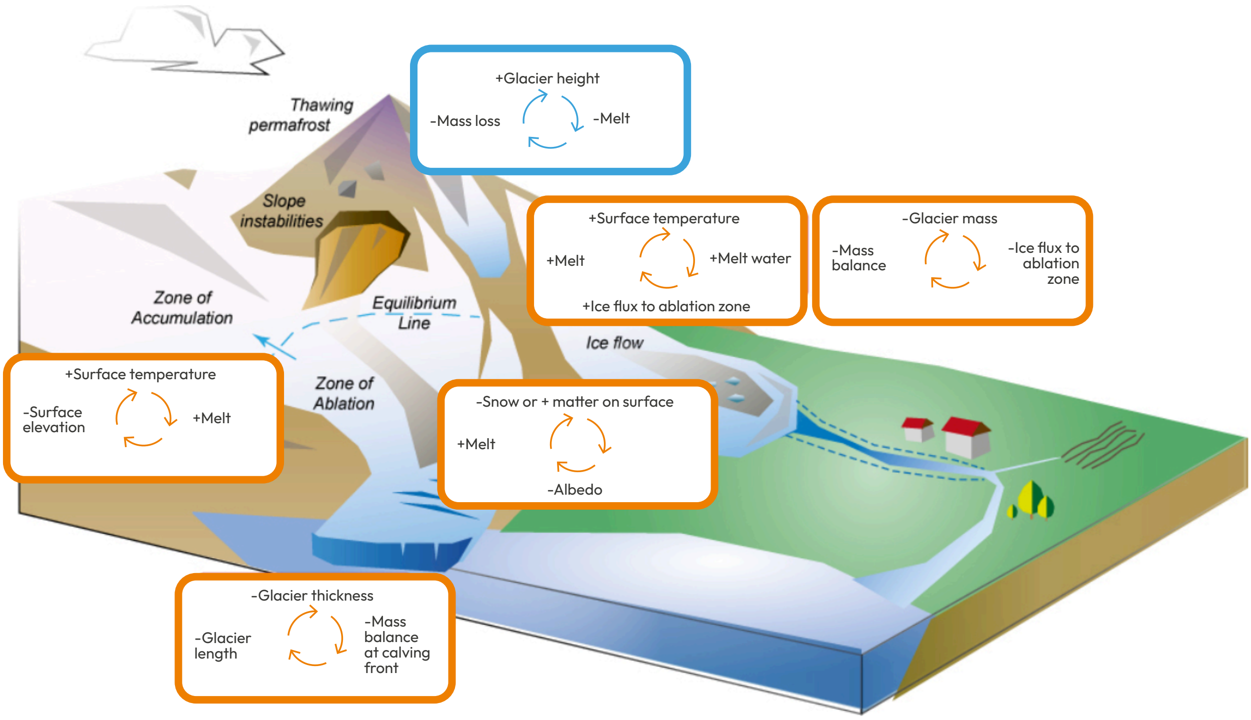

Figure 1.2.10: Terminology and some of the key feedbacks related to mountain glacier retreat. 'Positive' amplifying feedbacks that amplify ice loss are shown by red boxes, and 'negative' damping feedbacks that limit ice loss are shown by blue boxes. Above the equilibrium line (dashed blue line) glaciers accumulate snow and therefore mass, and below they 'ablate' – i.e. melt and lose moss. Note that this depiction is limited to the most relevant and widely examined feedbacks – further self-amplifying or damping feedbacks may, however, exist.

Evidence for tipping dynamics

In glaciers close to the melting point, the physical nature of ice inherently involves nonlinear feedbacks, in particular related to interactions between ice and water such as enhanced subaqueous ice melt, heat transport into the ice, or lubrication at the glacier bed. Such feedbacks act typically on the spatial scale of individual glaciers (Figure 1.2.10).

Dynamic instabilities of glaciers such as surges or even catastrophic detachments, but also less pronounced ice velocity fluctuations, can be related to increased melt-water production through positive/amplifying feedback mechanisms (Truffer et al., 2021; Kääb et al., 2021). However, these processes are still not very well understood and there is little evidence so far indicating that such processes could act synchronously over entire glacier regions (Kääb et al., 2023). On a regional scale, loss of ice thickness appears to rather reduce glacier flow speeds (Dehecq et al., 2019). Significantly increased ice flux, such as through surges, transports ice from high-elevation zones characterised by low rates of ice melt (ablation) to low-elevation zones with high ablation rates.

In contrast, retreat rates of calving glaciers, most of them found in polar regions, are understood to be governed by a feedback where a thinning of the glacier tongue (the narrow floating part of a glacier extending into the sea or a lake) leads to loss of glacier grounding at a topographic pinning point (places where a ridge or valley narrowness slows down glacier flow). This loss of pinning leads to accelerated glacier retreat, associated with increased ice flow velocities, calving rates and further thinning of the tongue, until they stabilise again at a new pinning point or retreat out of the water (Strozzi et al., 2017; Kochtitzky et al., 2022a). Once a destabilisation threshold is passed through processes at the ice-ocean or ice-atmosphere interface, the retreat phase is largely self-perpetuating, independent of climatic conditions or their changes (Pfeffer, 2007). In turn, calving glaciers need typically substantial positive mass balances in order to advance through deep water to a new pinning point. Nonlinear enhanced retreats of calving fronts can be roughly synchronised on regional levels and are in fact a significant component of the current mass loss of polar glaciers, roughly 20-25 per cent (Kochtitzky et al., 2022b).

Glaciers impact atmospheric conditions at their surface by increasing local surface altitude, enabling a feedback between surface elevation and mass balance. Ice thinning can drop glaciers into higher melt ('ablation') zones, while a rise in the equilibrium line altitude (ELA – the elevation where local mass balance, i.e. snow input versus melt output, is zero) can shift glaciers into lower snow accumulation zones, with both potentially leading to disproportionately large shifts when large areas of glacier are concentrated in narrow elevation bands. These elevation feedbacks could possibly be regionally synchronised at similar global warming levels, for instance for Arctic ice caps. These effects are typically included in regional and global glacier mass balance models and thus in projections (Rounce et al., 2023; Marzeion et al., 2020).

Reduced glacier albedo, for instance from deposition of dust, black carbon or thin debris, but also through reduced snow cover, significantly increases glacier mass loss (Cook et al., 2017; Naegeli and Huss 2017). Related mass balance feedbacks can happen when years with particularly negative mass balance lead to enhanced accumulation of albedo-reducing matter on the glacier surface, enhancing in turn glacier ablation (Gabbi et al., 2015). Another type of positive/amplifying feedback is deposition of wind-driven dust originating from adjacent mountain areas, a process that is believed to increase with continued uncovering of glacial sediments from ice and snow. Such feedbacks involving albedo can be assumed to affect nearby glaciers in similar ways, and thus represent potential regional effects that are not included in large-scale models yet.

On local scales, abrupt permafrost thaw processes creating 'thermokarst' features (see 1.2.2.4) can be self-perpetuating by enhancing the ice melt in particular of low-angle glacier tongues with low ice flow speeds. Such processes particularly impact debris-covered glaciers, most prominently through the growth of supraglacial ponds on them. There is evidence that such thermokarst processes can enhance glacier ablation on regional scales (Kääb et al., 2012; Buri et al., 2016, Compagno et al., 2019).

Glacier shrinkage has a range of local to global effects. Several types of glacier hazards can increase in frequency and magnitude as a consequence of glacier retreat, such as debris flows or rock slides and rock avalanches (Hock et al., 2019). Slope instabilities and the uncovering of formerly ice-covered areas leads to increased mobilisation of sediments with both negative (e.g. sedimentation of river infrastructure) and positive (e.g. release of nutrients) downstream impacts. Also the formation of glacier lakes, and thus the potential for glacier lake outbursts, is associated with glacier retreat (Carrivick and Tweed 2016; Linsbauer et al., 2016).

Changes in glacier river runoff can have impacts on ecosystems (Bosson et al., 2023) and humans, in particular where dry-season water supply is to a large extent depending on glacier ablation. Whereas peak water – the shift from increased runoff from enhanced glacier melt to reduced runoff under continued shrinking of glacier ares – constitutes on regional scales a soft decadal-scale transition rather than a threshold (Huss and Hock 2018), drastic declines of dry-season glacier melt runoff can exert strong pressure on ecosystems, hydropower production and irrigation, for example (Hock et al., 2019). It is important to note that the significance of glacier runoff for downstream areas depends on the seasonally variable percentage of glacier runoff in comparison to other sources of runoff, such as liquid precipitation or snow melt (Kaser et al., 2010). Measurements and projections of glacier mass loss alone are thus only meaningful in relation to potential impacts as part of a seasonally resolved hydrological balance. On longer time-scales and regional spatial scales, pronounced regional glacier shrinkage (or even partial disappearance of glaciers) leads to a transition from glacier-dominated to paraglacial landscape systems, with fundamental changes in all abiotic and biotic processes in the region and its downstream areas (Knight and Harrison, 2016).

Such a transition to a paraglacial landscape system may exhibit threshold-like behaviour, if climate change is happening rapidly relative to glacier response times, which can span from decades to centuries (Jóhannesson et al., 1989, Haeberli and Hoelzle, 1995). The lagged response of glaciers can lead to a substantial disequilibrium between glacier extent and concurrent climate conditions, such that a large part of a glacier's mass is committed to be lost, even though this loss has not yet been realised. On the global scale, the committed mass loss for present-day glaciers is estimated around 30 per cent (Bahr et al., 2009, Mernhild et al., 2013, Marzeion et al., 2018), but regionally it can be substantially higher (~60 per cent in central/northern Europe and ~50 per cent in western Canada/US).

Sea level contribution represents the most global but also most integrating consequence of global glacier mass loss and does not show threshold behaviour because any positive/amplifying feedbacks acting at the glacier or regional scale are averaged out in the huge ensemble of individual glaciers (c. 200,000) (Hock et al., 2019, Marzeion et al., 2020, Hugonnet et al., 2021).

Assessment and knowledge gaps

Glacier shrinkage involves a number of nonlinear, self-perpetuating processes that mostly act on local scales. Few of these feedbacks seem to be able to reach magnitudes and regional synchronisations substantial enough to enhance regional glacier shrinkage in a nonlinear way. However, the potentially large disequilibrium between glacier extent and concurrent climate implies that, regionally, glaciers may be synchronously transitioning from one state to another, even if the individual glaciers' tipping points are distributed over a broad temperature range. Such effects might explain the almost synchronous retreat of Arctic tidewater glaciers (Kochtitzky et al., 2022a, Malles et al., 2023). Elsewhere, glacier shrinkage is mostly a reversible response to climatic change, despite the irreversible changes that may happen on local scales, such as glacier-related slope failures.

Glaciers can recover from mass loss, but may need much more time for recovery than for melt. Reversibility of biophysical or social downstream effects of glacier shrinkage also requires long timescales (Hock et al., 2019). It is also important to note that a number of negative damping feedbacks are involved in glacier response to atmospheric warming – most importantly the retreat of glaciers to higher elevations, where they experience lower melt rates, or the thickening of insulating debris covers related to increased production of debris associated with reduced ice cover and permafrost on adjacent mountain flanks (e.g., Compagno et al., 2022). We assess with medium confidence that, while glaciers are not tipping points on a global scale, at a regional scale they may be subject to self-sustained retreat tipping points.

A number of the aforementioned glacier feedback processes are not, or not adequately, represented in numerical models. This limitation of models is motivated by the complexity of the processes and the lack of ability to resolve the relevant local scale in the atmosphere and ocean models providing the boundary conditions for the glacier models. The current regional or global glacier projections are struggling to predict the integrated behaviour of local feedbacks and their interactions accurately, and the thresholds and timescales at which slow but nonlinear associated responses of glaciers might emerge are not well known. First results from recent advances in the representation of local feedbacks indicate so far that also in the future the positive/amplifying feedbacks are mostly relevant at the local scale, hardly affecting regional and global scale projections (Compagno et al., 2022, Malles et al., 2023).

1.2.2.4 Permafrost

Permafrost is defined as ground frozen for at least two consecutive years (Van Everdingen, 2005) (Figure 1.2.11). Permafrost underlies about 14 million sq km (15 per cent of the land surface area) in the Northern Hemisphere (Obu, 2021), mainly in Russia, Canada, the US (Alaska), and China (Tibetan Plateau). In addition, there is about 2.5 million sq km of relict permafrost in the Arctic shelf seafloor (Overduin et al., 2019), which was submerged by rising sea levels at the end of the ice age.

Figure 1.2.11: Thawing coastal permafrost in Arctic Canada, with person for scale. Credit: G. Hugelius, taken from Pihl et al., 2021

Permafrost landscapes are complex. They commonly exhibit an active layer, which is the uppermost layer of soil or ground that thaws during the warmer months of the year and freezes again during colder months (Figure 1.2.12). Permafrost is further characterised by factors such as variable topography, ground ice presence, vegetation dynamics, and soil climatic conditions. For example, the presence of hills, valleys and slopes affects the distribution and characteristics of continental permafrost at different spatial and temporal scales. The interaction and feedback between these factors contribute to the complexity of permafrost environments and suggest a variety of potential responses of the permafrost domain to climatic changes.

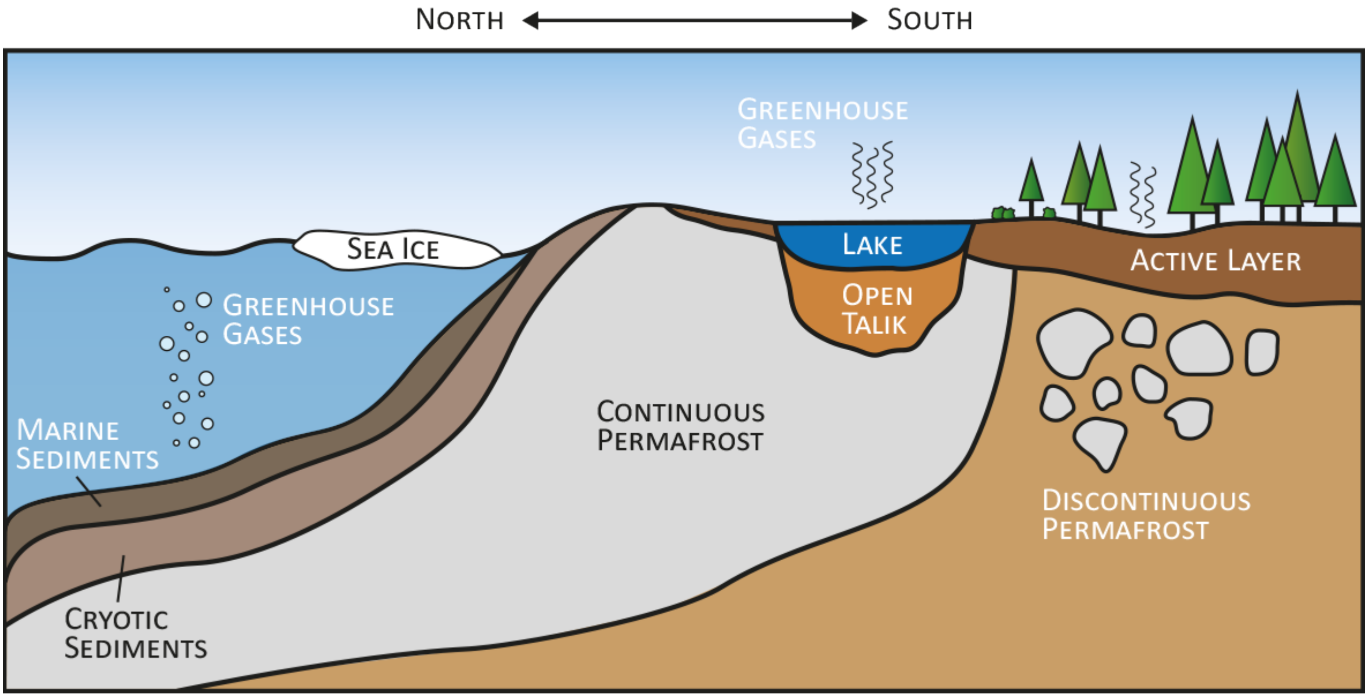

Figure 1.2.12: Schematic showing typical permafrost landscape features. Inspired by Lantuit et al., (2012).

Vast amounts of organic carbon and ground ice that accumulated during past cold climates in northern high latitudes are still preserved in permafrost today. The frozen conditions in permafrost soils prevent the microbial decomposition of organic material accumulated in the past during relatively warm summers. Currently, it is estimated that the upper three metres of permafrost soils contain about 1,035 ±150 GtC (Hugelius et al., 2014) or about 50 per cent more than today's atmosphere (Figure 1.2.13). Subsea permafrost stores additional organic matter estimated at between 560 (Sayedi et al., 2020) and 2,822 (1,518-4,982) GtC (Miesner et al., 2023). Further, permafrost also contains or caps large quantities of frozen methane and other gases. Such deposits are known as permafrost-associated gas hydrates and a conservative estimate suggested that about 20 GtC are currently locked in permafrost-associated gas hydrates (Ruppel, 2015).

Over the last four decades, the Arctic warmed almost four times faster than the rest of the globe (Rantanen et al., 2022). Ongoing climate change causes thawing of permafrost soils (Schuur et al., 2015, 2022; McGuire et al., 2018), which leads to the subsidence, erosion and potential collapse of the previously frozen ground in regions of diverse permafrost landforms. The degradation of organic matter and the dissociation of permafrost-associated gas hydrates are linked to the release of carbon dioxide (CO_2) and methane (CH_4) into the atmosphere as a consequence of permafrost thaw. This carbon loss is irreversible over several centuries.

These permafrost carbon emissions contribute to a positive climate feedback in which GHG emissions lead to additional warming, which, in turn, releases more GHG. This is called the permafrost carbon-climate feedback (Koven et al., 2011, Schuur et al., 2015, 2022, Canadell et al., 2021).

Current-generation climate models suggest a net positive impact of the permafrost carbon-climate feedback on global climate with estimates of additional warming of 0.05-0.7°C by 2100 (Schaefer et al., 2014, Burke et al., 2018, Kleinen and Brovkin, 2018, Nitzbon et al., 2023) based on low- to high-emissions scenarios, respectively. Methane emissions from permafrost could temporarily contribute up to 50 per cent of the permafrost-induced radiative forcing due to its higher warming potential (Walter Anthony et al., 2016, Turetsky et al., 2020, Miner et al., 2022). Overall, however, Canadell et al., (2021) summarise that "thawing terrestrial permafrost will lead to carbon release (high confidence), but there is low confidence in the timing, magnitude and relative roles of CO_2 and CH_4" of the permafrost carbon-climate feedback.

In addition, permafrost thaw impacts society in the permafrost region through changes at the land surface, e.g. wetting or drying of landscapes, ground subsidence due to melted ice, damaged infrastructure (roads, buildings, pipelines), and ecosystem changes such as ocean acidification or eutrophication (Hjort et al., 2018, 2022; Miner et al., 2021, Langer et al., 2023) (see Chapter 2.2 for societal impacts).

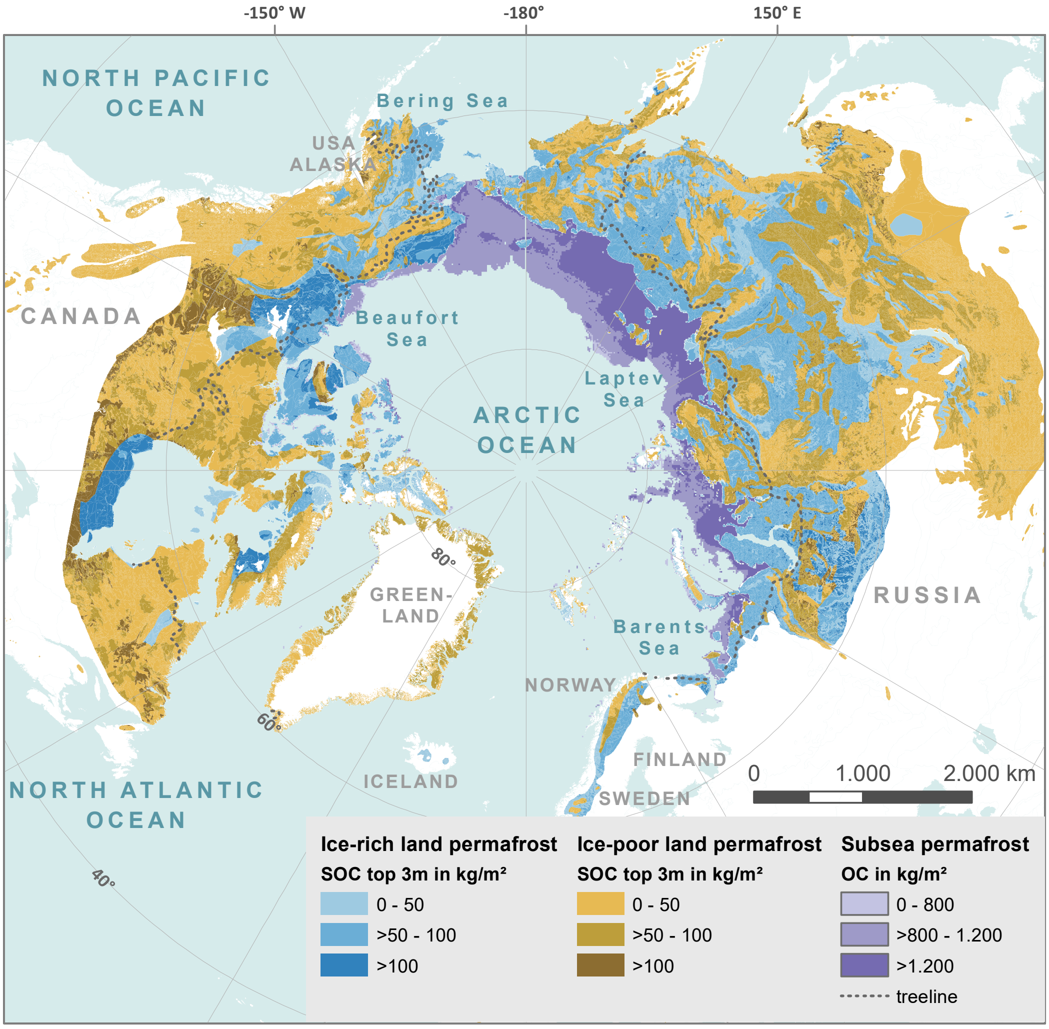

Figure 1.2.13: Map of estimated organic carbon storage ($kgCm^{-2}$) in the northern circumpolar permafrost region, combining terrestrial soil organic carbon contents (SOC, upper 3m) according to Hugelius et al. (2014) and subsea organic carbon contents according to Miesner et al. (2023). The terrestrial region is further divided into ice-rich and ice-poor regions according to Brown et al. (1997), where the ice-rich region is roughly coinciding with the areas susceptible to thermokarst and rapid thaw processes.

Evidence for tipping dynamics

Permafrost thaw is commonly denoted as gradual or abrupt. On land, gradual thaw occurs wherever the upper layer of thawed soil (active layer) gets successively deeper every year. Based on current projections, there is a high level of confidence that continued warming will result in ongoing, gradual declines in the volume of near-surface permafrost. It is anticipated that for every additional 1°C of warming, there will be a 25 per cent reduction in the global volume of perennially frozen ground found near the surface (Arias et al., 2021), which happens over the course of years to decades. The associated decomposition of permafrost carbon takes place on longer timescales, from centuries to millennia.

These models also suggest that the amount of carbon released from gradual thaw is roughly proportional to the amount of global warming in low- to high-emission scenarios, with the best estimate being 18 (3-41) GtC per degree of global warming (Canadell et al., 2021; Burke et al., 2017, 2018). Permafrost carbon release represents a relatively higher contribution to the remaining carbon budget for low-emission scenarios (Gasser et al., 2018; Kleinen and Brovkin, 2018), specifically when the permafrost carbon-climate feedback is taken into account in the carbon budget estimates (Canadell et al., 2021).

Abrupt or rapid thaw occurs where excess or massive ice is present in the ground and leads to the development of 'thermokarst'. When the ice melts and drains away, the land surface subsides. This leads to the development of characteristic landforms such as thaw lakes, thaw slumps, or eroding gullies and valleys (Figure 1.2.14). Their development is reinforced by increased heat conductivity of water and the decreasing stability of water body edges that further increases their size. Thus, these processes can permanently transform permafrost landscapes. Environments in which these processes are expected to occur are estimated to cover about 20 per cent of the present Arctic permafrost region (Olefeldt et al., 2016).

Thermokarst processes can occur in response to local disturbances or across regions experiencing rapid warming or extreme events, and positive/amplifying feedbacks can drive rapid permafrost loss (Nitzbon et al., 2020). Further, it is estimated that carbon emissions related to abrupt thaw processes could contribute an additional 40 per cent of emissions from newly formed features such as thaw slumps and thermokarst lake and wetland formation, which may double the radiative forcing from circumpolar permafrost-soil carbon fluxes (Turetsky et al., 2020; Walther Anthony et al., 2018). However, these processes are dependent on local environmental conditions that are unevenly distributed across the permafrost region (Olefeldt et al., 2016). Thus, despite the rapid nonlinear response at local-to-regional scale, the permafrost thaw and carbon emissions from thermokarst processes are likely to aggregate to a near linear response globally (Nitzbon et al., 2023).

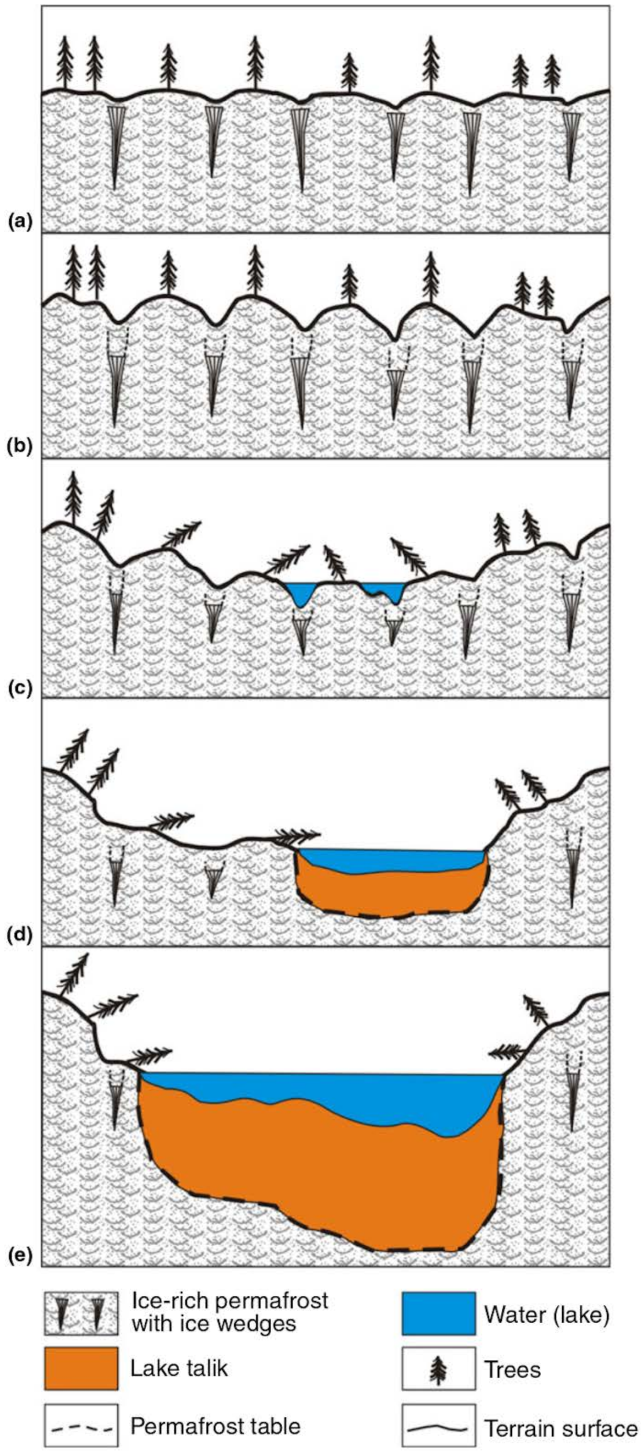

Figure 1.2.14: Schematic of abrupt thaw processes and landforms (thermokarst lake formation in ice-rich permafrost; from top to bottom) in continuous permafrost. Adapted from Grosse et al., (2013).

The loss of ground ice and the ecosystem changes are irreversible, with many local implications on topography and hydrology, including subsidence, drying or wetting, and changes in the microbial communities. In this context, microbial heat production was hypothesised as a possible self-reinforcing feedback on permafrost thaw (Khvorostyanov et al., 2008, Hollesen et al., 2015), but a consequential abrupt release of permafrost carbon through this 'compost bomb' mechanism (Clarke et al., 2021) is assessed to be unlikely. It would require organic carbon of very high quality and large quantity as well as comparably low ice contents, but such environmental preconditions are not prevailing over vast areas of the permafrost region. Accordingly, large-scale modelling studies found this effect to be of minor (Koven et al., 2011) or negligible (de Vrese et al., 2021) relevance to future projections of permafrost region carbon emissions.

While nonlinearity of the permafrost response to warming is exemplified in rapid thaw on local-to-regional scales, it is uncertain how these changes propagate to a larger scale. Some studies argue that an interaction of local feedbacks could lead to a quasi-linear response on a global scale (Schuur et al., 2015, Chadburn et al., 2017, Hugelius et al., 2020, Nitzbon et al., 2023), while others found multiple stable states in the permafrost system with potential nonlinear response on a large scale (de Vrese and Brovkin, 2021).

For the permafrost carbon-climate feedback to have large-scale tipping behaviour, it must be strong enough to cause self-sustaining permafrost loss beyond a certain warming threshold at either a global or subcontinental scale. Current AR6-based estimates yield a small positive amplification factor, indicating that the permafrost carbon-climate feedback is too small to be self-perpetuating on a global scale (Nitzbon et al., 2023). However, for future projections, both 'offline' permafrost models and Earth system models do not capture large-scale abrupt thawing throughout the Arctic.

Important processes such as interactions between fire, vegetation, permafrost, and carbon, as well as the potential for sudden releases through thermokarst phenomena, are currently not consistently considered (Natali et al., 2021). As a result, existing projections of permafrost thaw under various temperature thresholds are likely to be underestimates, indicating that the actual thaw potential may be greater than currently predicted.

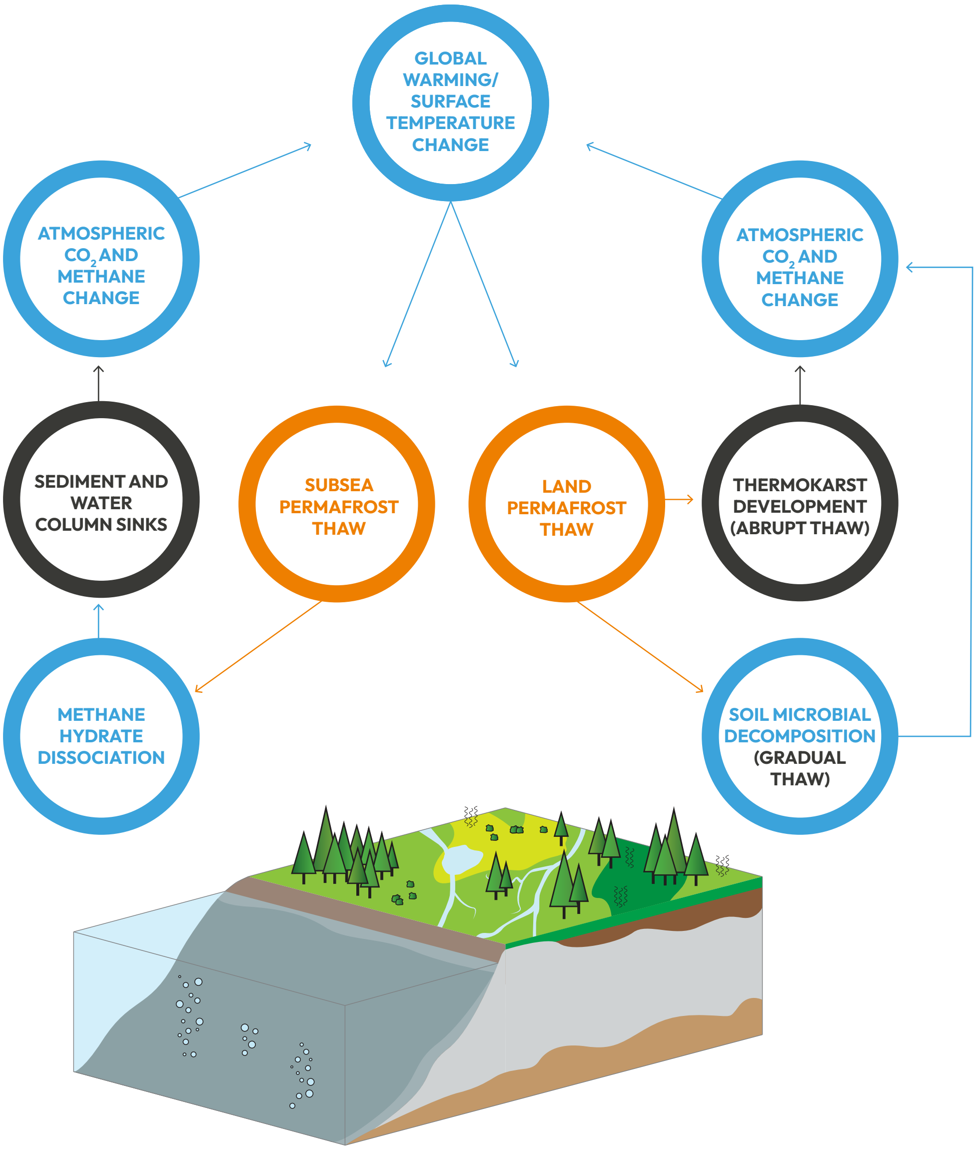

Figure 1.2.15: Schematic showing feedback processes related to land and subsea permafrost.

Since the flooding of the Arctic shelf after the last ice-age, the ocean floor has been exposed to relatively slow warming with small seasonal changes. Therefore subsea permafrost is thawing at a slow but continuous rate, leading to carbon emissions of 0.048 (0.025-0.085) Gt/yr (Miesner et al., 2023), an order of magnitude smaller than terrestrial permafrost carbon emissions (Figure 1.2.15). The disappearance of sea ice that has an insulating effect on ocean water temperature or major circulation changes in the Arctic Ocean may accelerate gradual thaw of subsea permafrost (Wilkenskjeld et al., 2022). However, this degradation process happens too slowly to support abrupt methane release (Reagan and Moridis, 2007; O'Connor et al., 2010). In addition, permafrost-associated gas hydrates within and below subsea permafrost are stabilised by the temperature and pressure conditions created by the permafrost. Permafrost thus acts as a lid on these GHG reservoirs and warming is expected to take centuries to penetrate them (Dmitrenko et al., 2011; Marín-Moreno et al., 2013). Some of these hydrates are relict deposits that are not necessarily stable under current conditions, but are self-preserving.

Subsea permafrost thaw only shows a delayed and dampened response to climate warming. In addition, microbial degradation rates are slow and strong methane sinks in both sediment and ocean likely limit net GHG emissions (James et al., 2016, Ruppel and Kessler, 2016). Another important aspect is the long timescale of permafrost thaw. Instantaneous changes in GHG emissions are quasi-linear, but committed changes on a centennial-to-millennial timescale could be nonlinear – as, for example, when a large area with frozen carbon storages is simultaneously affected by a strong warming. An example from palaeoclimate is a stepwise increase in atmospheric CO_2 concentration in response to an abrupt warming at about 14,700 years ago, plausibly explained by the permafrost thaw (Köhler et al., 2014).

Assessment and knowledge gaps

Accounting for its potential nonlinear response to warming, permafrost was considered a tipping system in numerous previous assessments (Armstrong Mckay et al., 2022, Fabbri et al., 2021, Yumashev et al., 2019, Schellnhuber et al., 2016, Steffen et al., 2018, IPCC AR6, Hamburg Climate Future Outlook). However, the aggregation of nonlinear or rapid local-to-regional permafrost degradation as a result of global warming results in a quasi-linear transient response of global permafrost extent on decadal to centennial timescales (Burke et al., 2020). The resulting permafrost carbon-climate feedback is likely positive, but current climate conditions do not support its self-sustenance, hence permafrost thaw is not expected to cause runaway global warming.

We conclude that permafrost exerts localised tipping points, which, however, do not aggregate to a large-scale tipping point at a global temperature threshold on decadal to centennial timescales. Similarly, subsea permafrost thaw happens relatively slowly, resulting in carbon emissions a magnitude smaller than from terrestrial permafrost. According to the strength of the available evidence, we have medium confidence in these assessments of both land and subsea permafrost.

The communication of a specific tipping threshold for permafrost could give a false sense of a temperature 'safe zone' at which permafrost is less vulnerable.

The effects of permafrost degradation are already seen today with implications for ecosystems and societies, where committed changes will continue to be relevant for centuries. Given the current modelling limitations, improvements in modelling permafrost dynamics will improve the confidence of evaluating permafrost stability, carbon loss, response linearity, and their impact on global climate.

1.2.3 Final remarks

With continued global warming, *all* parts of the cryosphere will be at increasing risk of further decline. For some parts of the cryosphere (like the ice sheets), this is likely to be characterised by tipping dynamics, while for others (like Arctic sea ice), it will occur gradually but surely, following the global warming trajectory. Due to the long response times of these systems, certain cryospheric elements are linked to committed long-term impacts. Major risks for each of the cryosphere elements for different levels of global warming are summarised in Figure 1.2.16. What is evident: despite the different dynamics and characteristics of ice sheet retreat, glacier decline, sea ice loss and permafrost thaw, the consequences of climate-induced changes in the cryosphere will be far-reaching and impact the livelihoods of millions of people.

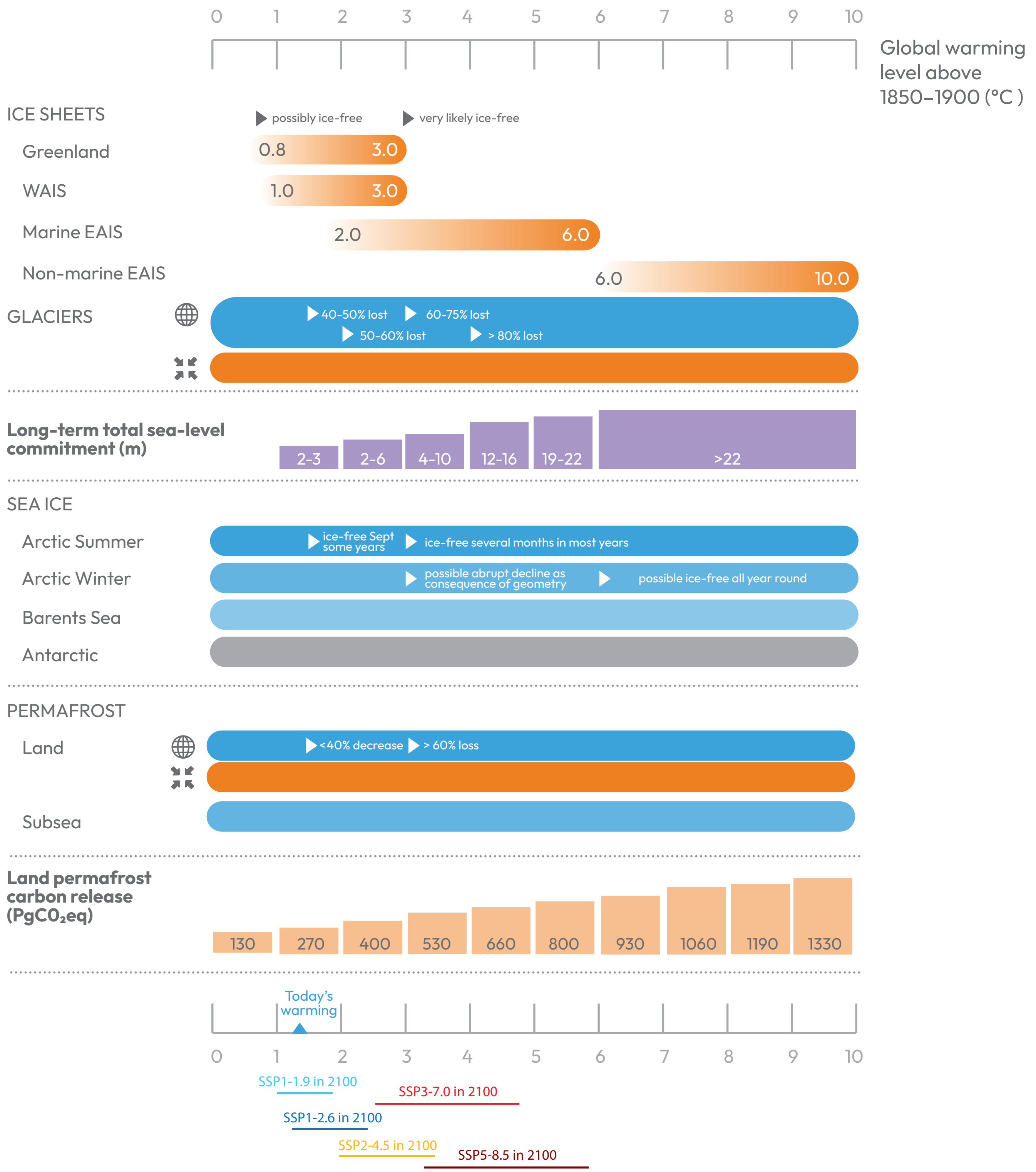

Figure 1.2.16: Increasing risks for cryosphere tipping elements with global warming. Potential thresholds (for ice sheets, glaciers, sea ice and permafrost) and impacts (long-term committed sea level rise and carbon release) are shown for different levels of global warming. Values for glacier thresholds, sea level commitment, Arctic summer sea ice, and land permfrost (for surface permafrost) are from Kloenne et al. (2023), land permafrost carbon release estimates are from Nitzbon et al. (2023), and SSP emission scenarios are from IPCC (2021). Sea level rise is 2000 yr commitment including thermosteric contribution with respect 1995-2014, and permafrost carbon release is relative to1850-1900." This figure is inspired by Kloenne et al., (2023).

Chapter 1.3 Tipping points in the biosphere

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Summary

This chapter assesses scientific evidence for tipping points across the biosphere, which comprises Earth's ecosystems. Human-driven habitat loss, pollution, exploitation and, increasingly, climate change are degrading ecosystems across the planet, some of which can pass tipping points beyond which a 'regime shift' to an alternative (and often less diverse or beneficial) ecosystem state occurs.

Evidence for tipping points emerges across many biomes. In forests, large parts of the Amazon rainforest could tip to degraded forest or impoverished savanna, while tipping in boreal forests is possible but more uncertain, and whether current temperate forest disturbance could lead to tipping is unclear. In open savannas and drylands, drying could lead to desertification in some areas, while in others encroachment by trees and shrubs could see these biodiverse ecosystems shift to a forested or degraded state. Nutrient pollution and warming can trigger lakes to switch to an algae-dominated low-oxygen state. Coral reefs are already experiencing tipping points, as more frequent warming-driven bleaching events, along with pollution, extreme weather events and diseases, tip them to degraded algae-dominated states. Mangroves and seagrasses are at risk of regional tipping, along with kelp forests, marine food webs and some fisheries, which are known to be able to collapse.

Together, these tipping points threaten the livelihoods of millions of people, and some thresholds are likely imminent. Stabilising climate is critical for reducing the likelihood of widespread ecosystem tipping points, but tackling other pressures can also help increase ecological resilience, push back tipping and support human wellbeing.

Key messages

- Evidence exists for tipping points in a variety of ecosystems, including forest dieback, tree and bush encroachment in savanna and grasslands, dryland desertification, lake eutrophication, coral reef die-off and fishery collapse.
- Several biomes (such as mangroves and the Amazon rainforest) are losing resilience and approaching key tipping thresholds, with current warming levels already triggering coral reef die-off tipping points in multiple regions.
- Ecosystem tipping points can be driven by many different drivers (including, but not limited to, climate change) that interact in complex ways across many species and feedbacks, making it harder to assess whether tipping points may be imminent.

Recommendations

- Reduce pressure on global ecosystems through the urgent phase-out of greenhouse gas emissions as well as tackling exploitation, habitat loss and pollution.
- Promote ecological resilience through adaptive management, ecosystem restoration and inclusive conservation, supporting sustainable livelihoods and rights for Indigenous peoples and local communities, and improved governance of land and oceans.
- Address deep uncertainties around feedbacks controlling ecosystem tipping and the impacts of increasingly extreme events, plant adaptability and spatial variability through more and better-integrated observations, experiments, and improved models.
- Invest in observations (field and remote sensing) and experiments to monitor and detect declining ecosystem resilience and potential early warning signals.
- Foster greater data sharing and international collaboration, and co-design research to bring together researchers across natural and social sciences and Global North and South, as well as Indigenous and traditional ecological knowledge.