Section 1

# Earth system tipping points


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## Contents

# Section summary

Tipping points exist across the Earth system – the interconnected systems that support life on this planet, including the cryosphere (ice-bound domains), biosphere (the living world), ocean and atmosphere. It is often assumed that environmental systems respond relatively linearly to human-driven pressures (such as climate change, habitat destruction and pollution). However, in some systems, pressure beyond a threshold causes them to shift to a very different state, often abruptly or irreversibly, as a result of self-sustaining feedbacks – they pass a tipping point.

In this section we compile evidence for tipping dynamics across the Earth system, noting where certainty and confidence is low and more research is needed. We also review the potential for interactions between climate tipping points to trigger tipping cascades, and the scope for detecting early warning signals before tipping points in monitoring data.

There is evidence for tipping points across the Earth system, including in major ice sheets, which could lock in multiple metres of sea-level rise, in ecosystems like the Amazon rainforest, which could die back to a degraded state, and in major ocean circulation patterns, which could abruptly shut down. Monitoring and early warning signals suggest some of these systems are already destabilising, indicating that tipping points could be approaching. Interactions between climate tipping points are destabilising in most (but not all) cases, and could lead to tipping cascades that destabilise wider parts of the climate system.

However, how close some Earth system tipping points may be is uncertain, with threshold estimates often spanning a large range. Established models, though capturing past and current climate trends well, often have a limited resolution of some processes that are key for making more accurate tipping dynamics estimates. However, they do hold evidence for potential tipping, which is strongly supported by conceptual models and palaeo reconstructions, which show that certain systems likely tipped in the past.

Given that tipping is possible, and that human-driven emissions are rapidly pushing the Earth to a climate unseen in at least the past 120,000 years, this provides strong motivation for rapidly reducing human-driven pressures on the Earth system (see Sections 3 and 4). It lays the foundation for preparing adaptation plans for the societal impacts of Earth system tipping points that cannot be avoided (see Section 2). Even if some tipping points are reached, mitigation to prevent further tipping points remains critically important.

# Key messages

- **We identify more than 25 parts of the Earth system that have tipping points,** based on evidence from palaeoclimate records, observations, theory and complex computer models, including:
    - » In the cryosphere, evidence exists for large-scale tipping points in Greenland and Antarctic ice sheets, and for localised tipping in glaciers and permafrost.
    - » In the biosphere, tipping points are present in a variety of ecosystems, including Amazon forest dieback, savanna and dryland degradation, lake eutrophication, coral reef and mangrove die-offs, and the collapse of some fisheries.
    - » In ocean-atmosphere circulations, there is evidence for tipping points in Atlantic and Southern Oceans overturning, as well as for the West African monsoon.
- **Multiple drivers are destabilising these systems.** Climate change is a key driver for most, as well as habitat loss (e.g. deforestation), exploitation (e.g. overfishing), and pollution (e.g. aerosols or nutrients) particularly in the biosphere.
- **Some Earth system tipping points could be very close already.** Coral reefs and some ice sheets could tip at current warming levels, and other systems' thresholds will soon be reached on current warming trends. Complex co-drivers, interactions, and feedbacks, as well as limited observational records, can make tipping thresholds difficult to assess for other systems, particularly in the biosphere.
- **Some climate tipping systems closely interact, and most interactions tend towards destabilising, making tipping 'cascades' possible.** There are large uncertainties around these cascades, but warming is approaching levels where they are becoming possible.
- **'Early warning signals' can sometimes indicate that a system is losing resilience** and so may be approaching a tipping point. Parts of the Greenland Ice Sheet, Atlantic Meridional Overturning Circulation, and the Amazon rainforest show such early warning signals, which is consistent with these systems approaching tipping points. However, these signals don't show for certain if or when a tipping point will occur.

# Recommendations

- **Prevent destabilisation of the Earth's tipping systems** through urgent and ambitious elimination of greenhouse gas emissions and reduction of other pressures such as deforestation, black carbon emissions and nutrient pollution.
- **Reduce deep uncertainties,** for example related to key processes and feedbacks like marine ice cliff instabilities, ecosystem responses to increasing extreme events and fine-scale ocean mixing, through further research and model intercomparison. Co-design research, bringing together the natural and social sciences, scholars across the Global South and North and multiple knowledge systems including Indigenous and traditional ecological knowledge.
- **Improve risk assessments of potential tipping cascades** through: i) representing more tipping system interactions in Earth system models, ii) large model ensembles to allow less common events to emerge, iii) studying possible cascades in ancient climate records, and iv) a fresh elicitation of expert knowledge to identify missed interactions.
- **Support development of novel and improved early warning techniques** (such as using machine learning) to detect declining resilience and other potential signs of tipping. Expand remote sensing capabilities and palaeorecords to improve datasets for early warning detection. Foster international data sharing and collaboration, and improve observational coverage in under-monitored regions such as Africa and Asia.

# Chapter 1.1 Earth system tipping points

## Introduction

In this section we scan different parts of the Earth system for evidence of tipping dynamics and assess whether they are likely to be tipping systems or not, providing confidence levels for each proposed tipping point and identifying knowledge gaps to be targeted with further research. We focus on the biophysical aspects of Earth system tipping points, with the societal impacts of these tipping points and adaptation to them explored in more depth in Section 2, and ways to govern the prevention of, and adaptation to, them examined in Section 3. We also consider how Earth system tipping points interact and potentially 'cascade' (where one tipping point triggers another, and so on) and assess the extent to which observations of these systems could give early warnings of impending tipping points.


**Figure 1.1.1:** Illustration of the Earth system, showing the different 'spheres'. The shown systems are a selected subset of the many components making up the Earth system.

The Earth system describes the interconnected complex system at the surface of the planet that sustains life (Figure 1.1.1). It is comprised of multiple subsystems (or spheres), including the cryosphere (ice-related systems, including ice sheets, sea ice, glaciers and permafrost), biosphere (global ecosystems), atmosphere, hydrosphere (water-based systems, including oceans, rivers and lakes) and the lithosphere (the Earth's solid surface) (Kump, Kasting, and Crane, 1999; Lenton, 2016). Together these subsystems and their interactions – referred to by the IPCC as the 'climate system' – determine the climate (the average long-term weather conditions at a place or across the Earth, usually measured over 30 years) (IPCC AR6 WG1 Annex VII).

At a smaller scale, ecosystems describe the complex systems composed of assemblages of living organisms and their physical environment in a particular location (e.g. a patch of rainforest in the Brazilian state of Amazonas), which at a larger scale form ecoregions, biomes and, ultimately, the whole global biosphere (Dinerstein et al., 2017; Keith et al., 2022). Humans, too, are a part of the biosphere, forming 'social-ecological systems' in which social and ecological dynamics have been inextricably long intertwined (Folke et al., 2016, 2021; Ellis et al., 2021).

Evidence from modelling, observations, theory based on understanding of fundamental biophysical processes, and geological records of ancient climate change (referred to as *palaeorecords*) suggests some of the Earth's systems can exhibit tipping points and associated dynamics (Lenton et al., 2008; Armstrong McKay et al., 2022; Wang et al., 2023). For example, there are multiple self-reinforcing feedback processes in ice sheets that not only amplify the effects of human-caused global warming, but may also lead to self-sustained melting beyond a critical warming threshold (Robinson et al., 2012; Garbe et al., 2020). Palaeorecords show that such collapses have happened before (Christ et al., 2021; Turney et al.,, 2020), and evidence from models and contemporary observations suggest some of these systems show increasing proximity to or may even be beyond tipping points (Feldmann and Levermann, 2015; Waibel et al., 2018; Rignot et al., 2014; Joughin et al., 2014; Boers and Rypdal, 2021).

Similar evidence for tipping points and destabilisation exists for ocean currents – such as the Atlantic Meridional Overturning Circulation (AMOC) (Böhm et al., 2015; Boers, 2021; Ditlevsen and Ditlevsen, 2023) – and ecosystems (Scheffer et al., 2009; Staal et al., 2020; Boulton et al., 2022). Tipping is often relatively rapid and irreversible, and has far-reaching implications for the climate, ecosystems and humans.

In this section we use the following tipping point definition to categorise proposed tipping systems, with key terms (defined in the Glossary) italicised:

**Box 1.1: Our Earth system tipping point (ESTP) definition:**

Tipping points occur when change in a *tipping system* (also known as a *tipping element*) becomes self-sustaining once a *forcing threshold* is passed, leading to a qualitative *state change* (e.g. an ecological *regime shift*) driven by one or more *positive/amplifying feedback loops*.

Climate tipping points, for example, occur when parts of the climate system reach global warming thresholds beyond which positive/amplifying feedbacks propel a shift to a totally different state, such as the inevitable collapse of an ice sheet or shutdown of a deep ocean convection site (Figure 1.1.2). Recent research suggests that five such tipping points may become likely beyond 1.5°C warming, including Greenland and West Antarctic ice sheet collapse, warm-water coral reef die-offs, overturning circulation collapse in the North Atlantic Subpolar Gyre, and widespread localised abrupt thaw in permafrost (Armstrong McKay et al., 2022). Earth system tipping points can occur due to a wider set of environmental drivers, including for example deforestation or nutrient pollution, as well as climate change.

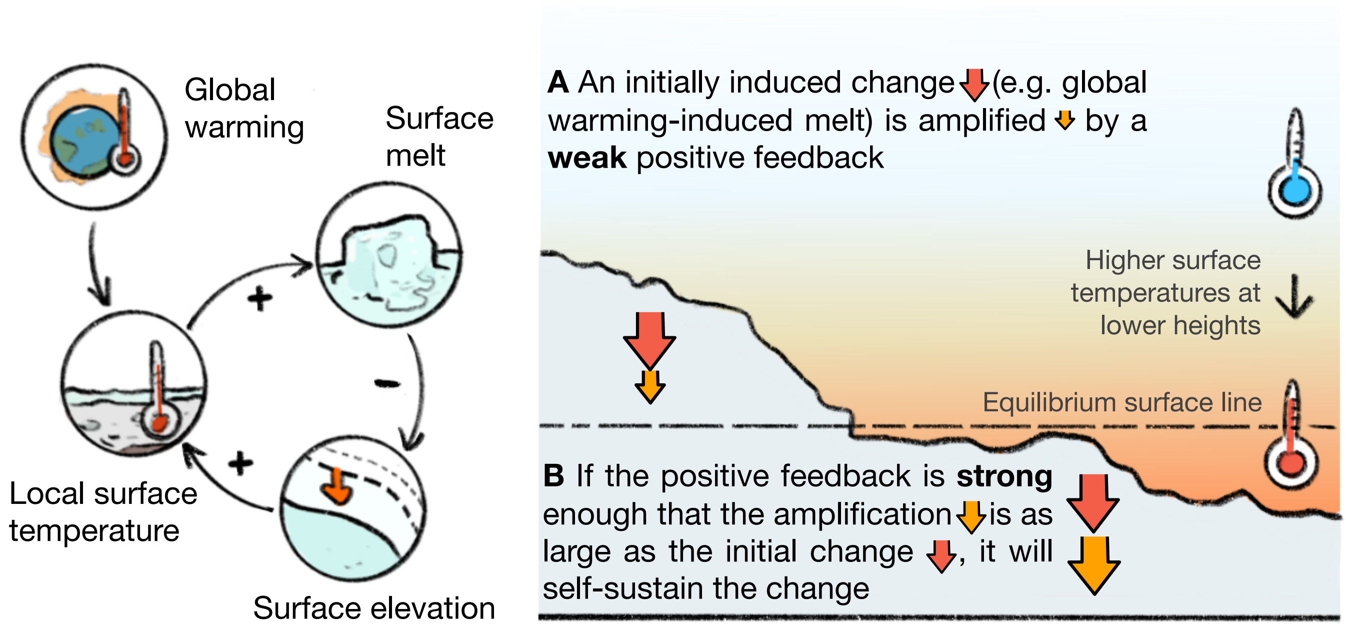


**Figure 1.1.2:** Self-sustaining change due to strong positive/amplifying feedbacks. Left shows one exemplary positive/amplifying feedback loop, the melt-elevation feedback that exists, e.g. in ice sheets. An increase in local surface temperatures leads to increasing melt, such that the melting ice sheet gets to lower heights. Since it gets warmer with lower altitude (atmospheric lapse rate), the local surface temperature is increased, restarting this circle. Such positive feedbacks amplify the initial change – however, if there is a critical threshold, beyond which the amplification leads to a change that is as large as the initial change, this leads to a vicious circle, self-sustaining the change.

We consider potential Earth system tipping systems in Chapters 1.2, 1.3, and 1.4 based on the scientific literature. In the cryosphere chapter (Chapter 1.2) we assess the ice sheets on Greenland and Antarctica, as well as sea ice in the Arctic and Southern Oceans, glaciers outside of polar regions, and permafrost. In the biosphere chapter (Chapter 1.3), on land we consider forests in tropical, temperate and boreal zones, as well as savannas, drylands and freshwater systems (lakes and rivers), and in the ocean we consider coral reefs, coastal and open ocean ecosystems. In the ocean and atmosphere circulation chapter (Chapter 1.4), we assess circulation in the North Atlantic and Southern Oceans, as well as atmosphere systems including monsoons, climate oscillations like the El Niño Southern Oscillation (ENSO), mid-latitude weather patterns like jet stream changes, as well as climate sensitivity and circulation linked to tropical clouds.

In many cases, the consequences of passing one tipping point make other connected tipping systems more or less likely to tip as a result. If passing one tipping point makes another tipping point more likely, then tipping points could cascade, with a chain of tipping points triggering each other. In Chapter 1.5 we present what is known about tipping point interactions in the climate system, including between the AMOC and ice sheets, Amazon and Arctic sea ice and between ENSO and coral reefs, Amazon and the West Antarctic ice sheet, and present some palaeoclimate case studies.

It may sometimes be possible to detect tipping points before they happen. Theory suggests that before some types of tipping point, subtle changes may be observable in the statistical properties of monitoring data, known as early warning signals (EWS). The most common type of EWS is critical slowing down, where natural fluctuations in an observed property of a system (such as temperature or tree cover) become bigger and longer, leading to larger values of variability (e.g. in variance) or self-similarity to recent values (e.g. in autocorrelation). In Chapter 1.6 we present different techniques and some case studies of detecting EWS before Earth system tipping points, and discuss both their limitations and future opportunities.

# 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.

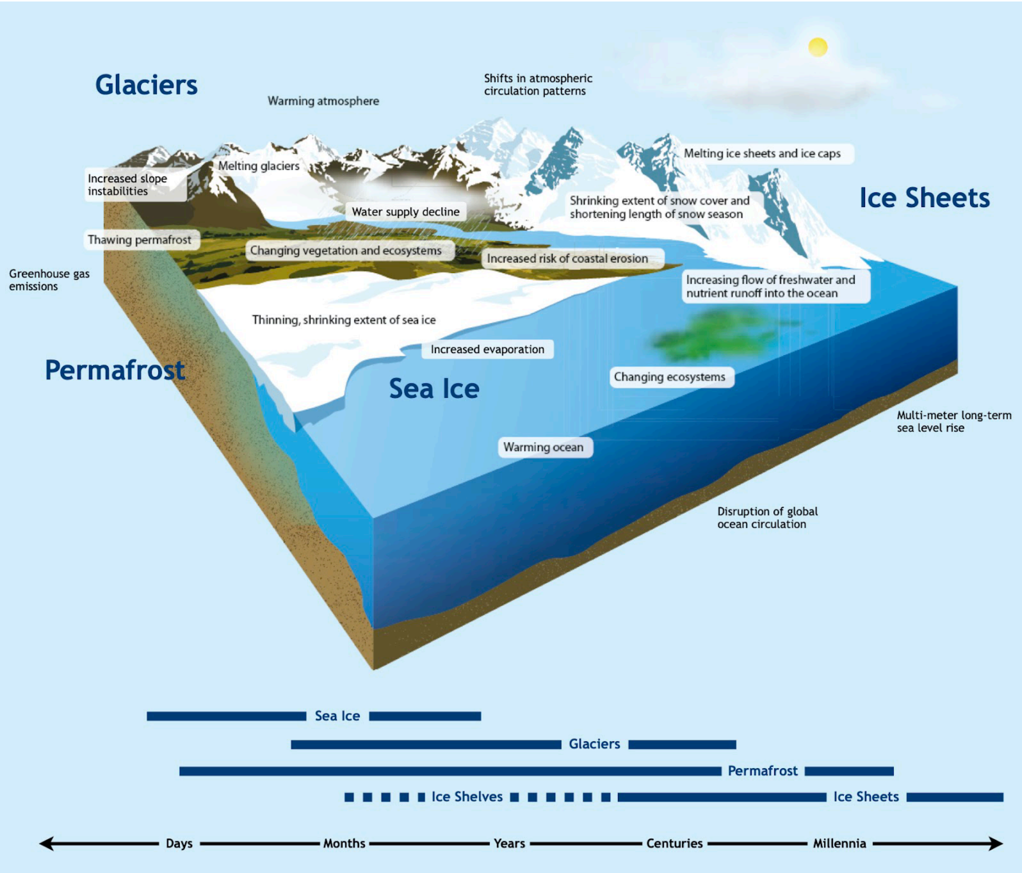


**Figure 1.2.1:** Key biophysical impacts resulting from crossing tipping points in the cryosphere. Diagram below gives approximate timescales of changes in the respective domain/system. Background graphic from: AMAP (2017).

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.

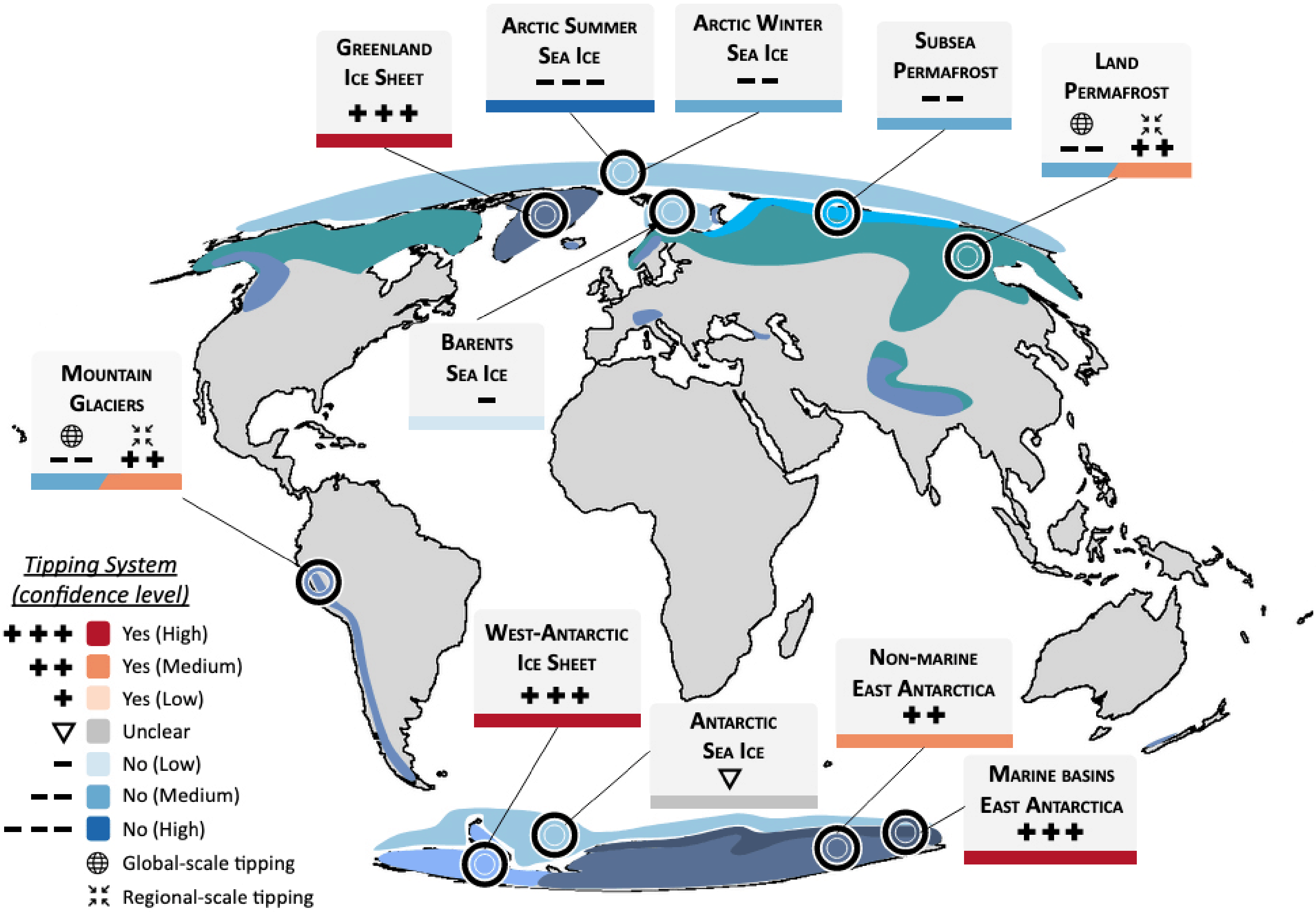


**Figure 1.2.2:** Map of cryosphere systems considered in this chapter (shading). The markers indicate which of the systems are in this report considered a tipping system (+++ high confidence, ++ medium confidence and + low confidence) and which are not (- - - high confidence, - - medium confidence and - 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 (↗)<br>DC: precipitation increase (↘)<br>DC: ocean warming and circulation changes (↗/↘)<br>DC: black carbon deposition (↗)<br>CA: sea ice decline (↗)<br>CA: atmospheric circulation changes (↗\|↘) | • Sea level rise (up to 7m) over centuries to millennia<br>• Disruption of global ocean circulation<br>• Substantial shifts in atmospheric circulation patterns | • PF: melt-elevation<br>• PF: melt-albedo | +++ | 0.8-3°C | +++ | **+++** |
| **West Antarctic Ice Sheet (collapse)** | DC: ocean warming and circulation changes (↗)<br>DC: atmospheric warming (↗)<br>DC: precipitation increase (↘) | • Sea level rise (up to 3m) over centuries to millennia<br>• Disruption of global ocean circulation<br>• Substantial shifts in atmospheric circulation patterns | • PF: marine ice sheet instability<br>• NF: glacial isostatic adjustment<br>• ?: melt-stratification | +++ | 1-3°C | +++ | **+++** |
| **Marine basins East Antarctica (collapse)** | DC: ocean warming and circulation changes (↗)<br>DC: atmospheric warming (↗)<br>DC: precipitation increase (↘) | • Sea level rise (up to 19m) over centuries to millennia<br>• Disruption of global ocean circulation<br>• Substantial shifts in atmospheric circulation patterns | • PF: marine ice sheet instability<br>• NF: glacial isostatic adjustment<br>• ?: melt-stratification | +++ | 2-6°C | +++ | **+++** |
| **Non-marine East Antarctic Ice Sheet (collapse)** | DC: atmospheric warming (↗)<br>DC: precipitation increase (↘) | • Sea level rise (up to 34m) over centuries to millennia<br>• Disruption of global ocean circulation<br>• 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 (↗)<br>DC: atmospheric circulation shifts (↗/↘)<br>DC: ocean warming (↗)<br>DC: ocean circulation shifts (↗/↘)<br>DC: black carbon deposition (↗)<br>DC: storminess increase (↗)<br>CA: ocean stratification increase (↘) | • Regional warming (polar amplification)<br>• Ecosystem disruption<br>• Impacts on ocean circulation<br>• Impacts on atmospheric circulations<br>• Increased evaporation | • PF: Ice-albedo FB<br>• NF: Snow FB<br>• NF: Growth FB<br>• 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 (↗)<br>DC: deposition of dust, black carbon, etc. (albedo) (↗)<br>DC: reduced snow (input and albedo) (↗)<br>DC: local thermokarst (↗) | • Water supply decline<br>• Ecosystem disruption (e.g. wetlands, water chemistry)<br>• Increase in number and size of glacier lakesIncrease in slope instabilities<br>• Transition from glacial to para-glacial landscapes<br>• Sea level rise | • PF: melt-elevation FB<br>• PF: calving front retreat<br>• PF-: ice-dynamic FBs<br>• NF: retreat to higher altitudes | + + (regional)<br>- - (global) | Regionally variable | - - | **++ (regional)**<br>- - (global) |
| **Permafrost** | | | | | | | |
| **Land permafrost (thaw)** | DC: atmospheric warming (↗)<br>CA: vegetation increase (increase albedo ↗, increase summer shading ↘, and vice versa for forest die-back)<br>CA: wildfire intensity increase (↗)<br>CA: precipitation increase (rain extremes, snow cover albedo ↗) | • Greenhouse gas emissions<br>• Landscape disruption<br>• Ecosystem disruption | • PF: carbon-climate FB<br>• PF: thermokarst development<br>• PF: summer soil drying<br>• PF-: vegetation interaction | - - (global)<br>++ (regional) | N/A | +++ (wrt carbon loss) - - - (wrt frozen soil) | **++ (regional)**<br>- - (global, on 10s-100s year timescale) |
| **Subsea permafrost (thaw)** | DC: ocean warming (↗)<br>CA: sea ice loss (↗)<br>CA: water pressure reduction (↗) | • Greenhouse gas emissions | • PF: Carbon-climate FB<br>• NF: sediment sink<br>• NF: water column sink | + | N/A | + + (w.r.t. gas hydrate dissociation)<br>+ + (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).

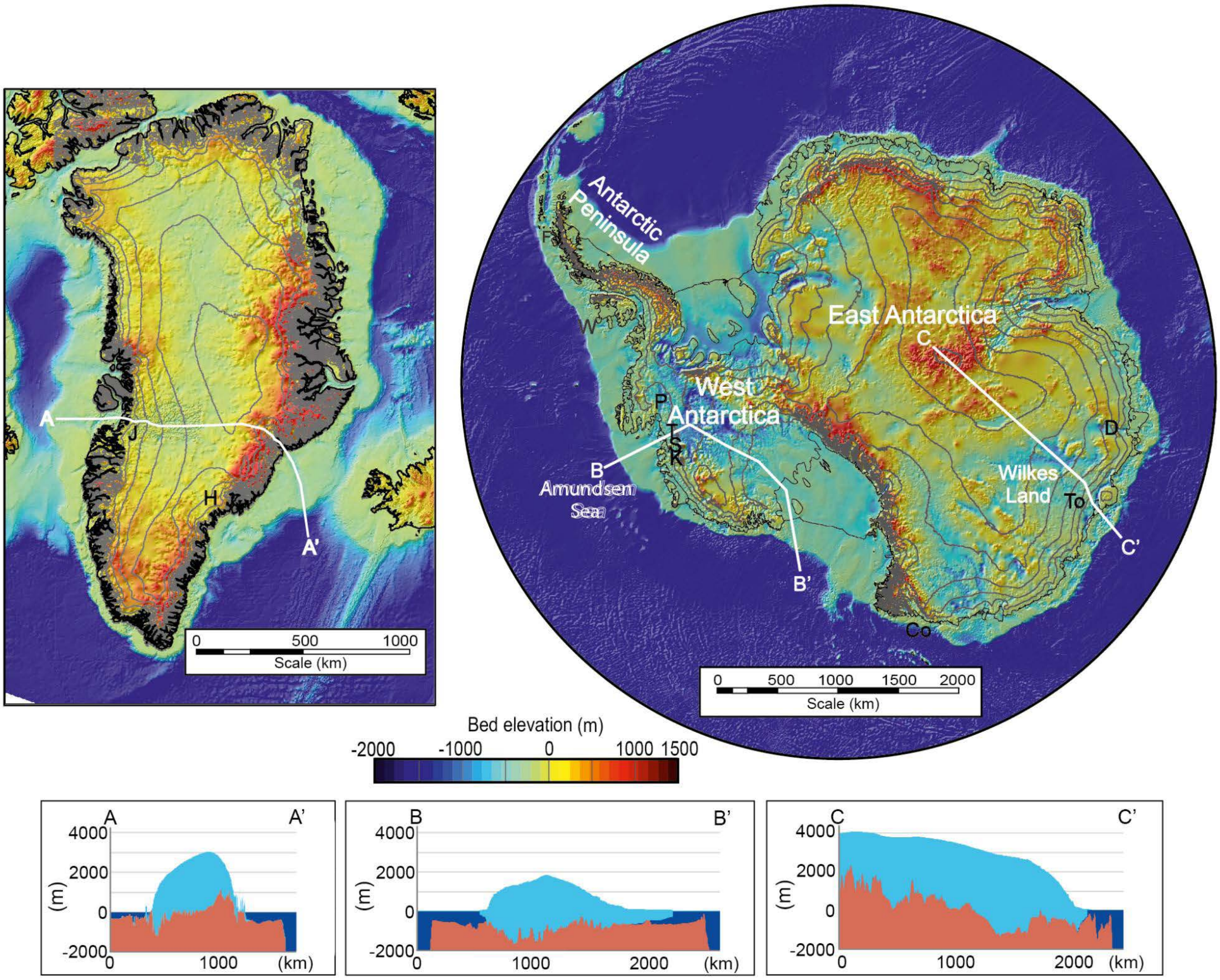


**Figure 1.2.3:** Greenland and Antarctic ice sheets. Given is the bedrock topography of the GrIS (left, based on Bamber et al., 2013) and the Antarctic Ice Sheet (middle and right, based on Fretwell, 2013) alongside cross sections marked in the maps by white lines. In marine ice sheet sectors (blue-green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

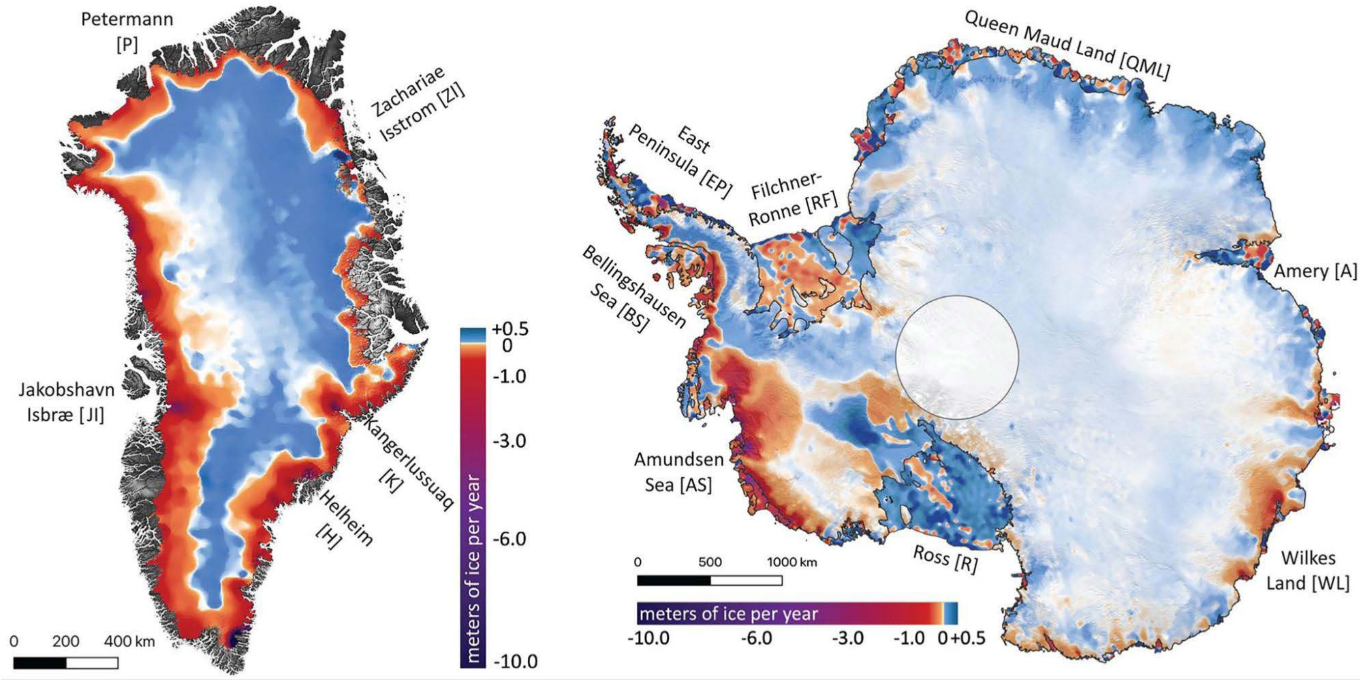


**Figure 1.2.4:** Observed mass change of Greenland and Antarctic ice sheets. Mass change (mass loss in red, mass gain in blue) between 2003 and 2019 for Greenland and Antarctica, given in metres of ice equivalent per year (from Smith et al., 2020).

**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).

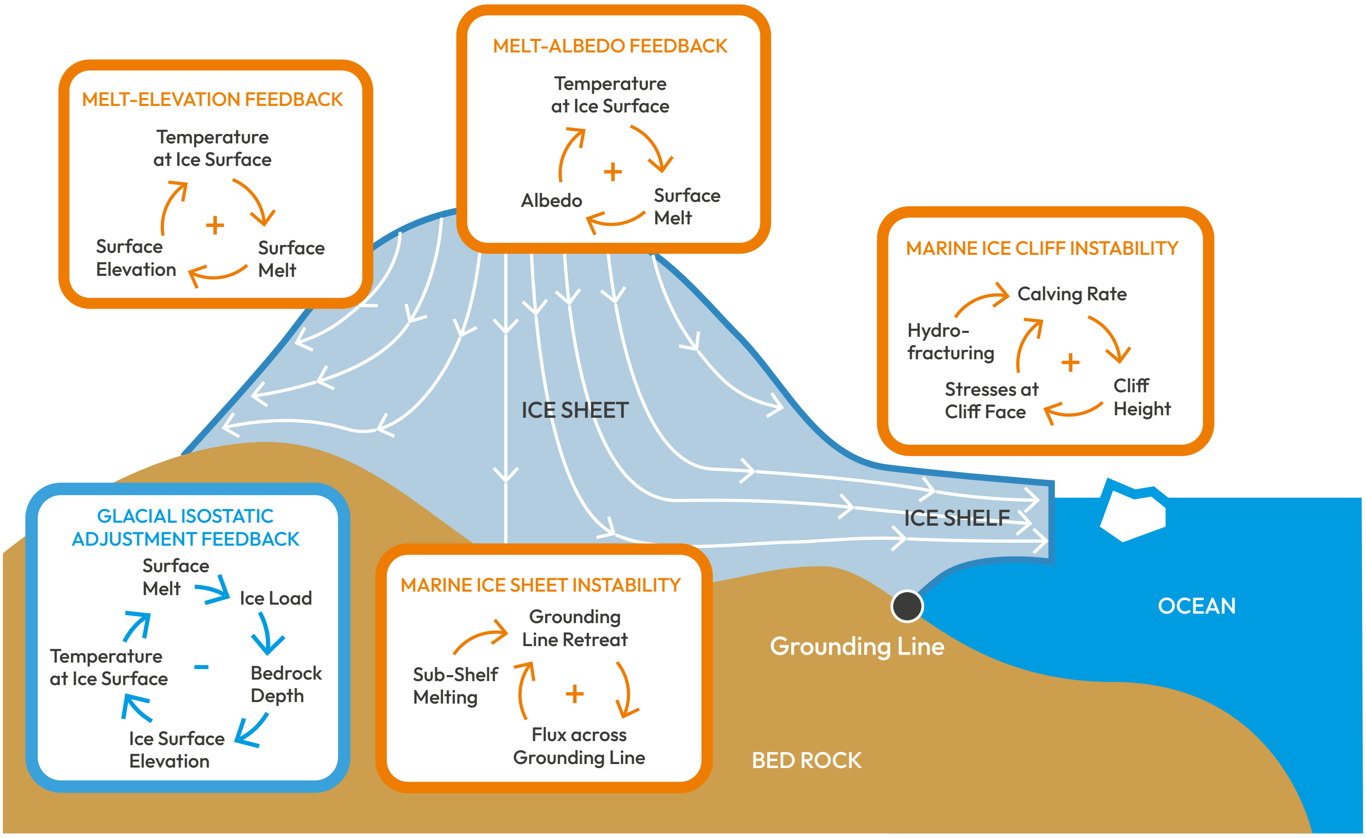


**Figure 1.2.5:** Schematic illustrating some of the key feedbacks in the ice sheet-climate system. Note that this depiction is limited to the most relevant and widely examined feedbacks, further self-amplifying or damping feedbacks may, however, exist.

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).

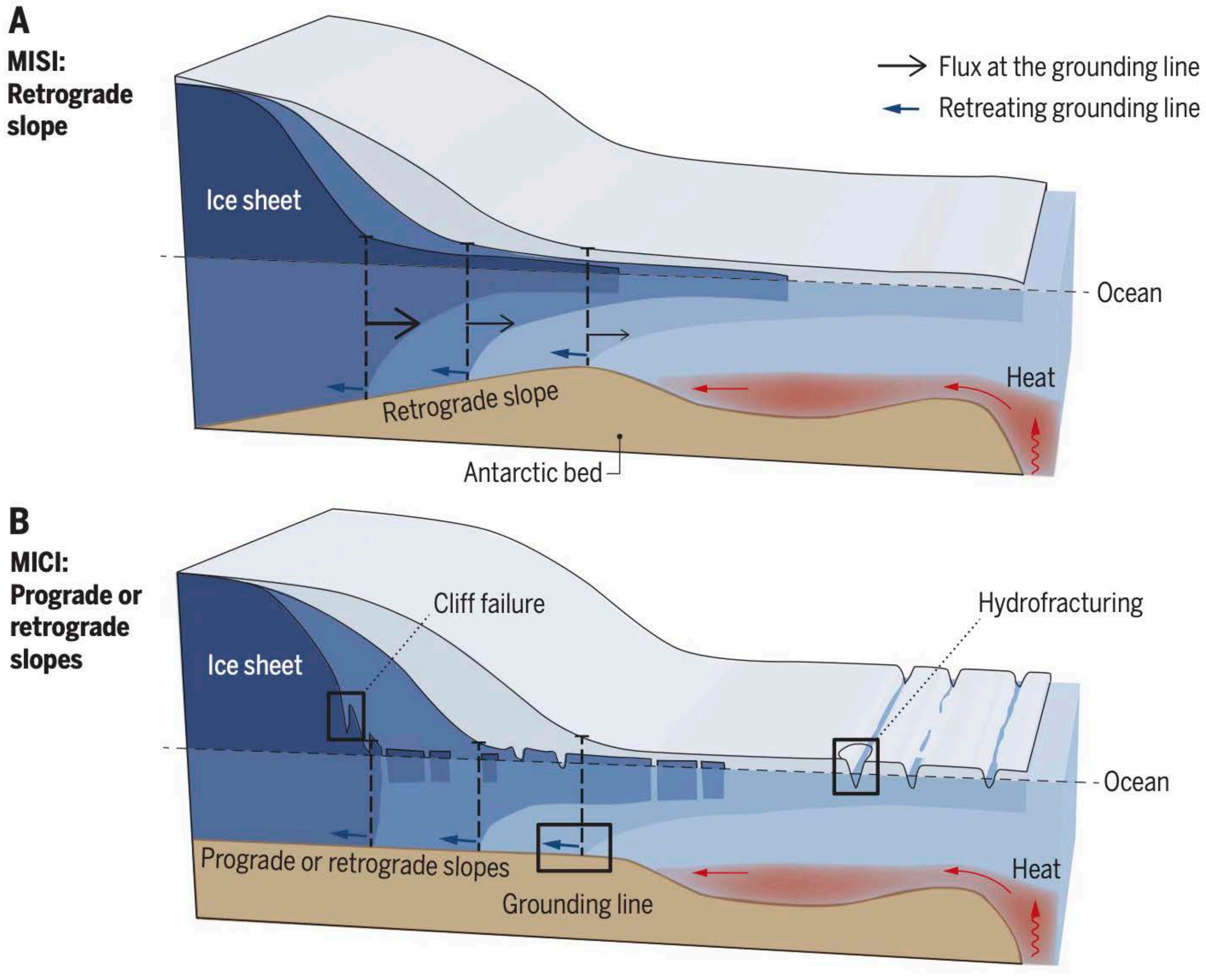


**Figure 1.2.6:** Schematic illustration of marine ice sheet instability (MISI; top) and marine ice cliff instability (MICI; bottom). From Pattyn and Morlighem (2020) .

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

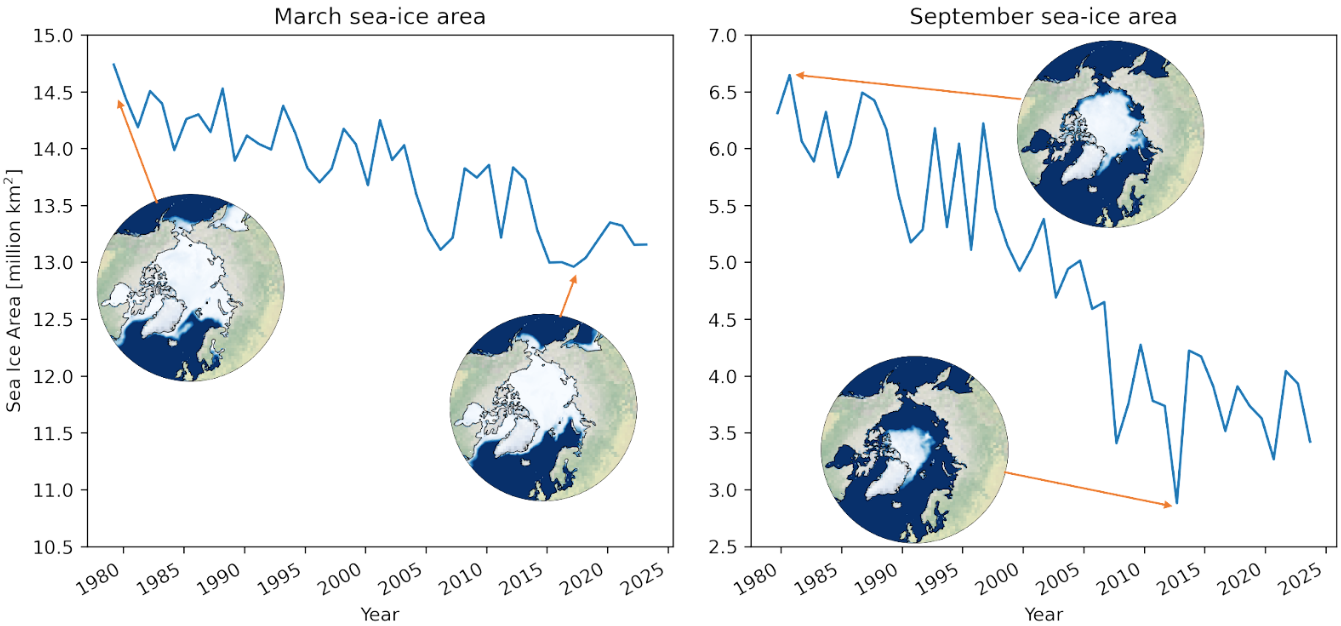


**Figure 1.2.7:** Arctic sea ice evolution 1979-2023. Time series of Arctic sea ice area, with insets showing sea ice concentration in selected years. March is usually the month of maximum sea ice area ('winter sea ice'), September is usually the month of minimum sea ice area ('summer sea ice'). Data: 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).

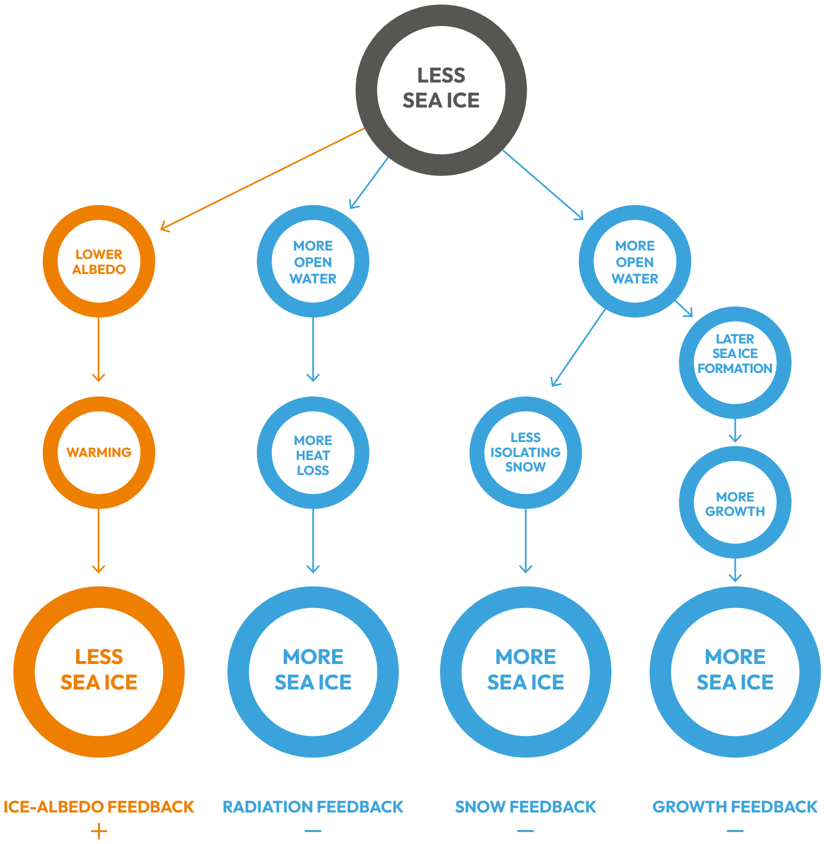


**Figure 1.2.8:** Schematic illustrating some of the key feedbacks related to Arctic sea ice loss. Note that this depiction is limited to the most relevant and widely examined feedbacks; further self-amplifying or damping feedbacks may, however, exist. Based on Notz and Bits (2016).

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).

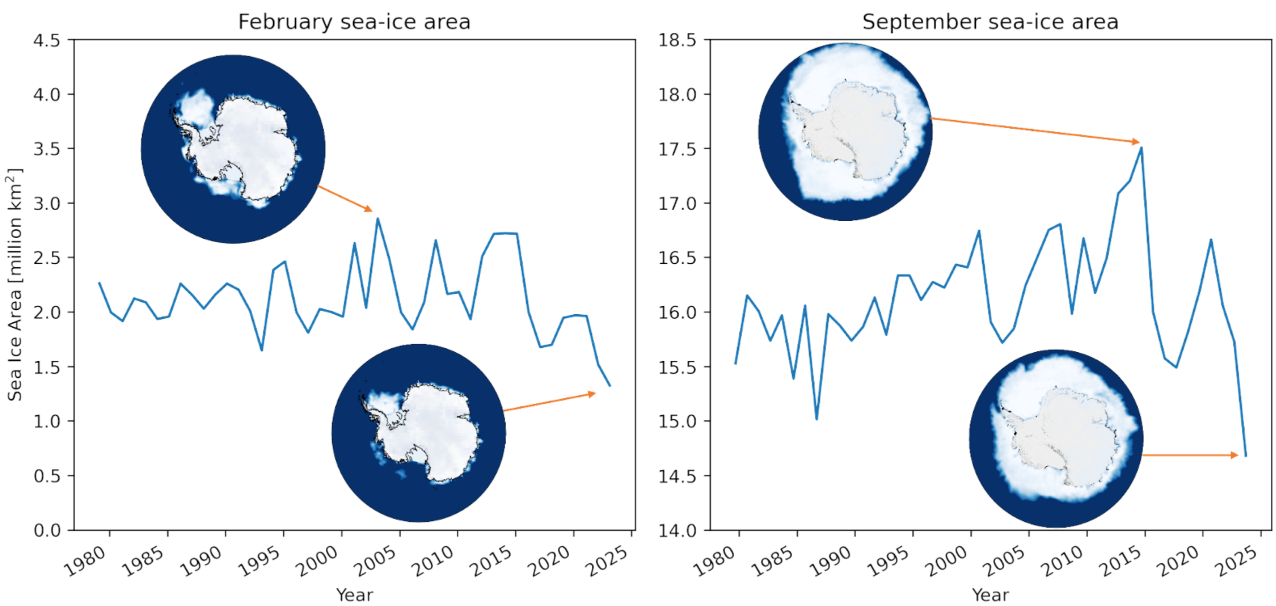


**Figure 1.2.9:** Antarctic sea ice evolution 1979-2023. Time series of Antarctic sea ice area, with maps showing sea ice concentration in selected years. February is usually the month of minimum sea ice area ('summer sea ice'). September is usually the month of maximum sea ice area ('winter sea ice'). 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).

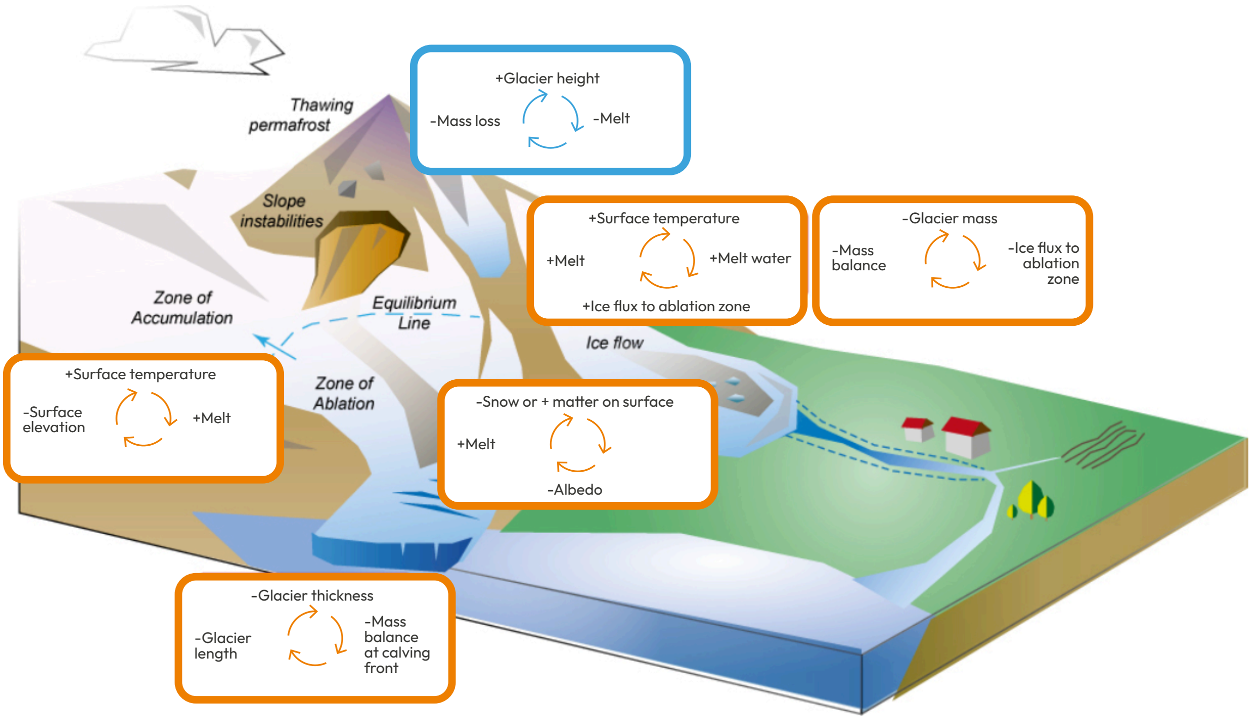


**Figure 1.2.10:** Terminology and some of the key feedbacks related to mountain glacier retreat. 'Positive' amplifying feedbacks that amplify ice loss are shown by red boxes, and 'negative' damping feedbacks that limit ice loss are shown by blue boxes. Above the equilibrium line (dashed blue line) glaciers 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.

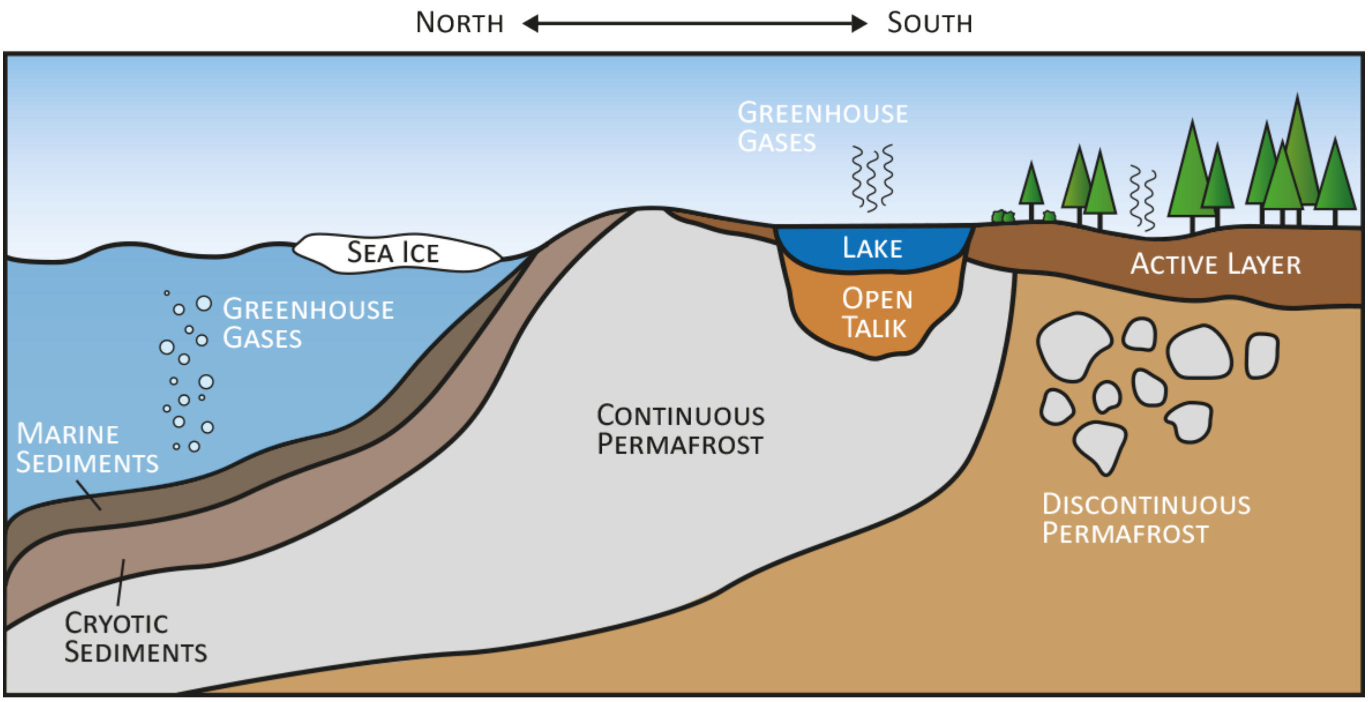


**Figure 1.2.12:** Schematic showing typical permafrost landscape features. Inspired by Lantuit et al., (2012).

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).

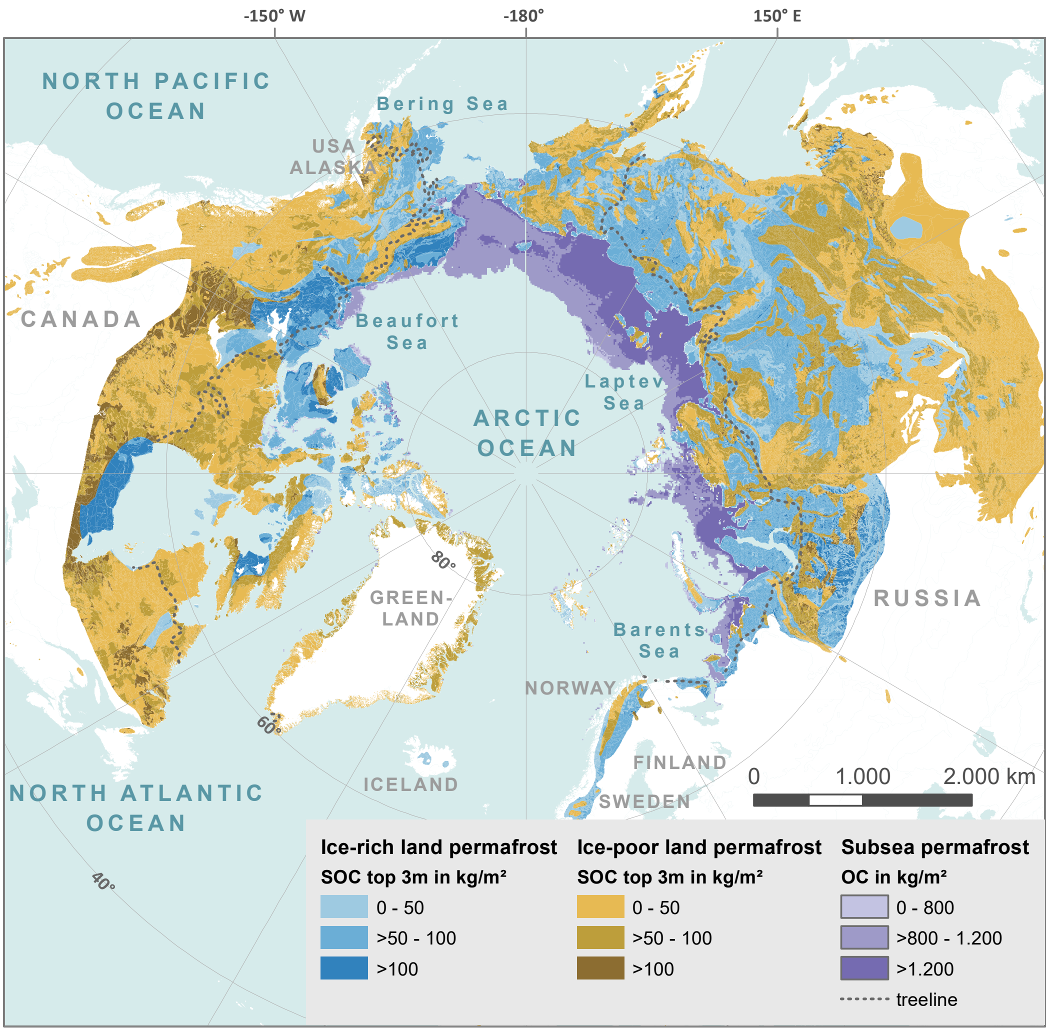


**Figure 1.2.13:** Map of estimated organic carbon storage (kgCm$^{-2}$) in the northern circumpolar permafrost region, combining terrestrial soil organic carbon contents (SOC, upper 3m) according to Hugelius et al. (2014) and subsea organic carbon contents according to Miesner et al. (2023). The terrestrial 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).

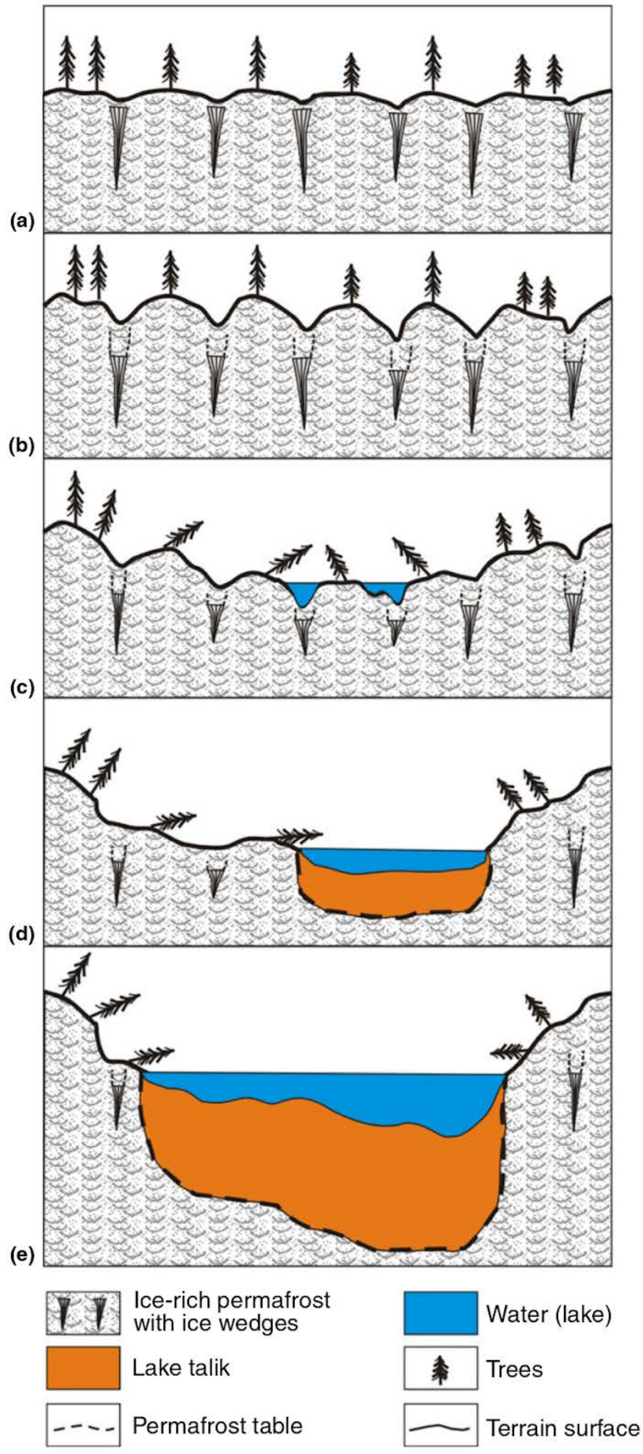


**Figure 1.2.14:** Schematic of abrupt thaw processes and landforms (thermokarst lake formation in ice-rich permafrost; from top to bottom) in continuous permafrost. Adapted from Grosse et al., (2013).

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.**

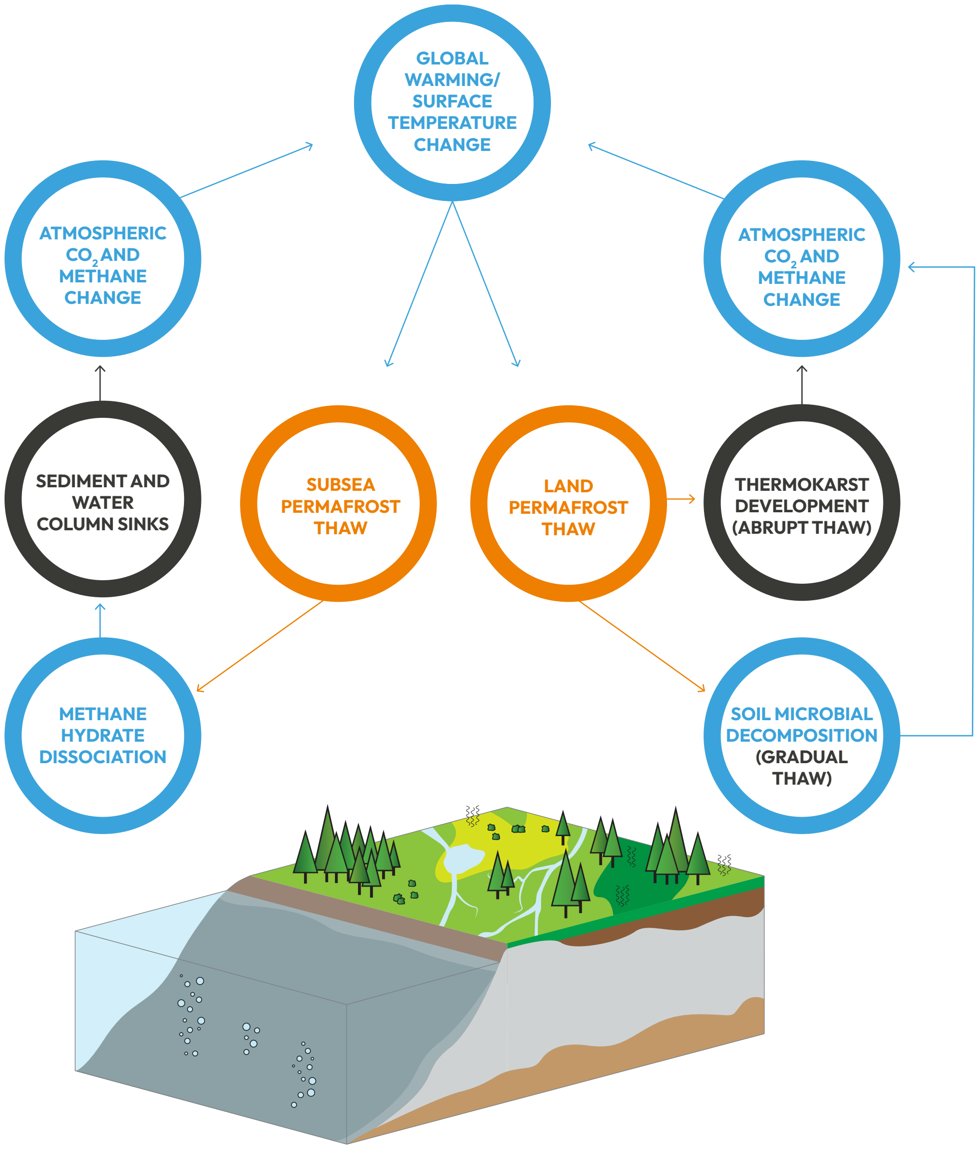


**Figure 1.2.15:** Schematic showing feedback processes related to land and subsea permafrost.

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.

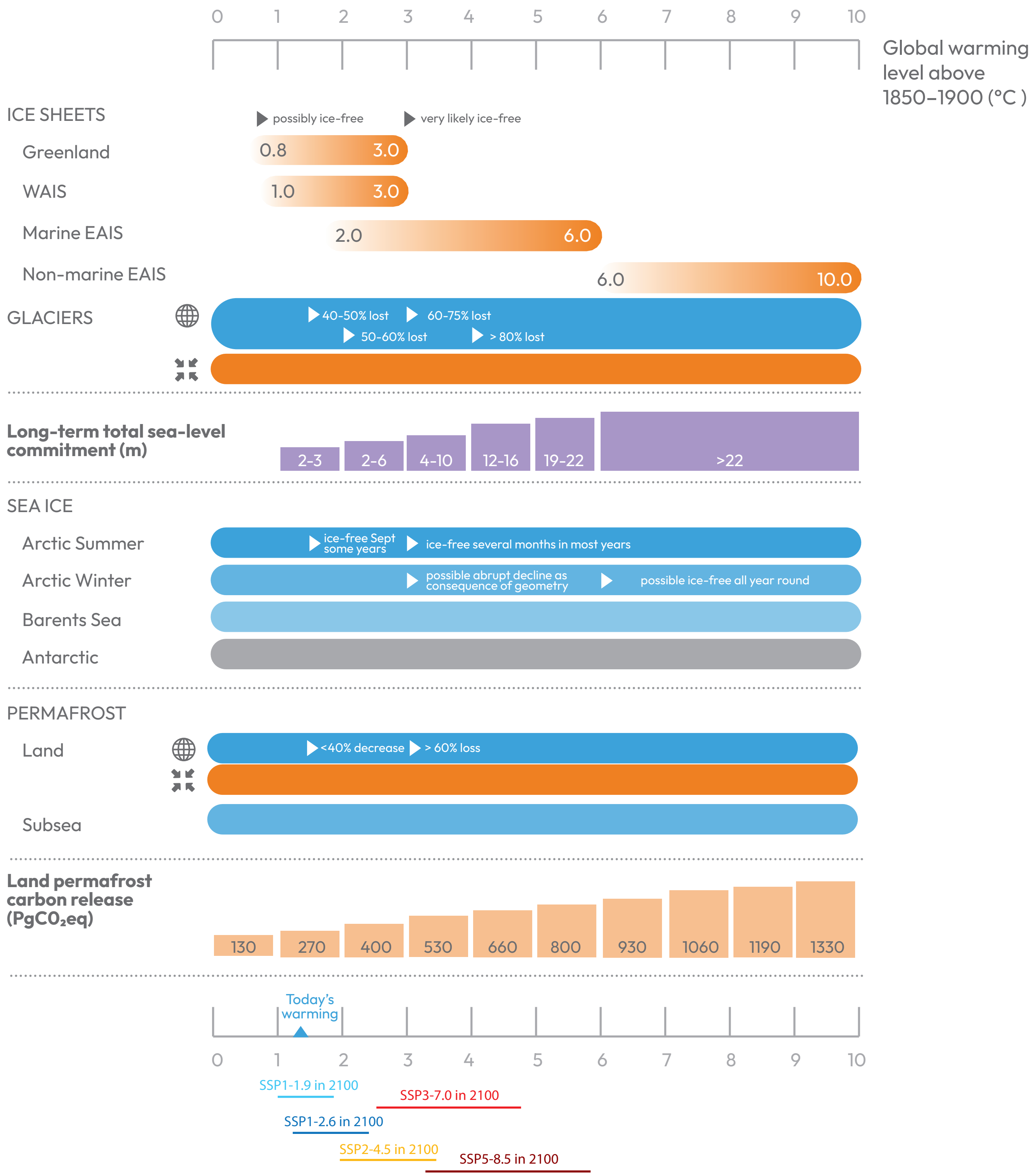


**Figure 1.2.16:** Increasing risks for cryosphere tipping elements with global warming. Potential thresholds (for ice sheets, glaciers, sea ice and permafrost) and impacts (long-term committed sea level rise and carbon release) are shown for different levels of global warming. Values for glacier thresholds, 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.

## 1.3.1 Introduction

The Earth's biosphere describes the sum of all global ecosystems. It forms a key part of the Earth system, driving the many biogeochemical cycles that maintain the climate system and keep Earth habitable (Kump, Kasting, and Crane, 1999). Ecosystems are the complex systems composed of assemblages of living organisms and their physical environment at the local scale (e.g. an area of rainforest in the Brazilian state of Amazonas).

At a larger scale, they form regional groupings (e.g. Madeira-Tapajós moist forest ecoregion in Dinerstein et al., 2017), ecosystem functional groups (e.g. tropical/subtropical lowland rainforests), biomes (e.g. tropical-subtropical forests), and ultimately the whole biosphere (Keith et al., 2022). Humans are also an integral part of the biosphere, with social systems being so closely intertwined with ecosystems that they can be seen as joint 'social-ecological systems' in which the dynamics of both interact as a single complex adaptive system (Folke et al.,2016; 2021; Ellis et al., 2021).

Ecosystems are being globally degraded by multiple human-driven pressures. At the species level, one million animals and plants face extinction (IPBES, 2019). Extinctions are happening at up to 100 times natural background rates averaged over the last century, leading some to assess that the Earth has now entered the sixth mass extinction event in the nearly 4 billion years of life's history (Barnosky et al., 2011; Ceballos et al., 2015). The Living Planet Index indicates that populations are declining in around half of vertebrate species, with an average decline across all species of 69 per cent since 1970 (WWF, 2022). The key drivers of biodiversity loss in order of importance are land and sea use change, direct exploitation, climate change, pollution, and invasive alien species (IPBES, 2019; Maxwell et al., 2016). Climate change is not currently the leading driver, but will become a substantial threat with further warming (IPBES, 2019). Global warming moving from 1.5 to 2°C increases the number of species facing the loss of most of their ranges from 4 to 8 per cent for vertebrates (e.g. mammals), 8 to 16 per cent for plants, and 6 to 18 per cent for insects, while 3.2°C of warming would increase these to 26, 44, and 49 per cent respectively (Warren et al., 2018). Together these losses are harming many ecosystems' ability to function and so threatening the critical ecosystem services that humanity relies upon, including providing food, clean water, and removing ~31 per cent of human-emitted $CO_2$ (Friedlingstein et al., 2022)..

As with many other complex systems, ecosystems have been proposed to feature nonlinear changes such as tipping points, beyond which dramatic shifts to a different ecological state are expected, further threatening biodiversity and bio-abundance (Scheffer et al.,2001, 2009). Ecosystems are also subject to many co-stressors with complex interactions, with changing disturbance regimes eroding resilience (e.g. Nystrom et al., 2000; Folke et al., 2004) and making tipping points easier to reach (Willcock et al., 2023). However, complex ecological and social-ecological dynamics crossing multiple scales can make it hard to discern tipping thresholds in observations (Schröder et al., 2005; Hillebrand et al., 2020; Spake et al., 2022). Organisms have agency that enables complex network and spatial dynamics to emerge – with human agency making social-ecological systems particularly complex – making ecosystem tipping dynamics often more difficult to detect and project relative to more physical systems (Kéfi et al., 2022; Rietkerk et al., 2021; Bastiaansen et al., 2022). Furthermore, while ecosystem functions or composition can have threshold responses to biodiversity loss or environmental change, in many cases responses remain relatively linear (Cardinale et al., 2011; Meyer et al., 2017; Hodapp et al., 2018; Strack et al., 2022).

Tipping at the global biosphere scale has been discussed (Barnosky et al., 2012; Hughes et al., 2013; Lenton and Williams, 2013) but is deemed unlikely, with local ecosystem shifts globally aggregating to relatively linear changes in response to human-driven pressures (Brook et al., 2013; Montoya et al., 2017; Rockström et al., 2018). Empirical evidence for tipping has, though, been found in multiple ecosystems from the local to regional scale – for example, in lakes, coastal zones, marine food webs, rangelands and forests (Scheffer et al., 2001, 2009; Folke et al., 2004; Walker and Meyer, 2004; Brook et al., 2013; Rocha et al., 2015; regimeshifts.org), and model evidence suggests tipping is possible in some biomes across sub-continental scales (Armstrong McKay et al., 2022; Wang et al., 2023). As such, ecological tipping points remain a useful concept (alongside gradual and nonlinear change) in understanding and managing ecosystems, despite being sometimes hard to observe in practice (Lade et al., 2021; Spake et al., 2022; Norberg et al., 2022).

In this chapter we follow the wider section's tipping point definition to categorise proposed tipping systems (see Box 1.1). In ecology, the terms 'regime shift' and 'critical transition' have been used interchangeably with 'tipping points', despite differences in meaning (Dakos, 2019). A *regime shift* refers to a shift in the current state of an ecological or social-ecological system from one partially stable state to another that is often large, relatively sudden (depending on system size and feedback timescales) and long-lasting, and entails a reorganisation in the structure and functioning of the system (Biggs et al., 2009; Maciejewski et al., 2019; Cooper et al., 2020). A *critical transition* refers to an abrupt shift in a system that occurs at a specific critical threshold in external conditions (Scheffer et al., 2009). In this Chapter, we use *tipping event* to describe the crossing of a tipping point (which is equivalent to critical threshold), and *regime shifts* to describe the resulting changes that unfold (equivalent to critical transition above). *Resilience* – the ability of ecosystems to maintain functioning in response to change and regenerate in the face of shocks, sometimes adapting and transforming in the process – is also a key concept, with declining resilience being a potential precursor to tipping (see Section 1.6) (Folke et al., 2004, 2016).

## 1.3.2 Current state of knowledge on tipping points in the biosphere

In this section we assess available scientific literature relating to tipping points in the Biosphere, as summarised in Figure 1.3.1 and Table 1.3.1. We focus on the following biomes: forests, savannas, drylands, lakes, coastal ecosystems and marine environments.

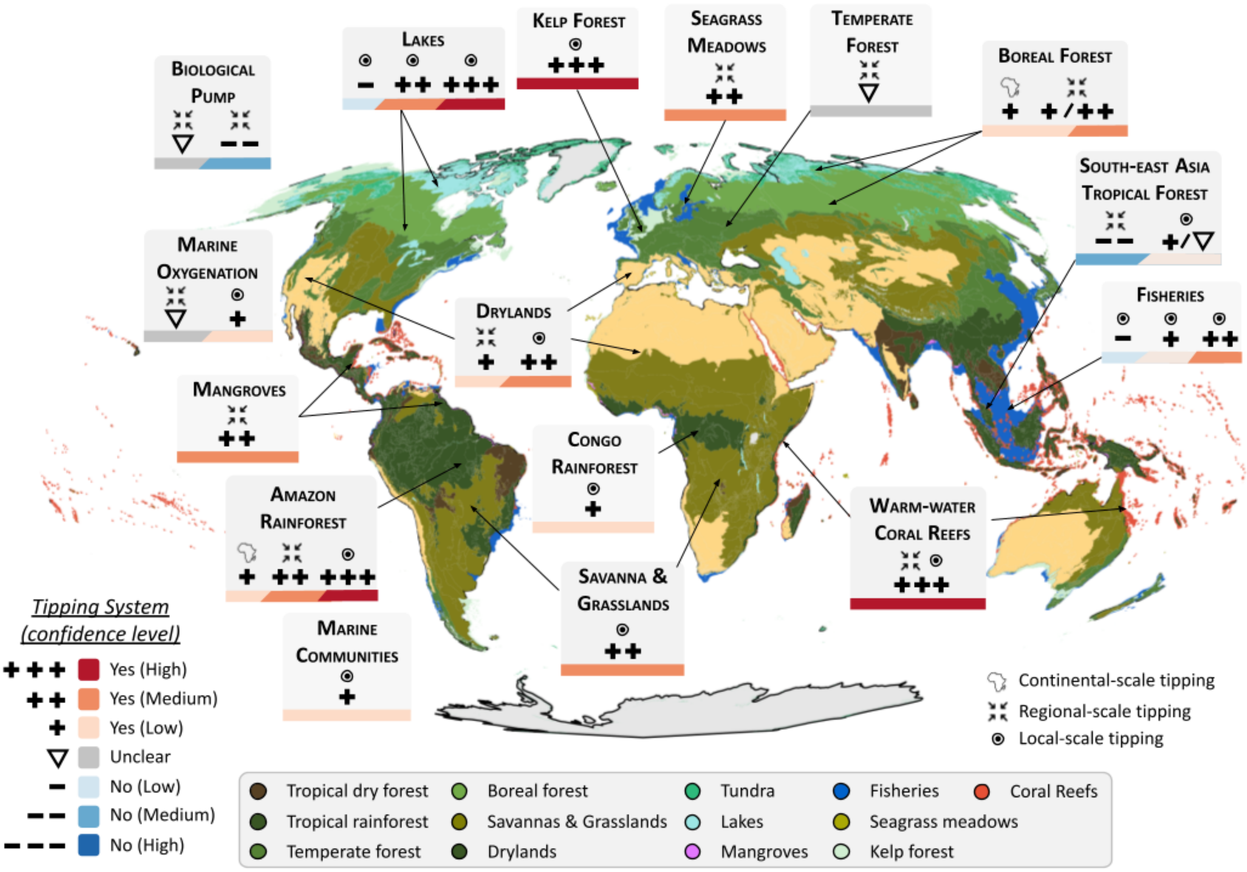



**Figure 1.3.1:** Map of biosphere systems considered in this chapter. Systems are marked by the coloured areas, with terrestrial biomes and mangroves based on biogeographic biomes (Dinerstein et al., 2017), and lakes and ocean biomes on IUCN functional biomes (Keith et al., 2022) (lakes are shown over other biomes for tundra only; fisheries are spread across the global ocean, but are marked only on key coastal seas for simplicity). Labels indicate which of the systems are in this report considered a tipping system (+++ high confidence, ++ medium confidence and + low confidence), which are not (- - - high confidence, - - medium confidence and - low confidence), and which are currently uncertain (▽).

**Table 1.3.1: Summary of evidence for tipping dynamics, key drivers and biophysical impacts in each system considered in this chapter**

| System (and potential tipping point) | Key drivers | Key biophysical impacts (see S2 for societal impacts) | Key feedbacks | Evidence base | Abrupt / large rate change? | Critical threshold(s)? | Irreversible? (decadal / centennial) | Tipping system? |
|---|---|---|---|---|---|---|---|---|
| Forests | | | | | | | | |
| **Amazon rainforest (dieback)** | **DC: atmospheric warming (↗)**<br>**NC: deforestation / degradation (↗)**<br>**DC: drying (↗)**<br>**CA: fire frequency/intensity increase (↗)**<br>DC: heatwaves (↗)<br>CA: ENSO intensification (e.g. Amazon, SE Asia) (↗)<br>CA: AMOC / SPG weakening / collapse (e.g. Amazon) (↗)<br>CA: terrestrial greening (↘, declining) | • Biodiversity loss<br>• Regional rainfall reduction (e.g. from Amazon dieback across Amazon Basin & Southern American Cone)<br>• Carbon emissions (amplifying global warming)<br>• Remote impacts on rainfall patterns all over the planet | Moisture recycling, fire, albedo | • Models<br>• Observations (local scale) | ++ | 1000-1250mm annual rainfall<br>-400 to -450mm max. accumulated water deficit<br>7-8m dry season length<br>~20-40% deforestation<br>~3.5°C (2-6°C) global warming | ++ | **+++ (local)**<br>**++ (partial dieback / regional)**<br>**+ (full dieback / continental)** |
| **Congo rainforest (dieback)** | | | | | + | ~1350mm mean annual rainfall; climate change increasing rainfall | + | **+ (local)** |
| **SE Asia rainforest (dieback)** | | | | | - | ~1550mm mean annual rainfall | - | +? (local)<br>- - (regional) |

| System (and potential tipping point) | Key drivers | Key biophysical impacts (see S2 for societal impacts) | Key feedbacks | Evidence base | Abrupt / large rate change? | Critical threshold(s)? | Irreversible? (decadal / centennial) | Tipping system? |
|---|---|---|---|---|---|---|---|---|
| **Boreal forest (southern dieback)** | **DC: drying (↗)**<br>**CA: fire frequency/intensity increase (↗)**<br>**DC: atmospheric warming (↗)**<br>**CA: permafrost thaw (↗)**<br>**CA: insect outbreaks (↗)**<br>NC: deforestation / degradation (↗)<br>DC: heatwaves (↗)<br>CA: terrestrial greening (↘)<br>CA: vegetation albedo (↗)<br>CA: sea ice albedo decline (↗)<br>DC: precipitation change (↘,↗) | • Biodiversity loss<br>• Carbon emissions from dieback, carbon drawdown from expansion<br>• Complex regional biogeophysical effects on warming - dieback = higher albedo (cooling) but less evaporative cooling (warming) & vice versa for expansion | Fire, albedo, moisture recycling | • Models<br>• Observations<br>• Experiments | ++ | ~4°C (1.4-5°C) | + [~100 yr] | **++ (partial / regional)**<br>**+ (continental)** |
| **Boreal forest (northern expansion)** | | | Fire, albedo, moisture recycling | • Models<br>• Observations<br>• Experiments | + | ~4°C (1.5-7.2°C) | + [~100 yr] | **+ (partial / regional)** |
| **Temperate forests (dieback)** | **DC: atmospheric warming (↗)**<br>**DC: droughts (↗)**<br>**DC: heatwaves (↗)**<br>CA: insect outbreaks (↗)<br>CA: windthrow (↗)<br>NC: deforestation & fragmentation (↗)<br>CA: fire frequency increase (↗) | • Biodiversity loss<br>• Carbon emissions<br>• Regional warming in summer due to less evaporative cooling, less cloud cover<br>• Less atmos. water supply<br>• Less groundwater recharge | Moisture recycling, soil moisture -atmosphere, interacting disturbances, albedo | • Models<br>• Observations<br>• Experiments | ++ | Widespread thresholds uncertain | - [decades] | ? (partial / regional) |
| **Savannas, Grasslands & Drylands** | | | | | | | | |

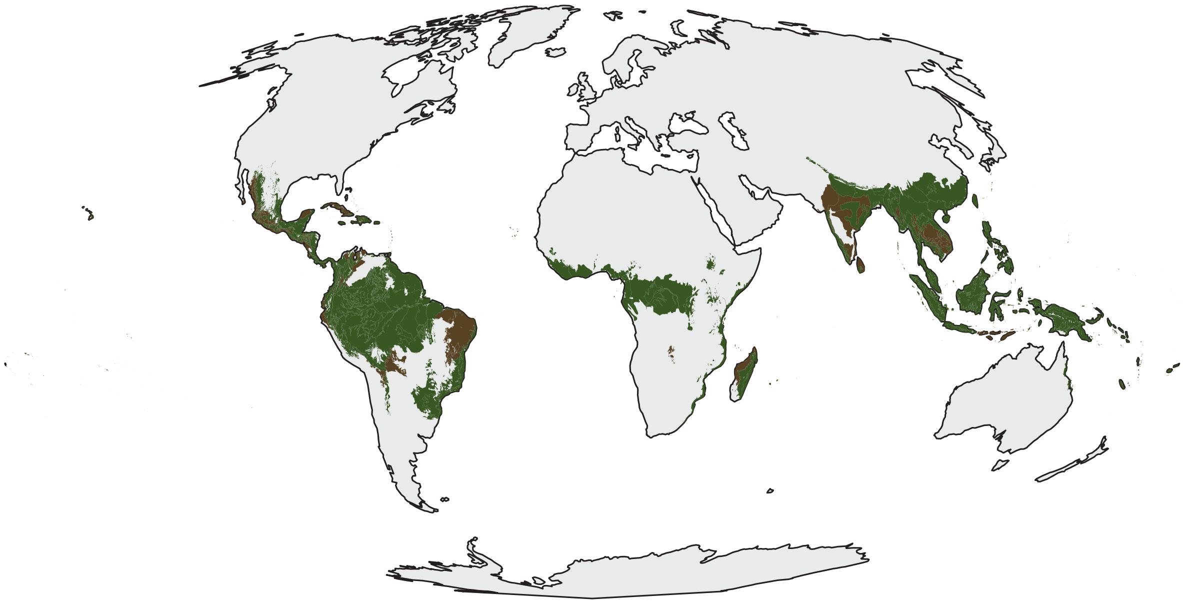
| System (and potential tipping point) | Key drivers | Key biophysical impacts (see S2 for societal impacts) | Key feedbacks | Evidence base | Abrupt / large rate change? | Critical threshold(s)? | Irreversible? (decadal / centennial) | Tipping system? |
|---|---|---|---|---|---|---|---|---|
| **Savanna & Grasslands (degradation)** | **NC: fire suppression (↗)**<br>**NC: overgrazing (↗)**<br>DC: increased precipitation intensity (↗)<br>CA: terrestrial greening (↗)<br>NC: afforestation (↗)<br>CA: ocean circulation shift (e.g. Sahe), ↗) | • Biodiversity loss<br>• Groundwater depletion (with encroachment)<br>• Nutrient cycle disruption<br>• Reduced fires (with encroachment) | Fire, grazing | • Models<br>• Observations (remote sensing & fieldwork) | + | Regionally variable mean annual rainfall; thresholds highly localised;<br>Fire percolation threshold ~ 60% flammable cover | ++ | **++ (local to landscape)**<br>? (regional) |

| System (and potential tipping point) | Key drivers | Key biophysical impacts (see S2 for societal impacts) | Key feedbacks | Evidence base | Abrupt / large rate change? | Critical threshold(s)? | Irreversible? (decadal / centennial) | Tipping system? |
|---|---|---|---|---|---|---|---|---|
| **Drylands (land degradation)** | **DC: drying (↗)**<br>**DC: atmospheric warming (↗)**<br>**NC: land use intensification (e.g. livestock, agriculture, urbanisation)(↗)**<br>DC: extreme events (heatwaves, floods) (↗)<br>DC: increased rainfall variability (↗)<br>CA: terrestrial greening (↘)<br>CA: insect outbreaks (↗)<br>CA: invasive species (↗) | • Biodiversity loss<br>• Aridification / Desertification<br>• Groundwater depletion (with encroachment)<br>• Regional rainfall changes<br>• Shift in species composition (e.g. shrub encroachment)<br>• Vegetation recruitment | Soil fertility, / moisture / microbes, vegetation structure, veg-rainfall, fire, herbivory | • Models<br>• Observations (current & historical)<br>• Field experiment | ++ | Aridity index (0.54,0.7 and 0.8) (limited reliability of aridity measures; lack of temporal evidences for some thresholds) | + (shorter timescales possible, e.g. via active restoration) | **++ (local to landscape)**<br>**+ (regional)** |
| **Freshwater** | | | | | | | | |
| **Lakes (eutrophication-driven anoxia)** | **NC: nutrient pollution (↗)**<br>DC: atmospheric warming (↗)<br>DC: precipitation changes (↗) | • Biodiversity loss<br>• Water quality declineIncreased GHG emissions | Anoxia-driven P release, trophic cascades | • Observations<br>• Models<br>• Experiments | +++ | 20-30 mg P/l<br>No clear warming/ rainfall thresholds | ++ (decadal) | **+++ (localised, widespread)** |
| **Lakes (DOM loading - 'browning')** | **CA: terrestrial greening (↗)**<br>**NC: afforestation (↗)**<br>DC: atmospheric warming (↗) | • Biodiversity loss<br>• Increased GHG emissions | Anoxia-driven P release | • Observations<br>• Models | + | >10 mg DOC/l | ++ (decadal) | **++ (localised, widespread in boreal)** |
| **Lakes (appearance / disappearance)** | **CA: permafrost thaw-related thermokarst formation / drainage (↗)**<br>**CA: glacier lake formation / drainage (↗)** | • Biodiversity loss<br>• Increased GHG emissions | (can be driven by thermokarst) | • Observations | +++ | As for permafrost thaw | +++ (centennial) | - (localised, widespread on tundra) |
| **Lakes (N to P limiting switch)** | **NC: nutrient pollution (atmos. deposition) (↗)** | • Biodiversity loss | N/A | • Observations | ++ | Related to elemental ratio | ++ (decadal) | - (localised, regions with high N-deposition) |
| **Lakes (salinisation)** | **DC: atmos. warming (↗)**<br>**DC: drought (in arid regions) (↗)**<br>**CA: water use intensification (↗)** | • Biodiversity loss<br>• Reduced GHG emissions | Salt release from sediment | • Observations | + | Species-specific salinity threshold | ++ (decadal) | - (localised, arid regions) |

| System (and potential tipping point) | Key drivers | Key biophysical impacts (see S2 for societal impacts) | Key feedbacks | Evidence base | Abrupt / large rate change? | Critical threshold(s)? | Irreversible? (decadal / centennial) | Tipping system? |
|---|---|---|---|---|---|---|---|---|
| **Lakes (invasive species)** | **CA: warming-driven range expansion (↗)**<br>**NC: human-mediated introduction (↗)** | • Biodiversity loss | N/A | • Observations<br>• Models | + | Cannot be defined | ++ (decadal - centennial) | - (localised, widespread) |
| **Coastal** | | | | | | | | |
| **Warm-water coral reefs (die-off)** | **DC: ocean warming (↗)**<br>**DC: marine heatwaves (↗)**<br>**CA: disease spread (↗)**<br>CA: ocean acidification (↗)<br>NC: pollution (nutrient / sediment) (↗)<br>NC: disruption (ships, over-harvesting) (↗)<br>CA: invasive species (↗)<br>DC: storm intensity (↗)<br>CA: sea level rise (↗) | • Biodiversity loss (ecosystem collapse, ~25% marine species have life stages dependent on coral reefs)<br>• Loss of commercial & artisanal fisheries, and other sectors<br>• Coastal protection loss | Thermal stress leading to symbiont expulsion, | • Observations<br>• Models | +++ | **Region and reef dependent:**<br>• ~1.2°C (1.0-1.5°C) GW<br>• Temporally variable heat stress (8-12 Degree Heating Weeks)<br>• Long-term consequences of >350 ppm atmospheric $CO^2$<br>• Acidification threshold uncertain | ++ (decadal) | **+++ (localised)**<br>**+++ (regionally clustered)** |

### 1.3.2.1 Tropical forests

Tropical forests cover around 1.95bn hectares (including degraded portions), and are key components of the Earth system (Pan et al., 2011) (Figure 1.3.2). They are home to a disproportionate amount of Earth's species (e.g. Slik et al., 2015; Pillay et al., 2021), store huge amounts of carbon (circa 471 ±93 GtC) in their soils and biomass, and, through evapotranspiration and their effect on cloud formation through production of aerosols and cloud condensation nuclei, have an overall cooling and moistening effect at regional scales (SPA, 2021; IPCC AR6 WG2 2021). They are also home to many Indigenous peoples and local communities, with a long history of human habitation and high biocultural diversity (Ellis et al.,2021).



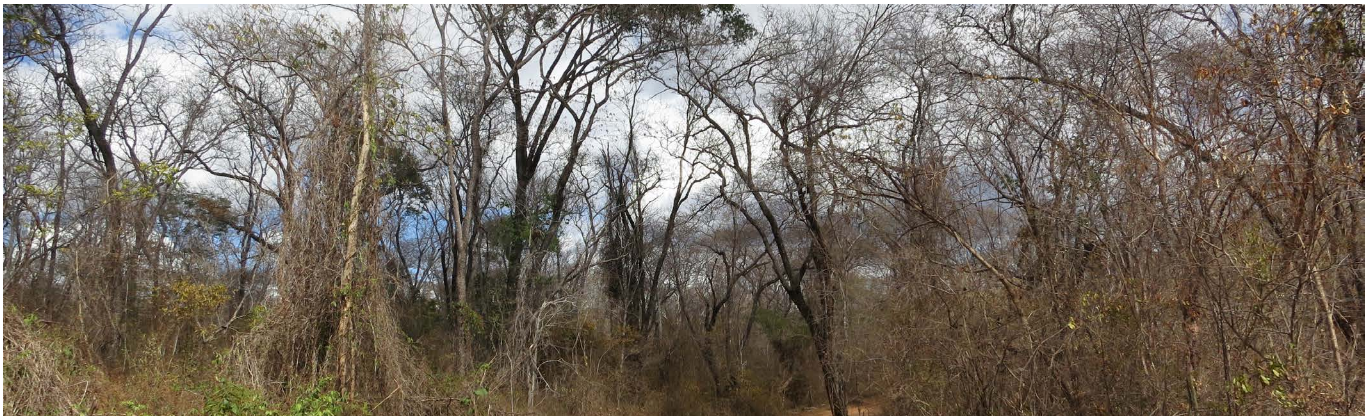


**Figure 1.3.2:** Top: map showing global extent of tropical forests, including tropical rainforests (dark green) and tropical dry forests (brown) (source: Dinerstein et al. (2017)) . Middle: photo of mature rainforest in Tapajós National Forest, Brazil (credit: Boris Sakschewski). Bottom: photo of arboreal Caatinga, a tropical dry forest formation in Eastern Brazil (credit: Kyle Dexter).

As well as experiencing deforestation and degradation due to land use change across the tropics (IPBES 2019), tropical forests in South America and Asia have been undergoing unprecedented climate-driven disturbances such as increasing dry season length and intensity, more intense and frequent rainfall and temperature extremes (Lapola et al., 2023; SPA, 2021). For instance, recent extreme droughts – mainly driven by climate variability modes such as the El Niño Southern Oscillation (ENSO) in 2014-2016 and the Atlantic dipole in 2005 and 2010 (e.g. Marengo et al., 2008; Marengo et al., 2011; Jimenez-Muñoz et al., 2016; see Chapter 1.4) – have caused extensive tree mortality, even up to 36 months after peak drought (e.g. Phillips et al., 2009; Phillips et al., 2010; Berenguer et al., 2021). Given the variability of forests across the tropics, their responses to global changes are likely to differ (Allen et al., 2017). Nonetheless, even subtle changes in their structure, composition and functioning could affect the global carbon and water cycles (e.g. Esquivel-Muelbert et al., 2019; Barros et al., 2019; Hirota et al., 2021).

Here we also consider deciduous and semi-deciduous forests (often referred to as dry forests) that coexist with evergreen forests in regions with around 1,000-2,000mm of annual rainfall, i.e. non-arid or dryland regions (Dexter et al., 2018). These dry forests may resemble (in terms of tree species composition) the dry forests in arid or dryland regions. However, because they exist in climates that can form continuous, high fuel-load flammable grass layers when canopies are opened (which is not the case in drylands), their dynamics are more comparable to neighbouring moist forests.

**Evidence for tipping dynamics**

Two positive/amplifying feedbacks are among the most plausible mechanisms that could lead to tipping dynamics in tropical forests, one at broader regional scales potentially causing large-scale forest collapse, and another at local scales potentially causing local forest collapse (Figure 1.3.3 and Box 1.3.2).

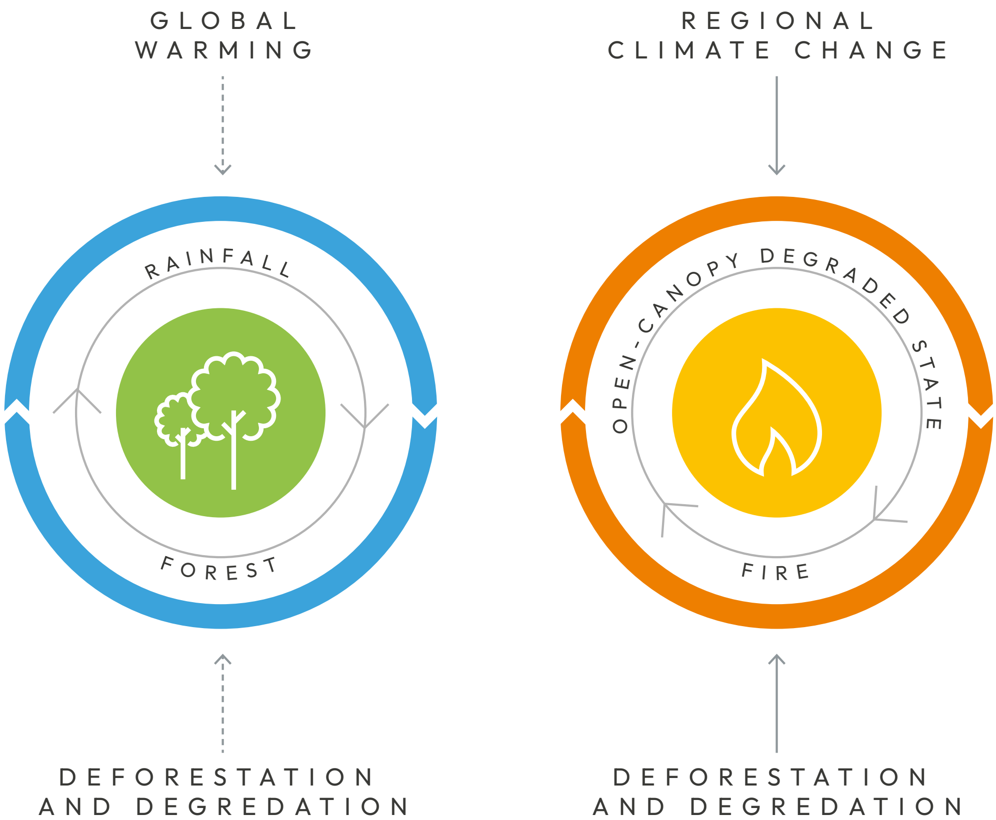



**Figure 1.3.3:** Diagram with positive/amplifying feedback loops that may cause large- and local-scale tipping events in tropical forests. (a) Regional climatic conditions are changing in response to global warming and also to deforestation, both of which contribute to weakening the forest-rainfall feedback. Reductions in rainfall cause water stress, increasing tree mortality and forest loss, further weakening the feedback, which could cause a large-scale forest collapse of the Amazon. (b) Interactions and feedbacks among the vegetation and fire can arrest the ecosystem in an open vegetation state, thus causing a local-scale forest collapse.

At regional scales, the forest-rainfall feedback is believed to be the dominant mechanism stabilising tropical forests by increasing annual rainfall levels and reducing its seasonal and interannual variability (Staal et al., 2020; Sternberg, 2001). However, under certain conditions it can instead amplify forest loss. Accumulated deforestation or forest loss reduces forest cover, which decreases evapotranspiration and moisture flow downwind, thus reducing regional rainfall (Smith et al., 2023). This in turn may increase tree mortality in downwind forest (Phillips et al., 2009; Berenguer et al., 2021), and beyond a threshold could lead to self-sustaining forest loss in drier areas of forest (Zemp et al., 2017; Staal et al., 2020) (Figure 1.3.3).

In the Amazon, on average, around 30 per cent of the water precipitating has been evaporated within the region beforehand at least once, but with large spatial differences: in the western Amazon, almost all precipitation has previously evaporated from the basin (Zemp et al., 2014; Staal et al., 2018). In the Congo basin, almost half of all precipitation originates from the Congo forest itself (Tuinenburg et al., 2020; Te Wierik et al., 2022). For Australian and Asian forests, evidence is still lacking, but this feedback likely has less effect on forest resilience due to less dependence on precipitation stemming from land evapotranspiration due to the major importance of monsoons (Staal et al., 2020).

At the local scale, evidence from across the tropics (Cochrane et al., 1999; Staver et al., 2011; van Nes et al., 2018) suggests that a fire-vegetation feedback can maintain the ecosystem in an open vegetation state: with less tree cover, fires spread more easily due to more flammable grassy fuels and because the air is drier in an open landscape without the local moistening effect of forest canopies. The resulting enhanced fire occurrence can in turn prevent the re-establishment of trees and maintain a more open vegetation state (Martinez-Cano et al., 2022; Drüke et al., 2023). This alternative open vegetation state could be either a natural savanna with native plant species (Flores and Holmgren, 2021; Beckett et al., 2022) or a degraded open-vegetation state when invasive plants are dominant (D'Antonio and Vitousek, 1992; Veldman and Putz, 2011; Malhi et al., 2014; Barlow et al., 2018) (Figure 1.3.3).

The effects of the fire-vegetation feedback are amplified by the regional forest-rainfall feedback (Staal et al., 2020). Moreover, forest loss may increase global warming by releasing carbon to the atmosphere, which further reduces regional moisture flows, causing more forest loss (Canadell et al., 2021). Also, climate change may change wind directions and residence times of moisture in a warmer atmosphere (Gimeno et al., 2021). Tropical forest loss also may change atmospheric circulation patterns (Portmann et al., 2022) and increase regional and global warming through reductions in cloud cover and evapotranspiration.

Among tropical forests, the Amazon forest has most evidence for potential tipping points. Analysis based on early warning signals (see Chapter 1.6) indicates that over 75 per cent of the Amazon has lost resilience since the early 2000s (Boulton et al., 2022). This decline is focused mostly closer to human disturbance, as well as in the drier south and east previously identified as 'bistable' (i.e. with two possible alternative states) due to the forest-rainfall feedback and thus is more vulnerable to tipping (Staal et al., 2020). While the Amazon has acted as a carbon sink due to $CO_2$ fertilisation, in mature forest this sink peaked and started declining in the 1990s (Hubau et al., 2020) and when including degraded forest (also predominantly in the drier south and east) the Amazon as a whole is now a carbon source (Gatti et al., 2021). Recent CMIP6 models indicate that localised shifts in peripheral parts of the Amazon forest system are more likely than a large-scale tipping event (IPCC AR6 WG1 Ch5, 2021; Parry et al., 2022). However, the latter cannot be ruled out (Hirota et al., 2021) because several compounding and possibly synergistic disturbances (e.g. combining an extreme hot drought with forest fires) may play a role in reducing forest resilience, with greater resilience loss closer to human activities (Boulton et al., 2022). Such synergies are generally not considered in Earth system models (Willcock et al., 2023).

A global warming threshold of ~3.5°C (2-6°C) has been estimated (Armstrong McKay et al., 2022), partly based on a few modelling studies that simulate some kind of nonlinear decrease in modelled properties of the Amazon forest, at least on small scales (Gerten et al., 2013; Drijfhout et al., 2015; Nobre et al., 2016; Boulton et al., 2017; Parry et al., 2022). However, most CMIP6 models do not include dynamic vegetation modules (Song et al., 2021; Canadell et al., 2021), which might make the forest artificially stable (Zemp et al., 2017). Models including deforestation, fire and dynamic vegetation have simulated widespread local-scale dieback (e.g. Cano et al., 2022; Parry et al., 2022), and also larger scale dieback in potential vegetation models (e.g. Salazar and Nobre, 2010).

Evidence pointing against a large-scale Amazon tipping point stems from palaeoclimate reconstructions suggesting that at least some parts of the Amazon forest have been resilient to past reductions in rainfall (Wang et al., 2017; Kukla et al., 2021) and temperatures as high as projected by climate models for the rest of the century (Steinthorsdottir et al., 2020). However, these were under more stable climate conditions (and before Pleistocene with different geographic effects on climate due to tectonics; (Brierley and Fedorov, 2016), with the current rate of warming far greater than during past climate changes (Zeebe et al., 2016; Osman et al., 2021). Geographically limited data means partial dieback elsewhere cannot be ruled out for drier intervals (Wang et al., 2017; Kukla et al., 2021), particularly in the drier south, where drying is currently leading to greater resilience loss (Boulton et al., 2022). Additionally, compounding disturbances are becoming increasingly widespread across the Amazon, even in remote central parts of the system, which is leading to resilience loss (Boulton et al., 2022) and could help trigger forest dieback at larger scales (Kukla et al., 2021; Wilcock et al., 2023).

Other tropical forests have evidence for local tipping points, but are less likely to cross them. The Congo has also been suggested as a possible tipping system (Staal et al., 2020) as it may also host a large area of bistable forest with some amplification by forest-rainfall feedback (Staver et al., 2011). However, because climate models indicate wetting across large parts of the Congo, it is not considered a tipping system in response to global warming (Armstrong McKay et al., 2022). The south-east Asian rainforests lack a strong regional forest-rainfall feedback and tend to have enough rainfall from ocean proximity for forests to remain stable, thus they are not considered a tipping system in relation to global warming (Armstrong McKay et al., 2022). Other tropical forests such as the Choco in Central America or Brazilian Atlantic Forests have not been assessed in detail.

Plants can reduce moisture transpiration in response to water limitation on very short timescales (hours to days), followed by water cycle feedbacks (weeks). Deforestation has a similarly fast effect on rainfall, as loss of trees can immediately reduce evapotranspiration. Large-scale forest dieback events in response to global warming can only be expected on the timescale of decades to centuries (Armstrong McKay et al., 2022). At a local scale, empirical evidence from the Amazon and from Africa has shown that forests can shift into savannas within a few decades after repeated fires (Flores and Holmgren, 2021; Beckett et al., 2022), and on larger scales tipping may occur faster (Cooper et al., 2020).

An Amazon tipping point would have global impacts from possibly large losses of carbon to the atmosphere. The best estimates suggest that a large-scale collapse of 40 per cent of the forest before the end of this century could lead to emissions of ~30 GtC and an additional global warming of ~0.1°C (Armstrong McKay et al., 2022). The Amazon dieback would also lead to substantial rainfall reductions across the Amazon basin and in to the Southern Cone of South America (Costa et al., 2021), and may also directly influence distant parts of the Earth system via 'teleconnections', for example to the Tibetan Plateau (Liu et al., 2023).

#### Assessment and knowledge gaps

The feedbacks that could contribute to tipping behaviour are relatively well understood in principle, yet there are large uncertainties surrounding the effects of climate and land use changes on these feedbacks. For instance, $CO_2$-fertilisation is expected to increase forest resilience locally, but it also increases water-use efficiency, reducing forest transpiration, and may thus weaken the forest-rainfall feedback and regional forest resilience (Brienen et al., 2020; Sampaio et al., 2021; Kooperman et al., 2018; Li et al., 2023). $CO_2$-fertilisation of tropical forests may also be overestimated in current Earth system models (Terrer et al., 2019; Hubau et al., 2020; Wang et al., 2020). Moreover, the actual thresholds and the extent to which tipping behaviour can be expected across heterogeneous landscapes and forest communities are much less certain (Levine et al., 2016; Longo et al., 2018; Sakschewski et al., 2021).

Considering only the Amazon as a rainforest tipping system, we have medium confidence in its potential for tipping of its bistable area (~40 per cent of the forest, predominantly in the drier south and east; Staal et al., 2020), with low confidence in the estimated tipping points and possibility of a large-scale collapse. The Congo may also be vulnerable to localised tipping (low confidence), but is unlikely to tip as a result of climate change, and localised tipping is possible but uncertain in south-east Asian rainforests.

Confidence in the tipping behaviour of tropical forests can be greatly improved through further development of models. Models can include dynamic vegetation modules and land use change to improve the representation of the forest-rainfall feedback, which would likely result in more drastic drying under high-deforestation scenarios (Parry et al., 2022). Incorporating fire dynamics in these modules would also likely result in a more bistable system (Drüke et al., 2023). In contrast, allowing for local vegetation adaptation (such as rooting depth) by including more plant types and traits in these modules would help better resolve the effect of landscape heterogeneity on tipping dynamics (Langan et al., 2017; Sakschewski et al., 2021), which may reduce the abruptness of the transition to an open degraded state (Levine et al., 2016). Efforts to increase ecological understanding of the feedback mechanisms and processes described here through observations (such as recent field studies on plant characteristics related to drought mortality throughout the Amazon basin (Tavares et al., 2023), or on the growth-survival tradeoff (Oliveira et al., 2021)) would help better understand forest dynamics and represent them in models.

### 1.3.2.2 Boreal forests and tundra

Boreal forests, also called 'Taiga', span around 1,135 million hectares, all located in the northern hemisphere (Pan et al., 2011) (Figure 1.3.4). They are vital for climate regulation, storing circa 272 (± 23) GtC, mostly below ground (Pan et al., 2011; Mayer et al., 2020). Management varies, but illegal logging constitutes a critical driver of boreal forest loss. Boreal forest growth is constrained by a short vegetation period, and their dynamics involve large-scale disturbances such as insect outbreaks and fire (with fire percolation dynamics important – see 1.3.2.4). While disturbance regimes differ in Eurasian and American forests, an overall increase in disturbances has been observed over past decades, fuelling worries about a wider loss of resilience.

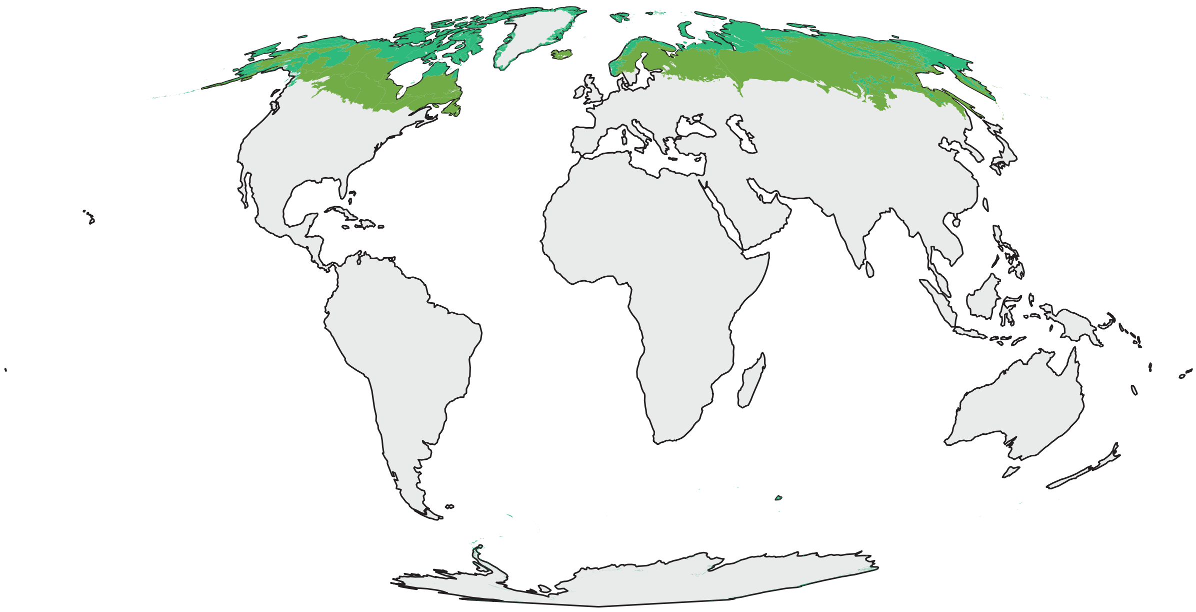


**Figure 1.3.4:** Top: map showing extent of boreal forests (light green) and tundra (blue-green) biomes (source: Dinerstein et al. (2017)). Bottom: photo of boreal forest and swamps, southern Norway (credit: Boris Sakschewski).

### Evidence for tipping dynamics

Boreal forest dieback has already been identified as a potential tipping element in the climate system in Lenton et al., (2008) and further assessed in the IPCC AR5 WG2 Report in 2014 and in the WG1 and WG2 reports of AR6 in 2021 and 2022. The IPCC SR1.5 (Hoegh-Guldberg et al., 2018) and also the most recent assessment by (Armstrong McKay et al., 2022) differentiate between southern boreal forest and northern tundra tipping points. The southern boreal forest tipping point refers to a dieback of southern boreal forests that lead to a state-shift to an almost treeless state (to steppe/prairie), while the northern tundra tipping point refers to an expansion of tree cover into currently treeless tundra ecosystems.

There is little additional evidence for a boreal forest tipping point since the assessment of Armstrong McKay et al., (2022). Significant losses in tree cover driven by fires and logging were identified for the southern boreal forests of North America between 2000 and 2019 (Rotbarth et al., 2023). In contrast, interior boreal forests have become denser. There has been no clear sign of a northward expansion of the boreal forests of North America (Rotbarth et al., 2023). Similarly, Burrell et al. (2021) found that the forests of southern Siberia might have approached a tipping point as fire regimes have intensified, causing widespread regeneration failure. Moreover, Siberian larch foliage is sensitive to warming, with temperatures potentially exceeding a threshold by 2050 after which forest dieback can be expected (Rao et al., 2023).

A range of mechanisms contribute to the feedback processes associated with boreal tipping points (see Box 1.3.2 for more on forest feedbacks). For the southern boreal forest, the recent surge in forest disturbances, such as the extreme forest fires in Canada in summer 2023, is noteworthy because they constitute a substantial change in forest dynamics and resilience that, combined with failure to regenerate, could initiate regional tipping. In particular, the southern trailing edge of boreal forests has been identified as prone to compound and interacting disturbances, including droughts, windstorms, fires, large herbivores and insect outbreaks (Frehlich and Reich, 2010). For instance, increasing water stress reduces tree resistance against insects, and increases the size and severity of wildfires.

Southern boreal tipping points are driven by forest dieback from disturbances (Lenton et al., 2008). Empirical evidence from satellite data suggests that disturbances are responsible for switches between states rather than causing gradual change (Scheffer et al., 2012; Abis and Brovkin, 2017). Rotbarth et al. (2023) confirm that processes dominating the dieback of southern boreal forests and the northward expansion of forests into tundra diverge and that a northward expansion is not compensating for declines in the southern boreal forests of North America.

Climate change will further intensify disturbance regimes (Seidl et al., 2017), with fire regimes expected to increase significantly in boreal forests (Velasco Hererra et al., 2022). In Canada, fire frequency could increase up to 50 per cent by the 21st century under climate change (Flannigan et al., 2013). A doubling of fire frequency and increased wind activity during the 21st century will likely cause a significant decrease in coniferous forests, potentially replaced by early successional broadleaved tree species (Anoszko et al., 2022; Liu et al., 2022).

The increase in fire could potentially modify the forest microclimate, so that subsequent fires and droughts become more likely, causing a change in vegetation dynamics. For instance, Whitman et al. (2019) found that drought after fire exacerbates regeneration failure. Overall, drought-induced mortality will likely rise more in western than eastern North American regions (Peng et al., 2010). Moreover, insects, such as mountain pine beetles might expand into North American boreal forests, causing changes in ecosystem dynamics (Safranyik et al., 2010; Jarvis and Kulakowski, 2015). For instance, severe defoliation could impede birch forest recovery (Vindstad et al., 2018).

If these changes in disturbances cause widespread mortality while, at the same time, forests fail to regenerate, the forest might tip into an almost treeless state. Stevens-Rumann et al., (2022) suggest that a combination of changing climate patterns and disturbance regimes could primarily cause regeneration failure in coniferous forests. Bailey et al., (2021) highlighted the importance of temperature-moisture interactions for successful seedling establishment at the upper treeline in the Southern Rocky Mountains. However, over the past decade, no seedling establishment occurred at any site, suggesting that a threshold for regeneration may have been passed. Regeneration failure of boreal forests might occur with warming alone (+1.6°C to +3.1°C increase in one local warming experiment), but temperature thresholds are reduced if an increase in temperature is combined with reduced precipitation (Reich et al., 2022).

The sensitivity of coniferous tree recruits to climate change is overall higher than for broadleaved tree regeneration (Reich et al., 2022; Stevens-Rumann et al., 2022). In addition, natural disturbances might more likely cause state-shifts of coniferous than broadleaved-dominated boreal forest (Thom, 2023) as broadleaved tree species have an overall higher resprouting ability than conifers (Thom et al., 2021). Topographic complexity and peatlands may act as refugia from fire (Kuntzemann et al., 2023; Rogeau et al., 2018), thus reducing the likelihood of regeneration failure and state shifts. If widespread mortality becomes an increasing issue in northern forests, reduced microclimatic buffering of forests to increasing temperature might accelerate the thawing of permafrost in the boreal biome, causing additional releases of greenhouse gases – further interacting with the climate system [See Chapter 1.2.2.4 on Permafrost].

An increase in abundance of woody plants and advancing shrublines into the Arctic tundra is likely as climate changes (Mekkonen et al., 2021). This shrubification driven by warmer climate is also accompanied by northward treeline migration. A recent review of more than 400 treeline site locations suggested that at about two-thirds of treeline sites' forest cover had increased in elevational or latitudinal extent (Hansson et al., 2021). Main drivers of treeline migration are an increase in the rate of seedling success through warmer summers and increased winter temperatures. The change from tundra and peatlands to boreal forests can be nonlinear. Experimental work in boreal peat bogs reveals positive interactions between shrub cover and tree recruitment in which shrub cover favours tree seedlings and, in turn, higher tree basal area fosters shrub biomass, potentially triggering tipping towards high tree cover (Holmgren et al., 2015). As with southern dieback, interaction with permafrost thaw is also likely, but is complex and currently uncertain.

There are no clear thresholds for boreal forest dieback beyond the initial estimates already presented in Armstrong McKay et al., [2022]. With low confidence, they estimate a southern dieback tipping point at a global warming threshold of ~4°C (1.4-5°C) and a tipping timescale of ~100 (50-?) years, and a northern expansion into tundra tipping point at an estimated global warming threshold of ~4°C (1.5-7.2°C) and a tipping timescale of ~100 (40-?) years. Regeneration failure of southern boreal forests might occur with warming alone, while those thresholds are even lower if precipitation amounts also decrease (Reich et al., 2022).

### Assessment and knowledge gaps

We assess with medium confidence that larger parts of boreal forests will approach a southern dieback tipping point and with low confidence that they will expand northwards as global temperatures increase by 3-4°C, if precipitation amounts and patterns remain similar. Yet, this threshold depends on multiple factors such as human and natural disturbances.

The capacity for adaptation and resilience is among the key uncertainties. Biodiversity, among other factors, might influence tipping dynamics as a diverse ecosystem may be more resistant to reaching tipping points, yet the effects of compositional and structural diversity require further investigation. Furthermore, although there is strong evidence and confidence in the increase of natural disturbances in boreal forests it remains uncertain whether they will truly lead to the transgression of a tipping point, pushing the southern range of boreal forests into an alternative, treeless state.

While in the southern boreal region the main mechanisms causing tipping points are relatively clear, for the northern tundra expansion tipping point the mechanisms sustaining large-scale abrupt state-shifts are not as evident. Disturbances in this region may be weaker and more localised, and the replacement of tundra by forest might occur more gradually. Yet, it is unclear if regeneration failure drives a self-sustaining feedback loop hindering recovery due to soil dryness or extreme conditions, causing a tipping point.

Further uncertainties linked to tipping points requiring further investigation include testing:

- interactions between climate, atmospheric forcing and disturbances;
- cascading and compounding disturbances;
- the existence of a 'fast-in, fast-out' behaviour of release and recovery in boreal forests;
- whether changes are self-reinforcing and perpetuating forest loss (or gain in the case of the northern tipping point);
- the extent of southern forest loss vs. northern forest expansion; and
- the role of human interventions, such as forest management on tipping dynamics.

### Box 1.3.2: Forest feedbacks that could lead to tipping

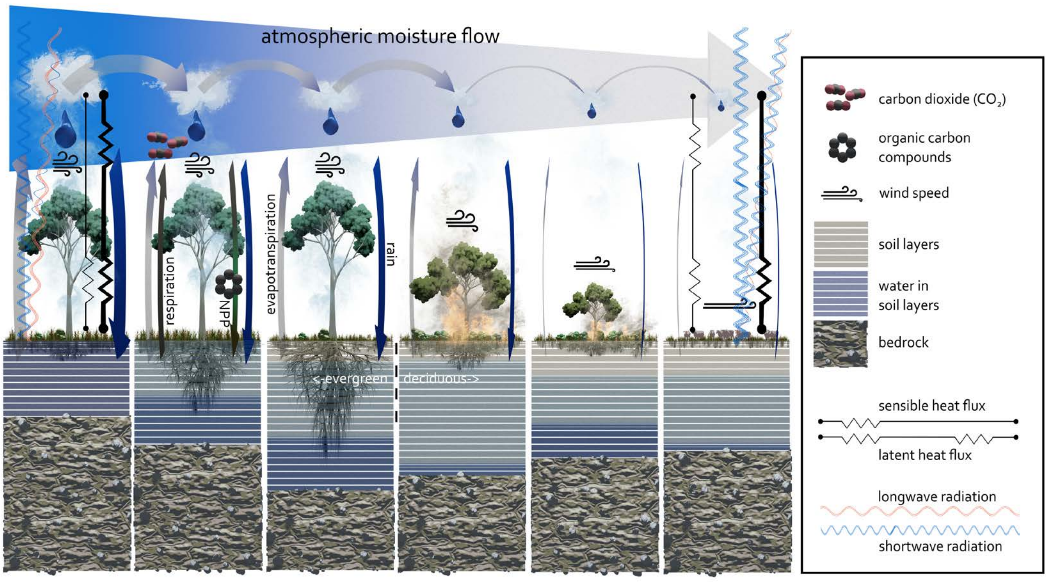



**Figure 1.3.5:** A conceptual regional transect from moist (left) to dry (right) localities depicting examples of local to regional feedbacks of forest cover with the land and atmosphere.

**Less forest leads to ...**

- less evapotranspiration (less productivity, less interception, less deep roots, etc.), hence reduced atmospheric moisture supply, and therefore reduced local and downwind precipitation, which leads to...
- less tree-produced volatile organic compounds (VOCs) serving as cloud condensation nuclei and therefore reduced local and downwind precipitation, which leads to...
- decreased roughness length of the landscape and hence increased wind speeds, leading to reduced residence time of moisture in the overall forest system, which leads to...
- decreased cloud formation due to less evapotranspiration, less VOCs, higher wind speeds leading to less reflectivity of sunlight, hence higher temperatures and therefore higher atmospheric water demand i.e. drought stress, which leads to...
- increased temperatures due to less evaporative cooling and decreased shading in canopy and ground proximity, hence higher atmospheric water demand i.e. drought stress, which leads to...
- more open canopy, drier understorey and less decomposition hence potentially larger pools of dead material to burn which all increasing fire probabilities, which leads to...
- higher windspeeds, less soil moisture and less soil retention capacity lead to higher erosion, which leads to...
- a surplus of atmospheric $CO_2$ by losing biomass carbon and losing a potential future carbon sink (a forest still capable of increasing biomass due to e.g. $CO_2$-fertilisation) and hence fueling global climate change, which leads to...

**... less forest**

### 1.3.2.3 Temperate forests

Temperate forests cover around 767 million hectares (16 per cent of the global forest area) and represent 34 per cent of global carbon sinks, storing around 119 GtC (Hansen et al., 2010; Pan et al., 2011) (Figure 1.3.6). In this report, we only consider temperate forests as defined in Figure 1.3.1. Mediterranean forests are covered under Drylands [see 1.3.2.5].

In most regions their spatial cover is highly fragmented following a long history of human land-use and forestry practices.

In fact there are only a few temperate forests which are considered 'intact' primary forest (Potapov et al., 2017; Sabatini et al., 2021) and the vast majority are managed by humans using vastly varying forest management techniques and intensities. Current managed temperate forests are often monocultures or mixtures of few tree species with relatively low biodiversity and structural diversity, optimised for high timber yields and certain wood features established under the assumption of stable climate and environmental conditions (instead of optimised for long-term forest resilience).

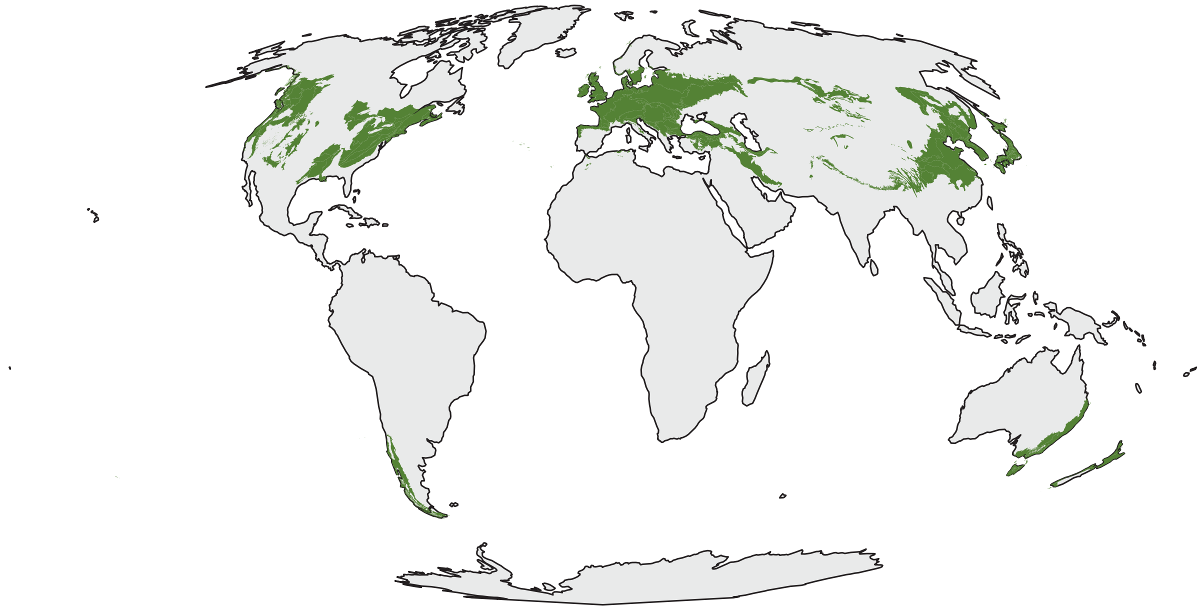


**Figure 1.3.6:** Top: map showing global extent of temperate forests biomes (green) (source: Dinerstein et al. (2017)). Left: photo of a mature temperate forest, Hainich National Park, Germany (credit: Boris Sakschewski,2022). Right: synchronous landscape-scale forest dieback (spruce monoculture) at Harz National Park, Germany (credit: Boris Sakschewski, 2023).

### Evidence for tipping dynamics

In recent years temperate forests globally have suffered enormous damages and losses caused by extreme heat waves and droughts in combination with secondary effects like insect outbreaks and fires (Allen et al., 2010; Buras et al., 2019; Senf et al., 2020; Zhang et al., 2021; Carnicer et al., 2021; Benyon et al., 2023, Forzieri et al., 2022). As many temperate forests are effectively plantations for wood production in most parts of the world, those impacts often occurred in a similar synchronised manner on regional scales. Embedded in landscapes dominated by human land use (segregated by roads, crops, power lines, etc.), many temperate forests feature reduced connectivity and hence less exchange of species or genetic material, which reduces resilience (Sabatini et al., 2021).

More importantly, the extremely low diversity reduces the forest's ability to cope with stress through mechanisms such as portfolio insurance effects or complementarity (Billing et al., 2020). Portfolio insurance effects refer to the idea that having a diverse portfolio of species can help protect the forest against stressors by spreading the risk among different species. Complementarity refers to the idea that different species work together in a complementary way to improve the overall functioning of the ecosystem. However, when there is low diversity, these mechanisms may not be as effective and hence a potential tipping of temperate forests might also be more abrupt than in natural systems. Still, it must be noted that effective support from forest management (by regenerating an area through planting or supporting natural regeneration) can in principle also alleviate some of the pressures that natural systems face.

Besides the clear devastating signals of temperate forest damage and dieback, past assessments have had difficulties classifying temperate forests as tipping systems. In a review by (Thom, 2023) many temperate forest ecosystems were identified as resilient and/or resistant to increasing disturbance regimes and unlikely to shift towards alternative states in the very near future at large scale. However, drastic changes under intensifying future pressures such as climate change cannot be ruled out. In accordance with these findings, the recent assessment of (Armstrong McKay et al., 2022) has categorised temperate forests as an uncertain potential regional impact tipping system.

So far self-amplifying feedbacks in temperate forest dieback were described for more localised landscape-scale stressors like bark beetle attacks and fire in the Boreal forest section (see 1.3.2.2 and Box 1.3.2) (Hlásny et al., 2021; Fettig et al., 2022). On larger spatial scales it remains less clear whether temperate forests might feature self-amplifying feedbacks strong enough to induce tipping behaviour. However, just as in the tropical zone, the principles of cascading moisture recycling also apply to temperate forests. Any loss of forest cover reduces atmospheric moisture supply, hence reducing precipitation downwind and increasing sensible heat, which can amplify drying and warming in the affected areas (Pranindita et al., 2021). The average net cooling effect of temperate forests compared to grassland was found to be 1-2°C, with maxima of up to 5°C (Zhang et al., 2020). A recent study integrating data and modelling results reports continental-scale cooling effects of regrowing temperate forests on abandoned agricultural areas (Huang et al., 2020).

Related to this, cloud formation probability was found to be higher above forests in comparison to other land cover types in the temperate region (Teuling et al., 2017). Therefore, recent forest damages could have decreased cloud cover during recent droughts and heatwaves further intensifying these events. Furthermore, soil moisture-atmosphere feedbacks related to droughts and heatwaves were reported for the temperate zone (Seneviratne et al., 2010; Jaeger and Seneviratne, 2011) and could indicate that droughts might self-propagate in space and time (Schumacher et al., 2022). A recent study for the US west coast suggests cascading effects of soil moisture and biomass during a multi-year drought (Au et al., 2023). The recent large-scale forest damages and losses in the temperate zone (Senf et al., 2020; Lloret and Batllori 2021) could mark the beginning of self-amplifying and potentially self-sustaining feedbacks, but further work is required to confirm this.

The most important mediator between soil moisture and the atmosphere is vegetation, and forests especially stand out since they access water in great depths via their root systems (Sakschewski et al., 2021; Singh et al., 2020; Fan et al., 2017). Hence, larger-scale forest damage or loss means losing this mediator, further decreasing atmospheric moisture supply and downwind rainfall. This becomes particularly significant when, during droughts, precipitation becomes increasingly dependent on water evaporated from land or transpired by vegetation due to altered atmospheric patterns (Pranindita et al., 2021).

Additionally, local mechanisms or secondary effects could increase the likelihood of nonlinear responses, thereby increasing the probability of reaching tipping points. For instance in a more open forest or simply due to warmer and drier conditions at the forest floor, fire occurrences and intensities can easily increase. Moreover, the suppression of forest regeneration can occur due to the invasion of highly competitive light-demanding plant species, forming ecosystems which potentially transpire less moisture back to the atmosphere.

In combination with reduced resilience and resistance due to human interferences, abrupt large-scale damage and dieback of temperate forests is conceivable. Early warning signals in satellite-derived biomass data hint towards such a destabilisation (Forzieri et al., 2022). Yet, large-scale tipping behaviour in temperate forests is not proven. If at all, this will certainly be region-specific and recent forest damage will illuminate such potential feedbacks in the near future.

### Assessment and knowledge gaps

It is uncertain if temperate forests have strong enough self-amplifying feedbacks like the Amazon rainforest and boreal forest to result in tipping, hence there is no evidence for larger-scale tipping and confidence is low. There is, however, a lot of evidence and medium confidence for abrupt changes with changing disturbances regimes.

Human forest management practices may have made temperate forests less resilient and therefore more susceptible to abrupt changes, but improved management can assist resilience and adaptation to climate change. Based on the impacts of current extreme events on temperate forests, it can be inferred that an increase in the intensity and/or frequency of such events could severely threaten existing forests in many areas, even without further climate change (Senf et al., 2020, Lloret and Batllori, 2021). The potential feedback to the water cycle requires further investigation. In particular, modelling studies should fully account for extreme events such as droughts, heatwaves and other important disturbances, their increasing frequencies and intensities as well as their potential impact on simulated vegetation and the resulting land-atmosphere feedbacks (Kolus et al., 2019).

### 1.3.2.4 Savannas and grasslands

Savannas and grasslands are characterised by the ecological dominance of grasses, sometimes with a substantial tree or shrub component (Figure 1.3.7). Savanna ecosystems are biodiverse and home to many people, but are being lost to a range of threats globally, especially because they are extensively targeted for agricultural conversion (Stevens et al., 2022; Strömberg and Staver, 2022). Even intact savannas are under threat, largely due to forest invasion or afforestation and woody encroachment, driven by grazing intensification and active fire suppression, and exacerbated by increasing atmospheric $CO_2$ (Stevens et al., 2017) and changing rainfall regimes (Kulmatiski and Beard, 2013). Active tree planting efforts further increase the threat to savannas from afforestation and woody encroachment.

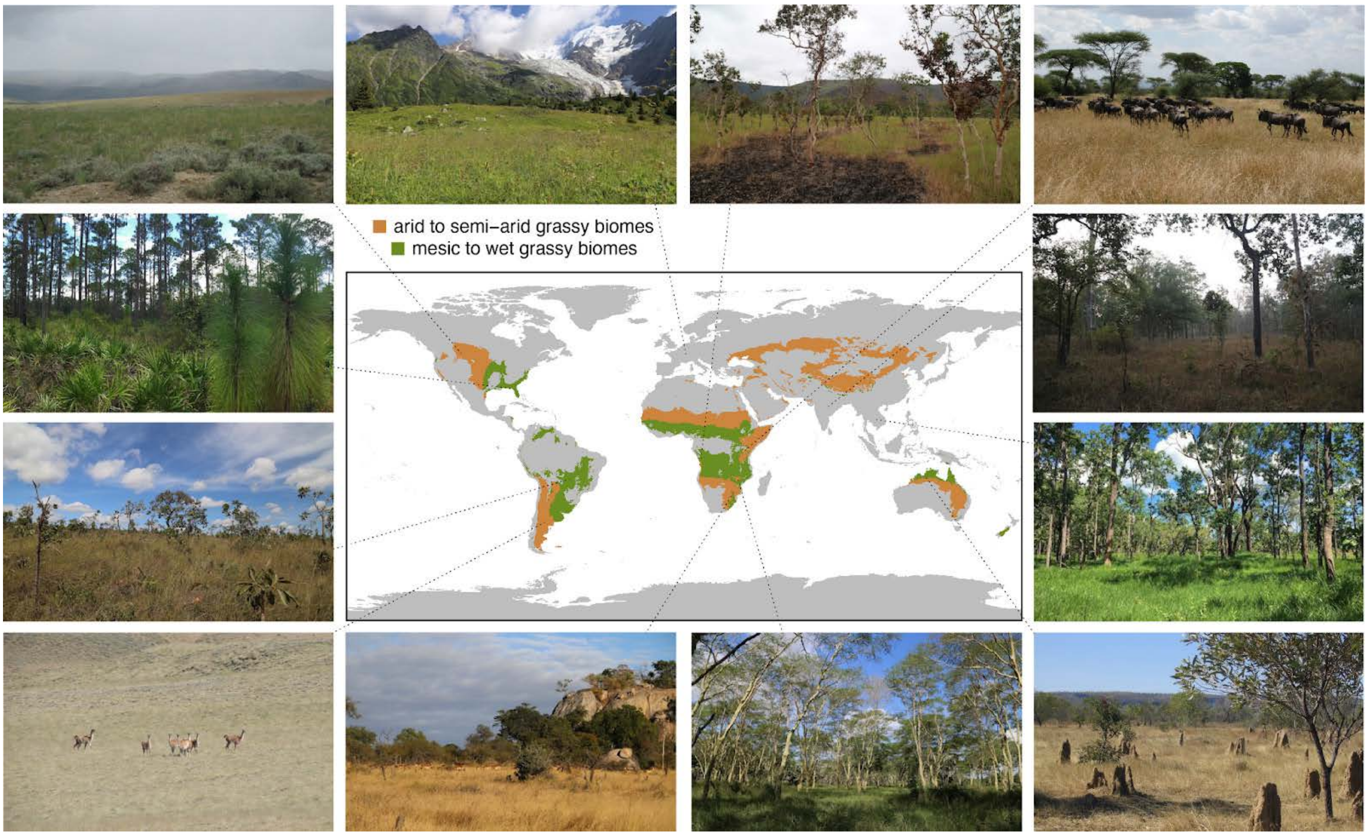



**Figure 1.3.7:** Global distribution of savannas and grasslands, showing semi-arid vs. mesic distributions (centre, from: Strömberg and Staver (2022), replotted from Dinerstein et al., (2017). Pictured, clockwise from top left, are native grassy ecosystems in 1) Montana near Dillon, USA; 2) Alps near Mont Blanc, France; 3) Pool Department, Republic of Congo; 4) Serengeti NP, Tanzania; 5) Pench NP, India; 6) Chhaeb Wildlife Sanctuary, Cambodia; 7) Kidman Springs Ranch, Australia; 8) Gorongosa NP, Mozambique; 9) Kruger NP, South Africa; 10) Santa Cruz Province near Lago Argentino, Argentina; 11) Instituto Brasileiro de Geografia e Estatística Reserve, Brasilia, Brasil; 12) Apalachicola National Forest, Florida, USA. Photo credits: Carla Staver, Caroline Strömberg, Naomi Schwartz.

Although the converse issue receives extensive attention (e.g. Amazon rainforest collapse), the issue of savanna vulnerability to tipping points is recognised (Staver et al., 2011b) but generally neglected in literature and assessments of tipping points in the Earth system (Armstrong McKay, 2022; Wang et al., 2023). Savanna vulnerability to desertification (corresponding to a self-sustaining loss of ecosystem productivity) is sometimes cited in tipping point syntheses, but the generality of this feedback has been questioned. For example, aridification observed in western Africa's Sahel during the 1970s and 80s has since reversed across much of the Sahel in response to a cyclic increase in rainfall (Nicholson et al., 1998; Prince et al., 2007).

#### Evidence for tipping dynamics

Savanna and forest are widely considered to be alternative stable ecosystem states in some climates (Staver et al., 2011a, 2011b; Hirota et al., 2011; Dantas et al., 2015; Aleman et al., 2020). In savannas and grasslands, an open tree canopy permits high grass productivity and thus the accumulation of grass fuel for frequent fires (Hennenberg et al., 2006; Lloyd et al., 2008). Fires in turn limit tree establishment (Higgins et al., 2000; Hoffmann et al., 2009), keeping the canopy open and creating a positive/amplifying feedback that potentially stabilises savannas in regions where forest is also a viable stable ecosystem state (Beckage and Ellingwood, 2008; Staver et al,. 2022a), although some apparent bistability may be the result of spatial climate variability (Good et al., 2015; Higgins et al., 2023).

The maintenance of savannas is thus dependent on fires across large parts of their range. This has meant that widespread fire suppression (active or passive via agricultural fragmentation or grazing intensification) has triggered woody encroachment and, in extreme cases, forest invasion (Stevens et al., 2017). These feedbacks between vegetation and fire frequency and intensity have also been implicated in accelerating the invasion of alien grasses that are more flammable and also tolerate higher fire intensities than native grasses (D'Antonio and Vitousek 1992; D'Angioli et al., 2022) (Figure 1.3.8). Fire-related feedback loops may not be as significant in drier savannas where herbivores or low water availability limit the accumulation of grass and thus fuel (Archibald and Hemson, 2016; Dexter et al., 2018), and further research is needed on the tipping dynamics of arid savannas and their potential alternate states (see Drylands 1.3.2.5).

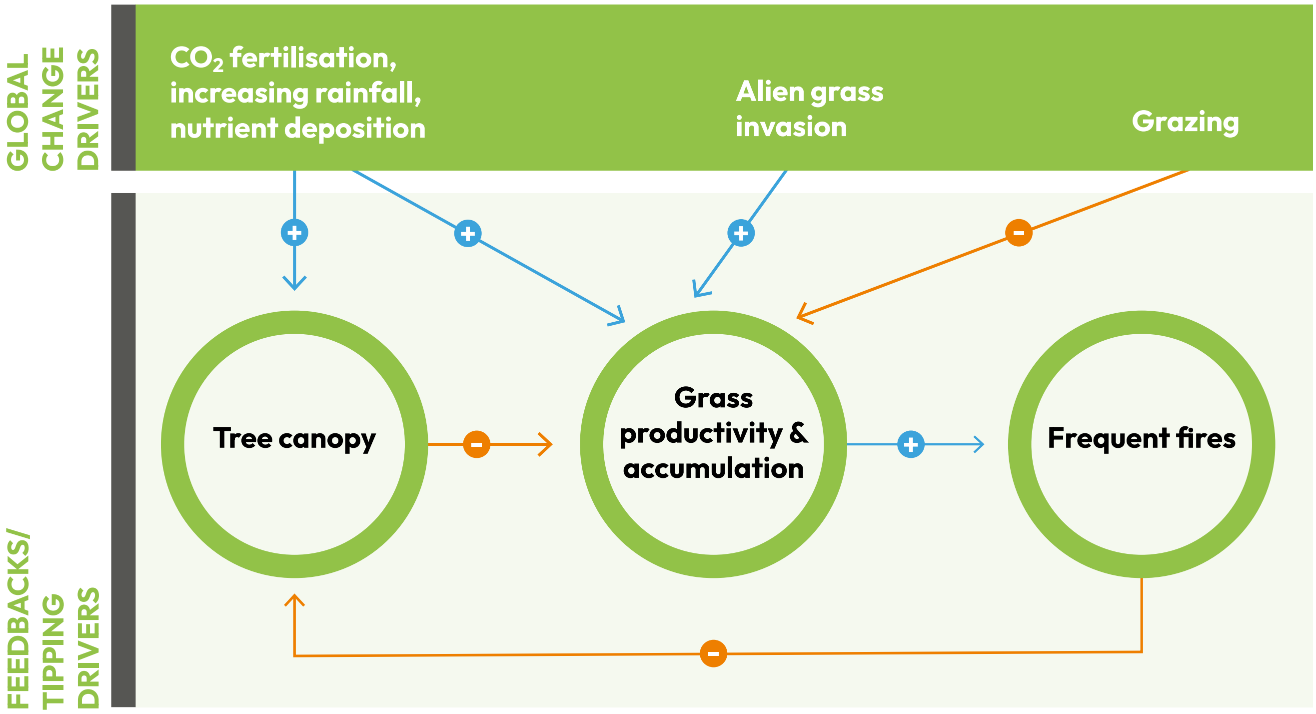



**Figure 1.3.8:** Key feedbacks that could lead to savanna tipping.

Several important thresholds are involved in this tipping point. First of all, fire spread is widely described as a percolation process (Loehle et al., 1996; Favier, 2004) – whereby a burning patch infects neighbouring or nearby flammable patches, thereby propagating fire in flammable landscapes. However, when not enough of the landscape is flammable (in this case, if trees shade grasses to prevent fuel accumulation), fires extinguish, with a clear threshold in fuel cover between 'connected' flammable vs. 'unconnected' non-flammable landscapes (Cardoso et al., 2022). In theory, this threshold can depend on the model used, but in practice, there appears to be a threshold in fuel cover of ~50-60 per cent, below which fire does not successfully spread (Archibald et al., 2009; Cardoso et al., 2022). Thus, fire suppression initiates woody encroachment or forest invasion, which can in turn decrease landscape flammability further, creating a cascade that results in the irreversible loss of open-canopy savannas.

The rate at which this happens – and ultimately the environmental space in which closed vs. open-canopy ecosystems are viable – depends also on environmental thresholds, but these are more widely disputed. A range of studies has defined the minimum required to sustain a closed forest canopy as ranging between 750 and 1,000mm mean annual rainfall (Sankaran et al., 2005; Staver et al., 2011b; Aleman et al., 2020), but more open but still fire-suppressing canopy can also form at much lower rainfall, for example in the Caatinga (Charles-Dominique et al., 2015; Dexter et al., 2018). The high-rainfall limit for savanna persistence is even less defined, as savannas can occur in areas with well over 1,600mm mean annual rainfall – for example, in the Llanos of Venezuela and Colombia (Huber et al., 2006) or the Beteke Plateau in the Republic of Congo (Nieto-Quintano et al., 2018).

Moreover, increasing atmospheric $CO_2$ is changing the relative photosynthetic efficiencies of 'C4' grasses vs. 'C3' trees (with C4 being the more efficient photosynthesis process) (Ehleringer and Björkman, 1977; Bond and Midgley, 2012) and is increasing plant water use efficiency across different plant types (Leakey et al., 2009; Norby and Zak, 2011). This has increased the rate of woody encroachment and forest invasion into savannas, suggesting that vulnerability of savannas to tipping points is accelerating and is not stationary with respect to climate (Higgins and Scheiter, 2012). For this reason, defining exactly how much global change might trigger savanna tipping points is not feasible (and indeed a single global tipping point may not exist).

Several lines of evidence provide support for the irreversibility of savanna-to-forest transitions. First, palaeoecological studies have suggested that reversible increases in rainfall can result in irreversible shifts from savanna to forest, consistent with hysteresis (i.e. where reversing the driver of change does not lead to recovery; see Glossary) (Karp et al., 2023). Second, and more directly, fire experiments have demonstrated that, while fire suppression causes savannas to transition to forest-like systems, introductions of fire into forests have much smaller effects (Gold et al., 2023), likely because closed forest canopies prevent fuels from accumulating to fuel intense savanna fires. This demonstrates that managed fire reintroductions are not sufficient to reverse forest encroachment (Gold et al., 2023).

Extreme fires can help reverse encroachment by forests when trees are fire sensitive (Silvério et al., 2013, Brando et al., 2014, Beckett et al., 2022) but extreme fires do not reverse woody encroachment (Strydom et al., 2023). In the case of savanna invasions by non-native grasses, irreversibility of transitions may be further exacerbated by resulting changes in nutrient cycling (Bustamante et al., 2012; D'Angioli et al., 2022). Together, these diverse lines of evidence suggest that savanna invasions, once initiated, may be rapid and irreversible.

The timescale of woody encroachment varies depending on environmental controls, but can happen in less than a decade, with accelerating vulnerability across savanna ecosystems due to rising $CO_2$ (Stevens et al., 2022) and widespread enthusiasm for climate mitigation via tree planting (Bastin et al., 2019, Fagan et al., 2022).

The climate impacts of woody encroachment and forest invasion are uncertain, however, due to substantial carbon in belowground pools in savannas (Zhou et al., 2022) and large uncertainty in how belowground carbon pools (root biomass and especially soil organic carbon) will respond to increasing woody cover (Veldman et al., 2019; Zhou et al., 2023). Hydrologically, there is evidence that an increasing tree fraction can increase rainfall interception and accelerate ecosystem water use, depleting groundwater recharge and streamflow, with implications for downstream water availability (Jackson et al., 2005, Honda and Durigan, 2016). Feedbacks with albedo (with woody vegetation being 'darker' than grass) have also been discussed, but little studied (Stevens et al., 2022).

#### Assessment and knowledge gaps

We have high confidence that Savannas are undergoing widespread degradation from woody encroachment, forest invasion, afforestation and alien grass invasion, high confidence that this is related to grazing intensification and active fire suppression and medium confidence that this is exacerbated by increasing $CO_2$ levels. These changes are increasingly difficult to reverse with the reapplication of fires (medium confidence), although sensitivity of invading vegetation to climate extremes is variable or unknown (Zeeman et al., 2014; Case et al., 2020).

Compounded by agricultural conversion and tree planting, this is rapidly eroding endemic savanna and grassland biodiversity (high confidence) (Smit and Prins, 2015, Andersen and Steidl, 2019, Wieczorkowski and Lehmann, 2022).

Overall, savannas are likely to feature tipping dynamics at local to landscape scales (medium confidence), although large-scale synchrony may be observed if global change drivers trigger tipping points. However, Earth system feedbacks associated with savanna degradation are highly uncertain (low confidence), with particular knowledge gaps about carbon and hydrological cycle outcomes. Potential tipping points in savannas and grasslands associated with herbivory represent another major knowledge gap.

### 1.3.2.5 Drylands

Drylands are hyper-arid, arid, semi-arid and dry-sub-humid climate zones (Figure 1.3.9) where rainfall is less than 65 per cent of the 'potential evapotranspiration' (i.e. the amount of evaporation that would occur if enough water were available) (Middleton and Thomas, 1992). They occupy over 46 per cent of the Earth's surface and host 38 per cent of the world's human population (more than 2 billion people) (Cherlet et al., 2018). Vegetation types include deserts, grasslands, shrublands, woodlands, savannas, Mediterranean forests and tropical dry forests (see 1.3.2.1 and 1.3.2.4 for tropical dry forests and savannas). Due to their extent and the chronic water deficit, these areas are of particular concern in the face of global changes, and so we assess them separately here, despite some overlap with the tropical forest and savanna and grassland biomes above.

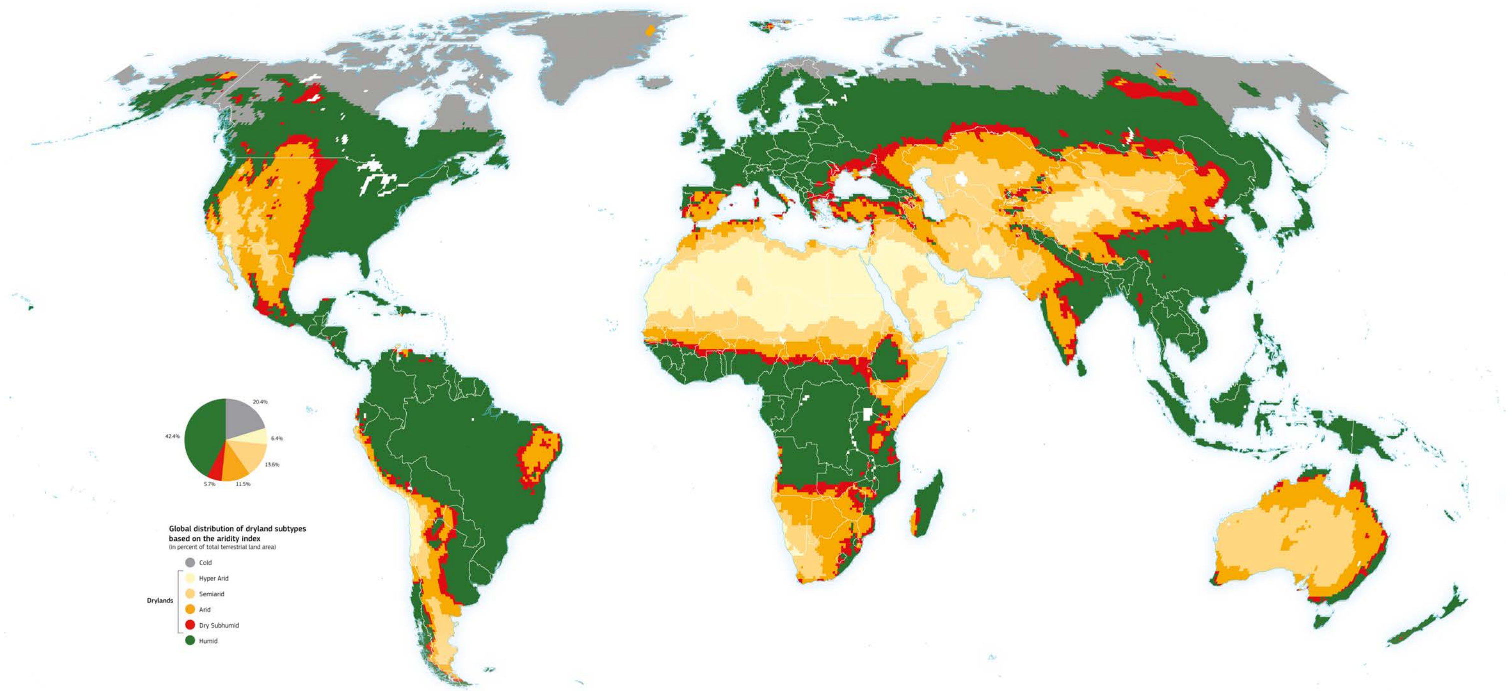



**Figure 1.3.9:** Global distribution of dryland subtypes based on the aridity index. Source: WAD3-JRC (Cherlet et al., 2018).

Recent estimates suggest that one-fifth of drylands are degraded as a result of climatic variations and human activities (Burrel et al., 2020; about 9 per cent in IPCC SRCCL, 2019). Major pressures on drylands (Cherlet et al., 2018) include:

- Climate change – for example, changes in precipitation, temperature, seasonal and interannual variability and frequency of extreme events. Projections indicate that some drylands might become more humid, whereas others may become drier (Huang et al., 2016; Pravalie et al., 2019). These expectations are uncertain though (Lian et al., 2021).
- Land use intensification – for example, grazing (the main use of drylands, at 62 per cent) (Cherlet et al., 2018), water extraction, deforestation, agriculture and urbanisation.
- Perturbations – for example, fires, insect outbreaks and biotic invasions.

The dynamics of drylands depend strongly on the interaction between these pressures, such as climate change and local perturbations (Rilig et al., 2023).

#### Evidence for tipping dynamics

Different lines of evidence point toward the existence of tipping dynamics in drylands, including past and current ecosystem transitions, bistability of dryland states at the global scale, thresholds along environmental gradients, and feedback mechanisms maintaining persistent dryland states.

Abrupt transitions have historically occurred in several dryland systems. Palaeo evidence reveals abrupt shifts into and out of African Humid Periods (Pausata et al., 2020), including a notable greening of the Sahara during the early to mid-Holocene, followed by its abrupt desertification around 5,500 years ago (Shanahan et al., 2015; Claussen et al., 2017; Hopcroft and Valdes, 2021, Claussen et al., 1999). Positive/amplifying feedback mechanisms between vegetation and the monsoon in North Africa are thought to be important (Charney et al., 1975). Climate change projections suggest 'Sahel Greening' might partially occur again in the future (Erfanian et al., 2016; IPCC SR1.5, 2018; Dosio et al., 2021). In dune systems, stratigraphic records covering 12,000 years have found coexistence of a vegetated, stabilised state and a bare active state in dune systems in northern China, with occasional sharp shifts in time between those contrasting states and hysteresis (Xu et al., 2020).

Shrub encroachment may also reflect tipping dynamics. Long-term data from Jornada Experimental Range (northern Chihuahuan Desert, New Mexico, USA) showed abrupt transitions from grasslands to shrublands triggered by a combination of climatic and human (i.e. overgrazing) factors during the last 150 years (Beltelmeyer et al., 2011; D'Odorico et al., 2012). Transitions from Mediterranean forests to shrublands have been reported under a combination of dry conditions, wildfires (Baudena et al., 2020; Acacio et al., 2009, Mayor et al., 2016) and herbivory (van der Wouw et al., 2011).

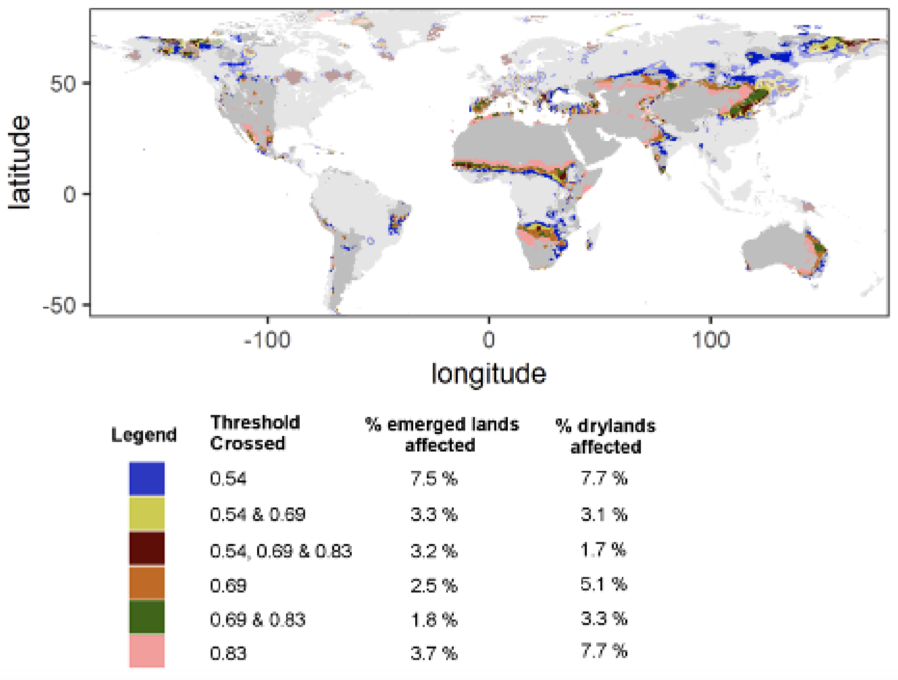



**Figure 1.3.10:** Map of drylands vulnerability to predicted changes in aridity for 2100 based on the IPCC RCP8.5 scenario (i.e. under the assumption of sustained increase in $CO_2$ emissions). Abrupt decays in plant productivity, soil fertility and plant cover were identified beyond aridity threshold values of respectively 0.54, 0.7, and 0.8 (Berdugo et al., 2020). The map displays areas that are expected to cross one (or several) of those thresholds in aridity level. Light-grey areas are areas that are not drylands today. Figure from (Berdugo et al., 2020). Satellite observations indicate that 5 per cent of drylands have experienced an abrupt loss of vegetation cover over the last 20 years, while 18 per cent underwent an abrupt increase in vegetation (Berdugo et al., 2022).

Evidence suggests that, in drylands, sequential abrupt shifts in plant productivity, soil fertility and plant cover occur at increasing aridity thresholds, respectively corresponding to aridity values of 0.54, 0.7, 0.8 (Berdugo et al., 2020) (Figure 1.3.10). A higher dependence of vegetation on water has been reported at aridity values of around 0.6 (Nemani et al., 2003), producing a decline in productivity with increasing aridity (Berdugo et al., 2020) and an increase in tree mortality events with hotter droughts (Hammond et al., 2022).

At aridity levels around 0.7, abrupt declines in vegetation are related with losses of soil fertility (Delgado-Baquerizo et al., 2013; Berdugo et al., 2020), changes in vegetation spatial structure, (Kéfi et al., 2007, 2011; Berdugo et al., 2017; Berdugo et al., 2019) which influences soil hydrological connectivity and resource loss at the landscape scale (Mayor et al., 2013; Rodriguez et al., 2018, Mayor et al., 2019), increases in the dominance of shrublands (Berdugo et al., 2020), and rapid shifts in the composition of soil microbial communities and soil functionality (Maestre, 2015; Lu, 2019; Delgado-Baquerizo, 2020; Zhang et al., 2023).

At aridity thresholds of 0.8, abrupt decays in plant productivity and vegetation cover occur (Berdugo et al., 2020) and can lead to a nonlinear increase in soil erosion (Mora and Lázaro, 2013; Elwell and Stocking, 1976; Francis and Thornes, 1990; Mayor et al., 2013).

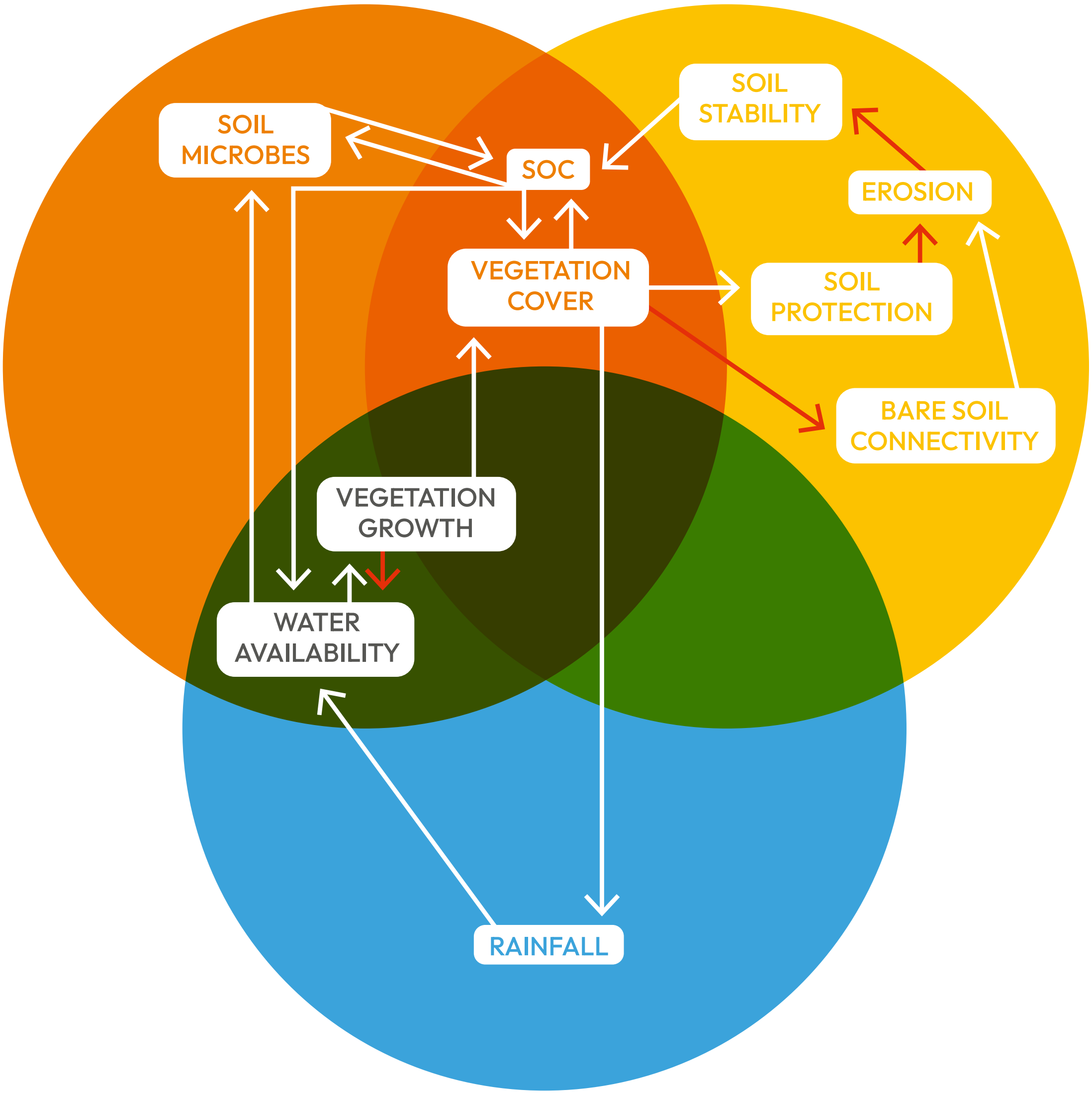



**Figure 1.3.11:** Schematic showing key feedbacks that could lead to dryland tipping. Coloured disks represent some of the main feedbacks described in the text (vegetation-rainfall in blue, biogeochemical feedback in red and ecohydrological feedback in blue). White arrows represent positive effects (an increase in the variable at the source of the arrow leads to an increase of the variable at the end of the arrow) and red negative effects. SOC stands for Soil Organic Carbon. See (Mayor et al., 2019) for a more detailed version of the ecohydrological feedback.

A number of feedback mechanisms are known to occur in drylands, operating across ecosystem elements and at different spatio-temporal scales. Theoretically, such feedbacks can lead to bistability and abrupt transitions between stable states in drylands (Holling, 1973; Noy-meir, 1975; May, 1977; Scheffer et al., 2001, Walker et al., 2004), although such alternative states do not necessarily exist in all regions (Ma et al., 2023). Several key feedbacks can be identified (Figure 1.3.11):

- **Soil microbial communities (biogeochemical feedback; small scale):** Microbial biomass and diversity in drylands are intricately linked to variations in water availability and organic matter (which change along the global aridity gradient; Zhang et al,.2023). Soil microbes, such as bacteria, fungal decomposers and mycorrhizal fungi, are fundamental for the breakdown of complex litter and organic matter. By decomposing organic matter, microbes are critical in the build-up of soil carbon stocks, which is essential in the maintenance of moisture in dry soils. Soil moisture, in turn, is needed for organic matter decomposition.
- **Plant-plant interactions:** In drylands, plants are known to facilitate the recruitment and growth of other plants, leading to the formation of vegetation patterns. The positive interactions between plants, i.e. facilitative effects, involve effects on microclimate, soil conditions and herbivores impacts:
    - **Plant-soil feedback (small to medium scale):** Plants enhance local soil conditions through several means, such as nutrient and water retention ('islands of fertility'), microclimate influence and erosion prevention (Aguiar and Sala 1999; Schlesinger et al., 1990; Rietkerk et al., 2000; D'Odorico et al., 2007). These processes boost vegetation growth and contribute to the formation of spatial patterns.
    - **Ecohydrological feedbacks (medium scale):** Plants aggregate and form spatial patterns of vegetation patches interspersed in a matrix of bare soil (Aguiar and Sala, 1999). The spatial connectivity of the bare soil (runoff-source areas) affects the redistribution of water, nutrients and sediments at the patch and landscape scale, which in turn shapes vegetation cover and spatial pattern (Mayor et al., 2013, 2019). These local (patch) and global (landscape) connectivity-mediated feedbacks affect the productivity and resilience of the ecosystem (Mayor et al., 2019).
    - **Vegetation-herbivore feedback (medium scale):** Herbivores graze/browse on palatable plants, which stimulates regrowth (McNaughton 1983); they then keep eating at the same places because the resprouts are soft and more easily digestible. This in turn allows the recruitment of unpalatable plants in areas without herbivores. An excess of grazing on palatable plants can prevent regrowth and lead to vegetation transitions from diverse, palatable to unpalatable dominated plant communities (Cingolani et al., 2005).
- **Vegetation-fire feedback (medium to large scale):** Fire can facilitate a transition from forest to shrublands. Shrublands recover faster and burn easier, generating a positive/amplifying feedback (e.g. dry Mediterranean regions in Portugal (Acacio et al., 2009), Spain (Baudena et al., 2020)). Replacement of native Mediterranean forests by pine forest plantations or invasion by exotic non-woody plants can contribute to this feedback (e.g. central Chile) (Pauchard et al., 2008; Gomez-Gonzalez et al., 2018) (see also: Tropical [1.3.2.1] and Boreal [1.3.2.2] forests).
- **Vegetation-rainfall positive/amplifying feedbacks (large scale):** Vegetation is largely controlled by local climate, but modelling studies suggest that it can also influence regional precipitation by modifying the atmospheric energy and water budget (Charney, 1975; Dekker et al., 2007). This large-scale albedo-precipitation and evapo-transpiration–precipitation feedback could have significant implications for ecosystem resilience. (see also: Tropical [1.3.2.1] and Boreal [1.3.2.2] forests).

Global dryland assessments suggest two different ecosystem states can exist at intermediate aridity levels ('bistability'). Drylands with aridity levels between between 0.75 and 0.8 (i.e. in the transition zone between semi-arid and arid drylands) may be in one of two different states, with higher and lower vegetation cover, with large contrasts in soil fertility, nutrient capture and nutrient cycling (Berdugo et al., 2017). Observing different ecosystem states across an area with similar conditions does not in itself prove those ecosystems are bistable. However, the global tendency for these two states to emerge, combined with our understanding of feedbacks in these ecosystems and observations of threshold responses, suggests that these could represent alternative stable states in these ecosystems.

Hysteresis, where reversing the driver of change does not lead to recovery (see Glossary), can also be evidence for alternative stable states and tipping dynamics in dryland ecosystems. In Spain (NE, Ebro Valley), past overgrazing was found to interact with droughts to explain the lack of secondary succession or even decreasing normalised difference vegetation index (NDVI, a remote sensing index for vegetation cover) trends (Vicente-Serrano, 2012). Some long-term field studies provide evidence for hysteresis in drylands. For example, in the northern Chihuahuan Desert (US), grasslands shifted into shrublands dominated by Creosote Bush (*Larrea tridentata*) during a prolonged drought combined with overgrazing, but the recovery of grass productivity did not occur in subsequent wet years (Bestelmeyer et al., 2011). Results also suggest the possibility of crossing critical thresholds for irreversible degradation (i.e. 20 per cent plant cover in Gao et al., 2011).

Long legacy effects are consistent with the existence of hysteresis in drylands. For example, palaeoclimatic legacies, e.g., from the Last Glacial Maximum, influence soil biodiversity (Delgado-Baquerizo et al., 2017), function (Ye et al., 2019) and forest distribution (Guirado et al., 2022). For example, drylands with a wetter past now have greater levels of function and forest coverage than what would be expected for current climatic conditions (Ye et al., 2019).

The reversibility of ecological transitions in drylands is challenging because plant growth rate is strongly limited by water scarcity and local disturbances. However, it is noteworthy that fast vegetation recovery during rainy periods has been observed at local and regional scales (Holmgren et al,.2006a, 2013). Studies have also found recovery of drylands to strong grazing pressure even at low cover levels in case of favourable weather conditions (Bestelmeyer et al., 2013). Coupling passive and active restoration of drylands to favourable climate swings can open windows of opportunity for dryland recovery (Holmgren and Scheffer 2001, Holmgren et al., 2006b, Sitters et al., 2012).

Timescale for transitions are about weeks to months for tree heat and grazing, months to decades for shrub encroachment Bestelmeyer et al., 2011; Tabares et al., 2019 and abrupt vegetation loss due to droughts Berdugo et al., 2022), and a few decades for the desertification of the Sahara (Shanahan et al., 2015; Claussen et al., 2017; Hopcroft and Valdes 2021; Claussen et al., 1999).

#### Assessment and knowledge gaps

Dynamical evidence of tipping points in drylands is challenging to find due to the slow dynamics of these ecosystems. Altogether, the knowledge of past transitions shows that relatively rapid changes have occurred in drylands, in particular in terms of vegetation cover, species composition and soil communities, leading to important changes for biodiversity and ecosystem functioning. Further, evidence from positive/amplifying feedbacks between different components of ecosystems, thresholds values in stressors (aridity, fire frequency, grazing) and hysteresis (lack of recovery) suggests the likelihood of future recurrence. We assess that dryland ecosystems can feature local to landscape-scale tipping points towards land degradation (medium confidence) with climate and land use change.

Core ecological questions remain, mainly on the mechanisms by which abruptness appears in drylands. We need long-term dynamical records. This is in particular true for soils; which is very relevant given that several thresholds involve soil transformations (particularly soil fertility losses). This lack of evidence in soils is even more difficult to address given that soils are themselves a slow component in an already slow ecosystem type and, unlike vegetation, can not be assessed with remote sensing. Quantification of the thresholds for herbivory pressure, fire frequency, and logging along aridity gradients is also necessary. Crucially, we need to improve our description and incorporation of social-ecological feedbacks in drylands (Reynolds et al., 2007). Indeed, dryland ecosystem transitions are associated with important social pressures and livelihood dependency, especially in developing countries, making social-ecological feedbacks critical to understand (see Section 2; Walker et al., 2004; Reynolds et al., 2007).

Several biotic mechanisms (e.g. local negative plant-patch feedbacks – Mayor et al., 2019) that confer resilience to dryland ecosystems are still not sufficiently explored, such as plant plasticity or adaptability to drought. Some mechanisms might be able to counteract abrupt changes; for example $CO_2$ fertilisation may confer higher water use efficiency to plants, thus opposing stress caused by lack of water (Zhu et al., 2016) and possibly counteracting aridification (Peñuelas et al., 2017; Zhang et al., 2022). Also, we can refine our understanding of windows of opportunity for restoration in drylands (e.g. taking advantage of temporarily favourable climatic conditions; Holmgren and Scheffer 2001; Holmgren et al., 2006b; Sitters et al., 2012; Walker and Salt 2012).

### 1.3.2.6 Freshwater ecosystems

*The scientific content of this chapter is closely based on the following scientific manuscript: Hessen et al., (in review) Lake ecosystem tipping points and climate feedbacks, Earth System Dynamics Discussion.*

Freshwater bodies such as lakes are common across most biomes, forming unique and sometimes isolated ecosystems (Figure 1.3.12). In natural sciences, the hysteretic behaviour of lakes (Scheffer et al., 2007) has informed the concept of tipping points at the ecosystem level, leading to the development of the alternative stable states theory in shallow lakes (Scheffer et al., 1993; Carpenter et al., 1999; Carpenter 2005). They represent archetypal case studies for how tipping points relate to theories of ecological stability and resilience that can underpin preventative management approaches (Andersen et al., 2009; Spears et al., 2017). Despite this, significant uncertainty remains on the geographical extent of tipping points in lakes and the wider relevance for the Earth's climate system.

Lakes are also good examples of social-ecological systems, with their ecological dynamics closely intertwined with the socio-economic dynamics of surrounding populations who often depend on them for key ecosystem services and adaptively respond to changes in lake condition (Martin et al., 2020). Given the global vulnerability of freshwaters and the pervasive nature of major pressures acting upon them (e.g. nutrient pollution and climate change), tipping points in these systems could have significant societal impacts, including on human and environmental health, food production and climate regulation. The capacity to detect discontinuous ecosystem responses to pressure changes in natural systems has been challenged (e.g. Hillebrand et al., 2020). Nevertheless, there are several studies that have reported real tipping points, i.e. shifts from one stable state to another in small shallow lakes (the most common lake type globally, Messager et al., 2016).

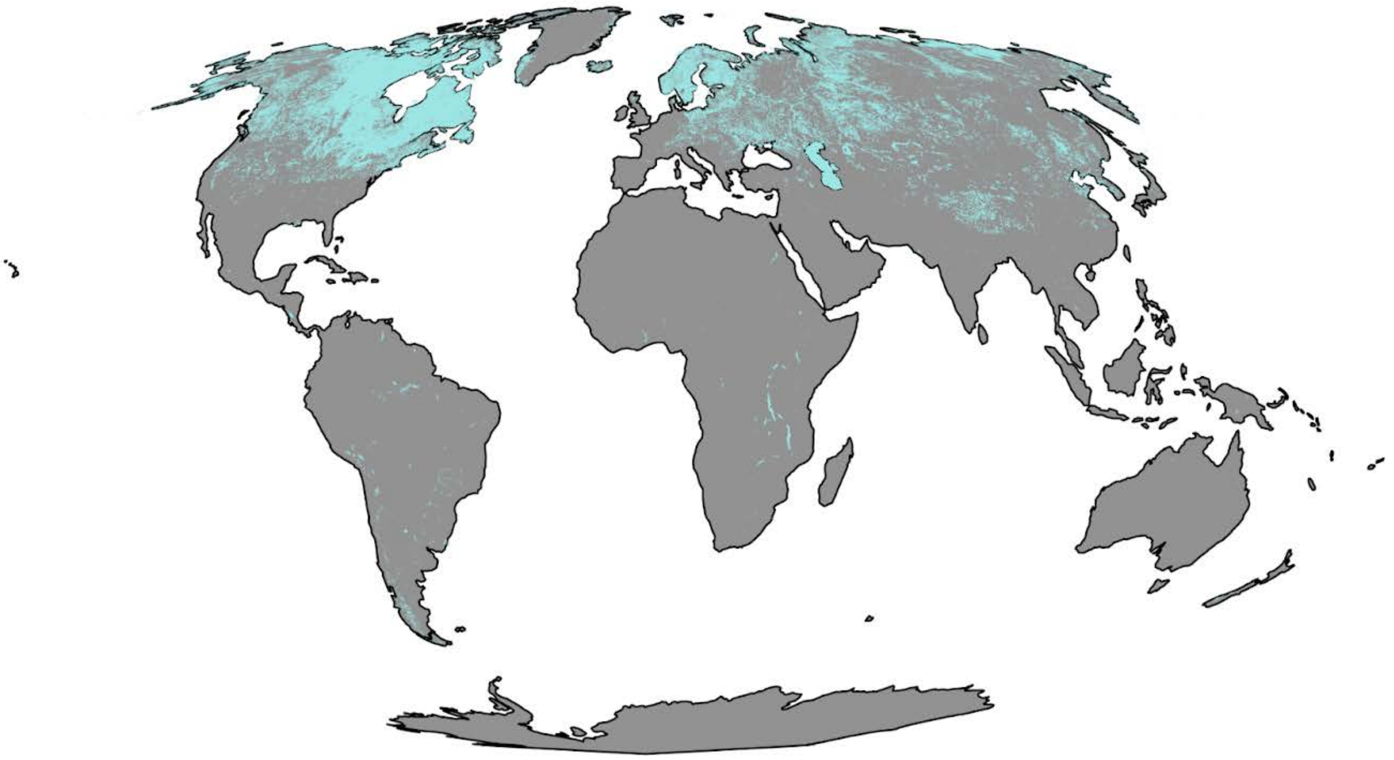


**Figure 1.3.12:** Top: map showing global distribution of lakes (light blue) (source: (Keith et al., 2022). Middle left: eutrophic urban lake receiving high organic matter loading leading to elevated $CH_4$ emissions, Bellandur Lake, Bengaluru City, India (photo: Laurence Carvalho). Middle right: boreal, brown water lake with deepwater anoxia and high emissions of $CO_2$ and $CH_4$ (photo: D.O. Hessen). Bottom left: Arctic pond at Svalbard, recently formed by permafrost thaw below Zeppelin mountain (photo: D.O. Hessen). Bottom right: thermokarst lakes in Yukon Flats, Alaska (photo: Sebastian Westermann).

Empirical analyses, process modelling and experimental studies are advanced for shallow lakes providing a good understanding of ecosystem behaviours around tipping points, typically starting with positive/amplifying feedback loops, then entering a runaway phase before finally the tipping point brings the system into a different stable state (Nes et al., 2016). For example, the well documented increase of phosphorus (P) loading across European lakes in the last century (e.g. from agricultural and waste water pollution) has uncovered critical loading thresholds beyond which lakes can shift rapidly from a clear water, macrophyte rich state to a turbid, phytoplankton rich state (Scheffer et al., 2001; Jeppesen et al., 2005; Tátrai et al., 2009), and vice versa when nutrient loading decreases.

Adding to such well-described and mechanistically well-understood changes, there is a wide range of local or single lake shifts that may be categorised as tipping points. The question remains as to whether tipping points are merely isolated phenomena in single lakes, or specific types of lakes, or whether they manifest, or will in the future, across geographically distinct populations of lakes experiencing similar environmental change, with the potential for regional or global extent (Figures 1.3.12 and 1.3.13).

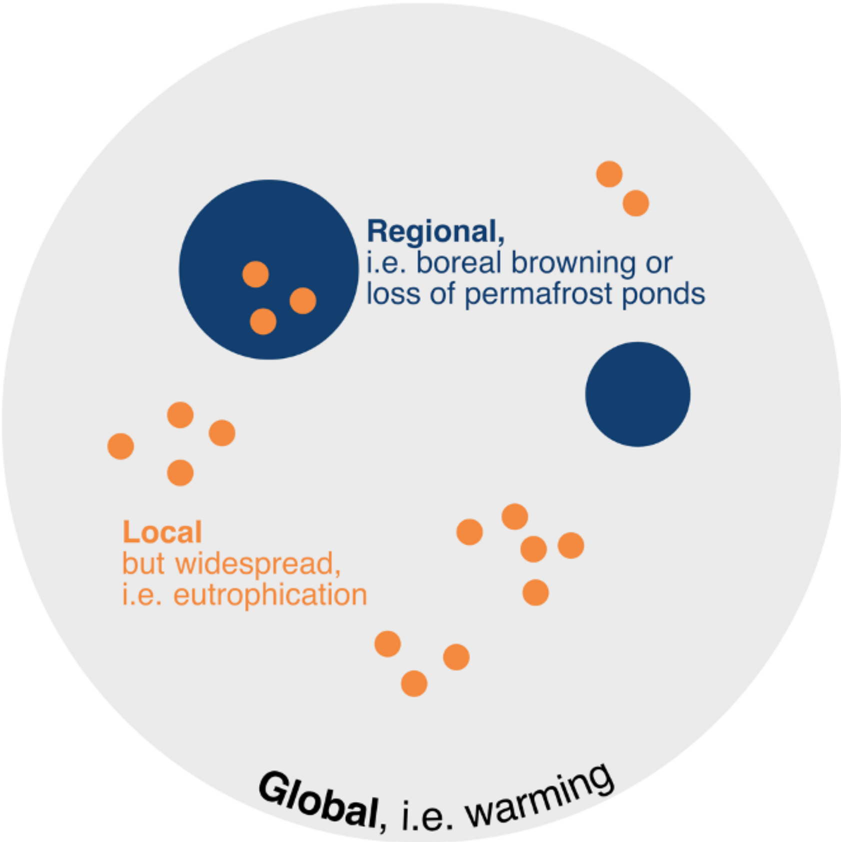



**Fig. 1.3.13:** Impacts at levels that may qualify for tipping points at relevant scales. Regional or biome-wise, effects could be loss of ponds and lakes due to permafrost thaw and/or increased loadings of DOM in the boreal biome or salinisation. Also, local but widespread changes such as anthropogenic eutrophication of lakes in populated areas would have large-scale impacts. Lakes worldwide show a warming trend, hence a global impact. Source: (Hessen et al., 2023, in review).

It is well established that lakes are sensitive to the effects of climate change, including warming and changes in precipitation and storminess (Meerhoff et al., 2022). Emerging evidence suggests that they may also play an important role in climate regulation, through both the emission of greenhouse gases (predominantly methane – Downing et al., 2021) and carbon burial (Anderson et al., 2020). It is therefore relevant to consider the extent to which potential tipping points may drive, or be driven by, climate change, leading to higher-level feedbacks to the Earth's climate system. In this context we will constrain the discussion to potential tipping points that are more generic, at least with some regional or biome-wise impact, and that could feedback to the climate, while not necessarily being driven or triggered by climate change per se.

Here, we adhere to tipping points as defined in this report (and matching Nes et al., 2016). Based on this we discuss candidate tipping points in freshwaters (Table 1.3.2), focusing on lakes and ponds, with the potential for global or at least regional or biome-scale relevance.

### Evidence for tipping dynamics

#### Eutrophication-driven anoxia and internal P-loading

The mobilisation of P from sediments, a process known as internal loading (Sondergaard et al., 2001), is well described and plays a key role in hysteresis in preventing lakes recovering from human-driven eutrophication (Boström et al., 1982; Jeppesen et al., 1991; Spears and Steinman 2020).

The process may be enhanced by lake warming, and there are feedbacks to climate since water anoxia and internal P-loading (which features the actual tipping point) could offset $CO_2$-fixation by increased release of GHGs. Consequent changes in biota also strengthen hysteresis (Brabrand et al., 1990), not least when cyanobacterial blooms develop. The phenomenon is local but widespread, and likely to increase as a result of global warming (Meerhoff et al., 2022). Increases in precipitation, and high-intensity rainfall events, are also expected to significantly increase runoff of P from agricultural catchments to surface freshwaters (Ockenden et al., 2017), further promoting eutrophication and its manifestations. Warming increases stratification and thermal stability promoting anoxia (Maberly et al., 2020; Woolway et al., 2020), internal fertilisation and increased GHG emissions. In addition to anoxia, there are other feedback mechanisms for lake eutrophication tipping points, such as the macrophyte-nutrient-algae-turbidity and macrophyte-zooplankton/fish-algae-turbidity loops (Wang et al., 2022). Shifts in trophic cascades, i.e. a top-down control of zooplankton and reduced grazing on phytoplankton, could also help drive eutrophication (Carpenter et al., 1985; Carpenter and Kitchell 1988). However, feedback to the climate is primarily related to anoxia.

### Increased loading of DOM and anoxia

Increased export of terrestrially derived dissolved organic matter (DOM) to lakes and rivers in boreal regions ("browning") is a widespread phenomenon partly linked to reduced acidification, but also driven by land use changes (notably afforestation) and climate change ($CO_2$-fertilisation of forests, warming and hydrology) (de Wit et al., 2016; Creed et al., 2018; Monteith et al., 2023). Wide-scale regime shifts in boreal lakes caused by increased loadings of DOM can promote a prolonged and more intensified stratification period (implications summarised above, described for DOM by Spears et al., 2017), amplified by warming. Increased terrestrial DOM loadings intensify net heterotrophy in the systems (i.e. through increased light attenuation and increased access to organic carbon) (Karlsson et al., 2009; Thrane et al., 2014; Horppila et al., 2023). While at present the thresholds around these effects have not been well constrained, the impacts may be significant at the global scale for GHG emissions (Tranvik et al., 2009) and regionally for coastal productivity (Opdal et al., 2019)

Both eutrophication and browning are to some extent driven by climate change, and warming of lakes will promote the effects by increasing thermal stratification, promoting anoxia which again promotes internal loadings of phosphorus, leading in some cases to self-sustaining change (i.e. tipping). Increased release of GHGs will serve as another feedback to the climate (Fig. 1.3.14).

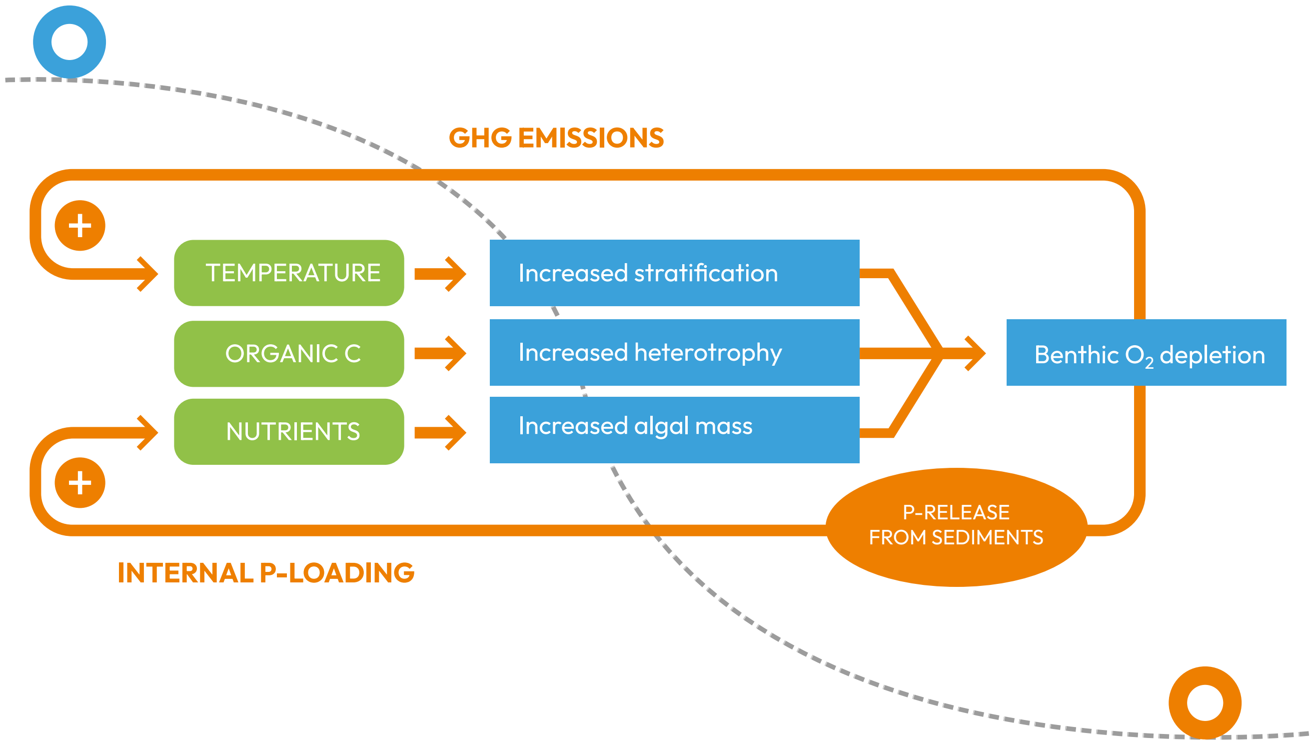



**Figure 1.3.14:** The interactive role of eutrophication, DOM-export (browning) and warming on lakes. Separately or combined they promote benthic $O_2$-depletions which cause an internal feedback by P-loading from sediments and a climate feedback via release of greenhouse gases. The potential shift between states (blue to red circle) is indicated. Adapted from: (Hessen et al., 2023, in review).

### Disappearance/appearance of waterbodies

A global reduction in lake water storage (Yao et al., 2023) and climate-related creation or, more frequently, disappearance, of water bodies is a large-scale concern (Woolway et al., 2022). For example, current and future permafrost thaw and glacier melting can both create new and drain old waterbodies, providing a strong link to the fate of the cryosphere (Smith et al., 2005; Olefeldt et al., 2021). Such small but numerous waterbodies over vast areas in the high Arctic may also serve as major conduits of greenhouse gases and historical soil carbon stocks to the atmosphere (Laurion et al., 2010) and play an important role in mediating nutrient delivery to the polar oceans (Emmerton et al., 2008), potentially affecting global productivity (Terhaar et al., 2021).

Despite the scale considered here, the extent of open water globally is relatively easy to quantify using remote sensing, and loss of waterbodies can be predicted from water balance and thresholds for permafrost thaw with high confidence. However, while representing a binary shift between two states, driven by climate, this should not be classified as tipping events as in most cases no self-sustaining feedback is involved. Lake appearance or disappearance can be driven by cryosphere tipping points though – for example, thermokarst lake formation or abrupt drainage due to permafrost thaw (Turetsky et al., 2020; Teufel and Sushama, 2019) (see Chapter 1.2) – and in such cases the lake forms part of a coupled thermokarst system capable of tipping.

### Switch from N to P-limitation

Regions receiving increased nitrogen (N) deposition may shift from prevailing P- to N-limitation (Elser et al., 2009). Conversely, increased N-loss by denitrification, eventually associated with increased internal P-loading, may shift systems from P to N-limitation (Weyhenmeyer et al., 2007). Changes in N- versus P-limitation of productivity are associated with changes in community structure, both for the phytoplankton and macrophyte communities, which could involve ecological tipping points. However, while the switch between N and P-limitation represents a binary switch with ecological consequences, it is not itself classified as a tipping point according to our criteria, as self-sustaining feedbacks have not been identified. There is currently weak evidence for this shift's impact on climate feedbacks.

### Salinisation

Salinisation is a prevalent threat to freshwater rivers, lakes and wetlands and is caused by a range of anthropogenic actions including water extraction, pollution and climate change (Herbert et al., 2015). It has severe consequences for aquatic communities (Short et al., 2016, Cunillera-Montcusí et al., 2022) with salinity thresholds likely strongly impacted by other stressors – including eutrophication (Kaijser et al., 2019). Salinisation has a strong societal impact, particularly related to domestic and agricultural water supply in arid and semi-arid regions (Williams et al., 1999). Salinisation tends to decrease $CH_4$ emissions (Herbert et al., 2015) and, in that sense, is a negative/damping feedback with respect to climate change. Salinisation may induce ecological regime shifts, for example leading to microbial mat dominance (Sim et al., 2006), and results in some hysteresis, with salinised sediments remaining salty also after the system is flushed with fresh water (Van Dijk et al., 2019), but is not in itself driven by self-sustaining feedbacks.

### Spread of invasive species

Freshwaters are especially vulnerable to species loss and population declines as well as species invasions due to their isolation. Substantial ecosystem changes by reinforcing interactions between invasive species and alternative stable states (i.e. macrophyte – aquatic plant – ***versus*** phytoplankton dominance, as described above) may occur (Reynolds and Aldridge 2021). The spread of several invasive species can be facilitated by climate change (Rahel and Olden, 2008) and may have some self-sustaining properties. Such changes could thus drive a regime shift for a given system, but in most cases are hypothetically reversible if the original driver (the invasive species) were removed. Species invasion is hard to predict and difficult to quantify, despite the risk of species ingress as ranges expand with climate change.

### Assessment and knowledge gaps

**Table 1.3.2: Candidate tipping events from the literature with potential to occur at local to regional scales, their association with climate change, and whether tipping points and hysteresis have been associated with them. Brackets indicate higher uncertainty. Bold entries represent categories that qualify as tipping points in this context, while the others are either simply binary shifts between states, threshold effects, or similar.**

| Type of event | Local | Regional | Climate driver | Climate feedback | Tipping event | Hysteresis |
|---|---|---|---|---|---|---|
| **Eutrophication-driven anoxia and internal P-loading** | **x** | | **x** | **x** | **x** | **x** |
| **Increased loadings of DOM** | | **x** | **x** | **x** | **x** | **(x)** |
| Disappearance/ appearance of waterbodies | | x | x | x | (x) (linked to cryosphere tipping) | (x) |
| Switch between N and P limitation | | x | x | (x) | | |
| Salinisation | | x | x | x | | (x) |
| Spread of invasive species | x | (x) | (x) | | | (x) |

Abrupt changes driven by warming, eutrophication or increased loadings of organic matter, leading to changes in the production to respiration ratio (i.e. systems shifting from net autotrophic to net heterotrophic), and/or onset of bottom-water anoxia have clear tipping dynamics (high confidence) and strong feedback to the climate via GHG emissions (Meerhoff et al., 2022) (Table 1.3.2). Whether the widespread effect of increased loading of organic matter (browning) in boreal lakes can drive tipping points is more of a knowledge gap, yet the feedback of lake browning to climate through increased GHG emissions is evident.

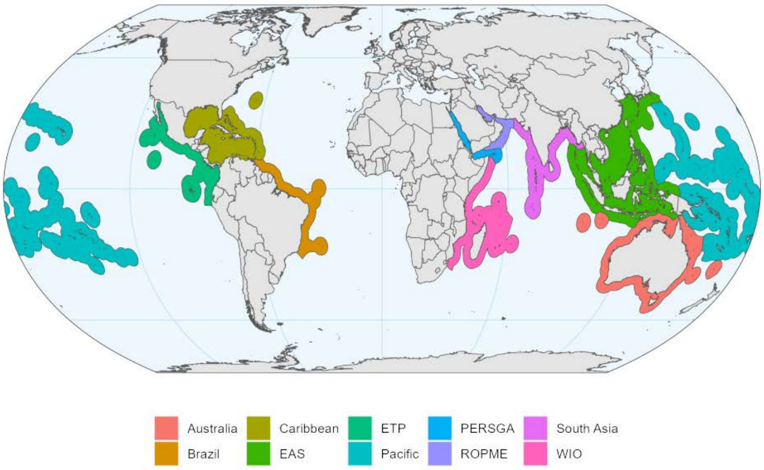
Loss of waterbodies residing on permafrost or suffering negative water balance and eventually complete disappearance represents a binary shift, which has major ecological consequences (Woolway et al., 2022) but is not considered a tipping event ***sensu stricto***. The same holds for other types of binary shifts, threshold effects or local changes. The role of warming as a catalyst on the changes driven by eutrophication and browning is a critical knowledge gap. Quantification of GHG release from lakes represents major feedbacks to climate, and to quantify the impact of eutrophication, browning and warming in this context should have high priority.

### 1.3.2.7 Coastal ecosystems

In this section we consider ecosystems bordering the land and ocean, covering the 'littoral' intertidal and subtidal zones. These zones include some of the most biodiverse and human-depended ecosystems on Earth, despite occupying globally tiny areas: warm-water coral reefs, mangrove forests and seagrass meadows. However, all face increasing pressures from increasingly frequent climate change-induced extremes compounded by habitat destruction, pollution and sea level rise.

**Warm-water coral reefs**

Warm-water coral reefs span the Earth's tropical and subtropical ocean, and are estimated to support over half a billion people for their livelihoods and over a quarter of marine species for part of their lifecycle (Wilkinson et al., 2004; Plaisance et al., 2011) (Figure 1.3.15). They can cross a threshold of ecosystem collapse when they cease to have sufficient cover (typically ~10 per cent) and diversity of hard corals to support the wide diversity of species taxa and ecological interactions typical of a coral reef (Bland et al., 2018; Darling et al,.2019; Sheppard et al,.2020; Perry et al,.2013; Vercelloni et al., 2020). Coral reef collapse is an ecological phenomenon at local scales; here we explore where localised coral reef collapse aggregates to the scale of regions, potentially irreversibly, and potentially to a global scale.



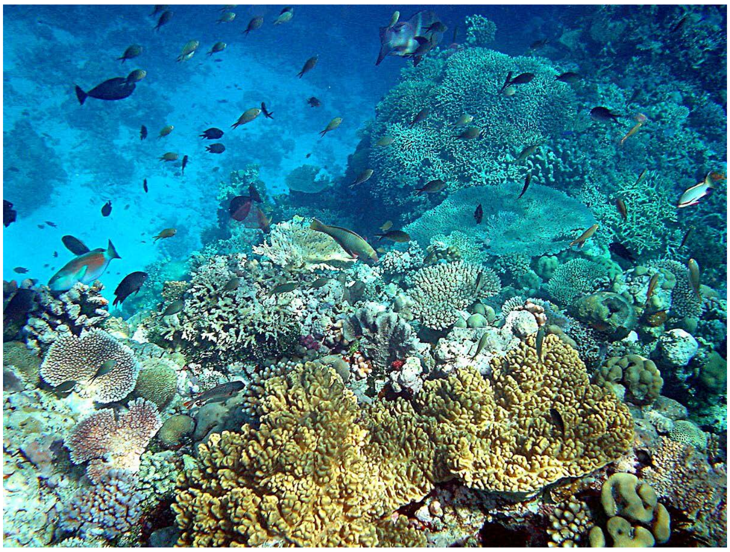



**Figure 1.3.15:** Global distribution of warm water coral reefs and key reef regions (top). ETP is the Eastern Tropical Pacific, PERSGA is the area included within the Regional Organization for the Conservation of the Environment of the Red Sea and Gulf of Aden, ROPME is the sea area surrounded by the eight Member States of the Regional Organisation for the Protection of the Marine Environment, and WIO is the Western Indian Ocean. A coral reef ecosystem in Papua New Guinea in 2003 (bottom). Credit: (Souter et al., 2021) and (top), Brocken Inaglory via Wikimedia (bottom).

Thermal stress, driven by increasingly warmer oceans and superimposed El Niño extreme events, is the primary driver of regional-scale mortality of hard corals (Hughes et al., 2017; Houk et al., 2020). Coral 'bleaching' occurs when thermal stress causes corals to expel the symbiotic algae that provides them with food (resulting in a characteristic loss of colour), and can result in death if it occurs frequently enough to prevent recovery (Hughes et al., 2018a, 2018b, Obura et al., 2022).

However, a wide variety of interacting and synergistic threats co-occur (e.g. ocean acidification, overfishing, pollution, invertebrate predators and sea level rise), generally lowering the thermal threshold for bleaching and/or mortality, bringing forward timing of collapse, or even surpassing thermal stress in local importance (Ban et al., 2013; Edmunds et al., 2014; Darling et al., 2019; Cramer et al., 2020; Dixon et al., 2022; Setter et al., 2022). Coral mortality may play out over weeks to a few months (for e.g. thermal stress-induced bleaching, for example), or years (for chronic threats such as diseases and land-based impacts), but prolonged failure to recover over a decade or more is necessary to qualify a coral reef as 'collapsed'.

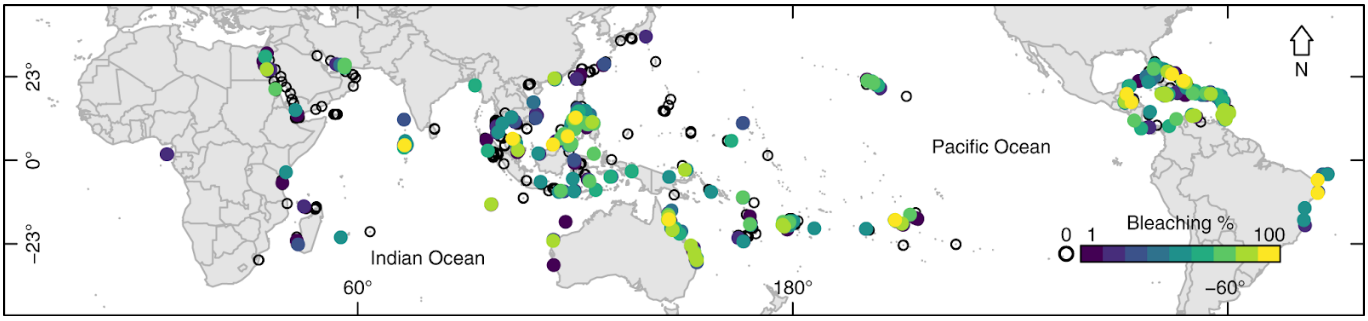



**Figure 1.3.16:** Map of recent coral reef bleaching distribution (as a percentage of the coral assemblage bleached at surveys from 1998 to 2017, with white circles indicating no bleaching. and coloured circles from 1% (blue) to 100% bleaching (yellow)) (top). Photos showing impact of coral bleaching in American Samoa before (left), during (middle), and after (right) the 2015 bleaching event (bottom). Credit: (top) (Sully et al., 2019), (bottom) from The Ocean Agency.

Localised coral responses to increasing stressor magnitude and intensity are now aggregating at scales exceeding 1,000km and manifesting as regional die-offs (e.g. western and central Indian Ocean, Great Barrier Reef, Mesoamerican Reefs) (Le Nohaïc et al.,2017; Amir, 2022; Muñiz-Castillo et al., 2019; Obura et al., 2022), with most reef regions having experienced multiple mass coral bleaching and die-off events (Darling et al., 2019; Cramer et al., 2020) (Figure 1.3.16). Around 50 per cent of global coral reefs are estimated to have been lost over the past 50-150 years (IPBES 2019), with estimated loss of 16 per cent in 1998 (Wilkinson et al., 1999) and measured loss of 14 per cent from 2009-2018 (Souter et al., 2020), but with high variance among regions.

Projected loss of coral reefs has been estimated in varied ways. Dominant projections are of 70-90 per cent loss of coral reefs at 1.5°C and ~99 per cent at 2°C warming (Cooley et al., 2022). The average year for projected global annual severe bleaching under SSP2-4.5 (a trajectory close to current projections) is 2045, which is delayed 30 years if corals can adapt to an additional 1°C of warming (UNEP 2020). A shift occurs from 84 per cent of reefs globally having 'good' thermal regimes in 1986-2019 to 0.2 per cent in 2100 at projections of 1.5°C, and 0 per cent at 2°C warming (Dixon et al., 2022). Finally, the proportion of reefs facing 'unsuitable conditions' increases from 44 per cent in 2005 to, under worst case scenarios, 100 per cent by 2055 under any one of several stressors, but by 2035 for cumulative stressors (Setter et al., 2022). Continued ocean warming over several decades (due to lagged ocean heat uptake) and sea level rise over centuries to millennia (due to thermal expansion and ice sheet melt, see 1.2.3) mean some reefs and other coastal ecosystems (see also Mangroves and Seagrassees) may be committed to eventually passing tipping thresholds even if emissions ceased soon (Abrams et al., accepted).

**Evidence for tipping dynamics**

Failure to recover from mass mortality shows evidence of having crossed a threshold for recovery, which we address for scales above approximately 1,000km, to regional and global scales. A key question is if coral reef decline globally is just an aggregate of regional events, so a linear/chronic decline process (Souter et al., 2021), or if there may be a global tipping point.

Observations on coral reef tipping points include the following:

- The first reported global bleaching event in 1998 was associated with atmospheric warming of ~0.6°C (corresponding to c. 350 ppm $CO_2$) with a strong El Niño on top (Veron et al., 2009), past which more frequent, intense and widespread coral bleaching and mortality has occurred.
- A very high risk of impact to corals was assessed by the IPCC as global mean warming levels crossed around 1.2°C (IPCC SR1.5 2018).
- Thermal bleaching tipping points are already being passed in the majority of coral reef regions (Cooley et al., 2022 – see Figure 1.3.17).
- The risk of ecosystem collapse is already predicted at high levels in all coral reef regions assessed. The MesoAmerican Barrier Reef is Endangered (Bland et al., 2018) and Western Indian Ocean coral reefs are Vulnerable to collapse, with two thirds of subsidiary ecoregions being Endangered or Critically Endangered due to projected warming (Obura et al., 2022).

## Where are we reaching tipping points in the ocean and what can we do about it?

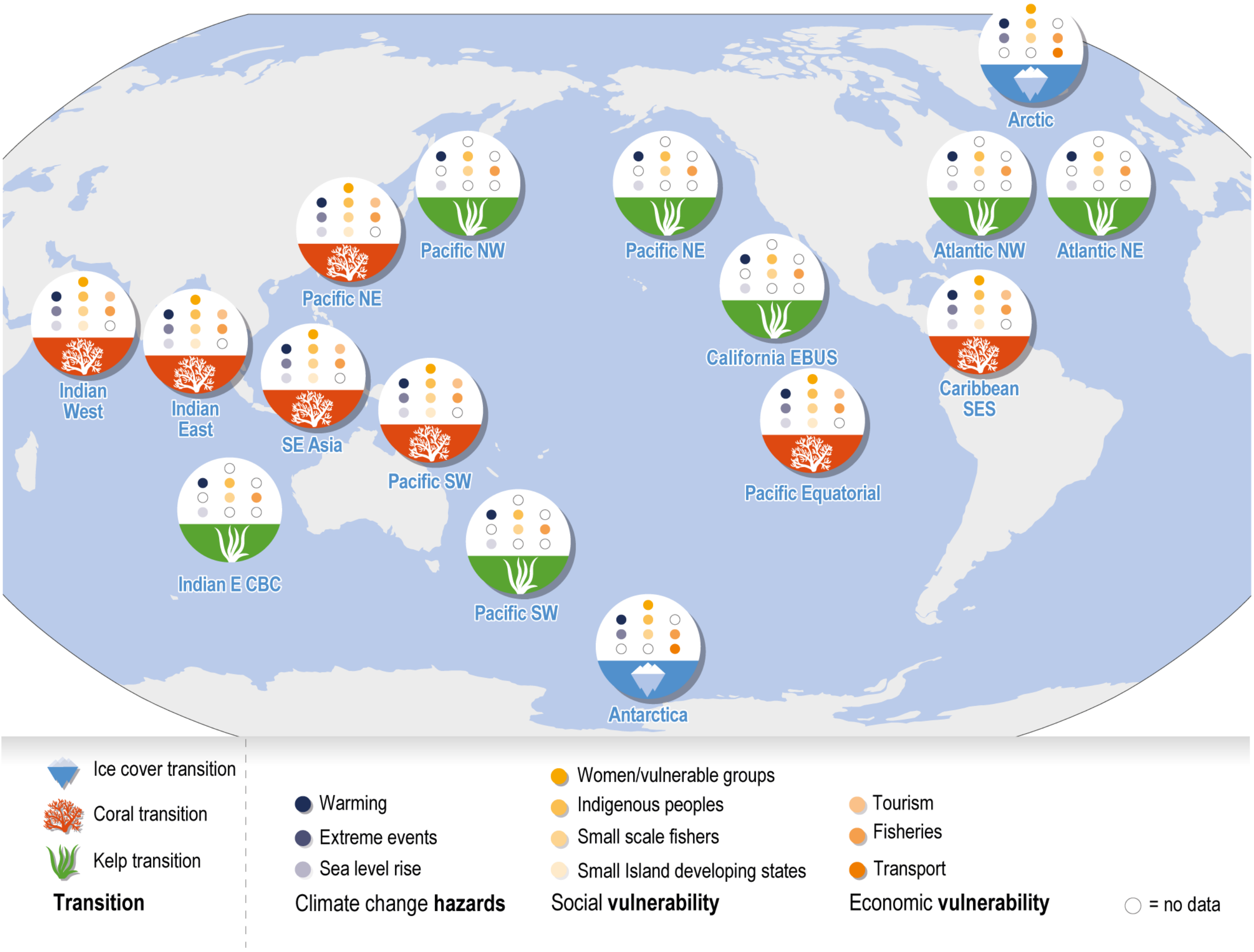



**Figure 1.3.17:** Tipping points have been passed in many ocean ecosystems, including coral reefs, kelp forests, and those associated with sea ice, with diverse socio-economic implications. [From FAQ3.3.1 in (Cooley et al,. 2022).

Elevated summer ocean heat maxima (over 1-2°C above site-specific individual coral acclimation thresholds) for weeks to months, and larger acute temperature spikes for several days, cause severe coral bleaching and mass mortality. Mass-mortality bleaching thresholds have been proposed at eight "Degree Heating Weeks" (a measure of how long and how much ocean temperatures are above normal) which is likely by ~2°C global warming (McWhorter et al., 2021), or at two bleaching events per decade (likely by ~1.5°C) (Frieler et al., 2013). Mass coral mortality repeated more than twice per decade and over hundreds to thousands of kilometres and larger, is increasingly recognised as giving insufficient time for recovery of impacted populations, and of ecological interactions (Hughes et al., 2018a, 2018b, Obura et al., 2022). However, estimating globally consistent warming thresholds is challenging given variation from individual corals to species and across all spatial scales in acclimation and adaptation ability. Other stressors reduce the ability of corals to resist thermal stress, thus bringing down tipping thresholds.

Increasing frequency and intensity of regional-scale coral mortality events past 1°C warming are suggestive that these coral reef regions have already passed regional bleaching tipping points (Cooley et al., 2022). The potential for thermal refuges for corals under likely future scenarios is doubtful (Beyer et al., 2018; Dixon et al., 2022; Setter et al., 2022) as very few or no reef areas are projected to remain below tipping thresholds of key stressors. The existence of putative refuges at greater depths (Bongaerts and Smith 2019) or higher latitudes (Yamano et al., 2011; Setter et al., 2022) are not strongly supported by recent work (Hoegh-Guldberg et al., 2017; Cooley et al., 2022). Ecological and biogeographical (spatial) positive/amplifying feedback loops prevent local recovery of coral reefs and promote expansion of reef collapses from local to regional scales when surviving corals and coral patches become too spatially separated for successful reproduction of adults, and supply of larvae from surviving to damaged reefs (Hock et al., 2017).

Coral reef decline does not substantially feedback to the climate system on policy-relevant timescales. However, localised surface cooling may arise through increased low level cloud albedo induced by sulphur compounds released by reef metabolism. Consequently, extensive coral die-offs could amplify local warming (Jackson et al., 2020).

### Assessment and knowledge gaps

Warm-water coral reefs have localised tipping points (high confidence) and are now experiencing regionally clustered tipping points (high confidence). Based on the evidence collected here, we suggest that the critical threshold of 1.5°C (range 1-2°C) (Armstrong McKay et al., 2022) should be adjusted, narrowing and lowering the range to 1-1.5°C, with a middle estimate of 1.2°C, marked by the multi-year global coral reef bleaching events of 2015-2017 (Cooley et al., 2022; Hoegh-Guldberg et al., 2018; Dixon et al., 2022; Setter et al., 2022). The co-occurence of additional synergistic drivers also support lowering the critical threshold (Willcock et al., 2023) and there is evidence of accelerating collapses at increasing spatial scales (Cooper et al., 2020).

The combined effects of long-term warming, sea level rise, ocean acidification and other stressors bears more investigation to identify the lower critical threshold for the coral reef tipping point. The potential for coral adaptation to warming is a critical but poorly known factor, and subject to high levels of variation locally. The potential effectiveness of restoration for coral reefs at scale, and with enhanced capacity to resist future threats, are both currently poor. The effect of climate migration on coral recovery is not known, with potentially positive effects at higher latitude (with in-migration), but negative at lower latitudes (with out-migration, but no replacement; Herbert-Read et al., 2023).

### Mangroves and seagrasses

Mangroves and seagrasses play vital roles in coastal societies and economies. They provide fundamental and hard-to-substitute ecosystem services such as support to fisheries, nutrient cycling, coastal protection and sediment trapping (Malik et al., 2015; Nordlund et al., 2016; Menéndez et al., 2020; Nabilah Ruslan et al., 2022; doAmaral-Camara et al., 2023, James et al., 2023). Located between the sea and the land, their unique dual nature exposes mangroves and seagrasses to climate drivers that arise in both systems (Lovelock et al., 2017a; Duke et al., 2017a, 2019), making them particularly vulnerable to climate change (Duke et al., 2022). Recent attention has focused on their climate mitigation services ('blue carbon') linked to their high productivities and long-term (millennia) storage of organic matter in their sediments, which positions them among the most dense carbon sinks on Earth (Donato et al., 2011; Alongi et al., 2016; Macreadie et al., 2021; Serrano et al., 2021).

While they occupy small areas (c. 140,000 sq km and uncertain c. 266,562 sq km for mangroves and seagrasses respectively in 2020; Bunting et al., 2022; McKenzie et al., 2020; Figure 1.3.18 and 1.3.19), they store up to 12.3 GtC and 3.8 GtC respectively (Macreadie et al., 2021). These ecosystems are natural sinks of $CO_2$, but when degraded they can release $CO_2$, $NO_2$ and $CH_4$, adding to the emissions of the estuaries they are embedded in (Rosentreter et al., 2022). Emissions derive from carbon stored long-term in sediments, which cannot be recovered in a lifespan and is therefore additional to the current atmospheric balance (Lovelock et al., 2017b; Schorn et al., 2021; Romero-Uribe et al., 2022).

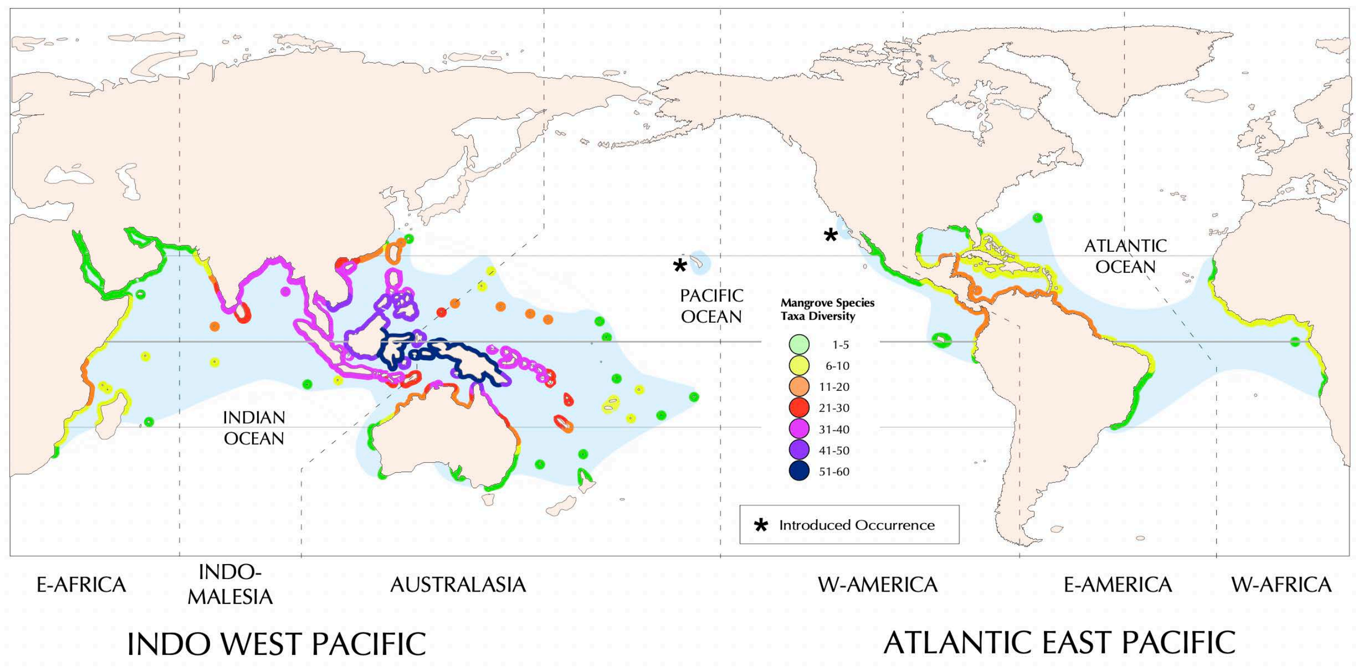


**Figure 1.3.18:** Upper panel: floristic distribution of mangroves in the world, with a marked diversity in the Wallacea region (Indo Pacific). Lower panel: white mangrove (*Laguncularia racemosa*) from Yucatan, showing the intricacy of mangrove roots, and their service as fish habitat, coastal protection against storms and sediment trapping. Source: (Duke et al., 2017) (top) and Jorge Herrera, CINVESTAV (bottom).

Mangroves and seagrasses are historically among the most human-threatened ecosystems in the world (Valiela et al., 2001; Waycott et al., 2009), with 35-50 per cent of mangroves' original cover now lost, mainly to aquaculture and agriculture (Richards and Friess, 2016, Goldberg et al., 2020; Hagger et al., 2022), while other factors including nutrient overload, invasive species, and ocean warming have led to a 19-30 per cent decrease of the original seagrass surveyed area (Waycott et al., 2009; Dunic et al., 2021).

In spite of this, the magnitude of their past and current feedback to global warming remains uncertain (Rosentreter et al., 2022). Under current rates of deforestation, estimates of global mangrove emissions by the end of the century range between 0.24 to 0.34 Gg $CO_2$e if foregone soil carbon sequestration is also included (Adame et al., 2021), which is comparable to the European Union's emissions in 2022. Southeast and South Asia (West Coral Triangle, Sunda Shelf and the Bay of Bengal) are projected to lead the emissions, followed by the Caribbean (Tropical Northwest Atlantic), the Andaman coast (West Myanmar), and northern Brazil (Adame et al., 2021).

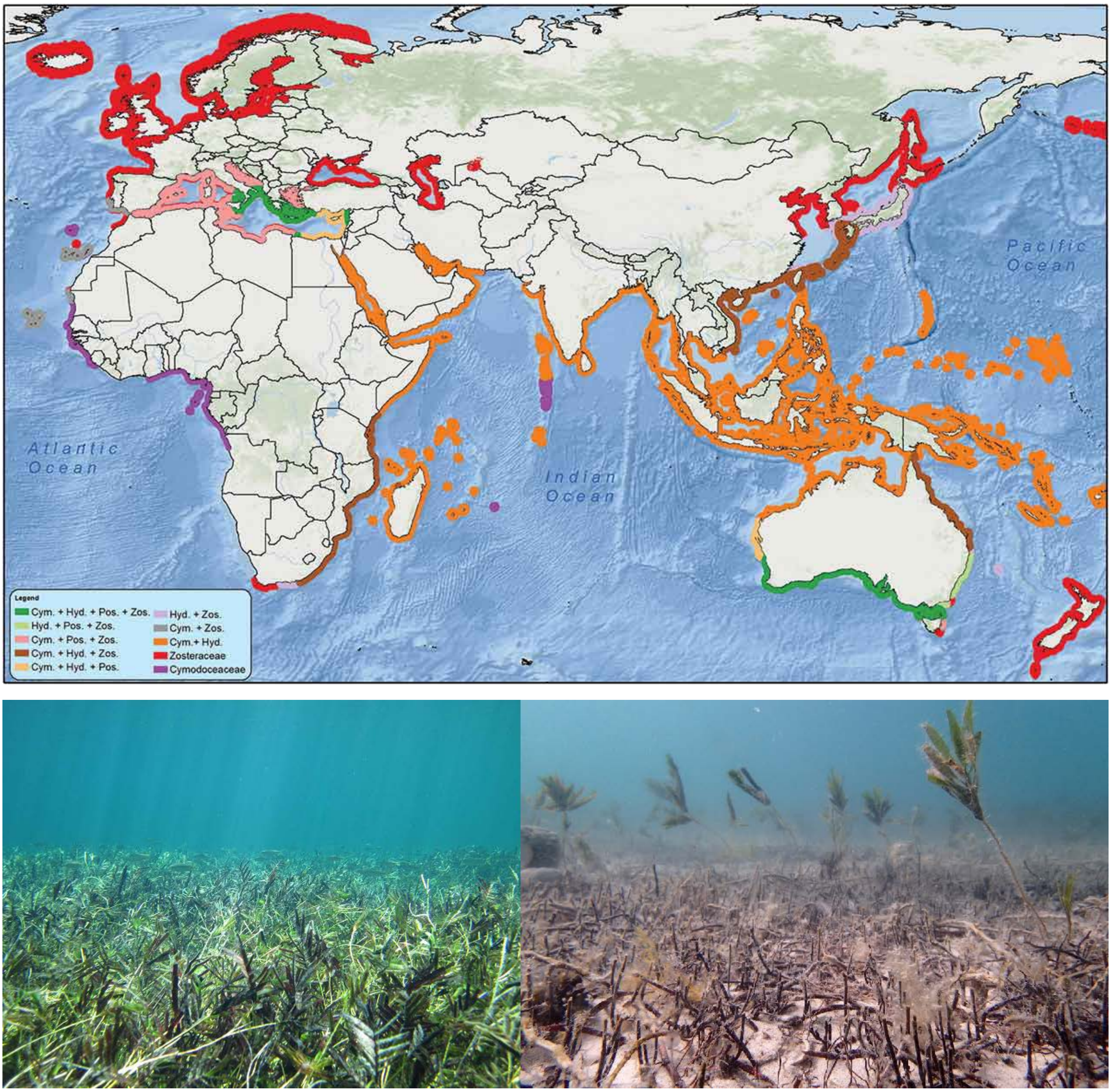



**Figure 1.3.19:** Upper panel: global distribution of seagrasses. Lower panel: Shark Bay temperate seagrass (*Amphibolis antarctica*) before the 2011 heatwave and after (2013). Revisits from 2012 to 2014 verify poor recovery of A. antarctica, and the slow expansion of the tropical seagrass *Halodule uninervis*, in sites with no recovery (30% of cover three years later). Source: IUCN, map created by T. Bakirman. Seagrass die-off: credit goes to the Shark Bay Ecosystem Research Project and (Nowicki et al., 2017).

### Evidence for tipping dynamics

In spite of major historical habitat loss and degradation, there are not yet generalised signs of irreversible global transitions of mangroves towards alternative states such as tidal flats, and the remaining systems have so far retained large-scale stability in the tropics. Bistability is, however, observed in northern subtropical distributions with mangrove encroachment over tidal marshes where freezing events are now rarer (Feller et al., 2017; Hesterberg et al., 2022). Observational data also suggests rainfall-induced bistability of mangroves and salt marshes (Duke et al., 2019).

Scarce global monitoring prevents robust analyses of seagrass trends, but transitions (>50 sq km) towards unvegetated sediments have intensified in many coastal regions in the last two decades (e.g. Europe, Australia, US, Caribbean) (Waycott et al., 2009; Carr et al., 2012; Arias-Ortiz et al., 2018; Duarte et al., 2018; Kendrick et al., 2019; Cooley et al., 2022; MacLeod et al., 2023) (Fig 1.3.19). For temperate regions, bistability and tropicalisation of temperate seagrass species are observed in edge-of-range meadows, with uncertain stability trends (Bartenfelder et al., 2022). For tropical seagrasses, local resilience after disturbance has been observed when enough time and reduced pressures apply (MacLeod et al., 2023).

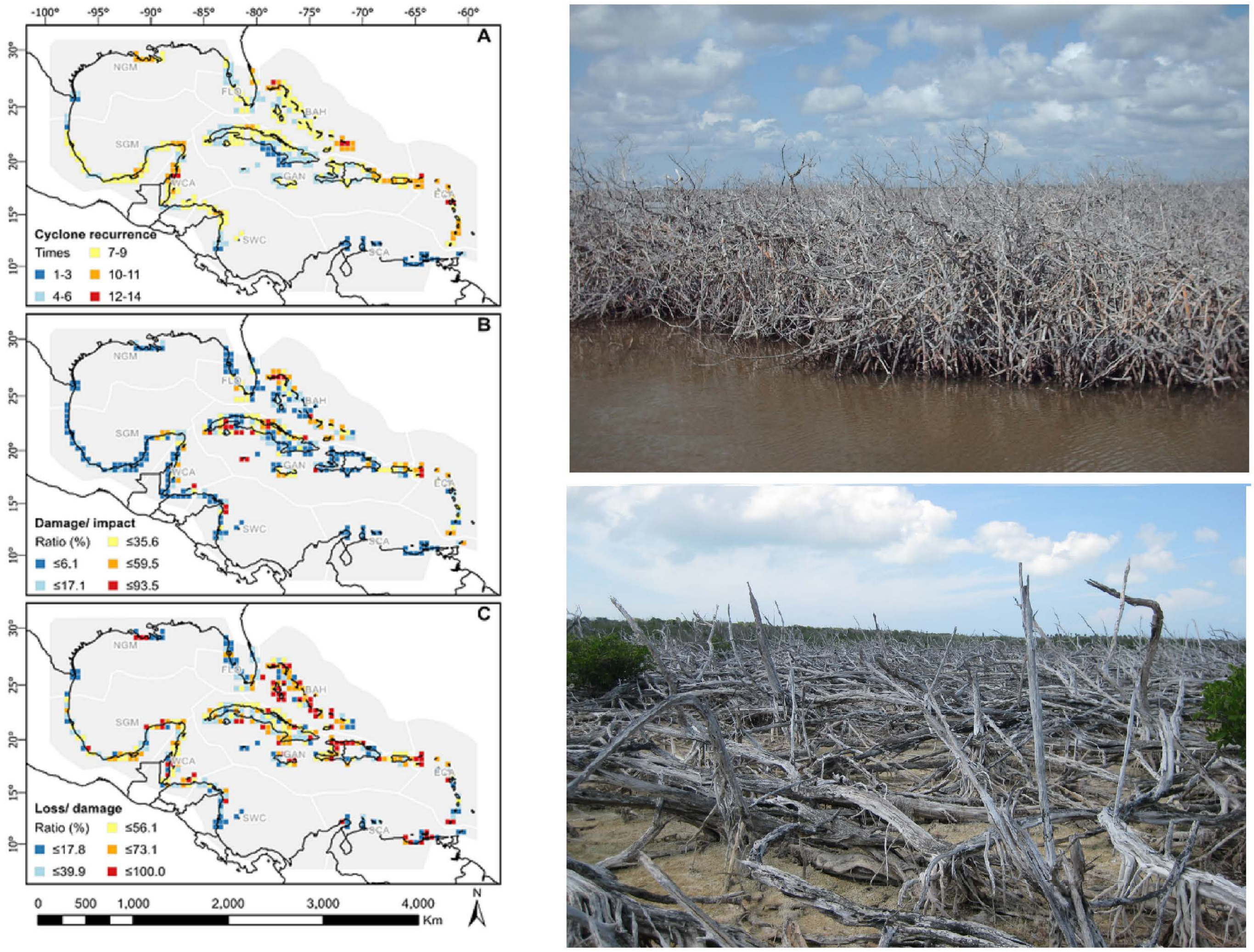



**Figure 1.3.20:** Left panel presents (A) the recurrence of tropical cyclones (from tropical storms to hurricanes category 5) in different subregions of the North Atlantic Basin (Caribbean, Gulf of Mexico, Mesoamerica), (B) percentage of pixels hit by a tropical cyclone where mangroves show damage six months after the pass of the storm (vulnerability), and (C) percentage of pixels that showed damage after a storm that do not show signal of recovery one year after being damaged (resilience). Right panel includes photos from mangroves hit by hurricanes in Yucatan. Sources: (Amaral et al., 2023) (left panel), Jorge Herrera, CINVESTAV (right panel).

While the resilience of these systems (particularly mangroves) does not yet seem compromised at the global scale, there is increasing evidence of region-dependent declines in resilience for both seagrasses (Dunic et al., 2021; Turschwell et al., 2021) and mangroves (Bergstrom et al., 2021, Friess et al., 2022; Amaral et al., 2023, Duke et al., 2023 in press). These responses relate to:

i) An increased exposure to more frequent and intense extreme events such as hazardous cyclonic activity (Figure 1.3.20), more frequent and intense El Niño (Figure 1.3.21) and marine heatwaves (Fig 1.3.19), which add to the long existing human pressures (nutrient overloads, land use changes, sedimentation rates, etc) and to the long-term environmental impacts that promote mangrove and seagrass mortality (including sea level rise, ocean acidification, ocean/atmosphere warming, regional drought, salinity, hypoxia, diseases and invasive species) (Waycott et al., 2009; Krauss et al., 2014; Lovelock et al.,2015; Feller et al., 2017; Duke et al., 2021; Friess et al., 2022; MacLeod et al., 2023).

ii) Shortened recovery times below re-establishment needs. Post-disturbance recovery has been reported to take ca. 10-20 years depending on the ecosystem service considered (Lugo 1980; Jimenez et al., 1985; MacLeod et al., 2023), with mangrove recovery taking c. 20 years (more on arid climates), and c.10 years for seagrasses. A decade has been considered the absolute minimum successful re-establishment time for both systems, if pre-disturbance conditions (hydrological stability and seed sources) were retained (Lugo 1980; Teutli-Hernandez et al., 2020, Duke et al., 2023 in press; MacLeod et al., 2023). Revisiting times are currently below these thresholds in many regions,

iii) Unprecedented increases in compound extreme events that precede, succeed, or coincide in time and space and amplify ecosystem responses (Allen et al., 2021). Along this line, magnified mangrove mortality due to drought-hurricane duos has already been reported in the Caribbean (Taillie et al., 2020; Amaral et al., 2023).

iv) Exposure to multivariable extreme pressures (Fig 1.3.22). While models frequently focus on a few independent-forcing variables, in reality multiple amplifying, synergistic or antagonic effects occur among stressors. As an example, El Niño combines multiple variables such as heat, drought, flooding, more extreme oscillations in sea level (e.g. Taimasas in the Indo-Pacific), and marine heatwaves, whose combined interaction amplifies mangrove and seagrass mortality.

Decreasing resilience enhances damages in coastal habitats, including severe losses of biodiversity, collapse of regional fisheries and aquaculture, and reduced capacity of habitat-forming species to protect shorelines, preventing re-establishment (Cooley et al., 2022). These make mangroves and seagrasses likely candidates for regional tipping points, with major social and economic consequences.

Additionally, lagged ocean warming (over decades) and sea level rise (over centuries) mean coastal ecosystems will continue to face increasing pressure after atmospheric warming stabilises, meaning tipping can be committed decades before it is realised (see warm-water coral reefs above).


**Figure 1.3.21:** Mangrove die-off in physiologically stressed mangrove systems after intense El Niño-driven droughts (2015-2016, 2019) combined with other interacting stresses (prolonged ocean retreat in the Indo Pacific, previous eutrophication in the Bay of Panama, timber extraction, etc). a) El Niño 2015-2016 effects over Australia's Gulf of Carpentaria (8,000 hectares of affected mangroves), b) mangrove die-off in the Maldives has been reported in 11 islands since mid-2020, c) mangrove die-off in the Bay of Panama (Juan Diaz site) after the 2015-2016 El Niño on an eutrophic, rapidly sedimented and colonised site. Sources: Norman Duke (James Cook University), Steve Paton (STRI-Panama), Save Maldives Campaign and Neykurendhoo Island Council (2020).

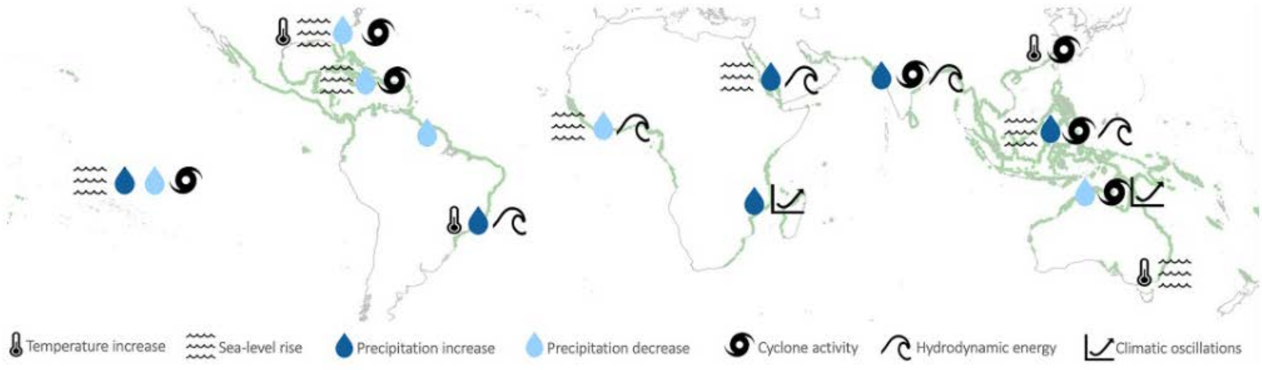



**Figure 1.3.22.** Regional differences in climate drivers (long-term trends and extreme events) leading to mangrove impacts. Combined with human and other environmental impacts, they are expected to lead to different regional tipping timings and degradation speeds. Source: (Friess et al., 2022).

On the potential tipping dynamics of coastal systems, the IPCC AR6 chapter on ocean and coastal ecosystems (Cooley et al., 2022) noted "irreversible phase shifts with global warming levels >1.5°C, making both systems at high risk this century even in <1.5°C scenarios that include periods of temperature overshoot beyond 1.5°C (high confidence). Mangroves, under SSP1-2.6, are expected to be unable to keep up with sea level rise by 2050, with ecological impacts escalating rapidly beyond 2050".

(Saintilan et al., 2020, 2023) found it very likely that mangroves were unable to initiate sustained accretion when relative sea level rise rates exceeded 6.1 (4-7) mm/year. This threshold is likely to be surpassed on low-latitude tropical coastlines within 3-5 decades under high-emissions scenarios (Sweet and Park 2014; Saintilan et al., 2020, 2023). For seagrasses, the IPCC AR6 (Cooley et al., 2022) projects contractions of temperate edge-ranges (e.g. ***Zostera costera*** seagrasses in the US would retract by 150-650km under RCP2.6 and RCP8.5, respectively and ***Posidonia oceanica*** in the Mediterranean Sea, which might lose as much as 75 per cent of their habitat by 2050 under RCP8.5 and become functionally extinct by 2100). Marine heat waves will escalate seagrass responses, with moderate responses to sea level rise (Cooley et al., 2022).

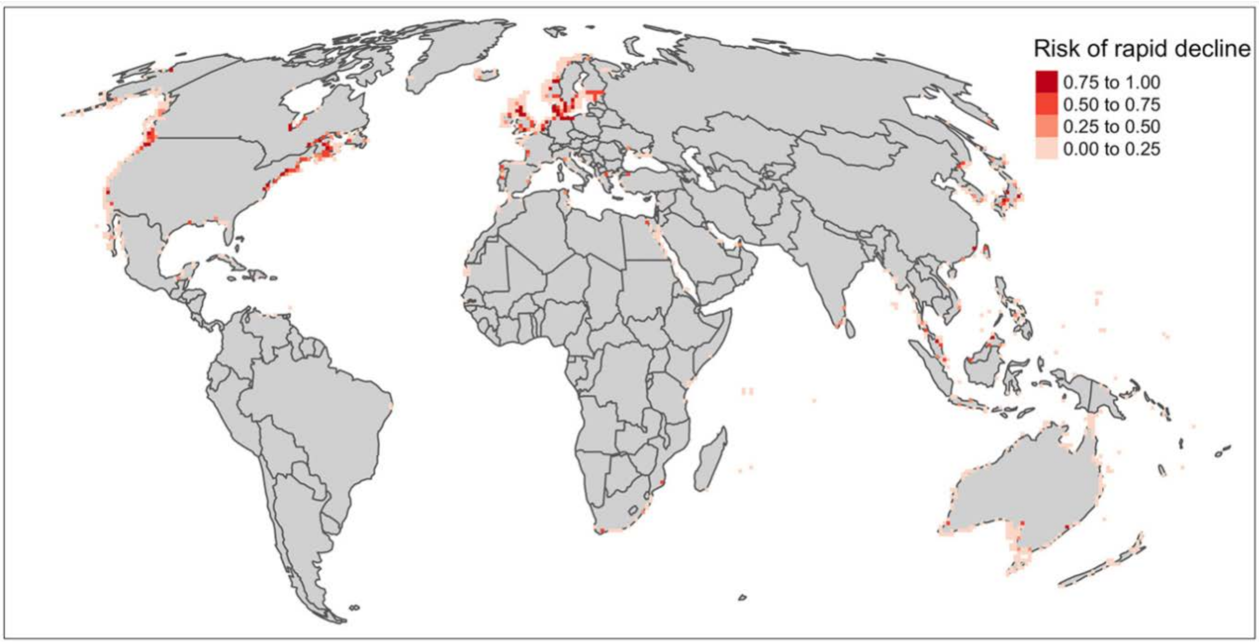



**Fig 1.3.23:** Rapidly declining trajectories of seagrass meadow extent (>25% loss from 2000 to 2010) predicted in 100×100 km grid cells. Sites are coloured by the probability of a site being ranked among the 10% of sites most likely to have a rapidly decreasing trajectory. Predictions were most strongly associated with high pressures from destructive demersal fishing and poor water quality. Source: (Turschwell et al., 2021).

**Assessment and knowledge gaps**

We conclude with medium confidence that, under current relative sea level rise projections, subsidence, expected increases in extreme events and coastal development (Cooley et al., 2022), tipping responses for mangroves are likely to be regionally visible by 2080 at temperature thresholds between 1.5-2°C (starting with physiologically stressed regions that host increasing extreme events – also medium confidence). Seagrasses are likely (medium confidence) to show region-dependent die-off responses earlier (by mid century) due to more intense and recurrent marine heatwaves, nutrient pollution and turbidity, at global temperature thresholds closer to 1.5°C (medium confidence).

We have high confidence that tipping responses will be region and site-dependent with diverse timings and degradation speeds. For mangroves, physiologically stressed regions such as arid or highly seasonal climates like the Middle East or the dry corridor of Central America, karstic systems such as the Caribbean, small islands, northern Australia, or the northern Coral Triangle are likely (medium confidence) to show tipping responses earlier than other regions such as the Indo-Pacific, South America or parts of the Indian Ocean, whose systems either have more species, are less exposed, or are less vulnerable to hazard exposure (e.g. there is more space for encroachment, or more refugia).

For seagrasses, temperate regions are predicted to be more vulnerable to tipping than warmer regions (Turschwell et al., 2021; Green et al., 2021; Cooley et al., 2022) (Fig. 1.3.23). Seagrasses in warm regions that are more exposed to water pollution, turbidity, extreme events (marine heat waves and cyclones), coastal development, salinity or invasive species are expected to tip earlier than seagrasses in other warm regions.

Compared to the IPCC AR6 report (Cooley et al., 2022), we highlight a higher confidence on the directional effects of storms on both mangroves and seagrasses towards regionally synchronous mortality (Carlson et al., 2012; Wilson et al., 2019; Taillie et al., 2020; Amaral et al., 2023; Duke et al., 2023 in press). Evidence also exists on decreased regional resilience in mangroves after cyclones (Amaral et al., 2023) and transitions to mudflat shifts in areas where storms combine with erosion co-stressors (Bhargava and Friess 2022). Similarly, warming responses in mangroves have a clearer directional trend, with extreme El Niño hot-droughts superimposed onto global warming and regional drought leading to well-known extended mangrove mortality in many regions (Jimenez et al., 1985), including recent reports of die-off in Australia (Duke et al., 2017a), Panama (Fig. 1.3.21) and the Maldives (Save Maldives Campaign and Neykurendhoo Island Council, 2020).

Current modelling does not yet properly cover extreme events or multiple drivers, nor their interactions (Cooley et al., 2022). These gaps are likely leading to an underestimation of their impacts on ecosystems and their long-term resilience thresholds. Resilience responses to enhanced stressors will be region- and site-dependent, but models still need data to properly represent key drivers per region and their interactions, as well as the thresholds of survival of regional ecosystems (Marba et al., 2022).

### 1.3.2.8. Marine ecosystems and environment

Climate change, pollution and overexploitation are affecting the marine environment at the physical, chemical and biological levels (e.g. Heinze et al., 2021; Jouffray et al., 2020; Bindoff et al., 2019). Pelagic marine ecosystems (defined as the water column from the surface ocean to the seafloor) as well as benthic marine ecosystems (defined as restricted on the seafloor) from the organism to the community level are changing at the same time as the ocean waters are becoming more warm, acidic and deoxygenated. In this section, we outline five potential tipping systems ranging from fisheries collapse and regime shifts in marine communities to ocean water hypoxia and the nonlinear weakening of parts of the ocean's biological pump (Figure 1.3.24).

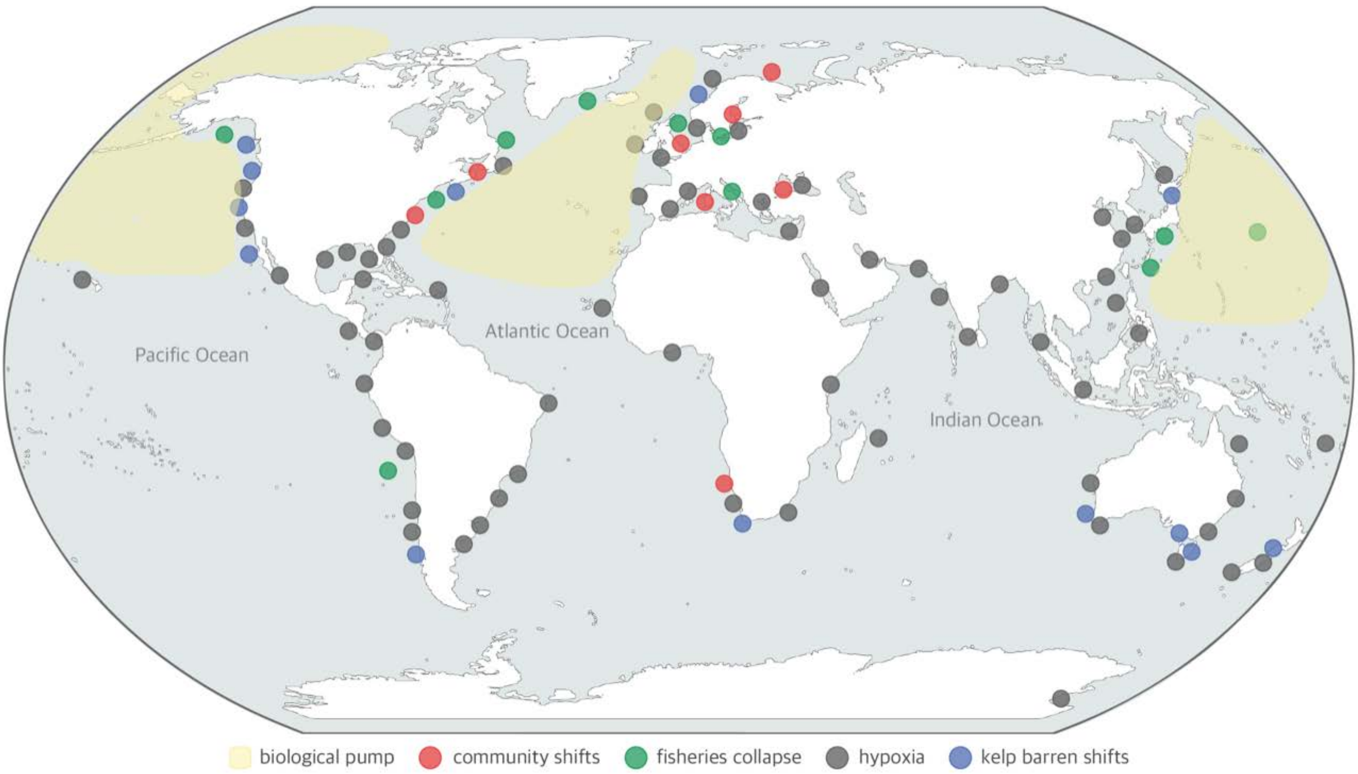



**Figure 1.3.24:** Locations of reported regime shifts and potential tipping points in the global marine environment. Redrawn and updated from (Blenckner and Niiranen, 2013).

### Evidence for tipping dynamics

#### Fisheries collapse

Over the past decades many fisheries have collapsed primarily due to over-exploitation, but they are increasingly threatened by climate change.

Fish stocks are defined as management units of a species; thus one fish species can have multiple stocks (e.g. more than 20 stocks in the North Atlantic are assessed for Atlantic cod, *Gadus morhua*).

**Figure 1.3.25:** A school of fish: Credit: iStock.com/armiblue.

Among more than 200 exploited fish stocks, 23 per cent of the species showed at least one stock collapse (biomass below sustainable reference points) (Pinksy et al., 2011). Concerningly, 40 per cent of the collapsed stocks present different regimes of productivity (different relationships between fishing and biomass at different productivity stages) (Vert-pre et al,.2013) that potentially indicate the presence of regime shifts and hysteresis. But, while for some species there is clear evidence of regime shifts (Atlantic cod stocks), for others more studies are needed (Frank et al., 2016; Sguotti et al., 2019).

Fish stock collapses can be due to different feedback mechanisms. The collapse of a stock can induce food web changes (i.e. trophic cascades) that, by modifying the other species of the community and their interactions, can maintain the population at a low level through predation or competition. For instance, large predators such as Atlantic cod may be successful because of the 'cultivation effect': adult cod prey on the juveniles of forage fishes (small pelagic fish which are preyed on by larger predators) that are competitors or predators of juvenile cod. Once the collapse in the biomass of cod occurs, the predation on the forage fish is released and these species start to thrive. Forage fish then prey on juvenile or recruit cod, thus maintaining the population in a depleted state. Examples of this particular dynamic can be found in Newfoundland and also the Baltic Sea (Walters and Kitchell, 2001). Another possible mechanism of hysteresis is the so-called Allee effect, which takes place when recruitment of a population (the process by which new organisms are added to a population) is positively correlated with its biomass.

This means that a minimum population size is needed for the population to grow; otherwise it collapses. Thus, if biomass collapses, recruitment will also drastically decline, limiting the capacity of the population to recover. The Allee effect has been shown to be one of the possible hysteresis mechanisms of 13 stocks of Atlantic cod (Winter et al., 2023).

It is difficult to detect specific thresholds in fisheries in general, since every species and every stock within each species is impacted by different levels of the same driver and may experience different pressures. However, it has been shown that, for Atlantic cod stocks, the threshold was created by the combination of multiple drivers, especially warming and fishing (Sguotti et al., 2019; Beaugrand et al., 2022). Specific thresholds need to be detected for every stock.

Beaugrand et al., (2022) have shown that rebuilding cod stocks may depend upon the fishing-environment interaction. When the environment becomes unsuitable at the same time fish stocks collapse, rebuilding the stock may take time or even be impossible so long as adverse environmental conditions persist. This provides an explanation as to why, despite the fishing moratorium near Newfoundland, a partial recovery took more than two decades (DFO 2018). Long-living, slow-growing species might be more prone to irreversibility. For instance, 16 out of 19 Atlantic cod stocks present regime shift dynamics due to fishing and warming and their recovery is hindered by the presence of hysteresis (Sguotti et al., 2019; Möllmann et al., 2022., Frank et al., 2016).

### Marine community shifts

Marine community shifts take place when abrupt changes cascade through several species or functional groups of an ecosystem, i.e. the change is not limited to a single species, as in a fish stock collapse, but can cascade all the way from top predators to phytoplankton (Figure 1.3.26).

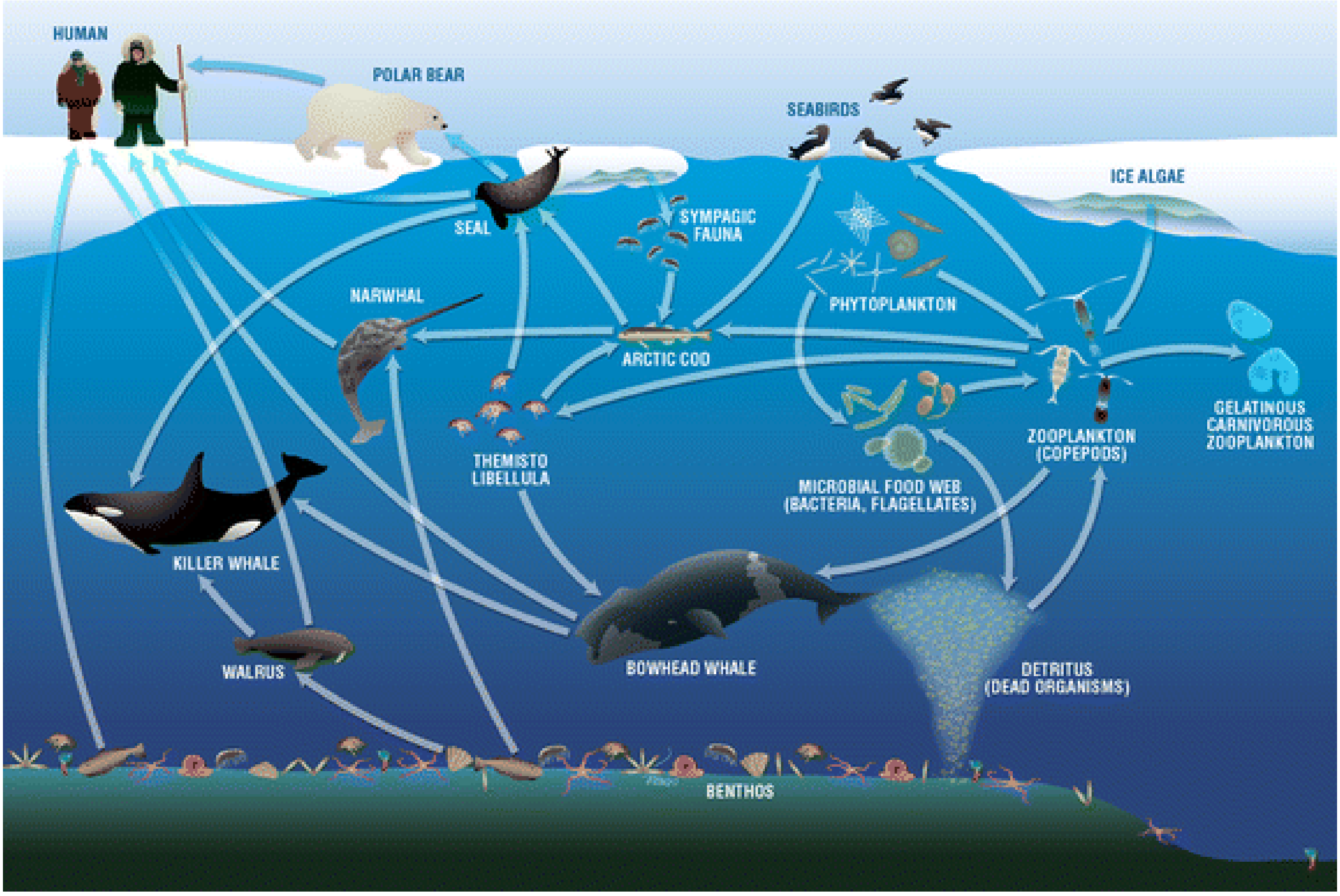



**Figure 1.3.26:** Schematic of a marine food web. Source: (Darnis et al., (2012).

Many community shifts have been reported in marine ecosystems (Conversi et al., 2015; Beaugrand et al., 2019; Möllmann et al., 2021; Ban et al., 2022; Sguotti et al., 2022). Some ecosystems have even experienced several marine community shifts, such as the Black Sea and Baltic Sea. In the Black Sea, the first major shift started in the end of 1960s with the overfishing of pelagic top predators, enabling surplus phytoplankton and jellyfish production during the following decades, and resulting in increased hypoxia (lack of oxygen necessary for life) followed by collapse of small pelagic fish and domination of jellyfish (Daskalov et al., 2017). In the Baltic Sea, the increased inflow of nutrients and organic matter resulted in the eutrophication of the main basins around the 1950s, enabling higher biological production, but also worsening hypoxia (Österblom et al.,(2007).

Community shifts related to tipping responses mostly occur when the system is controlled by a few key species through trophic cascade (Beaugrand et al., 2015; Daskalov et al., 2007, 2017). Trophic cascades can be environmentally induced or induced by anthropogenic pressures such as pollution or overfishing (Casini et al., 2009). The mechanisms at the origin of the apparent synchronicities among marine community shifts have been debated (Conversi et al., 2010a, Beaugrand 2015). Möllmann and Diekmann, (2012) suggested that multiple drivers, such as climate and overfishing, may interact in triggering ecosystem community shifts between alternative states. Reid and Beaugrand (2012) observed that, in many cases, the reported shifts coincided with major temporal changes seen in marine temperature anomalies. The interaction between climate-induced environmental changes and species' ecological niches (Beaugrand 2015; Beaugrand et al., 2019) may lead to a community shift. For such shifts, the existence of tipping is not needed as an explanation.

Another region of potential climate change-induced regime shifts is the Arctic Ocean. As summer sea ice declines, spring phytoplankton blooms are becoming possible, leading to Arctic ecosystems becoming more like the present North Atlantic and productivity increasing by 30-50 per cent (Yool et al., 2015). Warming and circulation changes can also lead to the spread of invasive species – for example in the Barents Sea and from the Pacific (Kelly et al., 2020; Neukermans et al., 2018; Oziel et al., 2020) (see 1.4.2.1). However, while these changes may trigger regime shifts, it is currently difficult to predict whether they will feature self-sustaining tipping dynamics.

Empirical thresholds for marine communities have been estimated in specific cases using ecosystem model-derived indicators of community status (e.g. Samhouri et al., 2010), but are in general challenging to identify. Evidence for irreversibility is anecdotal and case-specific. One example is shifts in the anchovy-sardine cycles (Schwartzlose et al., 1999) that occur worldwide. Such shifts appear to be triggered by changes in short and long-term climate conditions. In the Peruvian upwelling system, switches in climate cycles can thus correspond to tipping points for the community (Alheit and Niquen 2004; Chavez et al.,2003), with effects on the middle (decadal) to long (centuries) timescale (Salvatteci et al., 2018). Evidence for this system suggests that natural fluctuations and anthropogenic climate change may pose an increased risk of tipping toward irreversible changes to a community characterised by less desirable (from a social-ecological perspective) and less productive features (Salvatteci et al.,2022).

### Kelp forests

Kelp forests are mostly coastal ecosystems dominated by dense populations of large brown macroalgae (Figure 1.3.27). In recent decades, a significant number of these forests have undergone devastating collapses, resulting in their transformation into desolate and unproductive communities, called barrens. These collapses are primarily driven by overgrazing by sea urchins (Ling et al., 2015). However, additional pressures, such as marine heatwaves (McPherson et al., 2021), nutrient concentration (Boada et al., 2017) and sedimentation (Foster and Schiel, 2010), also contribute to its formation.

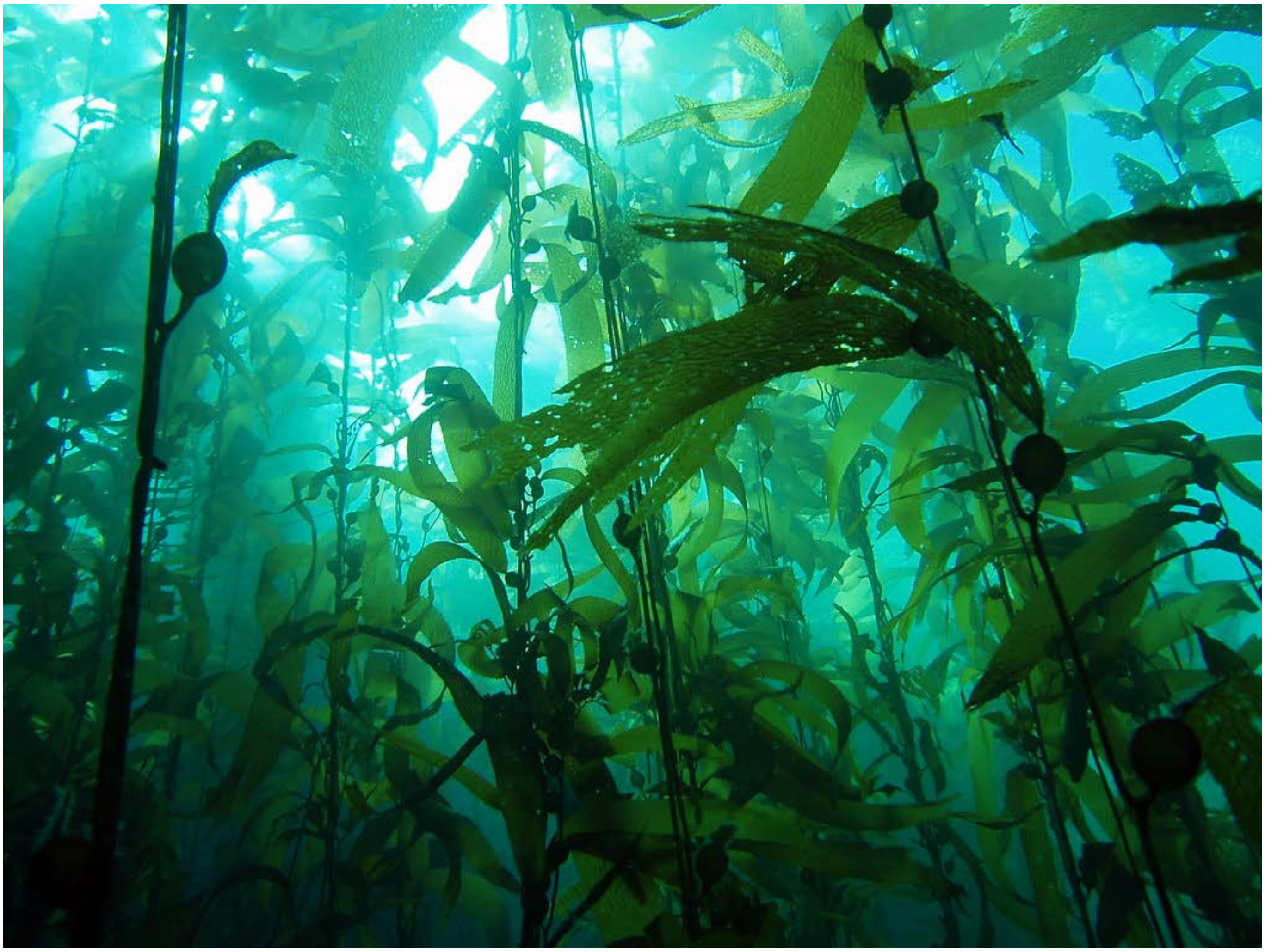


**Figure 1.3.27:** Kelp forest at Anacapa Island, California, 2010. Source: Dana Roeber Murray, flickr

Persistent, catastrophic regime shifts in coastal rocky communities transitioning between productive macroalgal beds and impoverished sea urchin barrens have been shown to occur worldwide (Ling et al., 2015). In many cases, such regime shifts exhibit nonlinear dynamics with hysteresis, where the transition shifts exhibit tipping points (Filbee-Dexter and Scheibling, 2018). Thresholds can be estimated empirically through a critical density of sea urchins (Ling et al., 2015), but such thresholds are influenced by biotic and abiotic factors.

Two feedbacks promote the stability of the barren state: processes that reduce kelp recruitment on barrens and processes that allow sea urchins to maintain high densities on barrens (Filbee-Dexter and Scheibling, 2018). For example, adult sea urchins seem to provide shelter and facilitate survival of urchin recruitment, offering a reinforcing mechanism. Similarly, barren conditions are kept open by intense grazing, reducing the chances of kelp recruitment.

Empirical studies have demonstrated the possibility of kelp forest recovery once sea urchin densities are limited (Smith and Tinker, 2022; Galloway et al., 2023). However, such recovery is influenced by abiotic factors such as marine heat waves, making kelp forest reversibility uncertain.

### Biological carbon pump

The biological carbon pump (BCP) refers to the suite of processes that remove ~50 Gt of carbon annually from the atmosphere and into marine biomass, transferring ~10 per cent of this into the deep ocean (Carr et al., 2006; Westberry et al., 2008; Fu et al., 2016). Without this flux, atmospheric $CO_2$ would likely be ~200 ppm higher than the present-day concentration (Henson et al., 2022).

The largest component of the BCP, the gravitational pump, is driven by sinking of organic matter, mostly from dead plankton and detritus such as faecal pellets (Figure 1.3.28) (Nowicki et al., 2022). This part of the BCP is expected to decline with warming as a result of reduced mixing between warming surface and colder deep waters (thermal stratification) leading to reduced nutrient supplies for surface algae (i.e. phytoplankton), as well as warming favouring smaller plankton species that contribute less sinking matter (Armstrong McKay et al., 2021). However, there is no known mechanism that would enable this decline to become self-sustaining, with changes scaling quasi-linearly with emissions in models, and it is therefore not considered to show tipping-point behaviour (Armstrong McKay et al., 2022).

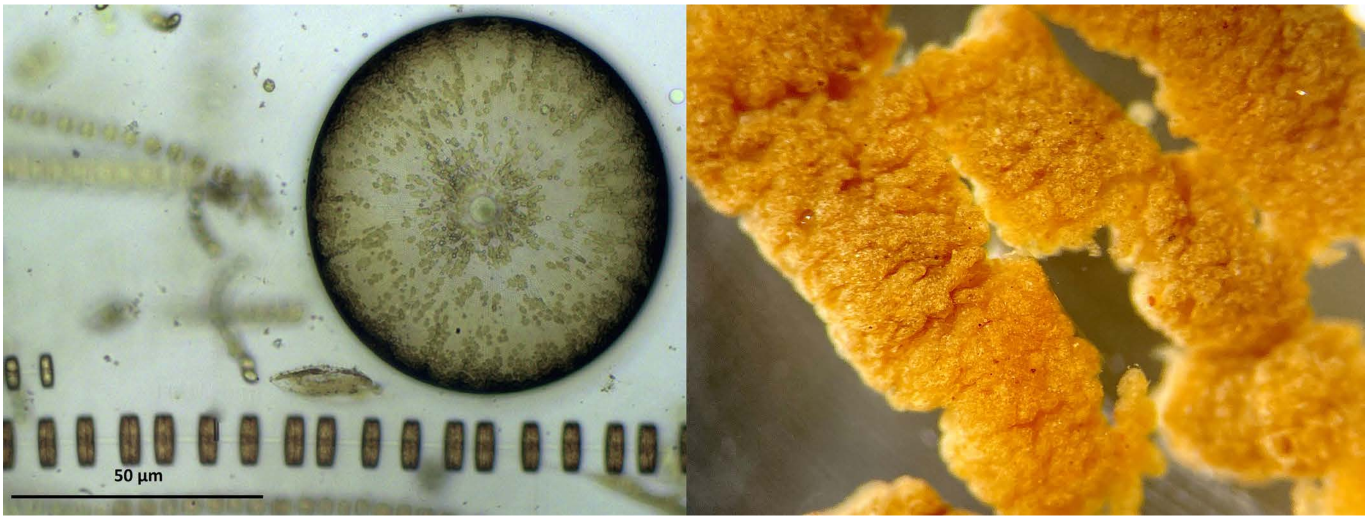



**Figure 1.3.28:** Left: the centric diatom *Coscinodiscus* sp. which is a large, lipid and carbohydrate-rich species that capitalises on peak nutrients during early spring. Image courtesy of Amanda Burson (British Antarctic Survey). Right: organic detritus produced by jellyfish from the subtropical South Atlantic, March 2023. Approximate width of pellets is 1.5mm. Image: Daniel Mayor on Instagram (accessed 2023).

A system that is more likely to show tipping-point behaviour is the seasonal lipid (fat) pump (SLP) (Jonasdottir et al., 2015). The SLP mainly occurs in high latitude oceans and is driven by the seasonal vertical migration of lipid-rich zooplankton (Figure 1.3.29) into the deep ocean, where they overwinter for ≥6 months, directly injecting carbon below the winter mixed layer.

A dramatic reduction in primary production via diatoms, for example, driven by changing nutrient supply patterns via increased stratification due to ocean warming, could result in zooplankton not consuming enough lipids to successfully overwinter and reproduce the following spring. Arresting the SLP would irreversibly change the ecological and biogeochemical functioning of high latitude ecosystems.

**Figure 1.3.29:** The marine copepod, *Calanus finmarchicus*, with its lipid sac outlined in red. Reproduced from (Mayor et al., (2020) and (Anderson et al., (2022).

Deep ocean warming will increase rates of respiration, meaning that lipid reserves may become exhausted before returning to the surface. This will interrupt recruitment and halt the SLP. The poleward migration of non-diapausing species (i.e. those that do not form an inactive life-form for parts of the year), as polar conditions ameliorate, could eventually mean that lipid-storing deep-diapausing zooplankton eventually disappear and the SLP collapse will be irreversible. However, the SLP was only described <10 years ago, and so our nascent understanding of its scale and complexity currently precludes the establishment of thresholds.

Other parts of the ocean biological pump could also result in nonlinear dynamics or tipping points. A recent paper found evidence that 'mixotrophs' – plankton that can both photosynthesise like algae and consume other plankton – can switch between a photosynthesis-dominant carbon sink state to a consumption-dominant carbon source state, with warming pushing them towards the latter and nutrient pollution making tipping dynamics more likely (Wieczynski et al., 2023). Mixotrophs are common in the ocean but their role in ocean and ecosystems and the biological pump is under-studied (Ward, 2019), making the impacts of these potential tipping dynamics unclear.

### Marine oxygenation

Coastal hypoxia is a regime shift that occurs when dissolved oxygen in water diminishes below levels detrimental to marine life. As a consequence, one of their symptoms are 'dead zones', areas of the oceans where fish and many other marine organisms (particularly in benthic communities) migrate outwards or die due to low oxygen levels.

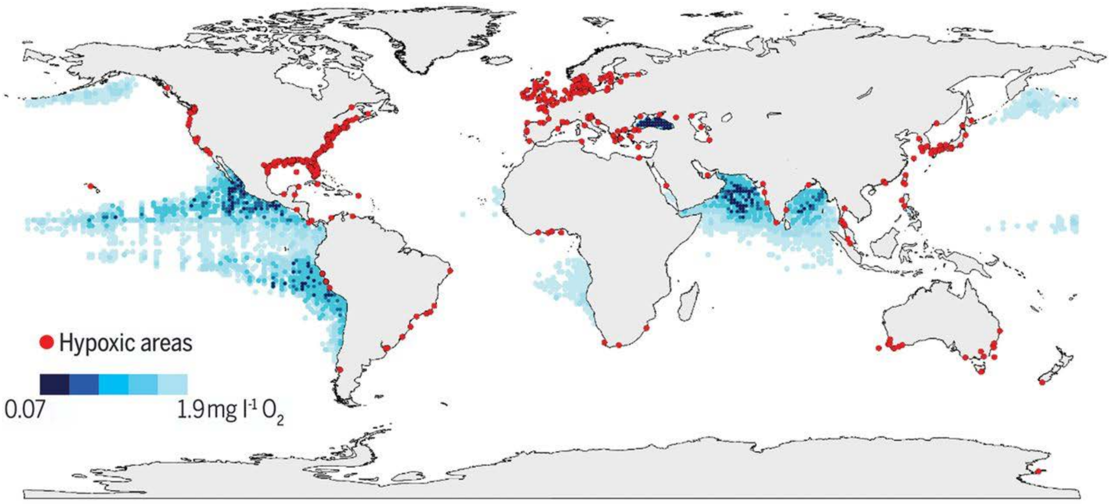



**Figure 1.3.30:** Map of known oceanic oxygen minimum zones (at 300m depth, blue) and coastal sites where anthropogenic nutrients have exacerbated or caused $O_2$ to decline to <2 mg/litre (red dots), becoming 'dead zones'. Source: (Breitburg et al., 2018).

While hypoxia is naturally occurring in some areas, hypoxic events have been increasing over the last few decades. A first global assessment of ocean deoxygenation documented over 300 cases mainly in the Atlantic coast of North America, the Caribbean, Mediterranean and Baltic seas (Diaz and Rosenberg, 2008). Subsequent assessments expanded to >500 case studies, from occasional hypoxic events to severe anoxia (Breitburg et al., 2018) (Figure 1.3.30).

The main mechanisms underlying coastal hypoxia are related to over-enrichment of nutrients like phosphorus and nitrogen coming from agricultural fertilisers, sewage or upwelling currents in the ocean. The latter are natural currents that bring nutrient-rich waters from the deep ocean to the surface, powering the primary producers (i.e. algae) and in turn productive food webs. In high-nutrient waters, algae can become over-abundant, consuming the available oxygen and causing the death of fish or other oxygen-dependent organisms. As they die, decomposers then further decrease available oxygen as they break down extra organic matter. Additional nutrients from fertiliser and sewage runoff on land is amplifying this process, increasing the number of hypoxic events and sites (Breitburg et al., 2018; Heinze et al., 2020).

At the same time, phosphorus can be released from sediment under low oxygen conditions, acting as a positive/amplifying feedback by further amplifying the growth of algae and the consumption of oxygen (Conley et al., 2002; Adhikari et al., 2015). Besides nutrients, climate change can exacerbate hypoxia by reducing oxygen solubility in water (Breitburg et al., 2018), and is projected to cause widespread deoxygenation over coming centuries to millennia via warming and enhanced land weathering delivering more phosphorus (Watson et al,.2017; Battaglia and Joos, 2018). Even if warming peaks and falls ('overshoot') Earth system models indicate that deoxygenation in the upper 1000 metres of the ocean is irreversible for multiple centuries (Santana-Falcon et al., 2023). Sea surface temperature can also change the strength of upwellings and thus the inflow of nutrients in coastal ecosystems.

Marine ecosystems with dissolved oxygen higher than >2 mL per litre sustain diverse ecological communities, and this level is considered normal (also known as 'normoxia'). Below this level the symptoms of hypoxia appear, including hypoxic events and dead zones. Anoxia occurs when levels of dissolved oxygen are below 0.5 mL per litre, which only a few microbial species are able to survive (Diaz and Rosenberg, 2008). Some dead zones and hypoxic events are reversible in scale of months to years. However, more and more areas are reported as chronically hypoxic, possibly irreversible in the timescale of ecosystem managers (centuries). Examples of severe hypoxia are dead zones in the Gulf of Mexico, central Baltic, Kattegat, Black Sea, and East China Sea (Breitburg et al., 2018).

### Assessment and knowledge gaps

**Table 1.3.3: Summary table of marine environment tipping points considered in this section.**

| System | Tipping system? | Timescale | Biophysical Impacts | Confidence | Gaps |
|---|---|---|---|---|---|
| **Fisheries [Small, fast-growing fish]** | No | Decades | Changes in entire trophic assemblage. A regime shift in one species could propagate the regime shift in many components of the ecosystem. Important especially in bottom-up and wasp-waist ecosystems. | Low confidence because too many different species | Need more coherent statistical approaches to identify tipping points and the presence of hysteresis. Also need more analyses on single species that look at tipping points in fisheries. |
| **Fisheries [Large, slow-growing fish]** | Depends on the stock | Decades | Changes in entire trophic assemblage. A regime shift in one species could propagate the regime shift in many components of the ecosystem. Especially important in top-down ecosystems. | Low confidence because too many different species and many different areas | Need more coherent statistical approaches to identify the tipping points and the presence of hysteresis. Also need more analyses on single species that look at tipping points in fisheries. |
| **Fisheries [Cod]** | Yes (in 16 out of 19 stocks) | Decades | Changes in the entire trophic assemblage, trophic cascade. | High confidence | In some cases there is the need to better understand feedbacks of hysteresis. |
| **Community shifts** | Yes | Decades | Changes in ecosystem function. structure and feedbacks that may affect how to best manage the system. | Low confidence - complexity from many different species and interacting drivers | Better understanding required on interplay of multiple drivers and species interactions. Tipping points difficult to identify and predict. |
| **Kelp forests** | Yes | Months to decades | Changes to community composition of fish and macroinvertebrates scaling up to trophic disassembly. | High confidence | Necessary to understand how key ecosystem properties, e.g. resilience or stability of kelp forests, evolve over the years. |
| **Ocean hypoxia** | Yes | Months/years to centuries. Reversible at surface, irreversible at depth for centuries to millennia | Major changes in ocean productivity, biodiversity and biogeochemical cycles. | Low confidence | Degree of self-sustaining change and hysteresis; influence of future climate change and nutrient use. |

Table 1.3.3 summarises our assessment of tipping dynamics (with confidence levels) along with biophysical impacts, timescales and knowledge gaps for marine ecosystems. We have high confidence that cod fisheries and kelp forests can pass tipping points, low confidence that some other large-fish fisheries, marine communities and potentially the lipid pump could also tip, and medium confidence that marine hypoxia could feature tipping dynamics. Knowledge gaps include limited understanding of complex species and driver interactions, limited ability to detect and project marine tipping points in practice, and how ecosystem resilience can change over time.

### 1.3.3 Final remarks

In this chapter we have assessed evidence for tipping dynamics across the biosphere, finding that many ecosystem tipping points are possible. Compared to tipping points in the cryosphere (Chapter 1.2) and ocean/atmosphere circulations (Chapter 1.4), biosphere tipping points tend to feature more co-drivers, including habitat degradation and loss, direct exploitation and nutrient pollution with often complex interactions (IPBES, 2019). Along with strong spatial variability, this often makes ecosystem tipping thresholds and risks more difficult to assess. However, these complexities also provide opportunities for action to avert tipping.

While climate change is a common leading driver, requiring urgent global emissions phaseout, compared to the cryosphere or ocean circulation it is more possible to directly increase the resilience of some at-risk systems. Actions such as ecological restoration and inclusive conservation, adaptive management and improved governance can help protect biodiversity and bio-abundance and so help to maintain key stabilising feedbacks that can help counter tipping (see Chapter 3.2). Such restoration and regenerative land use practices would also help to draw down some carbon from the atmosphere, helping to slow climate change (Girardin et al., 2021; Rockström et al., 2021). Such 'nature-based solutions' would not be enough to stop climate change though, which can only be achieved with a rapid cessation of greenhouse gas emissions.

Most ecosystems considered in this chapter can also be considered social-ecological systems, with people living within, and being integral to, the dynamics of these systems (Folke et al., 2016, 2021). While in some heavily degraded ecosystems restoration might entail minimising human impacts, in most places actions like supporting sustainable livelihoods for local communities can better help promote both ecological restoration and support human wellbeing in a way that makes both more sustainable in the long term (IPBES, 2019). The rights of Indigenous peoples – whose territories cover more biodiverse area globally than officially protected areas (ICCA Consortium, 2021) – must be respected, and their knowledge recognised as critical. Many other societal shifts are also necessary to underpin ecological restoration, including transformative changes to the global food system and commodity consumption (which together are key drivers behind much habitat loss and pollution – IPBES, 2019).

From a research perspective, we have identified several critical areas where improved knowledge could help us better understand biosphere tipping dynamics. In particular, deep uncertainties exist around the relative strength of feedbacks controlling ecosystem tipping dynamics, such as the complex interactions between ecohydrological and fire feedbacks in forest, savanna and dryland biomes. The role of increasing extreme event frequency and intensity in reducing and overcoming ecological resilience is also critical for ecosystems such as coral reefs, mangroves and forests, but it is not well resolved in models. Plant adaptability and spatial variability are also not well represented in models, despite being key factors adding complexity to ecosystem tipping dynamics. More observations, experiments and improved models, and integrations across these, are all required to address these issues.

Observations from field and remote sensing can also help monitor and detect declining ecosystem resilience, as well as potential early warning signals (see Chapter 1.6). Greater data sharing and international collaboration would improve both monitoring and understanding. Lastly, co-designing research with researchers from across the natural and social sciences, Global South and North, and from multiple knowledge systems including Indigenous and traditional ecological knowledge is critical for fully understanding ecological dynamics and the potential for tipping.

# 1.4 Tipping points in ocean and atmosphere circulations

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## Summary

This chapter assesses scientific evidence for tipping points across circulations in the ocean and atmosphere. The warming of oceans, modified wind patterns and increasing freshwater influx from melting ice hold the potential to disrupt established circulation patterns. We find evidence for tipping points in the Atlantic Meridional Overturning Circulation (AMOC), the North Atlantic Subpolar Gyre (SPG), and the Antarctic Overturning Circulation, which may collapse under warmer and 'fresher' (i.e. less salty) conditions.

A slowdown or collapse of these oceanic circulations would have far-reaching consequences for the rest of the climate system, such as shifts in the monsoons. There is evidence that this has happened in the past, having led to vastly different states of the Sahara following abrupt changes in the West African monsoon, which we also classify as a tipping system. Evidence about tipping of the monsoons over South America and Asia is limited, however large-scale deforestation or air pollution are considered as potential sources of destabilisation. Although theoretically possible, there is little indication for tipping points in tropical clouds or mid-latitude atmospheric circulations. Similarly, tipping towards a more extreme or persistent El Niño Southern Oscillation (ENSO) state is not sufficiently supported by models and observations.

While the thresholds for many of these systems are uncertain, tipping could be devastating for many millions of people. Stabilising climate (along with minimising other pressures, like aerosol pollution and ecosystem degradation) is critical for reducing the likelihood of reaching tipping points in the ocean-atmosphere system.

*The scientific content of this chapter is based on the following manuscript: Loriani et al., Tipping points in ocean and atmosphere circulations. Earth System Dynamics (submitted).*

## Key messages

- There is evidence for tipping points in the overturning circulations in the Atlantic and the Southern ocean, as well as for the West African monsoon.
- Short observational records, potential model biases towards stability, and limited resolution of various important feedback processes in models leave uncertainties, making an assessment of potential tipping difficult.

## Recommendations

- Prevent destabilisation of ocean and atmosphere circulations by urgent and ambitious reduction of greenhouse gas emissions and other pressures such as air pollution.
- Fill knowledge gaps and improve models to constrain projected impacts for the next decades and beyond. Reduce uncertainties. For example related to the resolution of small-scale processes and interaction of different systems.
- Invest in observations and palaeo reconstructions to detect early warning signs of tipping dynamics, and foster data sharing and international collaboration.

## 1.4.1 Introduction

The Earth's ocean and atmosphere form the flowing fluid parts of the Earth system that circulate around the planet. They drive the daily weather and climate patterns we see. On a global scale, the dominant circulations in the atmosphere are a consequence of regional differences in solar radiation (with poles less heated than the equator), Earth's rotation (redirecting winds) and thermodynamic properties (e.g. that warm air is less dense and rises).

Atmospheric circulation can be divided into several rotating cells: The 'Hadley cell' is formed either side of the equator by warm air rising near the equator (at the 'Intertropical Convergence Zone', or ITCZ) before sinking in both midlatitudes (at ~30° North or South). The midlatitude Ferrel cell sinks at mid latitudes and rises at high latitudes (~60° N or S), connecting to the polar cell rising at high latitudes and sinking at the poles. Diverted by Earth's rotation, surface winds tend to blow westwards (the 'easterly' trade winds) in the tropical cells, and eastwards ('westerlies') in the mid and high latitudes.

Over 70 per cent of the Earth's surface is covered by the global ocean, and is conventionally divided into the Atlantic, Indian, Pacific and Southern oceans. Ocean currents circulate water around the Earth as a result of pressure gradients driven by differences in temperature and salinity. This 'global thermohaline circulation', also known as the 'ocean conveyor belt', mixes the whole ocean over a roughly thousand-year timescale. Key components of this mechanism, connecting deep currents with those on the surface, are the sinking of cold and salty – therefore dense – water in polar regions as well as widespread 'upwelling'. The force exerted by atmospheric surface winds leads to basin-wide rotating 'gyres' of surface currents in the various ocean basins (Figure 1.4.1).

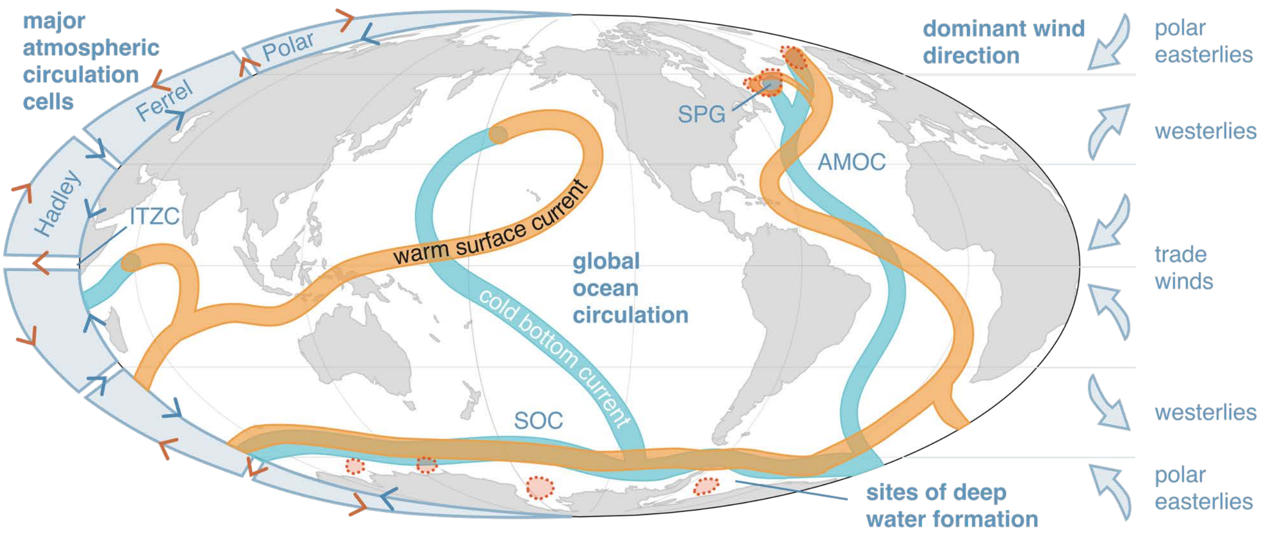


**Figure 1.4.1:** Atmospheric circulation cells, dominant wind directions, key ocean basins, surface currents and deep water formation sites. AMOC: Atlantic Meridional Overturning Circulation; SPG: Subpolar Gyre; SOC: Southern Ocean Circulation; ITCZ: Intertropical Convergence Zone.

Human-driven climate change is causing ongoing long-term changes in the ocean and atmosphere circulation. The effect of added greenhouse gases is to trap additional heat in the Earth system, driving atmospheric and ocean warming (with the latter accounting for more than 90 per cent of the heat trapped so far, (Fox-Kemper et al., 2021). There may also be changes in key circulation patterns, with increasing evidence that the Atlantic Meridional Overturning Circulation (AMOC) may be slowing (Dima and Lohmann 2010; Caesar et al., 2018; Rahmstorf et al,. 2015; Zhu et al., 2023). An extra seven per cent of water vapour can be held by the near-surface atmosphere with every degree of warming, leading to increasing precipitation in some regions (Zika et al, 2018). Evidence shows that heat extremes, heavy rainfall events and agricultural and ecological droughts are already increasing across every continent (IPCC 2021). As the ocean and atmosphere gradually warm, the range of natural variability around the baseline is shifting upwards, making formerly extreme events more common and formerly impossible events possible.

Evidence exists from geological records and model simulations that some of these circulation patterns could also feature tipping points, beyond which they may shift to a different state (Lenton et al,. 2008; Armstrong McKay et al. 2022; Wang et al,. 2023). Palaeorecords suggest deep water convection in the North Atlantic has abruptly shifted to a weaker or completely 'off' state during previous glacial cycles, with major climatic consequences – a pattern supported by some models (Böhm et al,. 2015; Iouville et al., 2021: Fox-Kemper et al., 2021). It has also been suggested that the Indian summer monsoon could shift to an alternative state as a result of aerosol emissions, counter to the general trend of monsoon strengthening with warming (Levermann et al. 2009; Doblas-Reyes et al., 2021), and as potential shifts in circulations in the southern hemisphere to El Niño-like mean conditions have also been proposed (Fedorov et al, 2006).

## 1.4.2 Current state of knowledge on ocean and atmosphere circulation tipping points

In this section, we assess available scientific literature on tipping points in ocean and atmosphere circulations. To this end, we focus on the following systems: ocean circulations in the Atlantic and the Southern Ocean; monsoons over West Africa, India and South America; tropical clouds and circulations; El Niño southern oscillation; and mid-latitude atmospheric circulations.

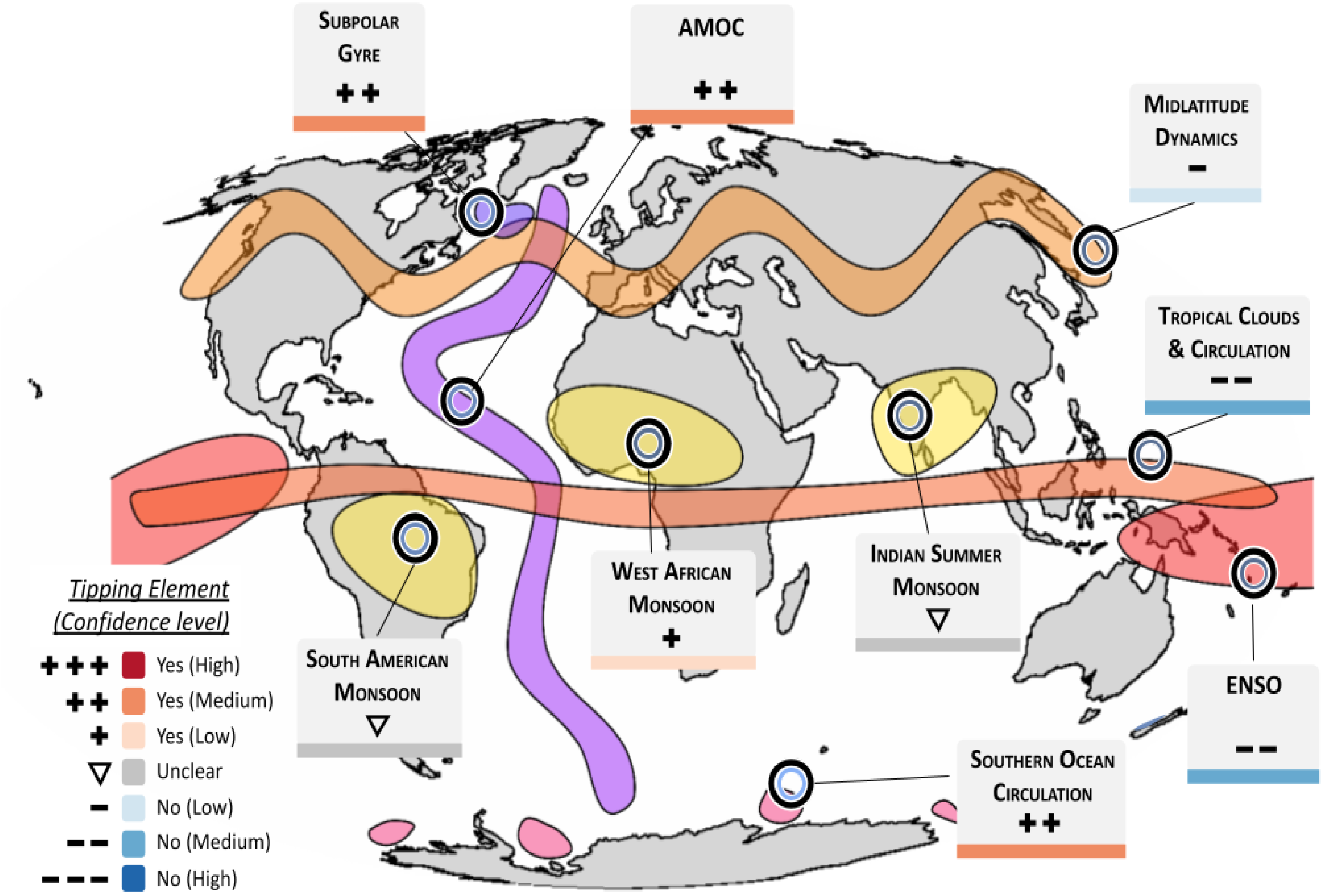


**Figure 1.4.2:** Potential tipping systems in ocean and atmosphere circulations considered in this chapter.

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 1.4.1: 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 (via direct impact of emissions on radiative forcing); **CA:** Climate-Associated driver (including second-order & related effects of climate change); **NC:** Non-Climate driver. Drivers can enhance **(↗)** the tipping process or counter it **(↘)**

| System | Key drivers | Key biophysical impacts *(see S2 for societal impacts)* | Key feedbacks | Abrupt / large rate change? | Critical threshold(s)? | | Tipping system? |
|---|---|---|---|---|---|---|---|
| **Ocean overturning circulation** | | | | | | | |
| Atlantic Meridional Overturning Circulation (AMOC)<br>Shutdown/collapse | • **DC: ocean warming (↗)**<br>• **DC: precipitation increase (↗)**<br>• **CA: Greenland ice sheet meltwater increase (↗)**<br>• **CA: Arctic river discharge increase (↗)**<br>• CA: sea ice extent & thickness decrease (↗)<br>• DC: regional aerosol forcing increase (↘)<br>• CA: regional ocean circulation changes (?) | • Cooling over Northern Hemisphere (up to 10°C over W/N Europe)<br>• Change in precipitation and weather patterns over Europe<br>• Change in location and strength of rainfall in all tropical regions<br>• Reduced efficiency of global carbon sink, and ocean acidification<br>• Reduced support for primary production in Atlantic oceans<br>• Deoxygenation in the North Atlantic<br>• Change in sea level in the North Atlantic<br>• Modification of sea ice and arctic permafrost distribution<br>• Change in winter storminess<br>• Reduced land productivity in Atlantic bordering regions<br>• Increased wetland in some tropical areas and associated methane emission<br>• Change in rainforest response in drying regions | • **Salt-advection (↗)**<br>• Sea ice melting (↗)<br>• Heat transport (↘)<br>• Temperature (↗)<br>• Surface heat flux (↗)<br>• Collapse of convection in the Labrador and Irminger Seas (↗) | Feedback-dependent:<br>Century (basin-wide salt advection feedback),<br>Few decades (North Atlantic salt-advection feedback),<br>< few decades (sudden increase in seaice cover in all convective regions) | **Salinity change/ freshwater/AMOC strength**<br><br>Thresholds likely path-dependent (depending on rate and spatial pattern) | ++<br>(centuries) | **++** |
| North Atlantic Subpolar Gyre (SPG)<br>Collapse | | • Increase in summer heat waves frequency<br>• Collapse of the North Atlantic spring bloom and the Atlantic marine primary productivity<br>• Increase in regional ocean acidification<br>• Regional long-term oxygen decline<br>• Impact on marine ecosystems in the tropics and subtropics | | Years to few decades | **Salinity change/ freshwater**<br><br>Global warming<br>1.1-3.8°C | ++<br>(decades) | **++** |

| System | Key drivers | Key biophysical impacts<br>***(see S2 for societal impacts)*** | Key feedbacks | Abrupt / large rate change? | Critical threshold(s)? | | Tipping system? |
|---|---|---|---|---|---|---|---|
| **Southern Ocean circulation**<br>Antarctic Overturning Collapse / Rapid continental shelf warming | • **DC: ocean warming (↗)**<br>• **CA: Antarctic ice sheet meltwater increase (↗)**<br>• CA: wind trends (↗)<br>• CA: Sea ice formation (↗)<br>• DC: precipitation increase (↗) | • Modification of Earth's global energy balance, timing of reaching 2°C global warming<br>• Reduced efficiency of global carbon sink<br>• Change in global heat storage<br>• Reduced support for primary production in world's oceans<br>• Drying of Southern Hemisphere<br>• Wetting of Northern Hemisphere<br>• Modification of regional albedo, shelf water temperatures<br>• Potential feedback to further ice shelf melt | • Density-stratification (↗)<br>• Meltwater-warming (↗) | ++ (AABW formation & abyssal overturning shutdown within decades) | Salinity change/ freshwater | ++<br>(cavity warming reversion would need 20th-century | **++** |
| **Atmosphere: Monsoons** | | | | | | | |
| **Indian summer monsoon (ISM)**<br>Collapse / Shift to low-precipitation state | • NC: increased summer insolation (↘)<br>• DC: increased water vapour in atmosphere (↘)<br>• CA: Indian Ocean Dipole events (?)<br>• CA: ENSO change (?)<br>• CA: North Atlantic cold SST (↗)<br>• DC: aerosol loading (↗)<br>• CA: Indian Ocean warming (↗)<br>• CA: low cloud reduction (↘) | • **Massive change in precipitation**<br>• Change in tropical and subtropical climates<br>• Biodiversity loss and ecosystem degradation | • Moisture-advection (↘) | Decades to centuries | Regional AOD level over Indian subcontinent (>0.25)<br>Interhemispheric AOD difference (>0.15)<br>AMOC slowdown (unknown threshold) | Uncertain; likely decades to centuries | **unknown** |
| **West African monsoon (WAM)**<br>Collapse or abrupt strengthening | • DC: increased water vapour in atmosphere (↗)<br>• NC: increased summer insolation (↘)<br>• NC: land-cover change (↗)<br>• CA: desertification (↗)<br>• CA: AMOC slowdown (↗)<br>• CA: regional SST variations (?)<br>• CA: High latitude cooling (↗)<br>• CA/NC: regional soil moisture variation (?)<br>• CA/NC: regional vegetation variation (?)<br>• NC: dust emissions (?) | • **Massive change in precipitation**<br>• Change in tropical and subtropical climates<br>• Biodiversity loss and ecosystem degradation | • Vegetation-albedo (↗) | Decades to centuries | Insolation changes in the Northern Hemisphere summers and surface albedo changes (unknown threshold)<br>Interhemispheric asymmetry in AOD (>0.15)<br>AMOC slowdown (unknown threshold) | Decades to centuries | **+** |

| **System** | **Key drivers** | **Key biophysical impacts** *(see S2 for societal impacts)* | **Key feedbacks** | **Abrupt / large rate change?** | **Critical threshold(s)?** | | **Tipping system?** |
|---|---|---|---|---|---|---|---|
| South American Monsoon (SAM) | • DC: increased water vapour in atmosphere (↗)<br>• NA: increased summer insolation (↘)<br>• CA: AMOC slowdown (↗)<br>• NC: Amazon deforestation (↗) | • **Massive change in precipitation**<br>• Change in tropical and subtropical climates<br>• Biodiversity loss and ecosystem degradation | • Vegetation-moisture (?) | Decades | Interhemispheric asymmetry in AOD (>0.15)<br>Extent of Amazon deforestation (30-50%)<br>AMOC slowdown (unknown threshold) | Uncertain; likely decades to centuries | **unknown** |

### 1.4.2.1 Atlantic circulation

### Atlantic Meridional Overturning Circulation (AMOC)

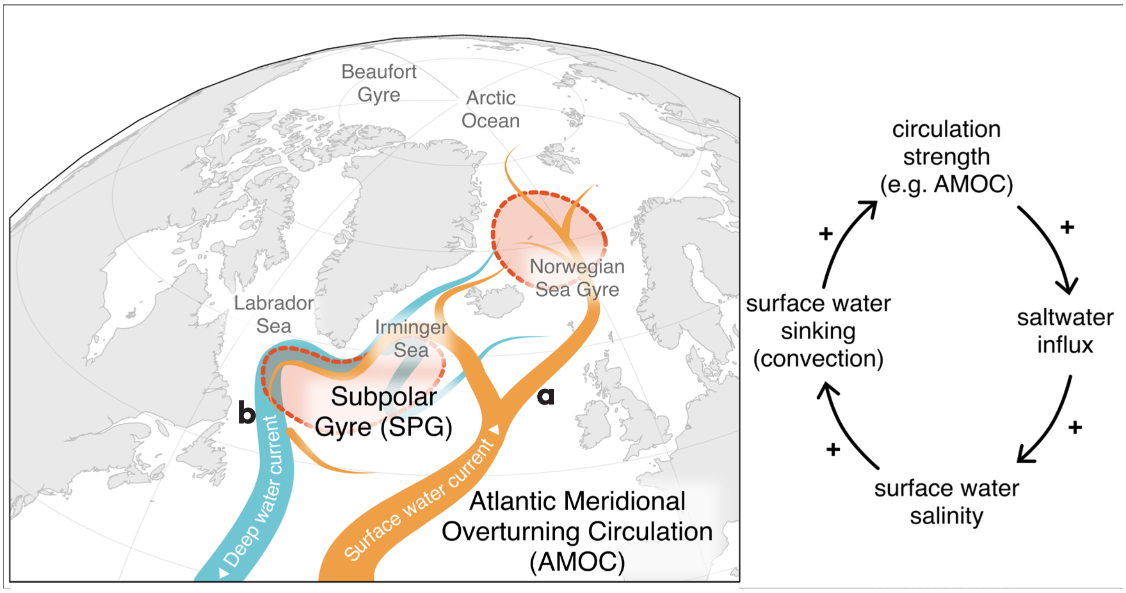


**Figure 1.4.3:** Overview over the major oceanic circulation systems in the North Atlantic. **a** The surface currents (orange pathways) are connected to deep ocean currents (blue) through sites where dense (cold, salty) water sinks, driving the overturning circulation (pink shading). **b** One critical feedback is the salt-advection feedback, in which the circulation strength determines how well the convection works, which in turn benefits the circulation.

**The Atlantic Meridional Overturning Circulation**

The Atlantic Meridional Overturning Circulation (AMOC) refers to a three-dimensional circulation present in the Atlantic (Figure 1.4.3a) whereby warm upper ocean waters ('upper branch') move northward from the tip of Southern Africa to the northern North Atlantic, where they cool, sink and return southwards as cold deep waters ('lower branch'). The AMOC moves heat from the South to the North Atlantic, helping to maintain the mild climate of western and northern Europe. Thereby it shapes the climate of the whole Earth, influencing, for example, the 1-2°C temperature difference between the Northern and Southern hemispheres, and the location and strength of rainfall across all tropical regions (Buckley and Marshall, 2016; Feulner et al., 2013; Marshall et al., 2014).

Fresh, warm water is less dense than cold, salty water. In the future, surface waters in the northern North Atlantic may become less dense. This will make it harder for the water in that region to sink, which will disrupt the connection between the upper and lower branches of the AMOC, causing it to weaken significantly or even collapse completely. Therefore, we need to monitor the processes which tend to warm and freshen the upper ocean at high latitudes. AMOC strength has only been observed directly since 2004 (Srokosz and Bryden, 2013), with more uncertain reconstructions based on observations such as surface temperature, which extend back in time before 2004 ('observational proxies'), or from palaeoclimate archives such as ocean sediment cores which extend back to prehistoric times ('palaeoclimate proxies') (Caesar et al,. 2018, 2021; Moffa-Sánchez et al., 2019). The lack of a sufficiently long observational record is a major issue for robust understanding of the AMOC.

The North Atlantic Ocean is freshening at subpolar latitudes (50-65°N), most strongly in the upper 100m, and warming, most strongly between 100-500 m water depth (IPCC, 2021). Both trends act to reduce AMOC strength. Greenland Ice Sheet melt is accelerating and releasing extra fresh water into the North Atlantic (Shepherd et al., 2020). In addition, Arctic sea ice is reducing in surface extent and thickness (Serreze and Meier, 2019) and overall Arctic river discharge is increasing (Druckenmiller et al., 2021), adding fresh water to the Arctic continental shelves and the high Arctic, and this riverine fresh water is potentially leaking into the North Atlantic from the Arctic. The North Atlantic is a region of high variability on interannual to decadal timescales (Boer 2000) and therefore subject to substantial climatic perturbations with the potential to trigger any underlying instability if a tipping point is approached.

Limited direct observations of AMOC strength make current trends uncertain, but there are some signs of ongoing weakening. Observational and palaeoclimate proxies suggest the AMOC may have weakened by around 15 per cent over the past 50 years (Caesar et al,. 2018) and may be at its weakest in 1,000 years (Caesar et al., 2021). However, the proxy data used in these studies have large uncertainties, and some other reconstructions show little evidence of decline (Moffa-Sanchez et al., 2019; Kilbourne et al., 2022). It is therefore difficult to confidently discern potential recent trends from natural variability, due to disagreement between published studies (Bonnet et al,. 2021; Latif et al., 2022; versus Qasmi, 2022).

The IPCC's most recent assessment is that the AMOC has weakened relative to 1850-1900, but with low confidence due to disagreement among reconstructions (Moffa-Sanchez et al., 2019; Kilbourne et al., 2022) and models (Fox-Kemper et al., 2021). For the future, the IPCC projects that it is very likely that the AMOC will decline in the 21st century (however with low confidence on timings and magnitude) (Figure 1.4.4a).

There is medium confidence (about 5 on a scale of 1 to 10) that a collapse would not happen before 2100, though a collapse is judged to be as likely as not by 2300. Hence the possibility of an AMOC collapse within the next century is very much left open by the latest IPCC report.

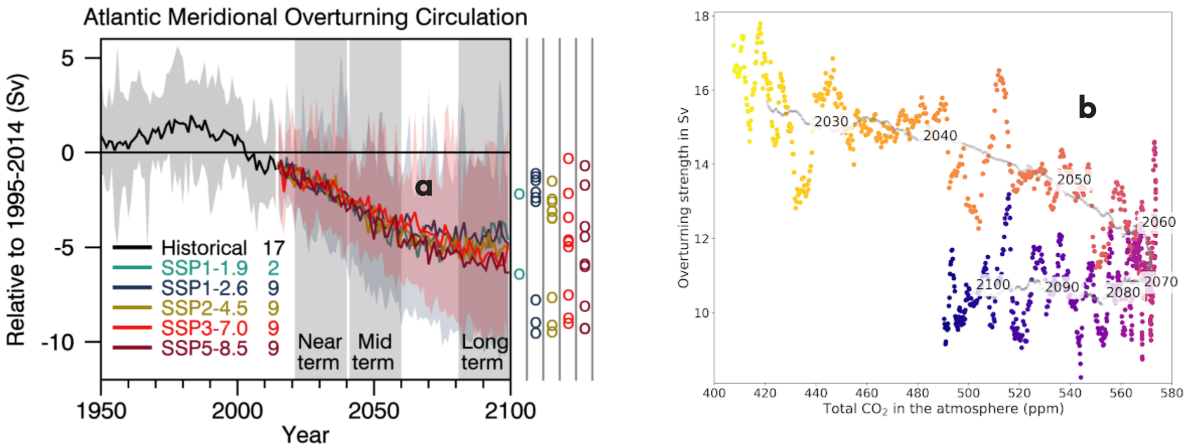


**Figure 1.4.4:** AMOC in CMIP models. **a** CMIP6 models showing gradual weakening of the AMOC during the 21st Century under all emission scenarios. Credit: Lee et al., 2021 l . **b** CMIP6 overshoot experiments (using UKESM; Jones et al. 2020) showing hysteresis - different states of the AMOC (vertical axis) for the same atmospheric CO² concentration (horizontal axis). Possible causes are delayed or nonlinear response to forcing or possibly bistability of AMOC. The AMOC strength is measured in 'Sverdrups' (Sv; i.e. a flow of 1 million cubic metres per second); colours from yellow to blue show model years from 2015 to 2100 respectively.

### Evidence for tipping dynamics

The AMOC has been proposed as a 'global core' tipping system of the climate system with medium confidence by Armstrong McKay et al., (2022). Palaeorecords indicate it has abruptly switched between stronger and weaker modes during recent glacial cycles (Figure 1.4.5).

Most of the time (including the warm Holocene of the past 12,000 years) the AMOC is in a strong, warm mode, but during peak glacials it sometimes shifted to a weak, cold mode instead (Böhm et al., 2014). It also occasionally collapsed entirely to an 'off' mode during 'Heinrich' events, in which iceberg outbursts from the North American Laurentide Ice Sheet temporarily blocked Atlantic overturning for several centuries.

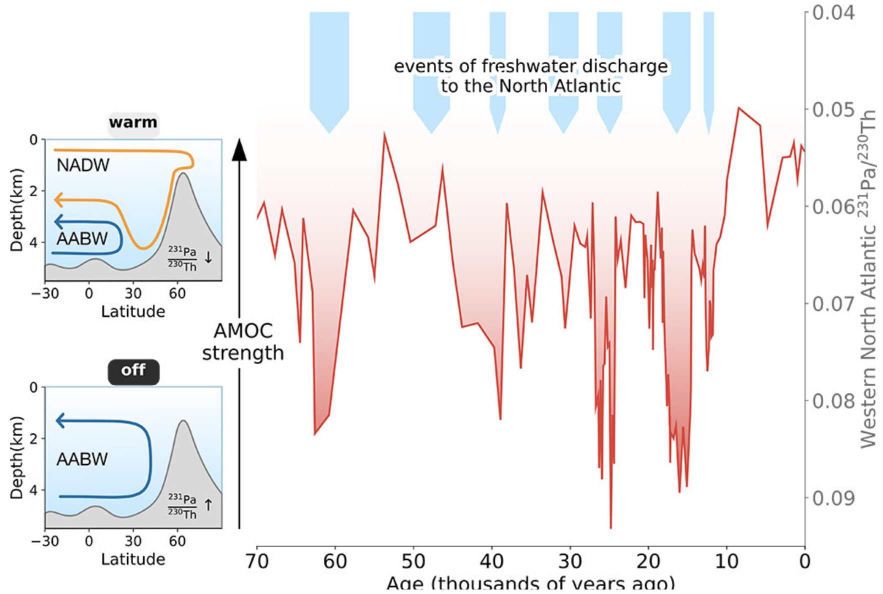


**Figure 1.4.5:** Different AMOC modes and palaeo-evidence.The diagrams on the left show two AMOC modes as indicated by sedimentary $^{231}Pa/^{230}Th$ in palaeorecords. NADW: North Atlantic Deep Water; AABW: Antarctic Bottom Water, adapted from Böhm et al., 2015. The timeline shows AMOC slowdown events during the last 70,000 years as recorded by sedimentary $^{231}Pa/^{230}Th$ data (McManus et al., 2004; Böhm et al., 2015) from the Western North Atlantic (Bermuda Rise, ca. 34°N). Sedimentary $^{231}Pa/^{230}Th$ from the Bermuda Rise is a proxy for AMOC strength that assesses the southward flowing North Atlantic Deep Water between ca. 3,500 and 4,500m water depth. The top of the panel marks the timing of past major events of freshwater discharge to the high latitudes of the North Atlantic that decreased AMOC strength (Sarnthein et al., 2001; Carlson et al., 2013; Sanchez Goñi and Harrison, 2010). The red shading highlights past AMOC slowdown events. There is also evidence of AMOC shifts during the last interglacial period, 116,000-128,000 years ago (Galaasen et al., 2014).

In two previous censuses of climate model projections, a shut-down of the AMOC over many decades was observed in a small minority of simulations (Drijfhout et al., 2015; Sgubin et al., 2017). This shut-down was preceded by decreases in subpolar surface air and ocean temperature and increased sea ice cover. Ultimately, deep mixing ceased to occur, destroying the connection between the surface and the deep ocean. There are, however, concerns that the AMOC may be too stable in CMIP-type climate models (Mecking et al., 2017; Liu et al., 2017), which suggests the CMIP multimodel ensembles may underestimate the likelihood of AMOC collapse ( Fox-Kemper et al., 2021).

Some recent studies have suggested that 'early warning signals' indicating destabilisation (see Chapter 1.6) can be detected in reconstructed 'fingerprints' of AMOC strength over the 20th Century (Boers, 2021), and if a tipping point is assumed then the collapse threshold could be reached during the 21st Century (Ditlevsen & Ditlevsen, 2023). These studies used observational proxies for temperature and salinity from the Northeastern subpolar North Atlantic, which are used as indirect AMOC fingerprints rather than direct measurements of AMOC strength. This gives long enough data to analyse for early warning signals, but using indirect proxies adds uncertainty. The model used by Ditlevsen & Ditlevsen (2023) to project collapse is also highly simplified with a tipping point assumed, and does not take into account the low-frequency variability of the AMOC, nor the presence of external forcings such as increasing greenhouse gases. So while signals in this dataset are consistent with approaching a tipping point, there are substantial uncertainties with this methodology (see also Michel et al., 2023 highlighting potential false warnings). Further potential early warning signals have been found from analysis of Northern Hemisphere palaeoproxies (Michel et al., 2022). Despite the caveats mentioned above, these results amount to a serious warning that the AMOC might be en route to tipping. However, the claim that we might expect tipping in a few decades is – in the view of the present authors – not substantiated enough.

AMOC stability is strongly linked to the 'salt-advection feedback' (Stommel, 1961, see Figure 1.4.3b). The AMOC imports salt into the Atlantic and transports it from the South Atlantic to the northern North Atlantic. If the AMOC weakens then less salt is transported to the northern North Atlantic, the surface waters freshen, which inhibits sinking, and the AMOC weakens further. The AMOC collapses seen in models (Drijfhout et al., 2015; Sgubin et al., 2017) were driven by this salt-advection feedback. However, the strength of this feedback, and the timescale over which it operates are governed by processes whose effects are quite uncertain. Although Figure 1.4.3a shows typical pathways of surface and deep water through the Atlantic, these are an average picture over many decades. Individual water parcels may get caught up in basin-scale surface or deep recirculations, smaller-scale eddies and meandering currents. There is no definitive evidence though from models or observations that these systematically impact the salt advection feedback.

Additionally, changes in the AMOC have other impacts on salinity – for instance through affecting evaporation and precipitation patterns (Jackson., 2013; Weijer et al., 2019). These other feedbacks can temporarily mask, and may even overcome, the salt advection feedback, potentially changing the stability of the AMOC (Jackson, 2013; Gent, 2018). It is difficult to characterise these processes and feedbacks from observations alone due to insufficient data coverage both in time and space, so we are dependent on numerical models. However, many studies have used reduced complexity models, which may not capture all the potential feedbacks, and even the current generation of climate models have quite low spatial resolution and do not well characterise narrow currents, eddies and processes such as horizontal and vertical mixing (Swingedouw et al,. 2022).

Armstrong McKay et al. (2022) estimated with low confidence a global warming threshold for AMOC collapse of ~4°C (1.4-8°C). In our view, the range is a better indication of the uncertainty in the different model responses rather than a relationship to global warming, as the likelihood is probably less dependent on temperature, but strongly depends on salinity changes and the strength of opposing feedbacks on the freshwater budget. Studies with climate models have found that adding freshwater can cause the AMOC to collapse and not recover in some models. Since many climate models might be biased towards stability, however, these studies use an unphysically large amount of freshwater to explore the sensitivity (Jackson et al., 2023). Although adding freshwater causes a collapse, they show the threshold is dependent on the strength of the AMOC and deep convection, rather than on the amount of freshwater added (Jackson and Wood, 2018; Jackson et al, 2023). AMOC collapse may also be more sensitive to the rate of freshwater forcing than the total magnitude (Lohmann & Ditlevsen, 2021).

Hysteresis and bistability both refer to systems which can adopt one of two or more states for the same external forcing, such as $CO_2$ concentration (see Figure 1.4.4b and Glossary; Boucher et al., 2012). Commonly, this is explored by approaching the same external conditions with different trajectories in model simulations, e.g. increasing and reversing the forcing to study reversibility. Bistability involving a full collapse of the AMOC by artificially flooding the North Atlantic with freshwater has been demonstrated (or strongly implied) in theoretical models (Stommel, 1961) and climate models of reduced complexity (Rahmstorf et al., 2005; Hawkins et al., 2011). These types of numerical experiments study bistability through forcings that change slowly enough for the system to equilibrate, typically requiring long simulations and thus coarse model resolution for reasonable computational performance. In more complex models it is not possible to conduct experiments for long enough to demonstrate bistability or hysteresis, however weak states have been shown to be stable for at least 100 years in about half of a test group of CMIP6-type models (Jackson et al., 2023) and in a high-resolution ocean-atmosphere coupled climate model (Mecking et al., 2016). A recent study finds AMOC tipping in a CMIP-type model in response to gradually increasing freshwater release in the North Atlantic (Van Westen et al., 2023). AMOC bistability is model-dependent though, controlled by the balance of the positive and negative feedbacks that determine the salinity of the subpolar North Atlantic. It is not yet understood why the bistability occurs in some models and not others (Jackson et al., 2023). However, as previously mentioned, there is evidence that the present generation of climate models is too stable due to model biases in the distribution of ocean salinity (Liu et al., 2017; Mecking et al., 2017).

Not only is it difficult to prove system bistability, the complexity of the system and interaction with multiple drivers make it hard to assess collapse thresholds. It may be that realistic freshwater input is not sufficient to cause the transition, or that changing $CO^2$ alters the underlying system stability, thus increasing the critical freshwater threshold (Wood et al., 2019). Nevertheless, overshoot scenarios, where the $CO_2$ trend is assumed to reverse at some point in the future, provide some useful information about reversibility of the AMOC on human timescales. Figure 1.4.4b shows how the AMOC changes in the UKESM climate model under an overshoot emission scenario exceeding and returning to 500 ppm. Even if $CO^2$ concentrations return to 500 ppm by 2100 the AMOC is still only 77 per cent of the strength it was in 2050 also at 500 ppm. Although the AMOC does not collapse in this model, it seems unlikely that it will recover its former strength on human timescales.

The timescale of AMOC tipping was estimated by Armstrong McKay et al., (2022) to be 15-300 years, however this range is very dependent on the strength of the freshwater forcing applied in experiments, which in many cases is unrealistically large as compared to projected melting of the Greenland Ice Sheet and increase in precipitation and river runoff. Moreover, the assessment is also potentially impacted by the models being unrealistically stable. With a realistic forcing scenario, the timescale will depend on the feedbacks. A basin-wide salt advection feedback may have a century timescale, while if it is preceded by a local North Atlantic salt advection feedback it may be reduced to a few decades. Even faster timescales are possible when deep mixing is capped off by sudden increases in sea ice cover in all convective regions (Rahmstorf et al., 2001; Kuhlbrodt et al., 2001).

AMOC collapse would lead to cooling over most of the Northern Hemisphere, particularly strong (up to 10°C relative to preindustrial) over Western and Northern Europe. In addition, a southward shift of the Intertropical Convergence Zone would occur, impacting monsoon systems globally and causing large changes in storminess and rainfall patterns (Jackson et al., 2015). A collapse of the AMOC would influence sea level rise along the boundaries of the North Atlantic, modify Arctic sea ice and permafrost distribution (Schwinger et al., 2022; Bulgin et al., 2023), reduce oceanic carbon uptake (Rhein et al., 2017) and potentially lead to ocean deoxygenation (Kwiatkowski et al., 2020) and severe disruption of marine ecosystems (including changes in the North Atlantic Subpolar Gyre, see below), impacting North Atlantic fish stocks. See Chapter 2.2 for more discussion on impacts.

**Assessment and knowledge gaps**

Although the AMOC does not always behave like a tipping system in many ocean/climate models, palaeoceanographic evidence strongly points to its capability for tipping or at least to shift to another state that can be quasi-stable for many centuries (Figure 1.4.5). Tipping is also suggested in a recent study of several CMIP6 models (Jackson et al., 2023) and in another study which found that removing model salinity biases strongly increased the likelihood of tipping (Liu et al., 2017). This does not necessarily mean that tipping is likely in a future climate, since some of these scenarios specified unrealistic inputs of freshwater or GHG emissions. Nonetheless, although the likelihood for collapse is considered small compared to the likelihood of AMOC decline, the potential impacts of AMOC tipping make it an important risk to consider in framing mitigation targets, for instance.

The latest AR6 assessment states that we have only medium confidence that an AMOC collapse will not happen before 2100 (Fox-Kemper et al., 2021). This uncertainty is due to models having strong ocean salinity biases, absence of meltwater release from the Greenland Ice Sheet in climate change scenarios, and the possible impact of eddies and other unresolved ocean processes on freshwater pathways. However, a recent study with the PAGES2K database of climate reconstructions of the past 2,000 years suggests, using statistical methods based on dynamical systems theory, that we may be close to an AMOC tipping point (Michel et al., 2022), as do the studies of Boers (2021) and Ditlevsen and Ditlevsen (2023) cited above. AR6 also concluded that reported recent weakening in both historical model simulations and observation-based reconstructions of the AMOC have low confidence. Direct AMOC observations have not been made for long enough to separate a long-term weakening from short-term variability. Another recent study suggests that we will need to wait until at least 2028 to obtain a robust statistical signal of AMOC weakening (Lobelle et al,. 2020). Thus, the coming years will be crucial for detection of an AMOC weakening potentially leading to longer-term instability.

There are substantial uncertainties around how the AMOC evolves over long timescales, because of a lack of direct observations. More palaeo-reconstructions of AMOC strength, ocean surface temperature, and other AMOC-related properties with high temporal resolution, using appropriate proxies and careful chronological control performed for key past periods (e.g. last millennium, millennial-scale climate change events, previous interglacials), hold great potential to improve our understanding about the AMOC as a tipping point. Other open issues are to: (i) reconcile disagreements between palaeo-reconstructions and model simulations, and (ii) develop improved metrics for creating historical reconstructions and monitoring the AMOC.

Current climate models suffer from imperfect representation of some important processes (such as eddies and mixing) and from biases which can impact the AMOC response to forcings. Hence we need to assess how important these issues are for representing AMOC stability, in particular, to understand how different feedbacks vary across models and are affected by modelling deficiencies. Given these issues, a robust assessment of the likelihood of an AMOC collapse is difficult, but based on the evidence presented, we assess that the AMOC features tipping dynamics with medium confidence. One potential way forward, given these uncertainties, is in developing observable precursors to a collapse that could be monitored.

### North Atlantic Subpolar Gyre (SPG)

The North Atlantic Subpolar Gyre (SPG) is an oceanic cyclonic (counter-clockwise in the northern hemisphere) flow to the south of Greenland (Figure 1.4.6). It is linked to a site of deep ocean convection in the Labrador-Irminger Seas, i.e. sinking of the subsurface ocean waters to great depths, contributing to the AMOC (Figures 1.4.3, 1.4.6-7).

There are indications for change in the SPG, as observations show that Labrador Sea Water (LSW) formed during oceanic deep convection events after 2014 was less dense than the LSW formed between 1987 and 1994 (Yashayaev and Loder, 2016), potentially influencing the AMOC. Moreover, the observed 'warming hole' over the North Atlantic can be explained by AMOC slowdown (Drijfhout et al., 2012; Caesar et al., 2018; also see AMOC above) and has also been linked to SPG weakening in CMIP6 models (Sgubin et al., 2017; Swingedouw et al., 2021). In these models, a collapse of the oceanic convection causes a localised North Atlantic regional surface air temperature drop of ~2-3°C. This cooling moderates warming over north-west Europe and eastern Canada in global warming scenarios, although it is smaller and less widespread than that associated with AMOC collapse.

A northward-shift of the atmospheric jet stream, which is predicted to take place with SPG weakening, means more weather extremes in Europe (which may be linked to the unusual cooling and heat waves in recent years) (Osman et al., 2021) and southward shift of the intertropical convergence zone (ITCZ, see Figure 1.4.1) (Sgubin et al,. 2017; Swingedouw et al,. 2021). These changes in the physical system may trigger changes in ecosystems with detrimental consequences for the North Atlantic spring bloom and overall Atlantic marine primary productivity. Neither of these return to the preindustrial state even if emissions reverse by 2100 in models for clarity (Yool et al., 2015; Heinze et al., 2023). This would impose a strong impact on fisheries and biodiversity, with wide societal implications (see Section 2). Last but not least, a transition between two SPG stable states has been suggested to explain the onset of the so-called 'Little Ice Age' in which colder conditions prevailed in Europe during the 16th-19th centuries (Lehner et al., 2013; Michel et al., 2022).

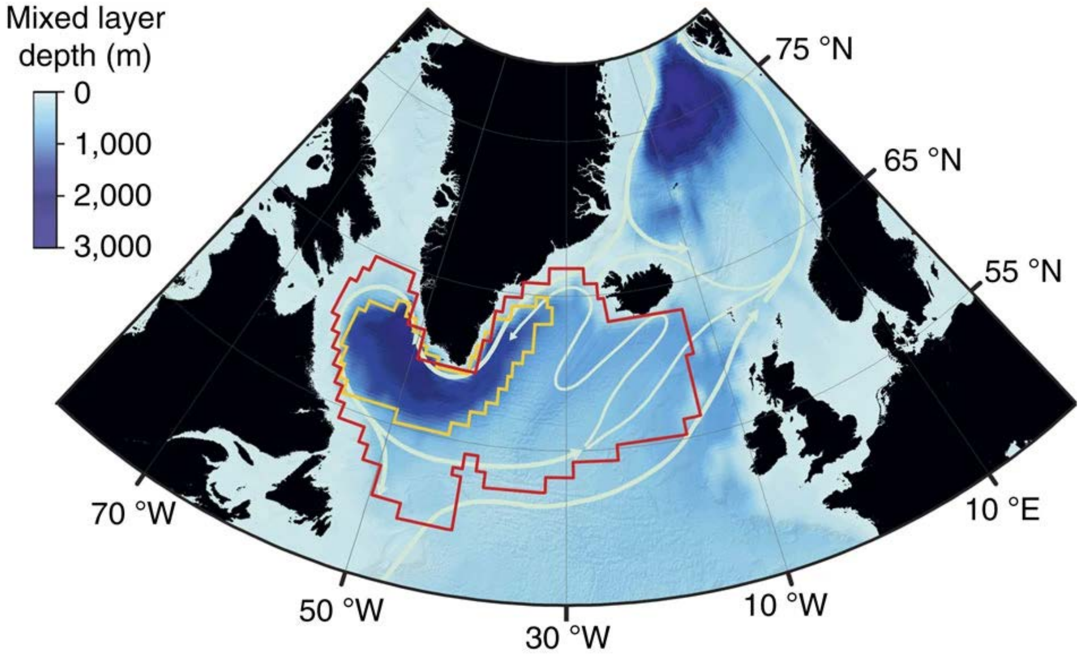


**Figure 1.4.6: Map showing the maximum ocean mixing depth in the North Atlantic** (light to dark blue), showing deep water convection sites driving the AMOC and SPG east and south of Greenland respectively (with the Labrador-Irminger Seas convection area bordered by yellow). The pale arrows show surface water currents, with the anti-clockwise subpolar gyre occurring within the red bordered area. Credit: Sgubin et al,. (2017).

Ventilation of LSW is accompanied by an uptake of oxygen. Starting in 2014, the convection in the Labrador Sea became more intense and reached depths of 1,500m and below. Consequently, oxygen in LSW is in general increased, but this increase did not penetrate the densest part of this water mass (Rhein et al., 2017). The oxygen concentrations in the deepest part of the LSW (around 2,000m) have decreased in the formation region and along the main export pathways (southward and eastward crossing the Mid-Atlantic Ridge) for more than 20 years. Most of the oxygen from the export of newly formed LSW has been consumed north of the equator (Koelling et al., 2022), and the long-term oxygen decline along the southward LSW pathway might have impacts on ecosystems in the tropics and subtropics over longer timescales (e.g., Heinze et al., 2023).

The potential shutting-down of winter convection in the Labrador Sea (see Figure 1.4.7a,b and Swingedouw et al., 2021) will also stop the production of Labrador Slope Water (LSLW). This water is next to the Labrador Sea continental slope and is lighter and less deep than LSW. It contributes to AMOC and the Gulf Stream and can influence variability of the Atlantic climate system overall (New et al., 2021). The LSLW is rich in nutrients and oxygen too, thereby affecting the ecosystems on the North American continental shelf and shelf slope (e.g. Claret et al., 2018) and might affect tropical and subtropical marine ecosystems on a timescale of several decades. Furthermore, the SPG takes up large amounts of atmospheric carbon and exports it to the deep ocean (Henson et al., 2022).

Shallowing of the SPG (Sgubin et al., 2017; Swingedouw et al., 2021) would directly increase regional $CO_2$ uptake but negatively impact marine biology, for instance threatening the habitat of cold-water corals in the area due to higher acidity with more $CO_2$ dissolved in the water (Fröb et al., 2019; Fontela et al., 2020; García-Ibáñez et al., 2021). Weakening or collapse of the SPG would reduce the amount of carbon-depleted intermediate water being upwelled and newly carbon-enriched water being convected, reducing export of anthropogenic $CO^2$ to the deep ocean (Halloran et al. 2015; Ridge & McKinley 2021), which in turn might lead to an increase of atmospheric $CO^2$ concentration in the long term (Schmittner et al. 2007). Declining SPG strength may also be reducing the currently high phytoplankton productivity in this area (Osman et al., 2019; Henson et al. 2022), reducing the amount of biologically fixed carbon to deeper water too.

Changes in the overall Atlantic ocean circulation (AMOC and SPG) can impact the spread of Atlantic water into the Arctic and affect marine ecosystems there. Summer sea ice decline reduces light limitation, rendering Arctic ecosystems more similar to the present North Atlantic (Yool et al., 2015). Increased seasonal phytoplankton blooms will deplete nutrients in the ocean, but increased inputs from rivers and coastal erosion can alleviate this, with Arctic primary production (i.e. the turnover photosynthesising plankton biomass) projected to increase by about 30-50 per cent in this century. Invasive species can also extend further into the Arctic habitat due to warming and current changes, e.g. in the Barents Sea and from the Pacific (Kelly et al., 2020; Neukermans et al., 2018; Oziel et al., 2020; Terhaar et al., 2021) (see also Chapter 1.3).

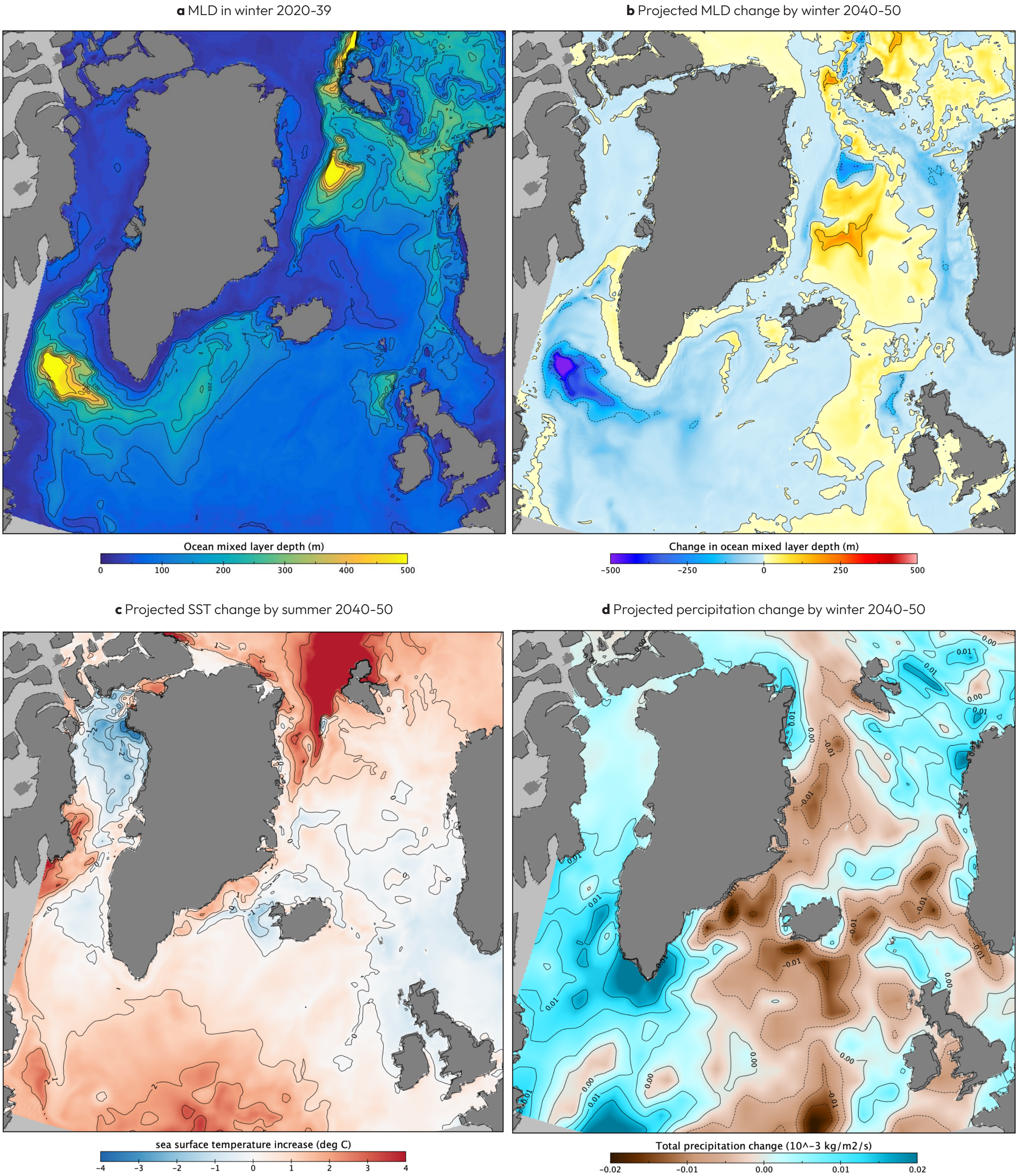


**Figure 1.4.7: a** Winter ocean mixed layer depth (MLD) as indicator of ocean convection in winter 2020-30 (January-March). **b** Changes in projected MLD by winter 2040-50. **c** Change in summer sea surface temperature (SST) and **d** winter total atmospheric precipitation, respectively, projected by winter 2040-50. NEMO-MEDUSA 1/4 degree high resolution model results using ssp370 CMIP6 scenario 2015-2099. High-resolution simulations are courtesy of Drs Andrew Coward, Andrew Yool, Katya Popova and Stephen Kelly, National Oceanography Centre, UK. Also see Swingedouw et al., (2021) for the IPCC CMIP6 model results.

In the North Atlantic, the AMOC can be defined as north-going warm 'limb' and saline upper waters and south-going, colder, denser deep water 'limb' (Frajka-Williams et al., 2019). In contrast, in the Subpolar North Atlantic and the SPG, the AMOC features a third 'limb' of a cold, fresh western boundary current with the origin in the Arctic Ocean and Nordic Seas (Bacon et al., 2023). This is likely linked with the deep convection and winter oceanic mixing in the Labrador, Irminger and Iceland seas, injecting waters into the deep, southward-flowing limb of the AMOC (Bower et al., 2019). Changes in SPG circulation are associated with the shallowing of the oceanic mixed layer and convection (Figure 1.4.7a,b) in the SPG and link the predicted future weakening of the North Atlantic subtropical gyre and a strengthening of the Nordic Seas gyre, pointing to the influences of the upstream changes in the Arctic on the North Atlantic (Swingedouw et al., 2021).

#### Evidence for tipping dynamics

Potential convection instability in the Labrador and Irminger Seas and the wider SPG is believed to be linked to lightening of the upper ocean waters due to reduced salinity (e,g., due to increased precipitation, Figure 1.4.7d), thus increasing 'stratification' – i.e. reduced mixing between layers of the water column. Warming (Figure 1.4.7c) also plays a role and could contribute to convection collapse (Armstrong McKay et al., 2022). Freshening and warming make surface waters more buoyant and thus harder to sink, which, beyond a threshold, can abruptly propel a self-sustained convection collapse (Drijfhout et al., 2015; Sgubin et al., 2017). This process can result in two alternative stable SPG states (Levermann and Born, 2007), with or without deep convection (Armstrong McKay et al., 2022). Similar to the AMOC, SPG stability is also strongly linked to the salt-advection feedback. When the SPG is 'on', it brings dense salty waters from the North Atlantic drift into the Irminger and Labrador Seas, allowing deep sinking and convection to occur (Born & Stocker, 2014; Born et al., 2016). When convection decreases due to stratification, the SPG weakens, less salty North Atlantic water flows eastwards, and the convection is further weakened, which eventually leads to convection collapse in some models. SPG collapse leads to cooling across the SPG region, and so will impact marine biology and bordering regions.

A freshwater anomaly is currently building up in the Beaufort gyre – a pile-up of fresh water at the surface of the Beaufort Sea in the Arctic – due to increased input from rivers, sea ice and snow melting as well as the prevailing clockwise (anticyclonic in the northern hemisphere) winds over the sea (Haine, et al., 2015; Regan et al., 2019; Kelly et al., 2020).

There is a considerable risk that this freshwater excess might flush into the SPG, disrupting the AMOC (Zhang et al., 2021). The most recent changes in Beaufort gyre size and circulation (Lin et al., 2023) suggest flushing might occur very soon or has already started. The SPG system has recently experienced its largest freshening for the last 120 years in its eastern side due to changes in the atmospheric circulation (Holliday et al., 2020). In contrast, so far there is only limited evidence of Arctic freshwater fluxes impacting freshwater accumulation in the Labrador Sea (Florindo-Lopez et al., 2020). An increased freshwater input into SPG water mass formation regions from melting of Greenland's glaciers can also inhibit deep water formation and reduce the SPG and AMOC (Dukhovskoy et al., 2021).

Although SPG changes are apparently linked to the AMOC the SPG collapse can occur much faster than AMOC collapse, on the timescale of only a few decades (Armstrong McKay et al., 2022). Armstrong McKay et al. (2022) estimated global warming threshold of ~1.8°C (1.1 to 3.8°C) for the SPG collapse (high confidence) based on climate models from CMIP5 and CMIP6. Abrupt future SPG collapse is diverse in the CMIP6 models, occurring as early as the 2040s (~1 to 2°C) but in only a subset of models. However, as these models better represent some key processes, the chance of SPG collapse is estimated at 36-44 per cent (Sgubin et al., 2017; Swingedouw et al., 2021).

#### Assessment and knowledge gaps

Similar to Armstrong McKay et al., (2022), the SPG is classified as a tipping system with medium confidence. A global warming threshold for tipping that could be passed within the next few decades, and an estimated tipping timescale of years to a few decades, raise reasons for concern. Furthermore, cessation of deep water production from other sources in the Labrador and Nordic Seas and the Arctic could also present other potential tipping points in the future North Atlantic (Sgubin et al., 2017).

### 1.4.2.2 Southern Ocean circulation

Two main tipping points in the Southern Ocean have been discussed in the past, which both could have large and global climate consequences. The first is the slowdown and collapse of the Antarctic Overturning Circulation; the second is the abrupt change in ocean circulation on the Antarctic continental shelf, leading to suddenly rising ocean temperature in contact with the Antarctic ice shelves fringing the ice sheet.

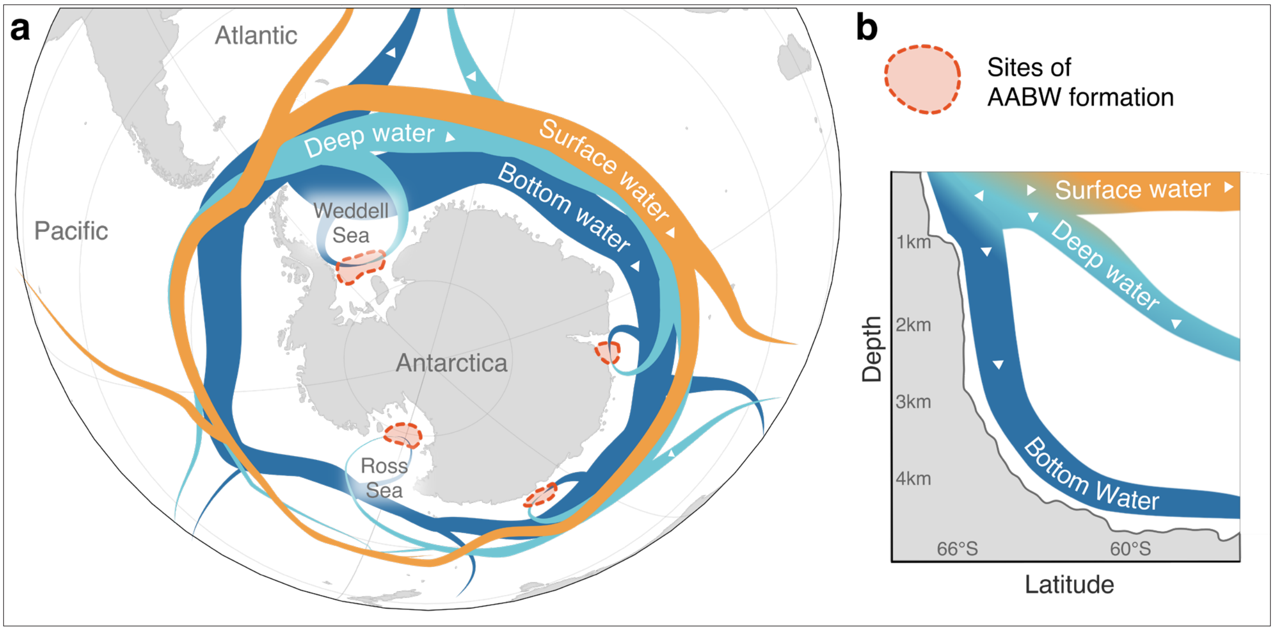


**Figure 1.4.8: Circulations and potential tipping systems in the Southern Ocean**. Adapted from Li, et al (2023) and IPCC SROCC Fig CB7.1

Along with the AMOC, the Antarctic overturing circulation constitutes the second branch of the global ocean overturning circulation linking the surface to the deep ocean (Figures 1.4.1 and 1.4.8), forming Antarctic bottom water (AABW) through sinking of the shelf waters around the Antarctic continent. A key mechanism is brine rejection from sea-ice formation: very salty water that is left behind when ocean water freezes, which causes the ambient liquid water to become heavier and sink. This is maintained by offshore winds blowing away from the Antarctic continent, pushing sea ice away from the coast and forming areas of open water (so-called polynyas) supporting brine rejection. The formation of AABW sustains the operation of the lower branch of the Antarctic overturning circulation (Figure 1.4.8 and Abernathey et al., 2016).

In contrast to our understanding of the AMOC, any changes related to the future of the Antarctic Overturning Circulation have remained at low or medium confidence due to a persistent lack of process understanding (Fox-Kemper et al., 2021; Heuzé et al., 2021; Purich and England 2023). However, evidence of its ongoing decline has escalated in recent years, both from observations (Gunn et al., 2023; Zhou et al., 2023; including record low sea ice extent in 2022-2023) and numerical models (Lago and England, 2019; Liu et al., 2022; Li et al, 2023), linked to the changes in melt water, wind trends, sea ice transport and water mass formation (Holland et al., 2012). (For the analysis of potential tipping in Antarctic sea ice, please see Chapter 1.2.)

Change or collapse in the Antarctic Overturning Circulation has the potential for widespread climate and ecosystem implications within this century. The Southern Ocean surface temperature is set by a delicate balance between ocean overturning strength, upper ocean stratification (the degree of mixing between ocean layers), and sea ice cover. The Antarctic Overturning circulation affects cloud feedbacks and has been shown to be a key regulator of Earth's global energy balance, so much so that it is the main control on the timing at which the 2°C global warming threshold will be reached for a given emission scenario (Bronsealer et al., 2018; Dong et al., 2022; Shin et al., 2023).

Reduced Antarctic overturning can also shift global precipitation patterns, resulting in drying of the Southern Hemisphere and wetting of the Northern Hemisphere (Bronsealer et al., 2018). Reduced Antarctic overturning also reduces the efficiency of the global ocean carbon sink, leaving more nutrient-rich water at the seafloor (Liu et al., 2022), and also affects global ocean heat storage (Li et al., 2023). Amplifying feedbacks to further shelf water warming and ice melt are also possible (Bronsealer et al., 2018; Purich and England, 2023; Li et al., 2023).

#### Evidence for tipping dynamics

Different generation climate models consistently project a slowing or collapse of the Antarctic overturning under a warming climate (Heuzé et al., 2015, 2021; Lago and England, 2019; Meredith et al., 2019; Fox-Kemper et al., 2021; Liu et al., 2022). However, our confidence in these models to assess change in Antarctic overturning is limited due to known limitations in the representation of dense water formation (Purich and England 2023). Limitations come also from the lack of representation of increased Antarctic ice sheet meltwater in most models (Fox-Kemper et al., 2021). Armstrong McKay et al., (2022) identified the Antarctic Overturning Circulation as a potential but uncertain tipping system in the climate system, but gaps in process understanding meant a threshold remained uncertain. They estimated it to be prone to collapse at a global warming level of 1.75-3°C based on Lago and England, (2019).

Specifically designed model experiments aiming to bridge some of these limitations, in combination with evidence from observed changes (Gunn et al., 2023; Purkey and Johnson, 2013), confirm that we are currently heading toward a decline and possible collapse of the Antarctic Overturning Circulation (Li et al., 2023, Zhou et al., 2023). The rapidity of this decline might even be underestimated, according to recent observations (Gunn et al., 2023). The sensitivity of the overturning to increases in upper ocean stratification is also consistent with palaeo evidence. Observations from marine sediments suggest that AABW formation was vulnerable to freshwater fluxes during past interglacials (Hayes et al., 2014; Huang et al., 2020; Turney et al., 2020) and that AABW formation was strongly reduced (Skinner et al., 2010; Gottschalk et al., 2016; Jaccard et al., 2016) or possibly totally curtailed (Huang et al., 2020) during the Last Glacial Maximum and earlier transient cold intervals.

Local water mass characteristics and associated circulation regimes on the Antarctic continental shelf are setting the rate of ice shelf melt rates in ice 'cavities', the regions of ocean water covered by floating ice shelves. Relatively warm water reaching the continental shelf in west Antarctica causes high basal melt rates with severe consequences for the ice shelf, ice sheet dynamics, and sea level rise (Naughten et al., 2023). In contrast, the largest ice shelf cavities in the Weddell and Ross Seas are not exposed to this relatively warm water, and consequently have melt rates orders of magnitude smaller than in West Antarctica. Despite this, the Weddell and Ross Sea ice shelf cavities have been shown to exhibit tipping behaviour (Hellmer et al., 2012; 2017; Siahaan et al., 2022). Models show that they are prone to sudden warming of their cavity under future climate change, dramatically increasing basal melting with important consequences for global sea level rise (Hellmer et al., 2012; 2017; Siahaan et al., 2022). Once tipped into a warm state, such cavities could be irreversibly maintained in such a state, even when forcing is reduced (Hellmer et al., 2017). However, it remains unclear what threshold would need to be crossed to tip those cavities from a cold to warm state, and it may only occur under extreme climate change scenarios.

#### Assessment and knowledge gaps

In summary, the combination of process-based understanding and observational, modelling and palaeoclimate evidence suggests that Antarctic Overturning Circulation will continue to decline in the 21st Century. There is increasing evidence for positive amplifying feedback loops that can lead to the collapse of the overturning, with widespread global climate and ecosystem consequences. Closely linked to this is a potential tipping in continental shelf water temperature, driven by amplifying meltwater feedbacks once a regional temperature threshold is crossed. We therefore classify the Southern Ocean Circulation as a tipping system with medium confidence. However, its potential tipping thresholds remain uncertain.

### 1.4.2.3 Monsoons

Monsoon circulations are large-scale seasonal changes in the direction and strength of prevailing winds driven by insolation (incoming solar radiation) and local temperature differences between land and ocean. Their dynamics are strongly influenced by the seasonal migration of the Intertropical Convergence Zone (ITCZ), the regional band in the tropics where the trade winds from the northern and southern hemisphere converge and rise as part of the tropical atmospheric overturning circulation (see Figure 1.4.1). The term 'monsoon' was historically associated with summer precipitation over South Asia; however, monsoon systems affect other parts of the globe such as East Asia, Africa, Australia and the Americas.

Historically, monsoons were seen as large-scale sea breeze circulations driven by land-sea heating differences due to seasonal changes in incoming solar radiation. Currently, a perspective of a global monsoon has emerged (Trenberth et al., 2000; Wang & Ding, 2008), where the monsoon systems are seen as interconnected and driven by localised seasonal and more extreme migrations of the ITCZ (Gadgil, 2018; Geen et al., 2020, and references within). Monsoon regions in the world experience heavy precipitation in the summer months, and the global monsoon system is an integral part of the global hydrological cycle, contributing ~31 per cent of total precipitation over the globe (Wang and Ding, 2008).

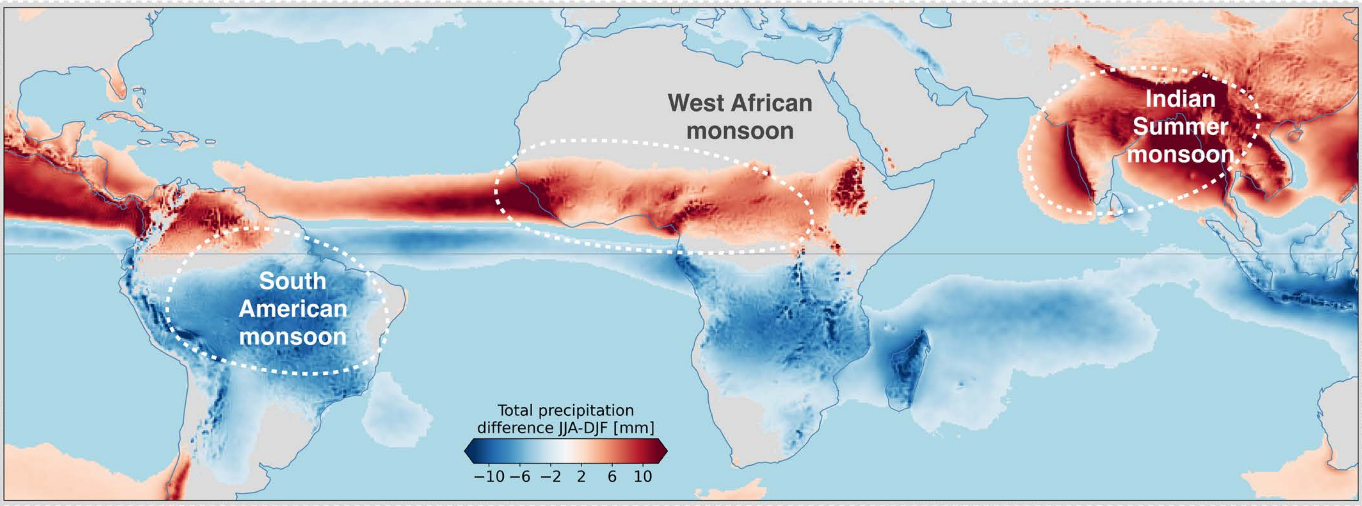


**Figure 1.4.9:** Monsoon systems. Shown is the total precipitation difference between Northern hemisphere summer (June-August, JJA) and winter months (December-February, DJF), highlighting the dominant precipitation patterns over South America (SAM), West Africa (WAM) and India (ISM). Generated using Copernicus Climate Change Service information (Hersbach et al., 2023), with monthly averages over 1980-2010.

There is a recent intensification trend in global monsoon precipitation, mainly due to enhanced northern hemisphere summer monsoon (Wang et al., 2012). It will likely continue in the future (high confidence, IPCC 2021, by ~1-3% per °C warming) because of increased water vapour related to warming driven by increased $CO_2$ in the atmosphere (Hsu et al., 2013; Lee and Wang, 2014; Chen et al., 2020; Ha et al., 2020; Wang et al., 2019); although a few studies conversely show that climate warming may lead to a weakened global monsoon circulation (Hsu et al., 2012, 2013). Climate simulations also project expansion of global monsoon domain areas with increasing $CO_2$ (Wang et al., 2020; Paik et al., 2023) and increased frequency of monsoon precipitation extremes in the 21st Century (Chevuturi et al., 2018; Ali et al., 2020; Ha et al., 2020; Katzenberger et al., 2021).

Monsoon precipitation is vital for agrarian populations and livelihoods in vast areas of South Asia, Africa and South America, and changes to it could expose almost two thirds of the global population to disastrous effects (Wang et al,. 2021). Hence it is crucial to understand the dynamics and potential nonlinear changes or tipping behaviour of monsoon systems under a changing climate. Here the 'tipping' of monsoon systems refers to a significant, feedback-driven shift in the precipitation state of the monsoon, with implications for the regional and global climate and ecosystems. In this discussion we assess if the major regional monsoon systems (West African, Indian and South American) show any evidence of nonlinear (tipping or abrupt) responses to climate forcings based on available literature.

### Indian summer monsoon (ISM)

During the summer season over South Asia (June-September), winds from the south west carry large amounts of water vapour from the Indian Ocean to the Indian subcontinent and cause heavy precipitation in the region, providing ~80 per cent of the total annual precipitation (Figure 1.4.9). ISM precipitation shows considerable intra-seasonal, interannual and decadal variability, many times with precipitation extremes (leading to droughts, floods) during the season, and years and decades with above and below (in drought years) normal precipitation. Indian monsoon variability is strongly influenced by ocean-atmosphere interactions such as El Niño Southern Oscillation (ENSO, see Chapter 1.4.2.5), Indian Ocean Dipole events (irregular changes in the temperature gradients in the Indian Ocean, Cherchi et al., 2021; Chaudhary et al., 2021; Hrudya et al., 2021), and cooler temperatures in the North Atlantic (Borah et al., 2020).

ISM precipitation declined in the second half of the 20th Century, attributed mainly to human-driven aerosol loading (Bollasina et al., 2011) and strong Indian Ocean warming (Roxy et al., 2015). Recent studies (Jin and Wang, 2017) suggest it has revived since 2002, linked to enhanced warming over the Indian subcontinent due to reduced low clouds, resulting in an increased land-ocean thermal gradient. Future projections suggest increases in the ISM precipitation in future warming scenarios (by 5.3% per celsius of global warming, according to CMIP6 models, Katzenberger et al., 2021) and a longer monsoon duration (Ha et al., 2020).

### Evidence for tipping dynamics

Many periods of abrupt ISM transitions have been identified in past monsoon records in association with high-latitude climate events (Schulz et al., 1998; Morrill et al., 2003) such as during Heinrich events (glacial outbursts that temporarily shut down the AMOC – see 1.4.2.1) (McManus et al., 2004; Stager et al., 2011), the Younger Dryas (a temporary return to more intense glacial conditions 12,900-11,700 years ago; (Cai et al., 2008; Carlson 2013), and several periods during the more recent Holocene (Gupta et al., 2003; Berkelhammer et al., 2012; Yan and Liu, 2019). However, the mechanisms of such abrupt transitions are not clearly understood. Efforts have been made to identify any Indian monsoon tipping mechanisms using simplified models (Zickfeld et. al., 2005; Levermann et al., 2009).

An internal feedback mechanism, a 'positive moisture advection feedback' (Zickfeld et. al., 2005; Levermann et al., 2009; Schewe et al., 2012), has been suggested as responsible for abrupt transitions simulated using these analytical models. In this feedback, the atmospheric temperature gradient between the land and cooler ocean in summer leads to the onshore transport of moist air (advection), which then rises, forms clouds and condenses into rain. The phase transition from vapour to liquid warms the surrounding air (through the release of latent heat, or 'diabatic heating'), increasing the land-ocean temperature gradient and sustaining this monsoon circulation. Any forcing that weakens this pressure gradient can therefore lead to monsoon destabilisation (Zickfeld et al., 2005). If monsoon winds weaken, advection and condensation reduce, and the threshold for a monsoon tipping is reached when the diabatic heating fails to balance the heat advection away from the region (Levermann et al., 2009).

Contrarily, follow-up studies (Boos and Storelvmo, 2016) challenge occurence of any tipping in these simplified models, and rule out any abrupt monsoon responses to human-driven forcings in the future, and instead attribute past monsoon shifts to rapid forcings or vegetation feedbacks. Simplified models omit key aspects and feedbacks in the monsoon system (specifically, static stability of the troposphere in the models that simulated the monsoon tipping, (Boos and Storelvmo, 2016; Kumar and Seshadri, 2022). Hence, more studies using models that represent the complexities of the monsoon and palaeoclimate data are required for a clearer picture on any non-linear changes in the monsoon system.

Apart from climate change, aerosols pose another significant human-driven pressure on the Earth system. Aerosols influence the Earth's radiative budget, climate and hydrological cycle by reflecting or absorbing solar radiation, changing the optical properties of clouds, and also by acting as cloud condensation nuclei. An increase in anthropogenic aerosols has been attributed as the major reason for the decline of Northern Hemispheric summer monsoon strength from the 1950s to 1980s (Cao et al., 2022), due to its dimming effect.

A large increase in regional aerosol loading over South and East Asia (>0.25 Aerosol Optical Depth, AOD, Steffen et al., 2015) could potentially switch the Asian regional monsoon systems to a drier state. Further, hemispheric asymmetries in the aerosol loading (>0.15 AOD, Rockström et al., 2023), due to volcanic eruptions, human sources or intentional geoengineering, could lead to hemispheric temperature asymmetries and changes in the location of the ITCZ, significantly disrupting regional monsoons over West Africa and South Asia (Haywood et al., 2013; Rockström et al., 2023; Richardson et al., 2023). However, there is no direct evidence of aerosols causing a tipping of the monsoon systems, and uncertainties in threshold estimates are large due to complex aerosol microphysics and aerosol-cloud interactions. Hence, systematic observational and modelling approaches would be needed to reduce the uncertainties, as well as additional assessments of interhemispheric asymmetries in the aerosol distribution.

#### Assessment and knowledge gaps

The ISM system was earlier classified as one of the Earth's tipping systems (Lenton et al., 2008), based on the threshold behaviour of the monsoon in the past and the moisture-advection feedback (Levermann et al., 2009), but this was refuted by later studies (Boos and Storelvmo, 2016; Seshadri, 2017). Most recently, Armstrong McKay et al,. (2022) categorise ISM as an "uncertain potential [climate] tipping element" as global warming is not likely to cause tipping behaviour directly in ISM precipitation.

Based on this current literature, the chances for ISM exhibiting a tipping behaviour towards a new low-precipitation state under climate change are uncertain, warranting extensive studies on the subject. However, potential tipping behaviour in the AMOC (see Chapter 1.4.2.1, 1.5.2.5, and relation to global monsoon described in West African monsoon below) or increase in the interhemispheric asymmetry of aerosol loading in the atmosphere beyond potential threshold levels could lead to large disruptions to monsoon systems. This could cause calamitous effects on millions of people in the monsoon regions, even in the absence of tipping.

#### West African monsoon (WAM)

The West African monsoon (WAM) controls hydroclimatic conditions, vegetation and mineral-dust emissions of northern tropical and subtropical Africa, up to the dry Sahel region at the southern edge of the Sahara Desert (**Figure 1.4.9**). The strength of the monsoon shows large variations over a range of timescales from interannual to decadal and longer. Albedo (reflectivity of the Earth's surface) changes caused by human-driven land-cover changes and desertification (Charney et al., 1975; Charney, 1975; Otterman, 1974) can affect rainfall: a less vegetated surface with higher albedo increases radiative loss, thereby reducing temperature and suppressing the rising and condensation of moist air into rainfall (i.e. convective precipitation). Variations of sea surface temperatures (SSTs) in different oceanic basins can also drive interannual and decadal variability in WAM precipitation (Rodríguez-Fonseca et al., 2015). Other major factors that affect WAM variability are land surface variability such as variations in soil moisture (Giannini et al., 2013; Zeng et al., 1999), vegetation (Charney et al., 1975; Kucharski et al., 2013; Otterman, 1974; Wang et al., 2004; Xue, 1997), high-latitude cooling (Collins et al., 2017) and dust emissions (Konare et al., 2008; Solmon et al., 2008; Zhao et al., 2011).

#### Evidence for tipping dynamics

Palaeoclimate records underscore dramatic variations of the WAM in the more distant past, such as the periodic expansion of vegetation into the Sahara Desert during the so-called 'African humid periods' (AHPs) and linked to the emergence of ancient cultures along the Nile. Another example is the drought 200-300 years ago, which caused the water level of Lake Bosumtwi in Ghana to fall by almost four times as much as it did during the drought of the 1970s and 1980s. Large past variations of the WAM, such as those during the AHPs, raise the question of whether present-day anthropogenic global warming could have potentially significant impacts on the WAM. Although the nature and magnitude of radiative forcing were different during the AHPs than they are now (i.e. an external change in insolation due to orbital forcing versus an internal change from increased greenhouse gases), the fact that the AHPs occurred under a globally warmer climate than the pre-industrial period invites questions.

Some palaeoclimate archives show WAM precipitation changes that took place over several centuries (deMenocal et al., 2000; McGee et al., 2013), i.e. an order of magnitude faster than the orbital forcing. However, others show a much more gradual change (e.g. Kröpelin et al., 2008) with a time-varying withdrawal of the WAM from North to South following the insolation changes (Shanahan et al., 2015). Because of geographic variability of the African landscape and African monsoon circulation, abrupt changes can occur in several, but not all, regions at different times during the transition from the humid to arid climate (Dallmeyer et al., 2021).

By inducing latitudinal movements of the ITCZ, change in the AMOC is considered to play a role in shifts of global monsoon systems. Palaeoclimate evidence suggests that glacial meltwater-induced weakening of the AMOC during Heinrich events in the last glacial period led to abrupt Asian and African monsoon weakening (Mohtadi et al., 2014; Mohtadi et al., 2016). Similarly, the Younger Dryas led to a cool and dry state over Northern Hemisphere tropical monsoon regions. North Atlantic fresh water-hosing simulations using climate models (Lewis et al., 2010; Pausata et al., 2011; Kageyama et al., 2013) confirm these shifts in ITCZ can occur as a result of substantial glacial meltwater release. These influences of AMOC on the monsoon systems have also been studied in the context of the South American monsoon (see below). Hence, a collapse of AMOC (see Chapter 1.4.2.1) has the potential to cause disruptions to the regional monsoon systems and other tropical precipitation systems over Asia, Africa and South America (Gupta et al., 2003; IPCC 2021).

#### Assessment and knowledge gaps

Abrupt changes in one region can be induced by abrupt changes in others, a process sometimes referred to as 'induced tipping'. The AHP transition of the Sahara was slow with respect to timescales of individual humans and local ecosystems, but regionally rapid with respect to changes in the driver. Based on the record of large past variations of WAM precipitation patterns (including collapse), and the existence of positive amplifying feedbacks, we classify WAM as a tipping system with low confidence. This is in line with previous assessments (Armstrong McKay et al., 2022), in which a lower tipping threshold of 2°C global warming was estimated but attributed low confidence due to limited model resolution of vegetation shifts, and model disagreements in future trends. The timescale of abrupt shifts is estimated to range from decades as observed in CMIP5 models (Drijfhout et al., 2015) to centuries based on palaeorecords (Hopcroft and Valdes, 2021; Shanahan et al.; 2015). Potential additional destabilisation through AMOC weakening and atmospheric aerosol loading, and the far-reaching implications of WAM tipping, call for intensified research efforts on this system.

#### South American monsoon (SAM)

The South American monsoon (SAM) system is characterised by strong seasonality in precipitation, even though it does not show a reversal of low-level winds like in the Asian monsoon (Zhou and Lau, 1998; Vera et al., 2006; Liebmann and Mechoso, 2011; Carvalho et al., 2012). Studies are relatively few compared to the Asian and African monsoon systems, as it was not classified as a monsoon system until a couple of decades ago (Zhou and Lau, 1998).

A mature SAM system (from December to February) shows features such as enhanced northeastern trade winds, increased land-ocean thermal gradient and the development of an active convective zone (the South Atlantic Convergence Zone) (Figure 1.4.9; Zhou and Lau, 1998). The SAM system affects vast areas of tropical South America all the way to southern Brazil, and provides more than 50 per cent of the annual precipitation to these regions (Vera et al., 2006) including most of the Amazon rainforest. SAM precipitation varies from interannual to orbital timescales (Chiessi et al., 2009; Liebmann and Mechoso, 2011; Carvalho and Cavalcanti, 2016; Hou et al., 2020).

The influence of anthropogenic climate change on the SAM precipitation is ambiguous (Douville et al., 2021), and many CMIP5/CMIP6 models are noted for their poor representation of SAM precipitation (Jones and Carvalho, 2013; Douville et al., 2021). IPCC AR6 finds high confidence in delayed onset of the SAM precipitation since the 1970s associated with climate change, which could worsen with increased $CO_2$ levels (Douville et al., 2021). However, the projected future change in total SAM precipitation is uncertain, as the models show low agreement on the projections (Douville et al., 2021).

#### Evidence for tipping dynamics

Orbital timescale changes (i.e. over tens of thousands of years) in SAM precipitation seem to be largely controlled by changes in insolation and respond linearly to it (Cruz et al., 2005; Hou et al., 2020). Millennial-scale changes (i.e. over thousands of years) in the SAM are thought to be associated with variations in strength of the AMOC, as described for the West African monsoon above. In particular, palaeo evidence indicates that an increase in South American precipitation to the south of the equator followed weakening of the AMOC related to Heinrich events (Mulitza et al., 2017; Campos et al., 2019). Similarly, meltwater flux from the Laurentide Ice Sheet during the Younger Dryas may have led to a warm and wet state over tropical South America to the south of the equator (McManus et al., 2004; Broecker et al., 2010; Venancio et al., 2020; Brovkin et al., 2021). Earth system model projections of AMOC collapse impacts on the tropical rainfall in South America are model-dependent, but generally find a reduction in rainfall over northern South America and an increase over the southern Amazon (Bellomo et al., 2023; Nian et al., 2023; Orihuela-Pinto et al., 2022; Liu et al., 2020: see 1.5.2.4).

Further, deforestation over 30-50 per cent of the Amazon rainforest led to a tipping point in the SAM system in one model (Boers et al., 2017), causing precipitation reductions of up to 40 per cent in non-forested parts of the western Amazon. This reduction is caused by the breakdown of a positive amplifying feedback mechanism that involves latent heat of condensation over the Amazon rainforest due to transpiration (i.e. water lost from plants) and water vapour transport from the Atlantic. Reduced transpiration due to deforestation can no longer sufficiently provide water vapour to sustain the latent heat required, thereby reducing the inflow of oceanic water vapour, and leading to a monsoon tipping in this model (Boers et al., 2017). (see 1.3.2.1 for more on Amazon dieback)

#### Assessment and knowledge gaps

A combination of climate change and deforestation could lead to substantial changes in the SAM system, affecting many millions of people. Additionally, a decrease in AMOC strength could potentially trigger major changes in tropical South American precipitation (see 1.5.2.4). However, the current scarcity of research in the subject limits our ability to fully understand and assess the tipping potential of the system, and we classify the possibility of SAM tipping to be uncertain.

### 1.4.2.4 Tropical clouds, circulation and climate sensitivity

Clouds play an important role in the climate system, as they contribute to the regulation of Earth's energy budget linked to the amount of solar radiation trapped or reflected back to space (Figure 1.4.10). In general, high, thin clouds at several kilometres altitude have a two-fold warming effect on the climate: They have a high transmissivity for shortwave radiation (incoming sunlight) and low emissivity for longwave radiation (heat), meaning they allow most of the sunlight to reach the surface but block some of the heat escaping to space. In contrast, low, thick clouds reflect more sunlight, and also have a high emissivity for long-wave radiation, allowing more heat to escape, and so have a cooling effect. A changing climate, which causes changes in temperature, humidity and circulation patterns, affects the formation and dynamics of these clouds. This, in turn, can influence the climate and how much warming results from increased atmospheric $CO_2$ concentrations (i.e. 'climate sensitivity').

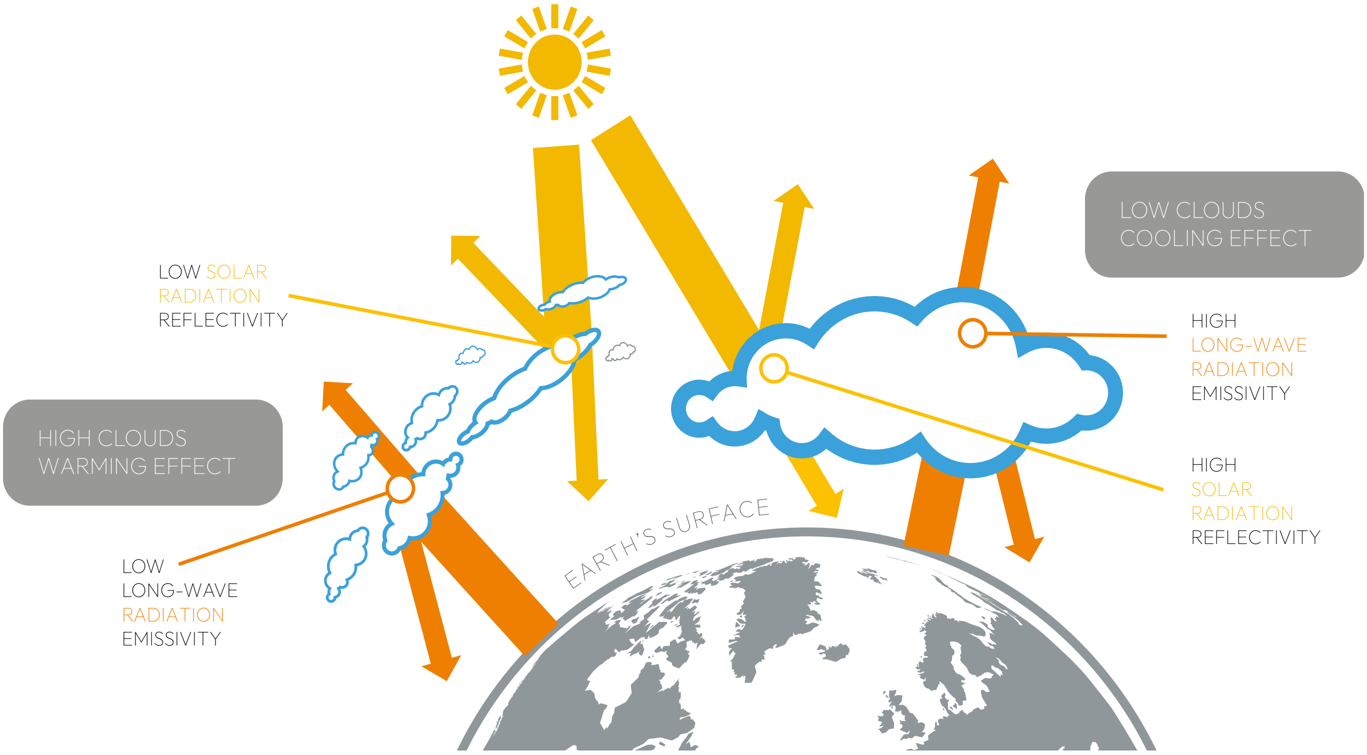


**Figure 1.4.10:** The role of clouds in regulating Earth's radiation budget.

### Evidence for tipping dynamics

Literature on cloud-induced tipping points is very limited. Yet cloud-forming processes exhibit strong hysteresis on weather timescales. Indeed, a cloud droplet forms when water starts to stick to a particle after a certain level of humidity (in which a so-called hygroscopic aerosol particle crosses a humidity tipping point into an unstable condensational growth phase); and precipitation, once initiated, is a self-reinforcing cascade where larger particles fall faster and hence grow faster by collisions. Coupling of these micro-scale processes to atmospheric dynamics can lead to spontaneous and irreversible transitions at the intermediate mesoscale – in particular, the transition of shallow cloud layers from closed to open-cell geometries (honeycomb-like cloud patterns formed by convecting air) (Feingold et al., 2015) and self-aggregation of deep convection (Muller et al., 2022). Both of these significantly decrease cloud cover and albedo, potentially enabling climate interactions. Could further coupling out to planetary scales produce climate-relevant tipping behaviour? Complicating this question is the fact that cloud-related processes are not well represented in current climate models, limiting their ability to guide us.

The most-discussed possibility has been the extreme case of a global climate runaway. If the atmosphere became sufficiently opaque to infrared (i.e. if it became harder for longwave heat energy to escape due to overcast high cloud, very high humidity, or CFC-like greenhouse gases filling in spectral absorption windows), the planet could effectively lose its ability to cool to space, producing a Venus-like runaway. Although general circulation models (GCMs) and palaeoclimate evidence suggest climate sensitivity rises as climate warms (Sherwood et al., 2020), calculations show virtually no chance of runaway warming on Earth at current insolation levels (Leconte et al., 2013).

A more plausible scenario is unexpectedly strong global positive amplifying radiative feedback from clouds and high climate sensitivity. Although presumably reversible, this would be serious. With respect to high clouds, suggested missing feedbacks (due to novel microphysical or aggregation mechanisms) have generally been negative (e.g. Mauritsen and Stevens, 2015).

Low clouds are a greater concern: one recent study using a multiscale atmospheric model found a strong and growing positive amplifying feedback from rapid disappearance of these clouds (Schneider et al., 2019), highlighting the possibility of nonlinear cloud behaviour and surprises (Bloch-Johnson et al., 2015; Caballero and Huber, 2013). Although various observations generally weigh against high-end climate sensitivities above 4°C per $CO_2$ doubling, they cannot rule them out (Sherwood et al., 2020).

A final possibility is surprising reorganisations of tropospheric circulation (i.e. in the lowest layer of the atmosphere). Innovative atmospheric models (Caballero and Carlson, 2018; Seeley and Wordsworth 2021) and geologic evidence (Tziperman and Farrell, 2009; Caballero and Huber 2010) have suggested possible 'super-MJO' (the 'Madden-Julian Oscillation' being the dominant mode of 'intraseasonal' variability in the tropical Indo-Pacific, characterised by the eastward spread of enhanced or suppressed tropical rainfall lasting less than a season) and/or reorganisation of the tropical atmospheric circulation in a warmer climate due to cloud-circulation coupling. These scenarios are supported by little evidence, but if they did occur they could massively alter hydrology in many regions. Poor representation of tropical low clouds has also likely inhibited coupled model simulations of decadal variability or regional trends (Bellomo et al., 2014; Myers et al. 2018), raising the possibility that, even if clouds cannot drive tipping points, they might amplify other tipping points in ways that are missing from current models.

### Assessment and knowledge gaps

In summary, concern about cloud-driven tipping points is relatively low. Cloud feedbacks will, however, likely affect the strength of climate responses, including for many tipping points. For example, they could potentially amplify variability, and current models may not be capturing this well. High climate sensitivity from strongly positive cloud feedbacks also cannot be ruled out.

### 1.4.2.5 El Niño-Southern Oscillation (ENSO)

The El Niño-Southern Oscillation (ENSO) is the dominant interannual mode of variability in Earth's climate. It originates in the tropical Pacific, where it affects sea surface temperatures (SST), trade winds, rainfall and many other climate variables. El Niño events typically happen every three to five years (hence the term 'interannual'). The tropical Pacific average climate is characterised by a strong east-west gradient along the equator of about 5-6°C, with warmer SSTs in the west and colder SSTs in the east maintained by easterly Pacific trade winds. During El Niño – the warm phase of this oscillation – this gradient weakens, while during La Niña – its cold phase – it intensifies (schematically depicted in Fig 1.4.11a). Both phases of this oscillation have far-reaching impacts on global climate and weather patterns, ecosystems and human health (e.g. McPhaden et al,. 2020).

The impacts of ENSO become especially pronounced during the strongest events, often referred to as extreme El Niños, defined as events with SST anomalies above a chosen threshold (for example 2 standard deviations as in Heede and Fedorov 2023a) (Fig. 1.4.11b). At their peak, these events can eliminate the east-west ocean temperature gradient along the equator, leading to a temporary collapse of the trade winds. Additionally, an extreme El Niño causes an increase in global mean surface temperature of up to 0.25°C (Hu and Fedorov 2017), contributing to the prevalence of heat waves around the globe. While only a few El Niño events reach large magnitudes, the global impacts of these events result in billions of dollars in damage (Callahan and Mankin 2023).

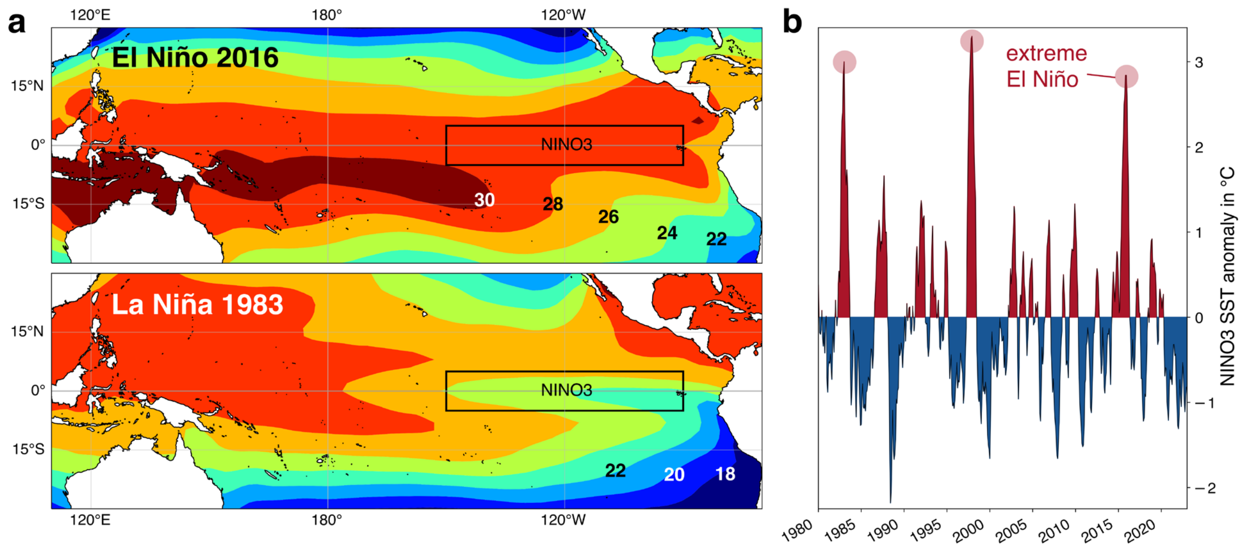


**Figure 1.4.11:** ENSO warm and cold phases and observational record. a Examples of strong El Niño (top) and La Niña (bottom) events seen in the tropical Pacific surface temperature (SST) distribution, with characteristic strong and weak SST gradient along the equator, respectively. **b** ENSO record since the 1980s. Note the three extreme events of the past four decades (1982, 1997 and 2015) and the weakening of ENSO variability between years 2000 and 2015. Temperature is averaged for the NINO3 region (5°C-5°N, 150°W-90°W) in the eastern equatorial Pacific. Based on NOAA Extended Reconstructed SST V5 data (Huang et al., 2017).

As this report was being written, a new El Niño event was announced (WMO, 2023), and will likely reach peak strength around the time of its publication in December 2023. At the time of writing, it is projected to be a 'strong' event, reaching ~2°C relative to neutral (CPC/NCEP/NWS, 2023).

#### Evidence for tipping dynamics

Extensive research conducted since the 1980s has significantly advanced our understanding of the physics behind El Niño, leading to improved predictive capabilities of climate models (L'Heureux et al., 2017). ENSO is now recognised as a large-scale, irregular, internal oscillatory mode of variability within the tropical climate system, influenced by atmospheric noise (Timmermann et al., 2018). The spatial pattern of ENSO is determined by ocean-atmosphere feedbacks, while its timescale is determined by ocean dynamics. In particular, it is a sequence of self-reinforcing feedbacks between SSTs, changes in zonal surface winds, equatorial upwelling and ocean thermocline depth that promotes the growth of El Niño anomalies (i.e. Bjerknes feedbacks, McPhaden et al,. 2020).

Coral-based proxy data indicate that the amplitude and frequency of ENSO events has gradually increased during the Holocene (Grothe et al., 2020; Lawman et al., 2022), possibly due to an increase in extreme El Niño events. All extreme El Niños in the observational record (1982, 1997 and 2015) occurred during the accelerated growth of global mean temperatures. This raises the question whether this trend is indicative of upcoming changes in the tropical Pacific to conditions with more frequent extreme El Niño events.

In the context of tipping points, the question arises: is there a critical threshold with an abrupt and/or irreversible transition to such a new state? Several recent studies (Cai et al,. 2018, 2022; Heede and Fedorov, 2023a) have indeed suggested that El Niño magnitude and impacts may intensify under global warming (**Figure 1.4.12**), even though there is still no model consensus on the systematic future change in ENSO, as IPCC AR6 and the results in **Figure 1.4.12** suggest.

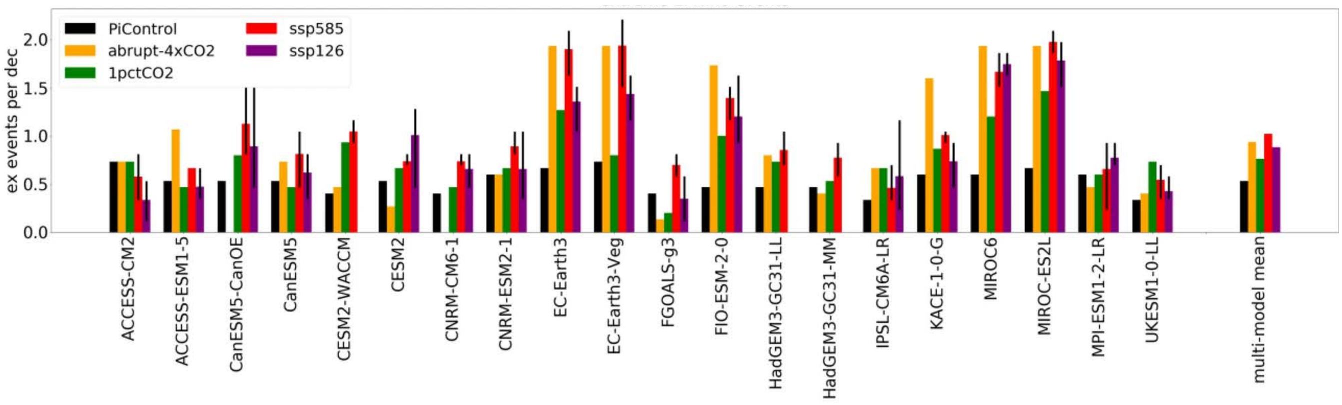


**Figure 1.4.12:** Overview of projected changes in extreme El Niño events in CMIP6 climate models. The bar chart shows the time-mean frequency of extreme El Niño events (the number of events per decade) for several idealised and more realistic global warming experiments (abrupt-4xCO2, 1pctCO2, SSP5-8.5 and SSP1-2.6) next to the pre-industrial Control simulation (piControl). From Heede and Fedorov, 2023a

It is projected that the eastern equatorial Pacific (EEP) will warm faster than the western part of the basin, leading to an EEP warming pattern or El Niño-like mean conditions, associated with weaker Pacific trade winds. Most climate model future projections exhibit this pattern (e.g. DiNezio et al,. 2009; Xie et al,. 2010; Heede and Fedorov 2021), and increased ENSO variability is prevalent in models that simulate stronger nonlinear (Bjerknes) feedbacks (Cai et al,. 2022). A recent comprehensive study of CMIP6 models and scenarios concluded that, although a common mechanism to explain a change in ENSO activity across models is missing, its increase under warming scenarios is robust (Heede and Fedorov, 2023a).

Furthermore, during the warm Pliocene epoch approximately 3-5 million years ago, when global surface temperatures were ~3°C above pre-industrial, the east-west SST gradient was indeed reduced (Wara et al,. 2005; Fedorov et al., 2006, 2013, 2015; Tierney et al., 2019). This state is often referred to as 'permanent El Niño-like' conditions, which does not indicate ENSO changes, but rather a consistent mean decrease in the east-west SST gradient. While debates on this topic are ongoing, estimates for this gradient reduction range from 1.5°C to 4°C, depending on the time interval, proxy data and the definition of this gradient.

#### Assessment and knowledge gaps

Therefore, there is a general expectation of a future reduction in the Pacific's east-west SST gradient by the end of the 21st Century. Together with other contributing factors, such as the strengthening of the MJO, the dominant intraseasonal mode in the tropical Indo-Pacific (Arnold et al., 2015; see 1.4.2.4), this reduction is expected to amplify ENSO (Heede and Fedorov, 2023a). Additionally, a warmer atmosphere can hold more water vapour, which could result in stronger precipitation and heating anomalies in the atmosphere, leading to greater remote impacts of El Niño events.

Consequently, the collective evidence implies an increase of El Niño magnitude and impacts under global warming. There is, however, insufficient indication for a critical transition associated with an abrupt or irreversible regime shift towards a new, more extreme or persistent, ENSO state, such that ENSO is considered with medium confidence not to be a tipping system (see also Armstrong McKay et al., 2022). However, it is well connected to other Earth system components (e.g. affecting tropical monsoon rainfall), thereby possibly playing a role in tipping cascades, linking different tipping elements via global teleconnections (see Chapter 1.5).

Notably, the projections of a future EEP warming pattern, weaker mean trade winds and stronger El Niño events contradict decadal trends in the tropical climate over the past 30 years or so. In fact, since the early 1990s, the Pacific trade winds have strengthened, and the eastern equatorial Pacific has become colder (e.g. Ma Zhou, 2016; Seager et al., 2022; Wills et al., 2022; Heede and Fedorov 2023b). Whether these trends reflect an ocean thermostat-like response to global warming, internal variability of the system, or both, remains an open question. Similarly, the magnitude of ENSO events has been generally weaker since the 2000s compared to the 1980s and 1990s (Fig. 1.4.11b; also Capotondi et al., 2015 or Fedorov et al., 2020).

Therefore, debates on the future of the tropical Pacific and ENSO revolve around the question of when the transition to a mean EEP pattern and weaker trade winds may occur, likely leading to a stronger El Niño and more frequent extreme events. Simulations with global climate models including strongly eddying ocean components (Wieners et al., 2019; Chang et al., 2020) and the currently developing 2023-2024 El Niño are expected to help reduce persistent model tropical biases in SST, precipitation and ocean thermocline, and to resolve some of the remaining issues.

### 1.4.2.6 Mid-latitude atmospheric dynamics

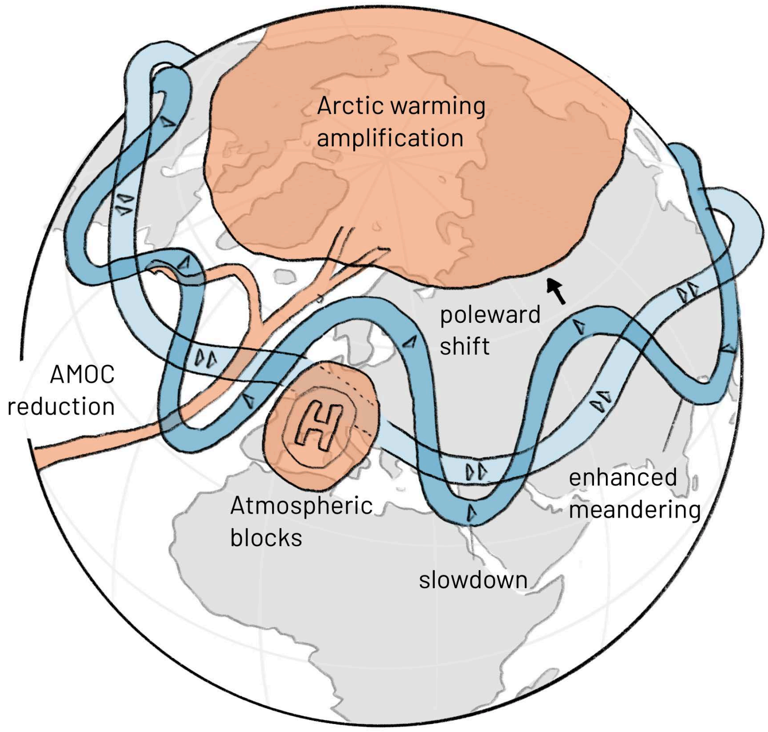


**Figure 1.4.13:** Potential changes in mid-latitude atmospheric circulations, exemplary for the Northern Hemisphere. Reduction of AMOC, atmospheric blocking events, Arctic warming and other drivers can modify the jet stream. Potential consequences are a northward shift, slowdown and enhanced meandering, related to increases in extreme weather phenomena.

Mid-latitude atmospheric circulation is characterised by a band of strong westerly winds (see Figure 1.4.1), with largest velocities at an altitude of 7-12km, forming the so-called northern polar 'jet-stream'. The jet serves as a separation of cold air masses at high-latitudes in the north from temperate air masses further south. Large meanders in the jet are referred to as planetary, or Rossby, waves. In most cases, these waves move over large distances and decline over timescales of a few days. When persisting for a prolonged time over the same location (referred to as 'quasi-stationary' waves) they can lead to high-impact climate extremes, including temperature extremes or heavy precipitation. An example is the record-breaking heatwave of 2021 in the North American Pacific Northwest (Bartusek et al., 2022).

Atmospheric features such as blocks (quasi-stationary high-pressure regions that divert, or 'block', the large-scale atmospheric flow on timescales of several days to weeks) are intimately linked to these persistent meanders in the jet. A widely discussed effect of climate change is a poleward shift of the mid-latitude jet, although this may be season and location-dependent (Oudar et al,. 2020), and smaller than previously thought (Curtis et al., 2020) (Figure 1.4.13).

#### Evidence for tipping dynamics

In climate models, the magnitude of the jet's shift strongly depends on the reduction of the AMOC (see Chapter 1.4.2.1). Models with a strong AMOC reduction in the future tend to project a much stronger poleward shift of the jet than models with a weaker AMOC reduction, making this the largest atmospheric circulation uncertainty in regional climate change projections (Bellomo et al., 2021).

Furthermore, it has been suggested that the mid-latitude flow might weaken, leading to more persistent and slower-moving weather patterns (Coumou et al., 2015; Kornhuber and Tamarin-Brodsky, 2021). A possible driver is Arctic amplification – namely the fact that the Arctic is warming more rapidly than the rest of the planet, partly driven by sea ice loss (see Chapter 1.2). This reduces the equator-pole temperature contrast, and could result in a weakening and enhanced meandering of the jet stream (Francis and Vavrus, 2015). While Arctic amplification is most evident during winter, such increase in waviness may also be occurring during the summer season (Coumou et al., 2018). However, evidence that the occurrence of large-amplitude atmospheric waves is increasing is debated (Screen and Simmonds, 2013; Blackport and Screen, 2020; Riboldi et al., 2020), and mechanisms which would reduce blocking in the future have also been proposed (Kennedy et al., 2016).

As part of this debate, it has been proposed that several weather extremes in recent decades were associated with a quasi-stationary, quasi-resonant wave pattern. This results from the interaction of climatological waves that are perpetually forced by orography (mountain geography) and land-sea contrasts with transient meanders of the jet stream (Petoukhov et al., 2013), given a set of favourable conditions (White et al., 2022). Petoukhov et al., (2013) also hypothesised that Arctic amplification and the associated weakened, wavier jet may provide increasingly favourable conditions for the occurrence of quasi-resonance. This can result in circulation features which accelerate regional extreme weather occurrence trends – for example, heatwave trends in Europe (Rousi et al., 2022), although the direction of causality is debated (Wirth and Polster, 2021). If recent extreme events are indeed associated with a resonance mechanism that only kicks in when the jet crosses a certain threshold in waviness, a tipping point might be involved. However, it is uncertain whether this would be associated with hysteresis and irreversibility or would just be a reversible, but abrupt, shift of the atmosphere towards enhanced large-amplitude mid-latitude waves.

More generally, there is no robust evidence that continued climate change and Arctic amplification will lead to a tipping towards a wavy-jet state, systematically higher amplitude and/or more frequent planetary waves, or blocks. Equally, there is no robust evidence that these hypothetical changes would be self-sustaining. Indeed, while a number of large changes in atmospheric dynamical features may occur under climate change, these are typically discussed as gradual changes, without explicit hysteresis or tipping behaviour. Similarly, there is no robust evidence pointing to tipping-like behaviour in the jet stream's latitudinal location, although gradual, long-term shifts may occur.

It should nonetheless be noted that atmospheric circulation responses to climate change are characterised by large model uncertainty and are possibly biased by the relatively low resolution of global climate models compared to, for example, weather-prediction models (Shepherd, 2019). In addition, some climate models show that tipping behaviour in atmospheric blocking, in the form of a self-sustaining, feedback-driven shift, is possible (Drijfhout et al., 2013).

**Assessment and knowledge gaps**

Although theoretically possible, there is thus no robust evidence for tipping point behaviour in mid-latitude atmospheric circulations in the near future. At the same time, a number of relevant physical processes are currently debated or ill-constrained. We thus evaluate, with low confidence, the mid-latitude atmosphere as not displaying tipping points.

The mid-latitude large-scale circulation itself may, though, still affect or be affected by tipping behaviour of other components of the Earth system to which it is coupled, such as the land surface, overturning ocean circulations (e.g., Orihuela-Pinto et al., 2022) or high-latitude cryosphere. Indeed, such interactions can lead to abrupt climate shifts. A recent example is the transition to hotter and drier conditions in inner East Asia, resulting from drier soils, a strengthened land-atmosphere coupling, and a contribution from large-scale circulation anomalies (Zhang et al., 2020). Furthermore, joint non-tipping changes in mid-latitude atmospheric dynamics, the associated surface climate, and other components of the Earth system, may lead to tipping point behaviour, for example in vegetation (Lloret and Batllori, 2021). This could in turn feed back onto the atmospheric circulation.

Due to such feedbacks and interactions between the atmospheric circulation and other components of the Earth system, and due to its role in weather and climate extremes, an improved understanding of the physical processes underlying changes in mid-latitude atmospheric dynamics under recent and future climate change appears pivotal in a tipping point context. Large model uncertainty in projecting abrupt regional atmospheric circulation changes conditioned by changes in the ocean, cryosphere or land surface would lend itself eminently for a storyline approach (Zappa and Shepherd, 2017). Tipping of atmospheric circulation, and associated weather extremes, would then be conditioned by threshold behaviour in other, connected systems.

Finally, we argue for the need to investigate whether recent, record-breaking weather extremes can be explained by the slowly changing likelihood distribution that belongs to the last decades, or whether they are signs of abruptly changing likelihood distributions. Such a shift in the distribution of extremes could be diagnosed using extreme value theory. Although a shift cannot be associated with a global tipping point, it would suggest that the extreme value distribution of (a) certain type(s) of extreme weather did witness regional tipping, whether or not reversible, in the sense of a large nonlinear change in response to a small and gradual change in forcing, potentially driven by self-sustaining feedbacks.

# Chapter 1.5 Climate tipping point interactions and cascades

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## Summary

This chapter reviews interactions between climate tipping systems and assesses the potential risk of cascading effects. After a definition of tipping system interactions, we map out the current state of the literature on specific interactions between climate tipping systems that may be important for the overall stability of the climate system. For this, we gather evidence from model simulations, observations and conceptual understanding, as well as archetypal examples of palaeoclimate reconstructions where propagating transitions were potentially at play. This chapter concludes by identifying crucial knowledge gaps in tipping system interactions that should be resolved in order to improve risk assessments of cascading transitions under future climate change scenarios.

The scientific content of this chapter is closely based on the following scientific manuscript: Wunderling, N., von der Heydt, A. et al.: Climate tipping point interactions and cascades: A review, EGUsphere [preprint], https://doi.org/10.5194/egusphere-2023-1576, 2023.

## Key messages

- Tipping systems in the climate system are closely interacting, meaning a substantial change in one will have consequences for subsequently connected tipping systems.
- A majority of interactions between climate tipping systems are destabilising. While confirmation or rejection through future research is necessary, it seems plausible/possible that interactions between climate tipping systems destabilise the Earth system in addition to climate change effects on individual tipping systems.
- We are quickly approaching global warming thresholds where tipping system interactions become relevant, because multiple individual thresholds are being crossed.

## Recommendations

- At least three approaches are needed to improve risk assessments for tipping cascades: (i) Time-series analysis of observations and palaeoclimate data, (ii) Earth system models designed for tipping system interactions, (iii) Risk analysis using large model ensembles.
- Palaeoclimate observations improve our understanding of tipping cascades, by studying past abrupt or transition events such as the Eocene-Oligocene Transition, Bølling-Allerød warm period.
- Besides direct interactions, additional indirect feedbacks (for example, via temperature) should be quantified in order to determine the risk for tipping cascades.

## 1.5.1 Introduction and definition

The tipping systems identified in the climate system generally operate not in isolation from each other, but connected either directly or mediated via changes in the overall climate (for example, global temperature) (Liu et al., 2023; Kriegler et al., 2009). Via such connections (see Figure. 1.5.1) tipping in one subsystem can therefore cause tipping in another, which we define as a tipping cascade (see Definition below) (Wunderling et al., 2021a; Klose et al., 2020; Dekker et al., 2018).

**Definition:**
Here we call the linkages between tipping systems and/or other nonlinear components as tipping interactions, which could have a stabilising or a destabilising effect. The most extreme case is the situation in which the tipping of element 'A' causes a subsequent tipping of element 'B'. In this report, we define a sequence of tipping events involving several nonlinear components of the Earth system as **tipping cascades** (Dekker et al., 2018; Wunderling et al., 2021a). These tipping cascades can come in various forms dependent on the ordering of tipping systems (e.g. Klose et al., 2021; Dekker et al., 2018). Eventually, a tipping cascade might result in a fundamental change in the Earth's equilibrium climate.

For example, disintegration of the Greenland Ice Sheet (GrIS) can lead to an abrupt shift in the Atlantic Meridional Overturning Circulation (AMOC), while an abrupt change in AMOC strength can lead to an intensification of the El Niño-Southern Oscillation (ENSO). Interactions between climate tipping systems could effectively lower the thresholds for triggering a tipping event as compared to those individual tipping systems in isolation (Wunderling et al., 2021a; Klose et al., 2020). Moreover, one or more tipping events could activate processes leading to additional $CO_2$ emissions into the atmosphere; permafrost thaw and forest dieback are typical examples of such additions of stored $CO_2$ into the atmosphere via positive amplifying feedbacks (Wunderling et al., 2020; Lenton et al., 2019; Steffen et al., 2018).

It is also conceivable that components of the Earth system, though not necessarily tipping systems in themselves, could mediate or amplify tipping in other components, thereby creating larger-scale impacts. As a result, some of these nonlinear components are also taken into account in this chapter. A prominent example is Arctic summer sea ice cover, which is not expected to show tipping behaviour ( Lee et al., 2021) (see 1.2.2.2), but can nevertheless trigger tipping events in the ocean-atmosphere-cryosphere system (Gildor and Tziperman, 2003). On the other hand, an abrupt transition in one tipping system may also stabilise other climate subsystems (Nian et al., 2023; Sinet et al., 2023) as is the case for a weakening AMOC decreasing local temperatures around Greenland (Jackson et al., 2015).

While most tipping systems that have been proposed so far are clearly regional (with some being large-scale), there are significant knowledge gaps with respect to their tipping probability, impact estimates and timescales, as well as their interactions. The potential of a tipping cascade that could lead to a global reorganisation of the climate system (Steffen et al., 2018; Hughes et al., 2013) remains therefore speculative. However, since multiple individual tipping point thresholds may be crossed during this century with ongoing global warming, and could lead to severe tipping system interactions and cascading transitions in the worst case, it is critical to review the current state of knowledge and reveal research gaps that need to be addressed (Armstrong McKay et al., 2022; Masson-Delmotte et al., 2021; Rocha et al., 2018).

## 1.5.2 Interactions between climate tipping systems and further nonlinear climate components

### 1.5.2.1 Interactions across scales in space and time

In this section, we lay out the current state of the scientific literature on the interaction processes between several tipping systems and some other nonlinear components of the Earth system. The summary is shown in Figures 1.5.1 and 1.5.3.

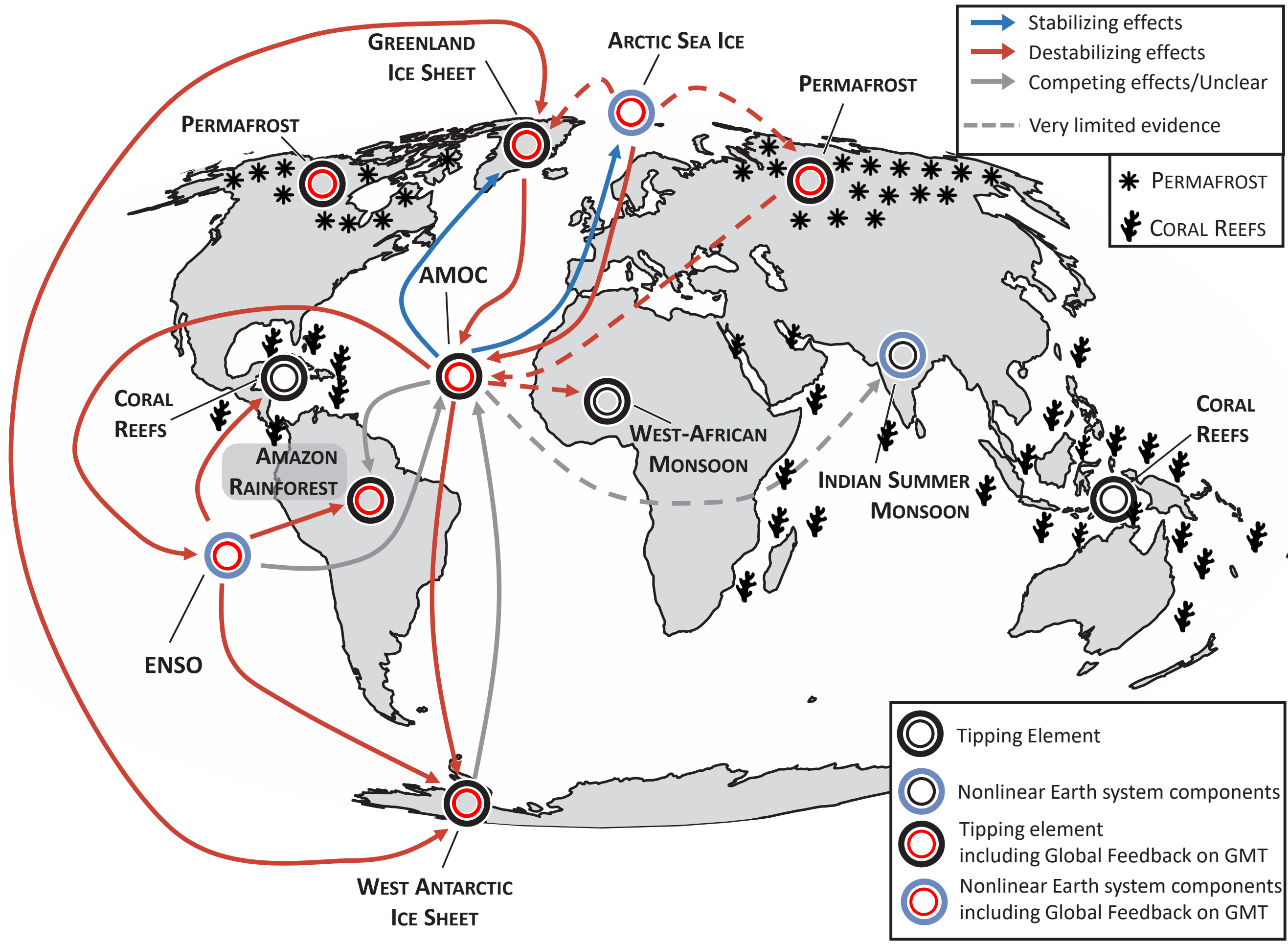


**Figure 1.5.1:** Interactions between established and more speculative tipping systems on a world map. All tipping systems discussed in this chapter are shown together with their potential connections. The causal interaction links can have stabilising (blue arrows), destabilising (red arrows), or unclear (grey arrows) effects. For some systems, it is speculative whether they are tipping systems on their own (such as ENSO or the Arctic sea ice) and they are denoted as such (blue outer ring) but they are included if they play an important role in mediating transitions towards (or from) core tipping systems. Tipping systems that exert a notable feedback on global mean temperature (GMT) when they tip are denoted by a red inner ring (for instance via albedo changes in case of a disintegration of the Greenland or West Antarctic ice sheets or Arctic sea ice, or via carbon release through tipping of permafrost or rainforests). This temperature feedback can be positive (i.e. amplifying warming, as likely for the permafrost, the Arctic sea ice, the Greenland and West Antarctic ice sheets, the Amazon rainforest and ENSO) or negative. Source: Wunderling and von der Heydt et al.

These systems are not isolated entities but interact across the entire globe (Figure 1.5.1). Not only do the interactions span global distances, but some tipping systems themselves can be of regional spatial scale (e.g. coral reefs or the GrIS), while others cover significant portions of the globe (e.g. the AMOC). Also, timescales differ vastly among the different climate tipping systems: some are considered fast tipping systems once the process has been initiated (in the order of years/decades to centuries, such as the Amazon rainforest and AMOC), while others are considered slow tipping systems (in the order of centuries to millennia, such as the GrIS).

These different spatial and temporal scales of the individual tipping systems are therefore also important for their interactions and are mapped out in Figure 1.5.2 (Rocha et al., 2018; Kriegler et al., 2009). The respective processes of the interactions can be found in Figure 1.5.3, alongside an estimation of the interaction direction and, if available, an estimation of their strength.


**Figure 1.5.2:** Interactions between tipping systems across scales in space and time. Temporal scales are transitioning times of a disintegrating tipping system from months up to millennia. Spatial scales denote the system size from sub-continental to (nearly) global scales. Transitioning times are taken from Armstrong McKay et al. (2022), and spatial scales from Winkelmann et al. (2022). The causal links can be stabilising (blue arrows), destabilising (red arrows), or unclear (grey arrows). Some tipping systems are particularly speculative (such as ENSO or the Arctic sea ice) and denoted as such (outer blue border). Tipping systems that exert a feedback on the global mean temperature (GMT) when they tip are shown with an inner red border. Adapted from: Wunderling and von der Heydt et al.

## 1.5.2.2 Interactions between ice sheets and the AMOC

The AMOC, Greenland Ice Sheet (GrIS), and West Antarctic Ice Sheet (WAIS) are key tipping systems and are threatened by increasing $CO_2$ emissions and temperatures (Armstrong McKay et al., 2022; Pörtner et al., 2019). Moreover, GrIS, AMOC, and WAIS interact on very different timescales, ranging from decades to multiple centuries. While some of those links might be stabilising, others are destabilising and would allow for the possibility of large-scale cascading events.

### Greenland Ice Sheet to AMOC

The AMOC depends on the formation of dense, salty water in the high latitudes of the North Atlantic. As GrIS melting increases (1.2.2.1), the associated discharge of salt-free freshwater in the ocean will decrease surface water salinity and thereby density, inhibiting the formation of dense waters and weakening the circulation. As less salt is transported to the North Atlantic, the salt-advection feedback implies a self-sustained freshening of the high latitudes of the North Atlantic, which, in the worst case, can result in the collapse of the AMOC (1.4.2.1). On top of this classic positive/amplifying feedback, there exists a wide range of other feedbacks related to the AMOC, either negative (heat advection feedback) or positive (evaporation feedback).

An overall destabilising impact of GrIS melting on the AMOC is mostly consistent across models, where adding freshwater in the North Atlantic (Jackson and Wood, 2018; Mecking et al., 2016; Stouffer et al., 2007), also in combination with increasing $CO_2$ emissions (Bakker et al., 2016; Swingedouw et al., 2006), leads to a substantial weakening of the circulation. Importantly, in the case of AMOC collapse, some models suggest it does not recover within century timescales (Jackson and Wood, 2018; Mecking et al., 2016). Note, however, that estimated melt rates of the GrIS are generally smaller than the amount of freshwater additions in models necessary to collapse the AMOC (Sinet et al., 2023, Jackson and Wood 2018), and it is currently a smaller contributor than increased Arctic precipitation.

### West Antarctic Ice Sheet to AMOC

In the case of freshwater release in the Southern Hemisphere originating from West Antarctica, different opposing processes are at play that could affect the AMOC. These effects have been identified to act on different timescales and depend on the state of the circulation (Berk et al., 2021; Swingedouw et al., 2009). First, the weakening of Antarctic Bottom Water (AABW; see 1.4.2.2) formation might lead to enhancement of the AMOC through the so-called 'ocean bipolar seesaw'. This describes the tendency for opposing temperature changes in the Southern and Northern Hemisphere, with ocean bottom water changes in response to ice sheet melt in either hemisphere taking a long time to affect the other hemisphere.

Second, the increase in wind intensity over the Southern Hemisphere, related to an increase in sea ice cover, might also help to enhance the AMOC (Li et al., 2023; Swingedouw et al., 2008). Third, the release of freshwater in the Southern Ocean might eventually reach the North Atlantic on a longer timescale (centuries), possibly weakening the AMOC. As a result, the impact of a WAIS collapse on the AMOC is still unclear, as most models show either a slight weakening (e.g. Stouffer et al., 2007; Seidov et al., 2005) or a slight strengthening (e.g.

Swingedouw et al., 2009) of the circulation. Notably, some studies also found that a sufficient freshwater release into the Southern Ocean allows for delaying an AMOC collapse (Sadai et al., 2020), or a recovery from it (Weaver et al., 2003).

**AMOC to ice sheets**

An AMOC collapse would decrease northward heat transport, leading to a substantial cooling of the Northern Hemisphere, and warming in the Southern Hemisphere (Pedro et al., 2018; Jackson et al., 2015; Stouffer et al., 2006). Cooling the high latitudes of the North Atlantic could stabilise the GrIS. Conversely, the related warming of the Southern Ocean represents a destabilising impact on the WAIS, being susceptible to these warmer ocean waters via the ice shelves and their buttressing effect on upstream ice flow (Favier et al., 2014; Joughin et al., 2014).

**Direct interactions between Greenland and West Antarctic ice sheets via sea level.**

It is known that an increase in sea level has an overall destabilising influence on marine-based sectors of ice sheets, possibly triggering or enhancing the retreat of their grounding line (Schoof, 2007; Weertman, 1974). In the case of ice sheet collapse, the induced sea level rise would vary locally depending on gravitational effects (with sea level falling near the former ice sheet as less water is attracted towards it), rotational effects, and mantle deformation (Kopp et al., 2010; Mitrovica et al., 2009). Overall, sea level rise is expected to negatively impact both the GrIS and WAIS, but more strongly the latter, where most of the bedrock lies well below sea level (Gomez et al., 2020).

### 1.5.2.3 Arctic sea ice interactions

**Interactions between AMOC and Arctic sea ice**

Changing Arctic sea ice cover can change AMOC strength in two main ways (Sévellec et al., 2017): First, it alters radiative heating and ocean-atmosphere heat loss via changing albedo. More precisely, as the Arctic sea ice area has substantially decreased over the past 40 years, especially during summer months (Masson-Delmotte et al., 2021), the open water fraction of the Arctic Ocean has increased and will continue to do so (Crawford et al., 2021). This has led to an increase in the absorption of solar radiation and to subsequent ocean warming, which can spread to ocean convection areas, affecting stratification and potentially weakening the AMOC. Second, the recent decrease in Arctic sea ice area together with ice loss from the GrIS has added freshwater to the Arctic Ocean. Although the trend in freshwater content has slowed during the past decade (Solomon et al., 2021), it could affect North Atlantic deep water formation and thus weaken the AMOC.

The AMOC can also affect Arctic sea ice via the transport of warm water to the North Atlantic Ocean, and subsequently to the Arctic Ocean via the Barents Sea Opening and Fram Strait. A weaker AMOC could result in lower ocean heat transport and increased Arctic sea ice area (Delworth et al., 2016). However, recent observations show that the ocean heat transport to the Arctic has increased, especially on the Atlantic side (Docquier and Koenigk, 2021; Polyakov et al., 2017; Onarheim et al., 2015; Årthun et al., 2012). Thus, the effect of a weaker AMOC may be merely to slow the pace of ongoing increases in ocean heat transport and the associated decrease in Arctic sea ice (Liu et al., 2020).

**Effect of Arctic sea ice on the Greenland Ice Sheet and Arctic permafrost**

Besides interacting with the AMOC, reduced Arctic sea ice cover could have a direct effect via regional warming on further high-latitude tipping systems such as the GrIS and Arctic permafrost (1.2.2.4). In the case of sustained Arctic summer sea ice loss, which may occur during the second half of this century (Niederdrenk et al., 2018) or sooner (Kim et al., 2023), additional warming levels are in the order of 0.3-0.5°C regionally over Greenland and the permafrost (Wunderling et al., 2020). Regional warming levels may be higher if Arctic winter sea ice also disappears under high-emission scenarios. Further, it has been found that regional Arctic sea ice loss has a limited effect for Greenland warming patterns and is mainly relevant for coastal parts of Greenland (Pedersen and Christensen, 2019).

At the same time, Arctic sea ice loss leads to increased coastal permafrost erosion (Hošeková et a., 2021; Casas-Prat and Wang, 2020; Grigoriev et al., 2019; Nielsen et al., 2020 and 2022). Abrupt changes in summer-autumn seaice retreat from the permafrost coast leads to an increase in waves, resulting in sudden increases in erosion rates (– about 50-160 per cent in the last 50 years (a two- to fourfold increase in hotspots in the Laptev and Beaufort Seas) (Irrgang et al., 2022). Thus, coastal permafrost collapse leads to a potential cascading risk of carbon releases locally to the Arctic ocean and the atmosphere of 0.0023–0.0042 GtC per year per degree celsius by the end of the century (Nielsen et al., 2022). The erosion causes changes in the shoreline, sediments, carbon, nutrients and contaminants in the coastal seas and offshore marine environment (Irrgang et al., 2022).

### 1.5.2.4 Effects of AMOC changes on the Amazon rainforest

The strength of the AMOC exerts a substantial influence on the climate of tropical South America – most importantly, on rainfall and its seasonal distribution (1.4.2.3). This in turn affects the state and stability of another potential tipping system in the Earth system: the Amazon rainforest.

The most important large-scale effect of the AMOC on Amazon rainfall works via the pattern of sea surface temperatures (SSTs) in the Atlantic, and the associated southward shifts of the Intertropical Convergence Zone (ITCZ) and the tropical rain belt. There is widespread agreement that a reduction or even collapse of the AMOC would lead to reduced SSTs in the North Atlantic and increased SSTs in the South Atlantic (Bellomo et al., 2023; Manabe and Stouffer, 1995). This southward shift would cause a substantial reduction in rainfall over northern South America, and an increase in rainfall over the southern Amazon rainforest as well as over northeastern Brazil, which is directly affected by the tropical rain belt (Jackson et al., 2015). Nevertheless, over the Amazon basin, rainfall change is uncertain and model-dependent (Ciemer et al., 2021; Swingedouw et al., 2013; Stouffer et al., 2006), resulting in a large uncertainty concerning the potential impact of AMOC weakening in the Amazon rainforest dieback.

Although different Earth system models have different biases in the location, shape and strength of the tropical rain belt, they generally agree on the AMOC collapse-induced increase in precipitation over the southern portion of the Amazon and northeastern Brazil (Bellomo et al., 2023; Nian et al., 2023; Orihuela-Pinto et al., 2022; Liu et al., 2020). Given that the forests in the southern half of the basin contribute mostly to the rainfall generation over the basin (Staal et al., 2018), one could speculate that this would lead to a stabilisation of the Amazon, given that a substantial fraction (24-70 per cent, Baudena et al., (2021) and references therein) of the rainfall of the basin is nonetheless produced by local moisture recycling. More generally, the full spectrum of rainforest stressors, including human-driven pressures such as land use changes driving deforestation, has to be taken into account when assessing AMOC effects over the Amazon rainforest (Lovejoy and Nobre, 2018).

### 1.5.2.5 Interactions between ENSO and tipping systems

The El Niño-Southern Oscillation (ENSO) is the most important mode of climate variability on interannual time scales, fundamentally affecting regional and global atmospheric and oceanic circulation (McPhaden et al., 2006). The response to climate change of ENSO itself is still debated, mainly because there are multiple (positive and negative) feedback processes in the tropical Pacific ocean-atmosphere system, whose relative strengths determine the response of ENSO variability (Timmermann et al., 2018; Cai et al., 2015; see 1.4.2.5).

Further, recent studies disagree about the future frequency of El Niño phases under global warming (Cai et al., 2021; Wengel et al., 2021). Although it is debated or even unlikely whether ENSO should be considered a tipping system in itself (Armstrong McKay et al., 2022), it exerts important effects on other tipping systems (for example, tropical monsoon rainfall). Through its global 'teleconnections' (i.e. links between widely separated climate phenomena), ENSO has the potential to influence multiple Earth system components including the AMOC, Amazon rainforest, WAIS, warm water coral reefs and tropical monsoon systems.

#### Interactions between ENSO and AMOC

Various physical mechanisms have been discussed to explain how a decline or complete shutdown of the AMOC could affect ENSO. An AMOC decline typically leads to cooling in North Atlantic surface temperatures, which affects the global atmospheric circulation, including the trade winds in the tropical Pacific. Therefore, many complex climate models project that AMOC decline leads to an intensification of northeasterly trade winds and a southward shift of the ITCZ, eventually leading to an intensification of ENSO amplitude through nonlinear interactions (Timmermann et al., 2007).

While the response of the trade winds and ITCZ to AMOC decline seems to be relatively robust within different climate models, the response in ENSO magnitude or frequency is much more model-dependent and thus uncertain. It should be noted that most complex climate models still exhibit severe biases in tropical temperature patterns, partly caused by not properly resolved oceanic processes (Wengel et al., 2021), which complicates the understanding of the fate of ENSO under global warming and AMOC changes.

The reversed pathway – i.e. ENSO impacting the AMOC – depends on several atmosphere-ocean processes which may not be adequately resolved in current state-of-the-art models. A relatively robust teleconnection exists between the El Niño phase and the North Atlantic Oscillation (NAO) (Ayarzagüena et al., 2018; Brönnimann et al., 2007). The relationship between the AMOC and the NAO in Earth system models depends on the subpolar North Atlantic background state; the AMOC is less sensitive in models that have extensive sea ice cover in the North Atlantic, while in models with less sea ice cover, the background upper ocean stratification largely determines how sensitively the AMOC reacts (Kim et al., 2023). As for ENSO, unbiased representation of the North Atlantic average state represents a significant challenge for state-of-the-art Earth system models, in part due to insufficient resolution of intermediate mesoscale ocean eddies.

#### Influences of ENSO on the Amazon rainforest

The frequency and amplitude of ENSO variability have changed on decadal to centennial timescales in the past (Cobb et al., 2013). In recent years, extreme El Niño events combined with global warming have become increasingly associated with unprecedented extreme drought and heat stress across the Amazon basin (Jiménez-Muñoz et al., 2016), leading to increases in tree mortality, fire and dieback (Nobre et al., 2016). Imposing the surface temperature pattern of a typical El Niño event in a global atmosphere-vegetation model suggests increased drought and warming in the Amazon (Duque-Villegas et al., 2019), which could enhance rainforest dieback (1.3.2.1) and transition regions of the Amazon rainforest from carbon sinks sources.

The destabilising effects from ENSO towards the Amazon rainforest are compounded by direct climate change effects and land use change and deforestation, often mediated by intensifying fires (1.5.2.4). Parts of the Amazon rainforest undergoing degradation and drying have already turned from a net carbon sink to a carbon source (Gatti et al., 2021). Further, it remains uncertain whetherthe vast Amazon rainforest would tip in its entirety or only partially, as it may have multiple intermediate stable states. In such a scenario, only specific areas in the rainforest margins might transition into degraded land (Rietkerk et al., 2021; Bastiaansen et al., 2020).

#### Influences of ENSO on the WAIS

Recent significant surface melt events on West Antarctica were associated with strong El Niño phases (Scott et al., 2019; Nicolas et al., 2017). It has been proposed that these melt events were caused by atmospheric blocking, eventually leading to warm air temperature anomalies over West Antarctica that pass the melt point of parts of the ice sheet (Scott et al., 2019). Using reanalysis data, satellite observations and hindcasting methods, strong indications have been found that the Ross and Amundsen Sea Embayment regions are most affected by El Niño phases (Scott et al., 2019; Deb et al., 2018).

Taken together, this adds to a growing body of literature that indicates a disintegration of the WAIS, especially along the Ross-Amundsen sector, would be favoured by strong El Niño phases, and tipping risks may increase if El Niño phases would become more frequent or intense under ongoing climate change (Cai et al., 2021; Wang et al., 2017; Cai et al., 2014; 1.4.2.5). This may be concerning in particular because the Amundsen region is where the most vulnerable glaciers of the WAIS are located, such as the Pine Island and Thwaites glaciers (Favier et al., 2014; Joughin et al., 2014).

#### Influences of ENSO on warm-water coral reefs

ENSO drives abnormally high SSTs (and seasonal summer heat waves), which are superimposed on already warming oceans. Anomalous heat destabilises corals, resulting in severe bleaching and mortality across multiple coral species on spatial scales exceeding thousands of kilometres (1.3.2.7). While ENSO is geographically modulated by other ocean dipoles (e.g. North Atlantic Oscillation, Indian Ocean dipole) (Houk et al., 2020; Krawczyk et al., 2020; Zhang et al., 2017), the Pacific signal is dominant and El Niño warm phases have been related to global episodes of extreme heat stress since the 1970s (1979/1980, 1997/98 and 2014-2017, for example) (Krawczyk et al., 2020; Muñiz-Castillo et al., 2019; Lough et al., 2018; Le Nohaïc et al., 2017).

As global warming progresses and oceans become significantly warmer, the incidence of mass bleaching can occur more frequently even without El Niño warm phases (Veron et al., 2009), with warmer conditions compared to three decades ago (McGowan and Theobald, 2023; Muñiz-Castillo et al., 2019). The global recurrence of bleaching has reduced to an average of six years (Hughes et al., 2018) – sooner than expected from climate models and satellite-based sea temperatures. While recovery from repeated bleaching events has been observed (Palacio-Castro et al., 2023; Obura et al., 2018), the proposed global mean warming thresholds of 1.5°C and 2°C would result in widespread reef die-off (70-90 and 90-00 per cent respectively loss of coral reefs globally) (Lough et al., 2018; Schleussner et al., 2016; Frieler et al., 2013), and lower thresholds of 1.0-1.5°C are argued for in this report (1.3.2.7).

#### Effects of AMOC and ENSO changes on tropical monsoon systems

Future climate projections show a weakening of the AMOC, which can be substantial in its impact on the regional and global climate (Pörtner et al., 2019; see 1.4.2.1). Indeed, model simulations of freshwater addition (via 'hosing experiments') in the North Atlantic show a clear southward shift of the ITCZ in response to the AMOC weakening and a decrease in northward oceanic heat transport (Defrance et al., 2017; Swingedouw et al., 2013; Stouffer et al., 2006). This shift of the ITCZ impacts the various monsoon systems worldwide (Chemison et al., 2022), as is also visible in palaeorecords (Sun et al., 2012).

For example, palaeo-reconstructions of a Heinrich event (a massive iceberg release causing further cooling in the North Atlantic region, 1.5.3.2) of the penultimate deglaciation between 135,000 and 130,000 years ago have been compiled, suggesting an increase in Indian summer monsoon rainfall (Nilsson-Kerr et al., 2019), but a subsequent reduction of the length of the monsoon rain season (e.g. Wassenburg et al., 2021). Summarised, a reduction of the AMOC strength, subsequent cooling of the Northern Hemisphere and southward shifts the ITCZ (Chemke et al., 2022) affect spatial rainfall patterns and amount of rainfall in the Northern Hemisphere semi-arid and tropical monsoon regions of West Africa and India/Asia.

An AMOC weakening has also been shown to strengthen the Indo-Pacific Walker circulation via cooling of the equatorial Pacific and warming of the Southern Hemisphere/Antarctic climate on a multi-decadal timescale (Orihuela-Pinto et al., 2022). The observed potential AMOC weakening during the last multiple decades might be partially affected by interannual ocean-atmosphere interactions, such as ENSO. These superimposed effects, operating across timescales, alter relationships between the ENSO and tropical monsoon systems and, thereby, regional rainfall patterns in a warmer climate (Mahendra et al., 2021; Pandey et al., 2020). For example, while the linear relationship between ENSO and the Indian summer monsoon rainfall has weakened, the ENSO-West African monsoon relationship has increased in recent decades (Srivastava et al., 2019).

However, ENSO and AMOC effects on tropical monsoon systems are still highly uncertain and should be further constrained using palaeoclimate reconstructions and Earth system models (see 1.4.2.3 for more on monsoon tipping).

### 1.5.2.6 Effects of permafrost thaw on the global hydrological cycle

Permafrost regions have accumulated substantial amounts of ice in their soils. With ground ice melting away in a warmer climate, permafrost landscapes experience abrupt thaw processes (1.2.2.4) and drastic hydrological changes, which are not fully understood yet. Hence, uncertainty exists about whether high-latitude regions might become wetter or drier in the future. They could turn into a wetter and cooler state with many freshwater systems and lakes, which support increasing land-atmosphere moisture recycling and cloud cover, reducing ground temperatures; or a drier state as newly formed lakes could drain, with less moisture recycling supporting less cloud cover and a warmer surface (Nitzbon et al., 2020; Liljedahl et al., 2016).

Which parts of the Arctic will be wetter or drier in the future is uncertain, but the differences between the potential Arctic hydroclimatic futures could be very pronounced. As recently shown by de Vrese et al. (2023), the drier and warmer permafrost state would lead to less sea ice, a reduced pole-to-equator temperature gradient, and a weaker AMOC. The drier Arctic state also shifts the position of the ITCZ, which results in higher precipitation in the Sahel region and potentially also in the Amazon rainforest. Increased forest and vegetation cover in these regions would be the consequence (de Vrese et al., 2023). Therefore, shifts in permafrost hydrology could affect climate tipping systems far beyond Arctic boundaries.

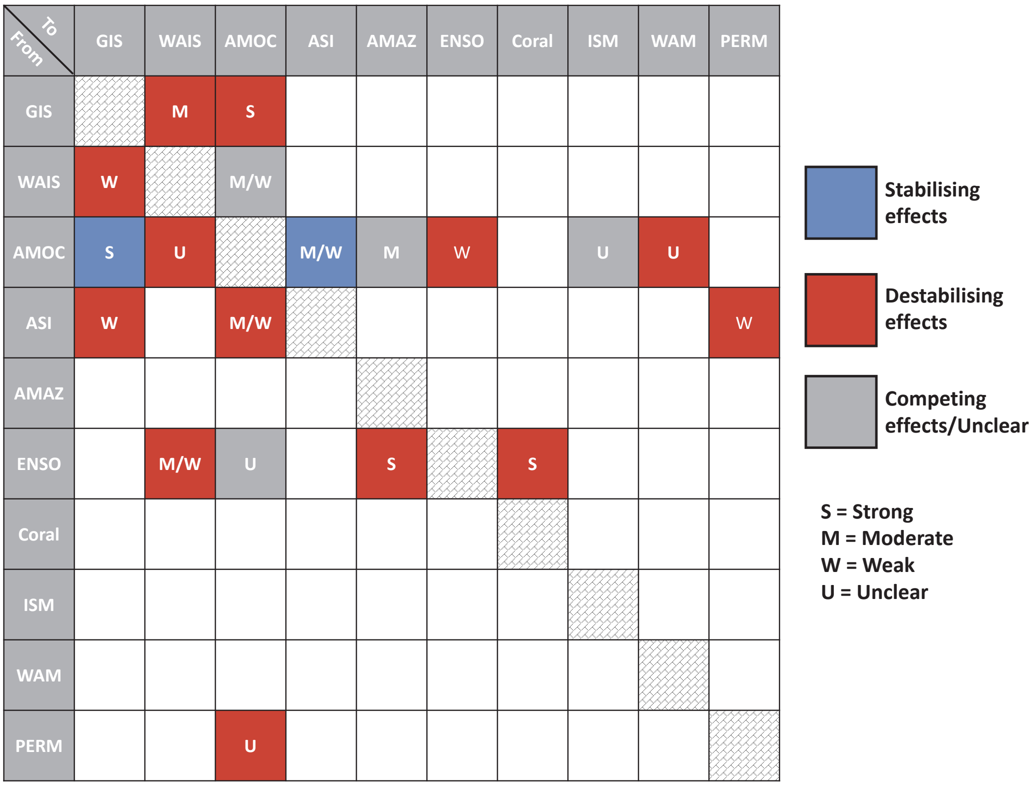


**Figure 1.5.3:** Matrix of links between elements (tipping systems and other nonlinear components) discussed in this chapter (see also Figs. 1 and 2). Columns denote the element from which the interaction originates, rows denote the tipping system to which element the interaction is pointing. We separate three different types of effects: A stabilising effect (blue box), a destabilising link (red box) and an unclear or competing link (grey box). White boxes denote no (or an unknown) link. Based on the recent literature, the strengths of the links are grouped into four groups: Strong (S), Moderate (M), Weak (W), and Unclear if a strength estimate is lacking (U). Abbreviations of the elements stand for: GrIS = Greenland Ice Sheet, WAIS = West Antarctic Ice Sheet, AMOC = Atlantic Meridional Overturning Circulation, ASI = Arctic Sea Ice, AMAZ = Amazon rainforest, ENSO = El Niño-Southern Oscillation, Coral = Coral reefs, ISM = Indian summer monsoon, WAM = West African monsoon, PERM = Permafrost. More details on each of the links can be found in Table 1 of the accompanying scientific review paper Wunderling and von der Heydt et al, from which this figure is adapted from.

## 1.5.3 Archetypal examples of interactions between tipping systems from a palaeoclimate perspective

### 1.5.3.1 Interactions in the distant past: the Eocene-Oligocene Transition

The formation of a continent-scale ice sheet on Antarctica during the 'Eocene-Oligocene Transition' about 34 million years ago is known as Earth's Greenhouse-Icehouse Transition. Following a cooling over tens of millions of years during the warm 'Eocene' period (c. 56 to 34 million years ago), this shift to a new cooler climate state in the 'Oligocene' period (c. 34 to 23 million years ago) would have been visible from space, as Antarctic forests were replaced by a blanket of ice and seawater receded from the continents, changing the shapes of coastlines worldwide. The climate transition had global consequences for Earth's flora and fauna, both in the oceans and on land (Hutchinson et al., 2020; Coxall et al., 2005).

This climate transition has been identified as a possible palaeoclimate example of cascading tipping points in the Earth system (Dekker et al., 2018; Tigchelaar et al., 2011). Examples of climate tipping systems in this case consist of the global ocean circulatory system, the Antarctic ice sheet, polar sea ice, monsoon systems and tropical forests. In a conceptual model, the first part of the Eocene-Oligocene Transition is attributed to a major transition in global ocean circulation, while the second phase reflects the subsequent blanketing of Antarctica with a thick ice sheet (Tigchelaar et al., 2011). The glaciation of Antarctica also produced a sea level fall of several tens of metres, causing shallow seaways to recede, turning many marine regions into continental habitats (Toumoulin et al., 2022; Lear et al., 2008), see Figure 1.5.4.

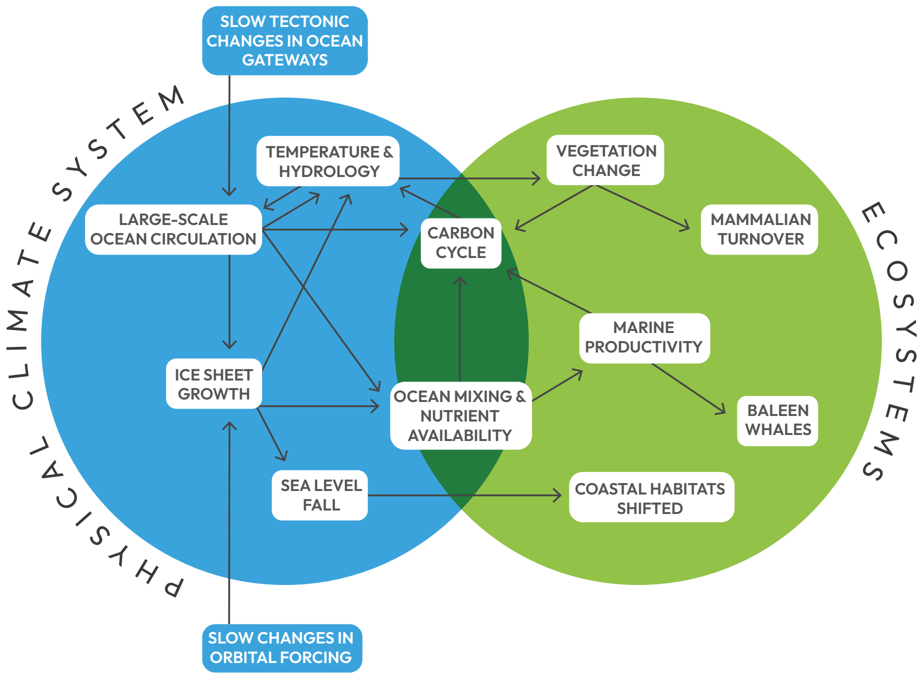


**Figure 1.5.4:** Conceptual linkages between changes in the Earth system associated with the Eocene-Oligocene Transition, 34 million years ago. External drivers were the slow changes in ocean gateways caused by tectonic plate movement, and slow changes in Earth's orbital configuration. The interactions and feedbacks within the Earth system act on different timescales, which makes the complete sequence of events complicated, but overall these processes resulted in Earth's Greenhouse-Icehouse Transition. There is a large uncertainty in all links portrayed. Adapted from: Wunderling and von der Heydt et al.

#### Ocean circulation

The global ocean circulatory system was showing tentative signs of change a few million years before the climate transition, likely caused by changing ocean gateways in the north Atlantic (Coxall et al., 2018). Isotope measurements suggest that a precursor to North Atlantic Deep Water reached the southern hemisphere close to the Eocene-Oligocene Transition, perhaps signalling the first onset of AMOC (Via and Thomas, 2006), but the exact timing remains uncertain.

#### Biosphere

Biomes in Earth's greenhouse state reflect warmer and wetter conditions than the icehouse state of the early Oligocene, but many of these seemed to have changed gradually as climate cooled in the Eocene, making it difficult to identify vegetation tipping systems following the glaciation of Antarctica (Hutchinson et al., 2020). The mammal fossil record, which is coupled to vegetation through diet, suggests more acute changes in the early Oligocene.

The Grand Coupure ('The Big Break'), is a long-known mammal extinction/origination event around the Eocene-Oligocene Transition, involving large-scale migrations of Asian mammals into Europe (Hooker et al., 2004). Thought to signal a combination of changing climate and floral changes, this abrupt faunal turnover might reflect the crossing of ecosystem tipping points caused by the crossing of a climate tipping point: a climate-biosphere tipping cascade.

In summary, Earth's Greenhouse-Icehouse Transition was likely associated with a range of interactions between components of the Earth system that are debated as potential tipping systems. Determining the extent to which these reflect a cascading series will require a major data-modelling effort, with improved correlations between marine and terrestrial records, and better constraints on the rate and magnitude of change within a range of tipping systems.

### 1.5.3.2 Interactions during and since the last glacial period

Here, we discuss three important palaeoclimate candidates for tipping interactions during and since the last glacial period.

**Dansgaard-Oeschger events**

Rapid, decadal-timescale Northern Hemisphere warming transitions known as 'Dansgaard-Oeschger' (D/O) events (Figure 1.5.5) occurred repeatedly during glacial periods throughout much of the late Pleistocene prior to the Holocene (Ganopolski and Rahmstorf, 2001). In general, these events consist of an abrupt (in the order of decades) warming from glacial to interglacial conditions, followed by gradual cooling over the course of hundreds to a few thousand years, before a rapid transition back to cold glacial conditions.

Evidence from Greenland ice cores and North Atlantic sediment records suggest that the abrupt cooling transitions were systematically preceded and possibly triggered by more gradual cooling across the high-latitude Northern Hemisphere (NGRIP project partners, 2004; Barker et al., 2015). The abrupt transitions from glacial to interglacial conditions were also preceded by more gradual changes elsewhere (for example, increasing Antarctic and deep ocean temperatures and decreasing dustiness; Barker and Knorr (2007)), leading to the idea that both types of transitions may be predictable to some extent (Lohmann, 2019; Barker and Knorr, 2016). Each event was also paired with rapid changes in ocean circulation, terrestrial hydroclimate, atmospheric composition and ocean oxygenation. The occurrence and interactions among many subsystems that show abrupt changes make it plausible then to consider it a cascade, and that such cascades are a common feature of late-Pleistocene climate variability.

During the abrupt warming phases of D/O cycles, an abrupt decrease of Arctic and North Atlantic sea ice cover likely contributed to the onset of convection and a rapid resurgence of a much weaker, and potentially even collapsed, AMOC (Gildor and Tziperman, 2003; Li et al., 2010: see 1.4.2.1). D/O-type changes in coupled climate models also feature a rapid disappearance of sea ice that precedes the abrupt AMOC strengthening (Vettoretti and Peltier, 2016; Zhang et al., 2014). Thus, the D/O warming events may potentially comprise a tipping cascade (Lohmann and Ditlevsen, 2021). However, such a cascading interaction may depend on the background climate state (i.e. only possible during glacial conditions), and it is unclear whether North Atlantic sea ice cover during the last glacial period can be considered a tipping system.

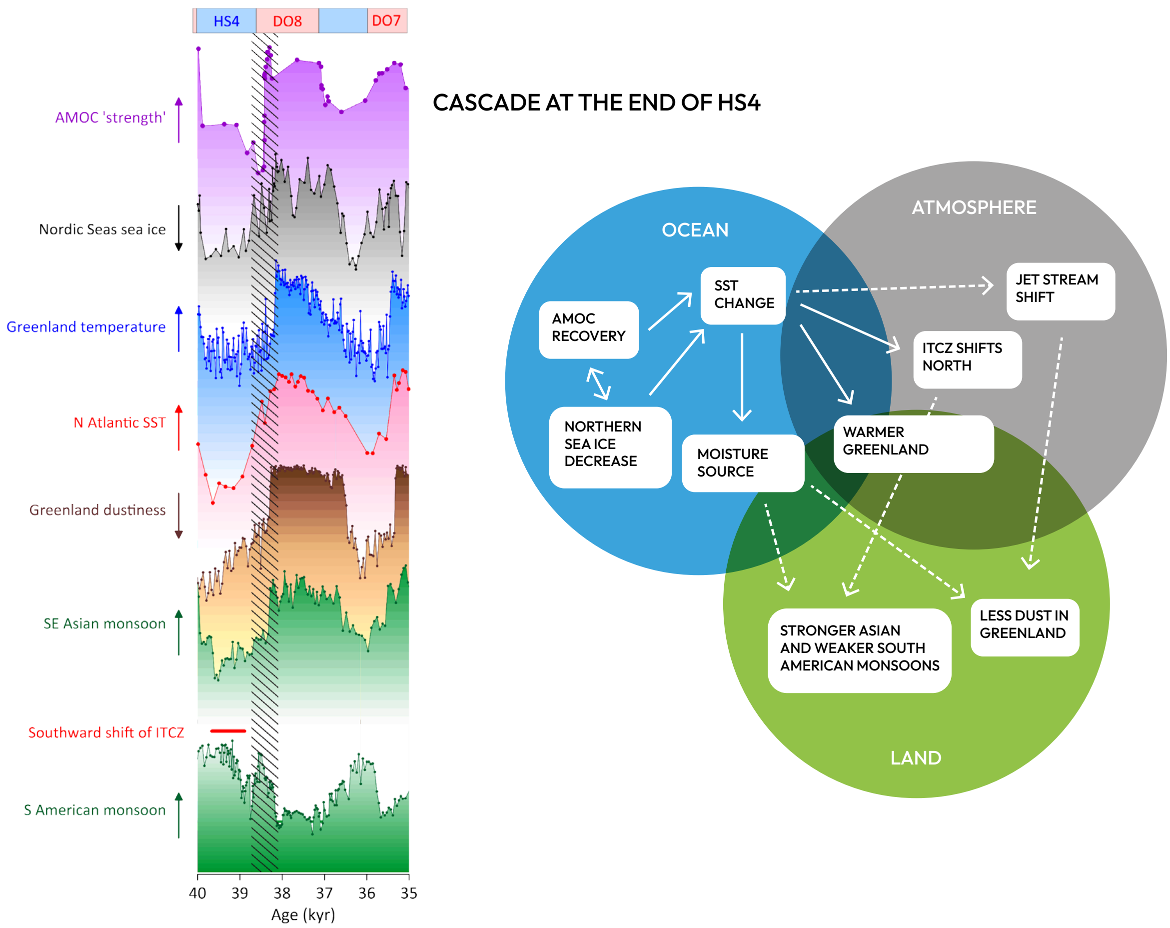


**Figure 1.5.5:** Interactions at the end of the Heinrich event 'Heinrich Stadial 4' (HS4). (a) Climate proxy indices spanning the transition from HS4 into Dansgaard-Oeschger (D/O) event 8 (time goes from left to right). From top to bottom: AMOC strength (Henry et al., 2016), Norwegian Sea ice cover (Sadatzki et al., 2020), Greenland temperature (North Greenland Ice Core Project members (NGRIP), 2004), North Atlantic SST (Martrat et al., 2007), Dust accumulation in Greenland (Ruth et al., 2007), Asian monsoon intensity (Cheng et al., 2016), South American monsoon intensity (Kanner et al., 2012). Horizontal red bar indicates period when ITCZ assumed a more southerly position (Wang et al., 2004). Hatched region spans the transition from HS4 to D/O8 and represents an estimate of the relative age uncertainty among the records shown (i.e. it is generally not possible to tell which changes occurred earlier or later within the overall sequence). Vertical arrows indicate the sense of increase for each parameter. (b) Interactions between ocean, atmosphere, and land during the end of HS4. Links with higher uncertainty are denoted by dashed arrows. Adapted from: Wunderling and von der Heydt et al.

**Bølling-Allerød**

Towards the end of the last 'ice age' glacial period, a very prominent climate event is recorded in numerous geological archives. The Bølling-Allerød (B/A) started 14,700 years ago with abrupt warming in the Northern Hemisphere (with temperature increase in Greenland by 10-14°C over a few years) in response to a reinvigoration of the AMOC (McManus et al., 2004) and lasted until 12,900 years ago. The B/A is an example of pronounced interactions between Earth system components and cascading impacts in the Earth system (Brovkin et al., 2021), potentially similar to a last D/O event during the ongoing deglaciation.

At the onset of the B/A, atmospheric $CO_2$ and $CH_4$ concentrations rapidly increased over a few decades (Marcott et al., 2014) in response to abrupt Northern Hemisphere warming and permafrost thaw (Köhler et al., 2014) and moisture changes (Kleinen et al., 2023). This was followed by fast changes in precipitation (e.g. Zhang et al., 2017) and vegetation composition (Novello et al., 2017; Fletcher et al., 2010). The trigger for the rapid amplification of ocean circulation and the associated abrupt impacts at the B/A transition has been a focus of debate, with opinions divided between an essentially linear response to the (possibly abrupt) cessation of freshwater forcing (Liu et al., 2009) versus a non-linear response to more gradual forcing (i.e. a tipping point – Barker and Knorr (2021); Knorr and Lohmann (2007); Chiessi et al. (2008)).

**Heinrich events**

While the exact causes and mechanisms of the B/A transition and D/O events are still under debate, Heinrich events are better understood. They occurred during some of the cold glacial phases mentioned above and were associated with major reorganisation of ocean circulation in the North Atlantic (for a review, see Clement and Peterson (2008)). During Heinrich events, large masses of ice were released from the Laurentide Ice Sheet, which at that point covered most of northern North America, leading to a dramatic freshening of the North Atlantic Ocean and enhanced suppression of deep-water formation and the AMOC (Henry et al., 2016). They can be understood as a phenomenon involving two tipping systems – the Laurentide Ice Sheet and the AMOC (referred to as 'binge/purge oscillator' – MacAyeal (1993)).

Heinrich events provide some, albeit not fully consistent, insights into the response of the Amazon rainforest to reductions in rainfall, and therefore shed some light on its resilience. Using isotopes from sediments, savanna intrusions into the Amazon rainforest have been found during repeated Heinrich events (Häggi et al., 2017). These intrusions occurred in northern Amazonia (Zular et al., 2019; Häggi et al., 2017) and validate the suggested decrease in precipitation over that region in response to AMOC weakening (Campos et al., 2019; see 1.4.2.3). While further palaeoclimate evidence showed that large parts of the Amazon rainforest were stable even when precipitation was relatively low (Kukla et al., 2021; Prado et al., 2013), in the present climate it is unclear how additional effects from deforestation (Zemp et al., 2017), future climate change (Wunderling et al., 2022) and increasing chances of fires (Drüke et al., 2023) will affect the stability of the rainforest in the future (1.3.2.1).

## 1.5.4 Interactions between tipping systems and planetary-scale cascades

Assembling the individual links mentioned in the sections before gives rise to the possibility of domino effect-style tipping cascades involving more than two elements. The likelihood of such domino effects clearly depends on the strengths of interactions between the tipping systems. These could lead to large changes at the regional and even planetary scale. A plausible palaeoclimate example are D/O events (section 1.5.3.2).

While unlikely, a major concern regarding the future may be that a cascade involving several tipping systems and feedbacks could lock the Earth system on a pathway towards a 'hothouse' state, with conditions resembling that of the mid-Miocene or even Eocene (around 4-5°C warmer, and sea level 10-60m higher compared to pre-industrial Holocene) (Burke et al., 2018; Steffen et al., 2018). Feedbacks that affect global temperature via albedo changes (through ice sheet or sea ice loss) and additional $CO_2$ and $CH_4$ emissions (through e.g. permafrost thawing or methane hydrates release) may lead to additional warming on medium to long timescales (Wunderling et al., 2020; Steffen et al., 2018). In a worst case (and unlikely) scenario, it has been speculated that a regional breakup of stratocumulus decks at atmospheric $CO_2$ levels above 1,200ppm could translate into a large-scale temperature feedback leading to a warming of roughly 8°C (Schneider et al., 2019; see 1.4.2.4).

Timescales are crucial when discussing hothouse scenarios. A potential hothouse state in the next few centuries seems implausible in light of the current state of research. For example, in climate projections up to 2100, CMIP6 models show no evidence of nonlinear responses on the global scale. Instead, they show a near-linear dependence of global mean temperature on cumulative $CO_2$ emissions (Masson-Delmotte et al., 2021). Similarly, in a recent assessment, it is concluded that a tipping cascade with large temperature feedbacks over the next couple of centuries remains unlikely and that, while the combined effect of tipping systems on temperature is significant for those timescales, it is secondary to the choice of anthropogenic emissions trajectory (Wang et al., 2023).

However, this does not completely rule out the possibility of a hothouse scenario in the longer term. Indeed, tipping events are not necessarily abrupt on human timescales. Positive/amplifying feedbacks could have negligible impacts by 2100, for example on global mean temperature and sea level rise, but still influence Earth system trajectories on a timescale of thousands of years (Kemp et al., 2022; Lenton et al., 2019; Steffen et al., 2018). Overall, this calls for experiments across the model complexity hierarchy. Earth system models of intermediate complexity in particular, and atmosphere-ocean general circulation models at coarse spatial resolution, offer an interesting trade-off as they include representations of most tipping systems while still allowing for long-term simulations.

Finally, spatial scales and patterns are relevant when it comes to risks of hothouse scenarios. Most examples of tipping cascades from palaeoclimate suggest that, while impacts are clearly global (e.g. greenhouse-icehouse transition, D/O events), the spatial expression of climate change (weather extremes, precipitation, seasonality) can vary greatly across the globe. Nevertheless, for societies, such cascades can be as dangerous as a global hothouse scenario, as are tipping cascades that do not lead to a hothouse but lock in other major harmful impacts such as a 'wethouse' scenario of tens of metres of sea level rise.

## 1.5.5 Final remarks

As anthropogenic global warming continues, tipping systems are at risk of crossing critical thresholds (Armstrong McKay et al., 2022). Several assessments have investigated the risk of crossing critical thresholds of individual tipping systems, whereas interactions between tipping systems are only more recently taken into account, mostly by conceptual models (e.g. Sinet et al., 2023; Wunderling et al., 2023b; Dekker et al., 2018).

Based on the current state of the literature, we conclude that tipping systems interact across scales in space and time (see Figure. 1.5.1 and 1.5.2), spanning from subcontinental to nearly planetary spatial scales and timescales from sub-yearly up to thousands of years. We find that many of the discussed interactions between tipping systems are of a destabilising nature (Figure 1.5.3), implying the possibility of cascading transitions under global warming. Of the 19 discussed interactions, 12 are assessed as destabilising, two are stabilising, and five are unclear (see Figure 1.5.1). Assessing the overall stability of the Earth system, and the possibility of a chain of nonlinear transitions, will however require more detailed assessments of their interactions, strengths, timescales and climate state-dependence.

While there is increasing research on individual thresholds of climate tipping systems, substantial uncertainties prevail in the existence and strength of many links between tipping systems. In order to decrease such uncertainties, we propose three possible ways forward:

(i) Observation-based approaches: Satellite observations, reanalysis and palaeoclimate datasets may be evaluated using correlation measures (Liu et al., 2023), or advanced methods of inferring causality (e.g. Runge et al., 2019; Kretschmer et al., 2016; Runge et al., 2015). In-situ monitoring is also very important for most of the tipping systems as well, and in particular for the biosphere (see Chapters 1.3 and 1.6).

(ii) Earth system model-based approaches: With recent progress, Earth system models of full or intermediate complexity could be used to evaluate interactions between climate tipping systems in detail at the process level, and quantify their interactions using specifically designed experiments (see Chapters 1.2, 1.3, and 1.4).

(iii) Risk analysis approaches: Since relevant parameter and structural uncertainties are large within Earth system models, analysing model ensembles with a considerable number of ensemble members is very helpful in order to comprehensively propagate uncertainties for risk assessments (Daron and Stainforth, 2013; Stainforth et al., 2007; Murphy et al., 2004).

(iv) Finally, all three approaches above have their limitations, and could probably benefit from direct expert input. Therefore, expert elicitation exercises on tipping system interactions remains of high value to update and move beyond early investigations of this kind (Kriegler et al., 2009).

To summarise, the approaches above (and likely more) are required to obtain more reliable estimates of the existential risks potentially posed by tipping events or even cascades (Kemp et al., 2022; Jehn et al., 2021). They could be used to inform an emulator model for tipping risks, taking into account properties of individual tipping systems as well as their interactions. In addition, there also exist large uncertainties, not only among the known interactions as discussed above, but also because not all interactions are known or quantified (i.e. known unknowns versus unknown unknowns).

Further, in certain systems there are forcings of non-climatic origin that could interact with climate change and lead to tipping, and thus to interactions and possibly cascades with other systems. For instance, land use change and specifically deforestation are threatening the Amazon and decreasing its resilience to climate change (e.g. Staal et al., 2020; Boulton et al., 2022) (1.3.2.1). Lastly, systems do not necessarily tip fully in one go, but can also have stable intermediate states (such as through the formation of spatial patterns). This has mostly been reported in ecological systems, but is not limited to them (Rietkerk et al., 2021; Bastiaansen et al., 2020).

Taken together, assessing and quantifying tipping system interactions better has great potential to advance suitable risk analysis methodologies for climate tipping points and cascades, especially because it is clear that tipping systems are not isolated systems. The relevance for developing such risk analysis tools to assess tipping events and cascades is clear given the potential for existential risks and long-term irreversible changes (Kemp et al., 2022).

# Chapter 1.6 Early warning signals of Earth system tipping points

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## Summary

This chapter focuses on the methods used to predict the movement of parts of the Earth system towards tipping points. It begins by introducing the theory of critical slowing down (CSD), a general phenomenon of slowing recovery from perturbations that happens in many systems being forced slowly towards a tipping point. Then, it describes the various methods that can be used to estimate the occurrence of CSD and the approach of a tipping point, beginning with methods based on changes over time in the system, spatial changes, or changes in network structure, up to more advanced modelling techniques, including AI.

These 'early warning signals' (EWS) can be used on data from a number of different sources, be these models, field experiments or remotely sensed data from satellites. The chapter considers various case studies that use real-world observations, to show how these methods are being used to predict losses in resilience in these systems. Finally, it explores limitations and potential solutions in the field of EWS, looking ahead to advances in data availability and what this could mean for predicting the movement towards tipping in these systems in the future.

## Key messages

- Early warning signals can be used to detect the potential movement of Earth's systems towards tipping points.
- The central western Greenland Ice Sheet, Atlantic Meridional Overturning Circulation, and Amazon rainforest all show evidence of loss of resilience consistent with moving towards tipping points.

## Recommendations

- EWS can provide an indication of a tipping point approaching, and should be taken as a chance to prevent it from happening.
- Results from models need to be leveraged to identify which specific variables are most likely to display EWS as tipping points approach, so that these can be monitored with empirical data.
- Further investigation is required to explore the utility of machine learning for EWS and to detect the drivers of conventional EWS.
- Openly available datasets from on-the-ground sensors and measurements as well as remote sensing products provide an avenue for this EWS detection, however careful consideration is required to ascertain which variables are most appropriate and the limitations of existing remote sensing data.
- Future work should look to design remote sensing studies and data acquisition strategies that minimise the potential for biassing EWS such that false indications occur.

## 1.6.1 Theory and methods of early warning signals

While tipping points are often abrupt, rapid and irreversible, and may come as a surprise after only modest and smooth changes beforehand, they are not always unpredictable. Given their potential for disruption, there have been numerous attempts to identify when a system may be losing resilience and approaching a tipping point. These approaches, often called EWS, rely on monitoring the changes in the underlying behaviour of these systems across time and/or space prior to a transition. While these indicators are well grounded in theory, there are limitations to consider when transferring them to real-world systems.

Here, we introduce the theory of **critical slowing down** – the phenomenon that allows most of the EWS detailed here to be used. We then go into detail about the various methods used to predict the movement towards tipping points. These concepts are illustrated with real-world case studies from targeted climate and ecological systems. Finally, we explore some limitations of these methods, some potential solutions, and look ahead to potential future research in this field.

### 1.6.1.1 Theory of critical slowing down

The majority of studies on early warnings of Earth system tipping points are based on searching for evidence of CSD. Essentially, if a system is forced towards a tipping point, the state it currently occupies starts to lose its stability as the restoring feedbacks that 'pull' the system back to that state after it is perturbed start to weaken. If the system is forced sufficiently slowly that it can remain close to steady state, this causes the system to respond more sluggishly to short-term perturbations, and thus 'slow down' (Wissel, 1984).

Figure 1.6.1 shows this concept visually using the 'ball in potential well' analogy. When the system is more stable (represented by the well with steeper sides) recovery from any given perturbation is faster (the ball returns faster). A system closer to tipping (represented by a shallower well) has a slower recovery from the same perturbation (the ball takes longer to return). Eventually, the restoring feedbacks of the system become so weak at a tipping point that the stability of the initial state is lost, and the system moves to a new stable state. Before that point a random disturbance may cause the system to exit its initial state early.

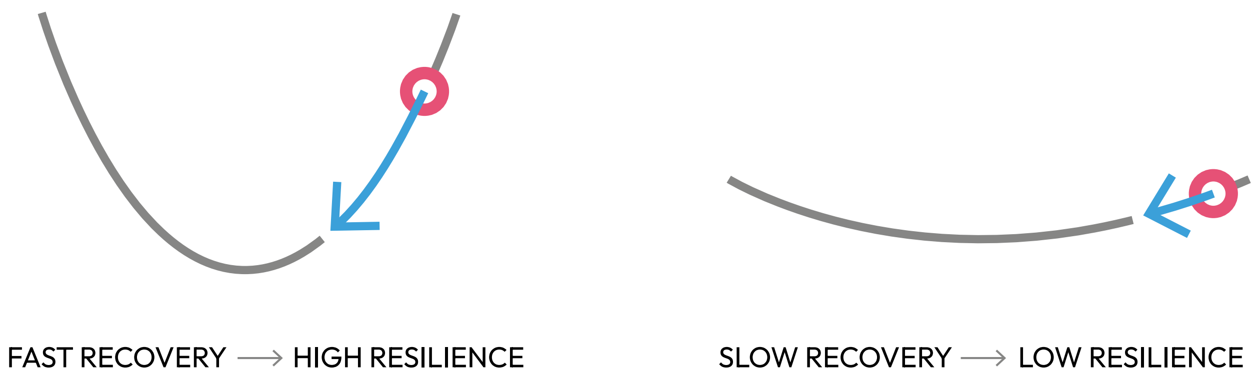


**Figure 1.6.1:** Using the 'ball in the well' analogy to compare a system that is (left) far from tipping, and (right) close to tipping. The system that is further away from tipping recovers faster from perturbations, the steeper sides of the well describing the stronger restoring feedbacks of the system. Close to tipping, the sides of the well are shallower, such that the system will take long to return from the same perturbation as the restoring feedbacks are weaker. Adapted from: Dakos et al. (2023).

The occurrence of CSD prior to a critical transition has been identified across numerous domains (Kubo, 1966; Kawasaki, 1966; Ferrell, 1970; Wissel, 1984; Dakos et al., 2023). In most cases, it mathematically involves the leading 'eigenvalue' of the system (which describes the strength of damping negative feedback) approaching 0 from below. However, in reality we typically do not have the equations that govern the system's dynamics, and as such have to estimate the occurrence of CSD with methods detailed in this chapter.

### 1.6.1.2 Temporal methods

One way to detect CSD is to measure the rate at which a system returns to its initial state following known disturbances. A resilient system with strong restoring feedbacks will return to its initial state faster than one which is near to a tipping point (Wissel, 1984). However, this method requires the occurrence of well-defined perturbations, as well as clear knowledge of when the equilibrium state of the system has been reached again, neither of which are always clearly defined in the real world. Hence statistical techniques are often used to detect CSD behaviour in the form of resilience loss of a system (resilience defined in this chapter as the ability to return to the equilibrium state) prior to a tipping point.

As a system approaches a tipping point and its recovery slows down, the system at each time step *t* is more correlated to the previous timestep, *t-1* (as shown in Figure 1.6.2). This can be measured with 'lag-1 autocorrelation' (or AR(1)) which measures a system's self-similarity through time, and tends towards 1 as a system experiences CSD prior to tipping to a different state (Scheffer et al., 2009); Visually, this can be viewed by observing a scatterplot of a section of the time series data of the system against the same section of time series lagged by one time point (Figure 1.6.2). When the system is far from tipping (left column Figure. 1.6.2), there is no relationship between the system now and with itself at the previous time step (i.e. low AR(1)). As the system approaches the tipping point, CSD means that there is a stronger correlation between the system now and with itself at the previous time step (and thus a higher AR(1)). Larger deviations in the red section of the time series can be seen, further showing this slowing down and increase in AR(1).

Similarly, as resilience is lost, a given perturbation will cause a greater movement of the 'ball' from Figure 1.6.1, meaning the variability (measured as 'variance') of the system is expected to increase. The system can sample more of its 'state space' (all the possible states the system can be in) due to the shallower well (Scheffer et al., 2009). Theory shows that because of CSD, AR(1) and variance should increase together in a characteristic fashion where their ratio remains constant. Hence an increase in both variance and AR(1) should be sought for robustness. However, there are other factors which can lead to a change in variance, such as a change in the variance of the system's external forcing (Ditlevsen and Johnson, 2010).

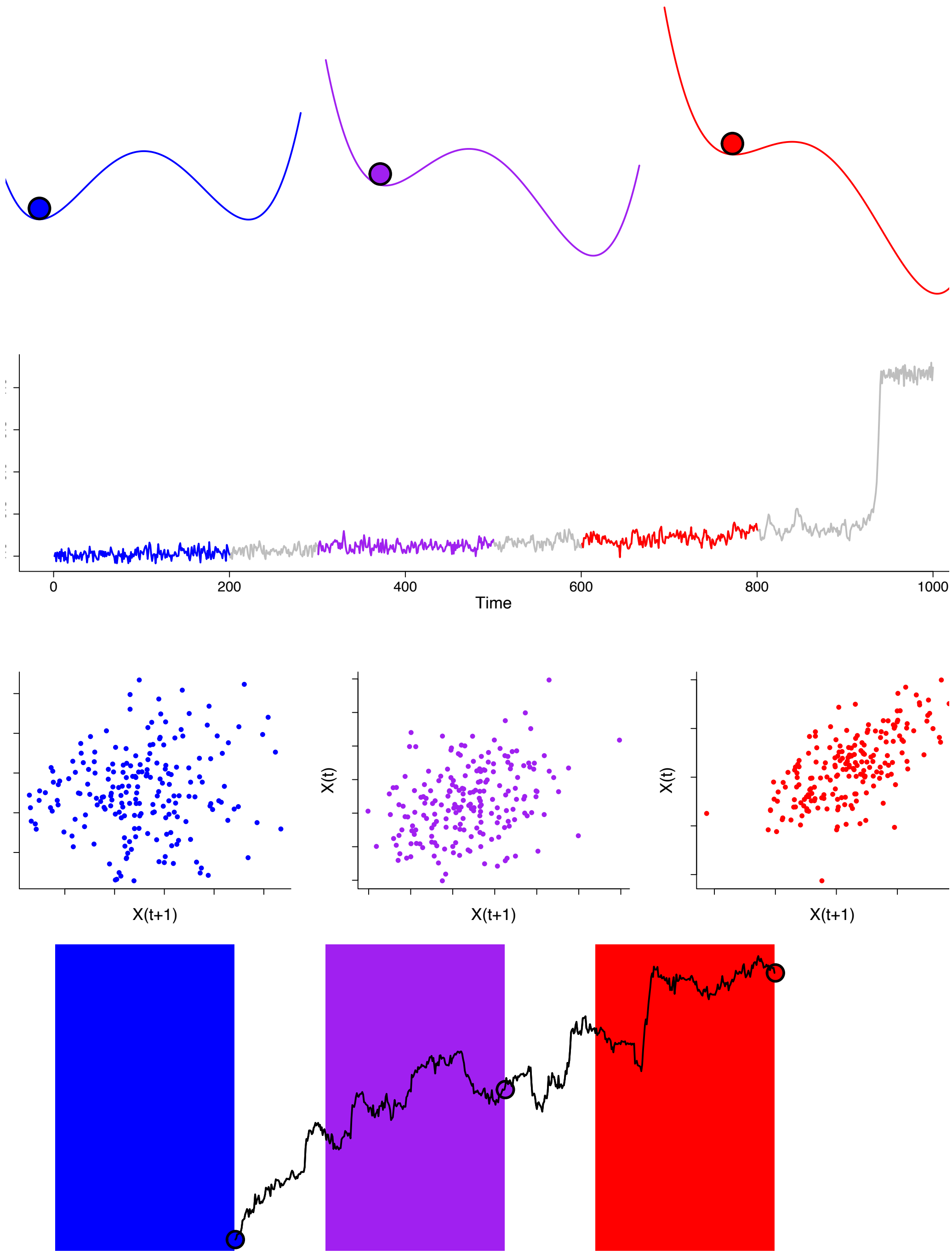


**Figure 1.6.2:** A comparison of the lag-1 autocorrelation (AR(1)) for a system that is far from tipping (blue), getting close to tipping (purple), and close to tipping (red). As the systems represented by the time series data approaches tipping (top row), there is no correlation between the time series and itself at the previous time point in the blue part of the time series, far from tipping. However, closer to tipping, in the purple and then red regions of the time series, there are correlations and thus higher AR(1) values. In the time series itself there are clear deviations towards the end compared to the beginning, suggesting CSD is occurring as the tipping point approaches. The EWS are calculated on a moving window (coloured regions in bottom plot). Here, AR(1) is shown at the end of the window used to calculate it, with examples shown as coloured points to match those windows on the detrended time series. Adapted from: Dakos et al. (2023).

For these time-based measures to be robust they require the time series data of the system to be stationary (i.e. the data's statistical properties do not depend on the time at which the data is observed). Multiple methods exist for removing non-stationarity, typically by detrending the time series (i.e. subtracting any trends). The EWS is then calculated on this detrended time series on a short section that shifts one time point at a time before recalculating (a 'moving window') (Figure. 1.6.2).

In our example tipping point, as the current steady state of the system is losing resilience and the probability to shift to an alternative state increases, the distribution of states of the system is expected to become increasingly skewed toward the alternative state (because of the increasing asymmetry of the potential well, see (Figure. 1.6.2). This can be quantified by the 'skewness' of the system, again measured on a moving window as described above. A change in skewness itself is not linked to CSD, but can be used as an EWS. 'Flickering' may also be observed before a tipping point, where sufficient noise can push a system temporarily into an alternative state before returning to the original with increasing likelihood as the system is approaching tipping (Wang et al., 2012; Dakos et al., 2013).

Once an EWS indicator has been calculated across a time series dataset, its tendency can be measured to determine if there is a movement towards tipping. Kendall's tau correlation coefficient is a common way to estimate trends. It is 1 if the time series of the indicator is always increasing (every value in the time series is higher than all of the previous values), -1 if it is always decreasing, and 0 if there is no overall trend. The significance of this trend can be calculated using null models that resample the time series such that the statistical properties of the system are maintained (e.g. mean and overall variance) but the memory of the system, such as changes in AR(1) and variance over time, are destroyed (Dakos et al., 2008). From these, the significance or p-value of an observed trend can be calculated. In practice, different detrending techniques and window lengths are used to test the robustness of EWS.

Particularly for temporal methods, there is a need to carefully consider the different timescales of the system. For EWS to work, a slow external forcing should move the system towards tipping, while faster 'noise' processes push the system away from equilibrium in either direction (noise can be thought of as short-term weather events in the climate system). This allows us to measure the return rate and other indicators on a time series from the system as it is monitored somewhere in between these timescales. The indicators may fail if, for example, the system is forced too fast towards tipping. This is further detailed in the Limitations section (1.6.1.6).

### 1.6.1.3 Spatial methods

Spatial analogues of the time-based EWS above exist too, allowing spatial information about a system within a single time step to be used. This can be considered as a space-for-time substitution, and eliminates the need for enough long-term data to have a moving window to measure AR(1) or variance on.

As a system approaches a tipping point, responding more sluggishly to external perturbations and sampling more of the state space, it is expected that there will be higher spatial autocorrelation and variance (Kéfi et al., 2014). This can be calculated, for instance, by Moran's I (Kéfi et al., 2014) and spatial variance (Guttal and Jayaprakash, 2009). The change in skewness observed in time series data also has a spatially analogous statistic (Guttal and Jayaprakash, 2009), noting again that this is not specifically related to CSD.

Some ecosystems have a clear self-organised spatial structure (e.g. drylands, peatlands, salt marshes, mussel beds; Rietkerk and van de Koppel, 2008). The emergence of such spatial patterns is thought to increase their resilience (Von Hardenberg et al., 2001), and could even allow them to evade tipping points altogether (Rietkerk et al., 2021). The size and shape of these patterns have been shown to change in a consistent way along stress gradients and have been suggested to be good candidate indicators of ecosystem degradation (Von Hardenberg et al., 2001; Rietkerk et al., 2002; Kéfi et al., 2007).

One of the most studied examples is the case of dryland ecosystems, where changes in the shape of the patch size distribution could inform us about the stress experienced by the ecosystem (Kéfi et al., 2007). As the stress level increases, the larger vegetation patches in the system fragment into smaller ones, which leads to a change in the shape of the patch size distribution from power law-like to a truncated power law (Kéfi et al., 2011). A number of metrics can be used to quantify the shape of the patch size distribution, such as the parameters of the best fit (e.g. the slope of the power law fit), the size of the largest patch in the system, or the power law range (Kéfi et al., 2014; Berdugo et al., 2017). We note that the use of spatial EWS is also dependent on some knowledge about the underlying system's spatial feedbacks (Villa Martín et al., 2015).

### 1.6.1.4 Network methods

Another way to monitor resilience loss in systems is to conceive of them as a network. In a spatial system, this would involve edges connecting neighbouring points. This framework can be applied to other, non-spatial, systems which are not necessarily linked in space but through other variables.

Multivariate systems (i.e. systems with multiple measurable variables) can pose problems for early warning signals. For instance, two different variables may give conflicting information, or obscure a clear signal (Boerlijst et al., 2013; Weinans et al., 2021). Multivariate systems relevant for climate science include examples such as interaction networks with different plant or animal species, or spatial systems where every grid cell can be represented as a variable in the system or a node in the network (Tsonis and Roebber, 2004; Donges et al., 2009).

Changes in network structure can show an approaching tipping point and have been observed in some systems, including climate (Lu et al., 2021) and lake systems (Wang et al., 2019). More generally, monitoring structural changes properties (e.g. connectivity, node centrality) in network systems (i.e. a network of interacting components, such as spatially connected sites, interacting actors, or species in a community (Mayfield et al., 2020; Cavaliere et al., 2016; Yin et al., 2016) can be used for EWS. Alternatively, correlations in time between components in multivariate systems has been used to construct an interaction network and analyse its structural properties (Tirabassi et al., 2014).

Once the nodes – or variables – are chosen, there are a number of ways the analysis can proceed. One such method evaluates network statistics. To create a network, the method calculates if the correlation between each set of two nodes is above a predetermined threshold and, if it is, connects the two nodes with an edge (a network connection). If this analysis is repeated on a moving window (measuring the correlation between two variables on a moving window like the temporal EWS), changes in the network topology (i.e. the arrangement of node connections) over time can be used as EWS. For instance, as the system moves towards a tipping point, the network will display a higher number of connections between nodes and an increase in variance in connections (Kuehn et al., 2013).

Unlike spatial methods, which examine a 'snapshot' of the system at a given time, these methods require the use of a time window to measure the changing structure on, and thus reasonably complete time series are needed. Another possible disadvantage is that, in some networks, the edges do not necessarily have a physical foundation (Ebert-Uphoff and Deng, 2012). Recent research explores a complementary approach where causal links are calculated instead of correlation links and the strength of the causal link works as the indicator of resilience (Nowack et al., 2020; Setty et al., 2023).

Alternatively, 'dimension reduction' techniques can capture overall network dynamics into a representative statistic. For instance, Principal Component Analysis (often referred to as 'Empirical Orthogonal Functions' (EOF) in climate science) can be used on a time series to get directions of change (Held and Kleinen, 2004; Weinans et al., 2019), although these linear projections may eliminate existing tipping points so care must be taken. It can often be used in spatial systems to detect the leading mode of variability over a region, such as a climate index like the Pacific Decadal Oscillation (Mantua and Hare, 2002). Next, data can be projected onto the leading principal component, effectively yielding a univariate (i.e. single variable) time series on which time-based univariate EWS can be calculated (Held and Kleinen, 2004; Bathiany et al., 2013; Boulton and Lenton, 2015).

From a network point of view, this analysis does not make any a priori assumptions about the interactions between the different network nodes, and is therefore quite flexible in its use. However, it requires large amounts of high-quality data to yield accurate results. The underlying assumption is that, as the system approaches the tipping point, the dynamics become more correlated, leading to a high explained variance of a PCA and clear directionality in the dynamics (Lever et al., 2020).

### 1.6.1.5 Model methods

As well as statistical and network methods that look for changing dynamics in a system, more complex methods can predict movement towards tipping points. One example is a generalised model approach, which integrates knowledge about the system into models and may allow us to estimate, for example, changes in the leading eigenvalue of the system once small model assumptions have been made (Lade and Gross, 2012). System-specific indicators can also be derived where understanding about processes in the system can help us to assess its resilience in novel ways (Boulton et al., 2013).

Machine learning (ML) techniques are now being applied to tipping point prediction. The documented success of neural networks for time series classification problems has inspired the development of similar ML methods specifically for EWS detection. There is a natural synergy to this approach in that the same CSD phenomena manifest across a wide range of systems approaching critical transitions, so the notoriously data-intensive task of training a neural network can be accomplished using plentiful synthetic data and still produce a result which can be applied to observational data (which is often more scarce and harder to label).

Deep learning models (which combine convolutional neural network layers with recurrent Long Short-Term Memory modules) have shown promise for EWS detection, outperforming methods using traditional statistical indicators (variance, AR(1), etc.) on a variety of test cases both real and simulated (Bury et al., 2021; Deb et. al., 2021). Furthermore, these models have exhibited success in inferring the *type* of oncoming bifurcation from observed pre-transition dynamics, and have performed well on rapid transitions in simple spatial models that evolve over time (Dylewsky et al., 2023).

Other ML techniques can also tell us something about how far systems are from tipping. For example, the 'random forest' method could be used to determine the factors that determine the AR(1) value in different areas of vegetation, and thus how close to tipping these areas could be, based on driving variables (Forzieri et al., 2022). Combining traditional EWS and ML techniques could provide some of the best prospects for monitoring systems that may be approaching tipping.

### 1.6.1.6 Limitations

There are some limitations to the EWS methods detailed here. These include, but are not limited to, availability of data used to monitor the systems, and the properties of the system assumed to be able to measure the EWS, such as the presence of a tipping point and the underlying timescales of the system.

Most importantly, it is worth noting that, for EWS to be used appropriately as an early warning of a tipping point, we require prior independent evidence that the system in question can actually exhibit tipping behaviour, as opposed to losing resilience with no tipping point or alternative state for the system to tip to. This evidence could come from established theory, models, or palaeo data. A subset of EWS can be used to monitor resilience in systems where a tipping is not necessarily expected, however these are not discussed here.

To make a robust assessment with EWS of a system, there is a requirement for high-quality data. Temporal EWS require a complete time series dataset which is sufficiently long to capture the relevant timescale of the system; infilling missing data points (which can be common in observational records) can interfere with the EWS, while shorter time series may not accurately detect changes in resilience.

Even with suitable amounts of data, there are inherent limitations associated with EWS. While theoretically, AR(1) should equal 1 when a system reaches a tipping point, these realworld systems are exposed to noise and can tip prior to this. Furthermore, the act of detrending the time series in the process calculating EWS changes the absolute value of AR(1). This means that, while EWS might tell us when a system is losing resilience, without a sufficiently dense dataset and knowledge about internal dynamics of the system, they cannot usually give a measure of the distance to a tipping point. However, robustly checking the tendency of these indicators (such as Kendall's tau) while varying the detrending technique and window length used to calculate the indicator on can provide useful information on the movement towards tipping.

Usually, it is assumed that a system approaching a tipping point is forced slowly towards it, and forced on shorter timescales by perturbations (which can be thought of as like weather in climate systems). It is generally assumed that this short-term noise is independent and identically distributed with a mean (average) of zero. This is unlikely to be the case in reality, with climate systems experiencing extreme weather events, for example, which are likely becoming more prevalent with the changing climate. Furthermore, extreme weather events would also increase the variance of the short-term noise over time, which also hampers the ability to use EWS indicators. To tackle this, we propose measuring EWS on the drivers themselves (e.g. on rainfall for vegetation systems) to check if changes in autocorrelation and variance in these are related to those found in the system being monitored.

Remote sensing products from satellite observations are a great resource of generally freely available data for using EWS on, and can enable complementary analyses to on-the-ground measurements of things that cannot be measured from space. They provide long records of climate systems, allowing us to create a long enough EWS indicator from which to get reliable results. However, due to sensor degradation and upgrades, it can be challenging to get a long time series from a single sensor, and products are often created from combined data sources. This can interfere with the EWS that we have described here, particularly AR(1) and variance, if this merging changes the signal-to-noise ratio (SNR) over time.

Newer sensors will measure with greater accuracy, increasing the SNR and in turn 'erroneously' increasing the AR(1) as far as an EWS is concerned, and a decrease in variance would also be expected. Anticorrelation between these two measures can show this is happening, whereas theory dictates that we should see an increase in both for a true EWS. In addition, newer remote sensors will also present shorter revisit times, as well as improved spatial resolutions, imposing the need to carefully consider the way data from different sensors are combined to produce long time series. Recently, we have become more aware of the effects of merging sensors and can prepare our analysis of these accordingly (Smith et al., 2023), such as only using data from a single sensor (Blaschke et al., 2023).

As well as questions around data availability and noise behaviour, the inherent timescale of the system being studied can hinder our ability to predict tipping points. While tipping is by definition a fast process, for slower-moving systems like the AMOC, the tipping event occurs over decades and it could therefore be difficult to detect the tipping point using EWS. Another example of this is the Amazon rainforest, where there is a slow decadal response of the forest based on climate change (1.3.2.1). It could take decades for dieback to occur even under a constant climate, such that a tipping point could be passed long before it is observed.

Part of the assumptions made around the occurrence of these EWS is that the system will approach a 'bifurcation' (a mathematically specific and common form of tipping point), rather than alternate forms of tipping. Alternatives include noise-induced tipping, where a system is shifted outside its stable state by a 'stochastic' (i.e. random) forcing, or rate-induced tipping, whereby a parameter changes too rapidly for the system to stay in the stable state (Ashwin et al., 2012). Rate-induced tipping can show some EWS (Ritchie and Sieber, 2016), such as threshold exceedance (detailed further in Chapter 2.5), while noise-induced tipping is generally unpredictable. For example, if the system is perturbed by something like an extreme weather event (e.g. a drought in the Amazon rainforest) such that it causes tipping by pushing the system past the ability for restoring feedbacks to return the system back to the previous state, CSD will not occur. However, bifurcation tipping and noise-induced tipping can be linked, whereby a system losing resilience approaching a bifurcation is more likely to be pushed to an alternate state by noise.

A related problem that may hamper EWS detection is that of cascading tipping points (see Chapter 1.5), where a tipping point in one system has a knock-on effect on another system, causing that to also tip. This can make it difficult for EWS to detect these tipping points, especially if the cascade causes instantaneous tipping points (a 'joint cascade') or happens soon after the first system tips (a 'domino cascade') (Klose et al., 2021).

**Box 1.6.1: Use of early warning signals beyond climate and ecological systems**

The EWS detailed here are not limited to use in climate and ecological systems; a recent study identified their use in other fields such as health, social systems and physical sciences (Dakos et al., 2023). Their utility in other domains is considered in later chapters in this report, specifically Chapter 2.5 - '*Early warnings of tipping points in socio-economic systems*' and Chapter 4.5 - '*Detecting "early opportunity indicators" for positive tipping points*'. Particular EWS of note which are used in these chapters include:

**Lag-1 autocorrelation (AR(1)) –** estimating critical slowing down (CSD), AR(1) in expected to increase as the restoring feedbacks of the system degrade such that the system slows and can be thought of as 'today is becoming more like yesterday'.

**Variance** – increases in variance are also expected due to CSD, as the system is able to sample more of the state space with weakened restoring feedbacks.

**Skewness** – not caused by CSD, skewness is expected to change as the potential well of the system changes shape, such that deviations or events become more pronounced towards the direction of tipping.

Full details of the indicators can be found in the relevant sections.

## 1.6.2 Case studies of empirically measured EWS

The EWS proposed above have been searched for in a number of real-world cases, using data from sources ranging from remotely sensed products from satellites to growth layers in marine bivalve shells.

A systematic review of academic papers that mention phrases associated with 'early warning' and 'tipping point', which we further filter based on using empirical data only, yields 229 studies, of which 33 are associated with the climate (Dakos et al., 2023); 22 of these climate studies find positive EWS, 1 negative, 9 mixed (from calculating EWS on different records and having conflicting results) and 1 inconclusive. These climate studies are further subsetted into palaeoclimate (12 total, 9 positive, 1 mixed, 1 negative), cryosphere (6 total, 3 positive, 2 mixed, 1 inconclusive), weather (3 total, 2 positive, 1 mixed), and modern climate, including AMOC collapse, El Niño, and monsoons, etc (12 total, 8 positive, 4 mixed). Overall, the most commonly used EWS are temporal AR(1) (17) and temporal variance (17 also, 13 of these using both together). Further details can be found in Dakos et al. (2023), and discussion of EWS beyond climate and ecological systems in Box 1.6.1.

Figure 1.6.3 below shows which climate systems have had studies searching for EWS of potential tipping points using empirical data. Below we detail some of these case studies specifically. We discuss where models suggest we may see EWS of climate tipping points and cases where empirical data has shown a loss of resilience in these systems. However, not all potential tipping points in the climate system have shown EWS in empirical data. In many cases this is due to observations being unavailable or the records being too short to see a significant movement towards tipping using EWS.

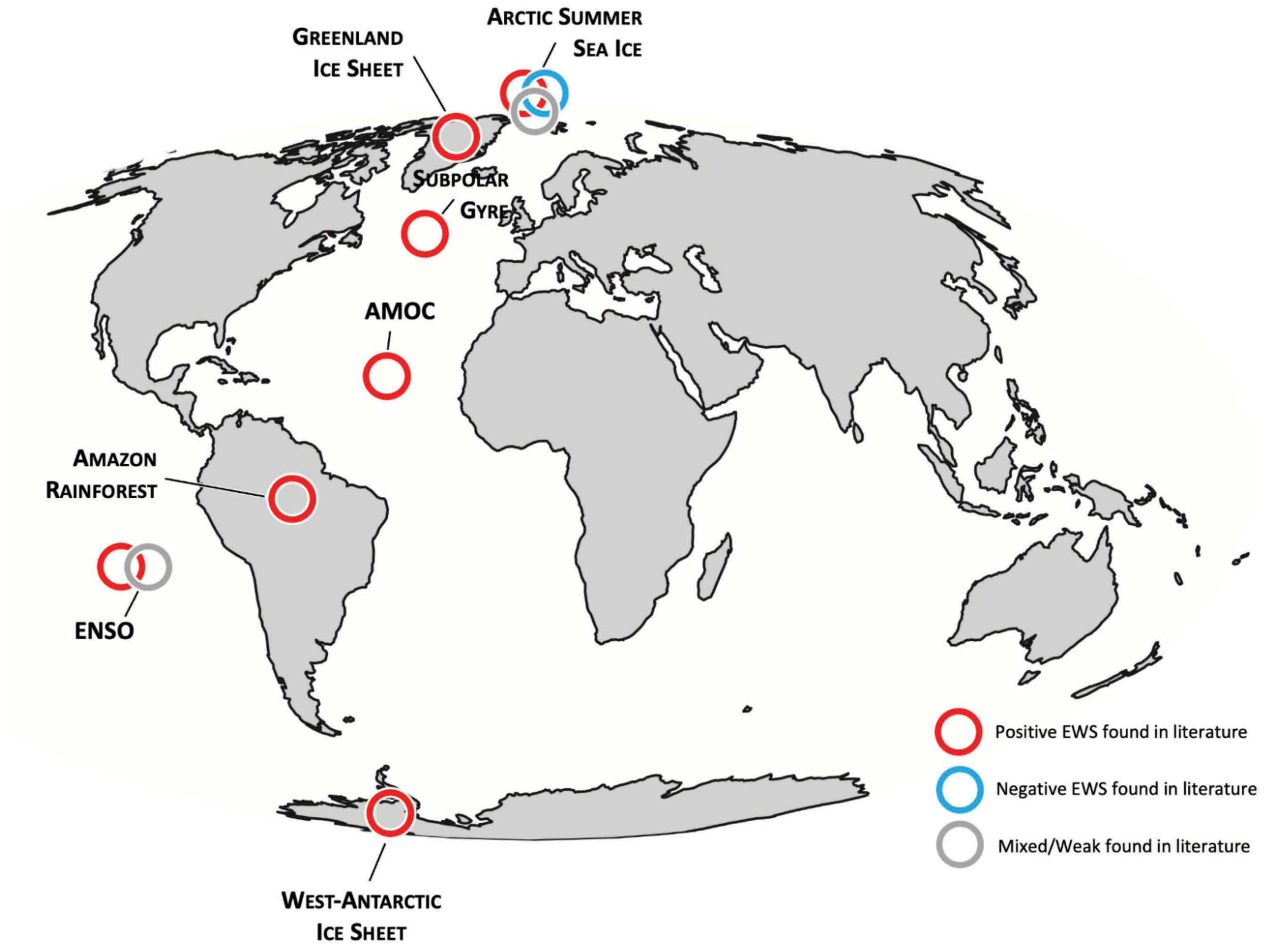


**Figure 1.6.3:** Map of studies that use empirical data to look for early warning signals (EWS) of tipping points in climate systems, and if they found evidence of EWS (red circles), no evidence (blue circles), or the evidence was unclear (grey circles). Specifically, these studies use real-world observations rather than being restricted to modelling studies.

### 1.6.2.1 Cryosphere: Ice sheets and sea ice

#### Ice sheets

Tipping points in the Greenland and West Antarctic ice sheets are detailed in Chapter 1.2, and several studies have looked for EWS. In West Antarctica, Rosier et al. (2021) searched for EWS for marine ice sheet instability on the Pine Island Glacier, identifying changes in recovery time and looking at the variance of the system state in a model. The EWS were applied to model output and successfully used to pinpoint tipping points. They find the tipping point that leads to total collapse of the glacier occurs at a +1.2°C ocean temperature increase, relative to initial conditions.

In Greenland, Boers and Rypdal (2021) found significant increases in variance and autocorrelation in detrended ice core-derived melt records from the central-western part of the Greenland Ice Sheet (GrIS), suggesting that this part of the ice sheet might be close to a tipping point. While they rule out that these EWS are directly caused by changes in temperature or precipitation, the exact mechanisms leading to the observed signs of stability decline remain unclear. The melt-elevation feedback (Levermann and Winkelmann, 2016) acts mostly on timescales longer than what can be captured by the data used by Boers and Rypdal (2021), so other positive/amplifying feedbacks related to much shorter timescales likely dominate.

As mentioned by Boers and Rypdal (2021) these include the melt-albedo feedback, related to snowline migration and albedo reductions once the uppermost, white firn layer has melted and the darker grey ice is exposed (Ryan et al., 2019), as well as thinning of outlet glaciers, which accelerates the ice flow upstream (Aschwanden et al., 2019).

#### Sea ice

Arctic sea ice loss has previously been proposed as a potential tipping system, but in this report both summer and winter sea ice loss are categorised as unlikely to feature tipping thresholds beyond which feedback-driven self-sustaining loss occurs, with other factors driving abrupt losses (1.2.2.2). In simplified models, however, Merryfield et al. (2008) found increasing variance and AR(1) in sea ice area before abrupt summer loss in a single column, two-season model. In another single-column model with a continuous seasonal cycle (Eisenman and Wettlaufer, 2009), Moon and Wettlaufer (2011) found that the destabilising ice-albedo feedback leads to CSD before the loss of winter sea ice.

While these results are apparently in agreement with expectations from simple dynamical systems, Arctic sea ice is an example of how additional caveats can obscure EWS, leading to 'false alarms'. Although attempts have been made using empirical data (Livina and Lenton, 2013), using total Arctic sea ice area as a variable could lead to misleading EWS, due to the different amounts of area masked by the continents in different climates (Goose et al., 2009; Eisenman, 2010). Moreover, several nonlinear feedbacks can dominate the recovery time and obscure CSD far from tipping; once ice gets thinner, its heat conductivity decreases, making its response to atmospheric temperature anomalies much faster (Thorndike et al., 1975). Also, a warmer Arctic means a longer period of open water after summer sea ice loss, which introduces a longer timescale. This effect is independent of the nonlinearity of the winter sea ice loss, and could cause EWS false alarms (Wagner and Eisenman, 2015; Bathiany et al., 2016a).

Alternative EWS which can also work in seasonal systems where the balance of feedbacks can change during the year (Moon and Wettlaufer, 2011) include measuring the amplitude and phase lag relative to the forcing (Williamson et al., 2016). Also, there are indications that abrupt loss of winter sea ice are still possible, but could potentially be predicted on the basis of the homogeneity in the ice-thickness distribution (Bathiany et al., 2016b; see 1.2.2.2).

### 1.6.2.2 Biosphere: Amazon rainforest dieback

Amazon dieback as a tipping point is observed in some modelled climate change scenarios (1.3.2.1). One such study shows that temporal EWS, such as increases in AR(1) and variance, are not necessarily good indicators of Amazon dieback in a number of HadCM3 GCM runs (Boulton et al., 2013). This is most likely because the Amazon is forced too fast and non-linearly for these statistical measures to work. Because of this, a system-specific indicator was suggested, looking at the sensitivity of ecosystem productivity anomalies to temperature changes, and then as a real-world measurable signal, the sensitivity of atmospheric $CO_2$ anomalies to these temperature anomalies. Both of these indicators worked well across the ensemble of runs.

Further work observes an increase in drying in the Amazon region across the recent CMIP6 model suite (Ritchie et al., 2022). An increase in the sensitivity of the temperature seasonal cycle amplitude to global warming is observed to be more prominent in locations that subsequently experience abrupt dieback shifts. The increasing sensitivity of the temperature seasonal cycle amplitude to global warming, therefore, has the potential to be used as a system-specific EWS for future dieback in the Amazon rainforest (Parry et al., 2022).

Real-world observational data has shown different results regarding the generic indicators discussed in this chapter, particularly the use of vegetation optical depth (VOD), a remotely sensed product that is strongly correlated with the amount of water content in the trees. Using this, increases in AR(1) and variance particularly since the early 2000s have shown a loss of resilience in the Amazon rainforest (Boulton et al., 2022). Using this same dataset, while modelling the water recycling network across the region (1.3.2.1), a network approach shows similar losses of resilience (Blaschke et al., 2023).

### 1.6.2.3 Ocean: Atlantic Meridional Overturning Circulation (AMOC)

The AMOC system has been identified as having the potential to collapse from its current strong state. One potential cause of this tipping point is the increased freshwater influx from a mix of increased precipitation and ice melt due to climate change (more detail on the mechanisms and likelihood given in 1.4.2.1).

EWS of AMOC collapse (Boulton et al., 2014) have been detected in a fully coupled climate model that was forced with a linearly increasing freshwater flux (Hawkins et al., 2011) – specifically increases in temporal AR(1) and variance in the strength of the overturning circulation. Furthermore, circulation strength at different latitudes are tested in this model, allowing the possibility to see where EWS may work best in the real world. A significant detection of the movement towards tipping could be seen up to 250 years in advance after 550 years of monitoring.

Direct measurements such as sea surface temperature and salinity across the Atlantic ocean can provide a real-world fingerprint of current AMOC strength. A recent study identified potential EWS of AMOC collapse using these measurements. In eight such indices, increases in AR(1) and variance are found over the last century and suggest that the AMOC could be approaching a tipping point to its weaker circulation mode (Boers, 2021). When extrapolated, these EWS of AMOC collapse give an indication of a mid-21st Century AMOC tipping point (Ditlevsen and Ditlevsen, 2023), although considerable uncertainty remains around these timelines (Ben-Yami et al., 2023; see 1.4.2.1).

Proxy records, such as bivalve shell increments, can provide an opportunity to measure early warning indicators prior to historical transitions. Recent work using three bivalve records has found that the North Atlantic Subpolar Gyre, a subsystem of the AMOC, destabilised prior to the transition into the Little Ice Age in the 14th century, with measurable EWS of AR(1) and variance prior to this transition (Arellano-Nava et al., 2022).

## 1.6.3 Recommendations and looking ahead

As detailed in this chapter, EWS can provide a way of detecting whether a system may be losing resilience and approaching a tipping point. This requires two elements: models to provide an indication of where tipping points and preceding EWS may be identified, and appropriate empirical data from the system (often remotely sensed) on which to calculate EWS. Future work in this field should be centred around this approach, and should aim to further our understanding of if (and when) a system may have a tipping point and increase the availability of tailored remote sensing products for empirically measuring these EWS.

### 1.6.3.1 Increasing data availability

Over the last decade, remote sensing data has gained greater prominence in assessing the possibility of climate and ecological tipping points (Dakos et al., 2023). This is linked to the increasing amount of open access data and the computational capacity to analyse it. Some datasets have been available since ~1972, thus giving us approximately 50 years of time series data to analyse. This provides users a long enough record from which to get statistically significant EWS (bearing in mind issues around merging data from new sensors; 1.6.1.6), and as such should be utilised as much as possible.

The use of remote sensed products can contribute in two different, and complementary, ways to detect EWS: direct and derived measurements. The use of direct observables, or low-level products, requires an advanced knowledge of the acquisition system to control and account for parameters that may affect the extraction of EWS in terms of the data's Signal-to-Noise ratio. Additionally, one could consider the use of derived measurements, or high-level products, which correspond to physical variables calculated from the aforementioned low-level products, such as NDVI ('normalised difference vegetation index') as a measure of vegetation greenness. These datasets can be more usable, but their second-order nature can present a source of uncertainty that may hinder the extraction of EWS. Nevertheless, both low- and high-level remote sensed data and products are considered in the extraction of domain-specific variables, such as climate, (Bojinski et al., 2014), ocean (Miloslavich et al., 2018) and biodiversity variables (Pereira et al., 2013).

The benefits of these growing and openly available remote sensing datasets are clear: new sensors are able to provide data with improved spatial resolutions (in the order of metres for optical and radar sensors) across very large areas, thus making possible improved analysis of both temporal and spatial EWS.

#### 1.6.3.2 Models and EWS of tipping points

Earth system models can provide information on where to look for temporal and spatial EWS with empirical data, as well as to help determine what processes are most appropriate to monitor.

Systematic efforts to identify tipping points in Earth system models, such as the new Tipping Point Model Intercomparison Project (TipMip), will help to catalogue which variables we should focus on for different tipping points. For example, examining simulated sea surface height, temperature and salinity data prior to modelled abrupt shifts in the subpolar gyre, while incorporating known uncertainties in remote sensing, could determine which remotely sensed data are most informative and where additional monitoring could add value.

The unprecedented amount of Earth observation data originating from remote sensing systems, field measurements and simulated data, coupled with innovative Earth system models and cutting-edge computing, has made possible the concept of an Earth 'digital twin' that can be studied in detail. This concept will allow us to explore the different components of the Earth system and natural and human-induced changes to identify EWS.

#### 1.6.3.3 Applications of AI for predicting tipping points

As alluded to earlier in this chapter, Artificial intelligence (AI), and in particular deep learning, is beginning to play an important role in tipping point prediction and EWS. With a wealth of data available now monitoring these systems, we can truly start to use these techniques alongside traditional EWS to attempt to fully understand Earth system tipping points. In addition to generic EWS derived from AI, we can conceive of system-specific AI-based EWS which are trained on models of specific climate tipping points. Eventually this might enable accurate prediction of critical thresholds in climate variables that would cause tipping, so that we can better predict when they would occur.

## 1.6.4 Final remarks

The EWS methods described in this chapter could be invaluable in understanding how climate and ecological systems are moving towards tipping points. A variety of different data types can be used to monitor the changing resilience of these systems, whether this be temporal, spatial, or more complex data. While many studies use models to test EWS, a number of studies already make use of real-world observations to do this. However, advances in machine learning techniques, in addition to longer and higher spatial resolution remote sensing products, can improve results. With these, we will be able to better determine with higher statistical accuracy if there are losses of resilience in systems, and in more systems than have been analysed so far.

# Chapter 1.7 Earth system tipping points synthesis

**In this chapter we present the key messages and knowledge gaps from this section on Earth system tipping points (ESTPs).**

## 1.7.1 Key messages

Several key messages emerge from this section. Firstly, there is **considerable evidence for tipping points existing in many parts of the Earth system** (see Table 1.7.1 for summary). Several tipping points are likely in the cryosphere, at a large scale in ice sheets and on a more local scale in permafrost and glaciers. In the biosphere, evidence for regime shifts and tipping points exist in many ecosystems such as in tropical forests, savannas, drylands, lakes, coral reefs and fisheries, and are often spatially complex. Tipping points in ocean circulation and monsoons are also likely to exist, but the proximity of their thresholds are subject to high uncertainty. In contrast, some other suggested ESTPs have been assessed as unlikely in this report, including for Arctic sea ice, global-scale permafrost or glacier tipping, some types of lake ecosystem tipping, tropical clouds and climate sensitivity, and the El Niño–Southern Oscillation (ENSO).

Secondly, we know that **we could already be very close to some Earth system tipping points**. Several cryosphere tipping points cannot be ruled out at 1.5°C of global warming, which will be reached even with aggressive mitigation. These tipping points become likely beyond the 2°C of warming that the Paris Agreement commits countries to stay well below, but which current policies are likely to substantially overshoot (Climate Action Tracker, 2022; Meinshausen et al., 2022; IEA, 2023). In the biosphere, deforestation in the Amazon combined with climate change-induced drying could lead to regional dieback, some drylands are close to degradation tipping points, and coral reef die-off is already occurring in many regions. In the North Atlantic Ocean, convection in the Labrador and Irminger Seas could collapse within Paris Agreement warming levels of well below 2°C, with severe impacts across the North Atlantic region. Early warning signals indicate that several systems, such as parts of the Greenland Ice Sheet, Atlantic meridional overturning circulation (AMOC) and the Amazon rainforest may be losing resilience, which could mean their tipping points are approaching (but exactly when is uncertain).

Thirdly, **complex and sometimes uncertain interactions between tipping drivers, components of the Earth system and key feedbacks make tipping dynamics difficult to assess for some systems.** For example, parts of the Amazon rainforest could die back as a result of climate change-driven drying as well as direct deforestation and degradation, but while their combination makes tipping likely sooner, the thresholds for the combined effect of these processes is difficult to estimate. For many of the systems considered, key feedbacks and processes that could be involved in tipping are not well understood and so are either represented simplistically or left out of models (such as fire feedbacks, land use change, and spatial variability in the Amazon), making future projections more uncertain. Short observational records for some systems make early warning signals less reliable as well. Many tipping systems closely interact through the climate, and evidence – particularly from palaeorecords – suggests that most interactions mean one system tipping it tends to destabilise connected systems. Model limitations mean there are large uncertainties around these potential tipping cascades, but as warming approaches the levels where some key tipping points become likely, the possibility of cascades is a growing risk that requires new approaches to assess.

Together, this evidence provides strong motivation for both rapidly reducing human-driven pressures on the Earth system, from eliminating greenhouse gas emissions (GHG) and deforestation to increasing social-ecological resilience (see Section 3) and preparing adaptation plans for the societal impacts of Earth system tipping points (see Section 2) should some tipping points occur despite mitigation efforts.

Importantly, our assessment does not suggest that crossing major tipping points could lead to runaway warming, with mitigation to prevent further tipping points being worthwhile even if some tipping points are reached. Equally, uncertainties around tipping point thresholds and interactions makes mitigation even more critical, as we cannot rule out tipping happening sooner than we currently expect.

## 1.7.2 Recommendations

Our section presents clear implications, as well as multiple research and institutional avenues for improving our understanding of Earth system tipping points.

Firstly and most importantly, **reduction of human-driven pressures on the Earth system is critical in order to prevent destabilisation of the Earth's tipping systems**. In particular, most ESTPs considered in this section feature climate change as a key driver, and as such urgent and ambitious action to reduce GHG emissions to zero would reduce the chance of passing these tipping points. Many ESTPs in the biosphere are also driven by habitat loss and pollution – for example, deforestation and degradation in tropical forests or nutrient pollution in lakes and coastal ecosystems. Reducing these pressures would make climate-driven tipping less likely, as would efforts to bolster ecological resilience in these systems through restoration, legal protection and supporting sustainable livelihoods.

Secondly, **key knowledge gaps can be addressed through improved observations and models of varying complexity**. Despite our growing understanding of key Earth system feedbacks and interactions, some are currently not well represented in many computer models. As a result, tipping dynamics and interactions between tipping systems are less likely to emerge in model simulations, making comprehensive risk assessments difficult. To this end, it is necessary to better understand key feedbacks and interactions and resolve them in models, for example processes like marine ice cliff instabilities, feedbacks between meltwater and ocean circulations, small-scale mixing processes in ocean and atmosphere, and interactions between ecohydrological and fire feedbacks or spatial variability in the biosphere. These shortcomings can be systematically explored in tailored model intercomparison projects (MIPs), which are an established cornerstone of climate assessments. Insights from such modelling initiatives, together with palaeo evidence, observations and conceptual understanding of natural processes, can help guide the development of simpler models. Since their reduced complexity allows them to be run more often, they can help better understand uncertainties, for example around tipping point interactions, and can support interim risk analyses, together with expert elicitations.

Thirdly, **improved palaeo reconstructions and observational data are key, both for developing better models and determining what systems may be at most risk**. Remote sensing from space has allowed for global monitoring of vegetation cover over the last few decades, while in the ocean the RAPID array has allowed AMOC strength to be monitored for the past 20 years. However, these datasets are not yet long enough to be sure whether trends or early warnings they are detecting are outside of their natural ranges, while many parts of the ocean and biosphere have low observational coverage (in particular in the Global South). Continued and expanded observations would help improve and extend this coverage, while developing novel and improved early warning techniques could help mine this data to detect declining resilience and potential early warning signals. Of equal importance is the improvement of palaeo reconstructions, which in many cases has demonstrated that different systems have tipped in the past, including many ice sheets and the AMOC. As the observational period reaches back only a few decades, palaeorecords are essential to extend our observations into the past, to improve our understanding of the tipping systems and potentially provide critical information needed for early warning. International data sharing and collaboration is also vital for improving monitoring of ESTPs, as is improving coverage in under-represented areas such as Africa and Asia.

Finally, it is clear that multiple different approaches are needed to understand the complexities of Earth system tipping points. As a result, integrating and co-designing research across natural and social sciences as well as other knowledge systems, including Indigenous and traditional ecological knowledge, can help better understand the drivers, dynamics and impacts of tipping in the Earth system.

**Table 1.7.1: Summary of key drivers, biophysical impacts and confidence in tipping dynamics for each system considered in Section 1.**

| System (and proposed tipping point) | Key drivers | Key biophysical impacts *(see S2 for societal impacts)* | Evidence for tipping dynamics? *(+ yes, - no, ? uncertain)* |
|---|---|---|---|
| **Cryosphere** | | | |
| **Ice Sheets (collapse)** | **DC: atmospheric warming (↗)**<br>**DC: ocean warming and circulation changes** (↗ GrIS, WAIS, EASBs / ↘ GrIS)<br>DC: precipitation increase (↘)<br>DC: black carbon deposition (↗)<br>CA: sea ice decline (↗)<br>CA: atmospheric circulation (?) | **Sea-level rise resulting in global loss of coastal land over centuries to millennia**<br>Disruption of global ocean circulation<br>Substantial shifts in atmospheric circulation patterns<br>New ecosystems on exposed land | **+++ Greenland**<br>**+++ West Antarctica**<br>**+++ Marine basins East Antarctica**<br>**++ Non-marine East Antarctica** |
| **Sea Ice (loss)** | **DC: atmospheric warming (↗)**<br>DC: atmospheric circulation shifts (↗/↘)<br>DC: ocean warming (↗)<br>DC: ocean circulation shifts (↗/↘)<br>DC: black carbon deposition (↗)<br>DC: storminess increase (↗)<br>CA: ocean stratification increase (↘) | **Regional warming (polar amplification)**<br>Ecosystem disruption<br>Impacts on ocean circulation<br>Impacts on atmospheric circulations<br>Increased evaporation | - - - Arctic summer<br>- - Arctic winter<br>- Barents Sea<br>? Southern Ocean |
| **Glaciers (retreat)** | **DC: atmospheric warming (↗)**<br>DC: deposition of dust, black carbon etc. (albedo) (↗)<br>DC: reduced snow (input and albedo) (↗)<br>DC: local thermokarst (↗) | **Water supply decline**<br>Ecosystem disruption (e.g. wetlands, water chemistry)<br>Increase in number and size of glacier lakes<br>Increase in slope instabilities<br>Transition from glacial to paraglacial landscapes<br>Sea-level rise | **++ (regional)**<br>- - (global) |

| System (and proposed tipping point) | Key drivers | Key biophysical impacts *(see S2 for societal impacts)* | Evidence for tipping dynamics? *(+ yes, - no, ? uncertain)* |
|---|---|---|---|
| **Permafrost (thaw)** | **DC: atmospheric warming (↗)**<br>**DC: ocean warming (subsea, ↗)**<br>CA: vegetation change (increase: albedo ↗, increase summer shading ↘; vice versa for dieback)<br>CA: wildfire intensity increase (↗)<br>CA: precipitation (rain extremes, snow cover albedo ↗)<br>CA: sea ice loss (subsea, ↗)<br>CA: water pressure reduction (subsea, ↗) | **Greenhouse gas emissions**<br>Landscape disruption<br>Ecosystem disruption | **++ land (regional)**<br>**- -** land/subsea (global, 10s-100s years) |
| **Biosphere** | | | |
| **Tropical Forests (dieback)** | **DC: atmospheric warming (↗)**<br>**NC: deforestation/degradation (↗)**<br>**DC: drying (↗)**<br>**CA: increasing fire frequency/intensity (↗)**<br>DC: heatwaves (↗)<br>CA: ENSO intensification (e,g, Amazon, SE Asia, ↗)<br>CA: AMOC/SPG weakening/collapse (e.g. Amazon, ↗)<br>CA: terrestrial greening (↘, declining) | **Biodiversity loss**<br>**Regional rainfall reduction (e.g. from Amazon dieback across Amazon Basin and Southern Cone)**<br>Carbon emissions (amplifying global warming)<br>Remote impacts on rainfall patterns all over the planet | **+++ (Amazon, local)**<br>**++ (partial dieback/regional)**<br>**+ (full dieback/continental)**<br>**+? (SE Asia, local)**<br>**- - (regional)** |
| **Boreal Forests (dieback / expansion)** | **DC: drying (↗)**<br>**CA: fire frequency/intensity increase (↗)**<br>**DC: atmospheric warming (↗)**<br>**CA: permafrost thaw (↗)**<br>**CA: insect outbreaks (↗)**<br>NC: deforestation and degradation (↗)<br>DC: heatwaves (↗)<br>CA: terrestrial greening (↘)<br>CA: vegetation albedo (↗)<br>CA: sea ice albedo decline (↗)<br>DC: precipitation changes (?) | **Biodiversity loss**<br>Carbon emissions (amplifying global warming) from southern dieback, carbon drawdown (reducing global warming) from northern expansion<br>Complex regional biogeophysical effects on warming - dieback = higher albedo (cooling) but less evaporative cooling (warming) and vice versa for expansion | **++ Southern dieback (partial/regional), + (continental)**<br>**+ Northern Expansion (partial/regional)** |

| **System (and proposed tipping point)** | **Key drivers** | **Key biophysical impacts** *(see S2 for societal impacts)* | **Evidence for tipping dynamics?** *(+ yes, - no, ? uncertain)* |
|---|---|---|---|
| **Temperate Forests (dieback)** | **DC: atmospheric warming (↗)**<br>**DC: droughts (↗)**<br>**DC: heatwaves (↗)**<br>CA: insect outbreaks (↗)<br>CA: windthrow (↗)<br>NC: deforestation and degradation (↗)<br>CA: fire frequency increase (↗)<br>NC: fragmentation (↗) | **Biodiversity loss**<br>Carbon emissions (amplifying global warming)<br>Regional warming in summer due to less evaporative cooling, less cloud cover<br>Less atmospheric water supply<br>Less groundwater recharge | ? (partial / regional) |
| **Savannas and Grasslands (regime shifts)** | **NC: fire suppression (↗)**<br>**NC: overgrazing (↗)**<br>DC: increased precipitation intensity (↗)<br>CA: terrestrial greening (↗)<br>NC: afforestation (↗)<br>CA: regional circulation changes (e.g. Sahel) (↗) | **Biodiversity loss**<br>Greater groundwater depletion (with shrub encroachment)<br>Nutrient cycle disruption<br>Reduced fires (with shrub encroachment) | **++ (local to landscape),** ? (regional) |
| **Drylands (regime shifts)** | **DC: drying (↗)**<br>**DC: atmospheric warming (↗)**<br>**NC: land use intensification (↗)**<br>DC: extreme events (heatwaves, floods) (↗)<br>DC: increased rainfall variability (↗)<br>CA: terrestrial greening (↘)<br>CA: insect outbreaks (↗)<br>CA: invasive species (↗) | **Biodiversity loss**<br>Aridification/desertification<br>Groundwater depletion (with encroachment)<br>Regional rainfall changes<br>Shift in species composition (e.g. shrub encroachment) | **++ (local to landscape), + (regional)** |
| **Freshwater / Lakes (regime shifts)** | **NC: nutrient pollution (↗)**<br>**CA: terrestrial greening (↗)**<br>**NC: afforestation (↗)**<br>DC: atmospheric warming (↗)<br>DC: precipitation changes (↗)<br>CA: permafrost thaw-related thermokarst formation/ drainage (↗)<br>CA: glacier lake formation/drainage (↗)<br>DC: drought (↗)<br>CA: warming-driven species range expansion (↗)<br>CA: water use intensification (↗)<br>NC: human-mediated species introduction (↗) | **Biodiversity loss**<br>**Water quality decline**<br>Increased GHG emissions from most (reduced for salinisation) | **+++ (eutrophication- driven anoxia , widespread localised)**<br>**++ (DOM loading, widespread localised in boreal)**<br>- (lake (dis)appearance, widespread localised in tundra)<br>- (N to P-limitation switch, localised in high N-deposition regions)<br>- (salinisation, localised in arid regions))<br>- (invasive species, widespread localised) |

| System (and proposed tipping point) | Key drivers | Key biophysical impacts *(see S2 for societal impacts)* | Evidence for tipping dynamics? *(+ yes, - no, ? uncertain)* |
|---|---|---|---|
| **Coastal - warm-water coral reefs (die-off)** | **DC: ocean warming (↗)**<br>**DC: marine heatwaves (↗)**<br>**CA: disease spread (↗)**<br>CA: ocean acidification (↗)<br>NC: water pollution (nutrient / sediment) (↗)<br>NC: disruption (ships, over-harvesting) (↗)<br>CA: disease spread (↗)<br>CA: invasive species (↗)<br>DC: storm intensity (↗)<br>CA: sea level rise (↗) | **Biodiversity loss (ecosystem collapse, ~25% marine species have life stages dependent on coral reefs)**<br>Loss of commercial and artisanal fisheries, and other sectors<br>Coastal protection loss | **+++ (localised).**<br>**+++ (regionally clustered)** |
| **Coastal - mangroves and seagrass meadows (die-off)** | **DC: increased climate extremes (e.g. tropical cyclones, marine heatwaves, El Niño intensity, droughts) (↗)**<br>**NC: habitat loss (agri/aquaculture) and degradation (fishing damage) (↗)**<br>**CA: sea level rise (esp. mangroves, ↗)**<br>**NC: nutrient pollution (↗)**<br>**NC: shoreline change (erosion, sedimentation) (↗)**<br>DC: ocean warming (seagrass, ↗)<br>CA: disease spread (seagrass, ↗)<br>NC: invasive species (seagrass, ↗) | **Biodiversity loss**<br>**Loss of coastal protection**<br>Loss of carbon sink/increased GHG emissions<br>Loss of water quality<br>Sediment salinisation<br>Subsidence<br>Enhanced sediment sulphide and methane releases<br>Hypoxia (seagrasses)<br>Reduced nutrient recycling | **++ (mangroves, regional)**<br>**++ (seagrasses, regional)** |
| **Marine ecosystems and environment (regime shifts)** | **NC: over-exploitation (↗)**<br>**DC: ocean warming (↗)**<br>**NC: water pollution (nutrients/sediment) (↗)**<br>NC: habitat loss (↗)<br>DC: marine heatwaves (↗) | **Keystone species collapse**<br>**Trophic cascades**<br>**Regime shifts**<br>Changes to carbon sequestration<br>Impacts on ocean biogeochemistry<br>Major changes in ocean productivity, biodiversity and biogeochemical cycles | **+++ (cod fisheries, regional)**<br>**+ (large fish fisheries, regional)**<br>- (small fish fisheries, regional)<br>**+ (marine communities, local)**<br>**+++ (kelp forests, local)**<br>? (lipid pump, regional)<br>- - (gravitational pump, regional))<br>**+ (marine hypoxia , ? (regional to global)** |
| **Ocean/Atmosphere Circulation** | | | |

| **System (and proposed tipping point)** | **Key drivers** | **Key biophysical impacts** *(see S2 for societal impacts)* | **Evidence for tipping dynamics?** *(+ yes, - no, ? uncertain)* |
|---|---|---|---|
| **Ocean overturning (collapse)** | **DC: ocean warming (↗)**<br>**DC: precipitation increase (↗)**<br>**CA: ice sheet meltwater increase (SO ↗, primary in the future for AMOC/SPG)**<br>**CA: river discharge increase (AMOC/SPG ↗)**<br>CA: sea ice extent and thickness decrease (↗)<br>DC: regional aerosol forcing increase (↘)<br>CA: regional ocean circulation changes (?)<br>CA: wind trends (SO,↗)<br>CA: sea ice formation (SO, ↗) | **Cooling, change in precipitation and weather over Northern Hemisphere**<br>**Change in location and strength of rainfall in all tropical regions**<br>Reduced efficiency of global carbon sink, and ocean acidification<br>Deoxygenation in the North Atlantic<br>Change in sea level in the North Atlantic<br>Modification of sea ice and arctic permafrost distribution<br>Change in winter storminess<br>Reduced land productivity in Atlantic bordering regions<br>Increased wetland in some tropical areas and associated methane emission<br>Change in rainforest response in drying regions<br>Modification of Earth's global energy balance, timing of reaching 2°C global warming<br>Reduced efficiency of global carbon sink<br>Change in global heat storage<br>Reduced support for primary production in world's oceans<br>Drying of Southern Hemisphere<br>Wetting of Northern Hemisphere<br>Modification of regional albedo, shelf water temperatures<br>Increase in summer heat waves frequency<br>Collapse of the North Atlantic spring bloom and the Atlantic marine primary productivity<br>Increase in regional ocean acidification<br>Regional long-term oxygen decline<br>Impact on marine ecosystems in the tropics and subtropics. | **++ (AMOC)**<br>**++ (Subpolar Gyre)**<br>**++ (Southern Ocean)** |

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