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.

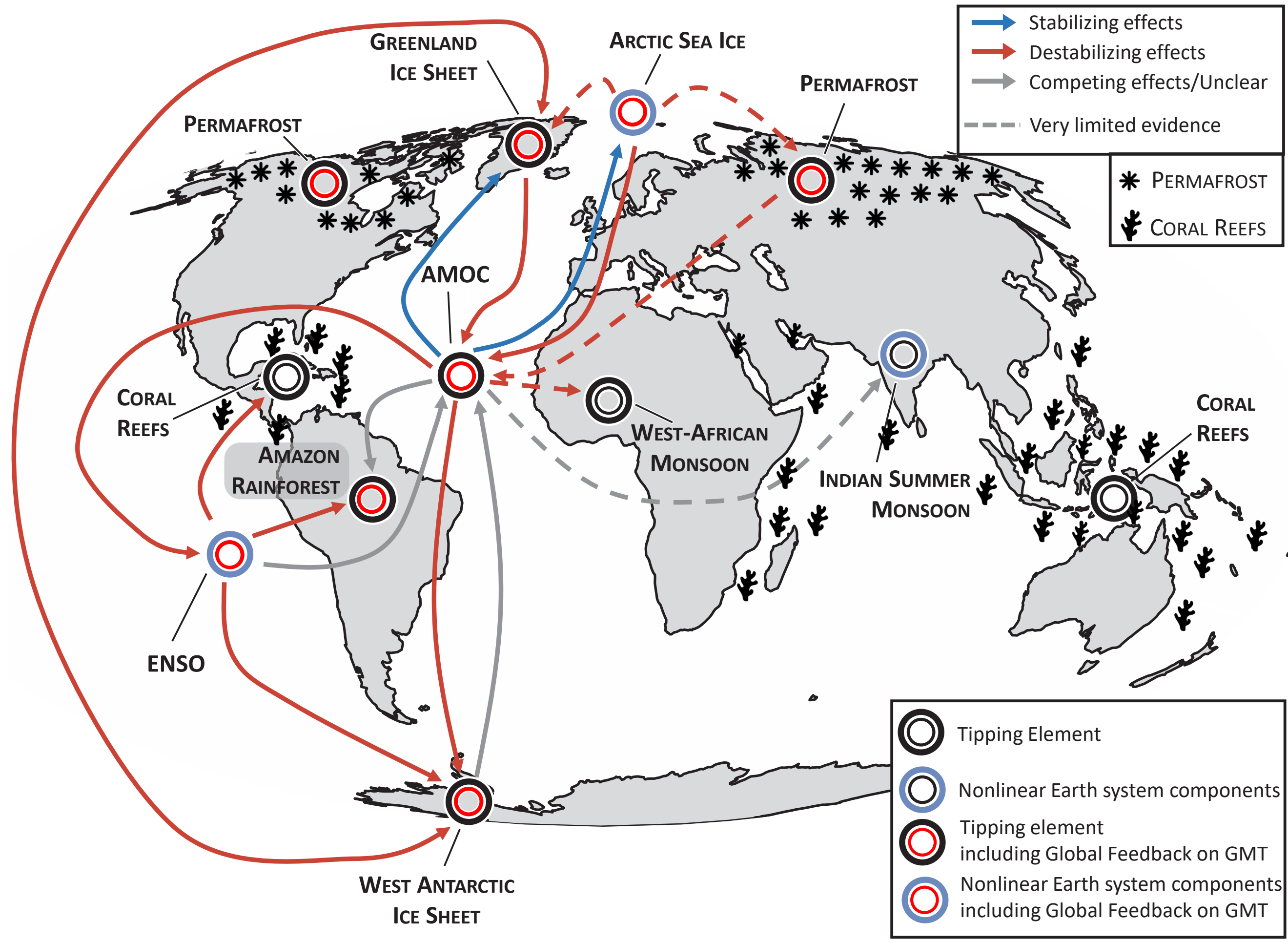

Figure 1.5.1: Interactions between established and more speculative tipping systems on a world map. All tipping systems discussed in this chapter are shown together with their potential connections. The causal interaction links can have stabilising (blue arrows), destabilising (red arrows), or unclear (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.

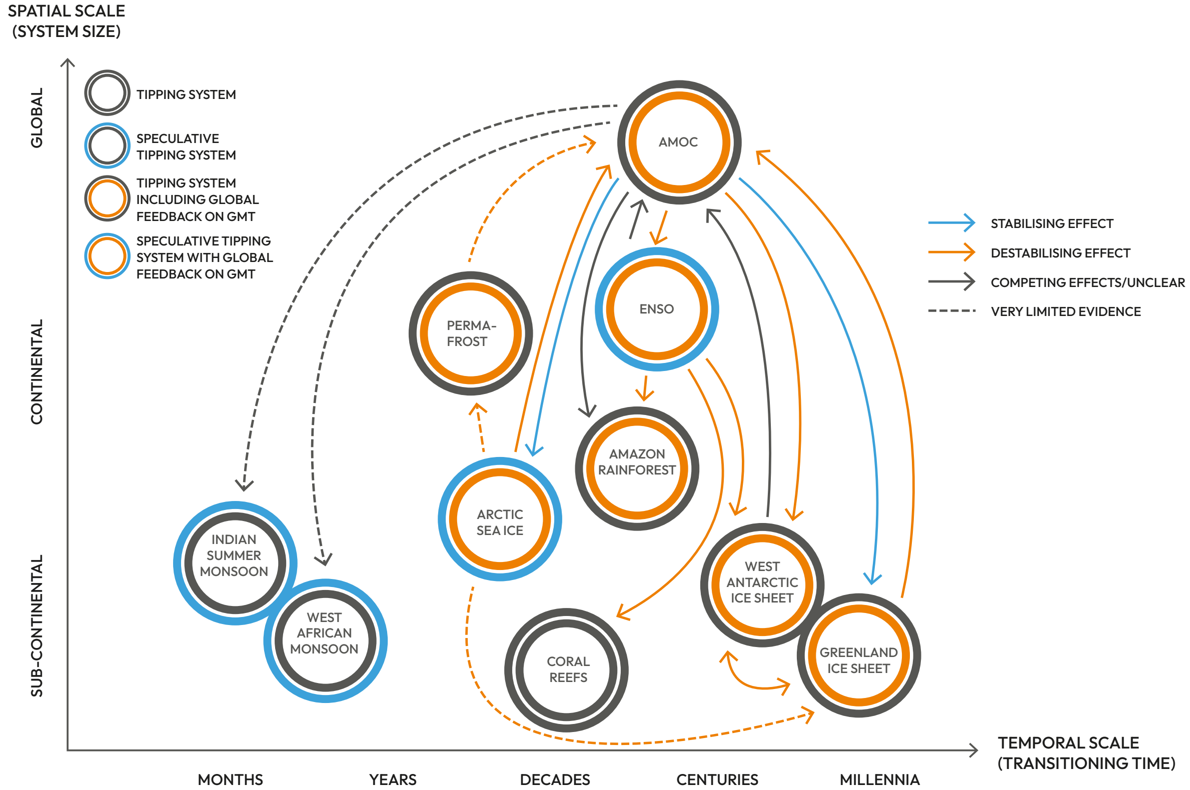

Figure 1.5.2: Interactions between tipping systems across scales in space and time. Temporal scales are transitioning times of a disintegrating tipping system from months up to millennia. Spatial scales denote the system size from sub-continental to (nearly) global scales. Transitioning times are taken 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.

From \ To	GIS	WAIS	AMOC	ASI	AMAZ	ENSO	Coral	ISM	WAM	PERM
GIS		M	S							
WAIS	W		M/W							
AMOC	S	U		M/W	M	W		U	U	
ASI	W		M/W							W
AMAZ										
ENSO		M/W	U		S		S			
Coral										
ISM										
WAM										
PERM			U							

Stabilising effects

Destabilising effects

Competing effects/Unclear

S = Strong
M = Moderate
W = Weak
U = Unclear

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.

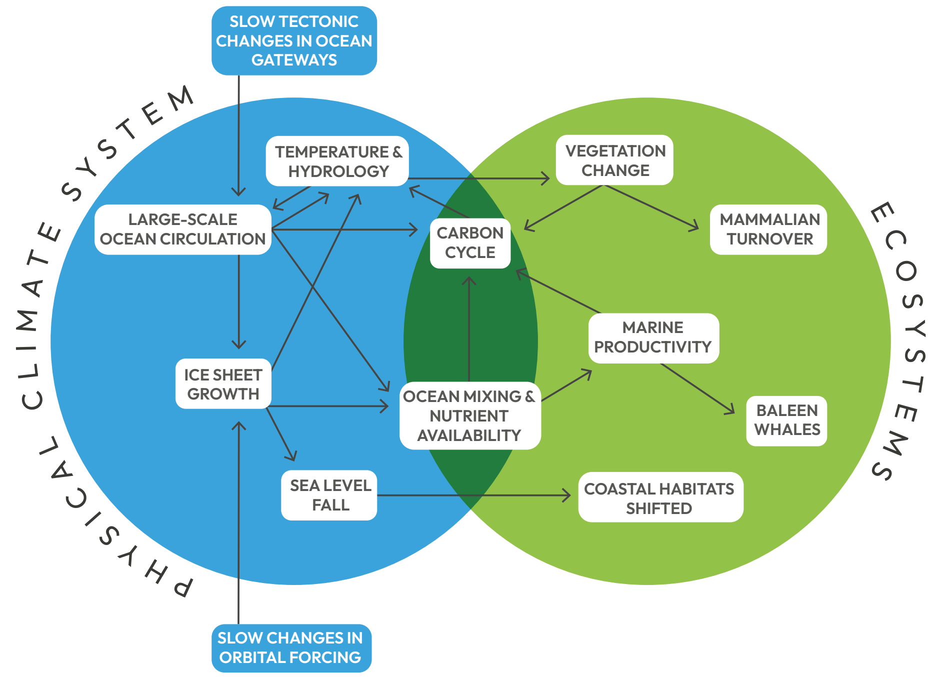

Figure 1.5.4: Conceptual linkages between changes in the Earth system associated with the Eocene-Oligocene Transition, 34 million years ago. External drivers were the slow changes in ocean gateways caused by tectonic plate movement, and slow changes in Earth's orbital configuration. The interactions 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.

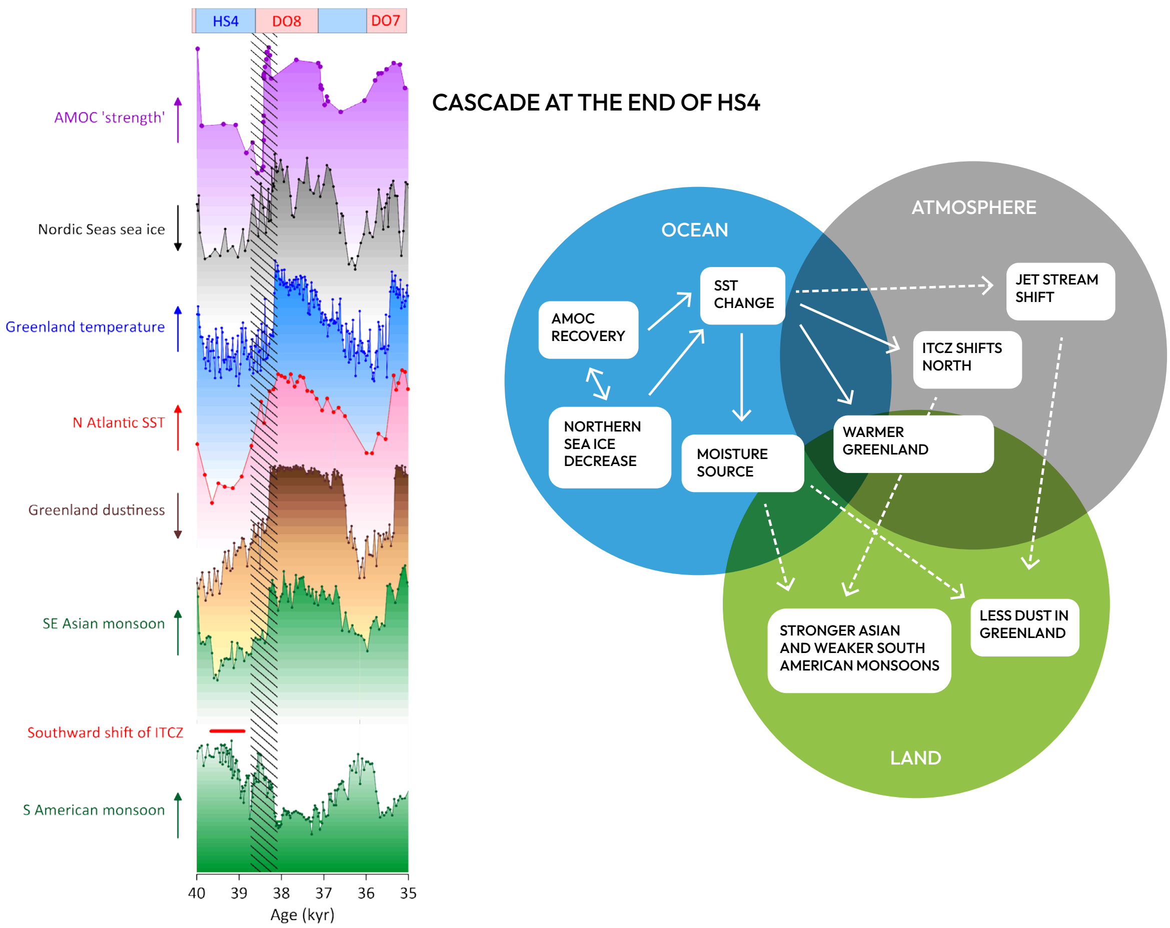

Figure 1.5.5: Interactions at the end of the Heinrich event 'Heinrich Stadial 4' (HS4). (a) Climate proxy indices spanning the transition from HS4 into Dansgaard-Oeschger (D/O) event 8 (time goes from left to right). From top to bottom: AMOC strength (Henry et al., 2016), Norwegian Sea ice cover (Sadatzki 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).