1.4 Tipping points in ocean and atmosphere circulations

Authors: Sina Loriani, Yevgeny Aksenov, Henk Dijkstra, Matt England, Alexey Fedorov, Gabriele Messori, Francesco Pausata, JB Sallée, Bablu Sinha, Steven Sherwood, Thejna Tharammal, David I. Armstrong McKay, Govindasamy Bala, Andreas Born, Sybren Drijfhout, Laura Jackson, Kai Kornhuber, Cristiano M. Chiessi, Stefanie Rynders, Didier Swingedouw

Summary

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

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

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

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

Key messages

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

Recommendations

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

1.4.1 Introduction

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

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

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

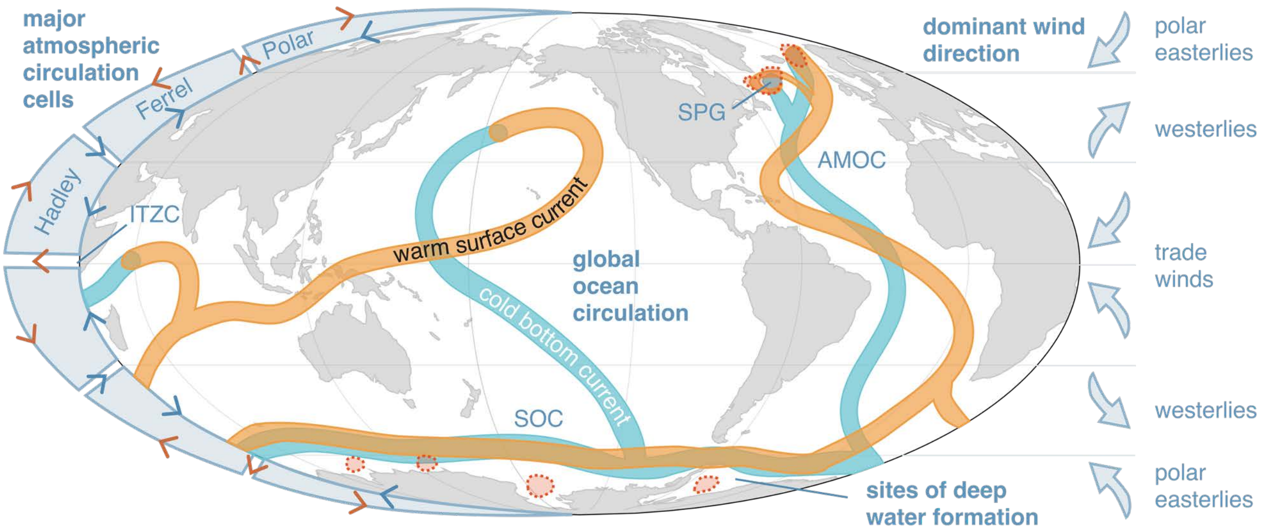

Figure 1.4.1: Atmospheric circulation cells, dominant wind directions, key ocean basins, surface currents and deep water formation sites. AMOC: Atlantic Meridional Overturning Circulation; SPG: Subpolar Gyre; SOC: Southern Ocean Circulation; ITCZ: Intertropical Convergence Zone.

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

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

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

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

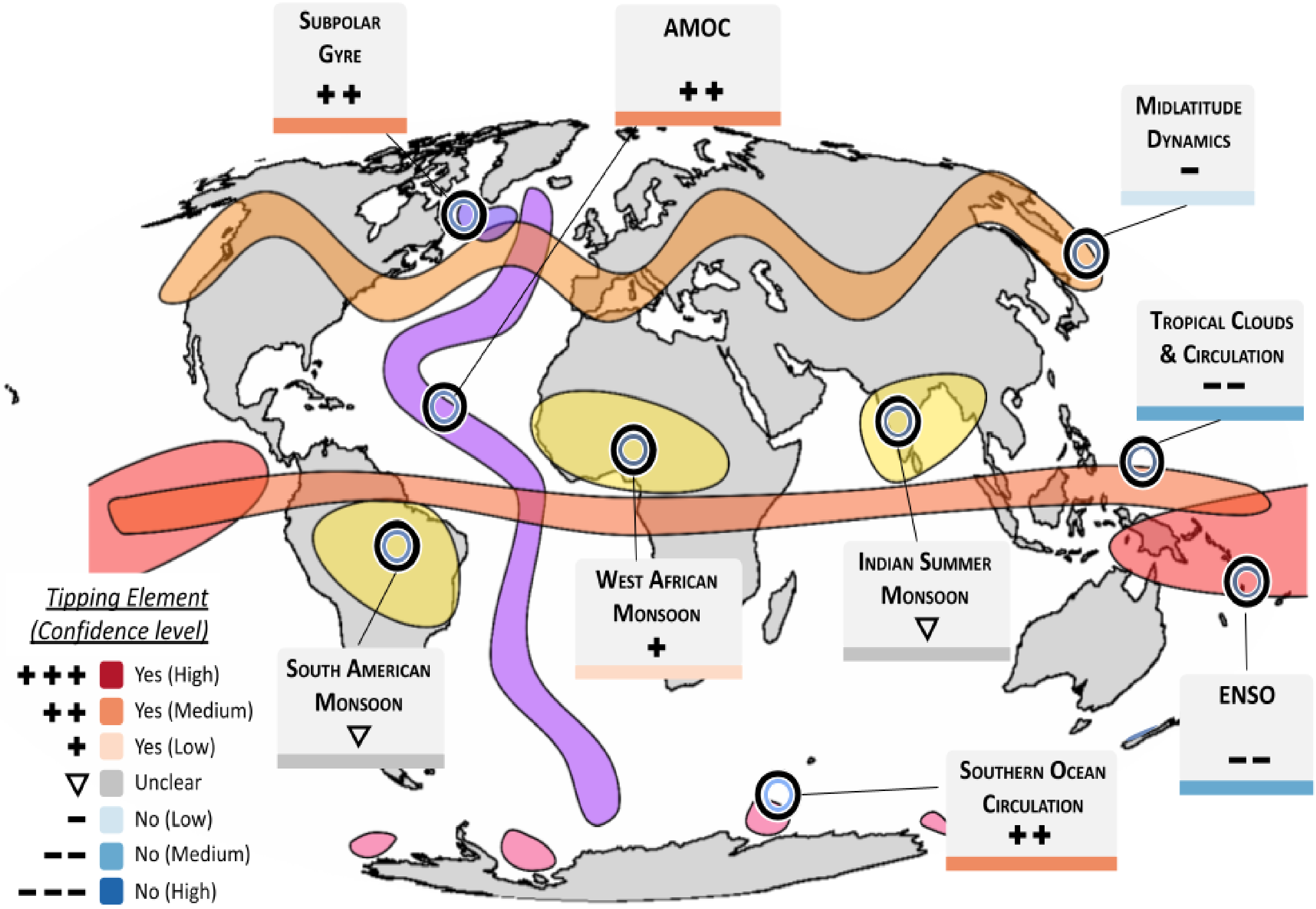

Figure 1.4.2: Potential tipping systems in ocean and atmosphere circulations considered in this chapter.

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

Table 1.4.1: Summary of evidence for tipping dynamics, key drivers, and biophysical impacts in each system considered in this chapter

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

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

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

System	Key drivers	Key biophysical impacts *(see S2 for societal impacts)*	Key feedbacks	Abrupt / large rate change?	Critical threshold(s)?		Tipping system?
Southern Ocean circulation Antarctic Overturning Collapse / Rapid continental shelf warming	• **DC: ocean warming (↗)** • **CA: Antarctic ice sheet meltwater increase (↗)** • CA: wind trends (↗) • CA: Sea ice formation (↗) • DC: precipitation increase (↗)	• Modification of Earth's global energy balance, timing of reaching 2°C global warming • Reduced efficiency of global carbon sink • Change in global heat storage • Reduced support for primary production in world's oceans • Drying of Southern Hemisphere • Wetting of Northern Hemisphere • Modification of regional albedo, shelf water temperatures • Potential feedback to further ice shelf melt	• Density-stratification (↗) • Meltwater-warming (↗)	++ (AABW formation & abyssal overturning shutdown within decades)	Salinity change/ freshwater	++ (cavity warming reversion would need 20th-century	**++**

Atmosphere: Monsoons

System	Key drivers	Key biophysical impacts	Key feedbacks	Abrupt / large rate change?	Critical threshold(s)?		Tipping system?
Indian summer monsoon (ISM) Collapse / Shift to low-precipitation state	• NC: increased summer insolation (↘) • DC: increased water vapour in atmosphere (↘) • CA: Indian Ocean Dipole events (?) • CA: ENSO change (?) • CA: North Atlantic cold SST (↗) • DC: aerosol loading (↗) • CA: Indian Ocean warming (↗) • CA: low cloud reduction (↘)	• **Massive change in precipitation** • Change in tropical and subtropical climates • Biodiversity loss and ecosystem degradation	• Moisture-advection (↘)	Decades to centuries	Regional AOD level over Indian subcontinent (>0.25) Interhemispheric AOD difference (>0.15) AMOC slowdown (unknown threshold)	Uncertain; likely decades to centuries	**unknown**
West African monsoon (WAM) Collapse or abrupt strengthening	• DC: increased water vapour in atmosphere (↗) • NC: increased summer insolation (↘) • NC: land-cover change (↗) • CA: desertification (↗) • CA: AMOC slowdown (↗) • CA: regional SST variations (?) • CA: High latitude cooling (↗) • CA/NC: regional soil moisture variation (?) • CA/NC: regional vegetation variation (?) • NC: dust emissions (?)	• **Massive change in precipitation** • Change in tropical and subtropical climates • Biodiversity loss and ecosystem degradation	• Vegetation-albedo (↗)	Decades to centuries	Insolation changes in the Northern Hemisphere summers and surface albedo changes (unknown threshold) Interhemispheric asymmetry in AOD (>0.15) AMOC slowdown (unknown threshold)	Decades to centuries	**+**

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

1.4.2.1 Atlantic circulation

Atlantic Meridional Overturning Circulation (AMOC)

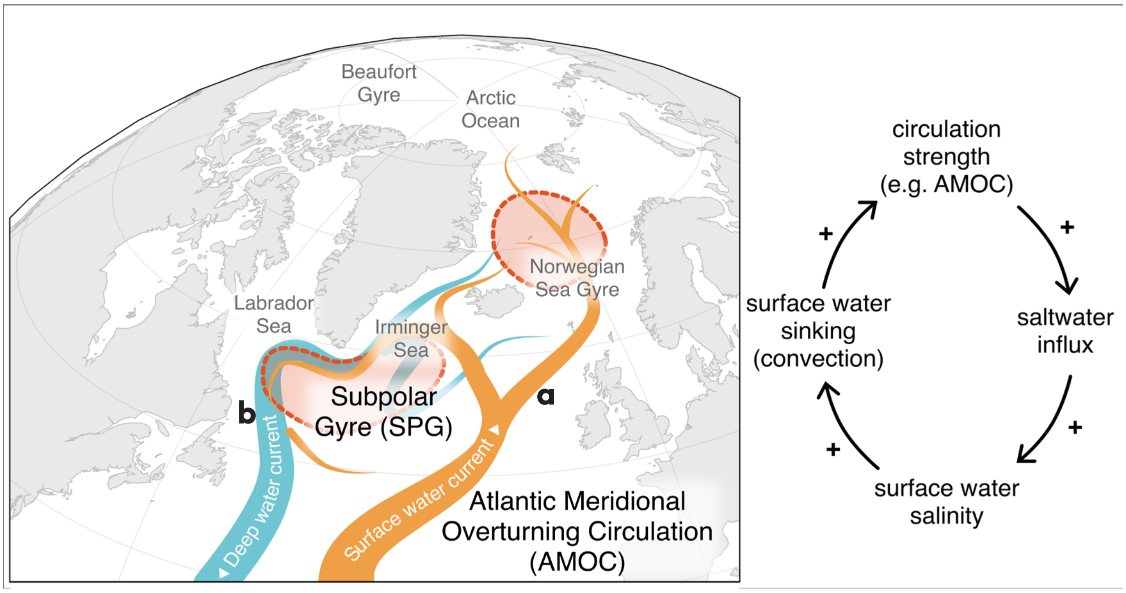

Figure 1.4.3: Overview over the major oceanic circulation systems in the North Atlantic. **a** The surface currents (orange pathways) are connected to deep ocean currents (blue) through sites where dense (cold, salty) water sinks, driving the overturning circulation (pink shading). **b** One critical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

The Atlantic Meridional Overturning Circulation

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

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

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

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

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

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

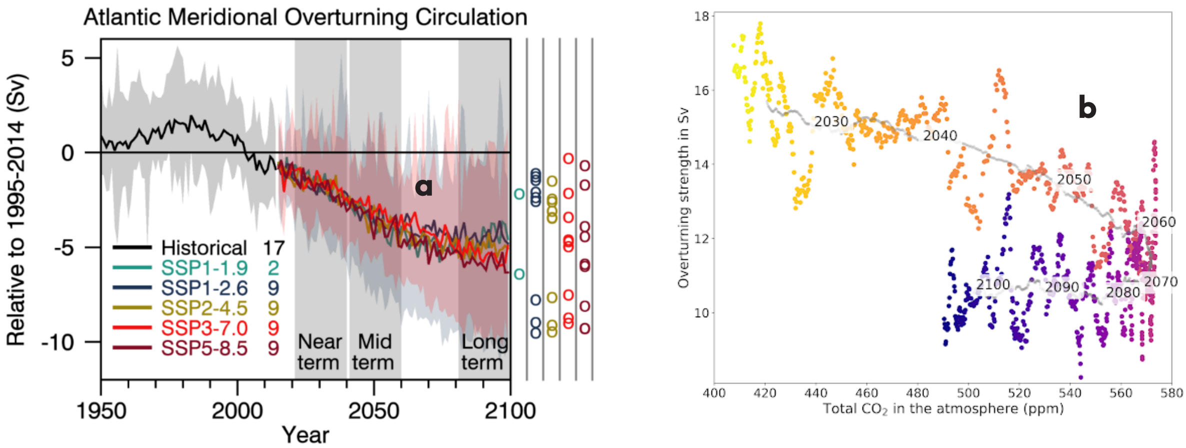

Figure 1.4.4: AMOC in CMIP models. **a** CMIP6 models showing gradual weakening of the AMOC during the 21st Century under all emission scenarios. Credit: Lee et al., 2021l . **b** CMIP6 overshoot experiments (using UKESM; Jones et al. 2020) showing hysteresis - different states of the AMOC (vertical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

Evidence for tipping dynamics

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

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

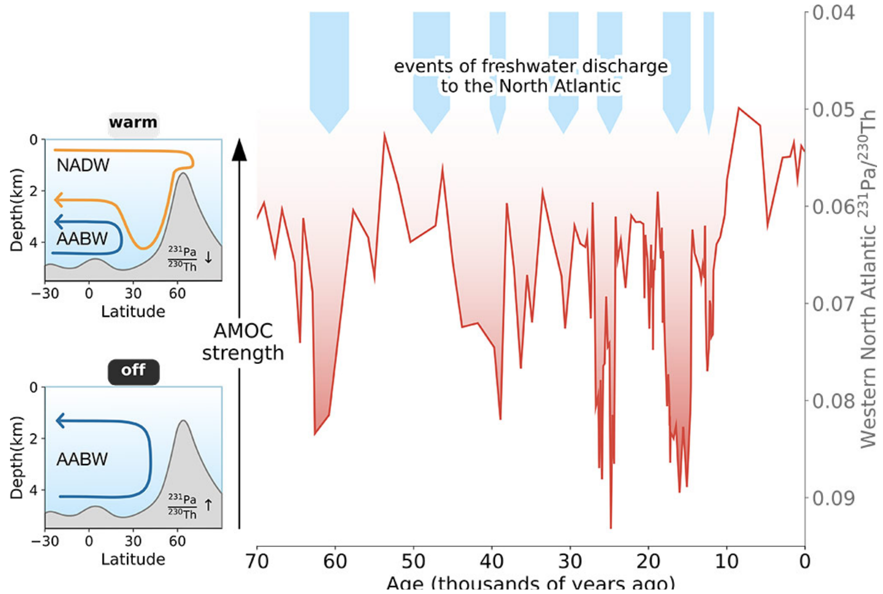

Figure 1.4.5: Different AMOC modes and palaeo-evidence.The diagrams on the left show two AMOC modes as indicated by sedimentary $^{231}Pa/^{230}Th$ in palaeorecords. NADW: North Atlantic Deep Water; AABW: Antarctic Bottom Water, adapted from Böhm et al., 2015. The timeline shows AMOC slowdown events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

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

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

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

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

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

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

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

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

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

Assessment and knowledge gaps

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

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

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

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

North Atlantic Subpolar Gyre (SPG)

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

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

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

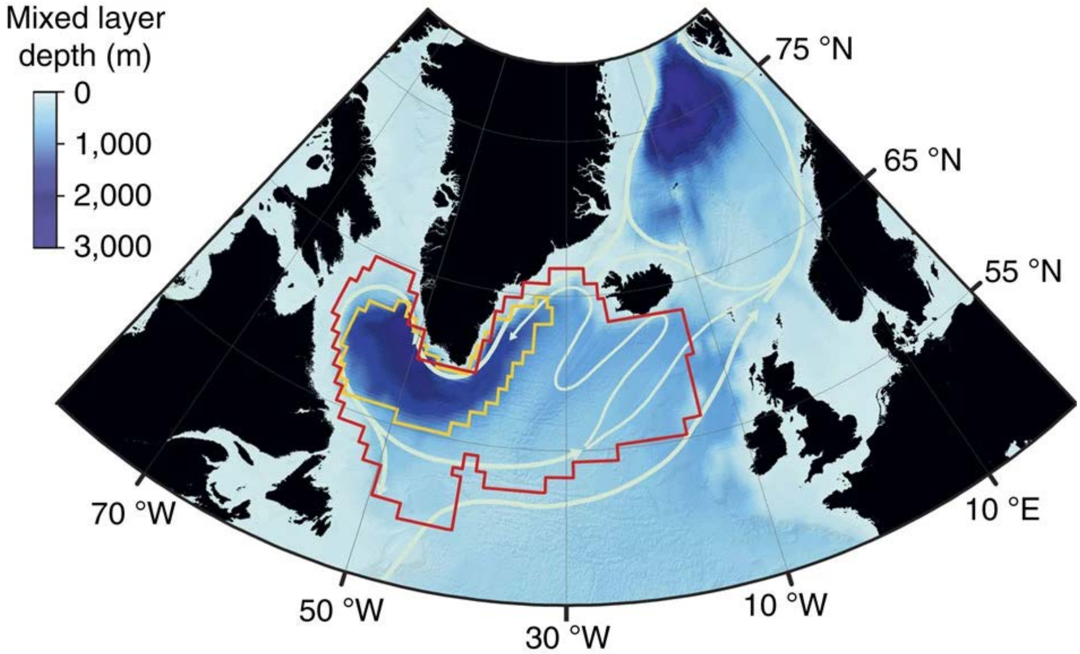

Figure 1.4.6: Map showing the maximum ocean mixing depth in the North Atlantic (light to dark blue), showing deep water convection sites driving the AMOC and SPG east and south of Greenland respectively (with the Labrador-Irminger Seas convection area bordered by yellow). The pale arrows show surface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

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

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

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

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

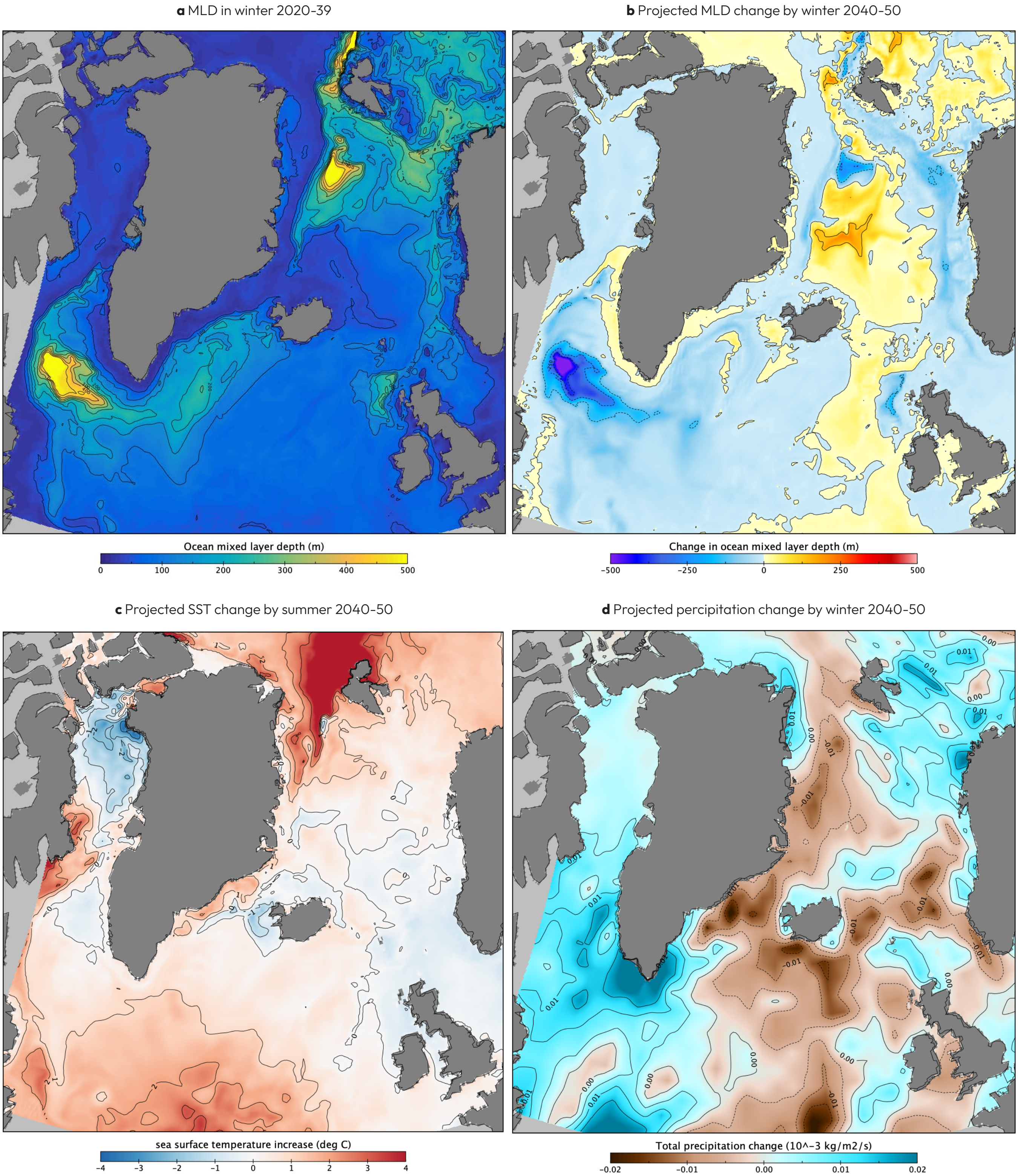

Figure 1.4.7: a Winter ocean mixed layer depth (MLD) as indicator of ocean convection in winter 2020-30 (January-March). **b** Changes in projected MLD by winter 2040-50. **c** Change in summer sea surface temperature (SST) and **d** winter total atmospheric precipitation, respectively, projected by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

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

Evidence for tipping dynamics

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

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

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

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

Assessment and knowledge gaps

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

1.4.2.2 Southern Ocean circulation

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

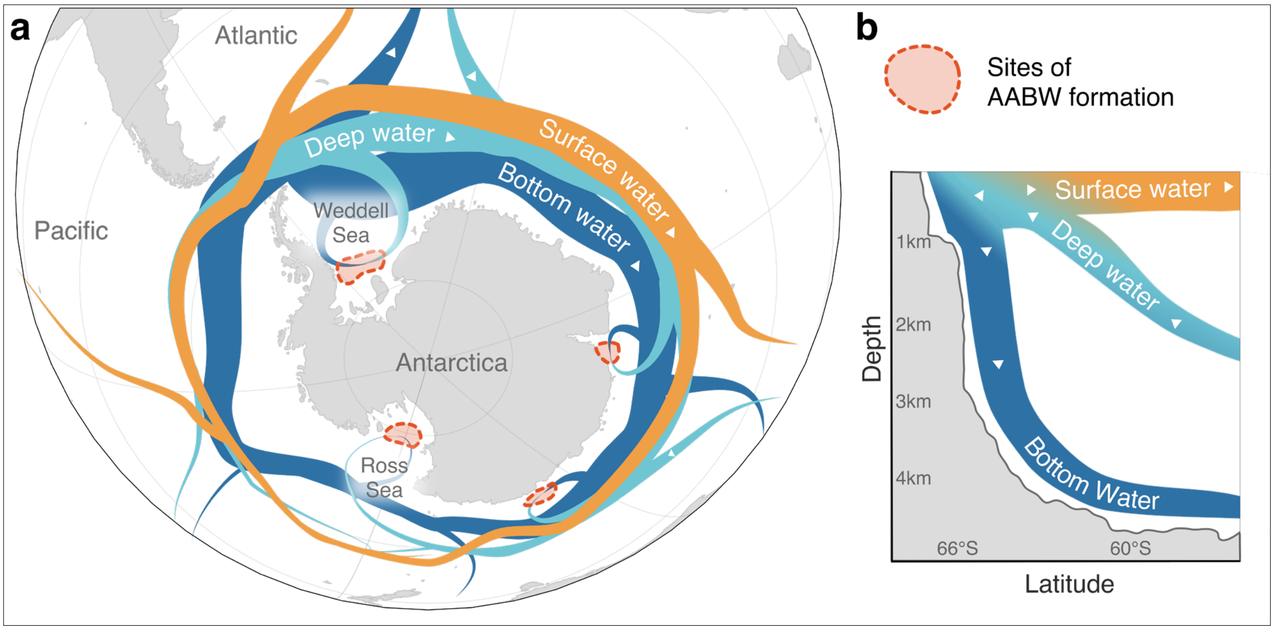

Figure 1.4.8: Circulations and potential tipping systems in the Southern Ocean. Adapted from Li, et al (2023) and IPCC SROCC Fig CB7.1

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

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

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

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

Evidence for tipping dynamics

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

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

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

Assessment and knowledge gaps

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

1.4.2.3 Monsoons

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

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

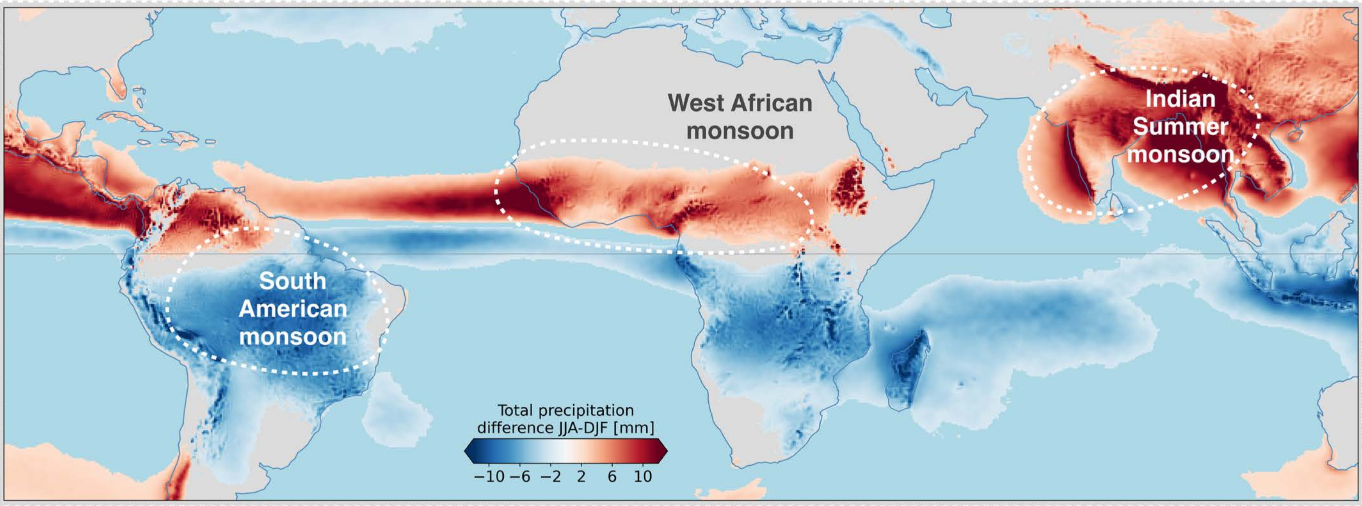

Figure 1.4.9: Monsoon systems. Shown is the total precipitation difference between Northern hemisphere summer (June-August, JJA) and winter months (December-February, DJF), highlighting the dominant precipitation patterns over South America (SAM), West Africa (WAM) and India (ISM). Generated using Copernicus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

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

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

Indian summer monsoon (ISM)

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

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

Evidence for tipping dynamics

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

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

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

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

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

Assessment and knowledge gaps

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

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

West African monsoon (WAM)

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

Evidence for tipping dynamics

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

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

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

Assessment and knowledge gaps

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

South American monsoon (SAM)

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

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

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

Evidence for tipping dynamics

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

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

Assessment and knowledge gaps

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

1.4.2.4 Tropical clouds, circulation and climate sensitivity

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

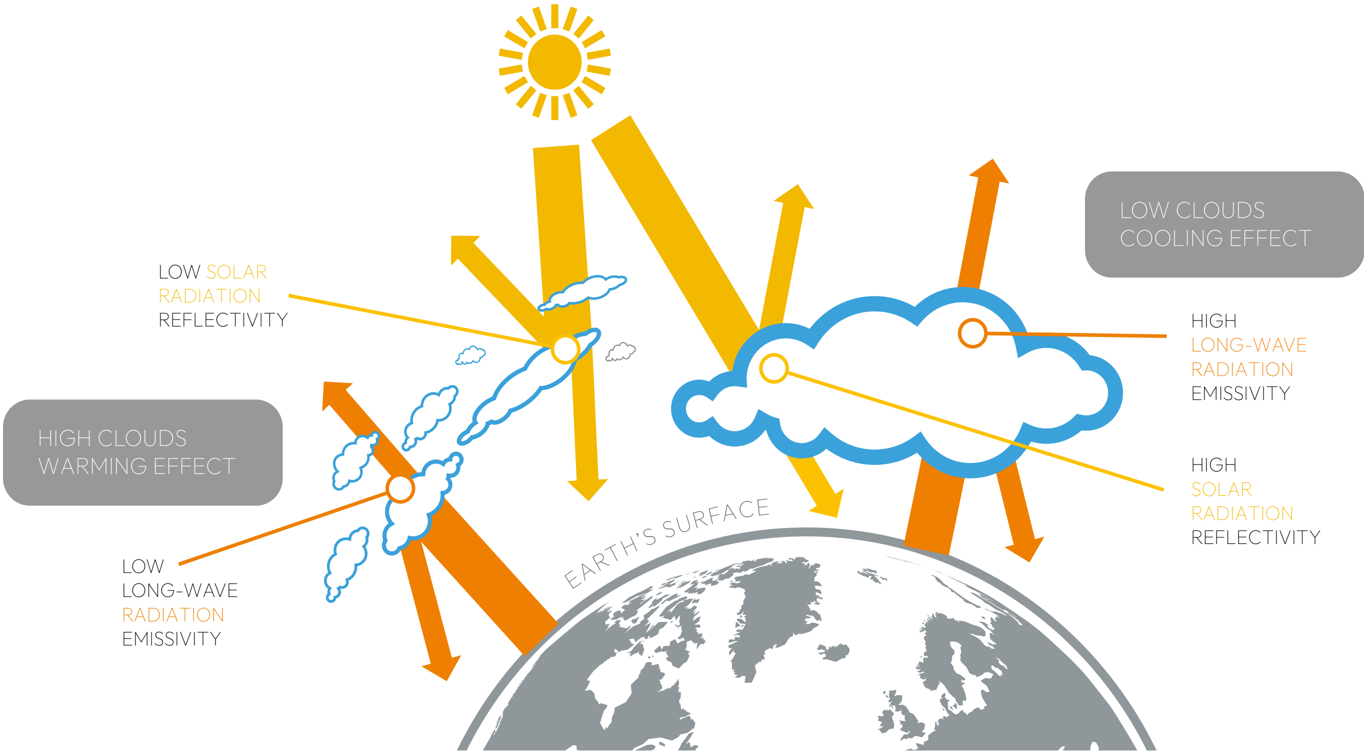

Figure 1.4.10: The role of clouds in regulating Earth's radiation budget.

Evidence for tipping dynamics

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

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

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

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

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

Assessment and knowledge gaps

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

1.4.2.5 El Niño-Southern Oscillation (ENSO)

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

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

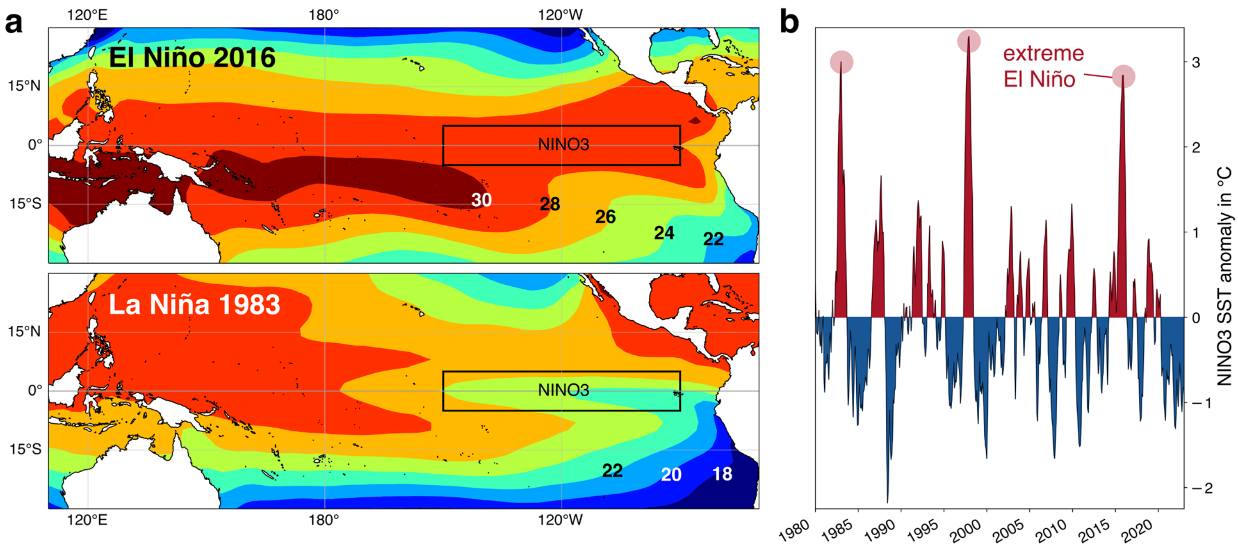

Figure 1.4.11: ENSO warm and cold phases and observational record. a Examples of strong El Niño (top) and La Niña (bottom) events seen in the tropical Pacific surface temperature (SST) distribution, with characteristic strong and weak SST gradient along the equator, respectively. **b** ENSO record since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

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

Evidence for tipping dynamics

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

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

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

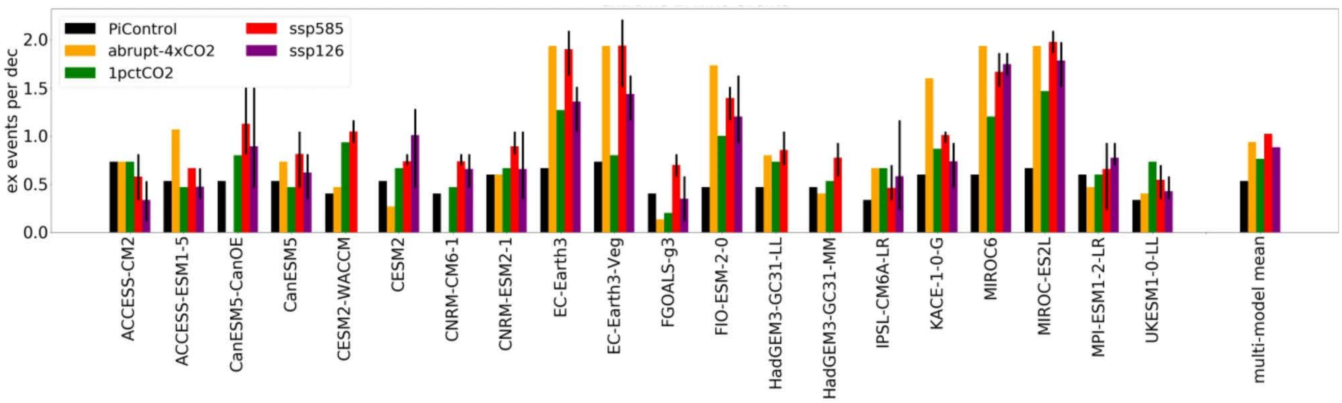

Figure 1.4.12: Overview of projected changes in extreme El Niño events in CMIP6 climate models. The bar chart shows the time-mean frequency of extreme El Niño events (the number of events per decade) for several idealised and more realistic global warming experiments (abrupt-4xCO2, 1pctCO2, SSP5-8.5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

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

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

Assessment and knowledge gaps

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

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

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

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

1.4.2.6 Mid-latitude atmospheric dynamics

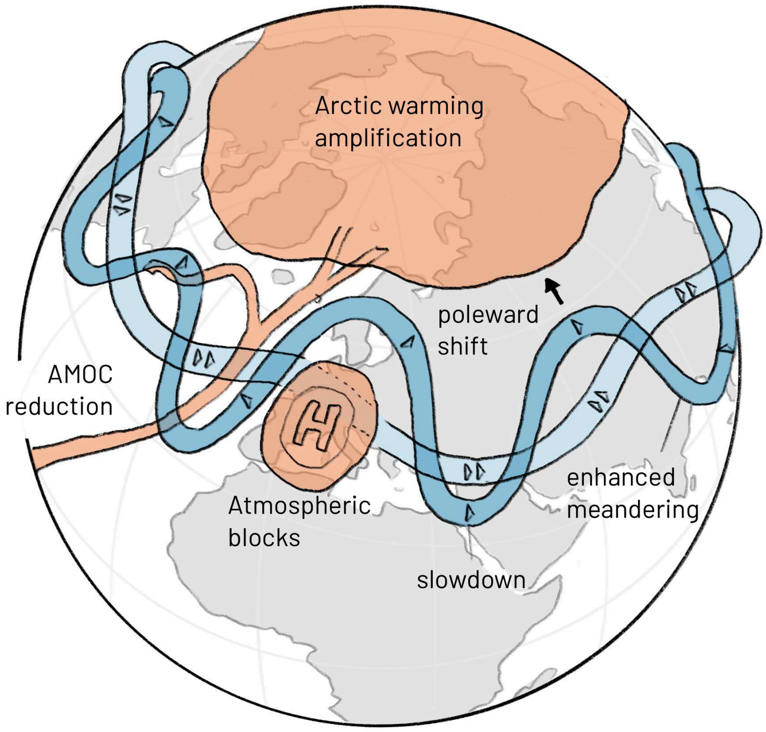

Figure 1.4.13: Potential changes in mid-latitude atmospheric circulations, exemplary for the Northern Hemisphere. Reduction of AMOC, atmospheric blocking events, Arctic warming and other drivers can modify the jet stream. Potential consequences are a northward shift, slowdown and enhanced meandering,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

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

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

Evidence for tipping dynamics

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

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

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

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

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

Assessment and knowledge gaps

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

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

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

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

Chapter 1.5 Climate tipping point interactions and cascades

Authors: Nico Wunderling, Anna von der Heydt, Yevgeny Aksenov, Stephen Barker, Robbin Bastiaansen, Victor Brovkin, Maura Brunetti, Victor Couplet, Thomas Kleinen, Caroline H. Lear, Johannes Lohmann, Rosa M. Roman-Cuesta, Sacha Sinet, Didier Swingedouw, Ricarda Winkelmann, Pallavi Anand, Jonathan Barichivich, Sebastian Bathiany, Mara Baudena, John T. Bruun, Cristiano M. Chiessi, Helen K. Coxall, David Docquier, Jonathan F. Donges, Swinda K. J. Falkena, Ann Kristin Klose, David Obura, Juan Rocha, Stefanie Rynders, Norman J. Steinert, Matteo Willeit

Summary

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

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

Key messages

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

Recommendations

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

1.5.1 Introduction and definition

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

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

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

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

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

1.5.2 Interactions between climate tipping systems and further nonlinear climate components

1.5.2.1 Interactions across scales in space and time

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

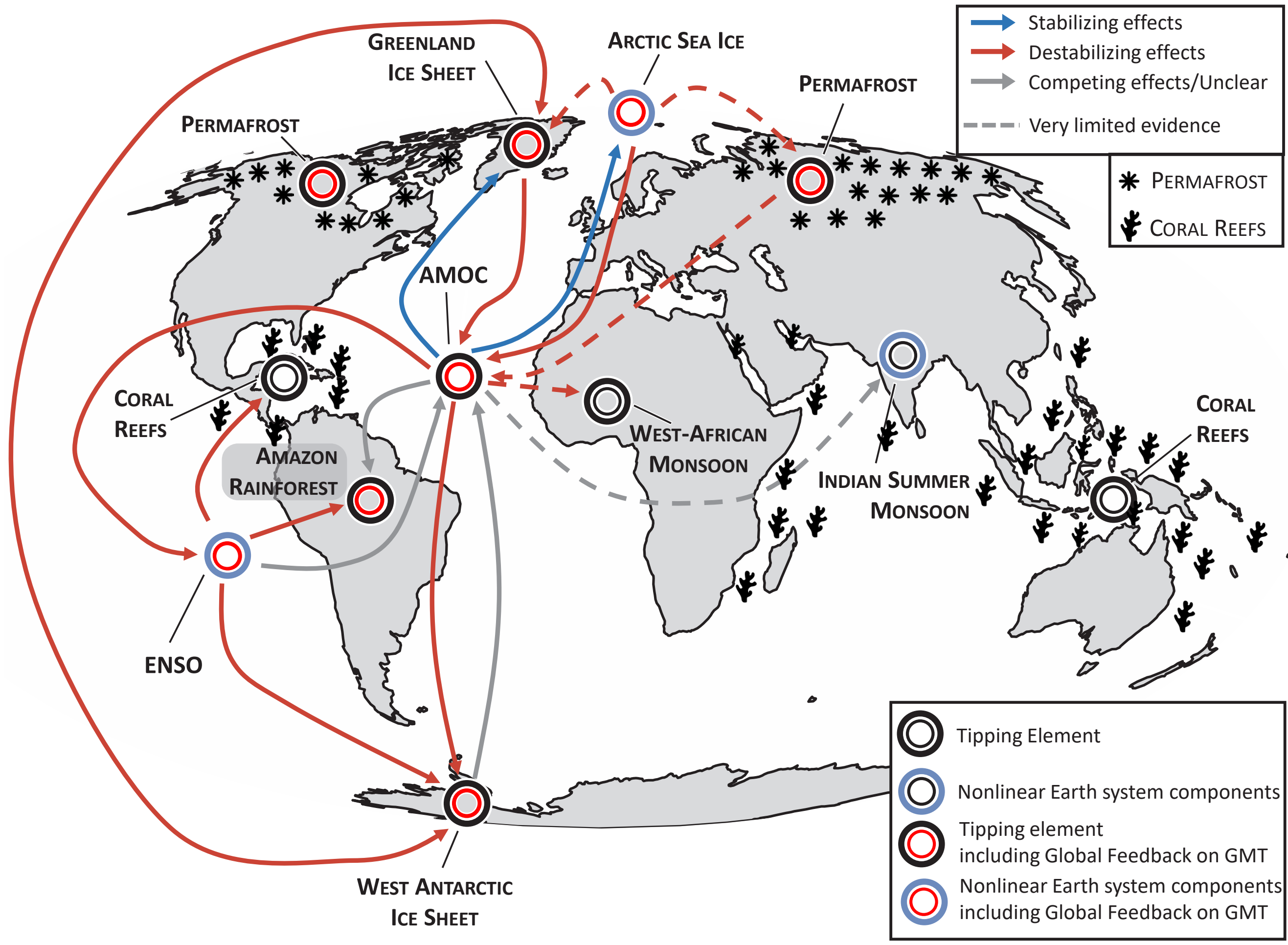

Figure 1.5.1: Interactions between established and more speculative tipping systems on a world map. All tipping systems discussed in this chapter are shown together with their potential connections. The causal interaction links can have stabilising (blue arrows), destabilising (red arrows), or unclear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

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

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

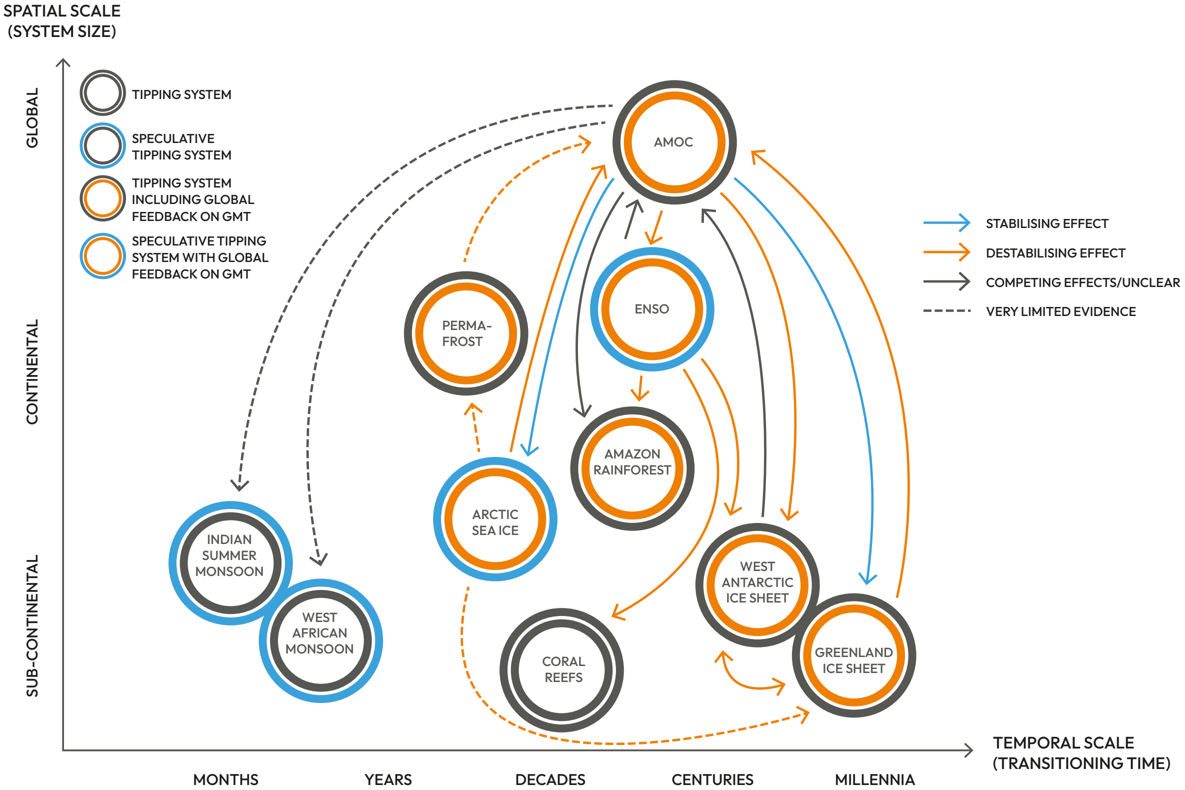

Figure 1.5.2: Interactions between tipping systems across scales in space and time. Temporal scales are transitioning times of a disintegrating tipping system from months up to millennia. Spatial scales denote the system size from sub-continental to (nearly) global scales. Transitioning times are taken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

1.5.2.2 Interactions between ice sheets and the AMOC

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

Greenland Ice Sheet to AMOC

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

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

West Antarctic Ice Sheet to AMOC

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

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

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

AMOC to ice sheets

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

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

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

1.5.2.3 Arctic sea ice interactions

Interactions between AMOC and Arctic sea ice

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

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

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

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

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

1.5.2.4 Effects of AMOC changes on the Amazon rainforest

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

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

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

1.5.2.5 Interactions between ENSO and tipping systems

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

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

Interactions between ENSO and AMOC

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

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

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

Influences of ENSO on the Amazon rainforest

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

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

Influences of ENSO on the WAIS

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

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

Influences of ENSO on warm-water coral reefs

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

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

Effects of AMOC and ENSO changes on tropical monsoon systems

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

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

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

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

1.5.2.6 Effects of permafrost thaw on the global hydrological cycle

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

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

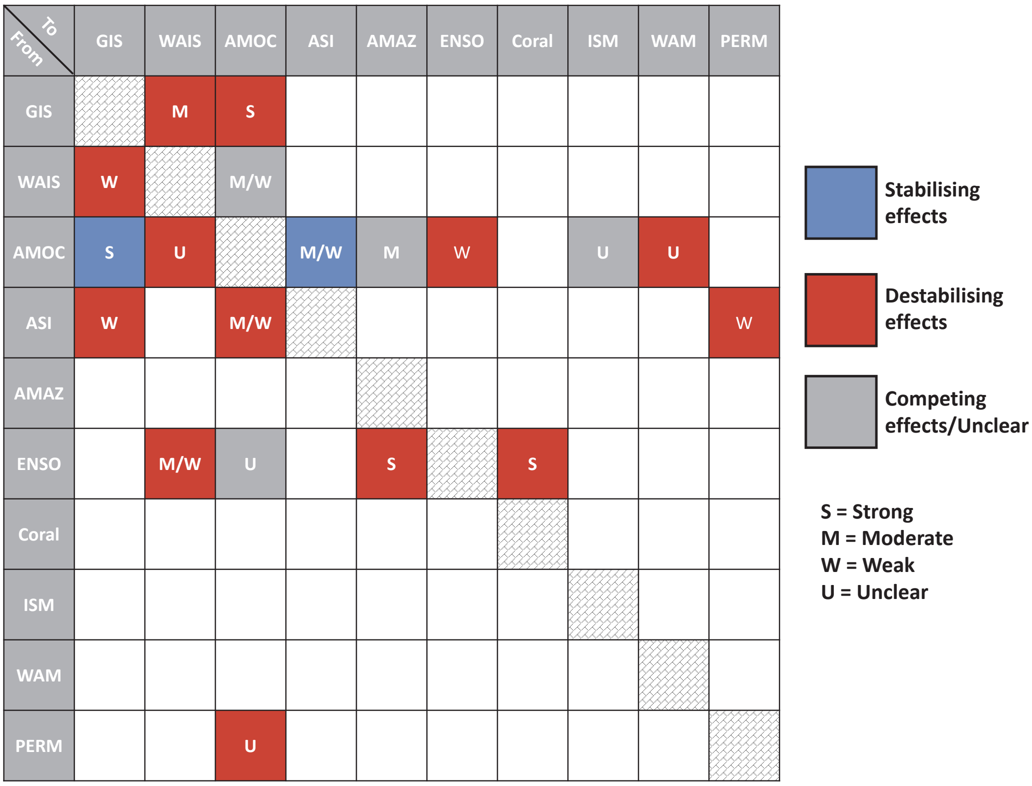

Figure 1.5.3: Matrix of links between elements (tipping systems and other nonlinear components) discussed in this chapter (see also Figs. 1 and 2). Columns denote the element from which the interaction originates, rows denote the tipping system to which element the interaction is pointing. We separate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

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

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

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

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

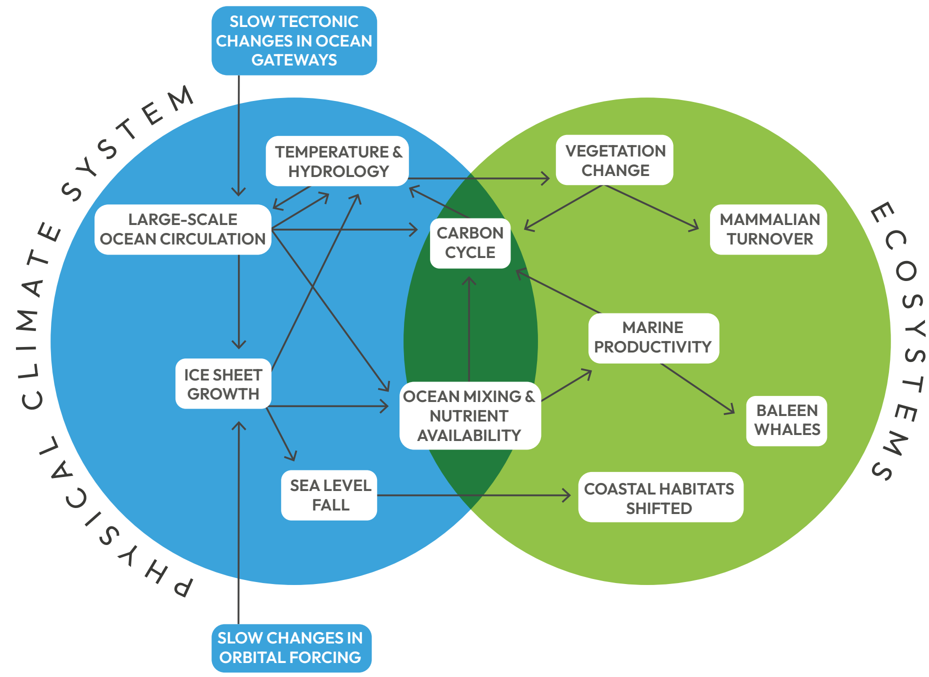

Figure 1.5.4: Conceptual linkages between changes in the Earth system associated with the Eocene-Oligocene Transition, 34 million years ago. External drivers were the slow changes in ocean gateways caused by tectonic plate movement, and slow changes in Earth's orbital configuration. The interactions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

Ocean circulation

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

Biosphere

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

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

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

1.5.3.2 Interactions during and since the last glacial period

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

Dansgaard-Oeschger events

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

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

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

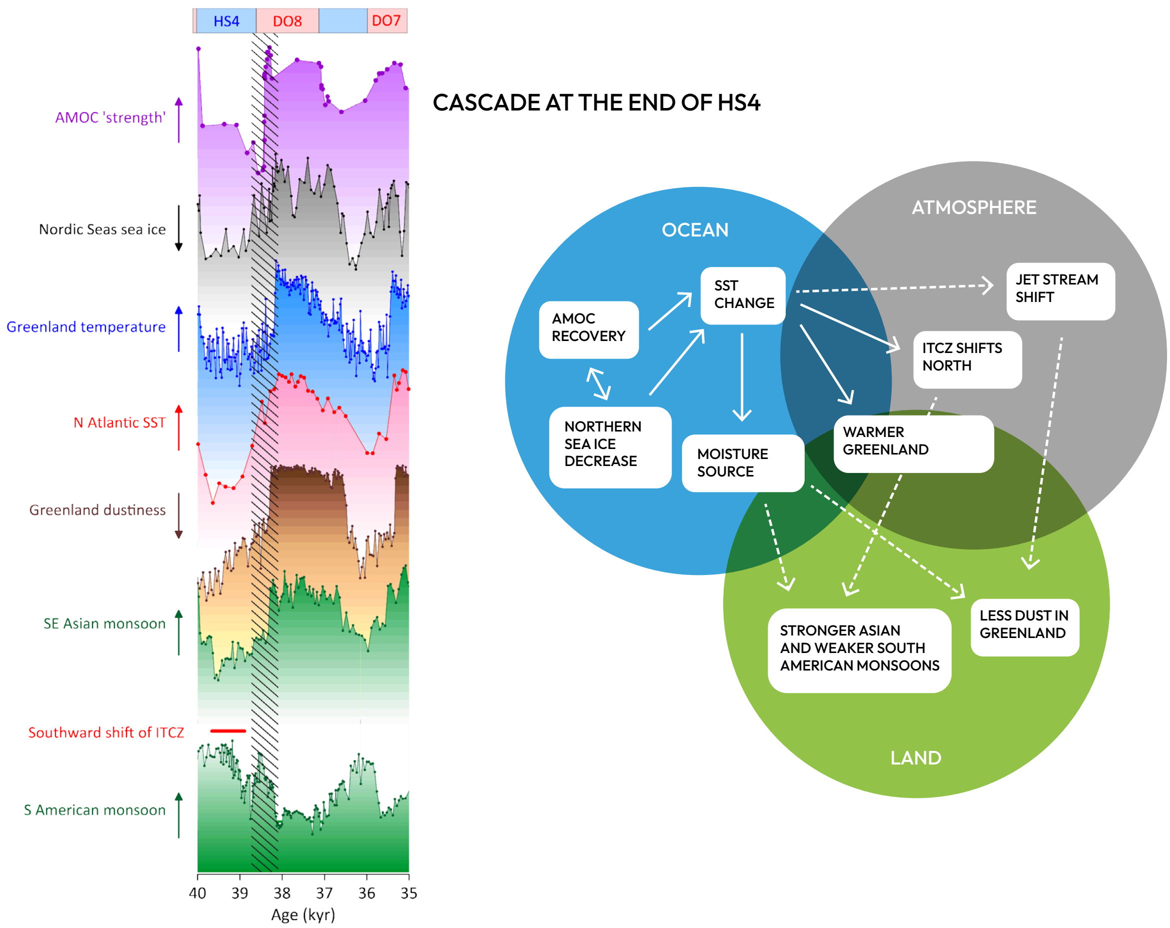

Figure 1.5.5: Interactions at the end of the Heinrich event 'Heinrich Stadial 4' (HS4). (a) Climate proxy indices spanning the transition from HS4 into Dansgaard-Oeschger (D/O) event 8 (time goes from left to right). From top to bottom: AMOC strength (Henry et al., 2016), Norwegian Sea ice cover (Sadatzki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

Bølling-Allerød

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

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

Heinrich events

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

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

1.5.4 Interactions between tipping systems and planetary-scale cascades

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

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

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

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

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

1.5.5 Final remarks

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

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

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

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

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

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

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

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

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

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