Chapter 2.2 Status of Earth system tipping points: What's new?

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Acknowledged: Daniel Mayor, Arie Staal, Els Weinans

Key messages

Cryosphere

- We have high confidence that ice sheets - from Greenland to West Antarctica - have tipping points leading to irreversible collapse, locking in long-term multi-metre sea level rise, and have been at risk since at least 1°C of global warming.
- While Arctic summer sea ice is unlikely to reach tipping points, we cannot rule out a tipping point for Antarctic sea ice which could already be underway, although this is highly uncertain.
- We have medium confidence in potential regional tipping in permafrost and glaciers, which would respectively amplify emissions and commit some regions to total deglaciation.

Biosphere

- The Amazon rainforest has faced two years of intense El Niño-induced drought, and the combined effects of deforestation and climate change put it at risk below 2°C of global warming.
- Warm-water coral reefs have experienced the worst bleaching event on record over 2023-25, and the central estimate of their thermal tipping point of 1.2°C global warming has been crossed.
- We now recognise river deltas and peat bogs as potential tipping systems, identify the potential for localised mangrove tipping with high confidence, and the potential for local-scale temperate forests tipping with low confidence.

Ocean/Atmosphere circulations

- Recent modelling supports convection in the Atlantic Meridional Overturning Circulation (AMOC) and subpolar gyre being capable of tipping, which cannot be ruled out at current warming levels, but limited models and observations means how likely they are to tip on current trajectory remains uncertain.
- In the Southern Ocean, dense shelf Water formation may be declining and could reach a tipping point, but understanding of its interactions with ice remains limited.
- Evidence has strengthened for no tipping dynamics in the 'jet stream', while recent modelling supports monsoons having tipping dynamics, but evidence remains limited.

Interactions

- Out of 20 climate tipping system interactions assessed, most are destabilising, but a few (e.g. AMOC on Amazon, West Antarctic Ice Sheet on AMOC) may have a stabilising effect.
- A vicious cycle may form where permafrost thaw could lead to amplified Arctic sea ice retreat, which may lead to enhanced inland permafrost degradation and so on.
- The AMOC is the key global mediator of tipping point interactions, featuring in 45 per cent of all assessed tipping point interactions.

2.2.1 Executive summary

Many parts of the Earth system can reach a point beyond which change in response to pressure can become self-sustaining, resulting in an often irreversible and abrupt shift to a very different state - what we refer to as a 'tipping point'. In this chapter we briefly summarise each proposed tipping system covered by the last Global Tipping Points Report, and reassess each based on relevant new scientific research published since the last report.

In the **cryosphere** - Earth's frozen reaches - we (the Global Tipping Points community) maintain high confidence in ice sheet tipping points, two of which have been at risk since around 1°C of warming, with potentially substantial consequences for future sea level rise. We also maintain medium confidence in local to regional tipping in permafrost and glaciers, with implications for amplified emissions and regional deglaciation. While Arctic summer sea ice decline is unlikely to reach a tipping point, we cannot rule out tipping for it in the winter, or around Antarctica, where sea ice has recently dropped for the first time.

In the **biosphere** - the living world - we are more confident in the potential for tipping in the Amazon at various scales, and note that combined with ongoing deforestation as little as 1.5°C of warming could trigger widespread dieback. Both the Amazon and coral reefs have suffered during the 2024-25 El Niño event, seeing the worst coral bleaching event on record and signs of die-off in many regions. We also have higher confidence in localised mangrove tipping, and now include peat bogs and river deltas as potential freshwater tipping systems.

In the **circulations of the ocean and atmosphere**, recent research strengthens the case for North Atlantic convection being capable of tipping, potentially at current warming levels, but large uncertainties remain on if and when they may tip in practice. Convection around Antarctica may also be weakening towards a tipping point, driven by warming and meltwater, but we are not sure how this in turn interacts with ice melt. We now include the East Asian summer monsoon as a potential tipping system, but remain confident that despite changes in response to warming the northern polar 'jet stream' as well as the El Niño Southern Oscillation (ENSO) and large-scale tropical circulations are unlikely to have a tipping point.

Tipping points do not exist in isolation - they **interact** in ways that can change the likelihood of their tipping. We have extended our previous analysis to cover more than twenty climate tipping system interactions, adding in for example interactions with subglacial basins in East Antarctica or interactions with the permafrost. Our newest science updates maintain the finding that the majority are destabilising. For example, a vicious cycle may form where permafrost thaw releases greenhouse gases, driving further warming and more Arctic sea ice retreat, which by making the Arctic darker amplifies warming, amplifying inland permafrost degradation, and so on. The AMOC emerges as a key global mediator of tipping point interactions, featuring in nearly half of all assessed tipping point interactions, including a few that may potentially have a stabilising effect, such as AMOC collapse's impact on the southern Amazon rainforest, and West Antarctic Ice Sheet impact on the AMOC.

2.2.2 Summary table

Table 2.2.1: Tipping assessment for each system considered, highlighting changes since GTPR23. Key: +++ (high confidence yes), ++ (medium yes), + (low yes), ? (uncertain), – – – (high no), – – (medium no), – (low no).

Domain	**Tipping system** **(& tipping dynamics)**	**GTPR25 tipping system assessment** **(bolded if changed vs. GTPR23)**
Cryosphere	**Ice Sheets (collapse)**	Greenland: +++ **[threshold updated]** West Antarctica: +++ Marine basins East Antarctica: +++ Non-marine East Antarctica: ++
	Sea Ice (loss)	Arctic summer: - - - **Arctic winter: ?** [was: - -] Barents Sea: - Antarctic / Southern Ocean: ?
	Glaciers (retreat) See 4.4 Mountain glaciers case study	++ (regional) - - (global)
	Permafrost (thaw)	++ (regional, land) - - (global, land / subsea)
Biosphere	**Tropical Forests (dieback)** See 4.1 The Amazon rainforest case study	Amazon: +++ (local) **[threshold updated]** ++ (regional) + (continental) Congo: + (local), **? (regional)** [regional added] SE Asia: ? (local), - - (regional)
	Boreal Forests (dieback / expansion)	Dieback: ++ (regional) + (continental) Northern Expansion: + (regional)
	Temperate Forests (dieback)	**+ (local)** [local added] ? (regional)
	Savannas & Grasslands (regime shifts)	++ (local to landscape) ? (regional)
	Drylands (regime shifts)	++ (local to landscape) + (regional)
	Freshwater (regime shifts)	Eutrophication-driven lake anoxia: +++ (widespread localised) Lake DOM loading ('browning'): ++ (widespread localised in boreal) Lake (dis)appearance: – (widespread localised in tundra) Lake N to P-limitation switch: – (localised in high N-deposition regions) Lake salinisation: – (localised in arid regions) Lake invasive species: – (widespread localised) **River deltas: + (localised)** [system added] **Peat bogs: ++ (localised)** [system added]
	Coastal ecosystems **(regime shift)**	**Mangroves: +++ (local)** [local added] ++ (regional) Seagrass meadows: ++ (regional) Kelp forests: +++ (local)
	See 4.3 Warm-water coral reefs case study	+++ (localised) +++ (regionally clustered)

Table 2.2.1: Tipping assessment for each system considered, highlighting changes since GTPR23. Key: +++ (high confidence yes), ++ (medium yes), + (low yes), ? (uncertain), – – – (high no), – – (medium no), – (low no).

Domain	Tipping system (& tipping dynamics)	GTPR25 tipping system assessment (bolded if changed vs. GTPR23)
	Marine (benthic & pelagic) ecosystems (regime shifts)	Cod fisheries: +++ (regional) Large fish fisheries: + (regional) Small fish fisheries: – (regional) Marine communities: + (local) Biological (lipid) pump: ? (regional) Biological (gravitational) pump: – – (regional) Marine hypoxia: + (local), ? (regional to global)
Ocean & atmosphere circulations	**Ocean overturning (collapse)** See 4.2 Atlantic Ocean circulation case study	Atlantic Meridional Overturning Circulation (AMOC): ++ Deep convection in North Atlantic Subpolar Gyre (SPG): ++ Southern Ocean: ++
	Monsoons (abrupt collapse / intensification)	West African: + Indian Summer: ? South American: ? **East Asian Summer: ?** [system added]
	Tropical clouds & circulation (reorganisation)	– –
	ENSO (extreme / permanent)	– –
	Mid-latitude jet (wavier)	–

2.2.3 Introduction

The Earth system describes the interconnected complex system at the surface of the planet that sustains life, including the cryosphere (ice-related systems, including ice sheets, sea ice, glaciers and permafrost), biosphere (all living things), atmosphere, hydrosphere (water-based systems, including oceans, rivers and lakes), and the lithosphere (the Earth's solid surface) [Kump, Kasting, & Crane, 1999; Lenton, 2016].

Evidence has accumulated from models, palaeorecords, and observations that change in many parts of the Earth system can under certain circumstances become self-sustaining once forced beyond a threshold, leading to a state change driven by positive (i.e. amplifying) feedback loops, and/or the weakening of negative/balancing feedback loops [Lenton et al., 2008; Armstrong McKay et al., 2022; Wang et al., 2023; GTPR23]. We refer to this situation as a 'tipping point', and the systems that this dynamic occurs in as 'tipping systems' (see Introduction of the report).

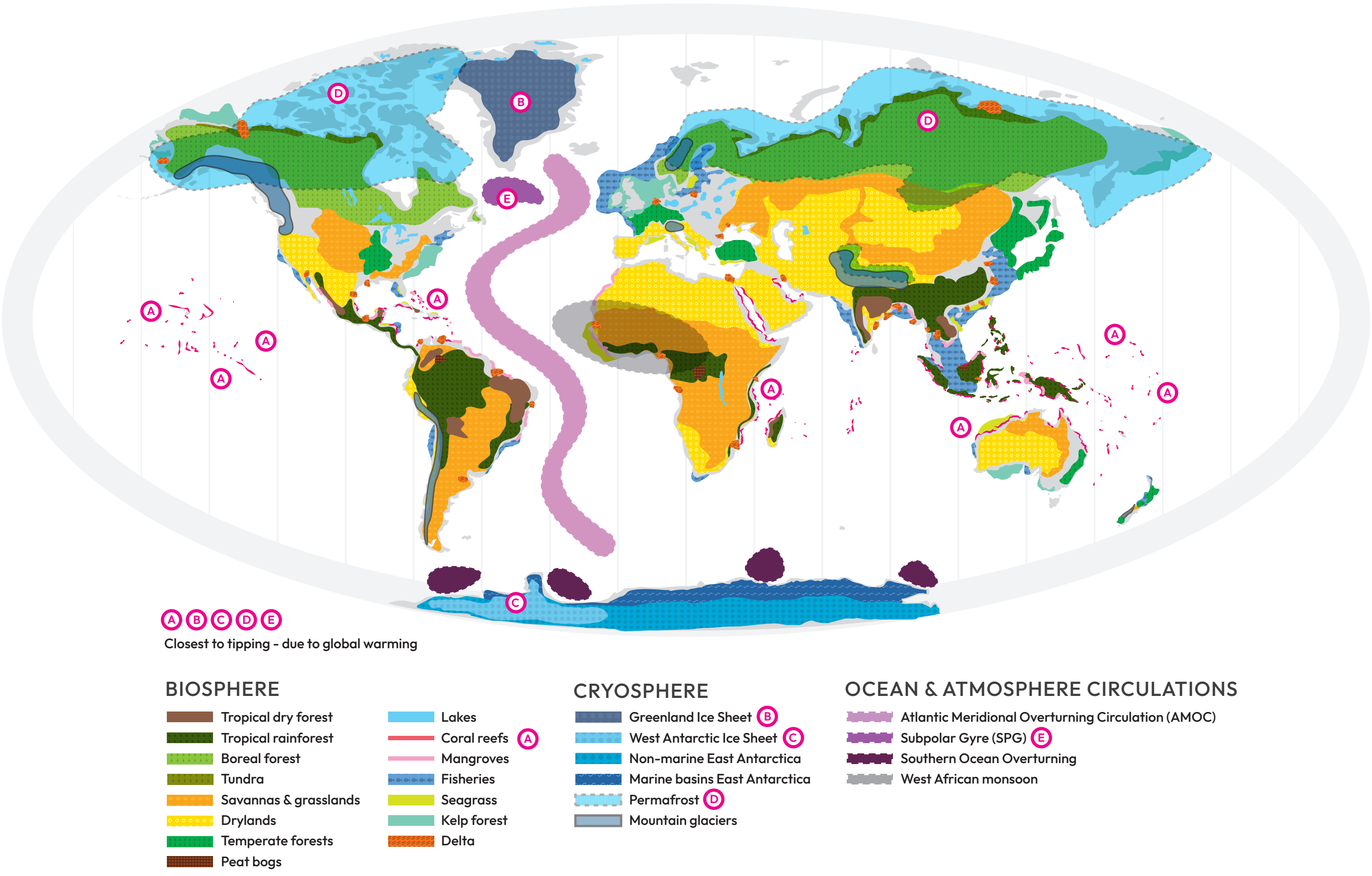

Figure 2.2.1: Map of potential tipping systems across the biosphere, cryosphere, and circulations in oceans and atmosphere.

In the first Global Tipping Points Report [GTPR23], over 120 Earth and environmental scientists assessed the biophysical evidence for tipping dynamics across the Earth system. We assessed the scientific literature for each proposed tipping system, and using collective expert judgement then judged if sufficient evidence exists for tipping dynamics in that system along with an associated confidence level (following the IPCC system; Mastrandrea et al. [2010]), and identified knowledge gaps to be targeted with further research. Based on this, we found evidence for potential tipping in many parts of the Earth system, across the cryosphere, biosphere, and ocean/atmosphere circulations, several of which may already be close to tipping thresholds due to a variety of anthropogenic pressures (Figure 2.2.1).

Since the last Global Tipping Points Report was published, new research has deepened our understanding of many of these systems. In this chapter, we briefly describe each system and our previous assessment of its tipping dynamics, before summarising insights from new research on it, and presenting any changes to the assessments made in GTPR23 (summarised in Table 2.2.1).

2.2.4 Methodology

There are many sources of information that Earth system science draws from. Broadly classified, there are direct observations (in-situ measurements and remote sensing), proxy records on historic or palaeo timescales (indirect reconstructions, e.g. inferring past temperatures via certain isotope concentrations) and models of differing complexity. The latter range from conceptual (e.g. box models) over component models (e.g. ice sheet models representing the relevant physical processes) to fully coupled Earth system models (e.g. the models participating in the Coupled Model Intercomparison Project, CMIP). All these lines of evidence come with their respective strengths and drawbacks, and can help construct a consistent understanding of a system [Boers, Ghil & Stocker, 2022].

In line with the tipping point definition above and as described in Loriani et al. [2025], our assessment reviews evidence for the presence of feedback loops that can drive self-perpetuating change beyond a threshold, leading to a state shift in that system. The confidence in our assessment increases with both robustness and agreement of evidence, following the IPCC system [Mastrandrea et al., 2010]. Table 2.2.2 summarises the criteria for different confidence levels.

Critically, this confidence rating concerns merely whether a system can tip under plausible future conditions (within coming centuries to millennia), not that it will tip in future, or could tip under any circumstances. Although these questions are tightly related, estimating whether a system is likely to tip requires quantification of the critical threshold and an assessment of whether that threshold will be transgressed, including a discussion of overshoots and timescales. All of these aspects are surrounded by considerable uncertainties for different systems.

Table 2.2.2: Summary of confidence level criteria for ESTP assessments in this chapter.

Confidence	**Criteria (from Loriani et al. [2025]**)
High (+++)	Multiple, independent lines of evidence consistently indicate the presence of feedback loops that can drive self-perpetuating change beyond a threshold on plausible future trajectories, leading to a state shift in that system. Strong palaeo analogues, consistent tipping behaviour in models across the hierarchy. If applicable, proxy and direct observations are compatible with the expected tipping dynamics.
Medium (++):	Multiple, independent lines of evidence indicate the presence of such feedback loops. However there are uncertainties in timing, magnitude or feedback strength. e.g. there are tipping dynamics in some models, and palaeo records hint at dynamics compatible with tipping. Support from observations is limited or contested.
Low (+):	Singular lines of evidence indicate the presence of such feedback loops. Tipping dynamics only emerge in specific models or under constrained assumptions. e.g. tipping is in principle conceivable via conceptual models, but there are no clear or only weak palaeo analogues. Limited demonstration of tipping dynamics in numerical models.
Not a tipping system	There is evidence indicating the lack of feedback loops that can drive self-perpetuating change beyond a threshold (with low/medium/high confidence).
Unclear	There is conflicting or limited evidence about the existence of such feedback loops.

2.2.5 Potential tipping points in the cryosphere

The cryosphere includes all of the ice-bound parts of the Earth, including ice sheets (separated into the Greenland and Antarctic ice sheets, the latter also subdivided into West Antarctic and East Antarctic ice sheets), mountain glaciers, sea ice, and permafrost.

In this section, we describe each of these systems, their evidence for tipping dynamics, and relevant new research in turn. Based on this, we have reassessed the status of several systems, including the threshold range for the Greenland Ice Sheet, updating Arctic winter sea ice from not a tipping system (low confidence) to uncertain, and discussing the potential implications of recent drop in Antarctic sea ice (Figure 2.2.2).

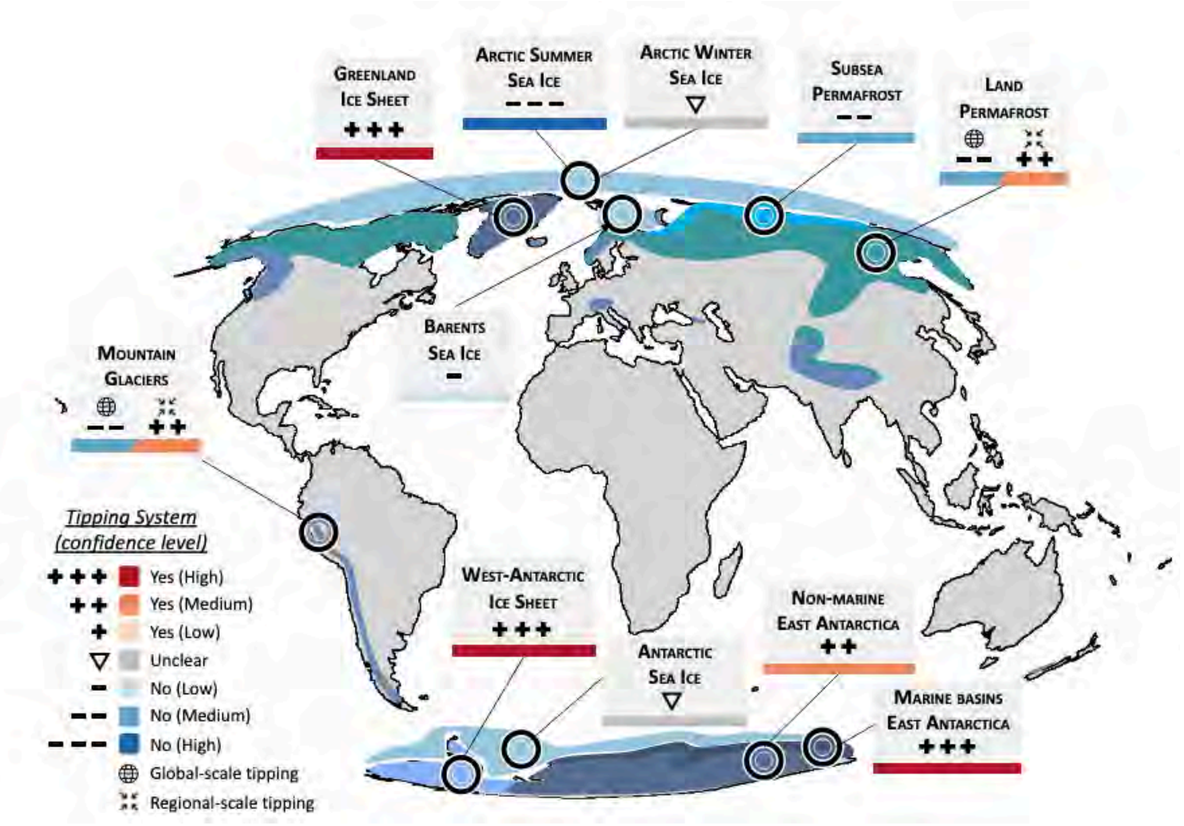

Figure 2.2.2: Map of cryosphere systems considered in this chapter (shading). The markers indicate which of the systems are in this report considered a tipping system (+++ high confidence, ++ medium confidence and + low confidence) and which are not (– – – high confidence, – – medium confidence and – low confidence), ∇ indicates systems for which a clear assessment is not possible based on the current level of understanding.

Greenland Ice Sheet

What it is

The Greenland Ice Sheet (GrIS) consists of the ice sheet covering the island of Greenland, and contains the equivalent of circa 7 metres of sea level equivalent (SLE) [Aschwanden et al., 2019]. Palaeoclimate evidence suggests that while it survived many recent interglacials (i.e. warm interludes between cold 'Ice Age' glacials, of which the current Holocene is the latest), it partly collapsed during the last interglacial 130,000 to 115,000 years ago (a.k.a. the 'Eemian', or 'MIS5e'), and may have collapsed during the 'MIS11' interglacial ~420,000 to 395,000 years ago when sea levels were likely 6 to 13 m higher than present [Alley et al., 2010; Schaefer et al., 2016; Rachmayani et al., 2017; Robinson et al., 2017; Christ et al., 2021]. While MIS11 may have reached a warmer peak than MIS5e (with up to ~2°C in the former and up to ~1.5°C in the latter relative to pre-industrial, albeit with high uncertainty and differing orbital configurations and regional patterns [Rachmayani et al., 2017; Fox-Kemper et al., 2021]), the larger estimated ice loss during the older interglacial is likely due to a longer duration of warmer climate [Robinson et al., 2017]. This sensitivity to only marginally higher warming than preindustrial indicates the presence of a tipping dynamic. Some models also support this, indicating that this is largely driven by melt-elevation feedback, in which ice mass loss leads to the ice sheet surface dropping to warmer altitudes, accelerating ice loss. Below a critical elevation, this process leads to inevitable collapse even if global warming were to halt or reverse [Boers & Rypdal, 2021]. If triggered, GrIS collapse would play out over millennia (1-10ky; A. McKay et al. [2022]), but would lock in a multi-metre sea level rise that would prove catastrophic to coastal areas. Early warning signals of tipping points in the Greenland Ice Sheet have been identified in empirical data [Boers & Rypdal, 2021], and ~30 cm sea level rise from Greenland may already be locked in, regardless of emissions scenario [Box et al., 2022].

What's new

In GTPR23, GrIS was assessed as a tipping system with high confidence, with high confidence too in it involving abrupt / large rate change and irreversibility over decadal/centennial timescales, and a threshold range of 0.8-3°C of global warming [Winkelmann, Steinert & Armstrong Mckay et al., 2023].

Since GTPR23, there have been several new publications relevant to this system. Recent observations have shown ongoing mass loss from the GrIS, with 196 ± 37 km3/yr of volume lost between 2010 and 2022 [Ravinder et al., 2024] and widespread accelerated calving across the GrIS from 1985 [Greene et al., 2024], as well as increased crevassing associated with an acceleration of ice flow at the marine-terminating sectors of the GrIS [Chudley et al., 2025]. There has also been an increase in extreme melting event frequency and intensity across the GrIS since 1950 [Bonsoms et al., 2024].

The irreversibility of ice loss from the GrIS in a fully coupled Earth system model of intermediate complexity is highlighted in Höning et al. [2024], which finds that once the southern GrIS has melted with a mass loss of greater than 0.4 m SLE, total regrowth would require atmospheric greenhouse gas concentrations below pre-industrial and timescales in the order of 10,000s years. While carbon dioxide removal technologies would be required to bring global warming levels down to below +1.0°C, such technologies at scale are hypothetical and strong ambitions to curb emissions are the most important way to prevent the GrIS from crossing a tipping threshold. Another recent study [Petrini et al., 2024] suggests that a global mean warming of 3.2-3.4°C above pre-industrial levels could result in a near full melt of the GrIS. Again this would unfold over 1,000s-10,000s of years. The higher the temperature forcing, the shorter the timescale of collapse. This analysis also suggests that the topography of the central west of Greenland may play a role in stabilising the GrIS [Petrini et al., 2024].

Based on this new research, the GTP community maintains its assessment of GrIS being a tipping system with high confidence, with an updated threshold range of 0.8-3.4°C.

Antarctic Ice Sheet

What it is

The Antarctic Ice Sheet (AIS) can be divided into the mostly marine-based West Antarctic Ice Sheet (WAIS), the marine-based sectors of East Antarctica, and the non-marine regions of the East Antarctic Ice Sheet (EAIS), on account of the differing dynamics and temperature thresholds associated with their tipping. The response of the Antarctic Ice Sheet—the largest source of long-term sea-level rise—to global warming remains poorly constrained, and large uncertainty regarding its future contribution to global sea-level rise remains [Seroussi et al., 2024; Levermann et al., 2020]. The AIS responds extremely slowly to changes in its surrounding climate, so that the full consequences of past and ongoing warming may take centuries or longer to fully unfold [Clark et al., 2016; Klose et al., 2024]. Nevertheless, Antarctica is already today losing mass and contributing to sea-level rise [IMBIE team, 2018], with losses projected to accelerate even if global temperatures were stabilised at today's levels [Reese et al., 2023].

The WAIS is separated from the EAIS by the Transantarctic Mountains and holds enough ice to raise sea levels by ~5 m. Unlike the EAIS, which largely rests on bedrock above the sea level, the majority of the WAIS rests on a bed well below sea level, making it especially vulnerable to ocean warming (either directly or by changes in ocean circulation). In contrast to the GrIS, much of the coast of Antarctica is fringed by floating ice shelves, which create a vulnerable point of contact with the ocean. These connections to the ocean, as well as the fact that most of its ice is resting below the sea level, make the WAIS much more vulnerable to climate warming than the EAIS. A collapse of the WAIS would lock in long-term sea-level rise on the order of 3-5 meters, depending on collapse extent [Garbe et al., 2020]. At current temperatures, a partial WAIS collapse may already be unavoidable in the long term [Reese et al., 2023]. Evidence exists for the collapse of the WAIS during past warm periods, such as the last interglacial [DeConto and Pollard, 2016; Sutter et al., 2016; Turney et al., 2020; Thomas et al., 2020; Weber et al., 2021] and the Pliocene when temperatures were ~2-3°C warmer than pre-industrial [Naish et al., 2009; Grant et al., 2019; DeConto et al., 2021]. Several potential mechanisms for WAIS tipping have been proposed. One likely candidate for WAIS loss is Marine Ice Sheet Instability (MISI), when the grounding line sits on a slope that deepens into the ice sheet interior. When the ice retreats inland the greater ice thickness means that more ice flows into the ocean This causes additional thinning and retreat, resulting in a self-accelerating loss until the ice reaches a stable point (often a higher bedrock elevation) [Weertman, 1974; Schoof, 2007; Mengel and Levermann, 2014; Feldmann and Levermann, 2015; Garbe et al., 2020]. Another proposed feedback mechanism is Marine Ice Cliff Instability (MICI), whereby ice shelf collapse creates inherently unstable tall marine-terminating ice cliffs, which in turn rapidly collapse and cause a self-reinforcing feedback of ice recession, which only terminates when ice cliff is buttressed or the water shallows [DeConto et al., 2021]. However, this feedback has not been observed in Antarctica (yet) and there is much less scientific consensus on this instability than MISI.

Marine-based ice-sheet sectors (like much of the WAIS) also exist in East Antarctica, including in the Wilkes, Aurora, and Recovery subglacial basins. They are potentially vulnerable to the same tipping dynamics as WAIS, including the MISI and the MICI [Morlighem et al., 2020; Stokes et al., 2022]. However, unlike the WAIS, palaeorecords of previous interglacials as well as models still carry large uncertainty with regard to their vulnerability. However, most studies suggest that the critical temperature threshold of EAIS marine-based sectors is above that of the WAIS or the GrIS.

Beyond its marine basins, the majority of the EAIS lies above sea level, containing an ice amount equivalent to ~34 meters of global sea level rise [Pritchard et al., 2025]. As they have no connection to the ocean, these parts of the ice sheet are not vulnerable to the instabilities of marine-based ice. However, although less vulnerable than the GrIS due to its thermal isolation and its location over the pole, the EAIS is susceptible to the melt-elevation feedback, which might cause self-sustained and irreversible ice loss in the long term (perhaps multiple millenia) if temperatures exceed 6°C of global warming [Garbe et al., 2020]. Under current projected warming scenarios, only modest ice loss is expected from this part of the ice sheet, and models suggest an excess of 10°C warming might be needed to lead to complete ice sheet loss [Garbe et al., 2020]. Crucially, if the ice sheet were to be lost, there would likely be strong hysteresis, requiring far greater cooling for the ice sheet to be restored.

What's new

WAIS was previously identified as a tipping system with high confidence, with a high likelihood of abrupt / large rate change and irreversibility on a decadal/centennial timescale, with an associated temperature threshold of 1-3°C. Marine-based East Antarctica was identified as a tipping system with high confidence, with a high confidence in abrupt / large rate change and irreversibility on a decadal/centennial timescale, with an associated temperature threshold of 2-6°C. Non-marine based East Antarctica was identified as a medium confidence tipping system, with a high confidence in abrupt / large rate change and medium confidence in irreversibility on a decadal/centennial timescale, with an associated temperature threshold of 6-10°C [Winkelmann, Steinert & Armstrong Mckay et al., 2023]. Since GTPR23, there have been numerous publications which are relevant to the science of ice sheet tipping points for the WAIS, marine-based and non-marine based EAIS, and across the whole Antarctica. We shortly highlight these new studies below for the AIS as a whole, and for West and East Antarctica separately.

Across the whole of Antarctica

On potential thresholds, an ice sheet model forced by different climate model simulations of the warm mid-Pliocene (3-3.3 million years ago) compared with warming stabilised at current levels indicates that WAIS collapse occurs with a modest 0.5-1°C of ocean warming above pre-industrial, while the East Antarctic Wilkes Subglacial Basin retreats at a higher level of around 3°C oceanic warming (depending on precipitation changes) [Blasco et al. 2024]. Using an ice-sheet model, Coulon et al. [2024] identify a threshold of +7.5°C warming above pre-industrial to amplify melt–elevation feedback across the Antarctic, leading to a complete collapse of the WAIS and retreat of the marine EAIS. Recent simulations in a coupled climate-ice-sheet model also support strong hysteresis of the Antarctic Ice Sheet driven by melt-albedo feedback with associated critical thresholds of atmospheric CO_2 levels for Antarctic Ice Sheet loss [Leloup et al., 2025]. Lastly, considering evidence from previous warm periods, ice sheet mass balance observations, and models, Stokes et al. [2025] recently argued that the current warming level is high enough to see substantial loss of ice sheets, and that a long-term limit of 1°C, or lower, above pre-industrial levels is necessary to avert substantial ice sheet loss.

On potential tipping dynamics, further modelling studies have supported the committed nature of sea level rise from Antarctic Ice Sheet loss [Alevropoulos-Borrill et al., 2024], including some degree of committed sea level rise from having potentially passed a tipping point in the Amundsen Sea sector of West Antarctica [Bett et al., 2024] (although a modelling study by Hill et al. [2023] found that a tipping point has not yet been crossed). The sensitivity of ice sheets to intrusion from warmer sea water is identified as an underestimated cause of tipping points in one model [Bradley & Hewitt, 2024]. Ice sheet modelling shows that basal water conditions can bring forward tipping points by up to 40 years [Zhao et al., 2025]. Atmospheric extreme events - often associated with atmospheric rivers, which are responsible for 50–70 per cent of extreme snowfall events in Antarctica, as well as being involved in the Larsen A and B ice shelf collapses during surface melting events - are not resolved by current-generation models, and could potentially result in faster melting than expected [Kolbe et al., 2025; Wille et al., 2025].

West Antarctica

There is further evidence that the WAIS is committed to long-term collapse at current or near current temperature levels [Van den Akker et al. 2025; Chandler et al., 2025], with each additional fraction of warming increasing the likelihood that collapse could be initiated much sooner. Recent studies show that the WAIS is particularly vulnerable to ocean warming and grounding line retreat [Hill et al., 2024; Rignot et al. 2024]. In the ISMIP6 ensemble of 16 ice flow models up to 2300, Seroussi et al. [2024] found a retreat of the WAIS leading to a rapid increase in sea level rise after 2100, reaching up to 4.4 m SLE by 2300 under high-emission scenarios. Offshore sediment cores also indicate that West Antarctica remained ice-free during the initial formation of the Antarctic Ice Sheet following the Eocene-Oligocene Transition around 34 million years ago, implying lower temperatures are required for WAIS formation as well as loss [Klages et al., 2024].

Recently, a study using three ice sheet models showed that the WAIS might be less vulnerable to MICI than previously thought [Morlighem et al., 2024]. Additionally, ice core data suggests that the West Antarctic Ronne Ice Shelf survived the Last Interglacial (approximately 125,000 years ago) when regional Antarctic temperatures were higher than today [Wolff et al, 2025], thus suggesting the WAIS may be less sensitive to MICI than previously suggested. Conversely, a study currently under review uses empirical data to suggest that MICI may be sensitive to parameters other than cliff height, such as ice thickness gradients, and if these parameters were better resolved in models more future cliff-calving may be projected [Needell, Walker and Bassis, under review]. Additionally, other mechanisms could still lead to a tipping point, and there is still uncertainty around the timing of this [Fricker et al., 2025].

East Antarctica

Hydrological feedbacks, associated with meltwater flowing beneath the ice sheet, which could accelerate ice loss in East Antarctica, are identified in a recent coupled ice sheet-subglacial hydrology model [Pelle et al., 2024]. This suggests that models without these feedbacks could underestimate future sea level rise. EAIS vulnerability is shown through the recent sudden disintegration of the Conger-Glenzer Ice Shelf, which is mapped with remote sensing data in Walker et al. [2024].

Summary

Based on this new research, the GTP community maintains its assessment of the WAIS being a tipping system with high confidence (with high agreement across robust evidence). We also assess that the current lower-end threshold estimate for WAIS of 1°C may be too high to be considered a safe long-term limit [Arthern & Williams 2017; Seroussi et al., 2017; Garbe et al., 2020; Golledge et al., 2021; Reese et al., 2023; Van den Akker et al., 2025; Stokes et al., 2025], and while the exact lower limit is hard to determine, a precautionary limit of 0.5°C would be appropriate. We also maintain our tipping system assessments for marine-based EAIS as high confidence, and non-marine based EAIS with medium confidence.

Sea ice

What it is

When seawater cools below the freezing point in each hemisphere's autumn to spring, it begins to form a layer of floating sea ice. Large areas of highly reflective (i.e. high albedo) white sea ice helps amplify regional cooling, and conversely reduced sea ice extent with global warming is one of the drivers of the 'Arctic amplification' of warming. This feedback was originally thought to lead to a tipping point beyond which sea ice loss becomes self-sustaining [e.g. Lenton et al., 2008], but more recent work instead expects quasi-linear sea ice loss with warming as a result of counteracting negative feedbacks serving to dampen ice loss [e.g., Gregory et al., 2002; Winton, 2006; Winton, 2008; Notz, 2009; Tietsche et al., 2011; Mahlstein and Knutti, 2012: Wagner and Eisenman, 2015]. However, there remains some possibility of sea ice tipping in certain regions and circumstances, such as around Antarctica [Winkelmann, Steinert & Armstrong Mckay et al., 2023].

What's new

In GTPR23, Arctic summer sea ice was assessed as not a tipping system with high confidence, Arctic winter sea ice as not a tipping system with medium confidence, and Barents sea as not a tipping system with low confidence. In contrast, Antarctic sea ice was assessed as unclear, with more evidence required to make an assessment [Winkelmann, Steinert & Armstrong Mckay et al., 2023].

Since GTPR23, several new publications have advanced understanding of sea ice tipping dynamics. Heuzé & Jahn [2024] showed through model simulations that the Arctic Ocean could experience its first entirely ice-free summer day before 2030 under scenarios of continued warming, emphasising that while ice loss is accelerating, it remains largely linear and influenced by acute warming events rather than tipping points. Selivanova et al. [2024] confirmed significant ongoing reductions in Arctic summer sea ice extent and thickness, projecting nearly ice-free summer conditions by the 2040s, indicating a regime shift to a thinner, more transient summer ice cover, but without identifying irreversible thresholds. However, species and ecosystems dependent on sea ice are less recoverable.

In the Barents Sea, Onarheim et al. [2024] documented recent localised thickening due to temporary cooler conditions, emphasising regional variability and responsiveness to short-term climatic fluctuations rather than sustained recovery or tipping behavior.

This is a mixed picture on whether Arctic winter sea ice might reach a tipping point or not. Recent winter ice reductions in the Barents–Kara Seas have resulted largely from anthropogenic forcing rather than feedbacks, though significantly amplified by internal climate variability [Siew et al., 2024]. Wunderling et al. [2024] reviewed potential tipping interactions, noting winter Arctic sea ice might exhibit threshold-like behavior linked to ocean–ice feedbacks and the Atlantic Meridional Overturning Circulation. However, they noted that current evidence does not yet support an irreversible collapse of Arctic winter sea ice. At the same time, abrupt reductions remain plausible at certain warming thresholds, even if these are not feedback-driven tipping events, as indicated in recent CMIP6 analyses [Terpstra et al., 2025].

Other recent work suggests the potential for tipping behaviour in winter sea ice. In a sea ice model, Hankel & Tziperman [2023] show a clear bifurcation: beyond a critical forcing the winter-ice equilibrium vanishes, driving an abrupt, hysteretic transition to a permanently ice-free Arctic. Observationally-constrained detection studies further show that CMIP6 still underestimates greenhouse-gas control on sea-ice loss, implying the threshold for year-round ice collapse lies closer to today's climate than CMIP5 suggested [Kim et al. 2023]. Finally, carbon-removal ensemble experiments indicate that even after CO_2 is drawn back to pre-industrial levels, most models retain an approximate 1 million km² winter-ice deficit — evidence of incomplete recovery and long-lived hysteresis [Yu et al. 2025].

Antarctic sea ice has shown alarming trends recently, with a gradual increase up to 2014 broken by a precipitous decline beyond natural bounds of variability since, rivalling Arctic losses [Abram et al., 2025]. 2023 saw a record-low extent, attributed to anomalously warm ocean conditions and unusual wind patterns [Espinosa et al., 2024]. Raphael et al. [2025] found compelling evidence of a structural regime shift in Antarctic sea ice since the mid-2010s, characterised by unprecedented consecutive low-ice events and decreased recovery capability, signaling potential tipping behavior. Recent improvements in satellite observations have also revealed a reversal in surface freshening in the Southern Ocean, with increasing salinity since 2015 associated with reduced stratification, which could accelerate sea ice loss through increased ocean heat loss [Silvano et al., 2025]. Abram et al. [2025] proposes that the Antarctic sea ice regime shift may feature self-perpetuating dynamics even below 2°C, but it is not yet clear if future projections of decline reflect this or lagged ocean warming.

Recent analysis of abrupt shifts in CMIP6 model results suggest abrupt shifts in Arctic summer sea ice occur in some simulations between 1.0 and 4.6°C, in winter sea ice between 2.4 and 5.4°C, and in Barents sea ice between 1.3 and 2.3°C [Terpstra et al., 2025; Angevaare & Drijfhout, in review] (see Table A2.2.1 in Appendix for details). Similarly, abrupt shifts are also detected around Antarctica in a number of simulations between 0.5 and 5.3°C. However, these abrupt shifts are not necessarily tipping points without confirming self-perpetuating dynamics.

Based on this new research, the GTP community maintains its assessment of Arctic summer sea ice and Barents Sea ice as not tipping systems. Arctic winter sea ice is assessed as uncertain regarding tipping point behavior, as while recent studies suggest there are possible threshold effects and rapid ice loss from significant warming, there is not yet strong proof that this would be irreversible due to self-perpetuating feedbacks. Improved modelling and longer observations of sea ice will help to clarify potential tipping dynamics. We also maintain our assessment of Antarctic sea ice potentially exhibiting tipping behavior as uncertain, reflecting that while there is mounting evidence of a fundamental and possibly irreversible shift in Antarctic sea ice conditions [Raphael et al., 2025], and there are feedbacks that could potentially sustain this shift [Silvano et al., 2025], there is insufficient research confirming the dynamics involved and the likely endpoint.

Mountain glaciers

For more on tipping points in glaciers, see the Mountain glaciers case study

What it is

Outside of the ice sheets of Greenland and Antarctica, ice bodies occur as mountain glaciers, gaining ice in higher altitudes before flowing to lower altitudes where they lose mass. In general glaciers are shrinking and projected to shrink further with global warming, but this mass balance is subject to various feedbacks with the potential for nonlinear responses to a changing climate as a result [Marzeion et al., 2018; Hock et al., 2019; Meredith et al., 2019; Rounce et al., 2023]. These feedbacks include changing flow rates from increased meltwater generation, increased retreat of calving glaciers from a warming ocean, drop of glacier surface elevation increasing melt rates and possibly also decreasing snow accumulation, and increased dustiness and surrounding vegetation reducing local albedo. Under certain circumstances the above feedbacks can result in self-sustained mass loss of individual glaciers [Winkelmann, Steinert & Armstrong Mckay et al., 2023]. Although these feedbacks act mainly on local scales, there is on average a tendency for regional similarities and thus synchronous transitions between different states of glaciers and their downstream impacts.

What's new

In GTPR23, mountain glaciers were assessed as a tipping system at the regional scale with medium confidence, but as not a tipping system at the global scale with medium confidence [Winkelmann, Steinert & Armstrong Mckay et al., 2023]. This assessment has since been confirmed by a number of further considerations and studies. The most recent global-scale compilation of glacier mass loss [The GlaMBIE team, 2025] found that glaciers worldwide lost 273 ± 16 gigatonnes annually from 2000 to 2023, with an increase of 36 ± 10 per cent from 2000–2011 to 2012–2023. These numbers correspond to a loss of between 2 and 39 per cent of regional glacier ice mass, about 5 per cent globally. The glacier mass loss found has already passed the IPCC AR6 lowest mass-loss projections over the period from 2000 to 2040. Glacier mass loss 2000–2023 is about 18 per cent larger than the mass loss from the Greenland Ice Sheet and more than twice that from the Antarctic Ice Sheet. A recent global glacier modelling intercomparison [Zekollari and Schuster et al., 2025] highlights the substantial regional diversity of already committed and further equilibrium response of glaciers worldwide, supporting the GTPR assessment of glaciers being tipping systems at regional scale, rather than global.

While GTPR23 focused mainly on processes of glacier dynamics and mass balance, also the atmospheric forcings behind glacier mass changes underlie nonlinear behaviour, that in turn can then cause nonlinear glacier development. Temperature-precipitation relations are weakly understood and quantified [Ding et al. 2014]. While in some glacier regions the increased humidity of the warming atmosphere leads to increased snowfall and accumulation, in other cases precipitation undergoes a transition to a higher percentage of the liquid phase [Hock et al., 2019]. In addition to changes in the regional atmospheric forcing, glacial landscape changes such as glacier area loss, exposure of rock and debris, formation of lakes, or increased vegetation cover will change temperature and wind patterns [Shaw et al. 2023]. The feedback of these changes on the glaciers themselves is poorly understood, including for instance the transition between sublimation-dominated to melt-dominated glacier ablation regimes [Marshall 2021]. The shift of polythermal glacier regimes (a mixture of ice zones at and zones below the pressure melting point) to temperate thermal regimes (all ice at pressure melting point) is also expected with atmospheric warming, but associated processes and consequences, for instance on meltwater refreezing, runoff, glacier dynamics and even mechanical glacier stability [Gilbert et al. 2018], are little understood [Marshall 2021].

Bolibar et al. [2022] suggest a number of nonlinearities in the relation between temperatures and snowfall, melt and in particular their positive and negative extremes that impact glacier mass balance. The combined impacts of these nonlinear changes in individual forcings can compensate each other towards quasi-linear behaviour. The latter overall forcing combines then with feedbacks related to glacier topography. In that context, Bolibar et al. [2022] point out the particularly important differences between mountain glaciers, which can retreat to higher average elevations where melt rates are reduced (negative, self-stabilising feedback, GTPR23), and flat glaciers (on global average the ones with largest ice volumes), where mass loss reduces average surface elevation and enhances melt rates (positive feedback, GTPR23). Studies for Alaska glaciers confirm GTPR23 findings that topographic controls, in particular elevation distributions, on surface mass balance of ice fields and glaciers can lead to tipping behavior on local to regional scales [Davies et al., 2022; Davies et al, 2024]. The causes and impacts of glacier tipping at the local- to regional scale are explored in the case study on tipping dynamics in an Alaskan (USA) glacial system.

Based on these considerations and new research, we maintain our assessment of mountain glaciers being a tipping system at the regional scale with medium confidence, but as not a tipping system at the global scale (medium confidence).

Permafrost

What it is

Permafrost consists of ground frozen for at least two consecutive years [Harris et al., 1988], and underlies about 14 million km^2 (15 per cent of the land surface area) in the Northern Hemisphere [Obu, 2021; Steinert et al. 2023]. Freezing prevents organic matter from tundra or boreal forest ecosystems entering the soil from decomposing, resulting in the buildup of over ~1000 GtC in the top 3m of permafrost soils on land [Hugelius et al., 2014]. However, global warming is leading to some of this permafrost beginning to thaw, allowing the preserved organic matter to degrade and emit greenhouse gases in the process, primarily as CO_2 but with a proportion as high-warming methane where permafrost is waterlogged [Walter Anthony et al., 2014]. Much of the carbon loss is likely irreversible due to the slow formation timescales of permafrost and sustained microbial decomposition of previously frozen organic matter, leading to continued carbon emissions over centennial to millennial timescales and reinforcing warming through a positive feedback loop [Schwinger et al. 2022; de Vrese & Brovkin et al. 2021; Park et al. 2025; Ji et al. 2025]. Furthermore, permafrost thaw does not occur uniformly, as some areas experience localised abrupt thaw, leading to rapid carbon loss and landscape destabilisation, i.e., degradation of ice-rich permafrost, and subsequent rapid slope slumping, ground subsidence and the formation of thermokarst landscapes. These processes could amplify emissions by 40 per cent under high emission scenarios but are not currently represented in Earth system models [Turetsky et al. 2020]. While large-scale permafrost thaw is gradual, these regional abrupt thaw processes involve positive feedbacks such as thermokarst formation that can lead to self-sustained thawing processes, allowing for localised tipping to take place [Nitzbon et al., 2020]. However, at the regional to global scale permafrost thaw is expected to aggregate to a quasi-linear response to global warming [Nitzbon et al. 2024].

What's new

In GTPR23, land-based permafrost was assessed as a tipping system at the regional scale (medium confidence), but not as a tipping system at the global scale (medium confidence). Equally, subsea permafrost was not identified as a tipping system (medium confidence) [Winkelmann, Steinert & Armstrong Mckay et al., 2023]. Since GTPR23, several new publications have advanced understanding of permafrost thaw in the context of tipping dynamics.

Evidence suggests that permafrost thaw remains characterised by multiple regional-scale tipping processes, including abrupt thermokarst lake formation and slope slumping, rather than exhibiting a single global tipping threshold [Nitzbon et al., 2024]. A recent analysis of CMIP6 model results identified individually small but widely distributed (collectively aggregating to over >1M km^2) abrupt shifts in land permafrost frozen soil moisture content in eight models between 1.0 and 3.8°C of global mean warming, and in soil frozen water content in 19 models between 1.2 and 3.3°C [Terpstra et al., 2025] (Table A2.2.1). Localised, abrupt thaw events [Webb et al. 2025] contribute cumulatively but gradually at the global level, thus reaffirming that permafrost degradation progresses incrementally and heterogeneously. The presence of multiple steady states in permafrost systems indicates the potential for local tipping of ecosystems and soil carbon storages on centennial timescales [Brovkin et al. 2025]. Further, Earth System simulations suggest that changes in permafrost hydrology can gradually, rather than abruptly, impact hydroclimate in the tropics and subtropics. The permafrost soil state and consequential carbon fluxes of such processes have the potential to be underestimated in low-resolution climate models [Schickhoff et al. 2024].

Recent findings from NOAA's 2024 Arctic Report Card [2024] revealed that Arctic tundra ecosystems have shifted to net carbon sources earlier than projected, driven by intensified permafrost thaw and record wildfire seasons. This transitioning of Arctic ecosystems from carbon sinks to sources is amplifying climate feedbacks at regional scales. A previously overlooked feedback mechanism linking permafrost thaw to reduced cloud cover, amplifying Arctic warming and its global impacts was identified by de Vrese at al. [2024].

Localised abrupt thaw expose deeper layers of organic material to microbial decomposition, significantly increasing emissions in a short period, and significantly contributing to methane emissions from inundated areas [Park et al. 2025], while emissions from deep Arctic lake sediment could be more substantial than previously thought [Freitas et al. 2025]. As such, abrupt thaw hotspots across the Arctic highlight the vulnerability of permafrost regions even under present-day conditions. Using the CESM2 model coupled with CLM5, Park et al. [2025] further determined that permafrost carbon emissions will persist for centuries even under scenarios of aggressive climate mitigation and net-negative emissions. Their model simulations indicate irreversible commitments to continued carbon release from thawed permafrost, further reinforcing the long-term implications of regional tipping processes.

A warming Arctic enhances the risk of causing localised abrupt thaw and increases its magnitude and therefore also increases the risk of tipping cascades, where permafrost degradation - in addition to its carbon-climate feedback - interacts with other Earth system components, such as boreal forest dieback [Alfaro-Sánchez et al. 2024], including wildfire occurrence [Kim et al., 2024] or a modified ocean circulation [Schwinger et al. 2022, Park et al. 2025, Steinert et al. 2025] and carbon uptake efficiency [Nielsen et al. 2024], potentially accelerating global climate change. Further, adaptive emission driven MPI-ESM simulations show that, while average annual permafrost carbon emissions of ~0.3–0.7 GtC/yr are small compared to present-day fossil fuel emissions (~10 GtC/yr), permafrost thaw can still reduce the carbon budget by ~11–13 per cent by 2300 under 2°C or 3°C warming scenarios [Georgievski et al., 2025].

Based on this new research, the GTP community maintains its assessment of land-based permafrost as a tipping system at the regional scale with medium confidence, reflecting ongoing evidence of localised tipping processes. It also maintains medium confidence that subsea permafrost is not a tipping system, as well as that land-based permafrost is not a tipping system globally, given the absence of a unified global threshold.

2.2.6 Potential tipping points in the biosphere

The ‘biosphere’ consists of all life on Earth and includes the major biomes of tropical, temperate, and boreal forests, more open ecosystems across grasslands, savannas, and drylands, and aquatic ecosystems across freshwater, coastal, and marine environments.

In this section, we describe each of these biomes, their evidence for tipping dynamics, and relevant new research in turn. Based on this, we have reassessed the status of several systems, including lowering the minimum warming threshold for the Amazon, adding river deltas and peat bogs as potential tipping systems, and clarifying local-to regional-scale dynamics in the Congo and temperate forests (Figure 2.2.3).

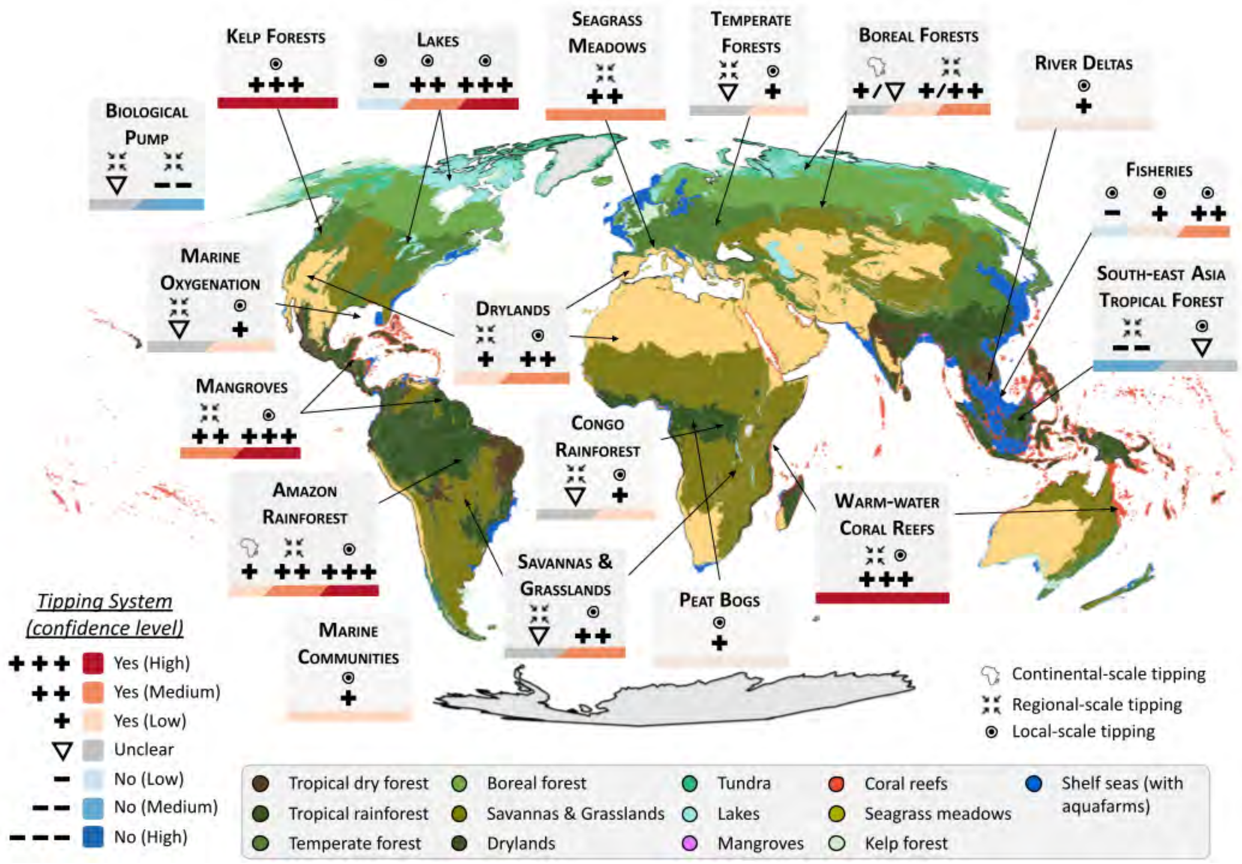

Figure 2.2.3: Map of biosphere systems considered in this chapter. Systems are marked by the coloured areas, with terrestrial biomes and mangroves based on biogeographic biomes [Dinerstein et al., 2017], and lakes and ocean biomes on IUCN functional biomes [Keith et al., 2022] (lakes are shown over other biomes for tundra only; fisheries are spread across the global ocean, but are marked only on key coastal seas for simplicity [Steinert, 2023]). 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 (▽).

Tropical forests

What it is

Known for their high levels of biodiversity [Slik et al., 2015; Pillay et al., 2021; Aguirre-Gutiérrez et al. 2025], tropical forests cover ~1.95 billion hectares and store substantial amounts of carbon within their biomass and soils (circa 471 +/- 93 GtC) [Pan et al., 2011; SPA, 2021; Ornetto et al., 2022]. They form a key part of the interconnected Earth system and have far reaching impacts on the climate through evapotranspiration and cloud formation. They are mainly threatened by deforestation, for example to create pastures, land-cover change, for example to plantations, and droughts and wildfires being worsened by climate change [Sternberg, 2001; Franco et al., 2025]. There are many tropical forest feedbacks, but two key positive feedback mechanisms, acting at different spatial scales, have been identified which may drive tipping points in tropical forests: the forest-rainfall feedback at regional scales and the fire-vegetation feedback at local scales [Flores & Staal, 2022; A. McKay, Sakschewski & Roman-Cuesta et al., 2023]. The forest-rainfall feedback operates at a regional level, with moisture being recycled throughout tropical forests, thus reducing the impact of rainfall variability on forest health; however deforestation and climate change-induced drought extremes can reduce this moisture recycling and force the forest towards a tipping point [Staal et al., 2020]. This feedback is present in the Amazon [Zemp et al., 2017; Staal et al., 2018] and Congo forests [Staal et al., 2020], however it is less important in Southeast Asian and Australasian rainforests where ocean-derived rainfall is plentiful. At a local level, the fire-vegetation feedback can cause a transition from tropical forests to a more open state (such as savanna or degraded dry forest). Less dense tree cover can lead to increased likelihood of fires due to the spread of grasses and local drier air, thus further reducing tree cover [Cochrane et al. 1999].

What's new

Global

Remote sensing by satellites of major forest disturbances (mainly wildfires and droughts) were recently used to calculate a hydrologic sensitivity index, establishing critical thresholds beyond which forest loss drives drastic changes in water yield and climate conditions [Domínguez-Tuda & Gutiérrez-Jurado, 2024]. In tropical rainforests, a threshold near 16 per cent tree cover reduction at local-to-landscape scales was identified, beyond which water yields notably decrease and warming trends intensify.

Amazon rainforest

For more on tipping points in the Amazon rainforest, see the Amazon case study

In GTPR23, the Amazon rainforest was assessed as a tipping system with high confidence at the local scale, medium confidence at the regional scale and low confidence at the continental scale, with medium confidence in it involving abrupt / large rate change, medium confidence in irreversibility over decadal/centennial timescales, and a threshold range for dieback of 1000-1250 mm mean annual rainfall, ~400 to ~450 mm maximum accumulated water deficit, a dry season length of 7-8 months, deforestation levels of 20-40 per cent and ~3.5°C (2-6°C) of global warming [A. McKay, Sakschewski & Roman-Cuesta et al., 2023].

Since GTPR23, the Amazon has faced droughts and extreme warmth associated with the 2023-24 El Niño event, which has led to water stress, increased fires, and reduced greenness [Jiménez et al., 2024]. A major recent synthesis of research on the potential for critical transitions (such as dieback) in the Amazon forest found that by 2050 the potential effects of compounding disturbances on Amazonian systemic resilience could see around 10-47 per cent of Amazon forests facing combined stresses beyond critical thresholds, potentially triggering irreversible regime shifts and exacerbating regional climate change [Flores et al., 2024]. Based on this, Flores et al. [2024] suggested precautionary limits of 1.5°C and 10 per cent deforestation (requiring restoration of 5 per cent of the biome) to avoid broad-scale ecosystem transitions, the latter reduced from the 20-25 per cent precautionary limit of Lovejoy & Nobre [2018].

Similarly, under review modelling suggests that while regional dieback would occur at 3.7-4.0°C without deforestation (in line with A. McKay et al. [2022]), deforestation reaching 22-28 per cent of current forest extent would reduce the warming threshold to 1.5–1.9°C of global warming and make dieback more widespread [Wunderling et al., in review]. Recent analysis of abrupt shifts in CMIP6 model results detected some abrupt shifts in Amazon vegetation between 0.9 and 5°C [Terpstra et al., 2025] (Table A2.2.1), but they were not consistent across models, and do not necessarily represent tipping dynamics. Running CMIP5/6 models until 2300 and including slower transitions reveals localised to regional scale dieback in nine out of twelve models, with location and thresholds (global warming 1.5-10.2 °C, local surface air temperatures >32.2 ± 4.8°C, precipitation <1394.3 ± 306.0 mm/yr) highly model-dependent [Melinkova et al., 2025].

A simpler land-surface-atmosphere model has found that under projected rainfall decreases deforestation of 45 and 55 per cent could trigger dieback, but this model did not feature spatial variation or more complex vegetation dynamics [Hajdu et al., 2025]. Conversely, Yoon & Hohenegger [2025] found that better representing atmospheric convection in a storm-resolving model limited rainfall's sensitivity to deforestation, although only in a short simulation. Under high emission scenarios, it has been estimated that more than 25 per cent of the forest in Central Amazonia could become a net carbon source under high emission scenarios, with drying trends reducing biomass and triggering regime shifts particularly if eastern Pacific temperatures rise >1.5°C (globally, >2.3°C) [Nath et al., 2024], while based on estimates of root zone moisture storage the area of forest at risk of regime shifts jumps by ~1.7–5.8 times (relative to <2°C warming) [Singh et al., 2024].

Potential 'early warning signals', in the form of 'slowing down' in system response to disturbances which can indicate resilience loss prior to a tipping point, have previously been identified in the Amazon in empirical data [Boulton et al., 2022] and models [Boulton et al., 2013; Bochow & Boers 2023]. However, recent studies have found a more heterogeneous or unclear response of forest resilience than earlier estimates, while other co-drivers of resilience loss can mask slowing down [Blaschke et al., 2024; Grodofzig et al., 2024; van Passel et al., 2024]. Despite this, more widespread resilience loss is still expected due to climate change in the future, and areas with greater deforestation or disturbance are already less resilient [Wang et al., 2024] and leading to greater seasonality in rainfall in those regions [Qin et al., 2025].

In GTPR23 the potential impacts of AMOC slowdown or collapse on the Amazon Rainforest were unclear, but several new studies have shone new light on this (see the Interactions & Cascades section).

Based on this new research, the GTP community maintains its assessment of the Amazon rainforest being a tipping system with low confidence at the continental-scale, medium confidence at the regional-scale, and high confidence at the local-scale (with robust evidence but low to medium agreement). While evidence has grown for larger-scale tipping, particularly on the basis of the assessment of Flores et al. [2024], and we now have high confidence in irreversibility over decadal/centennial timescales, continued model limitations and disagreements over location and thresholds limits current confidence level. However, we lower the minimum warming threshold from 2 to 1.5°C based on recent assessments of warming/deforestation synergies [Flores et al., 2024; Wunderling et al., in review; Melinkova et al., 2025], with the lower end of this range more likely when considering both warming and deforestation.

Congo rainforest

In GTPR23, the Congo rainforest was assessed as a tipping system with low confidence at the local scale, but unlikely to tip as a result of climate change, with low confidence in it involving abrupt / large rate change, low confidence in irreversibility over decadal/centennial timescales, and a threshold range of ~1350 mm mean annual rainfall [A. McKay, Sakschewski & Roman-Cuesta et al., 2023]. Since then, remote-sensing based assessments of local-scale bistability of high and low tree cover in the Congo have been refined [Zwaan et al., 2024]. While the hypothesis of local-scale bistability is supported [Staver et al., 2011; Aleman et al., 2020], the results indicate that transitions between closed forest and open savanna could instead pass through a state of coexistence, which would likely smoothen out tipping points between these states over larger spatial scales. In contrast, estimates based on root zone moisture storage project that the Congo Basin forest area at risk of critical transitions grows by ~0.7–1.7x under higher warming scenarios (relative to <2°C warming) [Singh et al., 2024].

Based on consistent evidence for bistability but continued limited agreement and evidence for wider scale tipping dynamics under future climate projections, we maintain our assessment of it being a low confidence tipping system at the local scale, and add that it is uncertain at regional scales. Overall, tipping may be more localised in comparison to the Amazon [Zwaan et al., 2024].

Southeast Asian rainforest

The Southeast Asian rainforest was assessed in GTPR23 as an uncertain tipping system at the local scale and not a tipping system with medium confidence at the regional scale, with low confidence in it involving abrupt / large rate change, low confidence in irreversibility over decadal/centennial timescales, and a threshold range of ~1550 mm mean annual rainfall [A. McKay, Sakschewski & Roman-Cuesta et al., 2023]. In the absence of substantial new research on this area, we maintain this assessment here.

Boreal forests & tundra

What it is

Boreal forests occupy ~1.14 billion hectares in the high latitude regions of the northern hemisphere [Pan et al., 2011]. They face disturbances from fire and insect outbreaks, as well as logging [Kuuluvainen & Gauthier, 2018]. Situated in an area with amplified climate change, there are two potential tipping points associated with boreal forests - one at its northern edge, where forest may expand into the tundra, and the other at the south, where the forest may dieback and transition to an open steppe/prairie landscape. Southern boreal forest dieback could be driven by the fire-vegetation feedback [Joos et al., 2001; Lucht et al., 2006; Lenton et al., 2008; Abis and Brovkin 2017; Rotbarth et al. 2023]. With high latitude temperatures increasing, there is evidence of increased survival rate of seedlings in the tundra and an advancing shrubline, with potential positive feedbacks (such as albedo and soil moisture feedback) leading to further forest expansion [Myers-Smith et al., 2011].

What's new

Dieback

In GTPR23, boreal forest dieback was assessed as a tipping system with medium confidence at the regional scale and low confidence at the continental scale, with medium confidence in it involving abrupt / large rate change, low confidence in irreversibility over decadal/centennial timescales, and a threshold of ~4°C (range 1.4-5°C) [A. McKay, Sakschewski & Roman-Cuesta et al., 2023].

Recent simulations found that boreal forests may be shifting from the current multi-stable state towards a unimodal semi-open state with 30-50 per cent tree cover in the coming decades [Rotbarth et al., 2025]. Such a shift would likely increase the risk of forest fires, leading to potentially substantial releases of stored carbon. However, these results are based only on the inferred relationship between tree cover change and mean annual temperature, with added stochasticity to represent process noise, and do not consider other important factors affecting the boreal forest dynamics, such as permafrost thaw or water and nutrient availability.

Warming will likely have consequences for boreal forest biodiversity and the likelihood of dieback. An average increase in tree species diversity by 12 per cent has been observed across boreal forests between 2000 and 2020 [Xi et al., 2024]. However, a negative impact was observed in areas of extreme warming (>0.065 °C /yr), suggesting that exceeding a certain threshold of warming could have detrimental effects. Repeated cycles of clear-cutting in boreal forests are also reducing old and large trees, deadwood diversity, and altering soil composition, causing a long-term decline in species richness [Lunde at al., 2025].

A recent satellite observations-based hydrologic sensitivity index [Domínguez-Tuda & Gutiérrez-Jurado, 2024] indicates that areas impacted by forest loss with a tree cover reduction higher than 18 per cent exhibited more pronounced warming trends and a rapid rise in hydrologic responses compared to areas with smaller losses. Similarly, it has been shown that tree growth in Eurasian larch forests is being increasingly limited by rising temperatures and the associated drought stress, leading to negative response to warming [Li et al., 2023]. Recent findings also show that across northwestern North America warming and disturbances are affecting vegetation resilience, which declined significantly in the southern boreal forest, including some regions exhibiting overall greening, but increased in much of the Arctic tundra [Zhang et al., 2024].

Based on this new research, we assess that while there is some increased evidence for dieback at the regional scale, we maintain our assessment at medium confidence in the absence of more evidence, and maintain low confidence at the continental scale.

Northern expansion

Boreal forest northern expansion was assessed in GTPR23 as a tipping system with low confidence, with low confidence in it involving abrupt / large rate change, low confidence in irreversibility over decadal/centennial timescales, and a threshold of ~4°C (range 1.5-7.2°C) [A. McKay, Sakschewski & Roman-Cuesta et al., 2023]. Since GTPR23, there have been several new publications relevant to this tipping system.

On the boreal forest's northern edge, transitional forests located between boreal forests and tundra are experiencing consistent increases in vegetation height and density [Montesano et al., 2024]. These changes are driven by Arctic amplification and are expected to continue through 2100 across all climate scenarios. Furthermore, a population of white spruce (*Picea glauca*) across an Arctic basin in North America has been documented advancing at rates that cannot be sustained by warming alone [Dial et al., 2022]. However, significantly reduced tree growth has also been found on thawing permafrost in the higher latitudes of North America [Alfaro-Sánchez et al., 2024], as trees need to invest more into remaining upright on destabilised grounds. Recent analysis of abrupt shifts in CMIP6 model results detected some abrupt shifts in boreal vegetation between 0.8 and 4.9°C, mostly for northward expansion [Terpstra et al., 2025] (Table A2.2.1), but they were not consistent across models, and do not necessarily represent tipping dynamics.

Based on this new research, while there is some agreement that climate change will likely induce tipping points in the expansion of the boreal forest, the still limited evidence base means the GTP community maintains its assessment of this being a tipping system with low confidence at the regional scale.

Temperate forests

What it is

Temperate forests make up 16 per cent of the global forest area [Hansen et al., 2010; Pan et al., 2011]. The majority of temperate forests are spatially fragmented and are managed, low biodiversity ecosystems [Potapov et al., 2017; Sabatini et al., 2021]. These management practices are likely to lead to large areas of forests with a lower resilience to perturbations, with forests exposed to droughts, heatwaves and pest outbreaks [Allen et al., 2010; Buras et al., 2019; Billing et al., 2020; Senf et al., 2020; Zhang et al., 2021; Carnicer et al., 2021; Benyon et al., 2023; Forzieri et al., 2022], with a potential of widespread dieback from these. Temperate forests may experience localised feedback dynamics from fire and bark beetle attacks in common with boreal forests (see above). Further investigation is required to establish the strength of the forest-moisture feedback in temperateforests. Large uncertainties remain around whether temperate forests may experience tipping points across a large scale [A. McKay et al., 2022], with some assessments suggesting this is unlikely at present [Thom, 2023].

What's new

In GTPR23, temperate forest dieback was assessed as a tipping system with low confidence due to limited evidence, with medium confidence in it involving abrupt / large rate change, low confidence in there not being irreversibility over decadal/centennial timescales, and uncertainty around any threshold ranges [A. McKay, Sakschewski & Roman-Cuesta et al., 2023].

Since GTPR23, there have been several new publications relevant to this tipping system. A recent satellite observations-based hydrologic sensitivity index [Domínguez-Tuda & Gutiérrez-Jurado, 2024] identified a threshold of around 46 per cent tree cover reduction for temperate coniferous forests, which leads to cooler climate conditions and higher water yield once surpassed. For Mediterranean woodlands, a threshold of roughly 54 per cent emerged, indicating relatively higher resilience but also rapid hydrologic shifts once that critical point is crossed. However, the degree to which these hydrological shifts involve tipping dynamics is unclear.

Several recent studies show new evidence of localised resilience loss and potential tipping points in different regions of temperate forests. In northwestern China, ecosystem productivity and photosynthetic efficiency have decoupled since 2010 [Zhang et al, 2024]. This indicates a loss of ecosystem resilience in these forests, which under rapid warming/drying flags a near-term dieback threshold in water-stressed regions. In Europe, the 2018-20 drought event led to a breakdown in standard forest dynamics in a German Beech forest [Mathes et al., 2023]. The drought's intensity potentially induced a nonlinear weakening of dominant trees, and future drought events of this or greater intensity could lead to a regime shift in such Beech forests.

Recent analysis of the relationship between forest fragmentation and ecosystem resilience revealed a clear nonlinear decline in resilience once forest connectivity dropped below critical thresholds (e.g. when number of patches per unit area increases beyond 0.89) [Fu et al., 2024]. In other words, when forest areas become too fragmented, their ability to recover from disturbances, such as droughts, fires, or storms, is significantly reduced. Notably, the study found that agricultural expansion had a more detrimental impact on forest resilience than urban development. While some degree of fragmentation may promote habitat diversity, exceeding certain fragmentation levels leads to a sharp and lasting decline in ecosystem stability, underscoring the importance of preserving large, contiguous forest areas.

While this new research supports the presence of tipping points in some temperate forest systems, and several temperate ecoregions have been subject to increasingly extreme heatwaves [Barriopedro et al., 2011; Sutanto et al., 2020; Lucarini et al., 2023], due to the evidence remaining limited the GTP community maintains its assessment of temperate forest dieback was assessed as an uncertain potential tipping system at the regional scale. However, we now assign low confidence to temperate forest tipping at the local scale.

Savannas & grasslands

What it is

Savanna and grassland are ecosystems dominated by grass cover intermixed, in savannas, with variable tree cover [Bond et al., 2008; Staver et al., 2018]. They face threats from conversion to agriculture [Stevens et al., 2022; Strömberg & Staver, 2022], woody encroachment [Stevens et al., 2017; Rosan et al., 2019], afforestation for carbon mitigation [Parr et al., 2024], and climate change via e.g. changes in rainfall variability [D'Onofrio et al., 2019], with major associated losses in ecosystem functions especially on the ground [Ding & Eldridge, 2024]. Savannas are distinct, biodiverse ecosystems, not degraded forest systems [Veldman & Putz, 2011; Veldman et al., 2013; Nerlekar & Veldman, 2020] or candidates for afforestation [Parr et al., 2024]. In some regions, savannas and forests represent potential alternative stable states [Hirota et al., 2011; Staver et al., 2011; Aleman et al., 2020], with open savanna states leading to a buildup of flammable grass material which can cause wildfires and limit tree growth. This open savanna-fire feedback loop can be disrupted by active and passive suppression of fires [Durigan & Ratter, 2016; Andela et al., 2017], which enables forest expansion into savannas [Stevens et al., 2017]. Palaeoecological evidence and fire studies have shown that this savanna to forest transition can be irreversible [Shanahan et al., 2008; Karp et al., 2023]. In some arid regions, savannas and grasslands also represent an alternative stable state to low vegetation cover with substantial bare ground [Hirota et al., 2011], discussed more fully in the Drylands section below.

What's new

In GTPR23, savannas and grasslands were assessed as a tipping system with medium confidence at a local-to-landscape scale. Tipping dynamics likely emerge over decades, resulting in low confidence in the possibility of abrupt / large rate change but with medium confidence in irreversibility over decadal/centennial timescales. Mechanisms involve decreases below ~60 per cent flammable cover that could prevent fire percolation, regionally variable and highly localised rainfall thresholds, and the influence of of CO_2 fertilisation. It is unknown to what extent savanna and grassland tipping points might scale up to emergent and synchronised events at larger regional scales, so this potential was assessed as uncertain.

Since GTPR23, there have been several new publications relevant to this tipping system. Higgins et al. [2024] synthesised several past studies arguing for widespread savanna-forest bistability, showing that a range of different approaches all produce savanna-forest bistability but that there is a substantial uncertainty in the climate thresholds associated with tipping points, consistent with our previous assessment. Several publications have also examined the possible contributions of spatial patterning and mosaics to avoiding tipping at larger scales in these systems [Zwaan et al., 2024; van der Voort et al., 2025]. Finally, a range of work showed that afforestation is accelerating potentially irreversible losses of savanna ecosystems [Loft et al., 2024; Parr et al., 2024] and has elaborated the potential for lost ecosystem services as a result of savanna encroachment [Ding & Eldridge, 2024].

Based on this new research, the GTP community maintains its assessment of savannas and grasslands being a tipping system with medium confidence at a local to landscape scale, and uncertain at the regional scale.

Drylands

What it is

Drylands consist of numerous vegetation types, including deserts, grasslands, shrublands, woodlands, savannas, Mediterranean forests and tropical dry forests, all defined by their aridity level (where the rainfall is lower than 65 per cent of the 'potential evapotranspiration', including hyper-arid, arid, semi-arid and pre-sub-humid climate zones) [Maestre et al., 2016; D'Odorico et al., 2013]. Some of these land cover types are covered in more depth in dedicated sections for 'tropical forests' and 'savannas', as well as the relevant feedback loops of vegetation-fire feedback and vegetation-rainfall feedback. Other feedbacks are possible, including at a small scale, e.g. microbial communities in soil influence the level of soil carbon stocks by decomposing organic. This aids moisture retention in dry soils which is in turn necessary for the decomposition of organic matter, thus forming a feedback loop. Interactions between plants can form important feedbacks in dryland, including around the formation of regular patterns which can occur from plants affecting local soil conditions, such as water and nutrient retention, which create 'islands of fertility' [Eldridge et al., 2024] and can create ecohydrological feedbacks at a large scale through plant connectivity. Additionally, excessive grazing by herbivores can drive changes in plant communities. Evidence exists for vegetation cover bistability in drylands around an aridity level (calculated as one minus the ratio of precipitation to potential evapotranspiration) of between 0.75 and 0.8 [Kefi et al., 2024], with different states of soil fertility, nutrient capture and nutrient recycling [Berdugo et al., 2017]. Studies have identified hysteresis in drylands through palaeorecords [Xu et al., 2020], remote sensing [Zhao et al., 2020] and field studies [Berdugo et al., 2017].

What's new

In GTPR23, land degradation in drylands was assessed as a tipping system with medium confidence at the local to landscape scale and low confidence at the regional scale, with medium confidence in it involving abrupt / large rate change, low confidence in irreversibility over decadal/centennial timescales, and three potential thresholds at aridity levels of 0.54, 0.7 and 0.8. Since GTPR23, there have been several new publications relevant to this tipping system.

On mechanisms, self-organised spatial vegetation patterns in drylands can enhance resilience to increasing aridity by facilitating resource redistribution. However, this ability is often lost in degraded ecosystems, with recent work showing this ability may break down beyond an extreme aridity level of 0.8 [Kéfi et al., 2024]. Land-atmosphere feedbacks involving existing drylands can also contribute to their own expansion. Warming and drying of air flowing over drylands can lead to reduced precipitation and increased atmospheric water demand in downwind humid regions, causing aridification [Koppa et al. 2024].

On the drivers of stress & thresholds, climate variability and seasonality were identified as significant environmental factors explaining abrupt changes in dryland Normalised Difference Vegetation Index (NDVI; a measure of vegetation greenness derived from satellite observations) [Berdugo et al., 2022]. Higher rainfall interannual variability is associated with increased vulnerability to abrupt shifts, while long-term trends in rainfall are a major driver of future abrupt shift susceptibility, with decreasing trends increasing the risk [Bernardino et al. 2025].

Aridity values reaching around 0.8 appears to be a threshold separating zones with contrasting dynamical behaviors for positive and negative abrupt shifts in NDVI [Berdugo et al., 2022]. Negative abrupt changes in NDVI are less likely after crossing this aridity threshold, while positive abrupt changes are more likely [Berdugo et al., 2022]. Kéfi et al. [2024] found bimodality in vegetation cover in a global dryland data set, which is consistent with the bistability predicted by dryland models and also supports a threshold at an aridity value of 0.8 (consistent with Berdugo et al. [2020]). Human activities, such as grazing, can amplify the effects of aridity on ecological thresholds. For example, grazing in China's drylands can act in synergy with aridity on dryland structure and functioning, and can therefore lower the aridity thresholds at which abrupt decreases in productivity, soil fertility, and plant richness occur [Li et al., 2023].

On the recent dynamics of dryland responses, while gains in vegetation productivity were more frequent than losses in the last two decades, 50 per cent of the areas experiencing significant changes in productivity showed abrupt (rather than gradual) changes in time [Berdugo et al., 2022]. Abrupt changes were more common among negative than positive NDVI trends and could be found in global regions suffering recent droughts, particularly around critical aridity thresholds.

On potential indicators of dryland tipping, in a study of global drylands lower functioning was found to be associated with impaired ability of the vegetation to self-organise into patchy spatial structure [Kefi et al., 2024]. Trends in spatial patterns observed along large scale dryland gradients matched model prediction, strengthening the idea that patterning is a mechanism of resilience and a possible indicator of ecosystem degradation. Similarly, a machine learning-based approach to detect early warning signals of abrupt shifts in ecosystem functioning was used in one of the world's largest dryland regions, the Sudano–Sahelian zone, and showed its applicability to identify regions that are more likely to undergo a future abrupt shift [Bernardino et al., 2025].

On projections of global dryland expansion, the contribution of reduced precipitation and increased evapotranspiration from existing drylands to ongoing dryland expansion was recently quantified, differentiating it from influences originating in other regions [Koppa et al., 2024]. This established that vegetation-climate feedbacks contribute to 50 per cent of the recent desertification of drylands areas, illustrating how existing dry regions contribute to the intensification and spread of aridity worldwide through feedbacks. However, an investigation of the effects of future climates on dryland productivity found that climate change might promote desertification in less than 4 per cent of current dryland areas, with the fertilisation effect from CO_2 emissions likely overcoming the projected increase in arid conditions in other areas [Zhang et al., 2024].

Based on this new research, the GTP community maintains its assessment of drylands being a tipping system with medium confidence at the local to landscape scale and low confidence at the regional scale.

Freshwater ecosystems

What it is

Freshwater ecosystems include lakes, ponds, wetlands, and rivers, all subject to major climate impacts. Lakes are present across much of the world and represent an iconic early example of ecosystem tipping points, hysteresis and resilience [Holling, 1973; Scheffer et al., 1993; Scheffer & van Nes, 2007]. Closely intertwined with the wellbeing of communities connected with them, lakes form part of a socio-ecological system which may be affected by rapid changes in lake state. Empirical evidence exists for tipping points in shallow lake systems [Scheffer et al., 2001; Tátrai et al., 2008], the most common type of lake globally. Anthropogenically driven eutrophication in lakes can be persistent, and difficult to reverse due to internal phosphorus-loading, whereby phosphorus stock-piled during periods of high pollution is mobilised from the sediment to surface waters following pollution reduction [Jeppesen et al., 1991; Spears & Steinman, 2020]. Lake warming enhances this effect, which together with nutrient pollution can also lead to an increase in greenhouse gases released from the lake, thus feeding back to global warming, but this may be countered to an extent by increased carbon burial processes [Anderson et al., 2020]. Critical phosphorus concentrations in shallow lakes are known to be highly variable with lake depth, retention time and fetch [Janse et al., 2008]. Modelled and empirical studies suggest thresholds in the range of 80–120 and 40–60 mg total phosphorus per m^{-3} lake water, respectively, for forward and reverse switches in both temperate and tropical systems [Wang et al., 2014; Springmann et al., 2018], though limitations in contemporary empirical data to explain non-linear relationships between chlorophyll a and nutrient concentrations in lakes requires further attention [Davidson et al., 2023].

Another potential tipping point for shallow lake systems involves increased levels of dissolved organic matter (DOM) from terrestrial sources due to land cover changes, such as afforestation, and a changing climate [Creed et al., 2018]. This process, also known as 'browning', occurs in boreal systems and can lead to increased stratification, net heterotrophy and anoxia, and ultimately increased greenhouse gas release [Jeppesen et al., 1991; Spears & Steinman, 2020], thus suggesting a potential positive feedback loop. The timescale of this 'browning' feedback loop is still uncertain [Hessen et al., 2024]. In permafrost regions, appearance or loss of waterbodies is tightly linked to permafrost thaw (e.g. thermokarst formation), and as such what could be considered a lake system tipping point is in fact a result of an underlying permafrost thaw tipping point [Hessen et al., 2024]. Rivers are not addressed in detail in this report, yet both extreme flood and droughts have major impact on wetlands, deltas, coastal areas and a renga of human activities. Glacial fed rivers clearly will be affected by disappearing glaciers [Milner et al 2008].

What's new

In GTPR23, eutrophication-driven anoxia in lakes was assessed as a tipping system with high confidence at the localised scale, with high confidence in it involving abrupt / large rate change and medium confidence in irreversibility over decadal / centennial timescales. DOM-loading, also known as 'browning', in lakes was assessed as a tipping system with medium confidence at the local scale, with low confidence in it involving abrupt / large rate change, medium confidence in irreversibility over decadal / centennial timescales, and a threshold range of >10 mg DOC/l. Other potential abrupt changes in lake ecosystems were identified in GTPR23, including disappearance / appearance of freshwater bodies, switch between Nitrogen and Phosphorus limitation, salinisation and the spread of invasive species. Some of the transitions to saline ecosystems are permanent and give rise to losses of biodiversity and changes in functions and services [Cunillera-Montcusí et al., 2022], but these were not classified as tipping points due to an absence of clear self-sustaining feedbacks. Several of these tipping points will imply positive feedback in terms of increased GHG emission [Rosentreter et al. 2021].

Since GTPR23, there have been several new publications relevant to this tipping system. Hessen et al. [2024] discussed candidate tipping points for lakes, while the concept of tipping points vs. 'tipping sets' have been explored using lake eutrophication as a case study [Mathias et al., 2024]. Lake drainage has been studied as a tipping point case [Liu et al., 2024], while the use of remote sensing for assessing lake tipping points has also been discussed [Gilarranz et al., 2022; Lenton et al., 2024]. Meanwhile, recent analysis found that compound weather extremes (heatwaves and extreme rainfall) in 2022 drove abrupt 'browning' shifts across multiple West Greenland lakes, altering biological and biogeochemical structure [Saros et al., 2025].

Recent works also address the link between permafrost thaw and expansion (or loss) of permafrost ponds linked to this. The widespread increase in thermokarst ponds has been linked to topography [Abolt et al., 2024], while other regions are prone to substantial loss of surface waters due to permafrost thaw, with Northern Sweden for example seeing thermokarst pond area and number decreasing by 6 and 27 per cent per decade, respectively, between 2003 and 2021 [Seeman & Sannel, 2024]. Of particular relevance is the work by Brovkin et al. [2025], arguing that permafrost and freshwater systems in the Arctic are inextricably linked in their tipping dynamics, and that hydrological changes in the permafrost region could have impacts on global hydroclimate.

GTPR23 only dealt with lakes, but given the prevalence of wetlands, their susceptibility to climate change (and other anthropogenic forcings), and not least their role as major greenhouse gas sources, they, should also be considered as potential freshwater tipping systems. For deltas, a number of drivers have been identified, also including cases of positive feedbacks, that may profoundly and rapidly change the physical and ecological properties of large deltas [Törnqvist et al. 2020; van de Vijsel et al. 2024]. The use of remote sensing has also been used for detecting regime shifts and loss of resilience in coastal wetlands, covering a salinity gradient from fresh to marine ponds and wetlands [Martinez et al. 2024]. For many wetlands, change in hydrology, water saturation and thus redox conditions due to degradation will shift systems from sinks to sources of CO_2, and determine the ratio between methanogenesis and methanotrophy. Zou et al. [2024] estimated a strong increase in greenhouse gas release from wetlands due to drought, and redox state of wetlands can be seen as a tipping point in the context of redox processes. However, tipping points for these systems implies irreversible losses of ecosystems with their key properties (rather than continued functioning with altered dynamics), and it is not always clear that positive feedback loops and hysteresis are as prevalent in all of these systems as for lakes.

Bogs, and notably peat bogs are globally important long-term sinks of carbon acting as major conduits of greenhouse gases, depending on temperature and water saturation. Waddington et al. [2025] recently argued that peatland ecohydrological resilience is a nonlinear function of water storage dynamics, with implications for carbon storage and fluxes when critical tipping points have been exceeded. Peatland in the Congo may have a rainfall-linked threshold, and has been suggested as a potential tipping system if drying led to carbon release [Crezee et al., 2022; Garcin et al., 2022]. Ombrotrophic (only precipitation-fed) peat bogs are highly vulnerable to rainfall and catchment properties, and have bistable properties linked to water table both in terms of carbon storage, greenhouse gas emissions and community composition [Lamentowicz et al. 2019; Loisel & Bunsen 2020]. The extent to which this bistability can reach tipping points to self-sustaining change is not always clear, but especially when it comes to carbon balance and hydrology, there is evidence that regime shifts can be driven by within-system feedbacks [Milner et al. 2020].

Based on this new research, the GTP community maintains its assessment of lakes being a tipping system with high confidence for eutrophication-driven anoxia (i.e. internal loading of phosphorus from sediments under anoxia) and maintains medium confidence for browning-related anoxia, noting growing evidence for the latter [Saros et al., 2025]. Loss or gain of permafrost-water bodies has a confidence level closely linked to permafrost thaw. We also add river deltas and peat bogs as potential local tipping systems, with low confidence for the former, and medium confidence for the latter.

Coastal ecosystems (mangroves, tidal wetlands, seagrass meadows, & kelp forests)

What it is

Despite their globally small area, coastal ecosystems such as mangrove forests, tidal saltmarshes, seagrass meadows, and kelp forests are highly biodiverse, and provide critical ecosystem services to many coastal areas [Nordlund et al., 2016; Menéndez et al., 2020; Cooley et al., 2022; doAmaral-Camara et al., 2023, James et al., 2023]. They face widespread degradation, primarily from habitat loss but increasingly from climate change impacts including increased weather extremes, severe storms & flooding, sea level rise, moisture and heat stress, and shifting climatological niches [Saunders et al., 2014; Bergstrom et al., 2021; Dunic et al., 2021; Cooley et al., 2022; Duke et al., 2022; Hagger et al., 2022]. For tidal wetland systems, there is strong evidence for bistability, with alternate states including mangrove or saltmarsh-dominated. Increased and compound pressures, strongly influenced by annual rainfall, are recorded as triggering the loss of one habitat form at the expense of the other [Feller et al., 2017; Duke et al., 2019; Duke et al., 2021; Hesterberg et al., 2022]. With more extreme and repeated damaging a point has been reached where re-establishment is threatened in some areas, leading to ecosystem collapse [Bergstrom et al., 2021]. Evidence is more limited for seagrass meadows, but also suggests that feedbacks can drive irreversible regime shifts to algal or unvegetated states in temperate and subtropical regions [Maxwell et al., 2017; Duarte et al., 2018; Kendrick et al., 2019; Cooley et al., 2022; Bartenfelder et al., 2022; Marba et al., 2022; Temmink et al., 2022]. Kelp forests can experience feedback-driven regime shifts to a barren state due to trophic cascades resulting from sea urchin dominance or climate-change intensified marine heatwaves [Ling et al., 2015; Filbee-Dexter & Werberg, 2018; Filbee-Dexter et al., 2020].

What's new

Mangrove forests & tidal wetlands

In GTPR23, mangrove forests and tidal wetlands were assessed as regional-scale tipping systems with medium confidence, with thresholds estimated to be reached by 1.5-2°C and late century alongside potential pollution and sea level rise rate thresholds.

Since GTPR23, research on mangroves and tidal wetlands has further established the extent to which mangroves are threatened. While mangroves overall have seen a greening trend since 2001 [Zhang et al., 2024], regionally there have been some large dieback events, with recent research analysing the El Niño-linked 2015-16 dieback event in the Gulf of Carpentaria [Duke et al., 2017], extreme events driving dieback in Australia, the Sundarbans, and Brazil [Sippo et al., 2018], and hailstorm-induced dieback in Mozambique [Machava-António et al., 2024]. The capacity of mangrove forests to re-establish has been compromised by rapidly rising sea levels coupled with increased severe weather and El Niño Southern Oscillation (ENSO) events, including ENSO impacts on sea level [Duke et al., 2022; Chung et al., 2023; Zhang et al., 2025], category 3+ cyclones [Duke et al., 2024], drought-hurricane compound events [Amaral et al., 2023], and unprecedented severe flood events and bushfires [Glasby et al., 2023]. In the future, one recent study estimated that the combination of climate-change intensified tropical cyclones and sea level rise would put around half of global mangrove area at high to severe risk of loss, and in particular those providing key services to people [Hülsen et al., 2025]. Mangroves and tidal wetlands can keep up with relative sea level rise rates up to a threshold of 4-7 mm per year through accretion, but at 2°C of global warming nearly all mangroves would be exposed to 4 mm per year and one third to 7 mm per year, and all nearly tropical and subtropical coastlines would reach 7 mm per year at 3°C [Saintilan et al., 2020; 2022; 2023].

Mangrove habitat is constrained within a very narrow elevational range between mean sea level and the highest tide levels. For the habitat to survive the vegetative structural elements must relocate, a process that can only be achieved over at least a decade, as the time needed for seedling establishment and growth to mature trees is essentially fixed [MacLeod et al., 2023; Duke et al., 2024]. As such, this capability can be overwhelmed, and this appears to be being surpassed in some regions [Duke et al., 2022]. Furthermore, the indirect impacts of climate change are also negatively affecting key functional groups in mangrove ecosystems, further reducing mangrove resilience [Ferreira et al., 2024]. However, landward mangrove expansion is still occurring in some regions, with for example the 2015-16 dieback event in northern Australia happening within a longer-term expansion [Asbridge et al., 2019], while some mangroves demonstrate continued resilience to tropical cyclones [Asbridge et al., 2025].

Based on this research, the GTP community maintains its assessment of mangroves being a tipping system at regional scales with medium confidence, while adding localised mangrove tipping at high confidence. While observations of particular mangroves failing to recover from increasingly extreme events are accumulating, there is strong regional variation in mangrove vulnerability [Rogers et al., 2019], and uncertainty globally in where sediment supply and landward migration can compensate relative sea level rise [Schuerch et al., 2018].

Seagrass meadows

In GTPR23, seagrass meadows were assessed as regional-scale tipping systems with medium confidence, with thresholds estimated to be reached by 1.5°C and mid-century, alongside potential pollution and sea level rise rate thresholds. Since then, new research has shown that in a conceptual mechanistic model of seagrass ecosystems, passing mortality thresholds results in a tipping point from seagrass meadow to a bare state, exposing the sediment to erosion, and reversing the meadow from carbon sink to source [Dakos et al., 2025]. This mirrors previous research indicating that seagrass meadows feature feedbacks that can drive self-sustaining regime shifts [Maxwell et al., 2017]. Research in the Gulf of Mexico has also shown how even in relatively undisturbed meadows, sea level rise can drive rapid seagrass loss [Capistrant-Fossa & Dunton, 2024]. Together this supports the confidence assessment of GTPR23, with further empirical research needed to assess regional variations in tipping thresholds and likelihood.

Kelp forests

GTPR23 assessed kelp forests as a local-scale tipping system with high confidence, with a timescale of months to decades. This assessment is supported by recent research focused on the effects of climate change on kelp ecosystems along the southeastern coast of Australia, particularly highlighting the invasion of overgrazing sea urchins that are expanding poleward due to warming waters [Ling & Keane, 2024]. The population of sea urchins has significantly increased here over the past 15 years, leading to the rapid emergence of incipient barrens, areas where kelp has been overgrazed. This suggests that half of the kelp beds within the affected region could collapse by around 2030, posing serious ecological concerns. Further work has also shown the increasing role of marine heatwaves in driving physiological tipping points [Leathers et al., 2024], while a sea urchin outbreak since 2014 on the Californian coast has led to a shift to a patchy mosaic of forest and barrens due to spatial heterogeneity in environmental conditions [Smith et al., 2024].

Other coastal systems

There are several other coastal ecosystems - such as mussel beds, oyster reefs, and salt marshes (cordgrass) - for which evidence exists for potential tipping dynamics (including bistability and self-sustaining regime shifts [Temmink et al., 2023]), that would benefit from targeted assessment here in future.

Warm-water coral reefs

For more on tipping points in coral reefs, see the coral reef case study

What it is

Shallow coral reefs in tropical and subtropical waters (hereafter 'warm-water coral reefs') are highly complex ecosystems built around the symbiotic relationship of reef-building corals and photosynthetic algae [Wilkinson et al., 2004]. Increasingly though, global warming means warm-water coral reefs are experiencing 'coral bleaching' events, during which sustained marine heatwaves triggers corals to expel their symbiotic algae due to heat stress [Hughes et al., 2017; 2018a; 2018b; Houk et al., 2020]. While natural bleaching events do occur, after which most corals recover, the increasing frequency and intensity of marine heatwaves - which has recently estimated to have increased by circa five times in frequency and intensity in the tropical Atlantic since 1982 [Rodrigues et al., 2025], and the 2023-24 global marine heatwave triggering catastrophic bleaching in previously less affected southern Great Barrier Reef [Byrne et al., 2025] - is increasingly preventing recovery between heatwaves, triggering mortality.

The loss of hard coral structure can trigger a wider ecological regime shift to an algae-dominated state, creating a localised tipping dynamic by which hard coral recovery would be impeded even if global warming were to be halted or reversed [Bland et al., 2018; Darling et al., 2019; Sheppard et al., 2020; Perry et al., 2013; Vercelloni et al., 2020]. Globally widespread die-off is expected by 1.5-2°C [IPCC SR1.5 2018; Cooley et al., 2022; Dixon et al., 2022; Setter et al., 2022; McWhorter et al., 2021; Frieler et al., 2013], but regional-scale coral reef mortality is already being observed as localised tipping becomes regionally synchronous [Le Nohaïc et al,.2017; Amir, 2022; Muñiz-Castillo et al., 2019; Obura et al., 2022]. Additionally, coral reefs face many other anthropogenic pressures, including pollution from nutrient and sediment runoff, increased weather extremes, overfishing and invasive species and diseases, which can also contribute to localised tipping [Ban et al., 2013; Edmunds et al., 2014; Darling et al., 2019; Cramer et al., 2020].

What's new

In GTPR23, warm-water coral reefs were assessed as a tipping system at the local and regionally-clustered scales with high confidence, with thresholds region- and reef-dependent but a global warming level of ~1.2°C (1.0-1.5°C) estimated for globally widespread losses. Since then, there has been a multi-year coral bleaching event, which has not been declared closed at the time of final text editing. Observations confirmed that 2023-2025 experienced the fourth global coral bleaching event on record, and the second within the past decade (following events in 2014-17, 2010, and 1998) [NOAA, 2024]. In this event, coral bleaching affected every ocean basin, with 83.7 per cent of corals experiencing bleaching-level heat stress by April 2025 (the greatest extent recorded, compared with 65.7 per cent in 2014-17) [NOAA CRW, 2025]. In 2024 catastrophic bleaching occurred in the previously less affected southern Great Barrier Reef, with mortality also affecting genera that are considered resilient [Byrne et al., 2025], while Coral Sea heat extremes were the worst for 400 years, putting the Great Barrier Reef at risk of near-annual bleaching [Henley et al., 2024]. Final global mortality figures are not yet available while the event continues, but around 14 per cent of coral reef was lost in the 2009-2018 period spanning the previous two global bleaching events [Souter et al., 2020].

Following GTPR23, Pearce-Kelly et al. [2025] explored the potential for coral tipping dynamics in response to various different stressors and their interactions, concluding that a warming threshold of 1.2°C (1–1.5 °C) as well as the long-term impacts of atmospheric CO_2 beyond 350 ppm were appropriate, noting these thresholds have already been passed, and warning that a comprehensive assessment of stressors and interactions has not yet been conducted and would likely result in lower threshold estimates. Cornwall et al. [2024] assessed the potential for ocean acidification to trigger tipping dynamics, and concluded that while evidence for direct physiological-level tipping dynamics is lacking, indirect ecosystem-level tipping is likely beyond 500 ppm of CO_2, particularly due to the differential impacts of acidification on calcification and growth to the detriment of calcifying (e.g. corals, molluscs, foraminifera) and benefit of non-calcifying (e.g. diatoms, fish, non-calcareous seaweed) organisms. Conversely, future decline in coral reef calcification due to climate change could increase the overall ocean carbon sink by around 7 per cent [Kwiatkowski et al., 2025].

Questions have been raised as to whether current coral decline projections are potentially overestimated. In a systematic review of the methods used to make projections of coral responses to climate change, Klein et al. [2024] found that most used deterministic rather than probabilistic approaches, limiting the ability to assess uncertainty, and that methods showing higher impacts (generally simpler 'excess heat' threshold models, often linked to estimated thresholds in 'Degree Heating Weeks', DHW, for which globally consistent values are difficult to establish) were disproportionately cited. Lab-based results also suggest that a broad range of coral species in the Indo-Pacific show sufficient heritability to allow for adaptation to both warming (with some coral species' DHW thresholds potentially increasing to those expected at c. 1-1.7°C) and acidification levels (c. -0.2 pH units) broadly consistent with the Paris Agreement, but would be insufficient for higher emission scenarios [Jury & Toonen, 2024]. However, these analyses do not explicitly account for interacting non-climate co-drivers, which could reduce adaptive capacities at the ecosystem level in the field, nor of in-situ applicability of lab results like these, while recent modelling has found that coral range expansion is too slow to counter future declines [Vogt-Vincent et al., 2025].

Based on the latest information, the GTP community maintains its assessment of warm-water coral reefs having localised to regional tipping points with high confidence [medium agreement, robust evidence]. This is based on well-documented evidence for coral reefs being vulnerable to regime shifts to various alternative states, and observations of increasingly widespread mortality in responses to increasingly frequent and intense marine heatwaves. We also maintain the warming threshold estimate of ~1.2°C (1-1.5°C), noting that widespread mortality is already being observed at current warming levels of ~1.4°C [WMO, 2025]. While laboratory tests suggest some coral species might have the adaptive capacity to cope with Paris Agreement-compliant warming [Jury & Toonen, 2024] this is not validated in in situ contexts. We also note that CO_2 levels above 500 ppm could directly trigger ecosystem tipping points via acidification [Cornwall et al., 2024], while in the long run CO_2 levels remaining beyond 350 ppm also threaten corals through long-term commitment to climate change [Pearce-Kelly et al., 2025].

Marine (benthic & pelagic) ecosystems

What it is

Marine ecosystems - from shelf sea to deep ocean, and sea floor to water column - face substantial pressures from multiple anthropogenic drivers, which has the potential to trigger irreversible regime shifts [Heinze et al., 2021; Jouffray et al., 2020; Bindoff et al., 2019]. Overexploitation combined with climate change could cause some fisheries to collapse [Sguotti et al., 2019; Beaugrand et al., 2022]. Similarly, warming, habitat loss, and pollution could result in community-wide shifts in wider marine ecosystems in benthic as well as pelagic environments [Conversi et al., 2015; Beaugrand et al., 2019; Möllmann et al., 2021; Ban et al., 2022; Sguotti et al., 2022]. Warming is also expected to result in a reduction in the biological pump - the transport of organic carbon from surface to deep waters - as ocean layers become harder to mix, although barring the polar seasonal lipid pump this is currently expected to be a relatively linear process [Jonasdottir et al., 2015; Armstrong McKay et al., 2021; 2022]. Finally, low oxygen 'dead zones' are expanding as a result of warming and nutrient pollution, with excess algae growth leading to deoxygenation and further amplified by sediment phosphorus release feedback [Diaz & Rosenberg, 2008; Breitburg et al., 2018; Heinze et al., 2020]. However, while many marine regime shifts have been observed and evidence exists for these ongoing changes and the potential for them to reach local to regional-scale tipping points in some cases, confidence is currently mixed due to limited understanding of potential thresholds and hysteresis in these systems.

What's new

In GTPR23, fisheries were assessed as a local-scale tipping system for some larger fish species with low confidence, and high confidence specifically for cod, with a tipping timescale of around a decade. Marine community shifts were assessed as a low confidence local-scale tipping system. The overall biological pump was not considered a tipping system (medium confidence), but specifically the seasonal lipid pump in the Arctic could be a regional-scale tipping system (unclear, decades). Finally, ocean hypoxia was assessed as a low confidence tipping system at the local scale (and unclear at regional-scale), with timescale varying from months to centuries.

Since GTPR23, there have been several new publications relevant to these systems. Vasconcelos et al. [2024] investigated the impacts of climate change on small pelagic fish communities in the Madeira Archipelago over a 40-year span (1980-2019). The research highlights how global warming has led to a regime shift in the small pelagic community, accounting for 88.9 per cent of the observed fluctuations in fish landings and life-history traits of two key species, *Scomber colias* and *Trachurus picturatus*. Cecapolli et al [2025] were able to identify contrasting regime shift dynamics across the three substocks of Atlantic cod in the North Sea, namely that the Southern North sea populations are now in a depleted state following a regime shift, whereas the other two substocks are either recovering or have not experienced a regime shift. Lastly, an analysis of three global datasets of 667 fish populations has identified abrupt shifts in productivity in almost 20 per cent of them [Cano et al., 2025], confirming and expanding on similar findings a decade earlier [Vert-pre et al., 2013]. Although the documented regime shifts of these papers support the GTPR23 assessments for marine systems, they also highlight the need for further research on the confidence for the extent of tipping point responses in benthic and pelagic marine systems.

2.2.7 Potential tipping points in ocean & atmosphere circulations

The ocean and atmosphere's circulations consist of the major flowing, fluid portions of the Earth system that transport water, air, and heat around the planet, and drive daily weather patterns. This includes the major overturning circulations in the ocean, as well as monsoon systems, mid-latitude atmospheric dynamics like the 'jet stream', tropical circulation patterns, and interannual 'oscillations' like the El Niño Southern Oscillation. In this section, we describe each of these systems, their evidence for tipping dynamics, and relevant new research in turn.

Based on this, we have reviewed the status of several systems, concluding that the Atlantic meridional overturning circulation (AMOC), convection in the North Atlantic Subpolar Gyre (SPG) and the circulation in the Southern Ocean are tipping systems with medium confidence, the West African Monsoon as tipping system with low confidence, and all other considered systems as uncertain or no tipping system with varying confidence levels (Figure 2.2.4). With respect to GTPR23, the East Asian Summer Monsoon has been added as an uncertain potential tipping system.

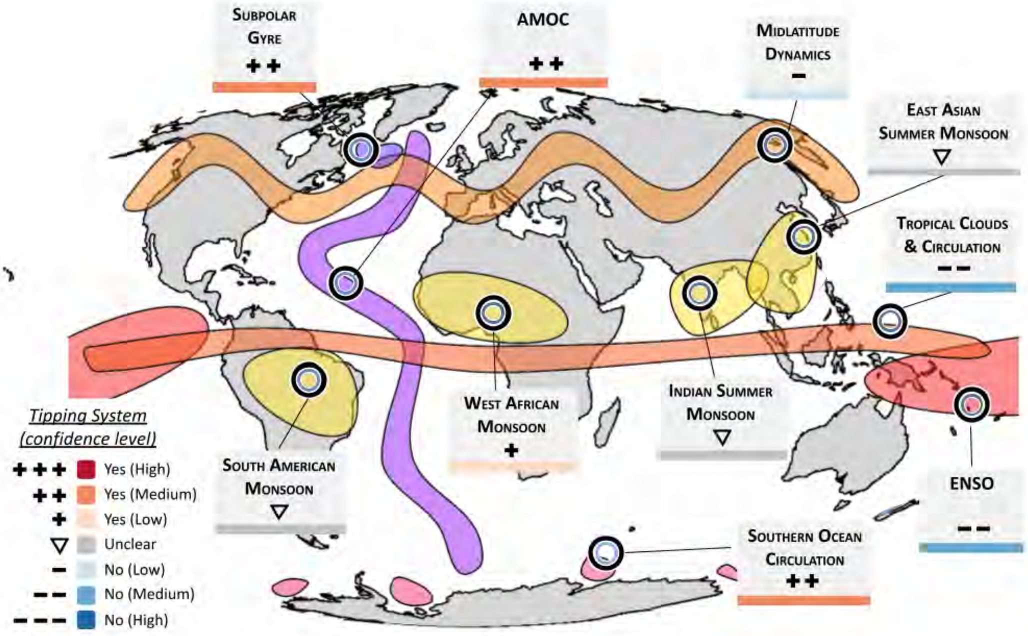

Figure 2.2.4: Potential tipping systems in ocean and atmosphere circulations considered in this chapter. The markers indicate which of the systems are in this report considered a tipping system (+++ high confidence, ++ medium confidence and + low confidence) and which are not (– – – high confidence, – – medium confidence and – low confidence), ∇ indicates systems for which a clear assessment is not possible based on the current level of understanding.

Atlantic meridional overturning circulation (AMOC) & deep convection in the North Atlantic subpolar gyre (SPG)

For more information on the AMOC and SPG tipping points, please see the AMOC case study

What it is

The Atlantic meridional overturning circulation (or 'AMOC') is the component of the Earth's ocean circulation driven by buoyancy loss at the surface in the north polar region. It describes the movement of warm surface water northwards in the Atlantic, followed by its cooling and densification, and subsequent sinking ('deep convection') in the seas around Greenland, helping to drive the circulation of the global ocean over the course of hundreds of years [Buckley & Marshall, 2015]. There are two main regions of deep convection: in the Greenland-Iceland-Norwegian Seas and in the Labrador-Irminger Seas. Climate change, however, can disrupt this circulation by warming and freshening water in the North Atlantic (via increased ice sheet runoff and precipitation), making it harder for it to get dense enough to sink [Arias et al., 2021, Liu et al., 2017]. Most model projections show a gradual decline initiated this century [Weijer et al., 2020; Bonan et al., 2025; Drijfhout et al., 2025], which is unprecedented at least in the last 6000 years [Gerber et al., 2025]. Palaeoclimate evidence suggests that the AMOC has more abruptly switched to weak or collapsed modes before that [Rahmstorf, 2024]. Some models also show weak AMOC states in future projections for extreme and intermediate forcing scenarios [Drijfhout et al., 2025], leading to strong North Atlantic cooling and substantial weather disruption in Europe and the Tropics [van Westen et al., 2025a; van Westen et al., 2025b], and the IPCC assessed in AR6 that many models are currently overly stable [Fox-Kemper et al., 2021; also Arumí-Planas et al. 2024, Dima et al., 2025 and Vanderborght et al., 2025]. The AMOC may already have weakened by ~15 per cent over the past 50 years [Caesar et al., 2018; Li and Liu, 2025; Michel et al., 2025], and some studies have detected statistical and physics-based 'early warning signals' (EWS) that could mean AMOC collapse begins within decades [Ditlevsen and Ditlevsen, 2023; van Westen et al., 2024], but whether tipping timings can be projected based on EWS has also been questioned [Latif et al., 2022; Ben-Yami et al., 2024, Terhaar et al., 2025].

Furthermore, the convection branch in the Labrador and Irminger Seas west and south of Greenland (forming part of the Subpolar Gyre, or 'SPG'), has been found to collapse separately to the rest of the AMOC (however being a precursor for AMOC weakening or tipping). This is suggested by palaeoclimate evidence during the transition into the Little Ice Age and by some models [Swingedouw et al., 2021; Arellano-Nava et al., 2022]. The risk of a SPG overturning collapse has been evaluated as 40 per cent in low- and medium-emission scenarios when selecting the CMIP6 models that best represent the observed oceanic stratification [Swingedouw et al., 2021].

What's new

In GTPR23, the AMOC was assessed as a tipping system with medium confidence, with a likely timescale of decades to centuries, and an uncertain path-dependent threshold [Loriani et al., 2023].

Since GTPR23, there have been several relevant publications. CMIP6-based projections show marked decline in the Arctic Beaufort Gyre, however the uncertainty of the gyre freshwater content trends is still large both in the models and observations [Athanase et al., 2025, Lin et al., 2023; Wang et al., 2025]. Increased freshwater export to the North Atlantic may weaken deep-water convection in key AMOC regions, but so far there is no published observational evidence of this happening. Various freshwater forcing experiments in models of varying complexity reaffirm evidence for potential AMOC collapse: In an ocean-only model, gradual freshwater forcing can trigger intermediate tipping events, leading to abrupt reorganisations of North Atlantic circulation [Lohmann et al., 2024; Lohmann & Lucarini, 2024] in multiple partially collapsed stable states. Four steady states are found using both surface freshwater forcing and changing CO_2 in CLIMBER-X [Willeit and Ganopolski, 2024]. A similar behaviour is observed in a high resolution model by Gou et al. [2024], where different convection areas reach tipping points independently. Van Westen et al. [2025] demonstrate that an AMOC collapse can still occur in a strongly eddying ocean-only model under freshwater hosing, indicating that ocean eddies do not prevent tipping. For the first time, an abrupt AMOC weakening has been simulated in Earth System Models without the need for externally imposed freshwater perturbations or an extreme emission scenario [NASA-GISS-E2-1-G in Romanou et al., 2023; MRI-ESM2-0 in Drijfhout et al., 2025]. This transition is triggered by stochastic variability in sea-ice transport and melting in the Irminger Sea. The stochastic bifurcation mechanism proposed in Romanou et al. (2023) has been explained by Boerner et al. [2025, in review] in terms of the collision between the `on' state of the circulation with the unstable circulation pattern associated with the `edge' state of the system.

Along similar lines, Gu et al [2024] find that nonlinear processes might amplify stochastic variability, leading to states with convection shutdown in the Labrador Sea. On the other hand, there have been recent studies that have identified contrasting mechanisms influencing the stability of the AMOC. Empirical modelling suggests that noise-induced tipping events are unlikely under present-day variability unless the AMOC is already close to a critical threshold [Chapman et al., 2024]. Furthermore, Baker et al. [2025] report a stabilising feedback present in CMIP6 models on a centennial time-scale, wherein sustained wind-driven upwelling in the Southern Ocean counteracts AMOC weakening—even under scenarios of extreme greenhouse gas forcing and freshwater input. However, these simulations are relatively short and may therefore correspond to a transient state. Also, the small AMOC strengths reported in several scenarios resemble the residual circulation persisting after collapse in extended simulations [Drijfhout et al., 2025; van Westen and Baatsen, 2025; van Westen et al., 2025]. The effect of the CO_2 ramping rate on the weak or collapse state of the AMOC has also been addressed [Hankel, 2024].

In terms of direct and proxy observation, Terhaar et al. [2025] suggest that the AMOC has not weakened in the last 60 years as derived from surface heat flux reconstructions serving as proxy indicator. Also, based on experiments with an idealised model, Zimmermann et al. [2025] advocate for cautious use of purely statistical early warning indicators since they may raise false alarms about approaching critical. On the other hand, a newly developed, physics-based observable early warning signal for AMOC tipping (representing the strength of the salt-advection feedback [Vanderborght et al., 2025]) suggests that the AMOC has destabilised over the past 40 years when applied to reanalysis and assimilation products [Van Westen et al., 2024], supported by deep-learning based reconstructions [Michel et al., 2025]. Furthermore, this indicator is biased positively in CMIP6 models, implying that these models may underestimate the risk of tipping [Arumì-Planas et al., 2024; van Westen and Dijkstra, 2024].

In GTPR23, like the AMOC, the SPG was assessed as a tipping system with medium confidence, however with a shorter timescale of years to decades, and a likely warming threshold range of 1.1 to 3.8°C [Loriani et al., 2023]. Since GTPR23, there have been new publications providing additional evidence for SPG as a tipping system. Proxy records from bivalve shells provide empirical evidence that the SPG crossed a tipping point during the transition into the Little Ice Age, with early warning signals appearing before the abrupt SPG weakening in the 14th century. The destabilisation was likely triggered by freshwater input from melting glaciers during the Medieval Warm Period, followed by an anomalous export of Arctic sea ice into the subpolar North Atlantic [Arellano-Nava et al., 2022]. A broader compilation of bivalve proxy records suggests that the subpolar North Atlantic experienced two periods of stability loss in recent times: one preceding the 1920s North Atlantic circulation regime shift, and the second in recent decades, indicating that the SPG region may be moving towards a tipping point [Arellano-Nava et al., 2025 in press].

Recent observations show that the SPG is undergoing strong freshening and reduced convection activity. The eastern SPG system has recently experienced its largest freshening of the past 120 years, primarily due to changes in ocean circulation [Holliday et al., 2020]. Two convective shutdowns in the Labrador Sea were observed in 2021 and 2023, with the latter being more intense. In 2023, convection shoaled to depths of less than 700 m, linked to extensive near-surface freshening driven by extreme Arctic sea-ice melt and enhanced by freshwater release from the Beaufort Gyre [Yashayaev, 2024]. This freshening also spread to the Irminger Sea, reaching depths of up to 1,500 m [Fried et al., 2024]. These events highlight the growing vulnerability of the SPG to crossing a tipping point.

Turning to model evidence, an analysis of abrupt shifts in CMIP6 models finds that 24 out of 57 models exhibit an abrupt SPG weakening, occurring within a global warming range of 0.3°C to 0.3°C, with a median critical temperature of approximately 1.5°C above pre-industrial levels [Terpstra et al., 2025]. In another study still under review, 'abrupt changes' and 'state transitions' are detected in the SPG in three models at 1.2-4.1°C and in 10 models at 1.2-5.6°C respectively (Table A2.2.1), and abrupt shifts are also detected in the AMOC in nine models between 1.1 and 3.9°C [Angevaare & Drijfhout, in review]. However, such abrupt shifts do not necessarily represent tipping points without confirming self-perpetuating dynamics. Menary et al. [2025] further classify different types of abrupt SPG changes, raising the need to define robust metrics and causal mechanisms [Falkena et al., in review] for the identification of abrupt events.

Based on this new research, the GTP community maintains its assessment of both AMOC and SPG being tipping systems with medium confidence, however acknowledging the increase in amount and agreement of evidence. For both, conceptual and low-complexity models suggest alternative stable states, further evidenced by palaeo records and process-based state-of-the-art climate models. Uncertainty persists around the magnitude of freshwater and thermal forcing that could initiate tipping, and the proximity to a potential tipping point.

Southern ocean overturning circulation

What it is

Ocean convection also occurs in the Southern Ocean (SO) around Antarctica, forming the second branch of the global ocean overturning circulation alongside the AMOC [Fox-Kemper et al., 2021; Heuzé et al., 2021]. Sea ice formation and strong offshore winds help produce dense salty water that sinks from the Antarctic shelves to the deep ocean, forming the Antarctic Bottom Water mass [Holland and Kwok., 2012; Abernathey et al., 2016]. The response of the SO overturning to global warming is less well studied than the AMOC, and is limited by the lack of Antarctic meltwater inclusion in models [Fox-Kemper et al., 2021; Heuzé et al., 2021; Purich & England 2023], but evidence has accumulated for an ongoing and continued decline [Lago & England, 2019; Liu et al., 2022; Gunn et al., 2023; Li et al, 2023; Zhou et al., 2023; Rosser et al., 2025 in review], and palaeo records suggest it has previously collapsed in response to meltwater pulses and could do so again [Skinner et al., 2010; Hayes et al., 2014; Gottschalk et al., 2016; Jaccard et al., 2016; Huang et al., 2020; Turney et al., 2020; Abram et al. 2025].

What's new

In GTPR23, Southern Ocean convection was assessed as a tipping system with medium confidence, with a likely timescale of decades and an unknown threshold [Loriani et al., 2023]. Additionally, there is the prospect of abrupt change in continental shelf circulation, leading to sudden rising ocean temperatures in contact with the Antarctic ice shelves [Li et al., 2023; Purich and England, 2023; Abram et al. 2025]. Cold ice shelf cavities in the Weddell and Ross Seas may be particularly vulnerable to warming under future climate scenarios, which could dramatically increase basal melting and contribute to sea level rise. This has been highlighted as another tipping mechanism [Hellmer et al., 2012; 2017; Siahaan et al., 2022; Naughten et al., 2023].

Over the past two years since GTPR23, a handful of studies have further investigated Southern Ocean convection and water mass change tipping systems, with much of the research effort concentrated on sea ice variability, but with some additional studies of indicators of ocean overturning change and future projections. On Southern Ocean convection and overturning, Gunn et al. [2025; in review] combines historical observations (1985-2024) and model projections of the upcoming decades (2041-2050) to assess changes in the abyssal Southern Ocean. They show that long-term freshening of the abyssal ocean has slowed and even reversed in some locations, as Dense Shelf Waters may no longer be reaching the abyssal ocean. Rosser et al., [2025; in review] demonstrate that the Southern Ocean overturning circulation collapses across nearly all CMIP6 models, even under strong mitigation scenarios.This disruption of the long-standing connection between the continental shelf and the abyss unfolds over decades and is driven by ice shelf melt and subsequent freshening. As the projected changes out to 2050 are already consistent with recent observations to 2025, there is evidence that model projections may underestimate the pace of change in the real system. This analysis indicates a potential tipping point in the ocean overturning connection between water over the shelf and water in the abyssal ocean.

Recent Antarctic sea-ice decline has raised further concerns that elements of the Antarctic ocean-ice system are changing more rapidly than first predicted. Since GTPR23 progress has been made in understanding its causes but uncertainty remains about the consequences for ocean convection. Using a reanalysis data product [Josey et al., 2024] and observations alongside models [Song et al 2024], recent work has shown that as sea ice cover is reduced, particularly in the Weddell, Bellingshausen, and Ross Seas, ocean heat loss to the atmosphere has doubled, also shifting the timing of peak heat loss. This intensification of winter heat loss leads to enhanced storm activity and also increased formation of dense surface water, but at locations far from the Antarctic shelf where dense shelf water is formed. A recent paper linking sea ice decline, heat loss, and a post-2015 surface salinity trend was misreported in the media as indicating an "SMOC reversal", but the implications for deep convection were not discussed in that study [Silvano et al., 2025]. Being remote from where Dense Shelf Water is formed, this effect is unlikely to contribute to changes in the abyssal overturning cell. However, advection and mixing of that off-shelf water mass may eventually impact the formation of dense water on the shelf, but the implications for Antarctic Bottom Water (AABW) formation remain unsubstantiated and is further complicated by decreased sea ice production, which diminishes AABW production. Nevertheless, this work highlights the potential interaction of different tipping points; namely, meltwater reduced AABW formation over the shelf [Li et al., 2023], contrasting changes in water masses off the shelf where sea ice loss has occurred [Josey et al., 2024]). In another study, using a coupled ocean-sea ice-ice shelf model forced by CMIP6 model-mean projected atmospheric conditions, Xie et al. [2025] find a decrease of more than half the rate of dense shelf water formation in the Ross Sea and a 300 metre thinning of AABW in the deep ocean by 2100. This reduction, a signature of an overturning slowdown in the region, is caused by the combined effects of meltwater-driven freshening with declining sea ice production.

A recent review of abrupt change around Antarctica finds evidence for emerging rapid, interacting and sometimes self-perpetuating changes in the Antarctic environment [Abram et al. 2025]. The study finds that Antarctic sea-ice coverage has reduced to levels far below its natural variability of recent centuries, with evidence that future changes could be more abrupt, nonlinear and potentially irreversible than Arctic sea-ice. The study also reviews evidence that the recent slowdown in the Southern Ocean overturning circulation is set to intensify this century, driven by increasing rates of ice melt around Antarctica. A slowdown of the SO overturning circulation in turn reduces the uptake of carbon by the oceans [Liu et al., 2022] and also causes further shelf water warming [Li et al., 2023; Purich and England, 2023], which are both expected to drive further ice shelf melt; an amplifying feedback that could lead to an eventual collapse in the SO overturning circulation.

Based on this new research, the GTP community maintains its assessment of the Southern Ocean overturning circulation as being a tipping system with medium confidence (high agreement, medium evidence), with evidence growing that the formation of Dense Shelf Water is declining and could reach a tipping point, but uncertainty remaining around the role of sea ice and interactions with ice shelf cavities.

Monsoons

What it is

Monsoons describe the large seasonal changes in the direction and strength of prevailing winds driven by seasonal insolation and local temperature differences between land and ocean, leading to heavy summer rainfall over land. Several subcontinental scale monsoon systems are recognised, including the Indian Summer Monsoon (ISM), the East Asian Summer Monsoon (EASM), the West African Monsoon (WAM), and the South American Monsoon (SAM). Today these are seen as being interconnected as part of one global monsoon system [Geen et al., 2020] strongly linked to the seasonal migration of the Intertropical Convergence Zone (ITCZ), where northern and southern hemisphere trade winds converge. The EASM however, extends to the subtropics, and is influenced by mid-latitude frontal systems and the jet stream as well as the Tibetan Plateau [Molnar et al., 2010; Son et al., 2019]. Rainfall projections are subject to high uncertainty in climate models, with projections of overall strengthening of the global monsoon precipitation in the future [Hsu et al., 2012]. Early tipping point studies suggested that the ISM could tip to a weak state as a result of aerosol emissions [Lenton et al., 2008; Levermann et al., 2009]. Similarly, the East Asian Summer Monsoon (EASM) was proposed to tip once an oceanic humidity threshold is crossed [Schewe et al., 2012]. However, more recent studies have cast doubt on both hypotheses [Boos and Storelvmo, 2016; Seshadri, 2017]. Palaeo records from the 'African Humid Period' also suggest that the WAM might interact with Sahel vegetation in a way that could tip this combined system to a wetter and greener state [Charney 1975]. In case of an AMOC collapse, the ITCZ and thus monsoons could strongly shift southwards, as evidenced by palaeo studies [Stager et al., 2011].

What's new

In GTPR23, the WAM was assessed as a tipping system with low confidence, and ISM and SAM assessed as unclear [Loriani et al., 2023]. Since GTPR23, there have been several new publications relevant to these tipping systems. By simulating a generic monsoon system on an aquaplanet, Katzenberger & Levermann [2025] introduce a new understanding that monsoons can be considered to undergo periodic tipping between stable 'on' and 'off' states. Changing climate conditions could thereby push the system towards a permanently altered state. Using an intermediate complexity model, Recchia and Lucarini [2023] showed the distinct response of the South and East Asian monsoon to anthropogenic forcings, emphasizing that aerosol forcing, rather than GHGs forcings, has the potential to strongly reduce the intensity of the monsoonal precipitation, with stronger impacts expected in the East Asian sector.

Furthermore, Loriani et al [2025] have included the EASM (not considered in GTPR23) in their assessment, classifying it as an uncertain tipping system based on limited understanding about potential tipping processes. Like the other monsoon systems, EASM could be subject to the effects of an AMOC collapse [Ben-Yami et al., 2024]. A recent analysis of CMIP6 models reveals a projected increase of extreme wet seasons frequencies, precipitation and interannual variability of the EASM [Katzenberger and Levermann, 2024], albeit not reporting major systematic shifts. Stronger evidence stems from proxy records indicating several abrupt and irreversible regime shifts since the Last Glacial Maximum [Lu et al, 2025], linked to abrupt shifts in the AMOC and Saharan vegetation. Finally, recent analysis of abrupt shifts in CMIP6 model results detected some abrupt shifts in the Indian Summer Monsoon (at ~0.8°C global warming) but not other monsoons [Terpstra et al., 2025] (Table A2.2.1), but they were not persistent, and do not necessarily represent tipping dynamics. Based on this research, the GTP community maintains its assessment of the West African Monsoon being a tipping system with low confidence. Similarly, for the South American and Indian Monsoon the assessment of an uncertain tipping system is maintained. The same classification of an uncertain tipping system is made for the now-added East Asian Summer monsoon.

El Niño southern oscillation ('ENSO')

What it is

The El Niño-Southern Oscillation (ENSO) is the dominant interannual mode of variability in Earth's climate [Timmermann et al., 2018]. Every three to five years, the 5-6°C difference between the warmer western tropical Pacific and cooler eastern tropical Pacific maintained by easterly trade winds becomes weakened, resulting in an 'El Niño' event (the warm phase of ENSO). Conversely, during a 'La Niña' event (the cold phase of ENSO) this temperature gradient intensifies. Both phases lead to substantial weather pattern shifts around the Earth, with substantial impacts on people and ecosystems [e.g. Holbrook et al., 2020; McPhaden et al,. 2020; Callahan & Mankin 2023]. Climate models differ, but despite overall stronger trade winds, colder eastern equatorial Pacific, and weaker ENSO events since the 1990s/2000s [Capotondi et al., 2015; Ma & Zhou, 2016; Fedorov et al., 2020; Seager et al., 2022; Wills et al., 2022; Heede and Fedorov 2023a], global warming is generally expected to result in an increase in ENSO variability, leading to more extreme El Niño and La Niña events [Cai et al., 2015; 2018; 2022; Heede and Fedorov 2023b; Wang et al., 2023]. Palaeorecords suggest that the amplitude and frequency of ENSO events has gradually increased during the Holocene [Freund et al., 2018; Grothe et al., 2020; Lawman et al., 2022] and that a 'permanent El Niño-like state' may have existed in the Pliocene 3 million years ago [Wara et al,. 2005; Fedorov et al., 2006, 2013, 2015; Tierney et al., 2019], leading to the suggestion that ENSO might feature a tipping point towards a permanent or extreme state [Lenton et al. 2008]. However, the evidence so far does not support a threshold beyond which a self-sustaining regime shift to such a state occurs [Loriani et al., 2023].

What's new

In GTPR23, ENSO was assessed as not a tipping system with medium confidence [Loriani et al., 2023]. Since GTPR23, there have been several new publications relevant to this tipping system. In particular, the 2023-24 strong El Niño event has helped to drive record-breaking global temperatures across land and sea [Jiang et al., 2025], with substantial implications for ecosystems. While a rare extreme, models indicate that such large jumps in temperature are not unexpected during strong El Niño event when combined with the current global warming trend [Terhaar et al., 2025], and may have been made more likely by the prolonged La Niña event (itself potentially linked to global warming [Wang et al., 2023]) preceding it [Raghuraman et al., 2024]. These results suggest a combination of underlying global warming, a strong El Niño, and the particular timing and pattern of this and the preceding La Niña could largely drive the observed the recent jump in global temperatures, rather than being a regime shift in global warming or a change in ENSO dynamics, which will be confirmed if temperatures revert to the long-term trend [Terhaar et al., 2025].

For the future, recent research continues to support an increase in ENSO variability and extreme El Niño event frequency, with an analysis of ENSO dynamics during past glacial changes indicating that cooler conditions led to less ENSO variability and fewer extreme El Niño events [Thirumalai & DiNezio et al., 2024; corroborating e.g. Brown et al., 2020], and therefore the inverse can be expected with future warming. Bayr et al. [2024] recently argued that ENSO could be considered a tipping system on the basis of stronger El Niño events potentially triggering other tipping systems via higher global temperatures, but this warming is temporary, and no self-sustaining state shift is observed or projected within ENSO's own dynamics, which is key to be considered a tipping system (rather than impacts on other systems). As such, the GTP community maintains its assessment of ENSO not being a tipping system with medium confidence.

Midlatitude dynamics

What it is

A key aspect of the mid-latitude atmospheric circulation is the 'jet stream', a band of strong westerly winds with largest velocities at an altitude of 7-12 km which separates cold polar air masses from temperate lower-latitude air masses. The jet features large 'meanders' (linked to planetary, or Rossby, waves) which normally move and dissipate but can become quasi-stationary, leading to high-impact climate extremes. A local example of such persistent weather features are atmospheric 'blocking' events, which are closely associated with several extremes including wintertime cold spells and summertime heatwaves [Kautz et al., 2022] and conditions favouring wildfires [Luo et al., 2025]. Global warming has likely already led to a poleward shift of the mid-latitude jet [Woolings et al., 2023] and a similar trend is expected to continue in the future [Oudar et al., 2020]. At the same time, amplified Arctic warming is leading to a reduced meridional temperature gradient between the high and low latitudes. This has been posited to slow down the mid-latitude circulation and the jet stream, increasing the latter's waviness. An enhanced waviness, in turn, would favour more persistent midlatitude weather and the connected extreme events [Kornhuber & Tamarin-Brodsky, 2021; Coumou et al., 2018]. Observations indeed evidence a slowdown of the boreal mid-latitude summer storm tracks over the last several decades likely associated with anthropogenic emissions [Chemke & Coumou 2024]. However, the strength of the poleward shift and the degree to which jet waviness and midlatitude weather persistence have increased and/or can be linked to Arctic warming has been disputed [Blackport & Screen, 2020; Riboldi et al., 2020]. Potential tipping points in jet stream dynamics have been suggested [Drijfhout et al., 2013; Steffen et al., 2018], but little evidence exists for a warming threshold beyond which such behaviour might become self-sustained.

What's new

In GTPR23, mid-latitude atmospheric dynamics was assessed as not a tipping system with low confidence [Loriani et al., 2023]. Since GTPR23, there have been new investigations relevant to this tipping system, with a focus on how a future, warmer Arctic will affect mid-latitude waviness. Key results concern future changes in the meridional temperature gradient, the role of Arctic sea-ice and the relationship between jet speed and jet waviness. Arnheim et al. [2025] analysed large ensembles of climate model simulations and concluded that climate change leads to a less wavy jet. They further found that the more recent set of simulations produced a weaker Arctic Amplification, displaying an enhanced reduction of waviness and blocking compared to an earlier model version. The effects of Arctic amplification on the jet stream and midlatitude circulation may be partially countered by Arctic sea-ice loss, which contributes to a slow-down of the wintertime North Atlantic jet stream [Jiang et al., 2025]. Finally, past arguments for a future, wavier mid-latitude circulation have mostly assumed that slower jets are more wavy and persistent, but the link between jet speed, waviness and persistence has recently been challenged by Baatelan et al. [2024] and Banderier et al., [2025], who found that weakened mid-latitudinal jet strength does not equate to increased waviness and that waviness can increase in the absence of clear persistence changes. In contrast, recent work has found evidence for increasing frequency of planetary wave resonance events in historical data [Li et al., 2025] and in future model projections [Guimarães et al., 2024], but with no indication of tipping dynamics. The latest research strengthens the evidence for the midlatitude atmospheric dynamics not being a tipping system. However, due to a low level of agreement in the literature the GTP community still assesses the midlatitude atmospheric dynamics as not being a tipping system with low confidence (medium evidence, low agreement).

Tropical clouds, circulation, & climate sensitivity

What it is

Clouds - and in particular tropical clouds - play an important role in the Earth's climate system by modulating how much incoming sunlight is reflected and how much heat from the surface is trapped [Forster et al., 2021]. In general, high thin clouds tend to let more light through but trap more heat, therefore having a net warming effect, while lower thicker clouds tend to reflect more light but let more heat through, having a net cooling effect. Global warming is affecting cloud dynamics though, with implications for the Earth's climate sensitivity depending on which type of cloud is more favoured. Cloud projections remain highly uncertain though due to poor process representation, but unexpected feedbacks remain possible [Caballero & Huber, 2013; Bellomo et al,. 2014; Bloch-Johnson et al.. 2015; Mauritsen & Stevens 2015; Myers et al. 2018]. One such feedback is the loss of subtropical stratocumulus cloud decks, which in one model featured a tipping point in CO_2 concentration beyond which abrupt loss and a global warming feedback of 8°C occurred [Schneider et al,. 2019]. Another possibility is an unexpected reorganisation in tropical circulation to a 'super-MJO' (Madden-Julian Oscillation) or superrotation state beyond some level of warming [Caballero & Carlson 2016; Seeley & Wordsworth 2021; Tziperman & Farrell, 2009; Caballero & Huber 2010]. However, cloud processes remain a key source of uncertainty in climate models, and little evidence is currently available to support these possibilities [Sherwood et al., 2020].

What's new

In GTPR23, tropical clouds and circulation was assessed as not a tipping system with medium confidence [Loriani et al., 2023]. Scientific literature published since then has not directly addressed tipping points in this system, such that the GTP community maintains its assessment of tropical clouds, circulation and climate sensitivity not being a tipping system with medium confidence.

2.2.8 Interactions between tipping systems

What it is

The tipping systems identified in the previous subsections are generally not isolated but interact across scales in space and time [Wunderling/von der Heydt et al., 2024]. Here, we define a tipping interaction as any linkage between two tipping systems that is destabilising, stabilising or where competing effects are at play. An example of a well established destabilising interaction is the linkage between GrIS and the AMOC, where meltwater from the GrIS destabilises the AMOC [see e.g. Weijer et al., 2019]. An example for a stabilising interaction is the interaction vice versa where a weakening (or tipping) AMOC leads to cooler temperatures around Greenland, which may be strong enough to slow down (or even stop) further GrIS melt [e.g. van Westen et al., 2024; Jackson et al., 2015].

What's new

Summary

We summarise the assessments of the GTP community in Figure 2.2.5. Based on these new and former assessments, the GTP community maintains the following three important conclusions on interactions and cascading transitions between tipping systems.

1. Tipping systems in the climate system are closely interacting, meaning a substantial change in one will have consequences for connected tipping systems.
2. We are quickly approaching global warming thresholds where tipping system interactions become relevant, because multiple individual thresholds are being crossed. These are at levels of 1.5-2.0°C of global warming.
3. While the pure pressure from global warming in its speed and also its magnitude dominates over the role of interactions (which need time to unfold) on tipping risks, the interactions between climate tipping systems seem to further destabilise the Earth system in addition to climate change effects on individual tipping systems.

The following three novel top-level conclusions including the newly assessed interactions are:

1. The majority of interactions between climate tipping systems are destabilising. However, new evidence shows that the interaction from AMOC weakening to Amazon rainforest may be stabilising as well as the interaction from West Antarctic Ice Sheet melt to AMOC stability. 15 interactions are assessed as destabilising, four as stabilising and one as unclear/competing effects (see Figure 2.2.5; not counting the interactions of very limited evidence, coloured in grey in Figure 2.2.5).
2. Although of unknown and/or limited strength, permafrost loss and Arctic Sea Ice decline may form a vicious cycle where less permafrost could lead to Arctic Sea Ice retreat and Arctic Sea Ice retreat may lead to enhanced inland permafrost degradation.
3. The AMOC has already been identified as an important mediator of interactions in GTPR23 but with the new evidence reported here, it can truly be stated that the AMOC is the global mediator of tipping point interactions. The AMOC alone features in 45 per cent (9 out of 20 interactions, not counting interactions of very limited evidence) of all assessed tipping point interactions (see Figure 2.2.5).

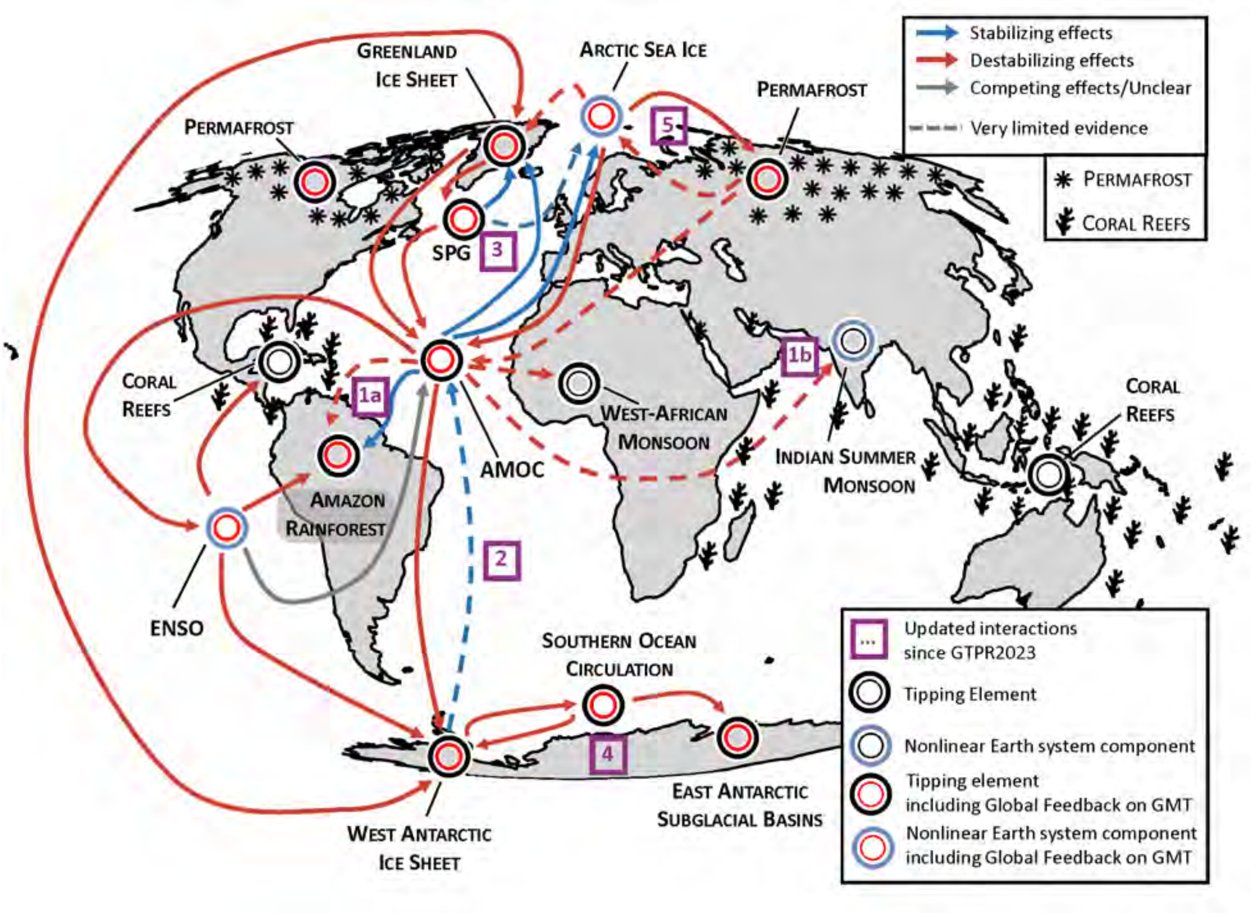

Figure 2.2.5: Update of tipping system interactions based on new evidence presented in this report. The updated interactions are denoted by purple squares and are: (1) Interactions from AMOC to global monsoon systems with the two updates (1a) AMOC→Amazon rainforest: AMOC weakening leads to increased rainfall in the southern part of the forest (blue arrow) but also to decreasing rainfall in the northern part (red dashed arrow); (1b) AMOC interaction with the Indian summer monsoon (red dashed arrow). (2) Interaction between AMOC and the West Antarctic Ice Sheet (WAIS) with limited evidence pointing towards a potential stabilising interaction from WAIS→AMOC (blue dashed arrow). (3) North Atlantic Subpolar Gyre (SPG) and its interactions with AMOC and the Greenland Ice Sheet. (4) Interaction from the Southern Ocean Circulation to the shelf regions of the Antarctic Ice Sheet and WAIS. (5) Interactions between Permafrost and Arctic Sea Ice (red arrows, now with higher evidence levels). The original figure has been updated from GTPR23 [Lenton et al., 2023] and Wunderling/von der Heydt et al. [2024].

Based on new evidence from recently published works presented here, we reassessed and for the first time evaluated the following five new interactions between climate tipping systems. Here, we focus on interactions where new evidence arose since the last Global Tipping Points Report 2023 [Lenton et al., 2023; Wunderling/von der Heydt et al., 2024].

1. Interaction from AMOC to global monsoon systems

a. AMOC → Amazon rainforest (via the South American monsoon)

The interaction between the AMOC and Amazon rainforest is complex. In the last report, it was assessed that a weakening AMOC would lead to a southward shift in the Intertropical Convergence Zone (ITCZ) that in turn could lead to increased precipitation in the southern Amazon region while intensifying drying trends in the north [e.g. Bellomo et al., 2023; Orihuela-Pinto et al., 2022]. These divergent patterns could stabilise forests in some regions while pushing others towards a degraded or savanna-like state. Indeed, most recent Earth system model and observational-data studies find that a weakening AMOC leads to an increase in rainfall in southern and eastern parts of the Amazon rainforest counteracting the global warming-related decrease in rainfall over these areas [Nian et al., 2023; Ben-Yami et al., 2024; Högner et al., 2025].

Another study confirms a vulnerability increase in northern Amazon forests due to a decrease in rainfall and increase in seasonality, based on palaeoclimatic pollen and microcharcoal data [Akabane et al., 2024]. As suspected, these studies suggest a regionally different response of the Amazon rainforest to a weakening AMOC that is now implemented in our updated tipping system interaction map (see Figure 2.2.5). In addition, seasonal precipitation changes after AMOC collapse may exert additional pressure on the rainforest [van Westen et al., 2024].

Importantly, the overall vulnerability of the Amazon rainforest is high: While the increased vulnerability in northern Amazon region combined with deforestation, wildfires and land-use change may have the capacity to trigger a systemic tipping point in the northern Amazon rainforest [Akabane et al., 2024], the effect in the southern Amazon region is different. Here, the effect of increased rainfall due to AMOC weakening is likely not of the same magnitude as drying effects from global warming [Nian et al., 2023; Högner et al., 2025] and strong deforestation, wildfires and land-use change [Albert et al., 2023; Lapola et al., 2023]. This means that global warming impacts likely decrease rainfall over the southern Amazon rainforest more than a weaker AMOC would lead to its increase.

b. AMOC interactions with the West African and Indian monsoon systems

CMIP6 models that feature a bistable AMOC show highly consistent changes in tropical monsoon systems following an AMOC collapse. The West African Monsoon and the Indian Summer Monsoon experience shorter wet seasons and longer dry seasons with mean annual precipitation decreases of 29 and 19 per cent, respectively [Ben-Yami et al., 2024]. Therefore, we classify the interaction from AMOC to both monsoon systems as destabilising with yet limited evidence (see Figure 2.2.5).

2.Interaction between AMOC and WAIS

Several recent studies report AMOC changes in response to meltwater from the Antarctic Ice Sheet and in particular from WAIS as the most vulnerable part of the Antarctic Ice Sheet. Two studies highlight the importance of both oceanic and atmospheric processes for AMOC stability [Shin et al., 2024; An et al., 2024]. On short timescales, AMOC weakening in response to meltwater input from WAIS is observed, driven by Rossby wave teleconnections and ITCZ-shift-induced precipitation changes [Shin et al., 2024; An et al., 2024], while ocean-driven strengthening may emerge on longer timescales [Shin et al., 2024]. However, under realistic meltwater input from both the Greenland and (West) Antarctic ice sheets, the additional meltwater from the Antarctic Ice Sheet can mitigate an AMOC slowdown, or even prevent its tipping [Sinet et al., 2025; Knight & Condron, 2025]. These studies use Earth system models of intermediate and full complexity (CLIMBER-X and CESM). An earlier expert elicitation assessed competing physical effects from WAIS melt to AMOC stability, including a physical mechanism for a stabilising interaction from WAIS disintegration to AMOC stabilisation [Kriegler et al., 2009]. In summary, we therefore decide to label the linkage between the West Antarctic Ice Sheet to the AMOC as stabilising (with very limited evidence for now).

3.North Atlantic subpolar gyre and its interactions

The North Atlantic subpolar gyre (SPG) has been classified a separate potential tipping system in the previous report [Loriani et al., 2023; [Loriani et al., 2023; Armstrong McKay et al., 2022]; Armstrong McKay et al., 2022], and the SPG convection has a proposed mechanism for bistability [Born and Stocker, 2014]. However, at the time of GTPR23, SPG collapse was not included in the chapter on tipping system interactions because not many studies had considered SPG connections to other tipping systems at that point in time. However, we reconsider this former assessment due to our following new lines of evidence of tipping point interactions: (1) Tipping of the SPG is associated with persistent ceasing of convection in the Labrador Sea and Irminger Sea ocean deep convection areas, with similar but smaller amplitude and more regional consequences for North Atlantic temperatures as an AMOC shutdown [Swingedouw et al., 2021]. Therefore, a shutdown of convection in the SPG cools the North Atlantic region and we expect a stabilising link from the SPG to the Greenland Ice Sheet and Arctic sea ice [Li and Born, 2019]. Importantly, changes in precipitation patterns can also contribute to the freshwater budget in the SPG region, and can therefore play a role in destabilising the SPG. (2) As the SPG convection regions significantly contribute to deepwater formation for the AMOC, a shutdown of convection in these areas could lead to an initial weakening of the AMOC [Neff et al., 2023; Rahmstorf, 1995] (i.e. destabilising link from SPG to AMOC). (3) Finally the tipping in the SPG is triggered by freshwater input that could be meltwater from a disintegrating Greenland Ice Sheet [Born & Stocker, 2014].

4.Interaction from the Southern ocean circulation to the shelf regions of the Antarctic Ice Sheet

Recent work has provided growing evidence for a potential transition from cold to warm ocean states in the Filchner-Ronne and Ross Ice Shelf cavities due to changes in the Southern Ocean Circulation. Using an ocean circulation model, Hill et al. [2024] found that temperatures in these cold ice shelf cavities could increase by 2 to 4°C, leading to significant increases in sub-shelf melt rates, building on earlier efforts suggesting that warming has been locked-in in certain regions of Antarctica [Naughten et al., 2023]. However, Hoffman et al. [2024] caution that such transitions may not occur uniformly. Their coupled Earth system model, which simulates ocean, land, sea ice, atmosphere, and ice sheet interactions, explores freshwater triggers for tipping points that could rapidly shift ice shelf cavities from cold to warm states, accelerating basal melt rates within a few decades. In addition, remote connections between melt fluxes at different ice shelves could lead to cascading effects, further destabilising ice shelves downstream. Another key driver expected for Antarctic ice shelf mass loss in coming decades are the ongoing positive trends and high variability in the Southern Annual Mode, which causes increased upwelling and subsurface warming and salinification close to ice shelves (and vice versa for negative phases), driving basal mass loss of 40 Gt/yr (around 40 per cent of the average loss) at one standard deviation of this climate mode of variability [Verfaillie et al., 2022; Osotoka et al., 2022].

5. Interactions with permafrost

a. Interactions between permafrost and Arctic Sea Ice

There is new evidence that suggests a possible destabilising linkage from declining permafrost to a decrease in Arctic Sea Ice [Nitzbon et al., 2024]. The reason is that inland permafrost degradation could increase the land-to-ocean heat transport via rivers [Wang et al., 2021] which in turn has a destabilising effect on Arctic sea ice [Park et al., 2020]. We add this interaction as an interaction with limited strength and very limited evidence (see Figure 2.2.5). Vice versa, additional evidence on top of GTPR23 [Lenton et al., 2023] shows that Arctic (winter) sea ice retreat leads to enhanced inland permafrost degradation, based on palaeoclimate [Vaks et al., 2020], and climate model studies [Lawrence et al., 2008]. Therefore, this link remains destabilising but its evidence level has increased.

b. AMOC → Permafrost

Additionally, limited new evidence suggests that an AMOC slowdown would lead to less northward heat transport, which would have a stabilising effect on land permafrost, particularly over northern Europe and Western Siberia [Park et al., 2025]. At the same time, an AMOC weakening or collapse would lead to rising sea levels across the northern Atlantic region and may weaken parts of the low-lying permafrost regions through flooding [Schwinger et al., 2023]. Vice versa, an AMOC recovery after a shutdown would increase the heat transport and destabilise high-latitude permafrost. Since an estimate of the strength of these competing effects is not assessed yet and may be limited, we do not add an additional link(s) from AMOC to Permafrost.

2.2.9 Appendix

Table A2.2.1: Global warming level (GWL) at which abrupt shifts have been detected in potential tipping systems in CMIP5/6 simulations, with previous tipping threshold estimates for comparison in part informed by CMIP5 abrupt shifts) and the number of models the shift features in [brackets].

System	Abrupt shift GWL thresholds			Tipping point GWL thresholds
	CMIP6 (1ptCO2): Terpstra et al. [2025]*	CMIP6 (SSPs): Angevaare & Drijfhout [in review]**	CMIP5: Drijfhout et al. [2015]***	A. McKay et al. [2022]****
North Atlantic subpolar gyre	0.5-2.9 [24/57]	1.2-4.1 (from abrupt change') [3]***** 1.2-5.6 ('state transitions') [10]	1.4-3.8 [5]	1.1-3.8 (based on Sgubin et al. [2017])
AMOC	-	1.1-3.9 [9]	1.4-1.9 [1]	1.4-8.0
Localised land permafrost	1.0-3.8 [8/46] (total soil moisture content) 1.2-3.3 [19/36] (soil frozen water content)	-	5.6 [1]	1.0-2.4
Amazon rainforest	- [-/7] (inconsistent small shifts 0.9-5.0)	-	2.5-6.2 [2]	2.0-6.0
Boreal forests	- [-/7] (inconsistent shifts 0.8-4.9)	-	7.2 [2]	1.4-7.2
Ind. s. monsoon	0.8 [1/57]	-	-	-
Antarctic sea ice	0.5-5.3 [11/53]	2.0-3.2 [4]	1.4-2.9 [3]	1.4-2.9
Arctic summer sea ice	1.0-4.6 [15/53]	1.3-2.3 ('abrupt changes') [7]	-	1.3-2.9
Arctic winter sea ice	3.3-5.4 [14/53]	2.5-5.1 ('state transitions') [22]	4.5-8.2 [5]	4.5-8.7
Barents sea ice	0.5-3.0	(included in ASSI)	1.5-1.7 [2]	1.5-1.7

*90 per cent interpercentile range, rounded to one decimal place

**68 per cent range; cannot be directly compared to Terpstra et al. [2025], as latter uses 1ptCO2 scenario rather than SSPs and uses different variables and abrupt shift detection methodology

***From Drijfhout et al. [2015] Table S2

****Literature-based synthesis range

*****Not directly equivalent to events of Drijfhout et al. [2015]